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ADC Robertson & GA Simpson




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SUMMARY

The Yarrol Project covered the area around Rockhampton in central coastal Queensland, and extended from Ridgelands to Monto. It mainly studied rocks of the Yarrol Province of the northern New England Orogen, bounded on the east by the Yarrol Fault and on the west by the Gogango Overfolded Zone and the Biloela Basin. Information on adjacent tectonic elements such as the Marlborough Block and Wandilla and Shoalwater Provinces was compiled from previous literature. The Yarrol Project joins with the South Connors-Auburn-Gogango Project to the north, west and south, and with the Bundaberg Project to the east.

The main challenge of the mapping project was to establish an easily recognisable, province-wide lithostratigraphic nomenclature. Earlier mapping was an extension of detailed mapping of restricted areas in university theses, and many subdivisions were based on biostratigraphy. Geological mapping under the Yarrol Project had the advantage of semi-detailed airborne radiometric and magnetic surveys, which outlined boundaries between stratigraphic units, some not obvious using other techniques. The mapping was supported by geochronological and biostratigraphic research listed in the Appendixes, and by about 430 whole rock chemical analyses.

The Princhester Serpentinite is the oldest rock unit in the area, having been originally formed as the mantle component of an ophiolite at an oceanic spreading centre in latest Proterozoic or earliest Cambrian time. Only its later history is related to the New England Orogen.

The tectonic evolution of the New England Orogen in central Queensland in the Late Silurian and Devonian continues to be the subject of debate. Traditionally, this has been regarded as a convergent plate margin related to a west-dipping subduction zone. The first point of controversy is whether the rocks formed in an island arc or on thin, juvenile crust of the continental margin. A second issue concerns the dip direction of the subduction zone if the island arc hypothesis is correct. The recent recognition of back-arc basin basalts has increased the possibility that the subduction zone dipped to the east. This summary is based on the interpretation (largely based on geochemical results) that the Upper Silurian and Devonian strata were deposited in an intra-oceanic island arc above an east-dipping subduction zone.

Fault blocks and inliers of Late Silurian to Middle Devonian age are distributed throughout the Yarrol Province. The chemistry of the volcanic and plutonic rocks in these outcrops is comparable with that of samples from modern island arcs. This, together with the submarine environment of deposition, favours an island arc setting. The possibility of earlier island arc development is raised by the occurrence of Ordovician limestones in a very small area within the Upper Silurian – Lower Devonian Calliope beds south of Gladstone. Each inlier or fault block has a different stratigraphic succession and volcanic geochemistry, and they probably represent remnants of more than one arc.

The geochemistry of volcanic and plutonic rocks in the Middle Devonian Capella Creek Group, forming a belt through Mount Morgan, is similar to that of modern primitive island arcs such as Tonga-Kermadec, implying the existence of a related back-arc basin. The uppermost Silurian to Lower Devonian Erebus and Calliope beds to the east of the Capella Creek Group can be explained as remnant arcs originally separated by a back-arc basin if the subduction zone dipped to the east. An east-dipping subduction zone also provides a simple mechanism for arc-continent collision. In the Mount Morgan region, felsic magmatism, both volcanic and plutonic, persisted over about 15 million years from the Middle to Late Devonian. The world class Mount Morgan gold-copper ore body was a VHMS deposit related to this magmatism. It has many features in common with a modern VHMS forming in a large submarine silicic caldera of the Izu-Bonin arc. The search for more VHMS deposits in the Upper Silurian – Middle Devonian rocks of the Yarrol Province should be concentrated on long-lived felsic volcanic centres.

An unconformity between the Upper Silurian – Middle Devonian sequences and younger strata, expressed as an angular unconformity near Mount Morgan and by an extended time break elsewhere, has long been considered to mark the timing of the arc-continent collision. Upper Devonian rocks that unconformably overlie the Middle Devonian Capella Creek Group were traditionally regarded as the earliest part of a transitional arc – forearc basin assemblage that formed along the continental margin above a west-dipping subduction zone, a tectonic setting that persisted until the end of Carboniferous time. However, detailed geochemical studies have shown that Upper Devonian basalts in the Hoopbound and Lochenbar Formations and Three Moon Conglomerate between Mount Morgan and Monto are typical of mature or evolved island arcs such as the Lesser Antilles, Marianas and Aleutians. As well, MORB-like back-arc basin basalts form a minor but important component of the Three Moon Conglomerate, located east of the Lochenbar Formation, and back-arc basin basalts occur along the west side of the Yarrol Fault east of Monto. The geological and geochemical evidence indicates that: (1) these Upper Devonian units represent mature island arc deposits that unconformably overlie the more primitive Capella Creek Group; and (2) the basalt compositions were related to or inherited from the same east-dipping subduction zone that produced the Middle Devonian and older rocks.

Intrusions and included blocks of island arc tholeiite composition within the Princhester Serpentinite which have given a Late Devonian age are linked to the basalts between Mount Morgan and Monto by their similar geochemistry, suggesting a common source and process in magma generation. The chemistry of the Princhester Serpentinite, particularly of the rare earth elements, can be explained by reaction between depleted peridotite (originally formed at a latest Proterozoic or earliest Cambrian oceanic spreading centre), and these magmas. The serpentinite is therefore the

mantle component of a Late Devonian supra-subduction zone ophiolite formed at a spreading centre above the east-dipping subduction zone.

The most widespread Upper Devonian unit, the Mount Alma Formation, was deposited in an intra-oceanic back-arc basin to the east of an active arc which sourced most of the volcaniclastic sediments. Remnant arcs within and along the eastern margin of the back-arc basin were local sources of sedimentary detritus, including large allochthonous limestone blocks. This pattern of deposition continued into latest Devonian (Famennian) time.

An inevitable result of the arc-continent collision in the Late Devonian was the reversal of subduction direction and formation of a new west-dipping subduction zone beneath the continental margin. This pattern of west-dipping subduction continued through all of the Carboniferous, and is recorded in parallel belts comprising an eastern accretionary wedge, a central forearc basin and remnants of a western continental margin arc that form the basic tectonic elements of the New England Orogen. Volcanism became more felsic, and the radiometric signature of volcaniclastic sediments increased markedly. As the forearc basin became established, the Devonian arcs were eroded, and they subsided to the point where they no longer influenced depositional patterns. The Lower Carboniferous Rockhampton Group consists mainly of volcaniclastic sediments derived from the western arc, but is characterised by the repeated development of oolitic carbonate banks along the western margin which supplied ooids to deeper water sandstones to the east. By the time the overlying Upper Carboniferous to lowermost Permian Lorray Formation was deposited, the volcanic arc was less active and more deeply eroded, so that the volcaniclastic component declined and the plutonic component increased, with another sharp increase in the radiometric signature.

To the east, an accretionary wedge assemblage developed throughout the Carboniferous, with strong Early Carboniferous provenance linkages to the Rockhampton Group. The bulk of the accretionary wedge is composed of volcaniclastic sandstones with a small but persistent component of calcareous ooids which were derived from the forearc basin to the west. Dating of detrital zircons shows that the volcaniclastic component of these rocks was sourced almost exclusively from the Carboniferous continental margin arc. The easternmost accretionary wedge rocks are interbedded quartose sandstones and thin dark grey shale layers, marking a distinct provenance switch. Although the Carboniferous volcanic arc continued to be a source, the majority of material came from cratonic areas of inland and northern Queensland.

The convergent cycle ended about the end of the Carboniferous, when a period of extensional tectonics was established. Over much of the area, older basement rocks were eroded and tilted to varying degrees, so that the contact with the Early to mid-Permian extensional package ranges from a disconformity to an angular unconformity. Extension was mild, and subsidence was neither extreme nor rapid. The Youlambie Conglomerate, the oldest unit of the extensional package, is characterised by conglomerate with clasts of granite and rhyolite. The Berserker Group and Rookwood Volcanics are also assigned to this tectonic setting, indicating that extension continued to mid-Permian time. The chemistry of felsic volcanics in the Berserker Group, as well as the presence of classic Kuroko-style mineralisation at Mount Chalmers, is indicative of a back-arc setting. The Rookwood Volcanics have typical oceanic back-arc chemistry, which indicates the absence of sub-continental lithosphere, and is consistent with the hypothesis that the Carboniferous forearc sequence was deposited on oceanic crust. The Kyle Mohr Igneous Complex is a bimodal granite-gabbro intrusion that was probably co-magmatic with the Rookwood Volcanics, and the Miriam Vale Granodiorite and Castletower Granite may be of the same age.

The position of the volcanic arc to the east of the Berserker Group and Rookwood Volcanics may be marked by the coeval Owl Gully Volcanics east and north of Monto, which have a strong subduction signature consistent with eruption in a continental margin arc or a mature island arc with a maximum crustal thickness of ~30km.

The peak of metamorphism in the accretionary wedge east of the Yarrol Fault may be of mid-Permian age, based on dating of metamorphic overgrowths on zircon from the higher-grade parts of the assemblage, and supported by preliminary dating of the Castletower Granite. This raises the possibility that flat to gently-dipping structures in the accretionary wedge were related to the mid-Permian extensional event rather than to the Hunter–Bowen Orogeny.

The extensional phase was followed by a brief sag phase represented only by the lower part of the Moah Creek beds in the Yarrol Project area.

The Hunter–Bowen Orogeny was a protracted mid-Late Permian to Middle Triassic event (265–230Ma) during which west-directed thrusts deformed the New England Orogen and the Bowen Basin to the west. Deformation was strongly partitioned from east to west, with the result that the orogeny had very different effects in different areas.

In the accretionary wedge, the only structures definitely produced by the Hunter–Bowen Orogeny are overprinting broad folds with variably developed steeply dipping axial plane cleavage.

The Yarrol Province is a thick-skinned fold-thrust belt with a detachment zone at a depth of several kilometres or more. It behaved as a relatively rigid crustal block that moved westward to cause the deformation in the Gogango Overfolded Zone and the Folded Zone of the Bowen Basin. Intensity of deformation was controlled by proximity to major faults and by lithology. Thrusting is most obvious along the western boundary with the Gogango Overfolded Zone. The age of the main folding event was Late Permian. Three later, out-of-sequence thrusts have been recognised, the main one forming the base of the Marlborough Block.

The Gogango Overfolded Zone is a stack of imbricate thrust sheets developed in rocks of the Bowen Basin immediately west of the Yarrol Province, which provides evidence for a high degree of contractional shortening. Bedding dips to the east at ~30°, and the cleavage, best seen in siltstones interbedded with sandstones, is steeper, giving a consistent vergence which indicates east over west transport. The age of thrusting and folding was probably the same as that in the Yarrol Province.

The Hunter–Bowen Orogeny was active throughout and was the main cause of development of the Bowen Basin as a major foreland basin. The rising thrust sheets in and adjacent to the New England Orogen became the dominant source of sediments for the basin, commencing in mid-Permian time and continuing until deposition ceased in the Middle Triassic. The elongate Middle Triassic Abercorn Trough, situated beneath the Mulgildie Basin west of Monto, may also have been a foreland basin, although it has been interpreted as a rift basin related to a tensional phase.

Extensive magmatism in Late Permian to Late Triassic time was largely coincident with the Hunter–Bowen Orogeny, and can therefore be regarded as syntectonic, although some plutons cross-cut folded rocks. Five small batholiths and several composite plutons are present. Granodiorite is the dominant composition, but a full range from gabbros (including layered gabbros) to syenogranites is present. Geochemically, the intrusives belong to the Clarence River Supersuite of the New England Batholith. The peak of intrusive activity was close to the Permian–Triassic boundary. The early magmatism was related to active subduction, with a change to extensional character and A-type chemistry in the Late Triassic. Mineralisation associated with the intrusions includes gold-bearing veins and porphyry-style copper–gold–molybdenum, but deposits are small and currently uneconomic.

Continental volcanics are exclusively of Triassic age, are flat-lying, and post-date the main folding event associated with the Hunter–Bowen Orogeny. Most show distinct subduction-related chemistry and can be matched with volcanics from the Southern Volcanic Zone of the Andes. Late Triassic volcanics are dominantly felsic, and show a transition from subduction-related to extensional character, similar to that displayed by the intrusions.

The Upper Triassic Callide Coal Measures represent the preserved fill of a half-graben controlled by faults along its western margin. Palaeocurrent data indicate a consistent south to south-eastward direction of sediment transport throughout the history of the basin.

In earliest Jurassic time, the extensive but thin sand sheet of the Precipice Sandstone was deposited by a series of braided streams that flowed from south to north. Much of the quartzose sand was derived from the Permo-Triassic granites. The younger strata of the Mulgildie Basin were deposited in fluviolacustrine environments, and the succession includes oolitic ironstones and the mid-Jurassic Mulgildie Coal Measures. The Mulgildie Basin was undoubtedly more extensive to the east, because its present eastern boundary is a major fault, the Mulgildie Fault, and the sequence thickens to the east. Only the northern end of this basin extends into the Yarrol project area, which is difficult to reconcile with the south to north streams that laid down the Precipice Sandstone.

Magmatic activity resumed close to the Jurassic–Cretaceous boundary, when a relatively small alkaline complex, including nepheline syenite, was emplaced between Calliope and Monto. Soon after, a marine incursion reached the Stanwell area in earliest Cretaceous (Neocomian) time. Only a thin, steeply-dipping sequence is preserved south of the Stanwell Fault, and there is no evidence for the existence of a substantial basin. This marine incursion was followed by extensive Early Cretaceous (Aptian, ~120Ma) within-plate volcanism in the same area. Initial trachytic to rhyolitic volcanism, comprising both extrusives and intrusive plugs, was succeeded by widespread basalt flows, and finally by a resumption of large scale eruption of trachytic and rhyolitic volcanics (Mount Salmon Volcanics). Still later in the Early Cretaceous (Albian), another marine incursion is recorded in the Stanwell Basin and in the Jim Crow Basin north-east of Rockhampton. The Stanwell Basin was deltaic, and may have marked the western limit of the marine incursion. No subsequent deposition is recorded in the Yarrol project area until the Eocene. Widespread normal faults in the Yarrol project area are believed to be at mainly of Late Cretaceous age.

Several lacustrine oil shale basins developed during Eocene time. They are relatively deep considering their small size. Most formed as half-grabens with a significant normal fault parallel to basement structural trends on the western side. Their formation has been related to spreading in the Tasman and Coral Seas, but the basins post-date sea-floor spreading. A younger limit for the age of the basin fill is provided by upper Oligocene basalt which overlies the Biloela Basin. The Tertiary basins contain very large resources of oil shale.

Within-plate continental basalts were erupted in the southern part of the Yarrol Project area in late Oligocene time (~25Ma). All of the basalt flows show normal magnetism; in contrast, related plugs have both normal and reversed magnetism, consistent with the rapid reversals during this time period.

Recent Cainozoic deposits accumulated in an erosional environment, and reflect the present topography. They are mainly restricted to narrow zones along streams, at the base of slopes, and along the coastline.

Keywords: Regional geology, stratigraphy, plutonic rocks, volcanic rocks, sedimentary rocks, structural geology, geochemistry, tectonics, New England Orogen, Yarrol Province, Wandilla Province, Shoalwater Province, Gogango Overfolded Zone, Marlborough Block, Marlborough Metamorphics, Craigilee beds, Calliope beds, Erebus beds, Capella Creek Group, Mount Dick beds, Mount Warner Volcanics, Raspberry Creek Formation, Ginger Creek Member, Marble Waterhole beds, Channer Creek beds, Mount Hoopbound Formation, Lochenbar Formation, Balaclava

Formation, Three Moon Conglomerate, Mount Alma Formation, Rockhampton Group, Curtis Island Group, Doonside Formation, Balnagowan Volcanic Member, Wandilla Formation, Shoalwater Formation, Lorray Formation, Youlambie Conglomerate, Yarrol Formation, Owl Gully Volcanics, Smoky beds, Rookwood Volcanics, Berserker Group, Lakes Creek Formation, Chalmers Formation, Sleipner Member, Ellrott Rhyolite, Warminster Formation, Moah Creek beds, Dinner Creek Conglomerate, Inverness Volcanics, Cynthia beds, Native Cat Andesite, Muncon Volcanics, Bobby Volcanics, Coulston Volcanics, Winterbourne Volcanics, Dooloo Tops Volcanics, Callide Coal Measures, Bushley Dacite, Annie Creek beds, Precipice Sandstone, Evergreen Formation, Hutton Sandstone, Mulgildie Coal Measures, Wycarbah Volcanics, Dalma Basalt, Mount Salmon Volcanics, Stanwell Formation, Jim Crow Basin, Alton Downs Basalt, The Narrows Group, Casuarina beds, Yaamba beds, Nagoorin beds, Biloela Formation, Rossmoya beds, Princhester Serpentinite, Mount Morgan Trondhjemite, Pomegranate Tonalite, Kyle Mohr Igneous Complex, Miriam Vale Granodiorite, Castletower Granite, Wattlebank Granodiorite, Bouldercombe Igneous Complex, Flaggy Quartz Monzodiorite, Umbrella Granodiorite, Moonkan Granite, Gavial Gabbro, Gracemere Gabbro, Quarry Gabbro, Bundaleer Tonalite, Moonmera Porphyritic Granodiorite, Kabra Quartz Monzodiorite, Craiglands Quartz Monzodiorite, Mount Seaview Igneous Complex, Galloway Plains Igneous Complex, Bocoolima Granodiorite, Sawnee Gabbro, Rocky Point Granodiorite, Dumgree Tonalite, Wyalla Granite, Redshirt Granite, Voewood Granite, Ridgelands Granodiorite, Bajool Quartz Diorite, Cecilwood Quartz Diorite, Glassford Igneous Complex, Monal Granodiorite, Littlemore Granodiorite, Lawyer Granite, Rule Gabbro, Robert Granite, Deception Quartz Monzonite, Munholme Quartz Diorite, Bororen Tonalite, Molangul Granite, Norton Tonalite, Pack Granite, Lookerbie Igneous Complex, Mannersley Quartz Microdiorite, Zig Zag Tonalite, Targinie Quartz Monzonite, Riverston Granodiorite, Diglum Granodiorite, Eulogie Park Gabbro, Kariboe Gabbro, Bucknalla Gabbro, Warrong Teschenite, Glassford Igneous Complex, Radley Nephelinee Syenite, Burns Spur Nephaline Monzosyenite, Tollbar Breccia, Judas Tracybasalt, Goondicum Gabbro, Mount Hedlow Trachyte, Palaeozoic, Ordovician, Silurian, Devonian, Carboniferous, Permian, Triassic, Jurassic, Cretaceous, Tertiary, Quaternary, Pleistocene, Holocene, Rookwood, Ridgelands, Rockhampton, Duaringa, Mount Morgan, Bajool, Gladstone, Banana, Biloela, Calliope, Scoria, Monto, SF55-16, SF56-13, SG56-1, 8850, 8851, 8949, 8950, 8951, 9048, 9049, 9050, 9051, 9148, 9149, 9150.

INTRODUCTION

The location of the Yarrol Project area is shown in Figure 1. The area roughly centres on Mount Morgan and extends from Ridgelands in the north to Monto in the south, and from the Biloela Basin and the Gogango Overfolded Zone in the west to the Yarrol Fault in the east, though it has been extended to the present coastline in the north. The rocks of the Curtis Island Group were investigated only cursorily during this project, and most of the information on them is from existing maps and reports (Kirkegaard & others, 1970; Dear & others, 1971; Willmott & others, 1986; Donchak & Holmes, 1991; Fergusson & others, 1988, 1990a,b, 1993, 1994; Leitch & others, 2003). Mapping was essentially confined to the Palaeozoic and Mesozoic rocks. The Tertiary and Quaternary geological units were mainly defined from airphoto interpretation and existing maps and reports.

The Yarrol Project covered all or parts of the following 1:100 000 Sheet areas: Rookwood (8851), Ridgelands (8951), Rockhampton (9051), Duaringa (8850), Mount Morgan (8950), Bajool (9050), Gladstone (9150), Banana (8949), Biloela (9049), Calliope (9149), Scoria (9048), and Monto (9148). The southern part of the Cape Capricorn 1:100 000 Sheet (9151) was included in the Gladstone 1:100 000 special sheet to cover all of Curtis Island. The Yarrol Project area adjoins the South Connors-Auburn-Gogango Project area (Withnall & others, 2009) to the north, west and south, and the Bundaberg Project area to the east.

The main regional centres are Rockhampton, Yeppoon, Biloela, Gladstone, Monto and Mount Morgan. Gladstone is the main port in the region. The area is serviced by the North Coast Railway, the Central Railway and the Yeppoon Branch Railway. The Bruce Highway, Capricorn Highway, Burnett Highway and Dawson Highway cross the area. Most of the area has good road access except in the areas with high ranges.

Historically the main industries were beef raising and mining. Beef is still a big industry in the area, but now the large mines and goldfields have largely ceased production, the only major operation being the Callide coal mine. Gladstone is an important industrial centre and major port. Dairying and crop and fruit growing are of lesser importance.

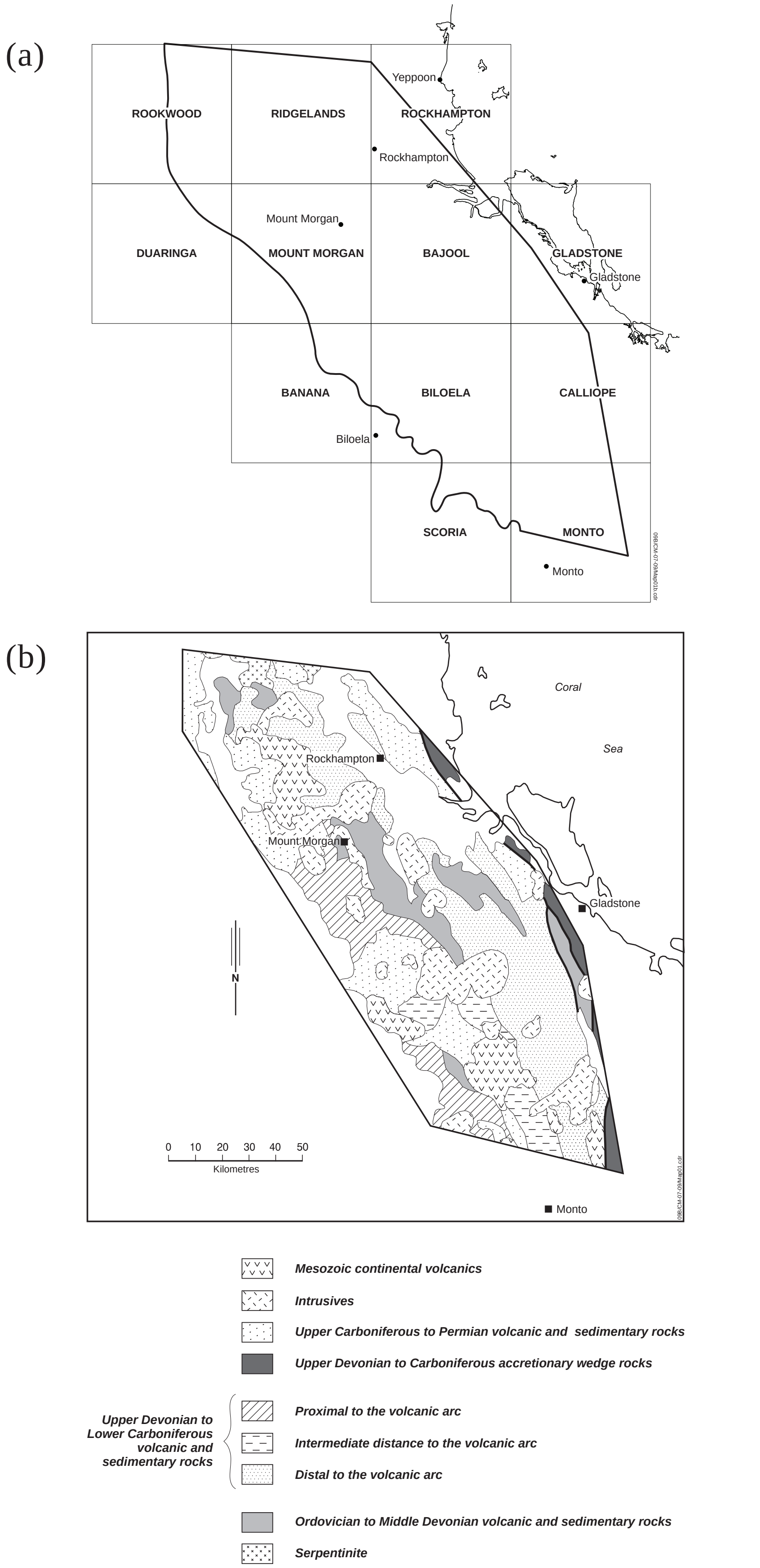


Figure 1. Location of the Yarrol Project area, showing (a) 1:100 000 sheet areas and (b) generalised geology (blank areas are Cainozoic cover). Mapping was concentrated on the Yarrol Province between the Biloela Basin and Gogango Overfolded Zone to the west and the Yarrol Fault to the east, but extended to the coastline of the Rockhampton and Gladstone 1:100 000 sheets.

OVERVIEW OF THE YARROL PROJECT

The Yarrol Project commenced in 1994 with a review by Murray (1994) which outlined the existing data, the main problems to be addressed, and the results expected by the mineral exploration industry. The Yarrol Project covered the core area identified in that study, and most of the remainder was mapped as the Bundaberg Project.

The project started with an intensive search for geological and mineral exploration information in Company Reports submitted to the Department of Minerals and Energy (presently the Department of Natural Resources and Mines) (Withnall & others, 1995). These Company Reports were summarised in a number of reports outlining the exploration history of the Yarrol Project area (Blake & others, 1995, 1996; Hayward & others, 1995; Osborne & others, 1997). The main field mapping was carried out from 1995 to 1997, accompanied by mineral occurrence mapping. All mineral occurrence data for the Yarrol Project have been published (Morwood, 2002a&b, 2003, Morwood & Blake, 2002, Burrows, 2004).

The geological mapping was supported by and based on semi-detailed airborne geophysical data (radiometric and magnetic) provided by the AIRDATA Project. The airborne geophysical data were integral to mapping many parts of the Yarrol Project area, and were particularly useful in distinguishing the various phases of many of the plutonic batholiths. They were also useful in highlighting radiometric differences between the uppermost Silurian to lowermost Carboniferous rocks and the Lower Carboniferous (Mississippian) Rockhampton Group, and between the Rockhampton Group and the Upper Carboniferous (Pennsylvanian) to Lower Permian units.

Initial results were published by the project team (Yarrol Project Team, 1997; Murray & the Yarrol Project Team, 1997; Blake & others, 1998; Fordham & others, 1998; Hayward & others, 1998; Morwood & others, 1998; Murray, 1998; Simpson & others, 1998; Crouch & Parfrey, 1998; Crouch, 1999a,b). A safari and excursion (Anonymous, 1997c; Yarrol and South Connors Project Teams, 1998) were designed to circulate the results to industry and academia. Mineral resource assessments of the Yarrol Province have been carried out by Scott (1998, 2006).

All data gathered during the project have been entered into the Geoscience and Resource Data Base (GRDB) in the departmental database MERLIN. The geology line work was entered by cartographers. All maps and reports have been produced digitally.

The 1:100 000 scale maps were produced in 2001, and revised versions incorporating corrections were released in 2006. Further changes are being made as a result of mapping in adjacent areas, and the re-examination of relationships amongst Cretaceous sediments and volcanics near Stanwell west of Rockhampton. A GIS package containing data from the Yarrol Project was released in 2002 (Geological Survey of Queensland, 2002), and has subsequently been updated to include results from the South Connors-Auburn-Gogango Project as the Central Queensland (Yarrol-Connors-Auburn) GIS (Geological Survey of Queensland, 2005).

An article by Bryan & others (2001) challenged the basic interpretation of the geological setting of the Yarrol Province, and was the subject of further debate (Murray & others, 2003; Bryan & others, 2003) and investigations involving the collection and chemical analysis of additional rock samples (Murray & Blake, 2005). The result was a different interpretation of the Devonian evolution of this region (Murray & Blake, 2005; Murray, 2007). Subsequent investigations have concentrated on Mesozoic rocks of the Stanwell area west of Rockhampton, and have produced more changes to the interpretations of geological units in that region. These new investigations and interpretations have delayed completion of this report.

Geochronological determinations were carried out by K-Ar (A.W. Webb), Rb-Sr (University of Alberta) and U-Pb (C.M. Fanning, University of Alberta), and results are included in Appendix 1.

In addition to work by the project team, palaeontological identifications of brachiopod faunas were carried out by Dr Sue Parfrey (Queensland Museum), Dr John Roberts (University of New South Wales), and the late Dr John Dear (Marlborough Gold). Dr John Jell (University of Queensland) assisted with the determination of the coral faunas. Results are either reported in the text or included in Appendixes 2, 3 and 4.

STRATIGRAPHY

PALAEOZOIC

Marlborough Metamorphics (Pzm_m)

(C.G. Murray)

Introduction

Only a small part of this unit occurs within the Yarrol Project area, at the northern edge of the Ridgelands 1:100 000 sheet area (Figure 2).

Distribution

The metamorphic rocks form generally north-west trending belts alternating with the Princhester Serpentinite within the Marlborough terrane. The largest belt is in the centre and is 10km wide. Smaller belts occur to both east and west.

Topographic expression

The Marlborough Metamorphics form a subdued topography with only low rounded hills. They contrast with the sharp-crested ridges developed on the serpentinite.

Geophysical expression

The central and eastern belts are similar, and have very low magnetism and low but patchy radiometrics. The western belt is also not magnetic, but has a high radiometric signature, reflecting a different origin.

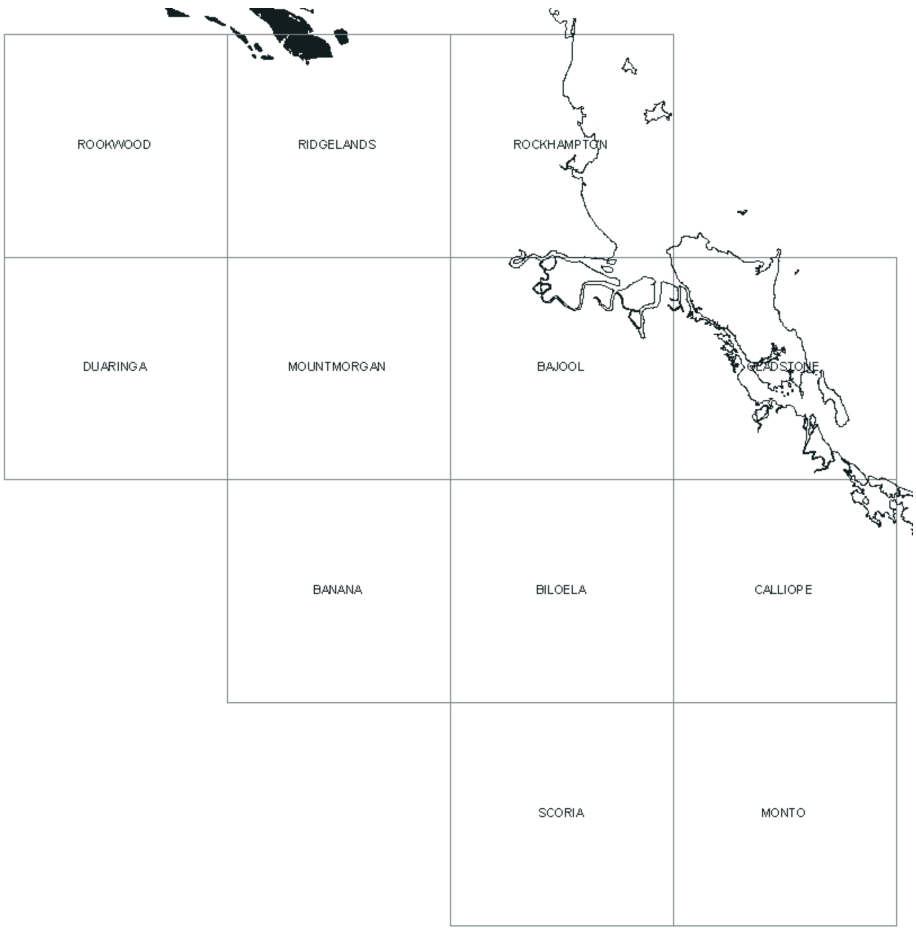


Figure 2. Distribution of the Marlborough Metamorphics

Lithology

The central and eastern belts of Marlborough Metamorphics are metasediments. The central belt is mainly of lower greenschist grade, and consists of chlorite schist and phyllite, with local muscovite- and higher grade biotite and amphibole-bearing schist. The eastern belt is composed of sediments which typically are contact metamorphosed by widespread intrusions north of Glen Geddes.

The different radiometric pattern of the western belt, now a biotite schist, suggests that it was originally a foliated granite.

Thickness

If the dominant foliation approximates bedding, the widest belt of metamorphics, which is fault bounded to both the east and the west, represents a thickness of over 6km. This suggests that the apparent thickness has been considerably increased by folding.

Structure

The Marlborough terrane is a thin-skinned (<2km thick) nappe sheet in which ultramafic and metamorphic rocks form alternating belts separated by moderately dipping to steep thrust faults and shear zones (Holcombe & others, 1997a,b). The overall structure was described as the lower part of an eroded duplex by Harbort (2001). The main fabric (S_1) in the metamorphics is parallel to the sedimentary layering where that can be recognised, and is locally overprinted by a second deformation (Kirkegaard & others, 1970; Henderson & others, 1993).

Age

Ages of ~245Ma have been obtained from the Marlborough Metamorphics (Henderson & others, 1993; Holcombe & others, 1997a,b), and are attributed to thrust movement, although Harbort (2001) regarded the thrusting event as prior to the age of an undeformed pluton at 253Ma.

The age of the sedimentary protolith is uncertain. Holcombe & others (1997b) considered that the rocks contain equivalents of the accretionary wedge sequence to the east, and if so they would probably be Carboniferous in age. Henderson & others (1993) expressed an opposite view that the metamorphic rocks could not be correlated with the accretionary wedge, and were an early Palaeozoic unit that formed basement to the younger forearc strata. The presence of beds of pure limestone (now marble) in the northern continuation of the main central belt of metamorphics on the access road to the telecommunication facility at Princhester (Kirkegaard & others, 1970, page 10) supports the latter view that the metamorphics cannot be correlated with the accretionary wedge, but it is unlikely that they are as old as early Palaeozoic. It is most probable that they were intercalated with the ultramafic mass during extension towards the end of the Carboniferous, and later emplaced and metamorphosed by thrusting (Murray, 2007).

Stratigraphic relations

The Marlborough Metamorphics are faulted against the Princhester Serpentinite and both are structurally emplaced over unnamed Permian strata previously assigned to the Berserker Group. They are overlain by Quaternary sediments.

Correlation with other units

Based on the presence of beds of pure limestone, the most likely correlation is with Devonian strata, although the Lower Carboniferous Rockhampton Group cannot be excluded. If the western belt originally consisted of foliated granites, it has no equivalents in the Yarrol Project area.

Economic significance

The notorious Canoona goldfield was located over the central belt of Marlborough Metamorphics and adjacent serpentinite. The source of the gold was probably the ultramafic rock.

ORDOVICIAN

Unnamed Ordovician (Ou)

(P.R. Blake)

Introduction

Two localities ~30km south-south-east of Calliope contain rocks of Ordovician age, based on conodont determinations (Figure 3).

One locality is at GR329813 7317553 in the Calliope 1:100 000 Sheet area, close to Santa Glen homestead. Limestone collected from this site yielded Late Ordovician conodonts, but comes from a conglomerate that also contains Devonian limestone clasts. The conglomerate is considered to be part of the Calliope beds based on the surrounding rock types.

The second locality, described here, comprises a limestone that appears to be interbedded with volcaniclastic sedimentary rocks ~1km west of Santa Glen homestead.

Distribution

The unnamed Ordovician rocks occur ~1km west of Santa Glen homestead at GR329227 7317607 in the Calliope 1:100 000 Sheet area.

Topographic expression

The outcrop area is too small to have any individual topographic expression.

Figure 3. Distribution of the unnamed Ordovician rocks.

Geophysical expression

The outcrop is too small to be represented on airborne geophysical images.

Lithology

The known Ordovician limestone is a fine-grained micritic limestone with no obvious macrofossils. It is several metres thick, and up to 100m long.

Nearby well-sorted fine- to medium-grained, and thin-bedded volcaniclastic sandstone and conglomerate was sourced from basaltic to andesitic lavas. The conglomerate is granule to pebble in grain size, moderately-sorted to well-sorted, and cemented by calcite. The clasts are usually well-rounded.

These volcaniclastic sedimentary rocks differ from the sedimentary rocks of the surrounding Calliope beds in that they are well-sorted and the clasts are well-rounded, and it is possible that they are part of the Ordovician sequence. The fine grained units can contain sponges, brachiopods and trilobites of Ordovician age (Peter Jell personal communication, 2011).

Environment of deposition

Given the restricted nature of the outcrop, all that can be determined is that the sequence was deposited in a marine environment.

Thickness

The exposed thickness is at most several tens of metres.

Structure

Bedding dips steeply to the west, with no evidence of folding.

Biostratigraphy and age

The conodonts extracted from the limestone are different to faunas found elsewhere in Australia, but based on comparisons to faunas found overseas they are interpreted to be Early Ordovician in age. Peter Jell’s (personal communicaton, 2011) determination on the brachiopods and trilobite support this interpretation.

Stratigraphic relationships

The relationship between the surrounding Calliope beds and the unnamed Ordovician rocks is unknown due to poor outcrop. The area is proximal to the Yarrol Fault, and cut by a large number of other faults. Therefore, the Ordovician outcrop is interpreted as a fault sliver, but it is also possible that it is an allochthonous block within the Calliope beds, similar to but larger than the limestones at GR329813 7317553.

Correlation with other units

These rocks are unique in the Yarrol Province, and are much older than any other sedimentary units. Cambrian to Ordovician sequences along the Peel Fault near Tamworth in the southern New England Orogen (Packham, 1969; Cawood, 1976) are possible correlatives.

LATE SILURIAN – MIDDLE DEVONIAN

Craigilee beds (SDc)

(G.A. Simpson, J. Domagala, & P.R. Blake)

Introduction

The presence of Lower to Middle Devonian rocks in the Mount Cassidy area 60km west-north-west of Rockhampton was first noted by Reid (1931a). Malone & others (1969) and Kirkegaard & others (1970) mapped their full extent, and collected fossils suggesting that part of the sequence is of latest Silurian age. The sequence was named the Craigilee beds by Talent & others (1982) in a discussion of the correlation of Silurian rocks in Australia, New Zealand, and New Guinea.

Figure 4. Distribution of the Craigilee beds

Mapping by Bradley (1994) and the Yarrol Project team outlined the internal stratigraphy of the Craigilee beds in more detail, and identified three unnamed subunits.

Distribution

The Craigilee beds occur in a number of inliers and fault blocks mainly east of the Fitzroy River (Figure 4). It is convenient to describe them as two discrete belts, an eastern group previously named informally as the Morinish beds (Bradley, 1994) (here referred to as the Morinish block), and a western group referred to collectively as the Mount Cassidy block.

In the Mount Cassidy block, the unit crops out as a north-trending linear belt 27km long extending northwards from about 7km south of Armagh homestead to about 4km north of Melrose homestead. It is up to 10km wide at the northern end, narrowing to the south, and is confined mainly to the eastern side of the Fitzroy River. The composite Morinish block is smaller: the main outcrop area is about 8km square, and is separated from a smaller exposure to the south by the Ridgelands Granodiorite. The Morinish and Mount Cassidy blocks are separated by the Devonian to Carboniferous Mount Alma Formation, the Lower Carboniferous Rockhampton Group, and the Upper Carboniferous to lowermost Permian Lorray Formation.

A small occurrence of fossiliferous Middle Devonian limestone surrounded by Permian rocks south of Mona Vale homestead (Kirkegaard & others, 1970, page 10) can probably be correlated with the Craigilee beds.

Derivation of name

The name is derived from Craigilee homestead east of the Fitzroy River in the central part of the Mount Cassidy block.

Topographic expression

The Craigilee beds form low to moderate hills, with strike ridges developed locally on sediment-dominated sequences. Most of the lower country has been cleared for grazing, but a dense vine scrub covers the area south of Melrose homestead.

Geophysical expression

Sediment-dominated parts of the unit are represented by medium to dark brown tones on images derived from radiometric data, reflecting a low response in all three channels. In contrast, where primary volcanic rocks are most abundant, the response is a dark red, indicating a relatively high potassium content. This strong potassium response is mainly due to felsic rocks, which have a K_2O content of more than 2% compared to 1% or less for mafic to intermediate compositions.

Overall, the Craigilee beds have a fairly uniform moderate to low total magnetic intensity, but they show more variation on images of the first vertical derivative. The Morinish block gives a rather bland and featureless pattern except along fault zones and adjacent to the Ridgelands Granodiorite where skarns are developed. The Mount Cassidy block has a much more prominent magnetic pattern, including a series of parallel south-east trending magnetic features that cross-cut the mapped geological units.

Lithology

Although the Craigilee beds occur in a number of separate outcrops, it has been possible to determine a generalised stratigraphic succession supported by biostratigraphic age determinations from limestones, but no formal subdivision has been attempted. They have been subdivided into a lower unit SDc_1, middle unit SDc_2, and an upper unit SDc_3 on the basis of rock type and sandstone composition. The current interpretation of the stratigraphic sequence differs from that suggested by Bradley (1994) because subsequent more detailed conodont analysis indicated that his correlations were incorrect.

In the Morinish block, unit SDc_1 comprises sandstones, siltstones, granular conglomerates, limestone, and minor volcanic rocks. Sandstones are feldspathic to feldspatholithic. Plagioclase is the main clastic component with lesser volcanic quartz and volcanic and sedimentary lithic grains. Limestone sequences composed of wackestone to packstone contain Early Devonian coral faunas including tabulate corals (*Heliolites, Favosites,* and *Alveolites*) as well as rugose corals (*Tryplasma*) and nautiloids. Intermediate to felsic volcanics and/or high level intrusions and rare basalts are present but form only a minor part of the sequence. Unit SDc_2 contains abundant porphyritic intermediate to felsic volcanics and/or high level intrusions, accompanied by ripple and cross-laminated sandstone together with mudstone and siltstone. The volcanics include local flow breccias and hyaloclastites as well as lavas. Point-count data on sandstones show a much greater proportion of lithic grains (mainly volcanic) and less quartz in unit SDc_2 than in SDc_1 (Bradley, 1994). Conglomerate lenses range from polymictic with volcanic and limestone clasts to oligomictic sourced exclusively from volcanic rocks. Unit SDc_3 is a sediment-dominated sequence of feldspatholithic sandstone, siltstone and mudstone, minor conglomerate, breccias and limestone. Conglomerates from the upper part of the unit contain clasts of limestone, andesitic and dacitic volcanics, and rare granite (Bradley, 1994). The source of the granitic clasts is not known as there are no granitoids of appropriate age in the area. Sandstones are moderately quartz-rich like those in the basal section SDc_1.

Malone & others (1969) considered that volcanics, mainly altered andesitic and basaltic flows (keratophyre and spilite), were the dominant rock types in the Mount Cassidy block, with subordinate sediments including prominent limestone lenses. Some of the rocks they described as crystal-lithic tuffs are now interpreted as volcaniclastic sediments. As in the Morinish block, the sequence has been divided into lower and upper units with a higher proportion of sedimentary rocks (SDc_1 and SDc_3), and a middle unit SDc_2 dominated by volcanics. The volcanics consist mainly of massive flows and/or high level intrusives of intermediate to mafic composition.

Unit SDc_3 is a sediment-dominated sequence consisting of volcanolithic breccia, sandstone, siltstone, silicified mudstone, minor polymictic conglomerate, and limestone. Included in this sequence is a notable matrix-supported breccia several hundreds of metres thick just east of Mount Cassidy. It is composed of volcanolithic sand to granule matrix with large clasts of pumiceous to vesicular, bomb-like blocks (to 30cm) with chilled margins. The large clasts are elongate to irregularly shaped and most are oriented with their long axis parallel to bedding (Figure 5). The matrix is sand to pebble-sized volcanic clasts with clast vesicles and interstitial cavities infilled with calcite and chlorite cement. Thin to medium granule to pebble beds separated by thin silty layers, interpreted as density current deposits, occur towards the top of this sequence. Minor volcanic flows with associated flow breccias and peperites occur towards the base of unit SDc_3 (Figure 6).

Figure 5. Massive, very thick volcanic breccia composed of angular to platy, pebble to cobble sized, scoriaceous clasts in a fine to very coarse sand sized matrix of rock fragments and subordinate plagioclase crystals, Craigilee beds.

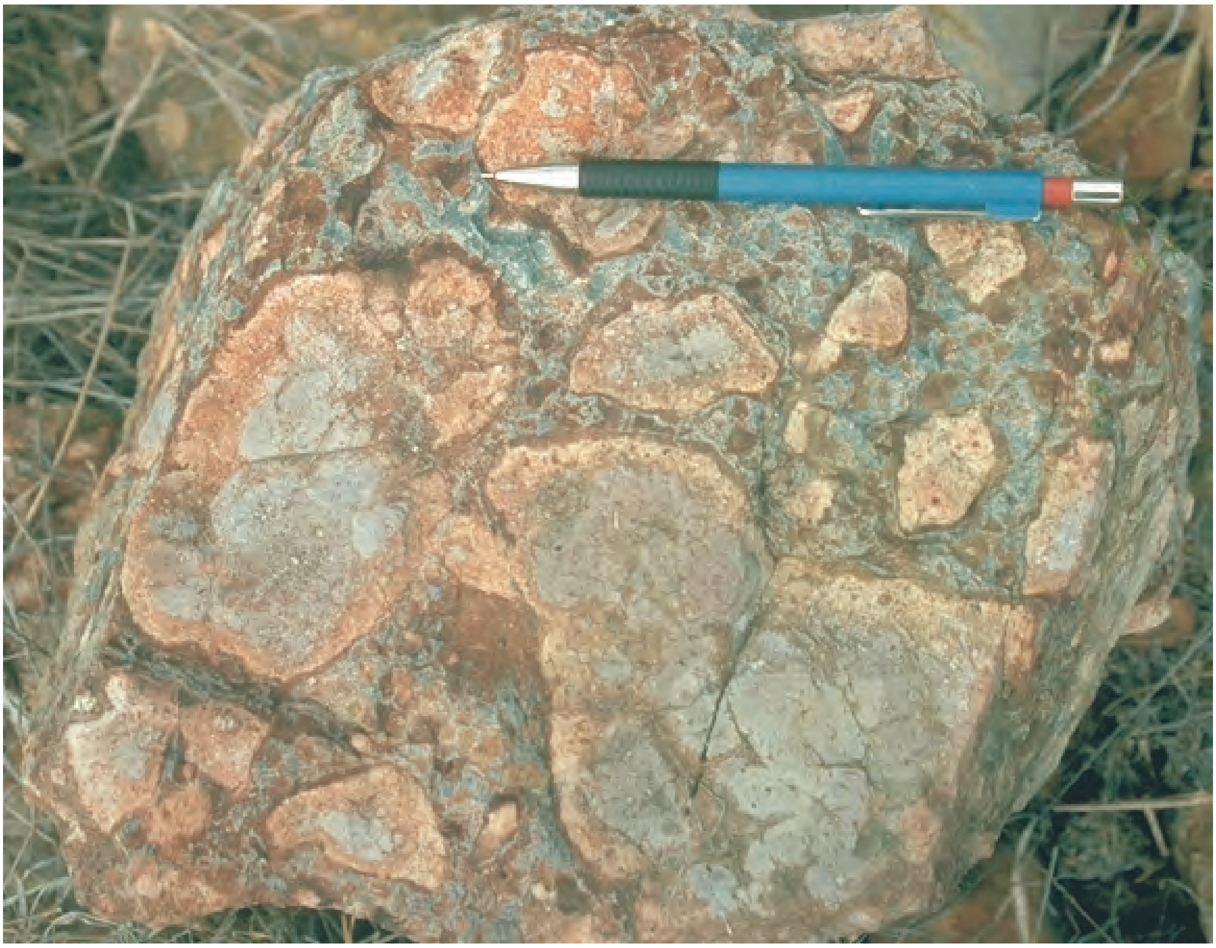


Figure 6. Volcanic breccia showing at least two phases of brecciation, Craigilee beds

The volcanoclastic breccia/arenites are typically medium to dark green-grey, thin to very thick bedded, generally clast supported, massive, with local normal grading, and horizontal laminations. Clasts are characteristically angular and commonly tabular in the pebble to cobble size range. Many of these tabular clasts are oriented with their long axis paralled to bedding and some pumiceous clasts resemble fiamme (pseudofiamme). The dominant clasts are intermediate to mafic volcanic rocks with textures ranging from richly porphyritic to aphyric. The breccia matrix consists mainly of sand size volcanolithic clasts of similar composition to the larger clasts, with minor pumice and crystals of quartz and feldspar.

The volcanoclastic arenites consist mainly of the three end member volcanic components of lithics, vitrics, and crystals, in various proportions. The lithic to vitric ratio is commonly difficult to determine due to the high degree of recrystallisation brought about by devitrification and diagenetic alteration. Plagioclase is the dominant crystal component with minor K-feldspar and rounded and embayed volcanic quartz. Limestone clasts are the most common non-volcanic component, commonly occurring in beds adjacent to limestone bearing strata.

The siltstones and mudstones are generally dark grey, thin to laminated, and may be ripple laminated. Soft sediment slump folds are locally common. Relict glass shards are present, as are carbonate microspheres and radiolarian-like siliceous spheres. Most of these fine-grained rocks appear to be silicified, as they generally produce a conchoidal fracture when struck with a hammer.

SDc_3^{cg} is a distinctive massive, polymict, clast supported, pebble to boulder conglomerate characterised by the common occurrence of fossiliferous limestone clasts (Figure 7). It is exposed as a laterally continuous, very thick bed towards the eastern margin of the unit. The dominant clast type is similar to that which occurs in the breccias (intermediate to basic, porphyritic to aphyric, variably vesicular volcanic clasts) with subordinate silicified mudstone and siltstone clasts. On weathered surfaces many of the volcanic clasts are pale and have a bleached appearance (Figure 8). Clasts are angular to rounded and locally imbricated.

Figure 7. Polymictic, clast supported conglomerate with clasts of andesite, limestone and silicified mudstone in a matrix of feldspar crystals and volcanic lithic fragments. The fabric is massive, but platy clasts are parallel to bedding, Craigilee beds.

Figure 8. Very thick, massive, cobble conglomerate with angular to rounded clasts of andesite and minor limestone and tuff in a matrix of feldspar crystals and volcanic lithic fragments, Craigilee beds.

Limestones within the Cassidy block are quite variable in composition and appearance. They range from autochthonous carbonate accumulations to allochthonous redeposited calcarenites, and vary from dark grey to light grey fossiliferous rudstone to medium bedded calcarenite to thin bedded marly limestone.

East of Armagh homestead, limestone in SDc_1 is lensoidal and extends several hundreds of metres along strike and is up to ~50m thick. Bedding ranges from thin to very thick and is defined by marly partings and orientation of *in situ* and locally reworked carbonate fossils. The limestone fabric ranges from mudstone to rudstone, to a poorly developed bindstone and bafflestone. The fossils include crinoids, corals (tabulate and rugose), stromatoporoids (amphipora and laminar), and brachiopods which are in growth or near growth positions. The matrix is characteristically dark grey and consists mainly of carbonaceous calcilutite with subordinate calcarenite. Bedding parallel styloseams are common and dolomitisation is locally developed. Geopetal structures indicate younging is eastwards.

A limestone assigned to SDc_2 about 2km south of Hillview homestead has a range of similar fabrics. It differs, however, in that it is a larger body, extending about 2km in an anomalous east-south-east direction with a thickness of up to 200m. The pale grey colour and partial recrystallisation are attributed to a nearby granitic intrusion. A coral fauna from this location indicates an Early Devonian age.

A different type of limestone occurs west of the Fitzroy River at GR801420 7427550 in SDc_1. It consists mainly of calcarenite which is relatively well stratified and well sorted. The beds are dark grey, thin to medium bedded, with normal and reverse grading, horizontal and ripple laminations, and wavy bedding (wave length of ~20cm, amplitude of ~2cm) (Figure 9). Beds consist mainly of sand size carbonate grains with minor siliciclastic (most probably volcanoclastic) grains. A crinoidal calcirudite was also observed. Recessive siliciclastic layers are commonly interbedded with the relatively resistant limestone beds.

An impure type of limestone crops out discontinuously towards the eastern margin of the unit in subunit SDc_3. This rock is best described as an impure calcilutite to calcareous mudstone. However, as it contains significant carbonate and is a distinctive unit, it is mapped as limestone. It is dark grey, thin to laminated, ranges up to 20m thick, and although no macrofossils were apparent, spicules and spherical calcisphere-like particles were observed.

Figure 9. Graded limestone bed, Craigilee beds

Geochemistry

Analyses of volcanics from the Craigilee beds were carried out by Bradley (1994) and as part of the Yarrol project. Murray & Blake (2005) used analyses of the more mafic samples to interpret the tectonic setting. They concluded that the geochemistry was consistent with formation in a supra-subduction zone volcanic arc. The Craigilee samples differ from other Devonian analyses in their strong enrichment in light rare earth elements (LREE). Normally this would suggest formation in a continental margin arc, but Nb, Ta, Zr and Hf contents, and in some cases Ti, are much too low. On the Th/Yb versus Ta/Yb discriminant diagram of Gorton & Schandl (2000), samples from the Craigilee beds plot in the oceanic arc field (Murray & Blake, 2005, figure 8). One sample from the Craigilee beds is almost identical in geochemistry to sample 11885 from the Middle Devonian Captain Rocks Formation of the southern New England Orogen, interpreted as an intra-oceanic arc by Offler & Gamble (2002), who attributed the enrichment in LREE to subduction and partial melting of sediments. Other rocks from the Craigilee beds display more enriched but parallel trends on multi-element and REE plots (Murray & Blake, 2005, figure 7). They show only modest increases in Ti/Zr and Sc/Y as Zr decreases, contrasting markedly with the Raspberry Creek Formation and the intrusives in the Mount Warner Volcanics (Murray & Blake, 2005, figure 6e,f). Ce/Y values suggest a crustal thickness of about 25km (Mantle & Collins, 2008).

Environment of deposition

The sedimentary rocks in the Craigilee beds are interpreted as submarine deposits. The presence of sub-aqueous volcanics, limestones, and fossiliferous sandstones throughout the succession indicates that the unit was deposited in a marine environment. Bradley (1994) described the depositional setting as a volcanically-active marine environment proximal to a volcanic arc. This interpretation is supported from point count data on sandstones and by the geochemistry of volcanic rocks.

Episodic but widespread submarine volcanism is inferred from the close association of the volcanic and marine sedimentary rocks and the presence of flow breccias and peperites. Proximal volcanism is inferred from the occurrence of the volcanic bomb breccia in the Mount Cassidy block. Geochemical evidence from lavas, breccia and/or high level intrusives in the unit suggests that the volcanic regime was arc-related.

Thickness

No continuous section could be measured through this unit due to faulting. On the basis of a composite section measured by Bradley (1994) the unit is ~3000m thick. Kirkegaard & others (1970) estimated a similar thickness of 2500m for a section south of Marble Ridges homestead.

Structure

Beds in the northernmost part of the Morinish block, separated by a north-west trending fault, dip uniformly to the south-west at steep angles, and are cut by a steep north-east-dipping cleavage (Kirkegaard & others, 1970; Bradley, 1994). Elsewhere, dips are more moderate and cleavage is only locally developed. Malone & others (1969) described the Mount Cassidy block as an east-dipping wedge, but dips are irregular. Detailed structural studies by Holcombe & others (1997a) have interpreted it as the core of a large antiformal structure, the Craigilee Anticline, that is a breached propagation fold related to westward thrusting. Intense folding is only locally developed in thin-bedded limestones. In both the Mount Cassidy block and the central part of the Morinish block, the unit appears to have been re-folded along east–west axes to produce arcuate bedding trends.

Biostratigraphy and age

Corals collected from a limestone near Mount Cassidy were identified as Lower or Middle Devonian (Whitehouse, 1931). McKellar (1969) described coral and brachiopod faunas in the unit, east of Armagh homestead, and assigned a latest Silurian age. The limestones contain the following corals, *Tryplasma wellingtonense*, *Aphyllum simplexum*, *Pseudamplexus princeps*, *Microplasma ronense*, *Xystriphyllum insigne*, *Favosites goldfussi*, *F. grandipora*, *F.* sp., *Squameofavosites bryani*, *S. nitidus*, *S. craigileei*, *Thamnopora boloniensis*, *Heliolites daintreei*, *H. comminus,* and *Syringopora jonesi* and was interpreted as Early Devonian (Emsian) by Blake (2010).

The subdivision of the unit is based partly on limestone ages determined from conodont faunas. A latest Silurian (Přídolí) age was confirmed for limestone towards the base of unit SDc_1 located east of Armagh homestead in the Mount Cassidy block, although corals from this locality appear to be restricted to the Early Devonian (Blake, 2005, 2010). Limestones of this unit in the Morinish block contain Early Devonian (Lochkovian to Emsian) conodont and coral faunas. Givetian ages were obtained for limestones in unit SDc_3 of both the Morinish and Mount Cassidy blocks.

However, the ages of some limestones are anomalous, possibly reflecting uplift and erosion during deposition of the Craigilee beds. Alternatively, the stratigraphy may be more complex than assumed for the simple subdivision into three informal subunits.

Stratigraphic relationships

Malone & others (1969) and Kirkegaard & others (1970) considered that the Craigilee beds were unconformably overlain by the surrounding Upper Devonian to Lower Carboniferous sequences. However, Morand (1993a) suggested that contacts were faulted, and most are mapped as faults on the geological maps of the Rookwood and Ridgelands 1:100 000 sheets. Only the eastern contact of the isolated outcrop of Craigilee beds south of the Ridgelands Granodiorite is mapped as an unfaulted boundary. Nevertheless, strong evidence in favour of an unconformable relationship is provided by the occurrence in the younger rocks of fossiliferous limestone clasts derived from the Craigilee beds (Kirkegaard & others, 1970, page 45).

The Craigilee beds are also faulted against the uppermost Carboniferous and Lower Permian Youlambie Conglomerate, and are intruded by a number of Permian to Triassic granitoids, including the Ridgelands Granodiorite.

Correlation with other units

Some degree of lithological correlation is possible between the Craigilee beds and the Capella Creek Group, but fossils from the Craigilee beds cover a much longer time span, from latest Silurian to Middle Devonian, the geochemistry of mafic volcanics is very distinct (Murray & Blake, 2005), and geophysical signatures differ. Therefore the Craigilee beds are no longer considered to be part of the Mount Morgan Stratigraphic Assemblage, where they were placed by the Yarrol Project Team (1997).

Economic significance

The Craigilee beds contain locally thick limestone beds.

Calliope beds (SDa)

(P.R. Blake)

Introduction

The Calliope beds were defined by Dear & others (1971), and were described as a succession of dark green intermediate and possibly basic lava, tuff and agglomerate, and grey limestone. Morand (1993a) combined the Calliope beds with the Mount Holly beds of Kirkegaard & others (1970) because of their proximity and lithological similarity. However, the Mount Holly beds have been re-mapped as the Erebus beds and the Mount Alma Formation. The Erebus beds differ significantly from the Calliope beds in both rock type and geophysical signature, while the Mount Alma Formation is younger.

Distribution

The Calliope beds occur as a discontinuous, elongate belt of rocks ~100km long and up to 5km wide extending southward from Yarwun on the Gladstone 1:100 000 Sheet area to east of Kalpowar on the Monto 1:100 000 Sheet area. The extension of the Calliope beds south from Many Peaks as a narrow complexly-faulted belt along the western side of the Yarrol Fault is based on the magnetic pattern shown on images produced from airborne survey data (Figure 10). This belt was previously mapped as Carboniferous rocks (Dear & others, 1971).

Topographic expression

The Calliope beds produce a topography consisting dominantly of low hills, and outcrop is generally poor.

Figure 10. Distribution of the Calliope beds

Geophysical expression

The Calliope beds have a very low response in all three channels on images of radiometric data, ranging from black to very dark tones. In contrast, the unit has a strong magnetic signature with linear features that are commonly parallel to the outcrop pattern. It is possible that some of these strong magnetic features are dolerite and microdiorite dykes, or are related in some way to the serpentinites along the Yarrol Fault to the east, but many of the conglomerates within the sequence also have high magnetic susceptibility values.

Lithology

The poor outcrop makes it difficult to determine the true abundance of the different rock types and their stratigraphic succession. As well, there may be some variation in composition along the length of the outcrop belt. Kirkegaard & others (1970) reported tuffs and flows of acid to intermediate composition in the north (although the tuffs are probably volcaniclastic sediments), and Morand (1993a) also considered that silicic volcanic rocks are common, whereas Dear & others (1971) described the sequence around Calliope as intermediate to mafic.

The most common rock types observed are conglomerate and breccia. They are usually poorly sorted and range from clast-supported to matrix-supported. Most of the clasts are granule to pebble in size with some cobbles, and were sourced from volcanics, dominantly andesite and basalt with rare dacite, and granite. The mafic volcanic clasts commonly have amygdales filled by calcite and chlorite, and sparse phenocrysts of plagioclase and rare augite. Some conglomerates contain clasts of limestone.

Minor rock types identified are sandstone, basaltic and andesitic lava, limestone, and rare siltstone. The sandstone is mostly medium- to coarse-grained, but some is fine-grained. The sandstones are lithic and are composed of basalt and andesite clasts with rare dacite. The lavas are mostly aphyric to sparsely plagioclase-phyric. The limestones, best exposed at Taragoola, are fossiliferous, containing crinoids, brachiopods, corals, gastropods and conodonts. The siltstones are massive to laminated, and contain very rare brachiopods, crinoids, and trilobites.

The belt south of Many Peaks consists mainly of basic to intermediate volcanics. The rocks are fine-grained, with small crystals of plagioclase and rare clinopyroxene in some samples. They are generally massive, with fragmental texture only apparent on larger exposures. Plagioclase-rich dykes of dioritic composition cut the sequence. Well-sorted volcaniclastic sandstone, siltstone and conglomerate are subordinate to the primary volcanics. All the volcanic and sedimentary rocks are strongly fractured.

Geochemistry

Morand (1993a) analysed two basalts and one mafic andesite (his samples 31722, 31724 and 31742) from the Calliope beds. From the results of standard discriminant plots, he concluded that these rocks, together with other samples from his Calliope Volcanic Assemblage, formed in a volcanic arc. The three samples from the Calliope beds fall mainly in arc fields on standard plots, but 31724 and 31742 are transitional to back-arc basin basalts because of their comparatively high TiO_2 contents (Figure 11). Offler & Gamble (2002, page 364) used additional rare earth element (REE) and high field strength element (HFSE) data for sample 31724 to interpret this rock as an island arc tholeiite. Sample 31724 has very similar patterns to the Late Devonian V2 tholeiite suite II from dykes in the Princhester Serpentinite (Bruce & Niu, 2000a; Murray, 2007), and can also be matched with samples from the Izu-Bonin arc (Figure 12). Its relatively high large ion lithophile element (LILE) contents, if primary, support an arc rather than a back-arc origin.

Three rocks from the belt south of Many Peaks were analysed, and appear to be back-arc basin basalts and andesites (Figure 11). The two most mafic samples have similar geochemistry to the Folly Basalt in the Nundle area (New England, New South Wales). Although Caprarelli & Leitch (2002) interpreted the geochemistry of the Folly Basalt in terms of a forearc environment, it is more similar to modern back-arc basin basalts. The tectonic setting of the Folly Basalt, immediately to the west of the Peel Fault, is the same as that of the ?Calliope beds south of Many Peaks. On a plot of Nb/Yb vesus Th/Yb, the compositions overlap with basalts from the modern Mariana Trough (Gribble & others, 1998; R. Stern & J. Pearce, personal communication).

Thus the very limited geochemical results from the Calliope beds and their possible extension south of Many Peaks suggest that the tectonic setting ranged from oceanic arc to back-arc basin. The flat REE patterns indicate that the crust was very thin (Figure 13).

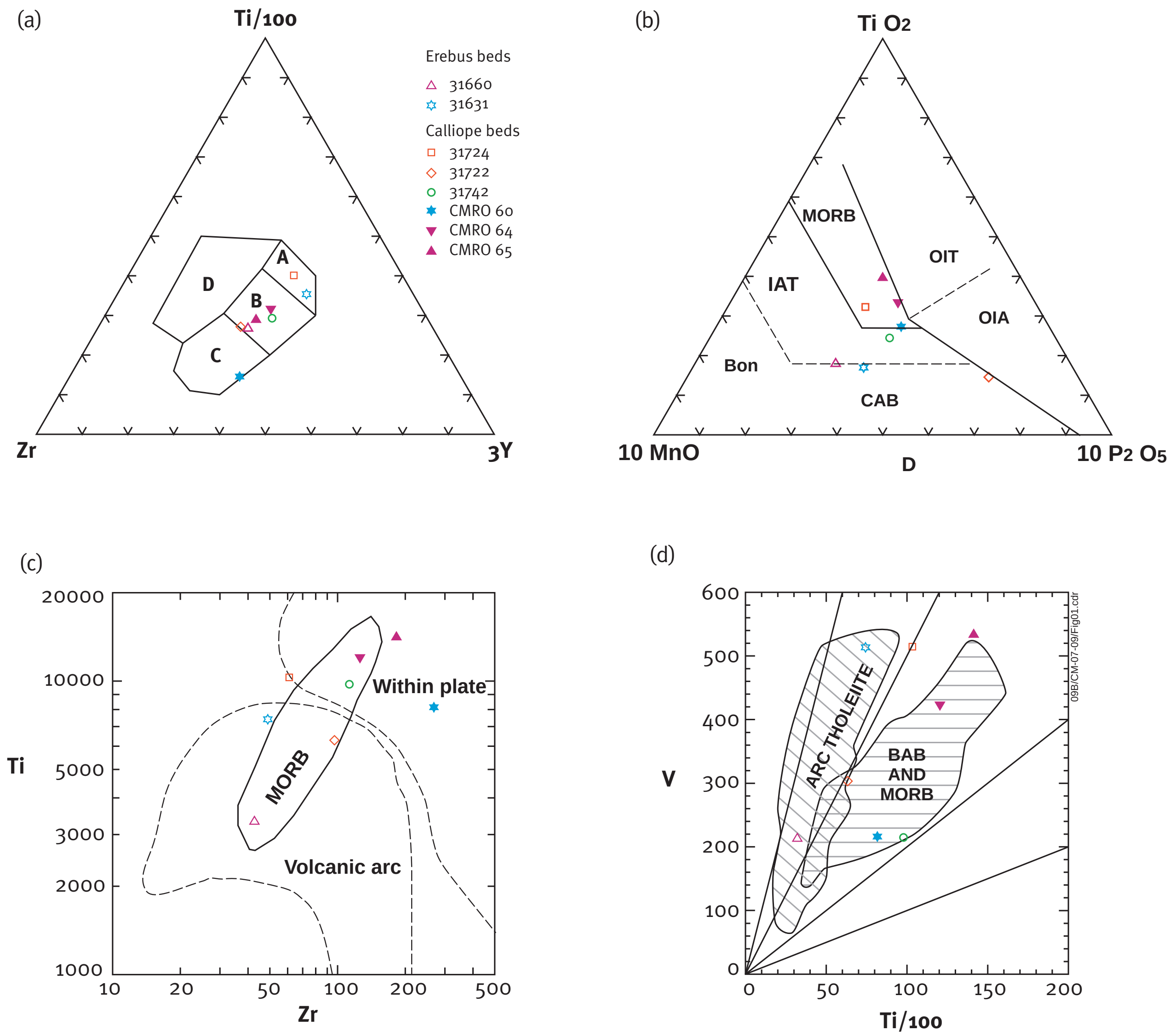


Figure 11. Analyses of basalts and andesites from the Calliope and Erebus beds (Morand, 1993a; GSQ) plotted on discriminant diagrams of (a) Pearce & Cann (1973): A = island arc tholeiites, B = island arc tholeiites, volcanic arc basalts and mid-ocean ridge basalts or MORB, C = calc-alkali basalts, and D = within plate basalts); (b) Mullen (1983): IAT = island arc tholeiite, CAB = island arc calc-alkaline basalt); (c) Pearce (1982); and (d) Shervais (1982): BAB = back-arc basin basalts). Analyses from the Erebus beds fall in fields defining island arc tholeiites or volcanic arc basalts, whereas those from the Calliope beds, particularly from south of Many Peaks, tend towards back-arc basin basalt compositions.

Environment of deposition

The presence of photic zone fossils is consistent with deposition in a shallow marine environment. Also, given that it contains lavas and coarse volcaniclastic sediments, the unit is assumed to have formed close to volcanic vents. The proposed tectonic setting for the Calliope beds is an oceanic island arc and flanking back-arc basin.

Thickness

The thickness of the Calliope beds is not known. The unit is bounded by faults and locally repeated by bedding parallel thrusts.

Structure

Little is known about the structure of the Calliope beds, largely due to the massive nature of all rock types. Kirkegaard & others (1970) reported moderate dips to the east. Detailed conodont work on limestone from the Taragoola mine shows repetition of conodont zones, indicating that the unit is repeated by a bedding parallel fault, probably a thrust. Some outcrops have a cleavage that is broadly parallel to the regional cleavage found in the Erebus beds and Mount Alma Formation.

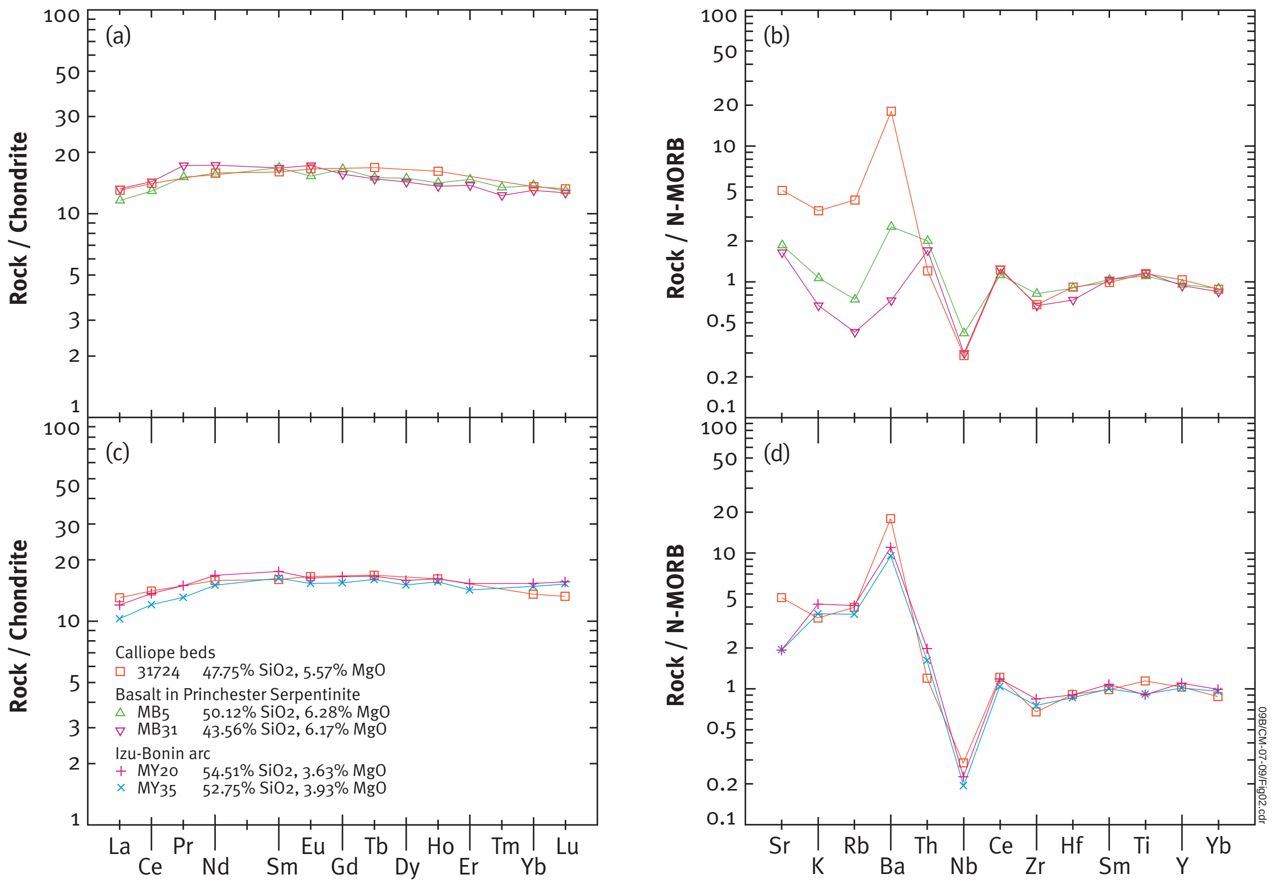


Figure 12. REE and spidergram plots comparing basalt 31724 from the Calliope beds (Morand, 1993a; Offler & Gamble, 2002) with (a,b) basalts MB5 and MB31 from the Princhester Serpentinite (Bruce, 1999), and (c,d) with basaltic andesites from the Izu-Bonin arc (Taylor & Nesbitt, 1998). Judging by the SiO_2 and MgO contents, the Izu-Bonin samples are more fractionated than the basalt from the Calliope beds. Chondritic REE values are from Sun & McDonough (1989), and N-MORB from Pearce (1983).

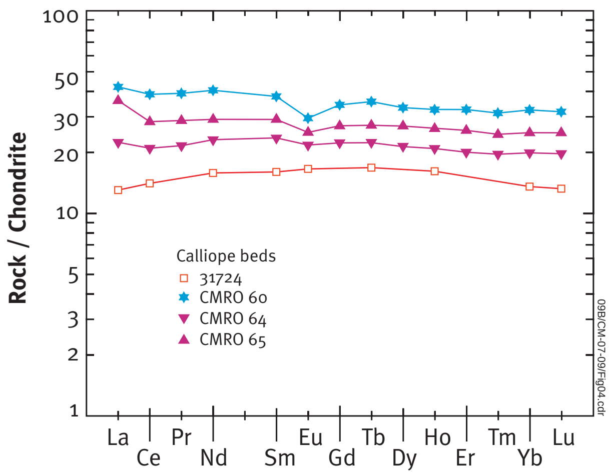


Figure 13. REE plots of basalts and andesites from the Calliope beds. The flat patterns indicate a thin, probably oceanic, crust. Chondritic REE values are from Sun & McDonough (1989).

Biostratigraphy and age

The Calliope beds is locally abundant in corals, but species diversity is low, with only *Aphyllum simplexum*, *Rhizophyllum parvum*, *Favosites forbesi*, and *F. grandipora* identified. This limited fauna indicated a general Early Devonian age but could not be more specific (Blake, 2010).

Conodont faunas indicate that the Calliope beds were deposited during the Early Devonian. Conodonts identified ranged from the *hesperius* zone to the *kindlei* zone indicating that the unit is mostly likely restricted to the Lockhovian to upper Pragian parts of the Early Devonian.

Stratigraphic relationships

The Calliope beds are bounded by faults over most of their length. Their base is not exposed. An unconformable contact with the Mount Alma Formation west of Gladstone was recorded by Kirkegaard & others (1970, pages 48–49) based on the occurrence of coralline limestone cobbles of Late Silurian to Early Devonian age. Conodont data confirm that limestone blocks at the base of the Mount Alma Formation in the Santa Glen area south of Mount Castletower are the same age as the limestones within the Calliope beds.

Correlation with other units

The Calliope beds are at least a partial time equivalent of the Erebus beds and Craigilee beds. However, the Calliope beds are distinguished from these other units by differences in rock types, geophysical signature, and basalt geochemistry (Murray & Blake, 2005).

These distinct differences support the idea that the Late Silurian to Middle Devonian units were not directly related to each other. Murray (2007) interpreted the Calliope beds as a remnant arc split off from the Capella Creek Group and Erebus beds by back-arc rifting and spreading above an east-dipping subduction zone.

Economic significance

Thick limestone lenses at Taragoola south-south-east of Calliope have been mined for various uses, including railway ballast and dusting in coal mines, but these deposits will be flooded by raising of the Awoonga dam.

Erebus beds (SDe)

(P.R. Blake)

Introduction

Kirkegaard & others (1970) mapped an extensive area of uppermost Silurian to Middle Devonian rocks, the Mount Holly beds, to the east of and in contact with the Capella Creek Group. Morand (1993a) changed the name to Mount Holly Formation and included the Calliope beds of Kirkegaard & others (1970) and Dear & others (1971) within the unit. Re-mapping with the aid of airborne geophysical survey data has revealed that most of the former Mount Holly beds is of Late Devonian and earliest Carboniferous age, and has been mapped as the Mount Alma Formation (Yarrol Project Team, 1997). The much smaller remaining outcrop of older rocks has been named the Erebus beds. The Calliope beds have been retained as a separate unit because they are very different in rock type and geophysical signature from the Erebus beds.

Distribution

The Erebus beds occur as a belt of rocks extending south-east from Raglan on the Bajool 1:100 000 Sheet area to Catfish Creek on the Calliope 1:100 000 Sheet area (Figure 14). Its extent is much less than that of the Mount Holly beds of Kirkegaard & others (1970).

Derivation of name

The Erebus beds are named after Mount Erebus which occurs at GR274796 7368496 on the Bajool 1:100 000 Sheet area within typical rocks of the unit.

Type section

No formal type section was measured for the Erebus beds, but the type area is approximately 3km west of Bracewell in the headwaters of Scrubby Creek on the Bajool 1:100 000 Sheet area, between GR282306 7355986 and GR280906 7356086. Most of the measured section described by Kirkegaard & others (1970)

Figure 14. Distribution of the Erebus beds

from Epala railway siding to Upper Horrigan Creek lies within the unit, but fault slices of Mount Alma Formation are intercalated with the Erebus beds in a complex structural zone at the western end between the Ulam and Queenslander Faults.

Topographic expression

The Erebus beds typically form hilly topography. The main creeks run approximately perpendicular to the regional cleavage, while most tributaries are parallel, giving the drainage an orthogonal appearance on airphotos.

Geophysical expression

The Erebus beds have a distinctive response on airborne geophysical survey images compared to adjacent units. They display a dark red colour indicating an elevated potassium content. The contrast with the very dark tones of the surrounding Mount Alma Formation enables the contact between the two units to be mapped with a high level of confidence. On aerial magnetic images the Erebus beds have a somewhat variable moderate to low signature, higher than that of the Mount Alma Formation. Most of the magnetic highs are parallel or sub-parallel to the regional cleavage and faults. Some of the visible magnetic linears correlate with mapped faults, and are possibly due to dykes intruding into planes of weakness.

Lithology

Outcrops seen within the Erebus beds are sandstone, conglomerate, breccia, minor limestone, and rare siltstone and rhyolite. The sandstones, breccias and conglomerates are volcaniclastic, sourced from dacitic and rhyolitic lavas. Clasts of basaltic to andesitic provenance are rare.

The sandstones are typically medium to very coarse-grained and poorly sorted; fine-grained sandstones are rare. They are mostly lithic, but some sandstones contain a significant proportion of feldspar grains. Quartz is present in most samples and usually makes up several percent of the rock. Bedding is rarely seen in sandstone sequences, particularly in the western part of the unit where deformation appears to be more intense. In the

Figure 15. Matrix-supported conglomerate containing clasts of pale felsic volcanics in a matrix of fine-grained sandstone, Erebus beds. The outcrop displays a sub-vertical cleavage. Field point PBRO 432, 4.5km east of Bracewell at GR290059 7355875 on the Bajool 1:100 000 Sheet area.

eastern part of the outcrop area, particularly around Raglan, the deformation appears to be weaker, thin to medium bedding is visible, and one outcrop displayed planar cross-bedding.

Most conglomerates and breccias are granule to cobble in grainsize and clast-supported, but matrix-supported examples do occur (Figure 15). Clasts are dominantly of dacitic to rhyolitic lava (see Figure 8b of Morand, 1993a). Kirkegaard & others (1970) mentioned that the Mount Holly beds contained rhyolitic to dacitic ash-flow tuffs, but these rocks are almost certainly volcaniclastic sediments rather than primary volcanics. It is possible that they were referring to the conglomerates, particularly the matrix-supported varieties, and that deformation had flattened some of the clasts giving them a fiamme-like appearance, making the overall rock resemble an ash-flow tuff. However, the rocks are deformed, and thin section examination found no trace of glass shards in any of the rocks collected.

Limestone is a minor component of the Erebus beds, but locally forms some very large outcrops. Most of the limestones show some level of recrystallisation, ranging from minor to almost entirely recrystallised marbles. The finer-grained limestones contain crinoids, corals, stromatoporoids, and brachiopods, and rare crinoid stems are even preserved in the marbles. The limestones are usually wackestones with the fossils suspended within micrite.

The siltstones are strongly cleaved and it is difficult to decipher sedimentary structures. Dacites and rhyolites are sparely to moderately porphyritic with phenocrysts of feldspar and minor quartz. Outcrop is insufficient to determine whether they are lavas or high-level intrusions.

Geochemistry

Morand (1993a) analysed two basalts (31631 and 31660) from the area now mapped as Erebus beds, although he was uncertain whether these represented flows or intrusions. He used standard discriminant plots to show that these samples formed in a volcanic arc setting (Figure 11). On the basis of additional REE data, Offler & Gamble (2002) interpreted sample 31660 as an island arc-type basalt, consistent with its lower content of LREE and some HFSE compared to basalts from primitive continental arcs such as the Southern Volcanic Zone of the Andes and the Eastern Volcanic Front of Kamchatka (Murray & Blake, 2005). In its geochemical

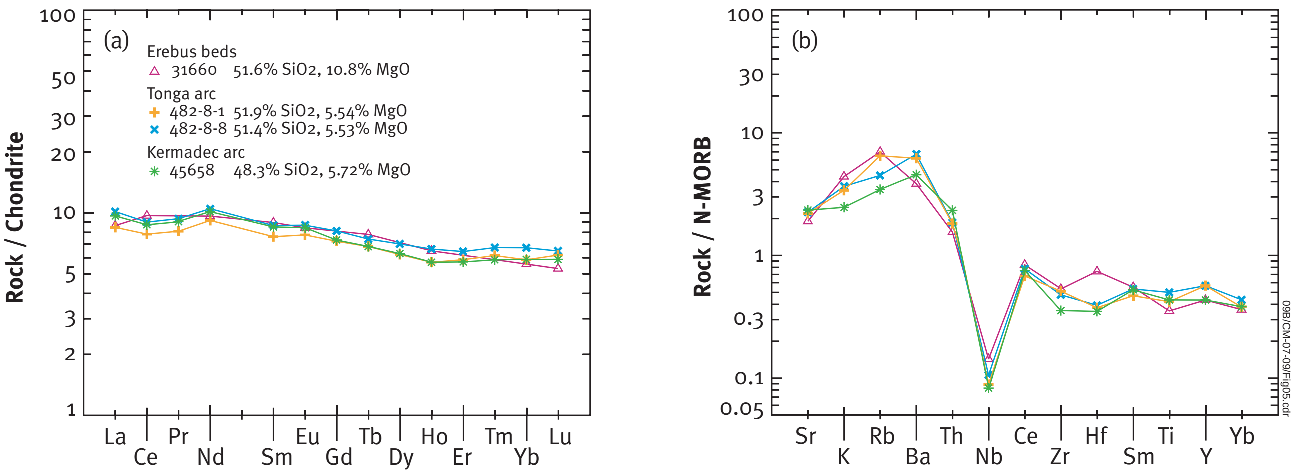


Figure 16. (a) REE and (b) spidergram plots comparing basalt 31660 from the Erebus beds (Morand, 1993a; Offler & Gamble, 2002) with basalts from the Tonga-Kermadec arc (Turner & others, 1997). The basalt from the Erebus beds has much higher MgO, and may be less fractionated, than the samples from the Tonga-Kermadec arc. Chondritic REE values are from Sun & McDonough (1989), and N-MORB from Pearce (1983).

pattern, sample 31660 resembles basalts from the Tonga-Kermadec arc (Figure 16). The REE pattern for sample 31660 conclusively demonstrates that, if it is an intrusive rather than a lava, it is not of Permian to Triassic age like many dykes in the region, and its geochemistry probably represents the true tectonic setting of the Erebus beds.

Environment of deposition

The Erebus beds were deposited in a marine environment. The presence of fossils such as corals suggests that it was within the photic zone, and the abundance of volcanic material indicates that it was proximal to a volcanic source that was probably an arc based on limited geochemistry. Therefore, the most likely environment of deposition for the Erebus beds is in a shallow marine environment near an island arc.

Thickness

Neither the base nor top of the Erebus beds is exposed, and the unit is folded and complexly faulted. Therefore it is not possible to give a reliable figure for the thickness of this unit. Some indication may be provided by the type section of the Mount Holly beds measured as 2600m by Kirkegaard & others (1970), most of which lies within the mapped Erebus beds.

Structure

The Erebus beds are folded and faulted and, in most areas, strongly cleaved. The cleavage has a consistent north-north-west strike, and moderate to steep dips to the east. It is best developed in the western part of the outcrop area in the vicinity of major faults, contrary to the conclusion of Morand (1993b) that folds become tighter and cleavage intensifies from west to east towards the Yarrol Fault. Bedding is generally steep.

Faults appear to be east-over-west thrusts, and the structures are consistent with the interpretation of Morand (1993b) that the unit is part of a west-vergent fold and thrust belt. Along their western margin, the Erebus beds are thrust over the Mount Alma Formation, and a belt trending south-south-east from Marmor between the Ulam and Queenslander Faults is a zone of complex fault imbrication in which it is not possible to map out the two units on a regional scale.

Although most structures are consistent with formation by a single fold-thrust event, evidence for later out-of-sequence thrusting is provided by the Bracewell Fault just north of Bracewell, where folded Mount Alma Formation has been thrust over the older Erebus beds.

Biostratigraphy and age

The Erebus beds have provided good faunas of both corals and conodonts. Kirkegaard & others (1970) concluded that limestones of two ages were present (latest Silurian to earliest Devonian and late Early to earliest Middle Devonian). The age range they suggested has been confirmed by subsequent conodont studies. More detailed collecting has yielded conodont faunas ranging from the *woschmidti* zone to the *costatus* zone, indicating that they range from upper Přídolí (latest Silurian) to lower Eifelian (earliest Middle Devonian). The deposition of the Erebus beds covered the interval from at least Late Silurian to early Middle Devonian.

The coral faunas appear to be more restricted in its time range and contains *Aphyllum simplexum*, *Pseudamplesux princeps*, *Mictocystis*? sp. cf. *M. endophylloides*, *Dohmophyllum* sp. *Embolophyllum* sp. *Blysmatophyllum* sp. *Papiliophyllum jelli*, P. sp. *Thamnophyllum* sp., *Favosites careyi*, *F. duni*, *F. gothlandicus*, *F. goldfussi*, *Favosites grandipora*, *Squameofavosites bryani*, *Heliolites amplusa*, *H. daintreei*, *H. comminus*, *Romingeria* sp. cf. *R. emergens*, and *Syringopora jonesi*, indicating an Early Devonian (Emsian) age (Blake, 2010).

The deposition of the Erebus beds covered the interval from at least Late Silurian to early Middle Devonian.

Stratigraphic relationships

The base of the Erebus beds is not exposed. The unit is surrounded and unconformably overlain by the Mount Alma Formation, although many contacts are faulted.

Correlation with other units

The Erebus beds are at least partial time equivalents of the Craigilee beds, Calliope beds and probably of the lower part of the Capella Creek Group. Despite the similarities in age and their proximity to each other, these four units are different in rock types, geophysical expression, basalt geochemistry and coral faunas (Murray & Blake, 2005). Therefore, the Erebus beds are considered to be a separate terrane, possibly not directly related to any of the other units of similar age within the area. Murray (2007) interpreted the Erebus beds as a remnant arc split off from the Middle Devonian Capella Creek Group to the west by back-arc rifting and spreading above an east-dipping subduction zone.

Economic significance

Limestone at East End is mined for cement production at Gladstone.

Capella Creek Group

(P.R. Blake & A. Taube)

These rocks (Figure 17) were named the Capella Creek beds by Kirkegaard & others (1970), who described them as a heterogeneous unit of acid to intermediate volcanics, tuffaceous lithic sandstone, mudstone, and limestone, and assigned a late Middle Devonian (Givetian) age. It was recognised that the lower part of the sequence consisted mainly of acid tuffs and the upper part comprised tuffaceous sediments, limestone, and andesitic rocks with minor acid volcanics, but the boundary between these two informal subdivisions was not mapped.

In his figure 1, Taube (1986) subdivided the Capella Creek beds, and differentiated the lower silicic section as the Mount Warner Volcanics. He retained the name Capella Creek beds for the upper part only. Morand (1993a) reverted to the usage of Kirkegaard & others (1970), including the Mount Warner Volcanics in a single unit that he renamed the Capella Creek Formation. The results of detailed company mapping along the Dee Range were summarised by Messenger & Taube (1994). They named the base of the sequence the Mount Dick beds, and subdivided the overlying Mount Warner Volcanics into three informal units, the footwall sequence, banded mineralised sequence, and hanging wall sequence. They also correlated the Mount Warner Volcanics with the host rocks of the Mount Morgan Au-Cu deposit, an isolated roof pendant within the Mount Morgan Trondhjemite that for many years had been referred to as the Mine Corridor Volcanics. The name Capella Creek beds was retained for the more mafic upper part of the succession.

Because the change in the usage of the name Capella Creek beds from the original definition by Kirkegaard & others (1970) would create confusion, and to take account of the results of subsequent detailed mapping, the unit was upgraded to Group status by the Yarrol Project Team (1997). The new name Raspberry Creek Formation was applied to the mafic upper part shown as Capella Creek beds by Taube (1986) and Messenger & Taube (1994).

Most of the current knowledge on this unit has been acquired from detailed company mapping over the last 20 years, mainly by teams led by A. Taube, with contributions from the Yarrol Project Team. Three formations and one member have been differentiated. From the base to the top, the three formations are the Mount Dick beds, Mount Warner Volcanics, and Raspberry Creek Formation. The Mount Dick beds and the Mount Warner Volcanics were the units recognised in the past as the lower silicic sequence, and the Raspberry Creek

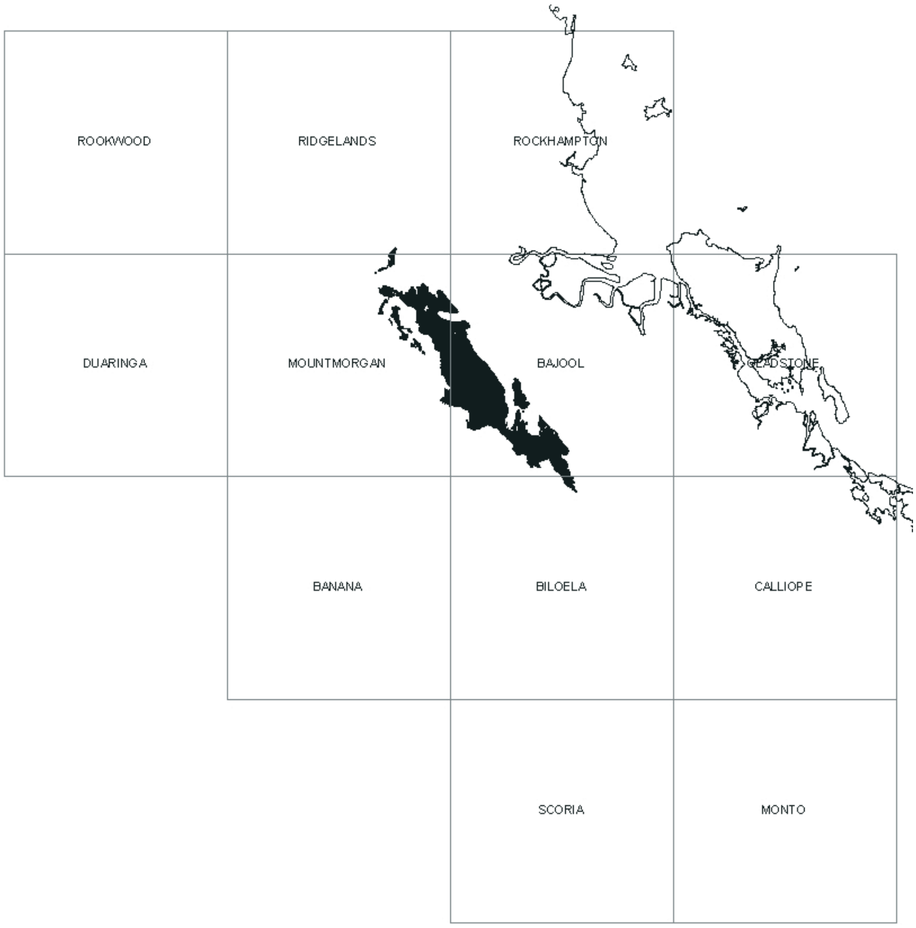


Figure 17. Distribution of the Capella Creek Group

Formation is the upper sequence containing andesitic to basaltic rocks. The Ginger Creek Member is a sequence of siliceous volcanics locally differentiated at the top of the Raspberry Creek Formation.

The Capella Creek Group, together with the Marble Waterhole beds, comprise the Mount Morgan Stratigraphic Assemblage, a revision from the original concept outlined by the Yarrol Project Team (1997).

Mount Dick beds (Dcd)

(P.R. Blake & A. Taube)

Introduction

The Mount Dick beds are the lowest unit of the Capella Creek Group. They were named and described in detail by Messenger & Taube (1994).

Distribution

The Mount Dick beds occur as a belt of rocks 13km long and up to 4km wide extending along the Dee Range from Station Creek to the south-east (from GR246100 7382200 to GR252100 7372200 on the Bajool 1:100 000 Sheet area) (Figure 18).

Derivation of name

The name is taken from the peak locally known as Mount Dick on the Dee Range 6km south of Station Creek (GR245300 7377650 on the Bajool 1:100 00 Sheet area). Although the summit of Mount Dick is within the footwall sequence of the Mount Warner Volcanics, the best exposures of the Mount Dick beds are located on its north-eastern slopes (Messenger & Taube, 1994). On the Mount Morgan 1:100 000 topographic map, the name Mount Dick is applied to a peak 1.5km to the west-north-west, but according to local property owners, this is actually Mount Warner.

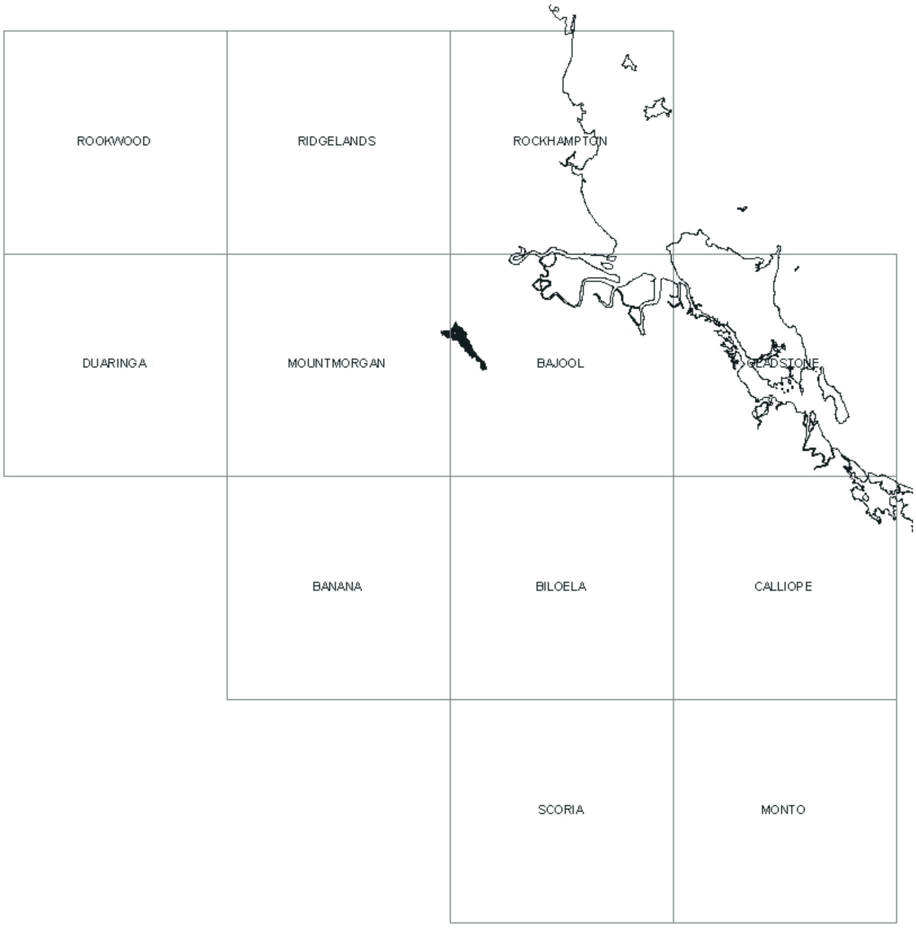


Figure 18. Distribution of the Mount Dick beds

Type section

No type section has been measured for the Mount Dick beds, although Messenger & Taube (1994) show two representative stratigraphic columns. The unit crops out poorly, and is strongly hornfelsed. A type area for the unit is a track up a spur of the Dee Range east of Mount Dick beginning at GR248500 7378450 on the Bajool 1:100 000 Sheet area.

Topographic expression

The Mount Dick beds form the steep eastern escarpment of the Dee Range, with rugged topography and deeply incised creeks.

Geophysical expression

No reliable airborne radiometric signature can be established for the Mount Dick beds due to the poor response resulting from the topographic relief along the eastern slope of the Dee Range. The unit appears to have a very low magnetic signature.

Lithology

The Mount Dick beds consist mainly of sediments with interbedded felsic lavas. Mapping indicates that sediments are dominant in the north, and lavas in the south. These interfinger east of Mount Dick, and may be in part lateral facies variants.

The lavas are fine-grained dacites with sparse feldspar and quartz phenocrysts, and local development of hyaloclastites (Messenger & Taube, 1994). In the type area towards the northern end of the unit, the most common rock-type seen was siltstone with minor sandstone, basaltic to dacitic lava, and rare skarns. Many of the siltstones have been silicified, and some are spotted hornfels due to metamorphism by the Bajool Quartz Diorite. They rarely display bedding. Sandstones range from fine- to very coarse-grained and are generally poorly sorted. They are volcaniclastic, dominated by lithic clasts from felsic lavas, with lesser quartz and feldspar.

Environment of deposition

The environment of deposition of the Mount Dick beds cannot be determined with certainty. Rare skarns possibly indicate the presence of limestones or calcareous sandstones, suggestive of a marine environment. No fossils have been found, and no characteristic bed-forms have been identified. However, because it is the lowermost unit of a Group that is interpreted to have been deposited in an island arc environment, and has broadly the same composition as the Mount Warner Volcanics, a similar depositional setting is assumed.

Thickness

The base of the Mount Dick beds is truncated by the Bajool Quartz Diorite, so the true thickness cannot be determined. Stratigraphic columns presented by Messenger & Taube (1994) show thicknesses of 500–600m.

Structure

The Mount Dick beds form the core of the regional north-north-west trending Gracemere Anticline, a broad structure with gently dipping limbs.

Biostratigraphy and age

No fossils have been found in the Mount Dick beds, but it underlies and is conformable with the Mount Warner Volcanics, which contain Middle Devonian corals and conodonts. The unit is probably also of Middle Devonian age, but may range into the Early Devonian.

Stratigraphic relationships

The Mount Dick beds are interpreted to be conformably overlain by the Mount Warner Volcanics. The base has been intruded by the Permo-Triassic Bajool Quartz Diorite.

Mount Warner Volcanics (Dcw)

(P.R. Blake & A. Taube)

Introduction

The Mount Warner Volcanics are the middle formation of the Capella Creek Group, and are part of the suite of rocks that Kirkegaard & others (1970) informally referred to as the lower part of the Capella Creek beds sequence containing silicic tuffs.

The Mount Warner Volcanics occur in two separate outcrop areas: the larger extends along the crest of the Dee Range, and the smaller, known for many years as the Mine Corridor Volcanics, is a roof pendant in the Mount Morgan Trondhjemite (Figure 19). Taube (1986), Messenger & Taube (1994) and Taube & others (2005) have subdivided both sequences and proposed correlations between them (Table 1).

The Mount Warner Volcanics are dominantly sourced from low-K rhyolitic volcanics, but material sourced from andesitic and basaltic volcanics occurs at the base of the formation, and sparse to common tonalite and trondhjemite clasts occur throughout.

Table 1. Correlations between successions in the Mount Warner Volcanics in the Mine Corridor and Dee Range

Mine Corridor Volcanics	Dee Range
Upper mine sequence (800+m)	Hanging wall sequence (35–200m)
Banded mine sequence (400m) Jasper sequence Dome complex	Banded mineralised sequence (390–750m) Crenulated cherty lavas
Lower mine sequence (650+m)	Footwall sequence (150–200m)

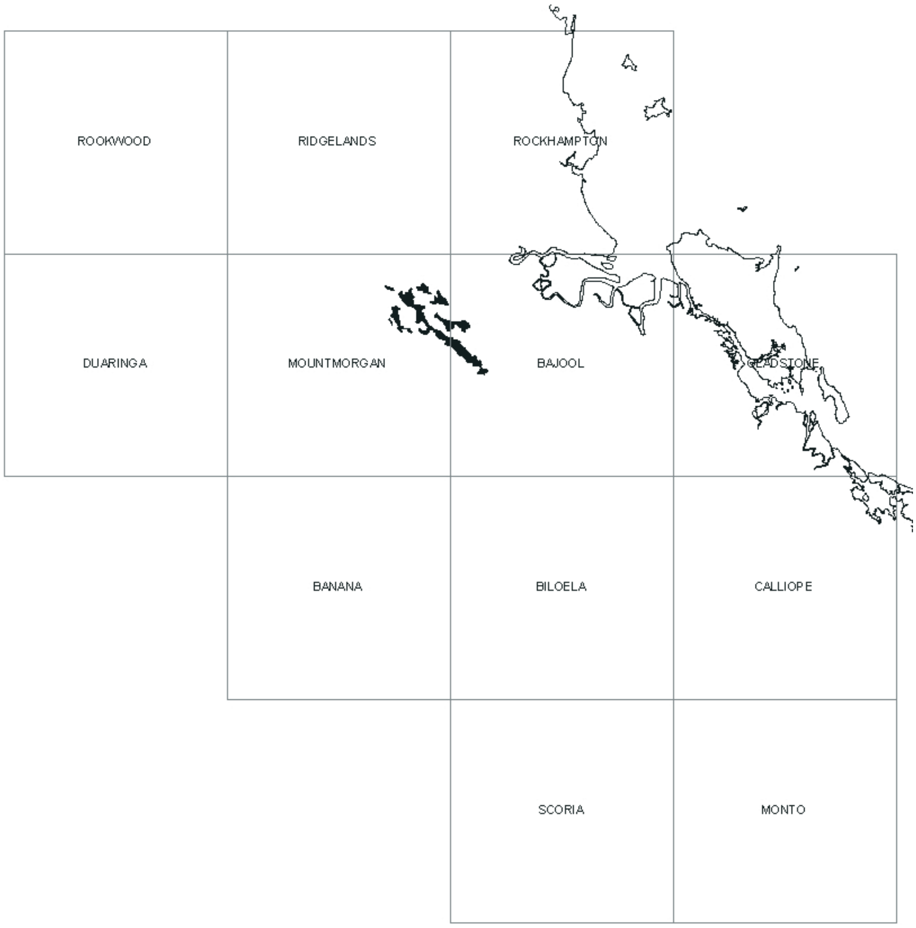


Figure 19. Distribution of the Mount Warner Volcanics

Distribution

The main outcrop of Mount Warner Volcanics is a south-east trending belt 30km long and up to 4km wide extending along the Dee Range from Moongan to south of Mount Hopeful. Isolated outcrops occur to the west of Bouldercombe, in the headwaters of Quarry Creek west of Moonmera, and as a roof pendant 7.5km long and up to 1.5km wide within the Mount Morgan Trondhjemite. The roof pendant hosts the Mount Morgan mine and has been named the Mine Corridor Volcanics in the literature.

Derivation of name

This formation was named by Taube (1986) after the peak locally known as Mount Warner, located within the Mount Warner Volcanics at GR244000 7378100 on the Mount Morgan 1:100 000 sheet. On the 1:100 000 topographic map, this mountain is labelled Mount Dick.

Type section

The type section is along the road leading to the Mount Hopeful TV towers, where the unit is well exposed and is almost 1000m thick.

Topographic expression

The Mount Warner Volcanics occur in the Dee Range, and form high, rugged topography. Creeks in the area are usually deeply incised with steep sides.

Geophysical expression

The Mount Warner Volcanics give a low radiometric response, although the data are of poor quality over the Dee Range outcrop because of the rugged topography and steep slopes. The low response reflects the low-K nature of all rock types in the sequence. The unit is also only weakly magnetic.

Lithology

The succession of rock types in the Mount Warner Volcanics in both the Mount Morgan and Dee Range areas has been described by Taube (1986), Messenger & Taube (1994) and Taube & others (2005).

In the Dee Range, the basal unit of the Mount Warner Volcanics is called the Footwall Sequence or Footwall Andesite (Table 1). This is dominated by sandstone and granule to pebble breccia (including hyaloclastites) sourced from andesitic and basaltic volcanics. Andesitic lavas have also been reported, and small gabbroic intrusions are common. This is the only unit within the Mount Warner Volcanics that is dominated by mafic compositions.

The Banded Mineralised Sequence contains a variety of rock types including sandstone, jasper, siltstone and rare peperite that can occur interbedded with each other. The sandstones range from fine- to coarse-grained and some are graded. They are dominantly sourced from rhyolitic volcanics, but some material is also sourced from granitoids as shown by clasts of graphically intergrown quartz and feldspar. The siltstones are light grey to purple in colour and generally laminated. The jaspers are well bedded to laminated, and commonly contain manganese nodules. The peperites were formed by rhyolite flowing into siltstone.

The lower part of the Banded Mineralised Sequence is dominated by the Crenulated Cherty Lavas, a name given to a unit of aphyric, flow banded rhyolitic lava. In many places the rhyolite has been silicified and has a ‘cherty’ appearance.

The Hanging Wall Sequence consists of sandstone and conglomerate with rare limestone. The sandstones are medium- to coarse-grained and contain abundant rhyolite clasts and feldspar and minor quartz and granitoid fragments. The conglomerates are mainly granule to pebble in grainsize, and most are poorly sorted and matrix supported. The conglomerates contain clasts of dacitic and rhyolitic volcanics, including pumice. These features give the conglomerates a tuffaceous appearance and they have been described in the past as ignimbrites. However, these rocks are now interpreted as ignimbrites that flowed into water and became debris flows. The limestones form small lenses and contain fossils of crinoids, corals and brachiopods.

Areas of undivided Mount Warner Volcanics are usually dominated by sandstones and conglomerates that resemble rocks within the Banded Mineralised and Hanging Wall Sequences.

In the Mount Morgan roof pendant, the rocks formerly named the Mine Corridor Volcanics consist mainly of massive crystal-rich pumiceous volcaniclastic mass-flow deposits (Messenger & others, 1997), with subordinate jasper, tuffaceous mudstone, siltstone and sandstone, and limestone lenses and bands. Quartz-feldspar porphyries with a fine-grained rhyolitic matrix may be domes or lava flows. Detailed mapping over the long history of mining at Mount Morgan has interpreted and subdivided the sequence in various ways (see Table 1).

Geochemistry

Cornelius (1968) carried out the first analyses of the Mount Warner Volcanics, from the vicinity of the Mount Morgan mine. He found that they were similar in geochemistry to the Mount Morgan Trondhjemite, particularly noting their low-K nature (Na_2O/K_2O ratios of 7.1 to 19.4), and concluded that the volcanic and plutonic rocks were related (Cornelius, 1969).

Messenger (1996) carried out detailed geochemical studies of both the Mount Warner Volcanics and Mount Morgan Trondhjemite at Mount Morgan. He concluded that they formed a co-genetic magmatic assemblage similar to the modern Kermadec arc: “The Mt. Morgan volcano-plutonic suite has mineralogical and geochemical affinities with low-K island arcs such as the Tonga-Kermadec arc, ie. high K/Rb, Rb/Zr and Sr/Nd, but low Rb/Sr, La/Yb and Ti/V. Trace element abundances of basaltic and gabbroic rocks are below those of average mid-ocean ridge basalt (MORB), being similar to the Kermadec section of the Tonga-Kermadec island arc. All rocks have low overall abundances of large ion lithophile (LIL) and high field strength (HFS) elements, and show LIL/HFS decoupling. The strong negative Nb anomaly in all rock types, and complimentary (*sic*) negative Ti anomaly in the felsic rock group, are features typically observed in modern subduction-related magmatic rocks. Strontium, neodymium and oxygen isotope compositions straddle the mantle array and overlie the fields for modern, SW Pacific island arcs with no evidence for the involvement of an evolved crustal source, ie. I_{Sr} = 0.70362 to 0.70431, $å_{Nd}$ (381Ma) = 6.3 to 9.0, and ä^{18}O = 6.1‰ to 8.1‰” (Messenger 1996, page iii). For the basaltic rocks, which were hypabyssal intrusives into the dominantly felsic volcanics, it should be noted that although the overall abundance of LIL elements is low, these elements are enriched relative to N-MORB, in contrast to the HFS elements that are moderately to strongly depleted.

Murray & Blake (2005) discussed the tectonic implications of the geochemical results for the Mount Warner Volcanics and Mount Morgan Trondhjemite. They favoured a setting in a primitive oceanic arc with an associated back-arc basin, such as the modern Tonga-Kermadec, Izu-Bonin, New Britain, and South Sandwich arcs.

The chemistry of exhalites within the Mount Warner Volcanics has been discussed by Peterson & Golding (1999, 2001).

Environment of deposition

The Mount Warner Volcanics were deposited in a marine environment, and given the coarse-grained nature and dominantly volcanic provenance of the sedimentary rocks, as well as the presence of rhyolitic lava and peperite, it is interpreted that they were deposited close to a volcanic source. Geochemical results suggest that this source was a primitive oceanic island arc (Murray & Blake, 2005), although Messenger (1996) and Messenger & others (1997, 1998) favoured an arc with thin continental crust located adjacent to the Australian continent.

Messenger & others (1997) proposed that the Mount Warner Volcanics in the Mine Corridor formed in a small fault-bounded extensional basin above a subvolcanic trondhjemitic magma chamber. A different model was put forward by Murray & Blake (2005). They suggested that the Mount Warner Volcanics were erupted from submarine felsic volcanoes like those in a number of modern island arcs, and that the Mine Corridor sequence at Mount Morgan was deposited in a large collapse caldera. The size of the modern submarine calderas, up to several kilometres across, is sufficient to incorporate the whole of the complex geology around the Mount Morgan mine. This model is similar to that proposed by Taube (1986) for the setting of the Mount Morgan Au-Cu deposit.

Thickness

The Mount Warner Volcanics are almost 1000m thick in the type section along the road to Mount Hopeful. The thickness is considerably greater at Mount Morgan, where up to 1900m of sequence is present (Messenger & Taube, 1994).

These thickness variations are consistent with the caldera model proposed by Murray & Blake (2005). The pyroclastic unit overlying the mineralised sequence at Mount Morgan (the Upper Mine Sequence) is at least several hundred metres thick. Along the Dee Range the equivalent unit, the Hanging Wall Sequence, is ~200m thick, becoming progressively thinner to the south-east (Messenger & Taube 1994). The much greater thickness of this unit at Mount Morgan is interpreted as intra-caldera fill, and the thinner deposits of the Dee Range as outflow facies.

Structure

The Mount Warner Volcanics dip at shallow to moderate angles to the south-west along the western limb of the Gracemere Anticline, although later faulting has produced variations in dip directions. In general they are only weakly deformed, but locally a distinct cleavage is developed, and clasts have been stretched, as in the Foot Wall sequence along the Dee Range (Messenger & Taube, 1994).

The southern part of the Mine Corridor sequence is folded into an open, gently plunging anticline. The northern part dips to the north-east at ~60°. The Mount Morgan Trondhjemite cuts these structures and appears to be unaffected by them, suggesting that the deformation was at least in part of latest Middle or earliest Late Devonian age (Taube, 1986; Golding & others, 1994a). Local doming around the Mount Morgan orebody was also of pre-Mount Morgan Trondhjemite age (Frets & Balde, 1975).

Biostratigraphy and age

Corals are fairly sparse in the Mount Warner Volcanics with only the following species identified, *Phillipsastrea*? *carinata*, *Temnophyllum stanesi*, *Tabulophyllum* sp., *Favosites goldfussi*, and *Alveolites suborbicularis* (Blake, 2010). This coral fauna does not provide a precise age of the unit, however, conodonts give a Middle Devonian (Eifelian) age for the Mount Warner Volcanics. Taube & others (2005) reviewed the conodont data, and concluded that the Banded Mine Sequence at Mount Morgan and the Banded Mineralised Zone of the Dee Range were equivalent, and both of late Eifelian (early Middle Devonian) age.

Stratigraphic relationships

The Mount Warner Volcanics are interpreted to overlie the Mount Dick beds conformably, and are conformably overlain by the Raspberry Creek Formation.

Figure 20. Water-filled open cut of the Mount Morgan gold-copper mine, viewed from the west

The sedimentary component of the Mount Warner Volcanics is similar to that in the Mount Dick beds, but the Mount Dick beds are dominated by siltstone with only minor sandstone, whereas relatively coarse-grained sandstone and conglomerate/breccia makes up a high proportion of the Mount Warner Volcanics.

Economic significance

The world-class Mount Morgan Au-Cu deposit (Figure 20) is hosted by a roof pendant of Mount Warner Volcanics within the Mount Morgan Trondhjemite. The orebody has been viewed either as an unusual pipe-like volcanic hosted massive sulphide (VHMS) deposit (Taube, 1986; Large, 1992), or as a replacement deposit related to emplacement and fractionation of the Mount Morgan Trondhjemite (Arnold & Sillitoe, 1989). These different models are important not only for understanding the origin of the Mount Morgan mineralisation, but also for predicting where similar deposits may occur. The host rocks contain unusual spherical objects which for many years were called hurgledurgles (Figure 21), but which have been identified as contact metamorphosed assemblages of chloritic alteration, including cordierite (now altered) and Ca-poor amphibole (England & others, 1998).

Reviews by Golding & others (1994a,b) supported a sub-sea floor replacement origin involving fluids derived from a combination of seawater and magmatic sources. They also confirmed an earlier observation by Lawrence (1972) that the Au-Cu orebody had been recrystallised, probably by the Mount Morgan Trondhjemite which must therefore have post-dated the main mineralisation. From his detailed study of the deposit, Messenger (1996) concluded that large-scale hydrothermal circulation responsible for the Au-Cu mineralisation was associated with submarine low-K volcanism, and that the Mount Morgan Trondhjemite was intruded during the waning stages of this circulation. While it seems unlikely that a hydrothermal system would continue to operate over the period between eruption of the Mount Warner Volcanics and intrusion of the Mount Morgan Trondhjemite (about 15my according to the time scale of Gradstein & others, 2004), there is clear evidence of ongoing magmatic activity of trondhjemitic composition during part of that time.

Precise determinations of Pb isotopes by Ulrich & others (2001, 2002) suggest that the Mount Morgan Trondhjemite was the source of the ore metals, rather than the volcanics that host the orebody. In view of the age relations, it is more likely that the source was a similar but earlier magma chamber of trondhjemitic composition.

Murray & Blake (2005) drew an analogy between the Mount Morgan orebody and a gold-rich sulphide deposit currently forming on the floor of the Myojin Knoll caldera in the Izu-Bonin arc (Fiske & others, 2001). This idea shares some common features with the model proposed by Taube (1986) involving cauldron subsidence and resurgence.

Figure 21. Contact metamorphosed assemblages of chloritic alteration in the host rocks of the Mount Morgan gold-copper deposit (England & others, 1998), previously called hurgledurgles, Mount Warner Volcanics.

Whatever model is accepted for Mount Morgan, either a composite model as suggested by Ulrich & others (2001, 2002) or a VHMS, the association with a long-lived felsic volcanic-plutonic centre in a submarine island arc environment seems indisputable. Such centres should therefore be regarded as exploration targets for Mount Morgan-style deposits.

Subeconomic stratabound mineralisation more characteristic of VHMS deposits occurs within the Mount Warner Volcanics in both the Dee Range and at Mount Morgan (Taube & McLeod, 1987; Taube & Messenger, 1994; Messenger & Taube, 1994; Perilya Mines NL, 1995).

Raspberry Creek Formation (Dcr)

(P.R. Blake & A. Taube)

Introduction

The Raspberry Creek Formation is the youngest formation within the Capella Creek Group. It comprises the sequence that Kirkegaard & others (1970) referred to as the upper part of the Capella Creek beds containing tuffaceous sediments, limestone, and andesitic rocks with minor acid volcanics.

Distribution

The Raspberry Creek Formation is by far the most extensive unit in the Capella Creek Group. It forms a belt 75km long and up to 14km wide extending south-south-east from near Kabra on the Ridgelands 1:100 000 Sheet area to Ayrdrie homestead in the northern Biloela 1:100 000 Sheet area (Figure 22).

Derivation of name

The Raspberry Creek Formation was named by the Yarrol Project Team (1997) after Raspberry Creek, which flows through the unit in its upper reaches ~15km south-east of Mount Morgan.

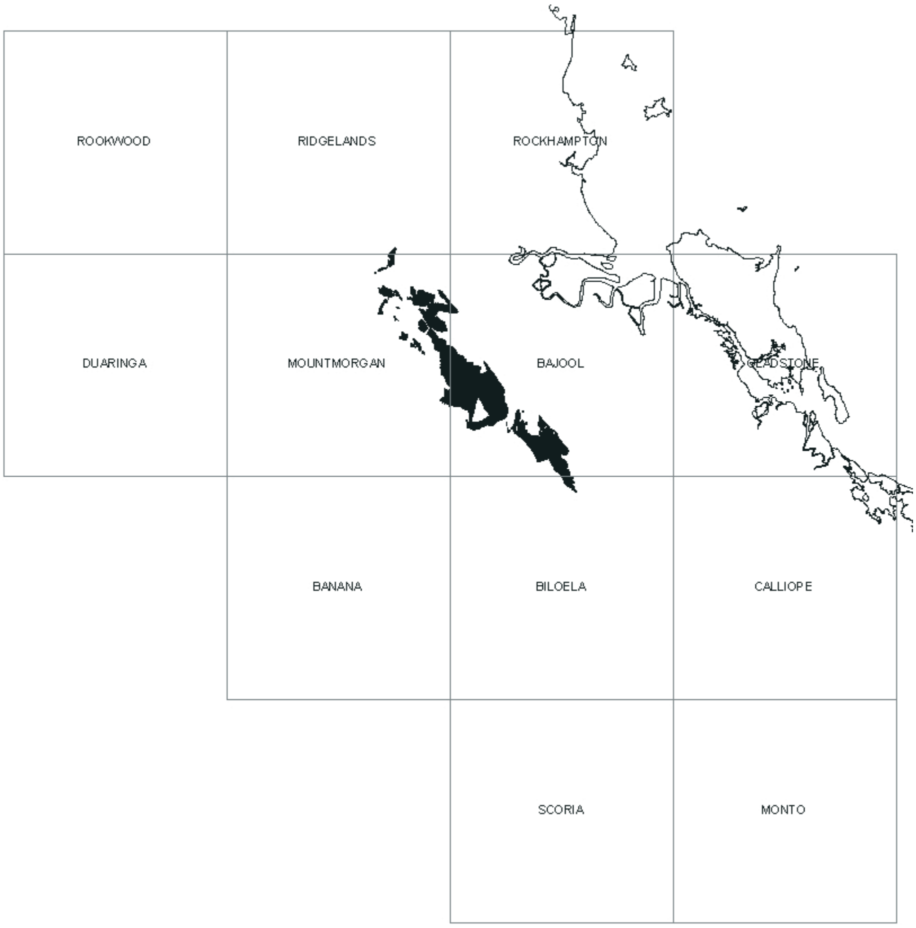


Figure 22. Distribution of the Raspberry Creek Formation

Type section

The type section was measured on the property "Belgamba", ~10km east-north-east of Mount Morgan. The section begins at GR243800 7387500 and finishes at GR244400 7387500 on the Mount Morgan 1:100 000 Sheet area and is 464m thick. It is possible that the Raspberry Creek Formation may be thicker in other parts of the Dee Range, but at this locality both the base and the top of the unit are exposed.

Topographic expression

The Raspberry Creek Formation crops out along the crest and western slopes of the Dee Range and forms hilly country dissected by steep-sided, deep creeks.

Geophysical expression

The Raspberry Creek Formation gives a low response for both radiometrics and magnetics, similar to the remainder of the Capella Creek Group, although the data are affected to varying degrees by the topographic relief of the Dee Range.

Lithology

The Raspberry Creek Formation consists of sandstone, conglomerate and breccia with local basalt lavas and minor siltstone and limestone, and rare accretionary lapilli tuff. Most of the sedimentary rocks within the Raspberry Creek Formation are volcaniclastic and sourced from basaltic and andesitic volcanics (Figure 23). Towards the top of the unit, clastic rocks sourced from dacitic and rhyolitic volcanics become more abundant. These usually are restricted in areal extent, and in most places are impossible to map out as a separate unit in a regional-scale mapping program. However, in the southern part of the Capella Creek Group they cover a relatively large area and have been mapped as the Ginger Creek Member of the Raspberry Creek Formation.

Most conglomerates and breccias are poorly sorted, and range from pebble to cobble in grainsize. Clasts of basaltic to andesitic volcanics are dominant with local limestone and granitoids ranging from granodiorite to

Figure 23. Pebble to cobble conglomerate composed of basalt to andesite clasts, Raspberry Creek Formation. Field point PBRO049 at the top of the waterfall in Station Creek east of Mount Morgan.

trondhjemite in composition (Hayward & others, 1999) (Figures 24 and 25). Sandstones are medium to coarse and rarely fine-grained, and graded bedding is locally well developed. They range from poorly to well sorted, and can be cross-bedded, including sparse examples of herring-bone cross-bedding. The sandstones are similar in composition to the conglomerates, but thin section examination reveals that some contain clinopyroxene grains and rare ooids. Siltstones are a minor rock-type overall in the Raspberry Creek Formation, but locally can form relatively thick sequences, and are commonly silicified. The tuffs are crystal and lithic-poor, with sparse to abundant accretionary lapilli.

Basalt lavas are most abundant in the Struck Oil area north-east of Mount Morgan (Hackman, 1995). On the surface they have weathered to a medium to pale grey colour, but in core from mineral exploration drilling they are dark grey to black, some with conspicuous plagioclase phenocrysts. They are interbedded with coarse sandstones sourced almost exclusively from basaltic volcanics. Basalts are absent from the eastern belt of Raspberry Creek Formation through Mount Gavial, and only andesitic compositions are present (Hackman, 1995).

Limestone is a minor part of the sequence, but is more common than in the Mount Warner Volcanics and the overlying Mount Hoopbound Formation, particularly towards the top of the unit. Most of the limestones are fossiliferous. Rare lenses of oolitic limestone are also present.

Geochemistry

Morand (1993a) analysed four basalts from the area covered by the Raspberry Creek Formation, and interpreted these as volcanic arc basalts using standard discriminant plots. Offler & Gamble (2002) presented REE and other trace element data for sample 31676 analysed by Morand, and favoured an island arc setting based on Ti/Zr and Th/Yb ratios and Zr contents. On standard discrimination diagrams (Figure 26), basalts from the Raspberry Creek Formation fall in the volcanic arc or island arc tholeiite fields. Trace and rare earth element contents are typical of basalts from modern island arcs such as Tonga-Kermadec and the Marianas (Figure 27). Together with 8 basalts from the Struck Oil district analysed by Hackman (1995) and sample 31676 from the same area, these rocks have relatively high Ti/Zr (range 116.4 to 296.6 and average 203.6) and Ti/Y values (range 306.3 to 375 and average 321.7) characteristic of fairly primitive intra-oceanic island arcs (Woodhead & others, 1993; see Murray & Blake, 2005, figure 6).

Figure 24. Trondhjemite clast in pebble to cobble conglomerate composed dominantly of andesite clasts, Raspberry Creek Formation. Field point MHRO 034 at the top of a waterfall in Station Creek east of Mount Morgan (see Hayward & others, 1999).

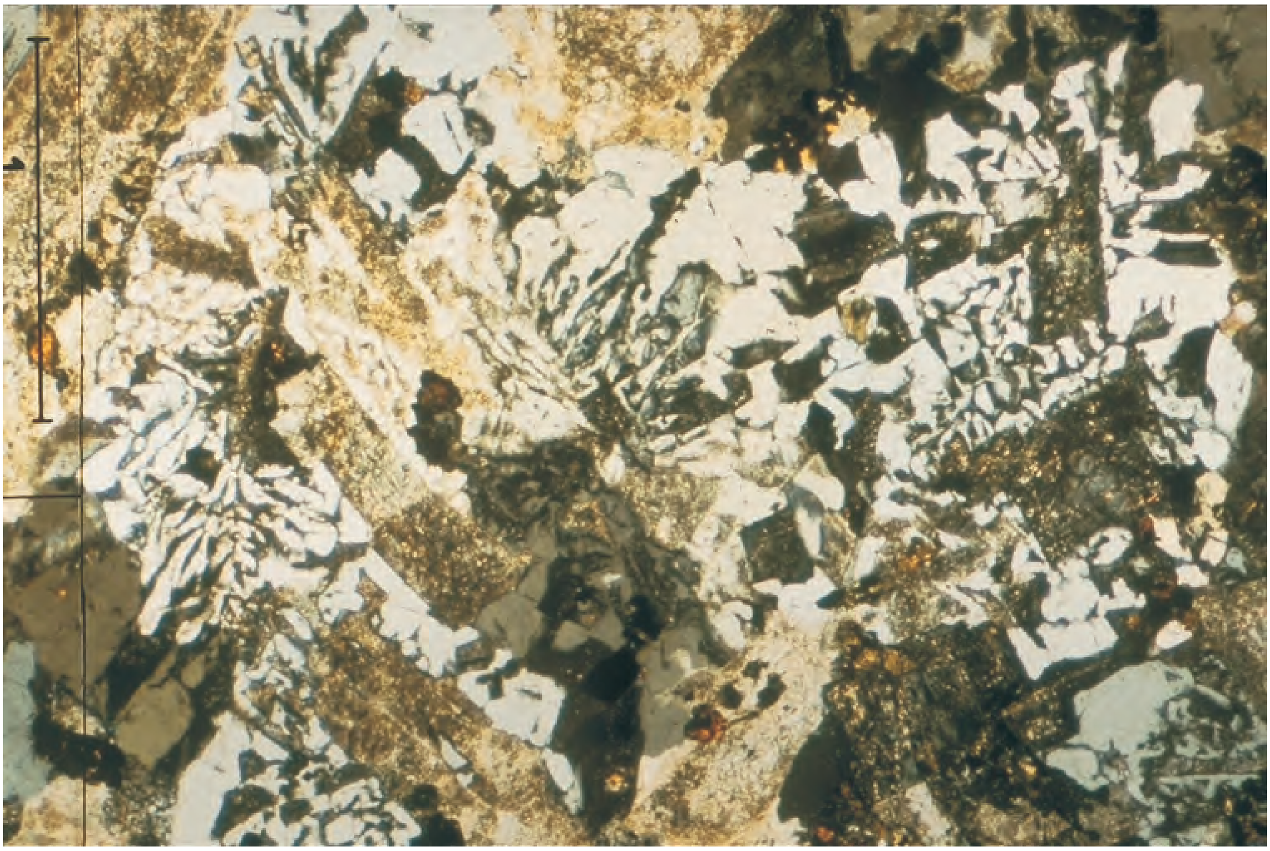


Figure 25. Photomicrograph of trondhjemite clast showing granophyric texture of plagioclase and quartz. Crossed polars. Scale bar is 1mm long. Clast from Raspberry Creek Formation.

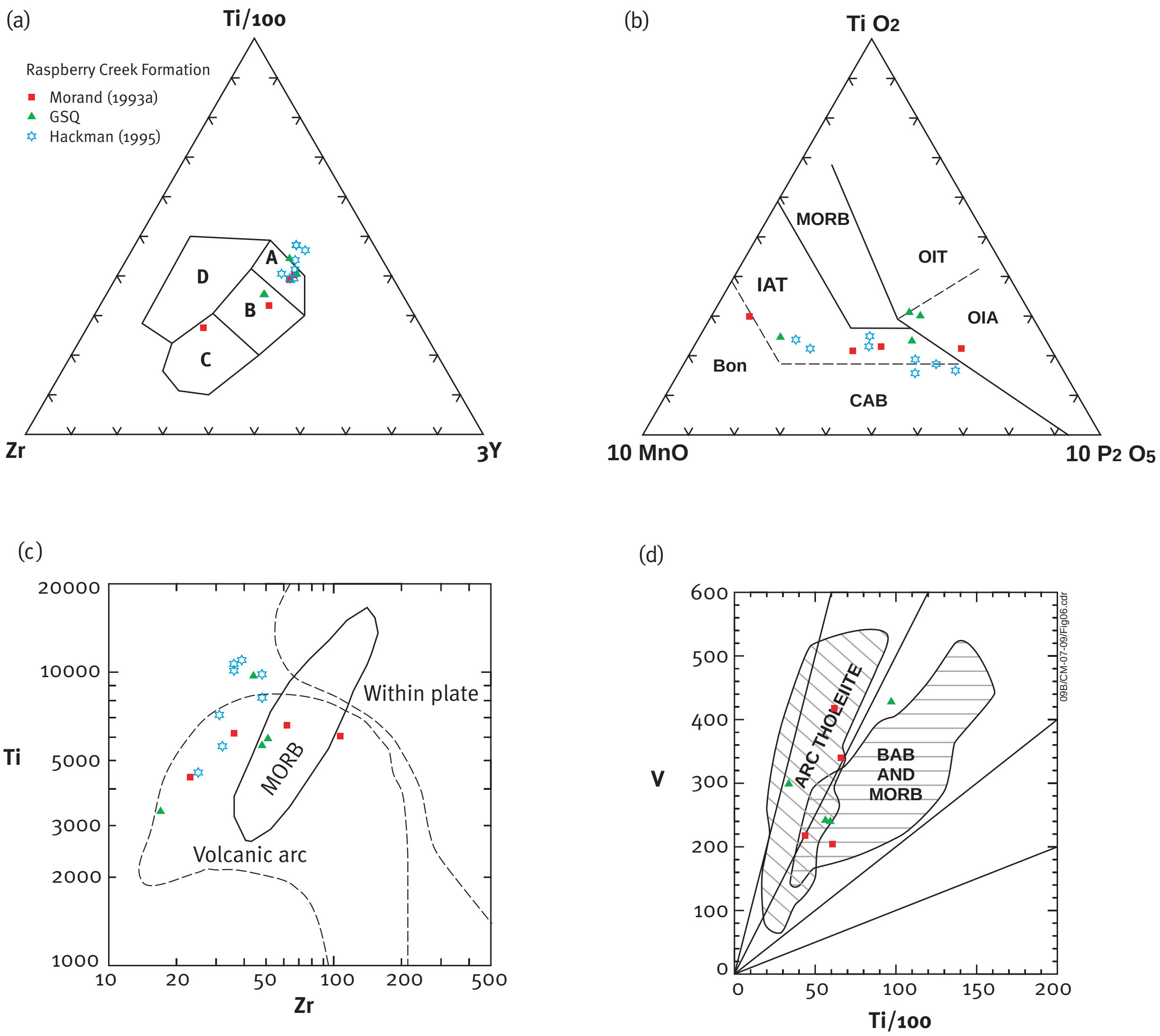


Figure 26. Analyses of basalts from the Raspberry Creek Formation (Morand, 1993a; Hackman, 1995; GSQ) plotted on discriminant diagrams of Pearce & Cann (1973), Mullen (1983), Pearce (1982) and Shervais (1982). Analyses fall in fields defining island arc tholeiites or volcanic arc basalts.

Environment of deposition

Corals, conodonts and brachiopods indicate that the Raspberry Creek Formation was deposited in a marine environment, probably in shallow water based on the presence of minor oolitic limestone and herring-bone cross-bedding. The abundance of coarse volcanic material suggests that it was proximal to a volcanic source, and basalt geochemistry is most consistent with an oceanic island arc setting.

Thickness

The type section of the Raspberry Creek Formation is ~465m thick. It is possible that other parts of the formation are thicker, but the maximum thickness cannot be estimated because the top is an unconformity and the unit is folded.

Structure

Together with the remainder of the Capella Creek Group, the Raspberry Creek Formation has been folded into broad folds with shallowly dipping limbs. The main structure in the Dee Range is a syncline parallel to and about 5km west of the crest of the Gracemere Anticline. A syncline outlined by limestone beds at Marble Mountain may be an extension of this structure. Cleavage is weakly developed, and is generally only apparent in fine-grained rocks. The cleavage becomes more strongly developed to the south, and between Marble Mountain and Ayrdrie homestead is visible in some outcrops of conglomerate. The cleavage dips steeply to the east-north-east and is roughly parallel to the regional trend in the Mount Alma Formation and Erebus beds to the east.

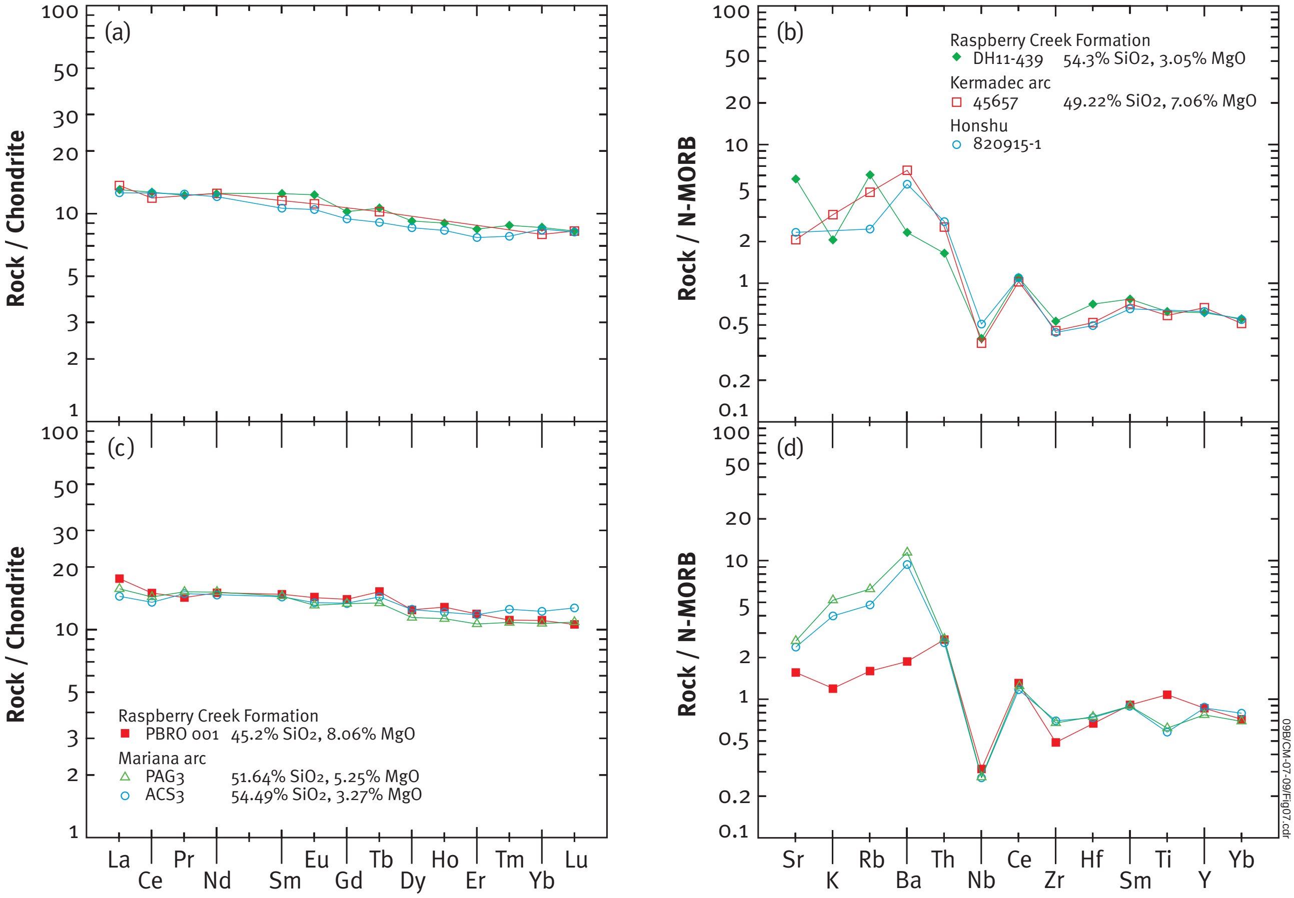


Figure 27. (a) REE and (b) spidergram plots comparing basalt and basaltic andesite from the Raspberry Creek Formation with basalts and basaltic andesite from the Kermadec, Honshu and Mariana arcs (Turner & others, 1997; Shibata & Nakamura, 1997; Elliott & others, 1997). Alteration is responsible for the different MgO contents of the samples from the Raspberry Creek Formation. Chondritic REE values are from Sun & McDonough (1989), and N-MORB from Pearce (1983).

The Raspberry Creek Formation is unconformably overlain by the Mount Hoopbound Formation. The contact has been observed at two localities along Emu Creek 23km south-south-east of Mount Morgan (Kirkegaard & others, 1970, pages 38–39; Leitch & others, 1992), and displays strong angular discordance. The significance of this Middle–Late Devonian unconformity has been the subject of debate. It has been attributed to collision of an exotic Early to Middle Devonian island arc with the Australian continent (Powell, 1984a, page 327), or to local changes in plate motion (Morand, 1993b) or extension (Messenger, 1996; Bryan & others, 2003) at a continental margin rather than docking of an exotic terrane. One of the critical points in the resolution of this problem is whether this tectonic event was associated with folding and cleavage generation. Some of the structural evidence provided by Morand (1993a,b) suggests that this might have been the case (see discussion in Murray & Blake, 2005), although he concluded that only a single fold-thrust event, the Hunter–Bowen Orogeny, was responsible for cleavage development in all units regardless of age. Folding in Devonian time is supported by relations between the Mine Corridor sequence and the Mount Morgan Trondhjemite at Mount Morgan (Taube, 1986; Golding & others, 1994a).

Field observations from the southern part of the Raspberry Creek Formation suggest that the cleavage is at least in part due to pre-Late Devonian deformation. At some localities in the Ayrdrie homestead area, the prominent cleavage in conglomerates of the Raspberry Creek Formation is not present in finer-grained rocks of the overlying Mount Alma Formation.

Biostratigraphy and age

The Raspberry Creek Formation contains a diverse coral fauna containing *Tryplasma*? *abyssum*, *Cystiphylloides immanum*, *Pseudomicroplasma australe*, *Metriophyllum*? sp., *Barrandeophyllum* sp., *Tabulophyllum* sp., *Breviphyllum mirourum*, *Sanidophyllum davidis*, *Acanthophyllum* (*A.*) *clermontense*, *Dohmophyllum clarkei*, *Grypophyllum jenkinsi*, *Xystriphyllum* cf. *magnum*, *Stringophyllum bipartitum*, *S. quasinormale*, *Sunophyllum* sp. cf. *S. proteum*, *Disphyllum stupendum*, *Amaraphyllum amoenum*, *Thamnopora* *boloniensis*, *Alveolites suborbicularis*, *Temnophyllum stanesi*, , *Alveolites caudatus*, *Heliolite porosa*, and *Multithecopora tubus* and indicates a late Middle Devonian (Givetian) age for the fauna (Blake, 2010).

The conodont faunas have been reviewed by Taube & others (2005), who concluded that they ranged from the post-*kockelianus* zone to the middle *varcus* subzone (latest Eifelian to middle Givetian) within the Middle Devonian.

Stratigraphic relationships

The Raspberry Creek Formation appears to conformably overlie the Mount Warner Volcanics. Where the dacitic to rhyolitic upper part of this formation can be mapped at a regional scale, it is separated out as the Ginger Creek Member. The Raspberry Creek Formation is intruded by the Mount Morgan Trondhjemite in the north-west, by the Bouldercombe Igneous Complex in the north, and by the Galloway Plains Igneous Complex in the south. It is overlain unconformably by the Balaclava, Mount Hoopbound, and Mount Alma Formations. The contact between the Raspberry Creek Formation and the Upper Devonian Mount Alma Formation on the eastern limb of the Gracemere Anticline north-east of Mount Morgan has been described as conformable (Golding & others, 1994a, page 96; Morand, 1993b, page 263), but there appears to be a slight angular discordance between the two units.

Ginger Creek Member (Dcrg)

(P.R. Blake & A. Taube)

Introduction

The Ginger Creek Member occurs at the top of the Raspberry Creek Formation. Outcrops are only locally large enough to be shown separately on regional scale maps.

Distribution

Rocks typical of the Ginger Creek Member can be identified in several areas over the whole length of the Raspberry Creek Formation. However, it has only been mapped out at three localities. The first is in a fault block with a triangular shape and sides about 5km long at Ginger Creek 25km south-south-west of Bajool, the second is an elongate belt 15km long and up to 3km wide along the Calliope Range 30km south of Bajool, and the third is an outcrop 8km by 2km surrounding the Ulam marble quarry 18km south of Bajool. All these occurrences are in the south-western corner of the Bajool 1:100 000 Sheet area (Figure 28).

Derivation of name

The name is from Ginger Creek on the western slopes of the Ulam Range, which flows along the eastern side of the triangular fault block of Ginger Creek Member.

Type section

No type section was measured, but the fault block near Ginger Creek is nominated as the type area.

Topographic expression

The Ginger Creek Member is a competent unit that forms rugged topography along the Ulam and Calliope Ranges and around Mount McCamley.

Geophysical expression

The Ginger Creek Member is geophysically indistinguishable from the other units in the Capella Creek Group. It has a very low response in all three channels of the airborne radiometric survey data, and also has a low magnetic signature. The low radiometric response suggests that the felsic rocks are low in K, similar to compositions in the Mount Warner Volcanics.

Lithology

Volcaniclastic sedimentary rocks dominate the Ginger Creek Member, with lesser fossiliferous limestone and rare dacitic to rhyolitic lava.

Figure 28. Distribution of the Ginger Creek Member

The volcaniclastic rocks consist of medium- to very coarse-grained sandstone and granule to pebble conglomerate and breccia. These sedimentary rocks are lithic to feldspatholithic, mainly sourced from dacitic and rhyolitic volcanics, and typically poorly sorted. Some contain rare fossils of brachiopods and corals.

The limestones are small lenses up to several metres thick and tens of metres long. They contain abundant crinoids, corals and sparse brachiopods.

The dacitic and rhyolitic lavas are sparsely porphyritic with phenocrysts of quartz and feldspar. Peperite formed by lava flowing into fossiliferous limestone was observed at one locality (Figure 29).

Environment of deposition

The presence of fossiliferous limestone containing photic zone fossils suggests that the Ginger Creek Member was deposited in a shallow marine environment, and the abundance of primary volcanic rocks and volcaniclastic sediments indicates deposition near a volcanic centre. Data from other units of the Capella Creek Group strongly support an oceanic island arc environment.

Thickness

The Ginger Creek Member is assumed to be at most a few hundred metres thick, but no section has been measured.

Structure

Most of the Ginger Creek Member is gently dipping and displays little evidence of deformation, except in the southern part of the Calliope Range where it possesses a relatively strong cleavage that dips steeply to the east-north-east.

Figure 29. Peperite in the Ginger Creek Member produced by dacitic lava intruded into limestone. The coral in the photograph is *Thamnopora boloniensis*.

Biostratigraphy and age

Corals collected from the Ginger Creek Member include *Stringophyllum quasinormale*, *Thamnopora boloniensis*, *Alveolites suborbicularis*, *Heliolite porosa* indicating a Middle Devonian age (Blake, 2010).

Conodonts recovered from the limestones have further refined the age, indicating an age in the range from the middle *varcus* subzone to the upper *hermanni–cristatus* zone, or middle to late Givetian. These are the youngest limestones collected from the Capella Creek Group.

Stratigraphic relationships

The Ginger Creek Member forms the top of the Raspberry Creek Formation. However, the three largest exposures are either separated from the Raspberry Creek Formation by Quaternary alluvium, or are in faulted contact. The Ginger Creek Member is overlain unconformably by the Upper Devonian to Lower Carboniferous Balaclava Formation, and is intruded by small plugs of Upper Devonian gabbro (**Dg_b**).

Correlation with other units

The Ginger Creek Member is most closely related, in both age and composition, to the Marble Waterhole beds on the Biloela 1:100 000 Sheet area.

Marble Waterhole beds (Dm)

(M.A. Hayward & P.R. Blake)

Introduction

The name Kroombit beds was first introduced by Dear (1968) to describe Late Devonian (Frasnian) andesitic rocks in the Cania area. Later, the definition was expanded to incorporate the fossiliferous limestones of

Figure 30. Distribution of the Marble Waterhole beds

Middle Devonian (Givetian) age in the vicinity of Marble Waterhole south-east of Biloela, as well as volcanics and epiclastic sediments to the west and north-west (Dear & others, 1971).

As they spanned the time interval of the Middle to Late Devonian unconformity in the Mount Morgan area, the Kroombit beds were considered to provide evidence against the regional nature of deformation of this age (eg. Morand, 1993b, page 268). However, an important result of mapping in the Kroombit Creek – Cania area is the recognition of two discrete units within what was previously mapped as Kroombit beds, namely the felsic Middle Devonian Marble Waterhole beds and the mafic Upper Devonian Lochenbar Formation.

Distribution

Rocks assigned to the Marble Waterhole beds form a north-westerly trending belt 20km long and up to 5km wide extending from near Blue Hills homestead (GR238165 7284986 on the Scoria 1:100 000 Sheet area) to the headwaters of Kroombit gully (GR272500 7303900 on the Biloela 1:100 000 Sheet area) (Figure 30).

Derivation of name

The name is from Marble Waterhole where cliffs of fossiliferous limestone bound Kroombit Creek (Figure 31).

Type section

A type section was laid out and logged in two parts. The first 24m was measured vertically through the limestone cliffs on the western bank of Marble Waterhole. The section extends only to where these rocks are disrupted by a microdiorite dyke that may have intruded along a pre-existing fault. To the south of this point the bedding dips consistently to the south-south-east at 25° to 40° and this is where the second part of the section was laid out and a further 210m of the sequence logged.

Figure 31. Limestone cliff beside Marble Waterhole, Marble Waterhole beds

Topographic expression

Mapped areas of Marble Waterhole beds form undulating to hilly country surrounding Kroombit Creek and its tributaries. Areas of massive limestone generally form resistant, rounded hill morphology that contrasts with the more dissected topography of other rocks of this unit.

Geophysical expression

The high resolution aerial geophysical data proved to be of very little assistance in subdividing the former Kroombit beds, or distinguishing between rock types within either of these units. On the total count radiometric image, the Marble Waterhole beds appear as an area of uniformly low-response. Similarly, the magnetic image is virtually featureless apart from a few subtle north-west trending structures that may reflect the numerous microdiorite/andesite dykes in the area.

Lithology

The unit contains, and is locally dominated by, large areas of fossiliferous limestone. However, it also comprises feldspatholithic sandstone, granule to pebble dacitic breccia/conglomerate and dacitic to rhyolitic tuff/ignimbrite, and rare andesite (Figure 32).

The limestones, which appear to occupy the lower part of the unit, contain abundant corals, which combined with conodont age determinations have returned a very precise age in the Givetian (late Middle Devonian) (Figures 33 and 34). The limestones are thick bedded, contain little in the way of rock fragments, but do locally contain abundant volcanic quartz and feldspar grains, and ooliths (Figure 35). Overall, the sedimentary sequence appears to be rhyolitic to dacitic, contrasting with the dominantly andesitic and basaltic composition of the Lochenbar Formation. Rocks of the Marble Waterhole beds are also generally finer-grained and less hematised than those of the Lochenbar Formation. This is reflected in the overall green-grey tone for rocks of the Marble Waterhole beds, in contrast with the purple to grey colour of most rocks of the Lochenbar Formation.

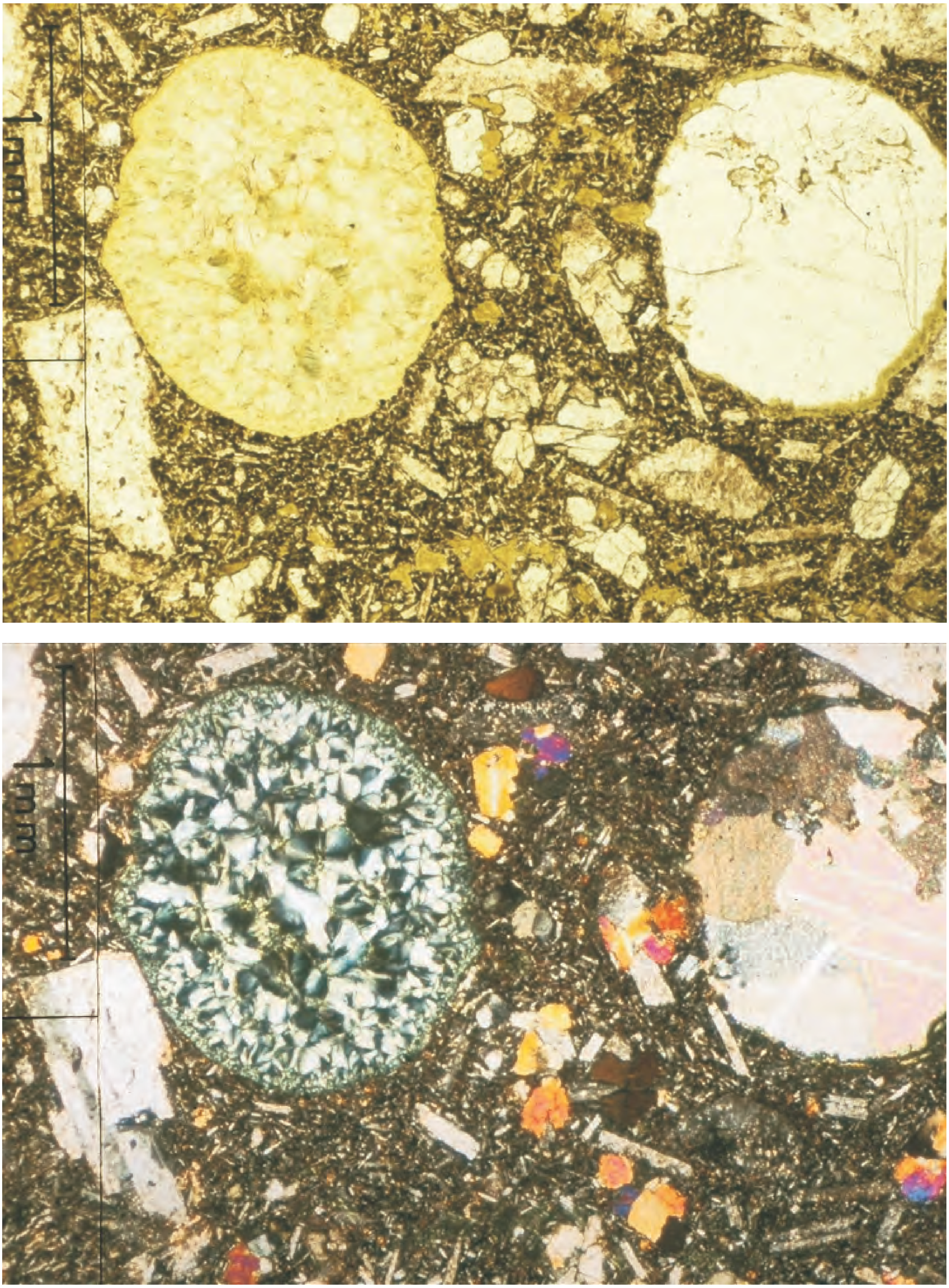


Figure 32. Photomicrographs of andesite with calcite- and chlorite-filled amygdales and microphenocrysts of plagioclase and augite, Marble Waterhole beds. Plane polarised light and crossed polars. Scale bar is 1mm long.

Figure 33. Corals in limestone beside Marble Waterhole, including *Calceola* sp., *Thamnopora boloniensis* (Gosselet), *Alveolites caudatus* Hill, and *Multithecopora tubus* Blake, Marble Waterhole beds.

Environment of deposition

Abundant thick-bedded coralline limestone and sparse brachiopod faunas indicate deposition in a shallow marine environment.

Thickness

A total section of 234m was measured near Marble Waterhole on Kroombit Creek, but neither the base nor the top of the unit is exposed. Elsewhere rocks of this unit are either poorly exposed or have been subjected to block faulting, rendering the estimation of thickness very subjective.

Structure

The structural complexity of the Marble Waterhole beds appears to be due to locally intense block faulting rather than to folding. The dip direction and azimuth of beds vary considerably within short distances. Numerous andesite, rhyolite and microdiorite dykes also intrude the unit, adding to the complexity.

Figure 34. The coral *Alveolites suborbicularis* Lamarck in limestone beside Marble Waterhole, Marble Waterhole beds.

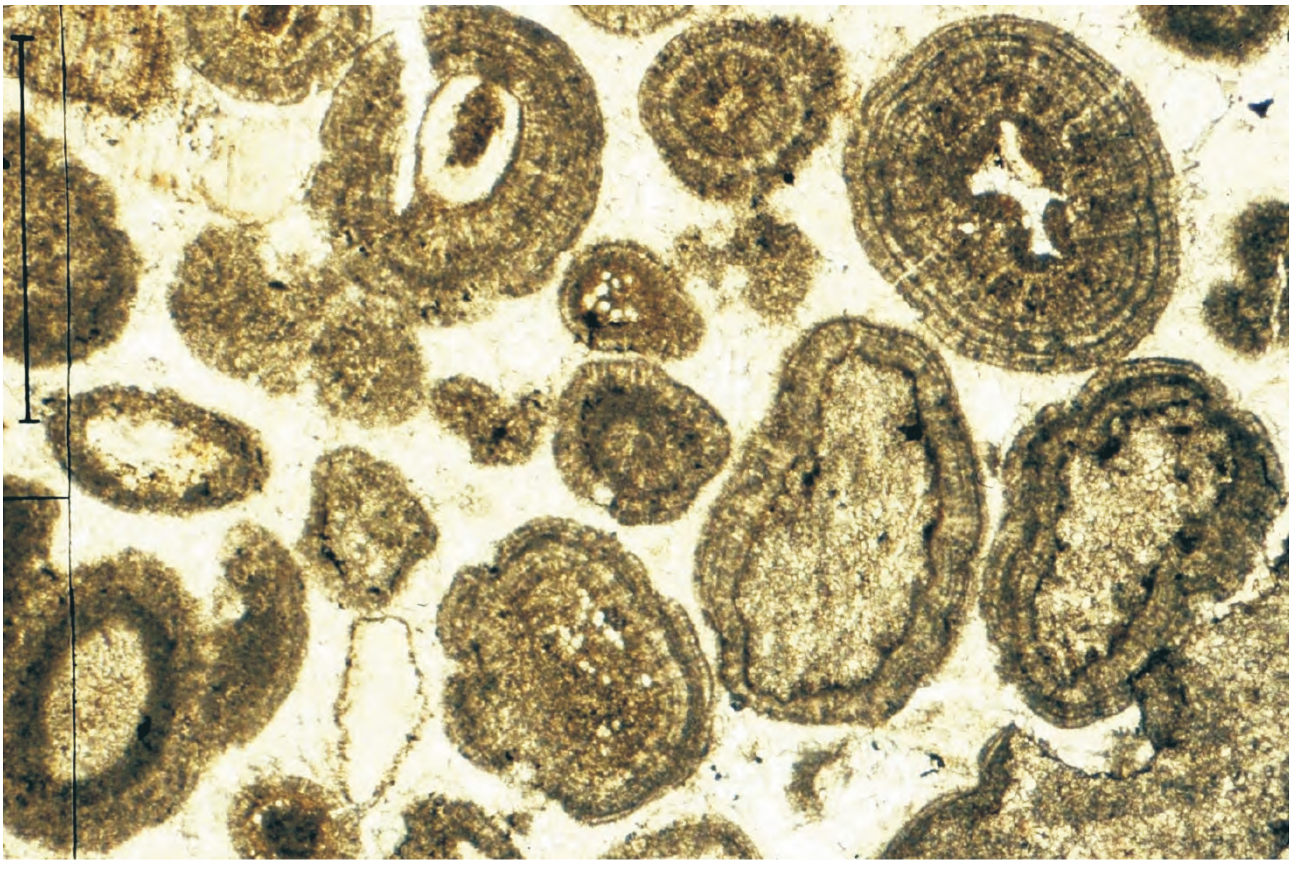


Figure 35. Photomicrograph of oolitic limestone showing that the ooids are not closely packed and that some have irregular shapes due to algal activity, Marble Waterhole beds. Plane polarised light. Scale bar is 1mm long.

Biostratigraphy and age

The rich coral fauna from the Marble Waterhole beds contains *Cystiphylloides immanum*., *Microplasma caespitosum*, *Pseudomicroplasma australe*, *Breviphyllum* sp., *Endophyllum columna columna*, *Sanidophyllum davidis*, *Tabulophyllum* sp., *Acanthophyllum clermontense*, *Dohmophyllum clarkei*, *Grypophyllum jenkinsi*, *Stringophyllum bipartitum*, *S. quasinormale*, *Sunophyllum* sp. cf. *S. proteum*, *Disphyllum stupendum*, *Amaraphyllum amoenum*, *Thamnopora boloniensis*, *Alveolites caudatus*, *Alveolites suborbicularis*, *Heliolites porosa*, *tubus*, and *Multithecopora* sp. This fauna indicates a Middle Devonian (Givetian) age (Blake, 2005, 2010).

The brachiopod *Atrypa desquamata* var *kimberleyensis* has also been found.

Samples of the limestones were dissolved for conodonts and results indicate that the limestones are confined to the *varcus* zone, confirming the Givetian age determined from corals.

Stratigraphic relationships

On the western side the Marble Waterhole beds are faulted against the Lochenbar Formation by the high angle Kroombit Fault. To the east they are unconformably overlain by felsic rocks of the uppermost Carboniferous and Lower Permian Youlambie Conglomerate and Triassic Winterbourne Volcanics, except at one locality where they are faulted against a small outcrop of the Rockhampton Group.

Correlation with other units

Lithological and geochemical similarities have been noted between the Marble Waterhole beds and dacitic rocks in the Ginger Creek Member of the Capella Creek Group, but a direct correlation has not yet been established. Conodont age determinations from limestone beds in these two units indicate that both units occupy a similar time range.

The Marble Waterhole beds are considered to be part of the Mount Morgan Stratigraphic Assemblage (Yarrol Project Team, 1997).

Economic significance

The Allen Creek or Mount Kroombit Mine is hosted by the Marble Waterhole beds. It is represented by several shafts, two adits and numerous pits and was worked at the turn of the century for copper. Abundant massive specular hematite and limestone lie at the surface around the mine area, as well as in Allen Creek. This area is currently held under a small mining lease.

In the 1960s and 1970s, Carpentaria Exploration Pty Ltd took out mining leases over small ?copper/base metal prospects (Karita and Tea Tree prospects) east-south-east of Kroombit Dam. However, most of these leases were surrendered shortly after grant and it appears that very little work was done in the area.

LATE DEVONIAN – EARLIEST CARBONIFEROUS

Channer Creek beds (Da)

(P.R. Blake)

Introduction

The Channer Creek beds were originally mapped as marine beds with Frasnian faunas (Maxwell, 1960a, figure 22). The name was introduced by Dear & others (1971), from whom the following description is taken.

Distribution

The Channer Creek beds form an outcrop 8.5km long and up to 2.5km wide along the western side of the Yarrol Fault in the southern part of the Monto 1:100 000 Sheet area, and extend south into the Mundubbera 1:250 000 Sheet area (Whitaker & others, 1974) (Figure 36).

Topographic expression

The Channer Creek beds produce a low hilly topography.

Figure 36. Distribution of the Channer Creek beds

Geophysical expression

The Channer Creek beds give a low radiometric response in all three channels and a strong magnetic signature. The magnetic response is similar to that of the Calliope beds that occupy an equivalent structural position.

Lithology

The dominant rock types are dark green, fine- to coarse-grained lithic sandstone and granule to pebble conglomerate. Light green argillite is rare. Whitaker & others (1974) record andesitic lavas and siliceous fine-grained sandstone as rock types on the Mundubbera 1:250 000 Sheet area.

The lithic sandstone is composed almost entirely of well rounded andesitic lava grains (55%) and subangular plagioclase feldspar grains (40%), cemented by chlorite. The conglomerates have granules and pebbles of andesite in a finer bonding fraction, or matrix, of lava grains, feldspar and chlorite (Dear & others, 1971).

Environment of deposition

The presence of corals indicates deposition in a marine environment.

Thickness

The Channer Creek beds are disrupted by faulting, and occur in a fault block with no top or bottom preserved, so their thickness is unknown.

Structure

The Channer Creek beds are disrupted by numerous faults.

Biostratigraphy and age

Dear & others (1971) assigned a Frasnian age based on corals collected from granule conglomerate.

Blake (2010) identified *Temnophyllum turbinatum*, *T. kroombitense*, *Piceaphyllum menyouense*, *Charactophyllum* sp., *Alveolites caudatus*, and *Heliolites porosa*. *Temnophyllum krombitense* & *Piceaphyllum menyouense* has only been found in the Frasnian units of the Yarrol province. Temnophyllum turbinatum has been considered Frasnian in age, but Brownlaw (2000) indicated that this coral is confined to the Givetian in the Canning Basin. Also, a single specimen of the Early to Middle Devonian Heliolites porosa was identified in a collection in the Queensland Museum.

It is possible that the Channer Creek beds range from Middle to Late Devonian in age, or more likely it is possible that they are wholly Late Devonian in age and the older corals were eroded from pre-Late Devonian units, an occurrence which has been identified in other parts of the Yarrol Province.

Stratigraphic relationships

The Channer Creek beds are faulted against the Rockhampton Group on their western side, and against the Doonside and Wandilla Formations to the east. Their stratigraphic relationships are therefore unknown.

Correlation with other units

On the basis of rock type and coral faunas, the Channer Creek beds can be correlated with the Lochenbar Formation and Mount Hoopbound Formation within the Yarrol Province. However, their structural position along the western side of the Yarrol Fault is anomalous. Their location and strong magnetic signature are similar to the uppermost Silurian – Lower Devonian Calliope beds.

Mount Hoopbound Formation (Dh)

(R.E. Randall, M.A. Hayward, & P.R. Blake)

Introduction

Upper Devonian rocks west, south-west and south of Mount Morgan were first mapped by Maxwell (1952, 1953). Kirkegaard & others (1970) revised the stratigraphy and described three separate units, the Dee Volcanics, Boulder Creek Grit and undifferentiated Upper Devonian. The Dee Volcanics were a dominantly andesitic volcanic unit of Frasnian age, consisting mainly of coarse volcaniclastic sediments and primary flows. The overlying Boulder Creek Grit, of Famennian age, was entirely sedimentary, and of finer grain size, although it still contained a large proportion of coarse to pebbly sandstone. The undifferentiated Upper Devonian unit was correlated lithologically with the Dee Volcanics, but was at least partly of younger age as indicated by the local occurrence of Famennian fossils (Kirkegaard & others, 1970, page 41).

Randall (1996) and the current mapping program recognised several horizons within the Dee Volcanic sequence that were lithologically identical to rocks assigned to the Boulder Creek Grit. The Boulder Creek Grit can therefore be considered as a somewhat thicker and younger example of a sedimentary facies that was repeatedly developed during deposition of the Dee Volcanics, and the two units have been combined as the Mount Hoopbound Formation, together with most of the undifferentiated Devonian rocks of Kirkegaard & others (1970). The Mount Hoopbound Formation contains both sediments and volcanics of Frasnian and Famennian age (Yarrol Project Team, 1997).

Distribution

The Mount Hoopbound Formation extends for 40km south from Mount Morgan to the Don River, and forms two belts with a combined width of up to 25km separated by the Balaclava Formation and Eulogie Park Gabbro (Figure 37).

Derivation of name

The unit takes its name from Mount Hoopbound 15km south-south-east of Mount Morgan, where thick sandstone and conglomerate beds form prominent westerly-dipping benches on its eastern slopes.

Figure 37. Distribution of the Mount Hoopbound Formation

Type section

The type section for the Mount Hoopbound Formation is the section measured by Randall (1996) along Raspberry Creek from the basal contact with the Mount Morgan Trondhjemite 1.5km south of Mount Hoopbound.

Topographic expression

The eastern belt of the Mount Hoopbound Formation, which extends from Mount Morgan in the north almost to Glengowan homestead in the south, forms moderately rugged country parallel to and west of the Dee Range. The southernmost portion of this belt forms part of the Gelobera Range. The western belt extends southwards from Sugarloaf Mountain to the Don River, and forms undulating country punctuated by low to moderate hills.

On aerial photographs, soils derived from the Mount Hoopbound Formation appear light brown to medium red-brown. Bedding traces are often clearly visible, particularly on east-facing slopes. The best example is on the eastern face of Mount Hoopbound, where west-dipping beds form prominent semi-circular benches.

Vegetation cover is densest in the east near the headwaters of McBride, Capella and Raspberry Creeks. Further west, moderate to heavy clearing has been carried out to promote better grazing pasture. Several prominent faults are clearly visible on air photos.

Geophysical expression

The Mount Hoopbound Formation produces dark red tones on red-blue-green images generated from airborne radiometric survey data, reflecting a consistent but relatively low potassium content. The potassium response is higher in an area along the Dee River south from Mount Morgan. The unit gives a low to moderate magnetic response. Linear north-west trending magnetic features were found during fieldwork to be related to mafic dykes.

Figure 38. Purple lapilli tuff in Raspberry Creek (field point RRRO 047, GR390 683). Purple colouration of the matrix is typical of thermal oxidation of hot emplaced pyroclastics (Cas & Wright, 1988). Note the reaction rims around aphyric andesite fragments. Mount Hoopbound Formation. Pen magnet is 140mm long.

Lithology

Rock types in the Mount Hoopbound Formation include dominant pebble to cobble andesitic breccia/conglomerate, accompanied by lithic sandstone, siltstone, lapilli tuff, ash tuff and andesite and basaltic andesite lavas (Figure 38). Brecciation is common, and the matrix of some samples is replaced by wairakite (Figures 39 and 40). Wilson (1980) reported thin rhyolitic flows from the middle and upper sections. Where the unit overlies the Mount Morgan Trondhjemite, a basal polymict conglomerate is developed with boulders up to 1m across, including trondhjemite clasts (Figures 41 and 42). The abundance of these clasts diminishes upwards through the sequence, and they are restricted to within 100m of the contact.

The Mount Hoopbound area was mapped in detail by Randall (1996), who recognised ten gravelly, fine-grained and primary volcanic lithofacies (Table 2) that he grouped into two facies associations: Dud_1 forms the base of the Mount Hoopbound Formation, and is overlain by Dud_2 (Figure 43).

The logs of three sections measured by Randall (1996) are presented in Figure 44.

Geochemistry

The geochemistry of basaltic andesites and andesites of the Mount Hoopbound Formation was discussed by Murray & Blake (2005), who placed them in their Lochenbar suite. The rocks are slightly to moderately enriched in LREE, and the HFSE are depleted relative to N-MORB to varying degrees, particularly Nb, Ta and Zr. For the Lochenbar suite (which also includes basalts from the Lochenbar Formation and the Three Moon Conglomerate), $(Th/Nb)_N$ ranges from 4.76 to 16.13 and averages 10.0. Ce/Y values for the Lochenbar suite suggest a crustal thickness of ~20km (Mantle & Collins, 2008). Compositions compare most closely with mafic volcanics from modern day evolved or mature island arcs such as the Lesser Antilles (Murray & Blake, 2005). For samples from the Hoopbound Formation, the contents of both REE and HFSE increase consistently with increasing SiO_2 and decreasing MgO. The ratios $(La/Sm)_N$ and $(La/Yb)_N$ remain constant when plotted against individual REE values, reflecting the parallelism of REE patterns as their overall level increases, and indicating that these trends are the result of fractionation (Murray & Blake, 2005, figure 12e). The geochemical data support a model in which basalts generated by partial melting processes from mantle source rocks underwent subsequent fractionation to produce andesites. The abundance of andesite produced by fractionation of basalt suggests the existence of a substantial upper crustal magma chamber.

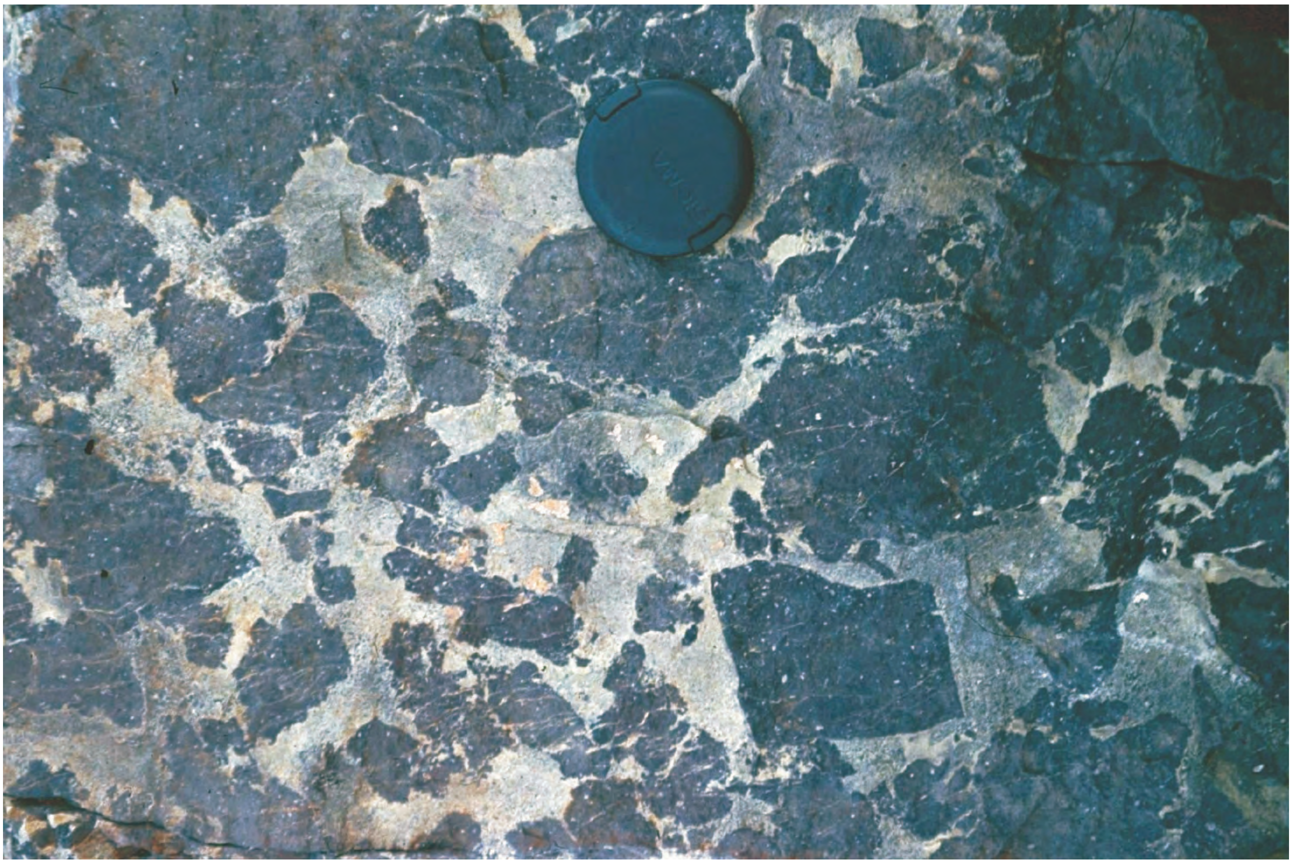


Figure 39. Hematised andesite fragments in brecciated andesite, Mount Hoopbound Formation

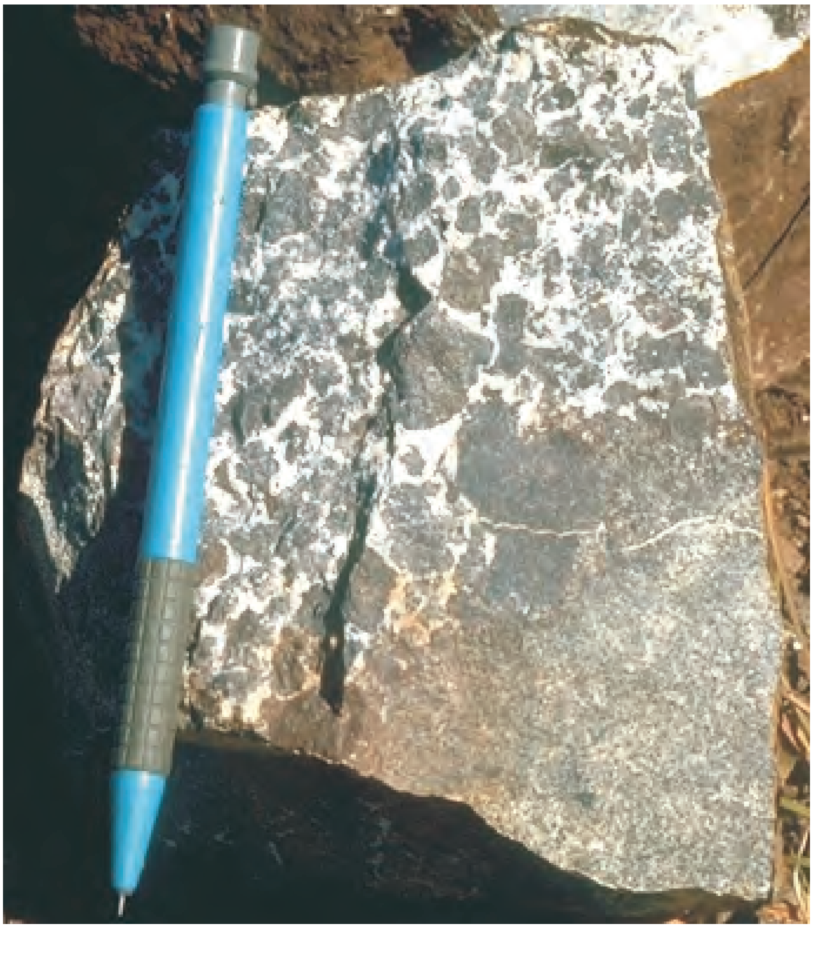


Figure 40. Brecciated and altered andesite with matrix of wairakite, Mount Hoopbound Formation

Figure 41. Polymictic conglomerate with clasts ranging from aphyric andesitic volcanics to equigranular, medium-grained trondhjemite, Mount Hoopbound Formation. Field point RRRO 088 at GR408 712 west of New Chum Creek.

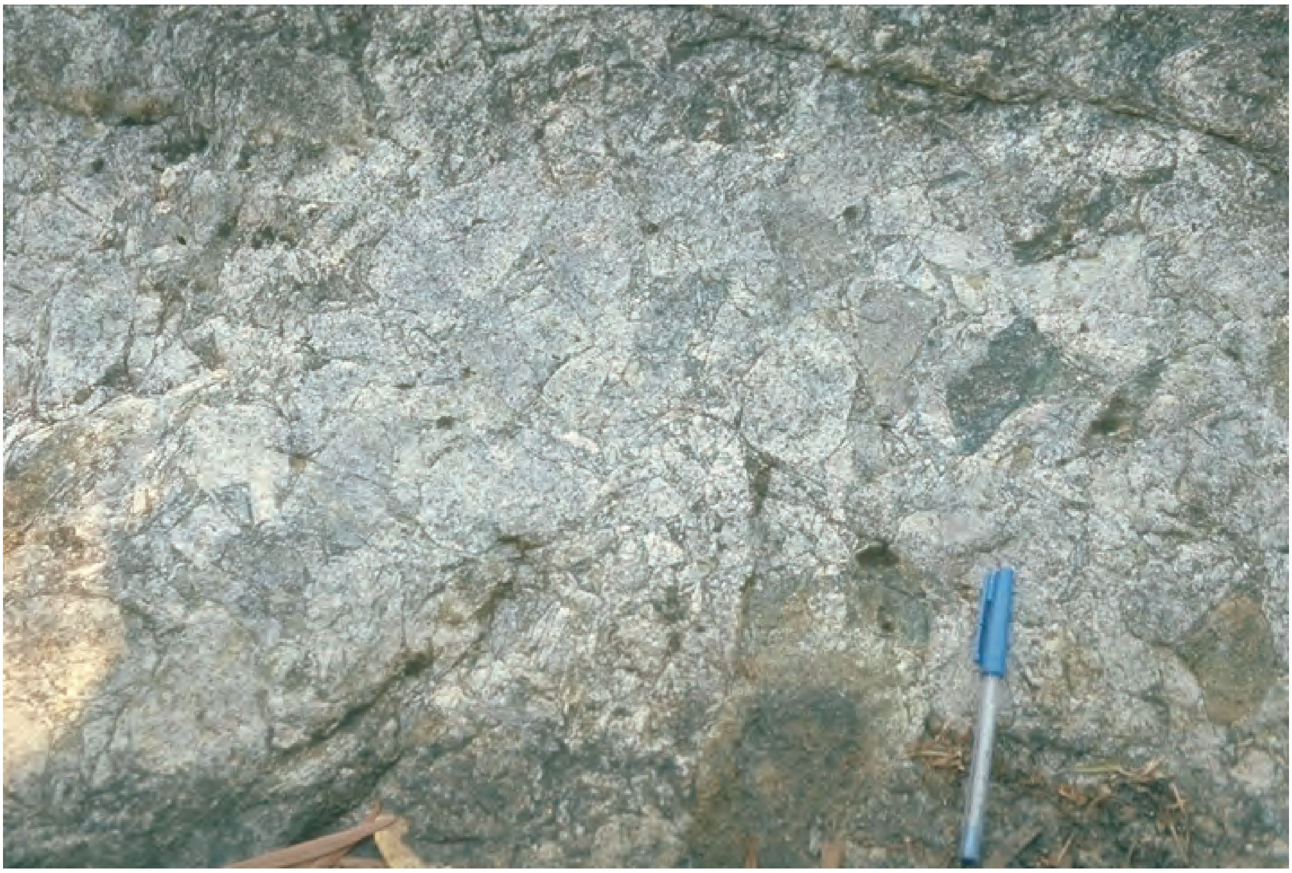


Figure 42. Closely packed polymictic conglomerate in Bull Creek. Clasts consist mainly of trondhjemite with subordinate andesitic volcanics, Mount Hoopbound Formation.

Table 2. Lithofacies in the Mount Hoopbound Formation

Facies Name	Description	Fossils	Interpretation
COARSE-GRAINED FACIES			
Polymictic conglomerate	Polymictic pebble to boulder, poorly sorted, generally clast supported conglomerate. Clasts generally composed of fine to medium silicic plutonics (trondhjemite, tonalite, granodiorite). Minor felsic and intermediate volcanics (rhyolite to andesite) and chert. Matrix of fine to coarse quartzose sand and ?andesitic lithic sand. Clasts are subangular to rounded.		These are probably talus cones which developed below gullies (White, 1981). They occur at the base of the high relief Mount Morgan Trondhjemite.
Monomict andesitic conglomerate	Light brown to purple-green, generally thick bedded, cobble to boulder, poorly sorted andesitic conglomerate. Clasts range from aphyric to dominantly porphyritic, and are mainly subrounded. Fabric ranges from clast to matrix supported. Clast imbrication poor. Channel fill deposits common.	Shallow marine fossils and limestone blocks present in some beds.	High energy terrestrial and shallow marine stream flow (Smith, 1986) and coastal fan deposits (Wasson, 1977; Wells, 1984).
Breccia/conglomerate	Purple-green, thin to thick bedded andesitic clastic deposits. Beds lower in the sequence contain rare to abundant trondhjemite clasts in fine to coarse ash (silt to coarse sand) matrix. Trondhjemite clasts often copper stained. Clast size diminishes in the southern portion of the area. Rare crudely fining up beds. Beds above those containing trondhjemite are generally monomict and are composed of andesite clasts in andesitic and crystal-rich ashy matrix.		High matrix proportion indicates probable debris flow (McPhie & others, 1993). These probably occurred as a result of slides and explosive eruptions (Cas & Wright, 1988).
Pebbly andesitic breccia/conglomerate	Large volumes of well bedded, generally pebbly but often coarser, angular to subrounded poorly sorted andesitic detritus. Commonly clast supported and exhibits no stratification.	Fossiliferous in some areas. Coral fragments and impressions most common; some brachiopods and gastropods present.	The deposits are poorly sorted and ungraded, and probably resulted from hypoconcentrated flood flow (Smith, 1986), beach and near shore processes (R.C.M. Smith, 1991).
FINE-GRAINED FACIES			
Lithic arenite	Cream-purple-green, fine to coarse, poorly to well sorted andesitic lithic arenite. Thin to very thickly bedded. Pebble to cobble, and rare boulder, clasts of aphyric to abundantly porphyritic andesite common. Interbedded pebble layers common, often infilling troughs. Trough cross-bedding common.	Carbonate-rich layers containing rare to abundant limestone clasts, with coral and shell fragments, common in the southern portion of the area.	Moderate to high energy fluvial to shallow marine deposits. Possible fan delta (presence of nearby land) tidal channel and shallow marine storm dominated shelf (Walker & Cant, 1984).
Siltstone	Purple to cream, well sorted, often fossiliferous siltstone. Zeolite infill of coral clasts at one locality.	Coral impressions common in places.	Fine-grain size and presence of marine fossils indicate low energy shallow marine deposits (Goldring, 1991).
PRIMARY VOLCANIC FACIES			
Lapilli tuff	Purple-green, often pumice-rich, poorly sorted lapilli tuff. Lapilli exhibit reaction rims and microscopic plastic deformation textures. Some outcrops show flattened pumice clasts, especially on weathered surfaces.		Primary pyroclastic pumice, block and ash flow deposits (Cas & Wright, 1988; McPhie & others, 1993).
Ash tuff	Purple, stratified, lithic and pumiceous ash tuff. Flattened and elongated pumice fragments common. Well layered. Low to moderate welding indicated in thin section. No cuspate glass shards observed.		Primary pyroclastic unwelded to poorly welded air fall ash tuff (Cas & Wright, 1988; McPhie & others, 1993).
Andesite lava	Dark grey aphyric to moderately porphyritic pyroxene andesite. No quench fragmentation observed in lava flows along Raspberry Creek. Autobreccia apparent in some outcrops. Generally poorly exposed.		Terrestrial and subaqueous lava flows (Cas & Wright, 1988; McPhie & others, 1993).
Hyaloclastite	Dark grey, flow banded, blocky fragments with dark grey ?andesitic matrix. Located on top of andesite flow which overlies fossiliferous arenite.		Hyaloclastite formed by lava flow into shallow water (McPhie & others, 1993).

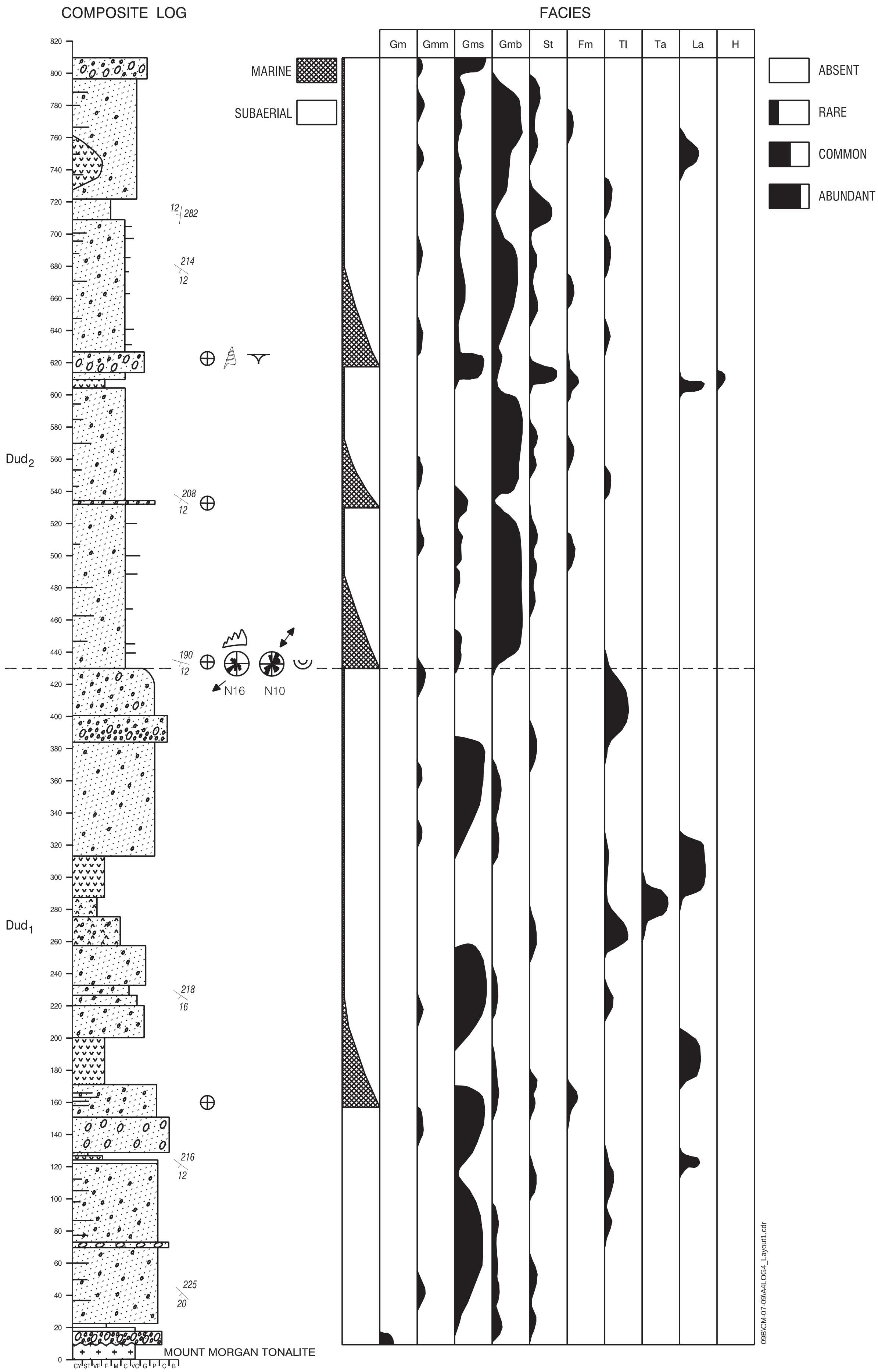


Figure 43. Composite stratigraphic section and facies abundances, Mount Hoopbound area. See Figure 44 for legend.

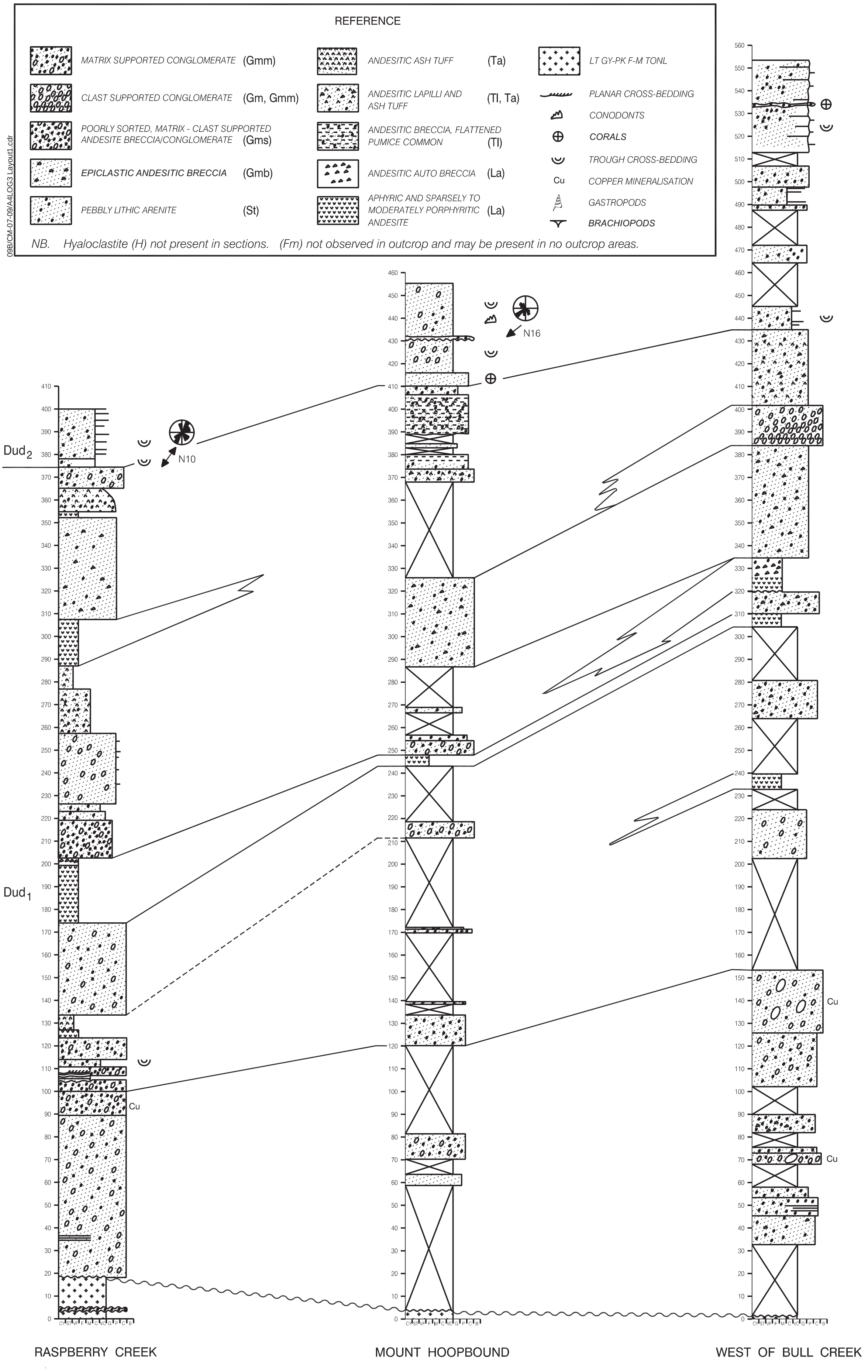


Figure 44. Measured stratigraphic sections and correlations

The volcanic setting appears to have been remarkably similar to that of the island of Montserrat in the Lesser Antilles. Montserrat is composed dominantly of porphyritic andesites with volumetrically minor basalts, and has been the site of major pyroclastic eruptions of andesitic composition (Druitt & Kokelaar, 2002). The geochemistry of basaltic lavas from Montserrat (Zellmer & others, 2003) matches analyses from Upper Devonian units south of Mount Morgan, including samples from the Mount Hoopbound Formation, very closely (Murray & Blake, 2005).

Environment of deposition

The presence of brachiopod, coral and conodont faunas throughout the formation supports earlier interpretations that these rocks were formed mainly in a marine environment. Herring-bone cross bedding and ripple marks suggest shallow to moderate water depth (Figures 45 and 46). Thick poorly sorted conglomerates and breccias within the unit are interpreted as debris flows.

Andesite hyaloclastite and altered rims of ragged breccia fragments indicate sub-aqueous quenching of primary volcanic material (Figure 47).

An andesitic ignimbrite on the flanks of Mount Hoopbound (Figure 48) suggests that part of the sequence was deposited subaerially.

The dominance of poorly-sorted volcanic detritus indicates a setting near an emergent volcanic terrain, and the presence of andesite and basaltic andesite flows demonstrates that volcanism was active during deposition. Randall (1996) proposed that the lower facies association Dud_1 was mainly terrestrial, and the succeeding facies association Dud_2 was largely shallow marine. A palaeogeographic setting in an island arc, with a volcanic island (or islands) like Montserrat at least partly beneath the Mount Hoopbound Formation, is consistent with the distribution and succession of rock types.

Thickness

A thickness of ~1000m is likely for the Mount Hoopbound Formation, as estimated by Morand (1993a). This estimate is supported by the section of 400m measured through the basal part of the unit along Raspberry Creek by Randall (1996). Extending this section to the boundary between the Mount Hoopbound and Balaclava Formations, and using outcrop width and average dips, the total thickness of the Mount Hoopbound Formation is calculated to be between 1000–1100m.

Structure

The Mount Hoopbound Formation dips to the south-west at shallow angles. Folding is expressed only as broad warps, and local steeper dips are probably the result of faulting.

Several large (10–15km long) faults are clearly visible on aerial photography and satellite and geophysical imagery. These faults appear to be normal faults with mainly vertical movement. Kirkegaard & others (1970) proposed that some were reactivated basement faults that originally pre-dated deposition of the Mount Hoopbound Formation, controlling the basement topography and causing thickness variations. Probably the best example is the fault trending west-south-west from Mount Morgan that forms the northern boundary of the Mount Hoopbound Formation.

Biostratigraphy and age

Previous reports on coral, brachiopod and conodont faunas from the Dee Volcanics and Boulder Creek Grit indicated that the sequence now included in the Mount Hoopbound Formation covers much of the Late Devonian (Frasnian to Famennian) (Druce, 1969b; Kirkegaard & others, 1970).

Corals identified in the Dee Volcanic by Blake (2010) include *Smithiphyllum petercolsi*, *Temnophyllum kroombitense*, *Thamnopora boloniensis*, *Alveolites suborbicularis*, and *A. caudatus* and are Late Devonian (Frasnian).

A poorly preserved cast of the gastropod ?*Michella* sp. and a medium sized turbiform gastropod, exhibiting three or more turns with strong spiral ornament of pustules, have also been found.

Randall (1996) reported results for conodont faunas extracted from limestone clasts from near the top of Mount Hoopbound and the bed of Fletcher Creek. The faunas ranged from the upper *rhenana* zone to the upper *postera* zone, which is middle Frasnian to middle Famennian.

Figure 45. Bi-directional cross-bedding in lithic sandstone in Raspberry Creek, Mount Hoopbound Formation

Figure 46. Pavement outcrop of ripples in lithic sandstone, Mount Hoopbound Formation

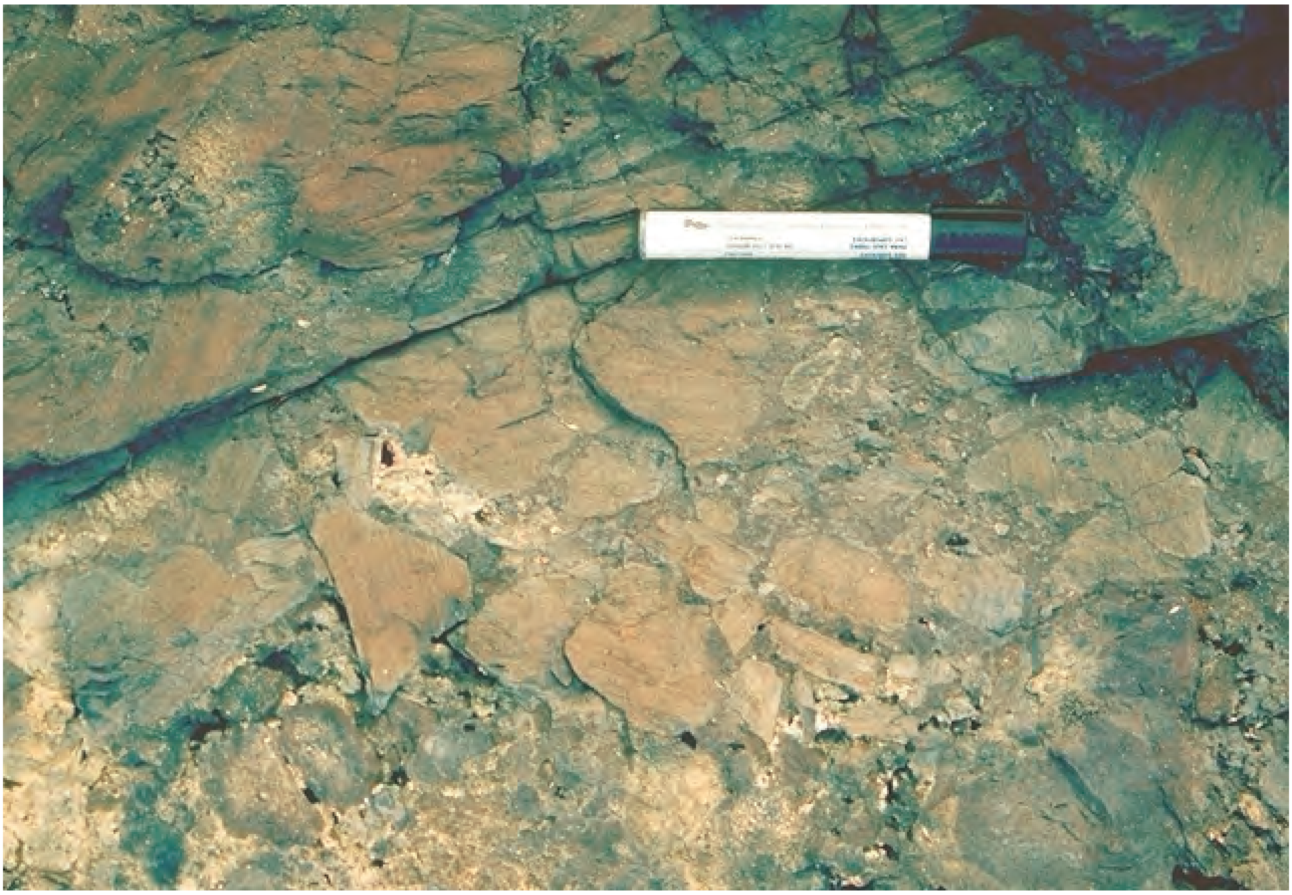


Figure 47. Andesitic hyaloclastite at GR379 691 in Fletcher Creek, Mount Hoopbound Formation. The clasts exhibit curvilinear shapes which are typical of quench fragmentation (McPhie & others, 1993).

Figure 48. Lithic-rich andesitic lapilli tuff on Mount Hoopbound (GR403 703), Mount Hoopbound Formation. The flattened pumice clasts probably indicate that it is an ignimbrite deposited by pyroclastic flow processes.

Figure 49. Large boulder of trondhjemite in andesitic breccia-conglomerate in the bed of Bull Creek at GR386 732. Trondhjemite clasts in this vicinity contain copper carbonate mineralisation, Mount Hoopbound Formation.

Stratigraphic relationships

Rocks of the Late Devonian Mount Hoopbound Formation unconformably overlie both the Capella Creek Group and Mount Morgan Trondhjemite. An angular unconformity with the Capella Creek Group was recorded by Kirkegaard & others (1970) and Leitch & others (1992). Morand (1993a) reported that there has historically been some difficulty in distinguishing these two units in areas where the angular discordance is small. However, Randall (1996) noted that the highly silicified nature of the Raspberry Creek Formation aided discrimination. Although both units are dominated by basaltic to dacitic rocks, the greater abundance of modal quartz in the Raspberry Creek Formation, and the common presence of limestone, are useful distinguishing features.

Contacts with the Mount Morgan Trondhjemite are marked by the presence of coarse basal conglomerate containing cobbles and boulders of trondhjemite (Brown & Wilson, 1965; Kirkegaard & others, 1970; Messenger & Golding, 1996; Randall, 1996). At Bull Creek, south-east of Mount Morgan, andesitic breccia to conglomerate near the base of the formation contains clasts of trondhjemite up to several metres across (Figures 49 and 50). Further south at New Chum Creek, the base of a conglomerate bed is dominated by cobbles and boulders of Mount Morgan Trondhjemite (Figure 41). The abundance of these clasts dramatically decreases up section, and they disappear within 100m of the contact. In a tributary of Capella Creek at the southern end of the Mount Morgan Trondhjemite, the base of the Mount Hoopbound Formation is a very coarse conglomerate with clasts of andesitic volcanics and trondhjemite up to 1m across (Figure 51).

The Mount Hoopbound Formation is both laterally equivalent to and overlain by the more felsic Balaclava Formation. It is intruded by unnamed Permian to Triassic granitic to dioritic intrusions, the Triassic Eulogie Park Gabbro, and by numerous microdiorite, andesite and diorite dykes.

Correlation with other units

The Mount Hoopbound Formation is at least partly equivalent in age to the Balaclava Formation, Lochenbar Formation, Mount Alma Formation, and Three Moon Conglomerate. In rock type and environment of deposition, the Mount Hoopbound Formation is most like the Lochenbar Formation.

Figure 50. Detail of the margin of the large trondhjemite boulder in Bull Creek, Mount Hoopbound Formation

Figure 51. Conglomerate with boulders of trondhjemite and andesitic volcanics near the southern end of the Mount Morgan Trondhjemite, Mount Hoopbound Formation.

Economic significance

Mafic volcanics of the Mount Hoopbound Formation in the Bull Creek – Mount Hoopbound area contain native copper mineralisation concentrated at the top of individual flow units (Wilson, 1980).

Lochenbar Formation (Dl)

(M.A. Hayward)

Introduction

All of what was previously mapped as Dawes Range Formation and the majority of what was previously mapped as Kroombit beds on the Monto 1:250 000 Sheet area is now assigned to the Upper Devonian Lochenbar Formation. Dear & others (1971) considered that the Lochenbar Formation formed the western limb of a faulted anticline, with Middle Devonian limestones now included in the Marble Waterhole beds on the eastern limb. The contact between Middle and Upper Devonian rocks is now interpreted as a fault. The recognition that the Kroombit beds of Dear & others (1971) actually consist of two very distinct units of different ages negates arguments put forward by Morand (1993b, page 268) for continuous deposition across the intra-Devonian unconformity reported by Kirkegaard & others (1970).

Distribution

The Lochenbar Formation occupies a south-east trending belt 50km long and up to 15km wide from near Callide Dam (Biloela — GR259000 7305900) in the north to the headwaters of Boolgal Creek (Scoria — GR297600 7273300) in the south (Figure 52). The best outcrops are in areas of moderate relief south of Lochenbar homestead (GR277800 7293800) and in the vicinity of Cania Dam. It is likely that the Lochenbar Formation persists under Tertiary basalt cover to the south of the main outcrop area.

Figure 52. Distribution of the Lochenbar Formation

Derivation of name

The unit is named after Lochenbar homestead at GR277800 7293800.

Type section

A 610m conformable sequence of rocks was measured and logged just south of Lochenbar homestead from GR276768 7293247 to GR276585 7292248. Rocks of the Lochenbar Formation continue to the south where they are covered by Tertiary basalt.

Topographic expression

The Lochenbar Formation produces fairly flat to slightly undulating country along the valley and slopes of Kroombit Creek, and moderately hilly topography south of Lochenbar homestead and east of Callide Dam.

The unit is quite unremarkable on aerial photographs. However, numerous north-west trending dykes are prominent. Residual cover, and basalt, both of Tertiary age, which overlie the Lochenbar Formation are readily distinguished.

Geophysical expression

The Lochenbar Formation has a uniform low response on airborne radiometric images, and a uniform low to moderate magnetic response. The uniformity of response makes internal discrimination impossible. Also, because rocks of the adjacent Marble Waterhole beds produce very similar responses to the Lochenbar Formation, the airborne geophysics is of little use in distinguishing between these two units. The Youlambie Conglomerate overlies both the Lochenbar Formation and the Marble Waterhole beds, and produces a very high response in all three channels of the radiometric data. This characteristic allows it to be readily distinguished from most other units with which it is in contact.

Numerous north-west trending linear features which are discernible on the total count radiometric image appear to correspond with Triassic rhyolite dykes, probably associated with the Winterbourne Volcanics. A parallel set of more prominent lineaments is clearly visible on the first vertical derivative magnetic image, and these correspond with microdiorite dykes of probable Permian to Triassic age.

The courses of Kroombit and Dry Creeks are clearly defined on total count radiometric images due to the strongly contrasting high response of the alluvial bedload (derived from felsic rocks in the Kroombit Tops area), and the very low response of the Lochenbar Formation and Marble Waterhole beds forming the banks and surrounds.

Lithology

Overall, the Lochenbar Formation is dominated by granule to cobble sized epiclastic (volcaniclastic) rocks of basaltic to andesitic, and lesser dacitic composition. However, some polymictic examples are also observed. Bedding is generally medium to thick or massive, although some finer-grained rocks display thin to medium bedding. The most abundant coarse clastic rocks are andesitic volcanic conglomerates and breccias (Figure 53). Not surprisingly, there is a correlation between grain size, sorting and bedding thickness, with the finer-grained rocks being generally thinner bedded and better sorted.

Basaltic andesites and basalts are typically dark grey to dark purplish grey, sparsely to moderately porphyritic, and sparsely to moderately amygdaloidal. Plagioclase feldspars commonly exhibit some epidote alteration.

A large proportion of the volcanic and volcanic-derived rocks exhibit a tendency to weather to a purple colour.

Two previously unrecorded large fossiliferous limestone bodies have been mapped within the Lochenbar Formation west of Kroombit Dam (at GR271100 7297900 and GR262600 7299500).

Geochemistry

The geochemistry of basalts and basaltic andesites from the Lochenbar Formation, as well as the Three Moon Conglomerate and Mount Hoopbound Formation, has been studied by Bryan & others (2001) and Murray & Blake (2005). Analyses from the Lochenbar Formation belong to the Lochenbar suite of Murray & Blake (2005), defined by means of REE and multi-element plots. The Lochenbar suite is slightly to moderately

Figure 53. Massive, matrix supported cobble conglomerate with well rounded clasts of basalt to andesite, Lochenbar Formation. Field point MHRO 877.

enriched in LREE, with $(La/Yb)_N$ ratios from 1.09 to 2.12, and slightly to moderately depleted in HFSE relative to N-MORB, particularly in Ta, Nb and Zr. Th contents are 1.7–5.8 times that of N-MORB, and $(Th/Nb)_N$ ratios average 10.0 and range from 4.76–16.13. Ce/Y ratios for the Lochenbar suite suggest a crustal thickness of ~20km (Mantle & Collins, 2008). The geochemical data for the Lochenbar suite support a model in which basalts of the Three Moon Conglomerate, Lochenbar Formation and Mount Hoopbound Formation were generated by similar partial melting processes from the same source rocks, but subsequent fractionation to produce andesites was restricted to the Mount Hoopbound Formation. Discrimination diagrams using immobile elements indicate that the basalts are volcanic arc tholeiites, or are transitional between tholeiites and calc-alkaline basalts.

Murray & Blake (2005) used the basalt geochemistry to determine tectonic setting in two ways: firstly by compilation of bivariant ratio plots; and secondly by direct comparison of individual analyses in REE and multi-element plots. Both methods demonstrate that the Lochenbar suite has compositions typical of evolved oceanic arcs such as the Lesser Antilles, Marianas, Vanuatu, and the Aleutians. However, they also show similarities to samples from the Eastern Volcanic Front of Kamchatka.

The Three Moon Conglomerate has more of a back-arc character than the Lochenbar and Mount Hoopbound Formations. Not only are some basalts comparable to those of modern back-arc spreading centres like the East Scotia Ridge, but the Lochenbar suite from the Three Moon Conglomerate also includes basalts and basaltic andesites that trend towards transitional arc-back-arc compositions. The fact that compositions become more arc-like to the west provides a strong argument against the simplistic Andean model for the Late Devonian envisaged by Day & others (1978), Henderson (1980) and Murray & others (1987), as well as the model of Bryan & others (2001, 2003) for a back-arc setting for the Upper Devonian rocks behind an eastern arc.

Environment of deposition

The presence of limestones combined with brachiopods found in some of the sediments indicates marine deposition, probably in a shallow marine environment.

Thickness

Dear (1968) logged a composite section of the four component members of the Dawes Range Formation, with a total thickness of 900m. He estimated the minimum thickness of the older Kroombit beds as 2000m.

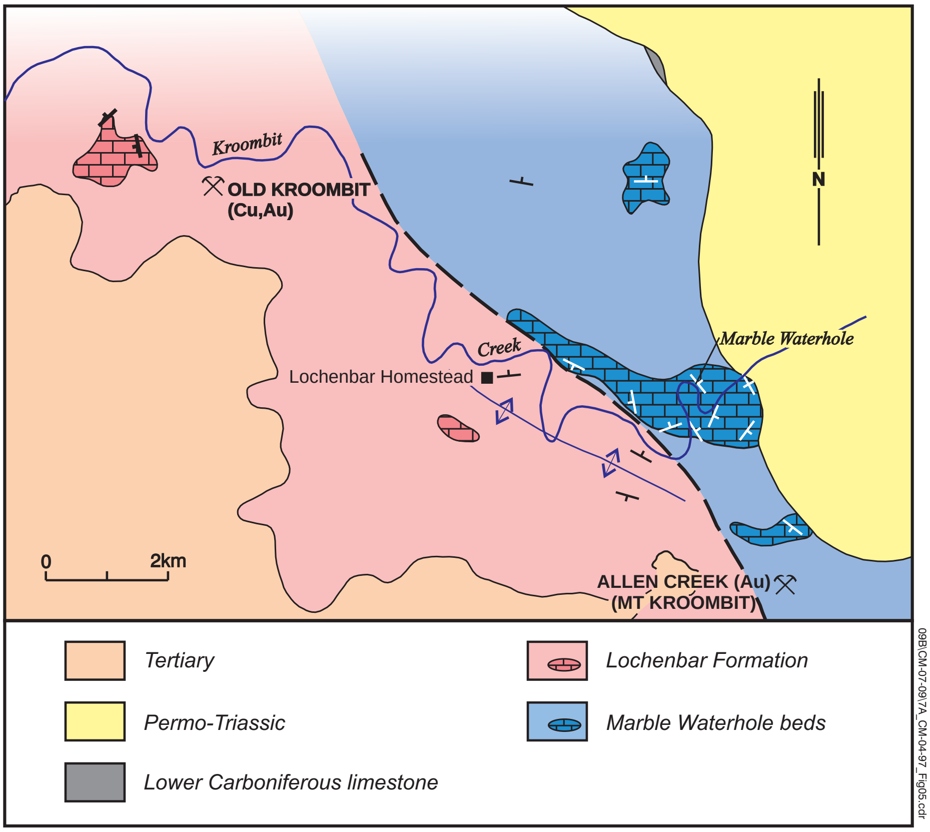


Figure 54. Structure of the Lochenbar Formation in the Kroombit Creek area

Recent mapping and logging of a continuous measured section in the Lochenbar homestead area indicates a minimum thickness of 1500m. The apparent inconsistency between this recent figure and the estimates by Dear (1968) may be accounted for by: (a) the sum of the four non-contiguous component member sections of the Dawes Range Formation gave an inflated thickness for that unit; and (b) the estimate of the thickness of the Kroombit beds by Dear (1968) included the Middle Devonian (Givetian) portion which has now been assigned to the Marble Waterhole beds.

Structure

In the Lochenbar homestead area, rocks of the Lochenbar Formation appear to be structurally quite simple, and have only been disrupted by gentle folding about an anticlinal axis parallelling Kroombit Creek (Figure 54). To the south-west of Lochenbar homestead, beds dip consistently 30° to 40° to the south-west. In the north-western area of exposure, south of Callide Dam, the strike tends to be more east–west and beds have been disrupted by another anticline with an east–west axis. In the Cania region, Dear (1968) invoked a great number of faults to separate blocks of Lochenbar Formation (mapped by him as Dawes Range Formation) containing discrete faunal assemblages. Subsequent mapping has supported the increased structural complexity of this area.

Biostratigraphy and age

The coral faunas in the limestones of the Lochenbar Formation are distinct from those in the Marble Waterhole beds, and were studied by Blake (2005, 2010). The Lochenbar Formation contains *Disphyllum caespitosum*, *Tabulophyllum* sp., *Smithiphyllum petercolsi*, *S. finseni*, *Temnophyllum kroombitense*, *Piceaphyllum menyouense*, *Charactophyllum* sp., *Thamnopora boloniensis*, *Alveolites suborbicularis*, *A. caudatus*, *A murrayi* and *Multithecopora* sp. and are Late Devonian (Frasnian).

Conodont samples from all limestones within this unit have returned faunas from the Late Devonian (Frasnian to ?Famennian). Dear (1968) proposed a Famennian age for the Dawes Range Formation in the Cania region, now mapped as Lochenbar Formation, based on detailed brachiopod determinations.

Stratigraphic relationships

Conodont determinations combined with subsidence curve plots of the Marble Waterhole beds and the Lochenbar Formation indicate the strong probability of at least a 2 million year hiatus between these units.

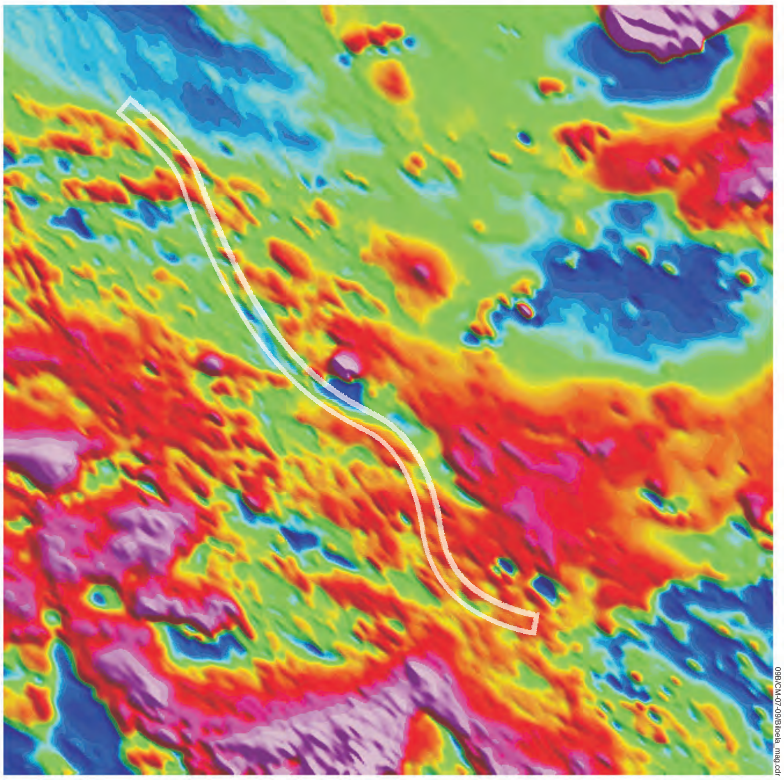


Figure 55. Kroombit Fault (rainbow colour image of total magnetic intensity)

The magnitude and timing of this hiatus directly coincides with the well documented gap between the Capella Creek Group and the Mount Hoopbound Formation in the Mount Morgan area, at the time of intrusion of the Mount Morgan Trondhjemite.

The nature of the boundary between the Marble Waterhole beds and the Lochenbar Formation is uncertain. No convincing evidence for an unconformity was found. The presence of a large, high ridge which parallels Kroombit Creek on the south-western margin of the Middle Devonian limestone, combined with a strong magnetic linear on magnetic images, suggests a fault in this position (Figure 55). This fault juxtaposes the Upper Devonian Lochenbar Formation against the Middle Devonian Marble Waterhole beds and has been named the Kroombit Fault.

To the north the Lochenbar Formation is unconformably overlain by the Youlambie Conglomerate, to the west it is unconformably overlain by Tertiary sediments and volcanics, and to the south it is intruded by the Wingfield Granite and the Kariboe Layered Gabbro.

Correlation with other units

The Lochenbar Formation is at least partly equivalent in age to the Mount Hoopbound Formation, Balaclava Formation, Mount Alma Formation, and Three Moon Conglomerate. It is most similar to the Mount Hoopbound Formation in its environment of deposition and volcanic composition.

Economic significance

The Old Kroombit mine is hosted by the Upper Devonian Lochenbar Formation. There are only a few pits in this area, and it would seem that production from this mine was limited.

Balaclava Formation (DCb)

(P.R. Blake & M.A. Hayward)

Introduction

As in other areas of the New England Orogen, the transition from Devonian to Carboniferous is marked by a change from mafic to felsic compositions in both volcanic rocks and in sediments derived from them. This

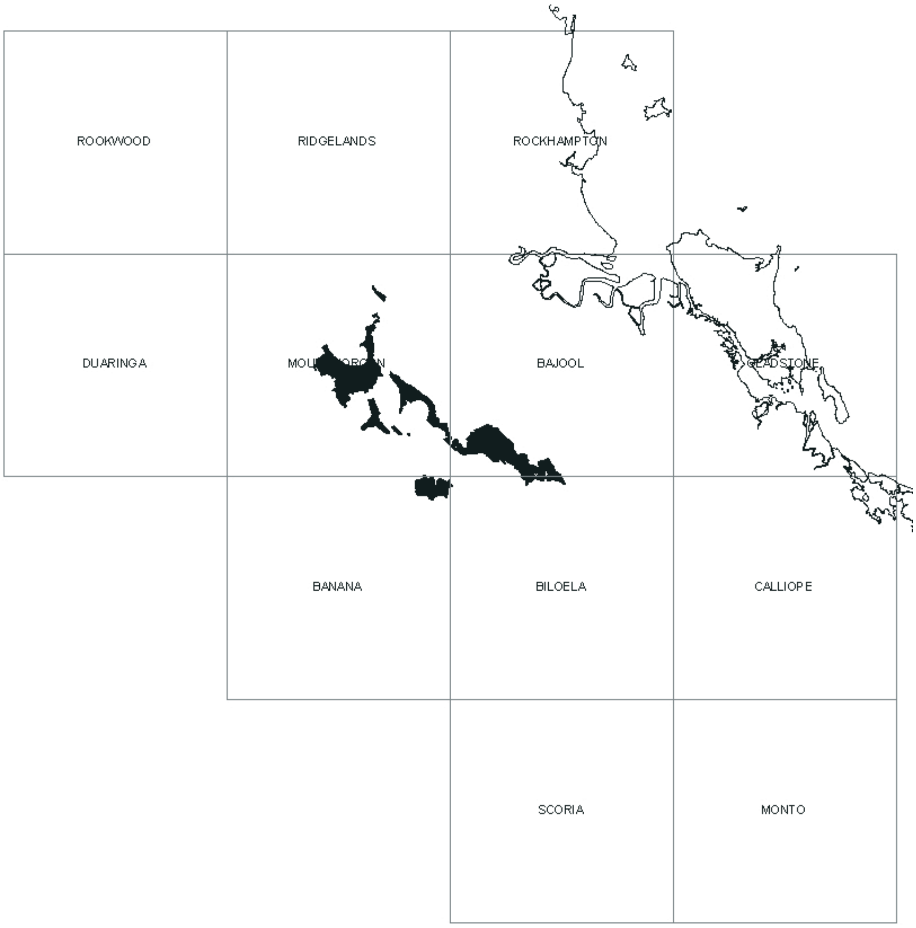


Figure 56. Distribution of the Balaclava Formation

change is reflected in a significant increase in radiometric response. The Balaclava Formation was established to include the oldest strata displaying an increased radiometric response that signifies this change. It combines the Lower Carboniferous Pond Formation and some rocks mapped as undivided Upper Devonian by Kirkegaard & others (1970) and Dear & others (1971).

Distribution

The Balaclava Formation extends as a discontinuous south-east trending belt along the western side of the Gracemere Anticline from near the headwaters of Quarry Creek (10km north-west of Mount Morgan) on the Mount Morgan 1:100 000 Sheet area to Gunpowder Creek in the northern part of the Biloela 1:100 000 Sheet area. Isolated outcrops also occur to the west of this belt, notably to the west and south of the Kyle Mohr Igneous Complex, and near the Don River 40km north of Biloela (Figure 56).

Derivation of name

The name is derived from the Parish of Balaclava in the Bajool 1:100 000 Sheet area, where typical rocks of the unit are exposed.

Type section

A type section has been measured near Grasstree Creek (Bajool 1:100 000 Sheet area GR254331 7352785). The section does not include the base or top of the formation, but it does include most of the common rock types including ignimbrites and pebbly sandstones. It also contains minor rock types such as fine-grained sandstone, fossiliferous siltstone, and rare flow-banded rhyolite. It is possible that the rhyolites are sills, but an origin as flows is favoured.

Topographic expression

The Balaclava Formation forms hilly country. In places it produces ridges with fairly prominent benches formed by the more competent beds within the sequence, and bedding trends can locally be seen on aerial photographs.

Geophysical expression

Rocks of the Balaclava Formation return a moderately high radiometric response in all three channels, consistent with their relatively felsic compositions. This contrasts with the very low response of the dominantly basaltic to andesitic volcaniclastic units such as the Mount Hoopbound Formation, Lochenbar Formation and Three Moon Conglomerate that are at least partial time equivalents of the Balaclava Formation. This characteristic radiometric response has enabled the inclusion within the Balaclava Formation of parts of areas mapped by Kirkegaard & others (1970) and Dear & others (1971) as Capella Creek Group and undivided Upper Devonian rocks. Field checking confirmed the presence of quartz-bearing, felsic rocks typical of the Balaclava Formation in these areas.

The magnetic response is quite variable, and the unit has a patchy appearance on images derived from airborne survey data.

Lithology

The sequence is dominated by sandstone and granule to pebble conglomerate, ranging from poorly to well sorted. The sandstones are feldspatholithic to lithofeldspathic. The poorly sorted sandstone beds are usually thick, poorly bedded and often have a tuffaceous appearance, but no glass shards have been identified in thin section. Therefore, they are interpreted as pyroclastic material deposited in water, the glass shards having been winnowed away. The well-sorted sandstones are well-bedded, often exhibiting cross-bedding, with rare outcrops displaying herringbone cross-bedding. Conglomerates are usually clast-supported and contain clasts of dacitic to rhyolitic lavas and ignimbrite, and rare granitic and sedimentary rocks. Less commonly, the conglomerates are dominated by volcanic clasts of andesitic to basaltic composition. These appear similar to rocks of the Mount Hoopbound Formation, but differ in that they contain at least minor amounts of free quartz (usually lacking in the Mount Hoopbound Formation), and hematised lithic fragments that are characteristic of the Mount Hoopbound Formation are absent or rare.

Lesser rock types include ignimbrite, and rare siltstone, oolitic limestone, and rhyolite lava. The ignimbrites are moderately crystal-rich with crystals of feldspar and minor quartz. They also are lithic-poor to moderately lithic-rich and are fiamme-poor. The lithic clasts are usually of dacitic to rhyolitic volcanics. Thin sections reveal that most of the ignimbrites are poorly to moderately welded, but some are strongly welded. The siltstones usually crop out poorly, and are typically fossiliferous containing abundant brachiopods, gastropods, lycopods and crinoids and sparse nautiloids, corals and bivalves (see Appendix 3).

Oolitic limestone occurs towards the top of the unit in the Gunpowder Creek area in the southern part of the Bajool 1:100 000 Sheet area. It is identical in both lithology and age, based on conodonts, to the oolitic limestones at the base of the Rockhampton Group. However, the limestones at Gunpowder Creek are a very minor part of the sequence, the bulk of which is typical of the Balaclava Formation.

The rhyolitic lavas are aphyric and strongly flow banded. The flow banding becomes contorted towards the top of the flows.

Geochemistry

Three siliceous volcanics (rhyolite and ignimbrite) were analysed from the Balaclava Formation. The samples have a significant subduction component, all being comparatively low in high field strength elements (HFSE), particularly Nb. The $(Th\text{-}Nb)_N$ gap is moderate, ranging from 17.5–21.9 (Figure 57). On the granitoid discrimination diagrams of Pearce & others (1984), the analyses plot in the volcanic arc granite field close to the boundary with ocean ridge granites (Figure 58). Pearce (1996) has pointed out that Rb, Nb and Y concentrations in volcanic arc felsic rocks are linked strongly to source, which is interpreted to be depleted mantle and a crustal component from the subduction process. Because the average composition of continental crust lies within the volcanic arc field, crustal assimilation does not have much effect. The fact that the Rb content is considerably lower than that from other subduction-related felsic volcanics in the Yarrol Project area (Figure 58) may be due to alteration, or it may suggest that they were erupted in a more primitive environment, possibly in the last phase of island arc development. Whether the tectonic environment was a mature island arc or thin continental crust, the felsic volcanics within the Balaclava Formation are interpreted to have been related to a subduction zone.

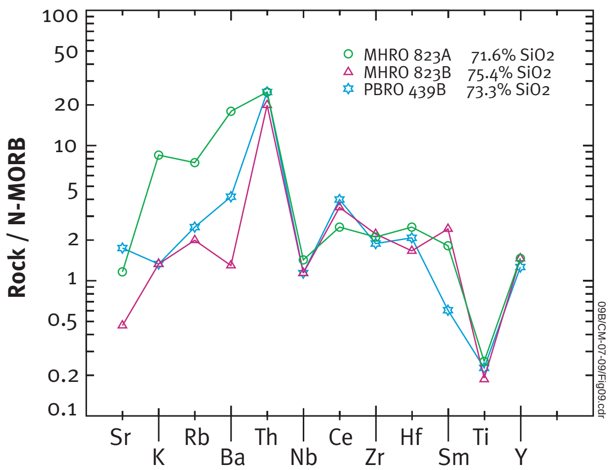


Figure 57. Spidergram plot of three rhyolites. N-MORB values are from Pearce (1983).

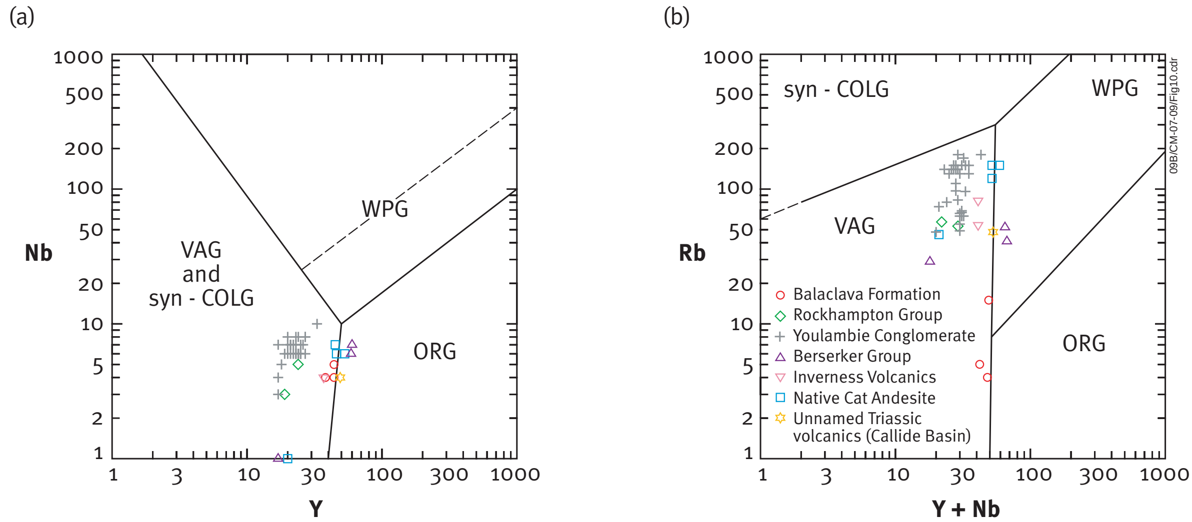


Figure 58. (a) Nb - Y and (b) Rb - Y+Nb discriminant plots of Pearce & others (1984), showing felsic volcanics from the Balaclava Formation, Rockhampton Group, Youlambie Conglomerate, Berserker Group, Inverness Volcanics, Native Cat Andesite and unnamed Triassic volcanics of the Callide Basin. VAG = volcanic arc granites, ORG = ocean ridge granites, WPG = within plate granites, and syn-COLG = syn-collisional granites.

Environment of deposition

The common occurrence of marine fossils within the sequence points to a dominantly marine environment. The local presence of herringbone cross-bedding indicates a shallow marine environment influenced by tides. However, the occurrence of flow banded rhyolites and welded ignimbrites suggests that there were periods of terrestrial deposition.

Given the evidence for shallow marine deposition, combined with the relative abundance of primary volcanics, the Balaclava Formation is interpreted to have been deposited in a shallow basin in close proximity to an arc, or within the arc itself.

Thickness

In the type area at Grasstree Creek, a thickness of ~2000m is estimated from the outcrop width and average dip measurements. Kirkegaard & others (1970) suggested that the Pond Formation was about 600m thick, but this figure does not include any Upper Devonian rocks.

Structure

The Balaclava Formation is not very deformed, with open folding and generally shallow dips to the south-west along the western limb of the Gracemere Anticline.

Biostratigraphy and age

The time range of the Balaclava Formation has mainly been determined from brachiopod faunas by Dr John Roberts from the University of New South Wales (Appendix 3), and is referred to the Early Carboniferous faunal zonation established by Roberts & others (1993).

At the base of the Balaclava Formation, brachiopod faunas contain abundant *Atrypa kimberleyensis*, indicating a Late Devonian (Frasnian) age. Other brachiopods found here include *Desquamatia* (*Synatrypa*) *kimberleyensis* (Coleman), *Nervostrophia bunapica* Veevers, *Cyrtospirifer* sp., *Tenticospirifer* sp., rhynchonelloids and spiriferids.

A fauna collected from about three quarters of the way up the sequence contains *Schizophoria* sp., *Rhipidomella australis* (McCoy), *Schuchertella* sp., *Leptagonia analoga* (Phillips), *Rugosochonetes magnus* (Maxwell), *Spinocarinifera kennedyensis* (Maxwell), *Productina globosa* Roberts, *Brachythyris davidi* (Dun), *Crassumbo kennedyensis* (Maxwell), *Prospira prima* Maxwell, *Unispirifer* cf. *striatoconvolutus* Qian, *Spirifer* sp. cf *S. sol*, *Kitakamithyris* sp., *Cleiothyridina segmentata* Roberts, ?*Schumardella* sp., *Hamburgia*? *hillae* Qian, *Mitchellinia* sp. This fauna is from the *Tulcumbella tenuistriata* Zone, which is late Famennian to early Tournaisian in age (Roberts & others, 1993). At approximately the same level of the stratigraphy a limestone sample returned conodonts from the early *Duplicata* Zone – early-middle *Crenulata* Zone.

Slightly higher in the sequence, brachiopods found include *Schizophoria* sp., *Retichonetes kennedyensis* (Maxwell), *Rugosochonetes magnus* Maxwell, *Acanthocosta* cf. *teicherti* Roberts, *Brachythyris davidi* (Dun), *Austrochoristites solidus* (Campbell & Engel), *Crassumbo kennedyensis* (Maxwell), *Crurithyris* sp., *Cleiothyridina* sp., *Shumardella* sp., *Angustispatulata campbelli* Qian, terebratulids and spiriferids. These are early to middle Tournaisian (Early Carboniferous) in age, slightly younger than the *Tulcumbella tenuistriata* Zone and older than the *Schellwienella burlingtonensis* Zone.

The brachiopod fauna collected in the Gunpowder Creek area (GR269095 7343998) is the youngest found in the Balaclava Formation. It contains *Schizophoria* sp., *Syringothyris australis* Maxwell, *Cleiothyridina* sp., *Mitchellinia* sp. and spiriferids. This fauna is from the Tournaisian (Early Carboniferous), is younger than the *Tulcumbella tenuistriata* Zone, and is either from the *Schizophoria* or *Schellwienella burlingtonensis* Zone.

Stratigraphic relationships

The Balaclava Formation unconformably overlies the Capella Creek Group. It is a partial lateral equivalent of, and partially overlies, the Mount Hoopbound Formation.

In the Gunpowder Creek area, the Balaclava Formation contains small lenses of oolitic limestone the same age as those at the base of the Rockhampton Group. Elsewhere the Rockhampton Group conformably overlies the Balaclava Formation.

The Balaclava Formation is also overlain by the uppermost Carboniferous to Lower Permian Youlambie Conglomerate. An angular discordance across the contact, and the absence of most of the Carboniferous sequence, supports interpretations of an extensional event close to the Carboniferous–Permian boundary (Holcombe & others, 1997b).

The relationship with the Pomegranate Tonalite is uncertain. Outcrop patterns suggest an intrusive relationship, but no obvious hornfelsing was seen.

Correlation with other units

The Balaclava Formation is largely equivalent in age to the Mount Hoopbound Formation, Lochenbar Formation, and Three Moon Conglomerate. Fossil evidence indicates that deposition of the Mount Hoopbound Formation and Balaclava Formation began at the same time, but the Mount Hoopbound Formation ceased accumulating first, and was eventually overlapped by the Balaclava Formation.

The Balaclava Formation also overlaps the Mount Alma Formation in age. The two formations never come into contact, being separated by the Capella Creek Group. Despite cropping out within 2km of each other, they appear to be sedimentologically isolated, suggesting that the Capella Creek Group was a palaeotopographic high in the Late Devonian, and separated at least the lower parts of the Balaclava Formation and Mount Alma Formation (Blake & others, 1998).

Figure 59. Distribution of the Three Moon Conglomerate

Three Moon Conglomerate (DCt)

(G.A. Simpson, M.A. Hayward & P.R. Blake)

Introduction

The Three Moon Conglomerate was defined by Dear (1968) in the Cania area and described as fine to coarse-grained purplish sandstone, poorly sorted conglomerate, siltstone, and minor basic to intermediate volcanics. Much more extensive outcrops of similar lithologies to the east, in the core of the Tellebang Anticline, were mapped by McKellar (1967) as Unit A of the Crana beds. In their report on the regional geology, Dear & others (1971, page 19) noted the similarity between the two units. Further mapping has confirmed that they can be combined as a single unit, and the name Three Moon Conglomerate is applied to the whole sequence. The remaining part of the Crana beds, Unit B of McKellar (1967), is now included in the Mount Alma Formation.

Distribution

The revised Three Moon Conglomerate crops out as a discontinuous north-west trending belt more than 110km long and up to 25km wide. The unit crops out in five main areas: the Dawson Highway at Collards Creek; north and south of Mount Seaview; Cania; along Monal Creek; and east of Tellebang Mountain east of Mulgildie (Figure 59).

Type section

The type section nominated by Dear (1968) "extends along the ridges north of Kroombit-Oaky Creek…then south into Three Moon Creek…along the south-west of the creek to the junction with First Crossing Creek to the base of the Cania Formation…". This locates the type section between GR294400 7279500 and GR297100 7274800 on the Scoria 1:100 000 Sheet area. The section runs along strike for some distance, is faulted at its base (northern end), and contains ~750m of true thickness. Some of the section is now submerged by Cania Dam but representative outcrops are still exposed along Kroombit–Oaky Creek.

A second representative section was identified in the headwaters of Callide Creek, and was measured by Bryan & others (2001, figure 5). Bryan & others (2001) divided this section into three formations (Monal volcanic facies association, Lochenbar beds and Three Moon Conglomerate), but regionally it can all be included in the Three Moon Conglomerate. Apart from the presence of pillow basalt at the base of the Callide Creek section, the rock types are similar to those in the type section nominated by Dear (1968).

Topographic expression

The Three Moon Conglomerate forms hilly topography.

Geophysical expression

The Three Moon Conglomerate corresponds with areas of low radiometric response in all channels, reflecting derivation from basic to intermediate volcanic rocks. In the Dawson Highway area 25km north-east of Biloela, the distribution of the Three Moon Conglomerate can be determined from images of radiometric data. These show narrow north-west trending linear belts of high radiometric response, now assigned to the Youlambie Conglomerate, surrounded by areas of low radiometric response representing the Three Moon Conglomerate. Field checking confirmed this interpretation. Small areas of moderate to low radiometric response correspond with sandstones, siltstone, and oolitic limestones of the Rockhampton Formation.

The moderate magnetic signature of the Three Moon Conglomerate contrasts with the extremely low response of adjacent sedimentary units, including the Mount Alma Formation, Rockhampton Group, and Youlambie Conglomerate, consistent with its mafic volcanic source.

Lithology

The unit contains fine to coarse-grained, purple sandstone and poorly sorted conglomerate (Figures 60 and 61), grey to green siltstone (Figure 62), minor basic to intermediate lavas including pillow basalt (Figure 63), and basaltic hyaloclastite (Figure 64). In many places the fine-grained sandstone and siltstone are thinly interbedded, and locally rhythmically interbedded, and resemble sequences of interbedded siltstone and sandstone in the Mount Alma Formation (Figure 65). The main clasts are basaltic volcanics (Figure 66). The sandstones locally contain minor amounts of quartz (up to 8%), but this is significantly less than the content

Figure 60. Massive, purple, pebble conglomerate with dominantly basaltic to andesitic clasts, Callide Creek, Three Moon Conglomerate

Figure 61. Massive, poorly sorted boulder conglomerate at the top of the section of Three Moon Conglomerate in Callide Creek. The clasts are dominantly basalt to andesite lavas.

Figure 62. Grey to green siltstone overlying purple volcaniclastic conglomerate, Callide Creek, Three Moon Conglomerate

Figure 63. Pillow basalt in Callide Creek (see Bryan & others, 2001), Three Moon Conglomerate

Figure 64. Basaltic hyaloclastite in Callide Creek (see Bryan & others, 2001), Three Moon Conglomerate

Figure 65. Thinly interbedded volcaniclastic sandstone and siltstone, Coal Road north-east of Biloela, Three Moon Conglomerate

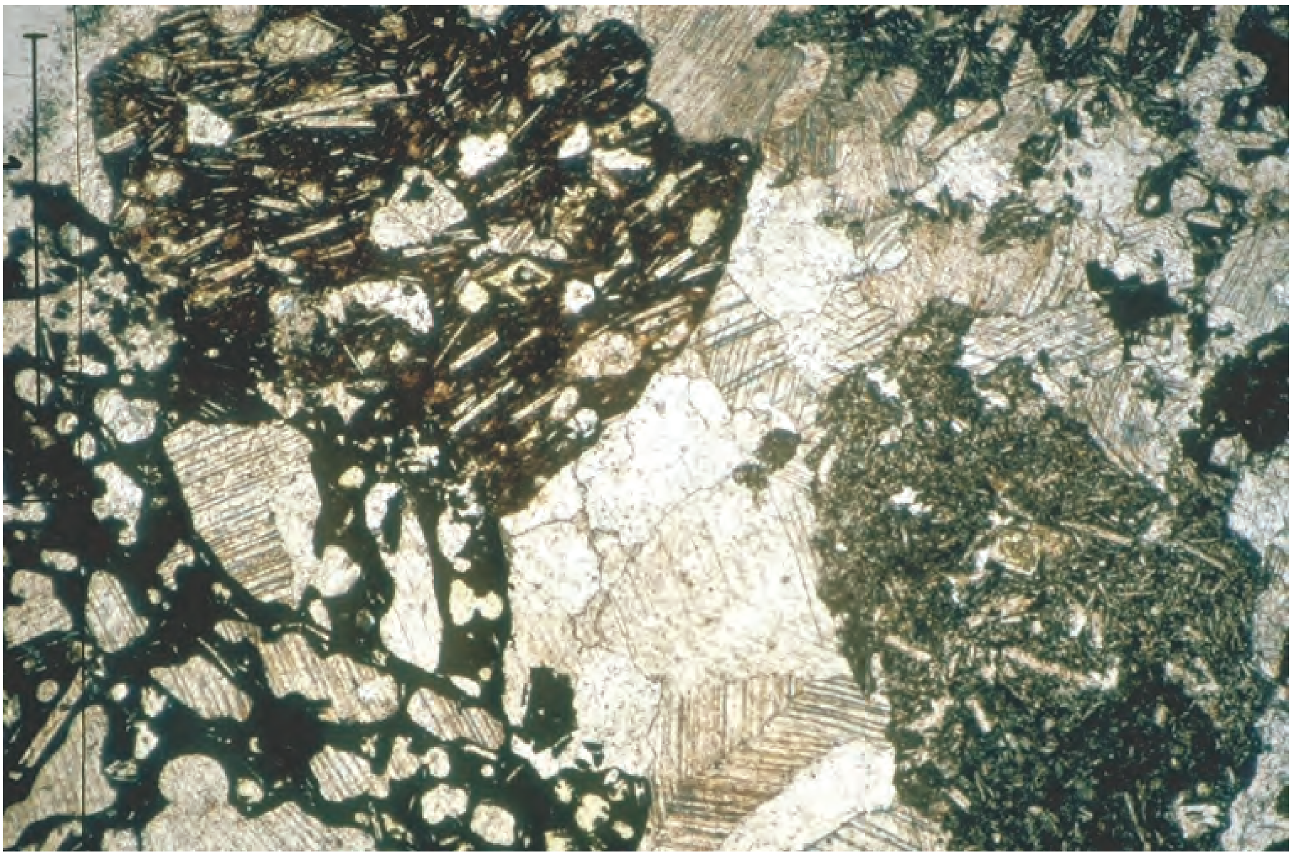


Figure 66. Photomicrograph of volcaniclastic sandstone showing abundant basalt clasts and calcite amygdales and cement, Three Moon Conglomerate. Plane polarised light. Scale bar is 1mm long.

Figure 67. Poorly sorted pebble to cobble conglomerate from the type section near Cania Dam, with abundant hematised siltstone clasts, Three Moon Conglomerate.

observed in more felsic units such as the Balaclava Formation, Rockhampton Group and Youlambie Conglomerate, and is not consistently observed.

Conglomerates consist of subangular to rounded clasts of purple, hematised, aphyric to plagioclase-phyric basaltic to dacitic volcanics in a lithic sandstone matrix. Many conglomerates contain hematised clasts of siltstone/mudstone (Figure 67). Sharp erosive contacts of only local extent have been observed at the base of some conglomerates (Figure 68), and scouring and grading occurs in sandstone beds (Figure 69). Boulder conglomerates with abundant clasts of fossiliferous Late Devonian (Frasnian) limestone occur in Dooloo and Pine Mountain Creeks. Rare nodules of micritic limestone were also observed in a rail cutting near Mungungo (GR315850 7260783) and in a road cutting near Dooloo Creek. No definite autochthonous fossiliferous limestones were found.

Geochemistry

The geochemistry of basalts and basaltic andesites from the Three Moon Conglomerate, as well as the Lochenbar Formation and Mount Hoopbound Formation, has been studied by Bryan & others (2001) and Murray & Blake (2005). Analyses form two groups on REE and multi-element plots, referred to by Murray & Blake (2005) as the Lochenbar suite and the Monal suite. The Lochenbar suite is slightly to moderately enriched in LREE, with $(La/Yb)_N$ ratios from 1.09–2.12, and slightly to moderately depleted in HFSE relative to N-MORB, particularly in Ta, Nb and Zr. Th contents are 1.7–5.8 times that of N-MORB, and $(Th/Nb)_N$ ratios average 10.0 and range from 4.76–16.13. Ce/Y values for the Lochenbar suite suggest a crustal thickness of about 20km (Mantle & Collins, 2008). The geochemical data for the Lochenbar suite support a model in which basalts of the Three Moon Conglomerate, Lochenbar Formation and Mount Hoopbound Formation were generated by similar partial melting processes from the same source rocks, but subsequent fractionation to produce andesites was restricted to the Mount Hoopbound Formation. Discrimination diagrams using immobile elements indicate that the basalts are volcanic arc tholeiites, or are transitional between tholeiites and calc-alkaline basalts.

The Monal suite, which occurs exclusively in the Three Moon Conglomerate, shows significant depletion of LREE with respect to Nd and Sm, and the La to Nd segments of their REE curves are very steep. HFSE ratios

Figure 68. Conglomerate-filled channel scoured in bedded siltstone, Coal Road north-east of Biloela, Three Moon Conglomerate.

Figure 69. Small scour at the base of a graded sandstone bed, Callide Creek, Three Moon Conglomerate

are close to those of N-MORB in multi-element plots, and $(Th/Nb)_N$ ratios are low, averaging 1.77 excluding one sample with a ratio of 8.34, the result of an unusually high Th content.

Murray & Blake (2005) used the basalt geochemistry to determine tectonic setting in two ways: firstly by compilation of bivariant ratio plots; and secondly by direct comparison of individual analyses in REE and multi-element plots. Both methods demonstrate that the Lochenbar suite has compositions typical of evolved oceanic arcs such as the Lesser Antilles, Marianas, Vanuatu, and the Aleutians. However, they also show similarities to samples from the Eastern Volcanic Front of Kamchatka.

The Three Moon Conglomerate has more of a back-arc character than the Lochenbar and Mount Hoopbound Formations. Not only are some basalts of the Monal suite comparable to those of modern back-arc spreading centres like the East Scotia Ridge, but the Lochenbar suite from the Three Moon Conglomerate also includes basalts and basaltic andesites that trend towards transitional arc-back-arc compositions. The fact that compositions become more arc-like to the west provides a strong argument against the simplistic Andean model for the Late Devonian envisaged by Day & others (1978), Henderson (1980) and Murray & others (1987), as well as the model of Bryan & others (2001, 2003) for a back-arc setting for the Upper Devonian rocks behind an eastern arc.

Environment of deposition

Dear (1968) reported indeterminate plant fragments throughout the unit. He recorded marine fossils from a single locality near South Kariboe Creek 6km north of Cania. Marine fossils including the brachiopod *Cyrtospirifer* occur towards the base of the sequence in the headwaters of Callide Creek, and limestone boulders containing Late Devonian corals and other marine fossils have been identified in the Dooloo Creek, Monal Creek, and Cania areas. The occurrence of primary sub-aqueous volcanic rocks (pillow basalts and hyaloclastite) at the base of the succession in the Callide and Monal Creek areas and the Cania area also (Cummings, 1998) suggests that the lower part of the unit was deposited in a marine environment, proximal to a volcanic source.

The Three Moon Conglomerate is interpreted to have been deposited entirely in a marine environment, transitional between the shallow marine to terrestrial environment of the Mount Hoopbound Formation and Lochenbar Formation, and the deeper marine Mount Alma Formation.

Thickness

The base of the Three Moon Conglomerate is not exposed. The entire Callide Creek section measured by Bryan & others (2001, figure 5) is 1300m thick, and is unconformably overlain by the Youlambie Conglomerate. The type section nominated by Dear (1968) in the Cania area is approximately 750m thick.

Structure

The overall structure of the Three Moon Conglomerate is characterised by open folding cut by normal and reverse faults. The folding was probably related to westerly-directed thrusting during the Hunter–Bowen Orogeny (see cross section in figure 2 of Bryan & others, 2001). The largest outcrop belt forms the core of the Tellebang Anticline east of Monto, a broad structure plunging shallowly to the south. This belt, like almost all outcrops of the unit, is fault-bounded.

In the Dawson Highway area 25km north-east of Biloela, fault blocks of Three Moon Conglomerate are surrounded by the Youlambie Conglomerate. The outcrop pattern is explicable in terms of doubly plunging, gentle folds offset by minor reverse faulting (R.J.Holcombe, personal communication, 1998). Strong cleavages in the area may be related to the reverse faulting.

Biostratigraphy and age

Dear & others (1971) listed a fauna in the Cania area containing *Schizophoria* aff. *resupinata* (Martin), *Leptogonia* sp., *Adolfia* ?*kennedyensis* (Maxwell), *Perissothyris* cf. *masonensis* Carter, and *Panenka* cf. *porteri* Benson & Dun. This fauna was identified as Early Carboniferous and considered to have affinities with faunas in the Pond Formation (now the upper part of the Balaclava Formation) on the Rockhampton 1:250 000 Sheet area described by Kirkegaard & others (1970). The most recent revision of the faunal succession in Carboniferous rocks of eastern Australia (Roberts & others, 1993) suggests an age straddling the Devonian–Carboniferous boundary.

Fossils found in the upper part of Callide Creek include *Cyrtospirifer* sp. and indicate that deposition commenced during the Late Devonian. This is supported by Late Devonian corals and conodonts from limestone clasts collected from conglomerates in the Dooloo Creek and Pine Mountain Creek areas.

Stratigraphic relationships

In the Cania area, Dear (1968) considered the Three Moon Conglomerate to overlie the uppermost Devonian (Famennian) Dawes Range Formation, now incorporated into the Lochenbar Formation. Contacts between these two units are faulted. Elsewhere, the base of the Three Moon Conglomerate is not exposed.

The Three Moon Conglomerate is overlain by the Rockhampton Group with apparent conformity in the Monal Creek, Dawson Highway, and Cania areas. Regionally it is unconformably overlain by the Lower Permian Youlambie Conglomerate.

Correlation with other units

The unit is a partial time equivalent of the Mount Alma Formation and Balaclava Formation, and possibly of the Mount Hoopbound Formation. It has been interpreted to overlie the Lochenbar Formation in the Cania area, although contacts are faulted. The Three Moon Conglomerate contains more abundant coarse-grained facies than the Mount Alma Formation and is distinguished by the presence of purple hematised clasts in the sandstones and conglomerates. The more consistent presence of minor quartz and the greater proportion of finer-grained facies and rip-up clast conglomerates can be used to distinguish the unit from the Mount Hoopbound and Lochenbar Formations. It had a more mafic volcanic provenance than the Balaclava Formation, which also contains much more prolific marine faunas.

McKellar (1967) and Dear & others (1971) mapped the sequence forming the core of the Tellebang Anticline east of Monto, and underlying the oolitic limestone-bearing Caswell Creek Group (now included in the Rockhampton Group), as the Crana beds. The Crana beds were subdivided into a near shore facies (Unit A) to the west, and a more distal, deeper water facies (Unit B) to the east, with a transitional boundary between them. Dear & others (1971, page 19) noted the similarity of Unit A of the Crana beds to the Three Moon Conglomerate defined by Dear (1968) in the Cania area. Subsequent mapping has confirmed that they can be combined as a single unit, and the name Three Moon Conglomerate is applied to the whole sequence. The remaining part of the Crana beds, Unit B of McKellar (1967), is now included in the Mount Alma Formation. The lack of hematised clasts in rocks of the Mount Alma Formation, combined with a more felsic volcanic source, suggests that this unit had a different provenance from the Three Moon Conglomerate.

Bryan & others (2001, figures 3 and 5) subdivided the sections in upper Monal Creek – Dooloo Creek, upper Callide Creek, and at Cania into three units: the Monal volcanic facies association at the base, overlain successively by the Lochenbar Formation and the Three Moon Conglomerate. On a regional scale, however, it is not possible to recognise these subdivisions. Further south along Monal Creek, basalts are interbedded with typical sediments of the Three Moon Conglomerate, demonstrating that they are not confined to the base of the unit. In Callide Creek, sedimentary facies within the hyaloclastite sequence are similar to those in the overlying portion of the Three Moon Conglomerate.

Economic significance

Elevated gold values occur in the Dooloo Creek area, but no economic deposits have been discovered.

Mount Alma Formation (DCa)

(P.R. Blake & G.A. Simpson)

Introduction

The Mount Alma Formation was named and briefly described by the Yarrol Project Team (1997). It combines several previously mapped units that were thought to range in age from latest Silurian to Early Carboniferous. These include: parts of undifferentiated Silurian–Devonian and Lower Carboniferous units on the Duaringa 1:250 000 sheet (Malone & others, 1969); undivided Lower Carboniferous rocks and much of the Lower Devonian Mount Holly beds on the Rockhampton 1:250 000 sheet (Kirkegaard & others 1970); most of the Lower Devonian Barmundoo beds and unit B of the Lower Carboniferous Crana beds on the Monto 1:250 000 sheet (Dear & others, 1971); and the Melrose beds west of Ridgelands (Simpson, 1995). Morand (1993a) had previously noted the similarity of the Barmundoo beds to adjacent Lower Carboniferous units, including the Crana beds.

Figure 70. Distribution of the Mount Alma Formation

Distribution

The Mount Alma Formation forms an arcuate, discontinuous belt 260km long and up to 30km wide from the Fitzroy River north-west of Ridgelands to Monto (Figure 70). Equivalent rocks extend onto the Mundubbera 1:250 000 sheet area where they have been mapped as the Crana beds (Whitaker & others, 1974).

Derivation of name

The unit is named after Mount Alma, which is on the southern part of the Bajool 1:100 000 Sheet area, at GR276653 7348507, and where typical outcrops of the unit are exposed.

Type section

The type section for the Mount Alma Formation was measured along the road leading to the radar installation at the top of Mount Alma. The section begins at GR279450 7349596 and finishes at GR278510 7349394 on the Bajool 1:100 000 Sheet area. This section does not include the top or the base of the formation, but was selected because of the excellent exposure and easy accessibility. Although the sequence here has volumetrically more conglomerate than is typical for most areas of Mount Alma Formation, the rock types present are typical and well exposed in the road cuttings.

A second representative section has also been measured along Takilberan Creek, south-east of Calliope on the Calliope 1:100 000 Sheet area at GR298301 7322411. This section exposes the top of the unit but not the base, and contains an unusually high proportion of coarse-grained sediments.

Topographic expression

Generally, the Mount Alma Formation has a subdued topography, with the best exposures found in creeks and road and railway cuttings. However, in areas where conglomerates are common within the unit (such as at Mount Alma where they form approximately half of the exposed sequence), the topography is very hilly.

Geophysical expression

On red-green-blue images derived from airborne radiometric survey data, the Mount Alma Formation gives very low values in all three channels, except towards the top of the formation in the core of the Alma Syncline where a weak red (potassium) signature can be seen.

Several radiometric anomalies approximately 10km north-west of Mount Alma, in contact rocks of the Cecilwood Quartz Diorite, are high in all three channels (potassium, uranium and thorium), and are attributed to skarns formed by granitic intrusions

The unit gives virtually no magnetic response, and features seen within the areas mapped as Mount Alma Formation are interpreted to represent either the underlying basement rocks or intrusions.

Lithology

The Mount Alma Formation is generally very uniform throughout its outcrop area, but appears to be finer-grained in northern outcrops compared with the type section. The base of the formation is an exception, however, because locally the rock types and their proportions differ from those typical of the unit as a whole.

Typical Mount Alma Formation is comprised dominantly of interbedded siltstone and fine-grained sandstone, with minor medium- to very coarse-grained sandstone and conglomerate. The siltstone and fine-grained sandstone are usually thinly and rhythmically interbedded (Figures 71 and 72). Fining-upwards sequences are a very common feature in the thin sandstone beds, as at Takilberan Creek at GR298301 7322411 on the Calliope 1:100 000 sheet. Ripple laminations and other current formed sedimentary structures are rare, but soft sediment deformation features are very common (Figures 73, 74 and 75). The top of one sandstone bed displays ripple marks with reversed asymmetry, possibly indicating that the bed was deposited as a turbidite. Marine macrofossils have not been found within these interbedded siltstone-sandstone sequences, and plants and trace fossils are rare (Figure 76). Microfossils such as radiolarians are commonly seen in thin sections of the siltstones but processing failed to recover any useful specimens. The interbedded sandstone and siltstone sequences crop out poorly.

Figure 71. Thinly interbedded sandstone and siltstone, type section, Mount Alma, Mount Alma Formation

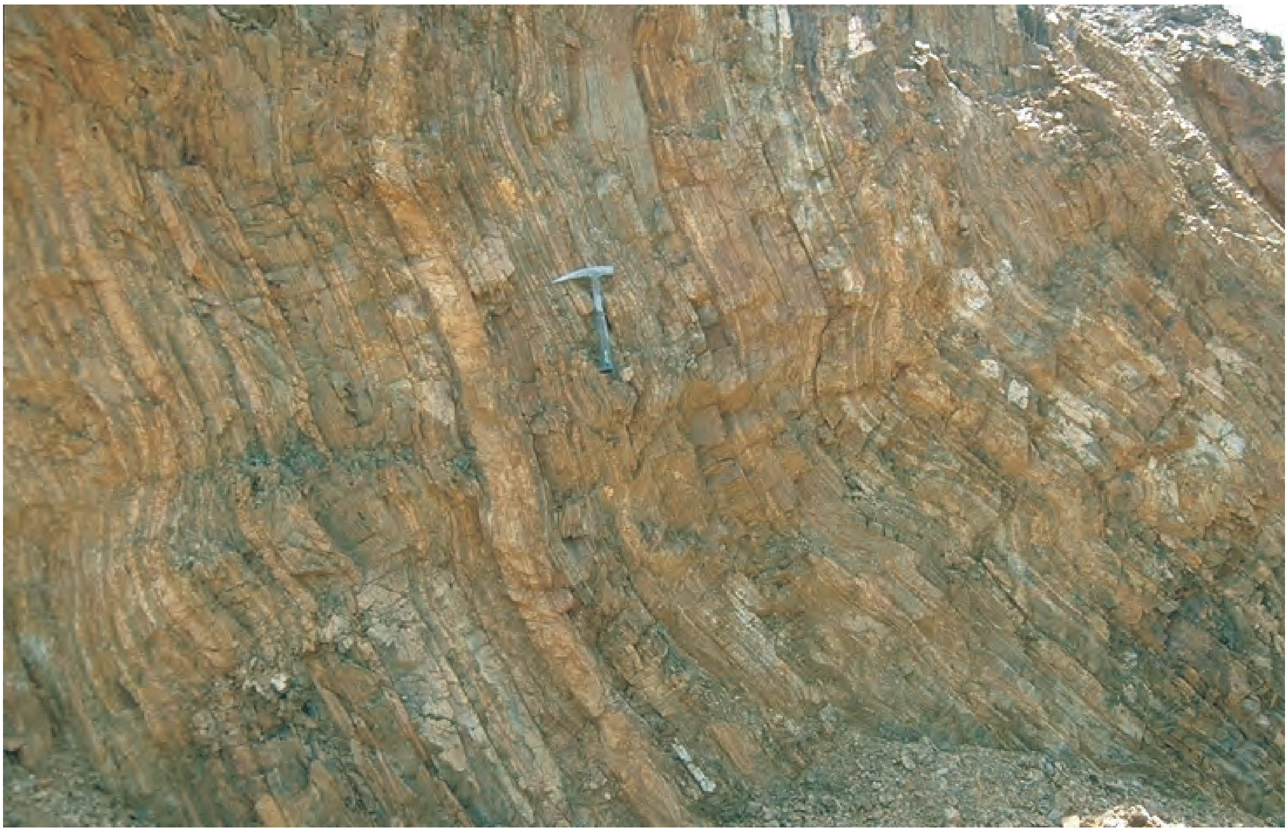


Figure 72. Interbedded sandstone and siltstone, Mount Alma Formation

Figure 73. Slump-folding and loading in siltstone, Mount Alma Formation

Figure 74. Soft sediment deformation in siltstone, Takilberan Creek, Mount Alma Formation

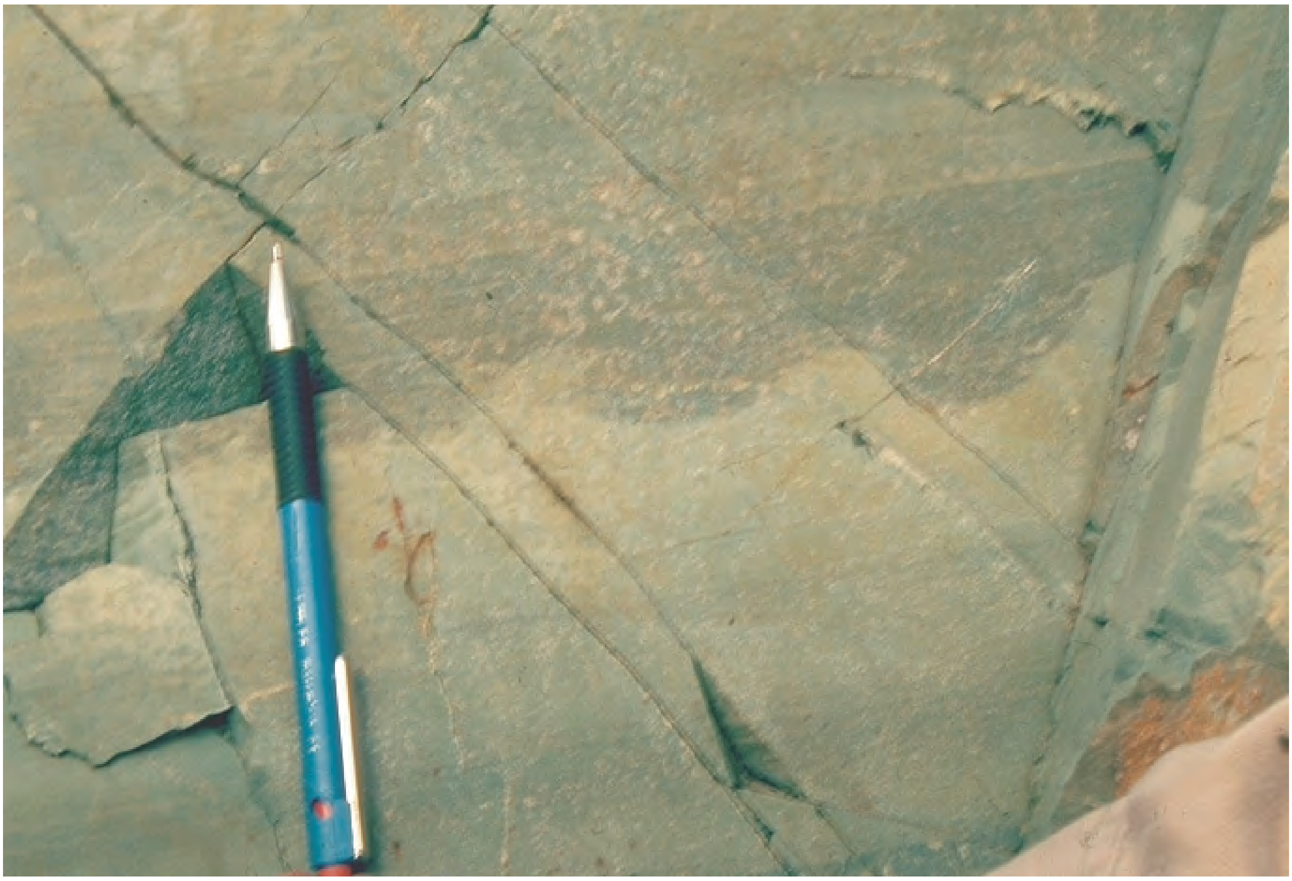


Figure 75. Flame structures, railway cutting, 16km west of Calliope, Mount Alma Formation

Figure 76. Lepidodendroid stem, Mount Salmon area, Mount Alma Formation

Conglomerates and medium- to very coarse-grained sandstones are a lesser component of the sedimentary sequence but are distinctive because of the widespread occurrence of siltstone rip-up clasts (Figures 77, 78 and 79). The conglomerates consist dominantly of subangular to subrounded clasts of intermediate to felsic volcanics and fine-grained sedimentary rocks. Rip-up clasts vary from sparse to abundant, and range in size from ~2cm to 3m long. Although they are usually a minor component of the Mount Alma Formation, locally the rip-up clast conglomerates constitute a large proportion of the unit. In the type section at Mount Alma and the representative section along Takilberan Creek, conglomerates make up almost 50% of the sequence. The conglomerates contain rare brachiopods, crinoids and corals as clasts. The rip-up clast conglomerates are very competent and crop out well, causing this rock type to be over-represented in outcrop compared to the interbedded sandstone and siltstone sequences. Medium to coarse-grained sandstones are best developed towards the base of the Mount Alma Formation, where they are more abundant than the typical sequences of interbedded siltstone and fine-grained sandstone.

Minor rock types within the Mount Alma Formation include thin autochthonous limestone, rounded clast conglomerates, and allochthonous limestone blocks. The autochthonous limestones are usually micritic with no obvious macrofossils, up to several centimetres thick, and occur within sequences of thinly interbedded siltstone and very fine-grained sandstone. The conglomerates are usually clast supported and contain well rounded, pebble to cobble sized clasts of granitoids, intermediate to felsic volcanics, and sandstone (Figure 80). Some conglomerates also contain sparse to abundant clasts of limestone.

Allochthonous Lower and Middle Devonian limestones are more common at the base of the Mount Alma Formation, where they form blocks up to a kilometre long, but some occur higher in the sequence within rocks that are considered more typical of the unit. Several features demonstrate that these limestones are allochthonous. They contain coral and conodont faunas of Early Devonian age that are the same as those in the Erebus beds, indicating their probable source. However, fossils collected from conglomerates at the base of the Mount Alma Formation indicate that deposition began in the Late Devonian. This significant age difference can only be explained if the limestones are allochthonous. Also, bedding within the limestone blocks is not parallel to bedding in the surrounding sediments. In addition, bedding in some limestones does not correspond to their outcrop pattern. An example is the limestone at GR290022 7353774 on the Bajool 1:100 000 Sheet, which is a flat lying block, whereas bedding readings within the block dip very steeply.

Figure 77. Rip-up clasts overlain by fining upward sandstone bed, and interlaminated siltstone and mudstone showing graded bedding, Mount Alma Formation. The succession may represent a partial Bouma sequence associated with distal turbidite facies.

Figure 78. Rip-up siltstone clast showing bedding at a high angle to the elongation of the clast, railway cutting, 16km west of Calliope, Mount Alma Formation

Figure 79. Rip-up bed/clast in coarse sandstone, Takilberan Creek, Mount Alma Formation

Figure 80. Matrix supported cobble conglomerate with rounded clasts mainly of felsic volcanics but including subordinate andesite and pale green quartz-veined siliceous rocks, Mount Alma Formation.

In the Mount Salmon area north-west of Rockhampton, the Mount Alma Formation is finer-grained, and is dominated by fissile micaceous siltstone and shale.

Environment of deposition

No wave-formed sedimentary structures have been found within the formation, indicating that the formation was deposited below wave base.

Photic zone macrofossils in the Mount Alma Formation occur within rip-up clast conglomerates, and in a limestone bed at the very top of the formation in the Cannindah area at GR325750 7248499 on the Monto 1:100 000 Sheet. The fossils within rip-up clast conglomerates include crinoids, brachiopods, and corals, and are interpreted to have been carried into the deeper water environment of the Mount Alma Formation from areas of shallow shelf deposition. Fossils at the top of the unit in the Cannindah area may also have been transported, since the brachiopod and crinoidal material is fragmentary. The contact between this limestone and overlying oolitic limestone of the Rockhampton Group may represent the onset of a regional shallowing event that marks the boundary between the Mount Alma Formation and the Rockhampton Group. The absence of *in situ* photic zone fossils, except possibly in the Cannindah Creek area, indicates that the Mount Alma Formation was dominantly deposited below the photic zone.

Palmetolepis dominates over *Polygnathus* in conodont faunas extracted from thin autochthonous limestones, and this ratio of abundance typically indicates outer shelf water depths (B.G. Fordham, personal communication).

This evidence, and further indicators such as the thickness of gravity flow deposits and the abundance of radiolarians in siltstone beds, suggests that the formation was deposited in an outer shelf to basinal environment.

Thickness

The type section, which is only a small part of the overall sequence, is 640m thick. Using the dip angle of the beds between the Queenslander Fault and the axis of the Alma Syncline, and assuming no repetition or removal of strata, the Mount Alma Formation is estimated to be between 3 and 5km thick. In the Belgamba region in the north-western corner of the Bajool 1:100 000 Sheet area, both the top and bottom of the unit are exposed, and a thickness of 2.3km is calculated from the width of outcrop and average dips.

Structure

Morand (1993b) interpreted rocks now mapped as the Mount Alma Formation as part of a west-vergent fold and thrust system with folds becoming tighter and cleavage more intense towards the Yarrol Fault to the east. This is probably an over-simplification of the variation in the intensity of deformation, which appears to be related to proximity to major faults within and marginal to the unit, such as the Ambrose, Bracewell, Ulam and Queenslander Faults. The Mount Alma Formation displays only weak deformation in most places, forming large, open synclines and anticlines, exemplified by the eastern part of the section along the Gladstone-Moura railway shown by Morand (1993a, figure 7). A gently south-plunging anticline occurs in Marble Creek north-west of Nagoorin (Dear & others, 1971, page 12). A weak cleavage best developed in the fine-grained sediments dips steeply towards the east-north-east, and is parallel or subparallel to the cleavage in the Erebus beds (Figure 81).

In some zones the deformation is more intense. An example occurs in a railway cutting at Bajool (GR259927 7382401), just west of the Queenslander Fault, where the Mount Alma Formation is deformed into relatively tight folds. Another series of isoclinal folds was noted by Dear & others (1971, page 12) in Diglum Creek just west of the Ambrose Fault which forms the eastern boundary of the unit. Also, in some localities the cleavage can be strongly developed and can even be seen in conglomerates.

Biostratigraphy and age

Brachiopods collected from the rip-up clast conglomerates at the base of the formation indicate that deposition of the Mount Alma Formation began at the beginning of the Late Devonian (early Frasnian). Conodonts extracted from thin, autochthonous limestones at different levels of the formation have also returned Late Devonian (late Frasnian to Famennian) ages. Conodonts from limestones at the top of the unit in the Cannindah area indicate a latest Devonian to earliest Carboniferous age (very late Famennian to early Tournaisian). Shelly faunas from just below this limestone horizon that were considered to be late Tournaisian by McKellar (1967) are now regarded as early Tournaisian, but not at the base of the Carboniferous (Roberts & others, 1993).

Figure 81. Small reverse fault parallel to cleavage in thinly bedded siltstone, wall of adit near Raglan, Mount Alma Formation

In the northern part of the outcrop area west of Ridgelands, conodonts were recovered from both the Mount Alma Formation and Rockhampton Group (Simpson, 1995). These conodonts revealed an upper Late Devonian (*crepida* to *trachytera* zone) age from the Mount Alma Formation and an Early Carboniferous (*duplicata* zone) age for the lower Rockhampton Group. These age determinations support evidence for a local disconformity in the area proposed by Simpson (1995).

Stratigraphic relationships

The Mount Alma Formation unconformably overlies the Erebus beds, a relationship clearly demonstrated by the occurrence of allochthonous limestone blocks. A similar relationship with the Craigilee beds is indicated by the presence of fossiliferous Siluro–Devonian limestone clasts in the sequence between Eight Mile and Ten Mile Creeks (Kirkegaard & others, 1970, page 45). The relationship between the Raspberry Creek Formation of the Capella Creek Group and the Mount Alma Formation on the eastern limb of the Gracemere Anticline has been described as structurally conformable (Golding & others, 1994a; Morand, 1993b, page 263). However, at Belgamba, in the north-western part of the Bajool 100 000 Sheet area and north-eastern part of the Mount Morgan 1:100 000 sheet area, bedding trends within the Mount Alma Formation appear to have a small angular discordance to bedding within the Raspberry Creek Formation. Further evidence for an unconformable relationship is shown by the strong development of cleavage in the Raspberry Creek Formation near Ayrdrie homestead in the southern part of the Bajool 1:100 000 Sheet area, whereas the overlying Mount Alma Formation in this area, much of it of finer-grain size, is not cleaved.

The Rockhampton Group overlies the Mount Alma Formation with apparent regional conformity. In the Cannindah area a sharp, regular, apparently conformable contact was observed between a wackestone to packstone of the Mount Alma Formation and an oolitic grainstone of the Rockhampton Group. Conodont ages from these two limestones show that there was no significant depositional hiatus between them. West of Ridgelands on the Rockhampton 1:250 000 Sheet, a local disconformity was interpreted by Simpson (1995) based on conodont age determinations and the apparent termination of a limestone bed of the Rockhampton

Group against the Mount Alma Formation – Rockhampton Group boundary. This disconformable relationship was not observed elsewhere.

On a regional scale, the contact between the Mount Alma Formation and Rockhampton Group is almost certainly diachronous. As well as the local presence of Early Carboniferous shelly faunas, some sandstones in sequences previously mapped as Crana beds, now included in the Mount Alma Formation, contain dispersed ooids (Batchelor, 1970), suggesting an Early Carboniferous age and contemporaneity with basal sequences of oolitic limestones in the Rockhampton Group.

Correlation with other units

The Mount Alma Formation is at least a partial time equivalent of the Three Moon Conglomerate, Balaclava Formation, and Mount Hoopbound Formation.

In some locations the basal part of the Mount Alma Formation resembles the Three Moon Conglomerate in the Dooloo Creek area. These units have several features in common, and where the Three Moon Conglomerate is mapped in contact with the Mount Alma Formation on the Monto 1:100 000 Sheet, this contact is interpreted as transitional.

The Mount Hoopbound and Balaclava Formations contrast strongly with the Mount Alma Formation in that they were deposited in shallow marine to terrestrial environments.

EARLY CARBONIFEROUS

Rockhampton Group (Cr)

(G.A. Simpson, G.E. Webb, & S. Lang)

Introduction

The Lower Carboniferous Rockhampton Group is the best studied stratigraphic unit in the Yarrol Province. It is dominated by sandstone and siltstone, but is characterised by the repeated development of oolitic limestones and ooid-bearing sandstones.

Previous authors named five Lower Carboniferous units characterised by ooid-bearing intervals in the Yarrol Province based on mapping done for their post-graduate theses. These units were the Neils Creek Clastics in the Mount Morgan area (Maxwell, 1953), Rockhampton Group in the Neerkol area (Fleming, 1960, 1967), Caswell Creek Group in the Cannindah Creek-Yarrol area (McKellar, 1967), Cannindah Limestone in the Old Cannindah area (Jull, 1968), and Cania Formation in the Cania area (Dear, 1968). Although the type areas of all five units derived from postgraduate thesis mapping are geographically widely separated, subsequent regional geological mapping (Kirkegaard & other, 1970; Dear & others, 1971) has extended the units laterally, so that they almost meet (Yarrol Project Team, 1997, figure 7). Regionally, the units clearly represent a single depositional package.

The faunas of the five units are similar, lithological variation is not distinctive, and they all are characterised by the presence of oolitic limestones and ooid-bearing sandstones. Conodont biostratigraphy suggests that, although the duration of deposition in individual areas may have varied, each sequence was deposited during a period of relatively continuous sedimentation in the early Carboniferous (Mississippian). The units are difficult to distinguish other than by their location. They were therefore combined in a single regional stratigraphic unit, termed the Rockhampton Group, by the Yarrol Project Team (1997). This extension of the Rockhampton Group throughout the Yarrol Province has not required any change to the initial concept of the unit as originally proposed by Fleming (1967) and derived from the Rockhampton Series identified by Daintree (1870).

Fleming (1967) recognised three major constituent stratigraphic units within the Rockhampton Group that he named the Gudman Oolite, Malchi Formation (including the Cargoogie Oolite Member), and Lion Creek Limestone. These formations cannot be reliably recognised outside of the Stanwell area, and everywhere else the ooid-bearing units are only mapped as undivided Rockhampton Group, although McKellar (1967) and Dear & others (1971) correlated the Herring Gully Limestone Member at the base of the Caswell Creek Group in the Yarrol Syncline with the Gudman Oolite.

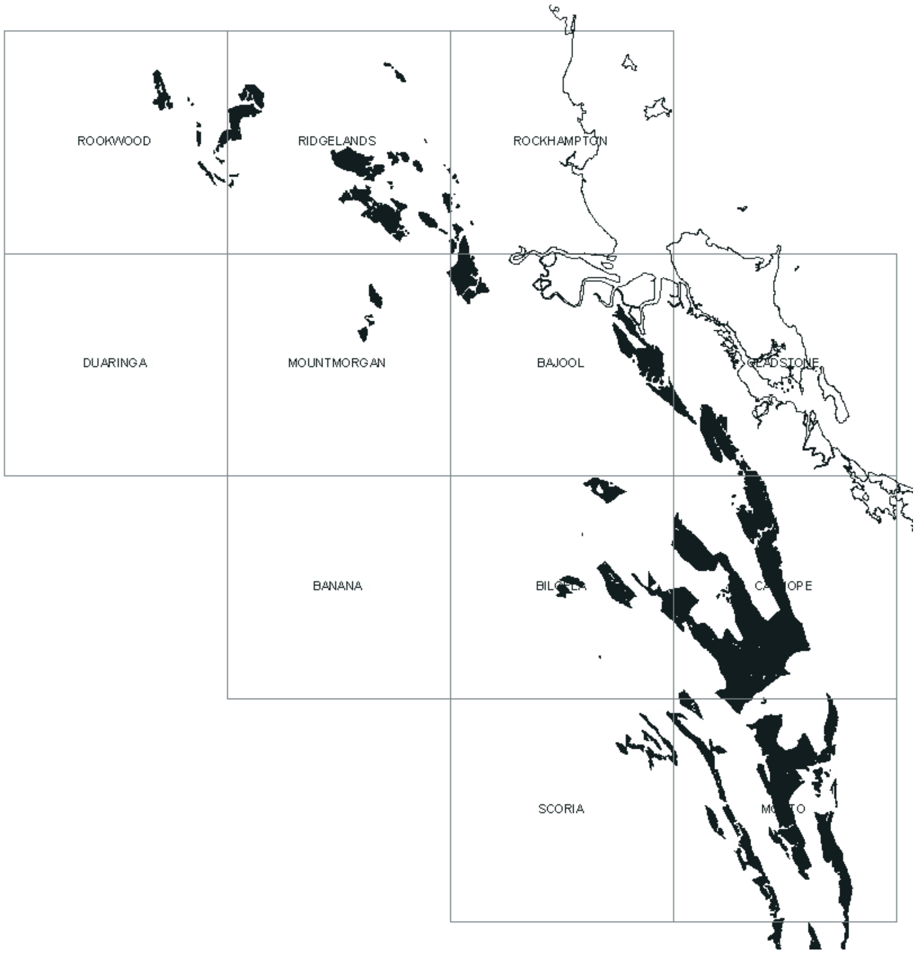


Figure 82. Distribution of the Rockhampton Group

Distribution

The Rockhampton Group is the most widely distributed unit in the Yarrol Province. It forms a discontinuous belt of scattered outcrops 260km long and up to 55km wide that continues south onto the Mundubbera 1:250 000 sheet (Figure 82).

Type section

Fleming (1967) nominated type sections for each of the constituent units in the type area at Malchi Creek near Stanwell. Subsequent studies have modified the upper boundary of the Group in the type section of the Lion Creek Limestone west of Mount Lion. The boundary has been moved up sequence (to the south-west) based on the radiometric signature and on the presence of ooid-bearing limestone and sandstone in the sequence above the Lion Creek Limestone. The limestone contains corals of coral assemblage D of Webb (1990), and is equivalent to the latest Visean FC5 limestones at the top of the Rockhampton Group in the Mount Salmon area to the west (Webb, 1990).

Topographic expression

The Rockhampton Group typically forms rolling hills with limestones and siliceous sandstones forming strike ridges extending over 1–3km. Therefore, the unit commonly shows defined bedding trends on aerial photographs. Limestone-dominated intervals typically show as grey to white ridges, commonly covered in dense scrub, whereas conglomerate beds produce a reddish brown soil. Overall the unit develops a thin, poor, grey-brown soil cover.

Geophysical expression

The Group has a moderate to high response in all three radiometric channels that is higher than any response from underlying units but lower than that from the overlying Lorray Formation and Youlambie Conglomerate. Some parts of the Balaclava Formation give a similar radiometric response. The Rockhampton Group has a low magnetic response except for some limestone intervals that have an increased magnetite content due to fluid interaction and skarn development.

Lithology

The Rockhampton Group consists of a suite of rocks ranging in age from Tournaisian to latest Visean (Lower and Middle Mississippian, or early Carboniferous). This suite includes calcareous, ooid-rich, medium- to coarse-grained sandstone; calcareous, fossiliferous, fine- to very coarse-grained sandstone; ooid-bearing fine to medium feldspathic to feldspatholithic sandstone; dark grey to grey siltstone and mudstone; cream to light blue-grey, cherty mudstone and siltstone; polymictic conglomerate; oligomictic conglomerate; oolitic limestone; bioclastic limestone; and pisolitic limestone. The Rockhampton Group is characterised by the development of oolitic and pisolitic limestones and ooid-bearing sandstones.

Dark grey to grey siltstone and mudstone and cream to light blue-grey cherty mudstone and siltstone dominate the Group. The light blue-grey cherty mudstone and siltstone distinguish the Rockhampton Group from the underlying Mount Alma Formation, which locally is dominated by fissile micaceous siltstone and shale. Intraformational siltstone rip-up clasts from the two units can be distinguished along similar criteria, and conglomerates containing rip-up clasts in the Rockhampton Group also commonly, and characteristically, have ooids in the sandy matrix.

Basic to felsic volcanic clasts are abundant in sandstones and conglomerates of the Rockhampton Group. Thin section analysis indicates a higher proportion of felsic and intermediate clasts compared to basic clasts in coarse-grained facies of the unit. This supports field evidence that the sedimentary rocks in the unit contain a higher proportion of quartz and felsic volcanic clasts than similar rocks in the underlying Three Moon Conglomerate and Mount Alma Formation.

The presence of shards in sandstones shows that volcanism was active throughout the deposition of the Rockhampton Group (Jell, 1961; Radke, 1971). However, primary volcanics are virtually restricted to the westernmost outcrops between Biloela and Monto, where they locally make up the dominant exposed rock types, and consist of andesitic to rhyolitic lavas and pyroclastics, including sparse rhyolitic ignimbrites (Dear, 1963, 1968; Radke, 1971; Webb, 1973, 1977; Cummings, 1998; Bryan & others, 2001). They prove the existence of a volcanic centre or centres in this area. McKellar (1966, 1967) found no volcanics in the western limb of the Yarrol Syncline, and no primary volcanics have been reported from the area north of Mount Morgan (Maxwell, 1952, 1953; Fleming, 1960; Kirkegaard & others, 1970; Simpson, 1995) or anywhere along the eastern margin (Jell, 1961; Kirkegaard & others, 1970; Willmott & others, 1986).

Facies

Simpson (1995) identified 11 sedimentary facies in the Rockhampton Group near Mount Salmon. These are described in Table 3, together with additional facies recognised mainly in the Gracemere–Stanwell area. This facies scheme has been extended to other areas. Table 3 also outlines the dominant processes involved in their deposition, and photographs showing features of these facies are shown in Figures 83 to 92.

Facies Associations

Six facies associations occur in the Rockhampton Group. The facies in these associations are considered genetically related because their common juxtaposition indicates that they were in close proximity to each other during deposition. The six associations identified are Gravity Flow, Coastal Gravelly Channel, Sandy Shelf, Muddy Shelf, Fossiliferous Muddy Shelf, and Carbonate Bank deposits.

Gravity Flow Deposits — Association 1

Description

Association 1 is an overall fining upward succession with an erosional basal contact followed by conglomerates of *Facies C1*, interbedded sandstones and siltstones of *Facies S1*, and mudstones and siltstones of *Facies F1*. Conglomerates at the base of the association contain evenly distributed, imbricated clasts of tuffaceous siltstone and mudstone in a sandy matrix of ooids, skeletal debris, and volcanolithic fragments. *Facies C1* is locally laterally extensive (up to 1km) but generally interfingers with oolitic limestone (*Facies L2*), or is interbedded with sandstone and siltstone of *Facies S1* and *F2*. Siltstone and sandstone of *Facies S1* and *F2* represent the development of A, B and C sections of the Bouma cycle. The association may be very thin (<10m) or thick (150m) and may contain all facies described in the association or just *Facies C1*. The association occurs rhythmically in Stratigraphic Sections 2 and 3 and to a lesser extent, Section 4 described by Simpson (1995) in the Mount Salmon area.

Table 3. Facies description and interpretation

Lithofacies	Description	Process and Depositional Environment
FINE-GRAINED FACIES		
Mudstone/Siltstone (F1)	Interlaminated mudstone and siltstone showing graded to massive bedding, and rare trace fossils (Figure 83). The facies is commonly associated with oolitic limestones and calcareous sandstones of *Facies L1*. The facies is more abundant in the more easterly outcrops of the unit in the Yarrol Project area (e.g. Gracemere and Cannindah Creek areas).	The occurrence of marine trace fossils associated with ooid-rich facies indicates the sediments were deposited in a *marine setting*. Graded bedding and interlamination in such fine-grained marine units is probably the result of settling from suspension.
Interlaminated Siltstone, Sandstone, and Mudstone (F2)	Interlaminated siltstone and mudstone and interbedded feldspatholithic arenite. Bedding is typically flat to wavy and may exhibit soft sediment deformation structures, graded bedding, and ripple cross-lamination. Grazing trails occur (possibly *Cruziana* ichnofacies) in some localities. Bouma sequences were identified in some areas, particularly when the facies was overlying intraformational conglomerates of *Facies C1*. These sequences include facies divisions F7 and F9 of Mutti (1992). Slumping was evident in some areas (Figure 84). This facies is more abundant in the Malchi Formation and in more easterly outcrops of the unit such as the Gracemere or Cannindah Creek areas. The feeding trace fossil *Lophoctenium*, characteristic of the *Zoophycos* ichnofacies, is present in turbidites of the easternmost part of the unit at Belmont north of Rockhampton (Figure 85)	Current ripple laminations suggest that tractional currents were operating during deposition. These structures and associated trace fossils indicate the sediments were deposited in *subtidal marine shelf conditions, near storm wave base*. In some instances, partial Bouma sequences were observed that indicate deposition from density currents. Similar deposits, however, have been observed on other ancient shelf deposits where they were interpreted as storm induced density currents (Reineck & Singh, 1980; Greenwood & Davis, 1984; Middleton & Hampton, 1976). Where these sediments exhibit graded bedding and minor bioturbation, but do not exhibit partial Bouma sequences, settling from suspension may have been the dominant process of deposition. Suspension, storm induced density currents, and gravity flows are all interpreted to have occurred episodically in the study area. The local presence of the *Zoophycos* ichnofacies suggests that the depositional environment deepened towards the east to outer continental shelf or upper continental slope (Seilacher, 1967), consistent with the occurrence of *Lophoctenium* in other Carboniferous sequences (Chamberlain, 1971; Eagar & others, 1985; Orr & others, 1996).
Fossiliferous Siltstone and Sandstone (F3)	Interlaminated siltstone and mudstone and interbedded feldspatholithic arenite. These beds are locally graded and are characteristically fossiliferous, particularly in the upper part of the sequence. They have a calcareous composition and contain a fauna that includes brachiopods, gastropods, crinoids, and bryozoa as well as feeding trails (Figure 86). Bedding is generally flat to wavy and may exhibit soft sediment deformation structures. The facies is more abundant in the lowermost and uppermost part of the Rockhampton Group including the Gudman Oolite and Lion Creek Limestone.	The presence of a substantial marine fauna and the dominance of feeding trails over burrowing trace fossils indicates that the sediments were deposited in relatively *quiet marine conditions, near or below storm wave base*. The presence of these fossils and graded bedding indicate that the facies was largely deposited by settling from suspension drift.
SANDY FACIES		
Interbedded Sandstone and Siltstone (S1)	Interbedded very fine to fine-grained feldspatholithic arenite and siltstone. The arenite beds contain minor ooids and exhibit symmetrical (Figure 87) and asymmetrical ripple cross-lamination and trace fossils. The siltstone is white and tuffaceous to grey and siliceous and the sandstone also contains a large amount (in excess of 30%) of angular, fresh, volcanic quartz and feldspar. The matrix is very fine and has a low birefringence. Some crystalline material in the matrix has the appearance of devitrified glass shards and most of the matrix may represent devitrified and altered material. The rocks commonly contain calcite cement (5-10%). Some beds and laminae show normal grading. Bedding is thin to laminar. This facies commonly has an overall coarsening–upward nature. There are higher amounts of quartz found in western outcrops of the unit compared to the east. Also, there is an increased abundance of quartz in the lower part of the sequence.	The abundance of angular beta-quartz and feldspar crystals, and bed thickness (30cm) of this facies suggest an ash - fall origin. The presence of minor (<10%) amounts of ooids, wave and current ripple laminations, and bioturbation suggests that the sediments were reworked and were deposited in a *shallow marine shelf environment at, or above fair weather wave base*. Development of current ripples is the result of tractional processes indicating that the reworking of sediments occurred during periods of moderate to high energy conditions. Associated graded bedding and feeding trace fossils indicate deposition in a lower energy shelf environment receiving sediment from suspension. It is likely that both suspension and tractional processes were operating simultaneously, as on modern shelves described by Walker (1984), and that feldspar and quartz crystals were concentrated during settling, as commonly occurs in modern volcaniclastic settings (Cas & Wright, 1988).

Table 3 (continued)

Lithofacies	Description	Process and Depositional Environment
Massive Sandstone (S2)	Massive fine to coarse-grained feldspathic arenite containing a minor amount of ooids and lithic fragments. There is little matrix but, in some instances, there is minor calcite cement. There are higher amounts of quartz in western outcrops of the unit compared to the east. Also, there is an increased abundance of quartz in the lower part of the sequence.	The abundance of fresh, angular feldspar and quartz crystals in massive sandstone is a result of concentration of ash-fall derived crystals reworked by settling through the water column and subsequent wave and tidal current action on the shelf. A *shallow marine shelf* interpretation of the depositional environment is supported by the presence of ooids in the rocks and minor calcareous cement. The massive nature of the facies is a result of the winnowing out of fine-grained material.
COARSE-GRAINED FACIES		
Intraformational Conglomerate (C1)	Large (2–13cm), angular, siltstone clasts with irregular boundaries containing mudstone rip-up fragments are characteristic of the facies, which contains dominantly intraformational clasts such as oolitic limestone, coral fragments, siltstone, and mudstone. Angular clasts within the facies are commonly imbricated with long axes parallel to flow, are evenly distributed throughout the bed, and are supported by a medium-very coarse-grained matrix of ooids, volcanolithic fragments, feldspar, and skeletal fragments (Figure 88). Volcanolithic clasts occur in some areas but are subordinate in abundance to the intraformational clasts. The facies generally crops out poorly and varies in thickness from 30m to 1m. The facies crops out discontinuously throughout the project area near the base of the succession, but commonly extends for up to 200m.	The angularity and imbrication of siltstone clasts and matrix supported texture of the facies indicate that deposition did not involve tractional movement. The homogeneous distribution of the clasts through the beds is characteristic of *gravity flow deposits* as described by Mutti (1992), Lowe (1982) and Middleton & Hampton (1976). Intraclasts have multiple sources, but the dominant sources were muds and ooid banks, which are inferred to have been forming on the shelf simultaneously. The most likely processes involved in deposition of the facies are debris flow or density currents, because these processes produce high flow intensities capable of imbricating clasts with a parallel-to-flow orientation. The deposits are similar to F1-F2 facies of turbidite systems as described by Mutti (1992). Similar gravity flow processes produce matrix supported textures (Middleton & Hampton, 1976; Lowe, 1982; Mutti, 1992; Shanmugam & others, 1995).
Polymictic Conglomerate (C2)	Thick, massive to lensoid polymictic conglomerate beds dominate the facies and contain subround to round granular to cobble-sized clasts of oolitic or reefal limestone, mafic volcanic fragments, rhyolitic volcanics, and siltstone set in a matrix of medium to very coarse, ooid-rich lithic arenite composed of volcanic fragments, ooids, and skeletal debris (Figure 89). Many volcanic clasts are oxidised (e.g. samples from field locality GSYB 351 at GR193622 7433133 on the Ridgelands 1:100 000 Sheet area). The sediments are poorly sorted, clast supported, and there is an overall fining upward trend. Low angle, planar, large-scale cross-stratification is common in the facies, particularly where conglomerates grade into very coarse sandstones (Figure 90). In the lower part of the Rockhampton Group the facies extends laterally for 3 to 4km. The facies has a maximum thickness of 90m and generally thins laterally over 1–2km. In areas where the conglomerates form lensoid bodies, lenses extend along strike for between ten and thirty metres. In the Cania area, pebbles from conglomerates in the upper part of the unit are dominantly purple felsic volcanics that are similar in appearance to rhyolitic volcanics in the same part of the succession. In the Cannindah Creek area, some small lenses consist predominantly of subangular to angular cobble-sized reefal boundstone fragments. In other areas, the conglomerates contain a wider variety of volcanic clasts including green to blue-green porphyry thought to be of intermediate to felsic composition. In the upper reaches of Monal Creek, the clasts are dominantly dacitic volcanics (Figure 91).	The rounded shape of oxidised volcanic clasts in the conglomerate indicates that these clasts were transported from a terrestrial source. The erosive nature of the basal contacts of the beds and their lensoid shape is consistent with the beds being channel bodies. The presence of volcanic fragments indicates a volcanic succession or hinterland was the source of some of the clasts and the freshness of these clasts indicates that they were deposited in close proximity to recently deposited volcanic and pyroclastic rocks. The presence of oolitic limestone clasts reflects erosion of older limestones and isolated ooids in the matrix indicate that some of the material was sourced from contemporary ooid banks. The deposit, therefore, formed by the erosion of a succession of rocks from various sources and this erosion involved incision of exposed shelfal deposits. The clast-supported, poorly sorted texture of the facies is consistent with deposition by tractional processes (Walker, 1984). This is confirmed by the channel-like nature of the beds and the low angle, planar cross-stratification interpreted to have resulted from the migration of channel bedforms. Therefore, this facies suggests subaerial erosion and incision of hinterland source rocks combined with erosion of additional material from unconsolidated shelfal deposits. Similar deposits have been described by Flint & others (1991) as fan-deltas and reworked fan deltas in the Andean forearc at Gatico, Chile. The deposits are also similar to facies described by G.A. Smith (1991) for volcaniclastic environments. The incision of the shelf and volcanic hinterland, the tractional nature of the processes involved, and the style of deposit formed are consistent with changes in marine shelf environments during periods of falling relative sea level predicted by Vail & others (1977) and Posamentier & Allen (1993). The coarse reefal debris may represent more proximal collapse of reef-edged platforms without incision.

Table 3 (continued)

Lithofacies	Description	Process and Depositional Environment
CARBONATE FACIES		
Clean Oolitic-pisolitic Grainstone (L1)	The facies consists predominantly of oolitic grainstone; pisolitic grainstone is less common, and pisoids represent giant ooids. Skeletal grains (mostly crinoid and bryozoan) and sand- to granule-sized feldspatholithic sediments may be relatively abundant, especially near unit boundaries, and peloids are typically rare. Ooid nuclei mostly represent skeletal debris or peloids, although volcanolithic, quartz, and feldspar nuclei occur in places. Larger fossils (corals, bryozoans, brachiopods) are generally abraded and uncommon. Bedding is generally fairly massive, and rare planar cross stratification has been observed. The facies commonly grades laterally or vertically into calcareous sandstones and grainstones of *Facies L3* (described below). However, the bases of some limestones are abrupt and progradational (e.g., Gudman Oolite, Stanwell area). Patch reefs (*Facies L5 and L6*) (described below) may occur in the more pure oolitic grainstone facies (e.g. Lion Creek Limestone and Gudman Oolite).	The dominance of skeletal ooid nuclei and typical sparsity of siliciclastic sediments suggest that the facies was generally deposited in an open water, shallow marine carbonate bank setting. Relatively isolated banks are suggested by the apparent lack of lateral continuity between major limestone units. However, areas where interfingering siliciclastic units occur may represent the upslope margin of a bank where it grades into the siliciclastic-dominated shallow shelf. Ooid formation is generally considered to occur in shallow, high-energy settings, and pisoids (approaching 20mm diameter in the Gudman Oolite) clearly represent shallow, tidally affected environments. Associated patch reefs also suggest high energy environments above fair weather wave base (Webb, 1989, 1998, 2005).
Sandy Oolitic Grainstone (L2)	The facies consists predominantly of dark gray to black oolitic grainstone that contains a variable amount of floating, silt- to pebble-sized siliciclastic sediment. Ooid nuclei may represent mostly skeletal debris or mostly siliciclastic sand. Skeletal grains are generally uncommon, but larger, abraded fossils (corals, crinoids) may occur. Large coral colonies (*Aphrophyllum*) occur in growth position in some cases (e.g., Cargoogie Oolite Member, Malchi Formation, Stanwell area; Webb, 1990). Bedding may be relatively massive, but is generally less massive than in *Facies L1*, and thin (10-20cm) beds may occur intercalated between black silts and shales that show graded bedding. Massive, interbedded calclithite beds (*Facies L7*) containing rounded oolitic grainstone cobbles also occur in some limestones. Contorted bedding, low angle cross bedding, and assymetrical current ripples occur less commonly. Cobble-sized, floating lithoclasts occur in some horizons. Examples of the facies are the Cargoogie Oolite Member (Stanwell area) and limestone in the Mount Salmon area referred to as FC3 by Webb (1990).	Graded and contorted bedding and interbedded calclithites suggest that the facies represents transported sediments derived largely from the shallow carbonate banks that produced *Facies L1*. The thicker, more massive units may be more proximal to the banks (e.g. limestone FC3, Mount Salmon area; Webb, 1990), whereas the thin limestones interbedded with black silts and shales appear to have been more distal, having been deposited farther down the slope by episodic sedimentation events (e.g., Cargoogie Oolite Member). Water depths varied depending on proximity to the shallow banks, but the most distal, episodic units were presumably deposited below storm wave base.
Cross-stratified Oolitic Grainstone and Calcareous Sandstone (L3)	This variable facies consists of strongly cross-stratified oolitic grainstone grading into calcareous volcanolithic arenite. Skeletal content is typically minor, except for scattered lag deposits of larger abraded fossils (corals, brachiopods, bryozoans). Planar, trough, and herringbone cross-stratification occur (Figure 92), with herringbone sets reaching 30-40cm in thickness. Bedding is flat to wavy. Ooid nuclei typically consist of rounded volcanolithic grains consisting of plagioclase phenocrysts in glassy groundmass. Quartz and feldspar crystals are angular, well preserved, and show no undulose extinction. Quartz crystals are typically embayed, beta types. Palaeocurrent data in the Mount Salmon area show a bipolar north-east–south-west trend, which was dominated by current directions from the north-east. The facies is commonly interbedded with silt- to sand-sized siliciclastics, some of which contain rare plant fossils and shallow channel conglomerates. An example of the facies is limestone in the Mount Salmon area referred to as FB2 by Webb (1990).	Trough, planar, and herringbone cross-stratification suggest deposition in shallow, wave, storm and tidally affected environments. The large percentage of volcaniclastic material, including rounded grains, and channelised conglomerates suggests a relatively near-shore shallow shelf position. However, pristine, euhedral feldspar and beta-type quartz crystals suggest that a portion of the volcaniclastic material was contributed as ash falls during intervals of proximal volcanic activity. Where these latter types of grains dominate the siliciclastic portion of the sediment, the depositional environment may have been farther offshore, near the edge of a shallow carbonate bank.

Table 3 (continued)

Lithofacies	Description	Process and Depositional Environment
Bioclastic Grainstone (L4)	The facies consists of coarse or fine, commonly well sorted bioclastic grainstone with abraded crinoid ossicles being the dominant allochem. Ooids may be absent, or may occur in increasing percentages as the facies grades into oolitic grainstone (*Facies L1*). However, for individual exposures, the ooid content is typically stable laterally. The ooids generally have skeletal nuclei. Bryozoans and calcareous algae may be abundant. The facies is generally fairly pure, but may contain volcaniclastic sand, especially where it grades into underlying or overlying units. Planar cross-stratification and current ripple lamination has been observed. Patch reefs (*Facies L5*) may occur in the facies (Webb, 1999). Examples of the facies include the upper limestone interval of the Gudman Oolite (Stanwell area), part of a limestone in the Mount Salmon area referred to as FC5 by Webb (1990), and the upper part of the Cannindah Limestone at Old Cannindah.	The well sorted, commonly abraded nature of the sediments and rarity of articulated crinoid stems suggests deposition in relatively high energy settings, above fair weather wave base, in an open shelf position. The general scarcity of siliciclastic sand suggests that the limestones may represent isolated banks similar to those upon which *Facies L1* formed, but the factors that controlled the occurrence of ooids versus crinoidal sands on such banks remain unknown. Occurrence of patch reefs in the facies supports the shallow bathymetric interpretation.
Reefal Microbialite Boundstone (L5)	The facies consists of poorly bedded, massive to irregular, light-coloured, fine-grained carbonate containing a variety of skeletal material, both in growth position and as transported debris. The fine-grained carbonate is characterised by microbialite textures. Stromatolitic (laminated) and thrombolitic (clotty) textures occur in the Tournaisian (e.g., Gudman Oolite reefs, Webb, 1998, 2005), but give way to thrombolite containing abundant lithistid sponges by the Visean (e.g., Lion Creek Limestone reefs, Webb, 1989; Old Cannindah reefs, Shen & Webb, 2005). Large colonial corals (fasciculate lithostrotionids and syringoporoid tabulates) occur scattered through the facies, and a highly diverse fauna and flora of calcareous algae, crinoids, sponges, solitary corals, and bryozoans occur in growth position. The facies occurs as isolated patch reefs within oolitic (*Facies L1*) or bioclastic (*Facies L4*) grainstone facies and is commonly associated with a halo of intraclastic debris derived from the reef. Patches of reefal coral biolithite (*Facies L6*) may, or may not occur within the facies. Examples include the lower part of the Gudman Oolite (Stanwell), Lion Creek Limestone (Stanwell), limestone FC5 (Mount Salmon; Webb, 1999), and the Cannindah Limestone (Old Cannindah).	Internal framework relationships and common halos of abraded intraclastic reefal debris suggest that the patch reefs were rigid structures in shallow, wave-affected environments consistent with the interpretations of facies *L1* and *L4*, with which they occur.
Reefal Coral Boundstone (L6)	The facies consists predominantly of colonial rugose and tabulate corals in growth position in a matrix of microbialite and coarse bioclastic rubble. The facies is massive, unbedded, light in colour, and generally somewhat less resistant than the oolitic grainstones (*Facies L1*) with which it commonly occurs. The facies invariably occurs within reefal microbialite biolithite (*Facies L5*), but the latter may occur without the present facies. The most abundant corals are lithostrotionids, *Aphrophyllum*, and syringoporoid tabulates (Webb, 1990). Coralla reach diameters over 1m. Calcareous algae and crinoids are also abundant in growth position, and sponges and bryozoans are less common. Calcimicrobes, such as *Renalcis*, occur in cavities. Pods of abraded reef talus occur within and around the facies. The facies so far appears to be restricted to Visean strata (e.g., limestone FC5, Webb, 1999; Lion Creek Limestone, Webb, 1989).	The facies was deposited with *Facies L5* in shallow, high-energy settings. Proximal, coarse talus beds associated with the facies attest to the high energy conditions (Webb, 1989). Maximum relief above the sea floor of the facies suggests that water depths were in some cases at least 2–3m.

Table 3 (continued)

Lithofacies	Description	Process and Depositional Environment
Intraformational Calclithite (L7)	The facies forms massive beds of pebble–cobble sized clasts of oolitic grainstone in ooid-rich matrix. The cobbles are subround and moderately to poorly sorted with a clean, well sorted, grainstone matrix. The rocks of the facies are clast-supported and massive. Bedding is typically thick (1–3m) and the facies may grade into *Facies L1* or *C2*. The facies rarely exceeds a lateral extent of 100–150m.	The size and rounding of the oolitic grainstone clasts indicates that, although they have been transported from the source, the distance travelled was not likely to have been great. The lack of siliciclastic material in both clasts and matrix, the lack of any association with finer-grained rocks, and the absence of any associated cross-stratification or sedimentary structures, indicates that the facies was not deposited by fluvial processes. The close association with ooid-rich limestones of *Facies L1* and *L2* is evidence of the proximity of these facies to each other and suggests that this facies formed in a *shallow, open water, marine shelf environment above fair weather wave base*. The size of the clasts, its relationship with *Facies L1*, and the clast-supported texture of the rocks suggest that the deposit formed during an episode of storm activity.
Carbonate mudstone (L8)	The carbonate mudstone facies consists of massive to medium bedded carbonate mud. Fossil are uncommon, but include crinoids. The facies is laterally associated with oolitic facies (*Facies L1*), but is contained within turbiditic interlaminated mudstones and siltstones (*Facies F2*). The facies is uncommon, having been observed only in the lower section of the Rockhampton Group in the eastern part of the outcrop area (e.g., parts of the Kolonga Creek Limestone at Cannindah Creek; McKellar, 1967).	The occurrence of *Facies L8* within muddy siliciclastic facies suggests deposition below fair weather wave base on a marine shelf, but lateral gradation into oolitic grainstones suggests that it was proximal to, but downslope from, shallower facies.

Figure 83. Graded bedding in interbedded siltstones and mudstones of Facies F1 at field point GSYB 55, GR194000 7427400. The facies commonly occurs in the Muddy Shelf Association, Rockhampton Group.

Figure 84. Slump fold in siltstone of Facies F2 at field point GSYB 63, GR193440 7247300. The facies commonly occurs in the Gravity Flow Association, Rockhampton Group.

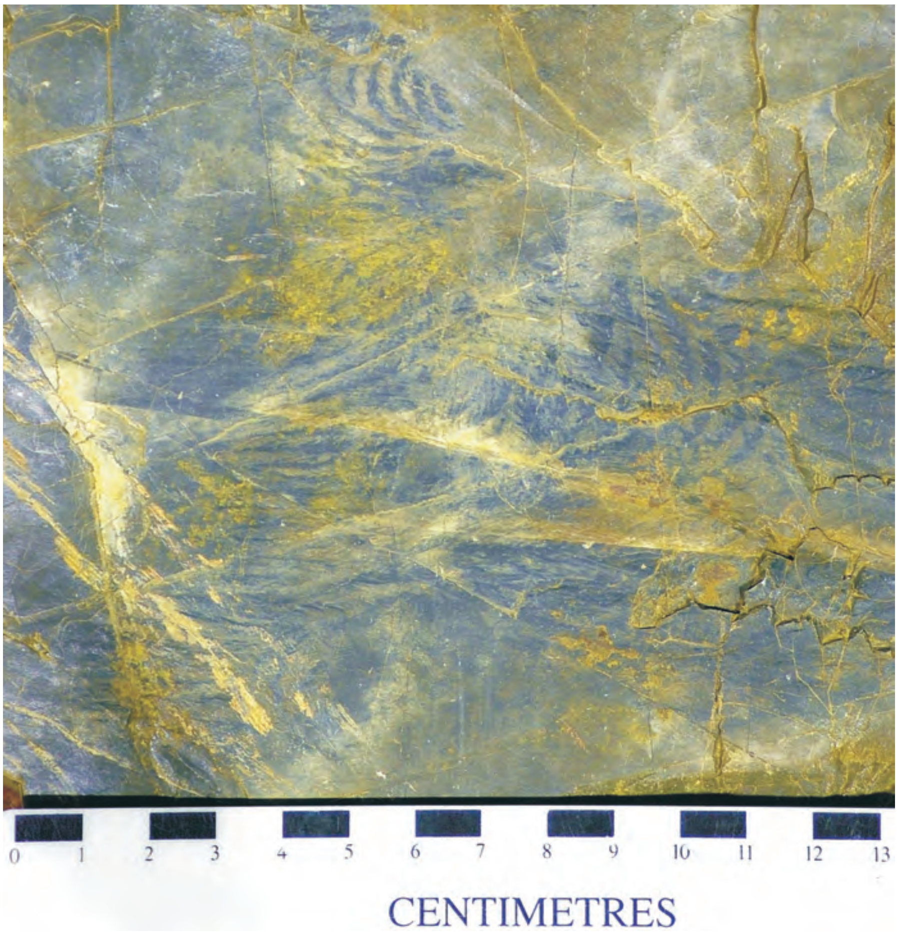


Figure 85. The trace fossil *Lophoctenium* from a small quarry at Belmont, ~10km north-north-west of Rockhampton. The rocks are interpreted as turbidites of Facies F2 deposited in deeper water towards the eastern limit of the Rockhampton Group.

Figure 86. Feeding trails on bedding surface of siltstone in Facies F3, Rockhampton Group

Figure 87. Sandstone from Facies S1 exhibiting planar lamination and symmetrical ripple laminations at field point GSYB 12, GR194000 7429350. The sedimentary structures indicate a shallow marine (above wave base) environment of deposition, Rockhampton Group.

Figure 88. Conglomerate from Facies C1 at field point GSYB 3, GR193550 7428000. The imbrication of evenly distributed matrix supported clasts indicates that gravity processes were dominant during deposition, Rockhampton Group.

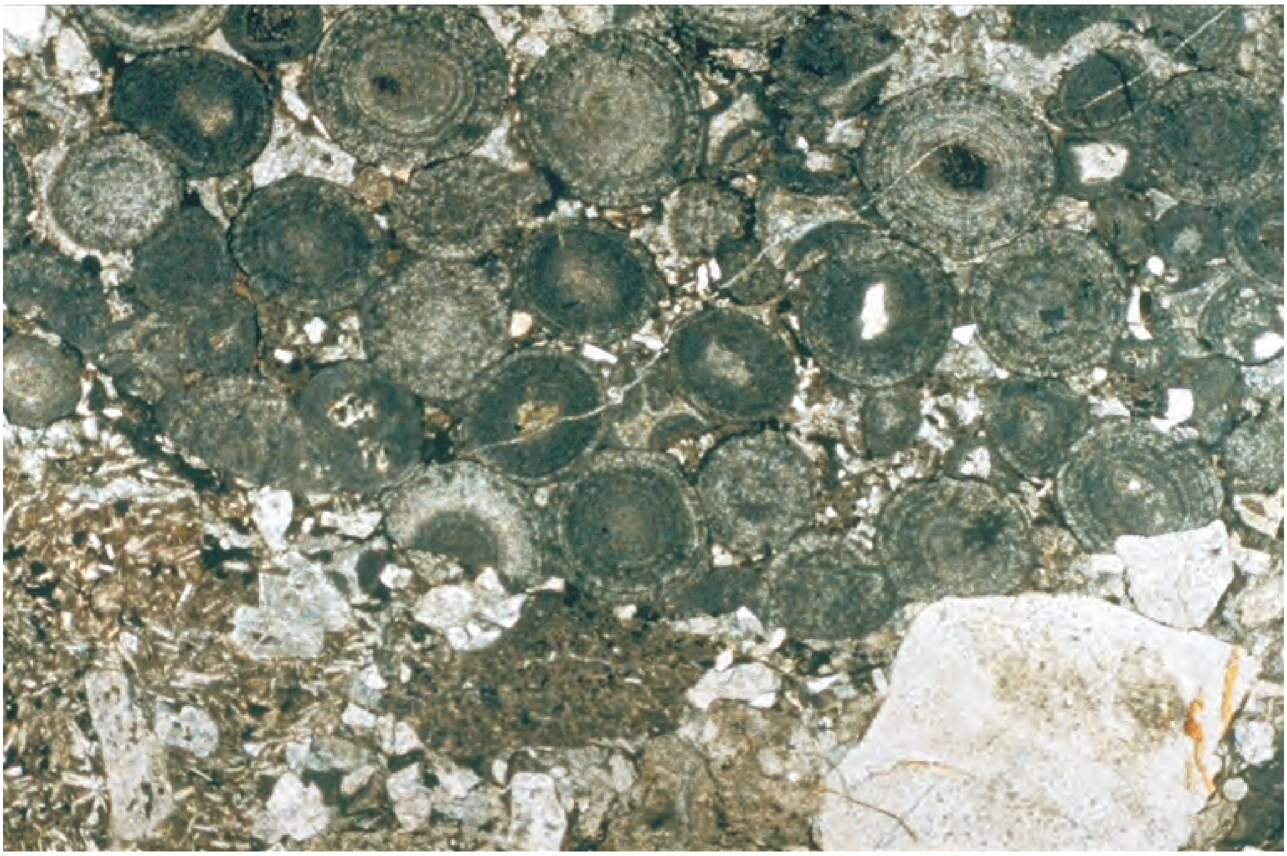


Figure 89. Photomicrograph of conglomerate from Facies C2 showing clasts of oolitic limestone and intermediate to felsic volcanics, Rockhampton Group. Plane polarised light.

Figure 90. Lensoid polymictic conglomerate from Facies C2 exhibiting low angle cross stratification, part of the Coastal Gravelly Channel Association, Rockhampton Group. The upper contact is with recessive-weathering sandstones and siltstones of Facies S1, and is interpreted as a sequence boundary. Field point GSYB 2, GR194000 7249400. Board marking scale is 40cm long.

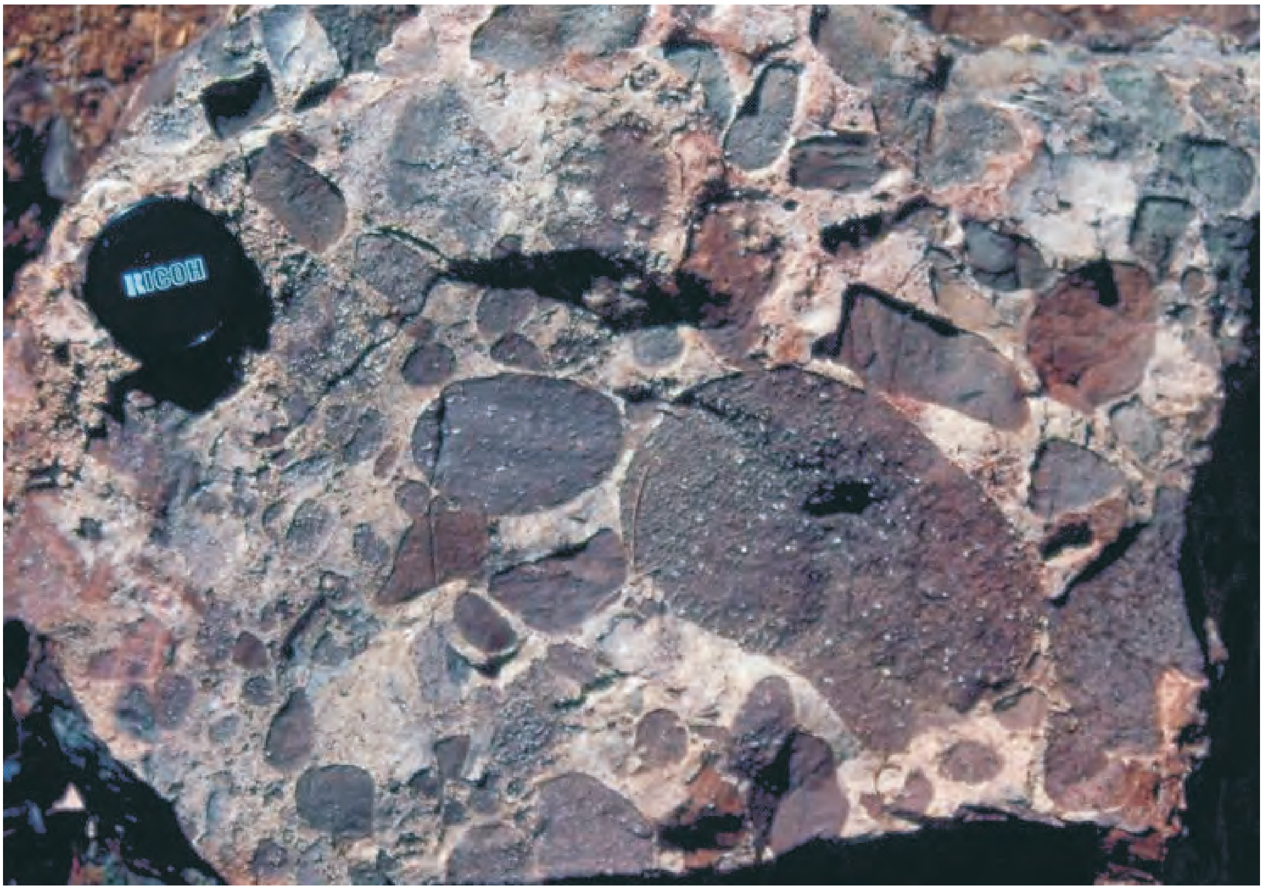


Figure 91. Poorly sorted, matrix supported cobble conglomerate, probably representing Facies C2, with rounded to subangular clasts of dacitic volcanics in a lithic-rich, sandy matrix with little or no quartz, Rockhampton Group. Field point GSMO 1266 in the upper reaches of Monal Creek.

Figure 92. Herringbone cross-bedding in oolitic limestone of Facies L3 of the Ooid Bank Association, Rockhampton Group. Cross-bedding indicates a shallow water, tidally influenced environment of deposition. Field point GSYB 5, GR195450 7431600.

Interpretation

The dominant process involved in this association is gravity flow resulting in matrix supported conglomerate (*Facies C1)* and finer-grained deposits of *Facies F2* and *S1*, which exhibit partial Bouma cycles. Oolitic limestones (*Facies L2*) are also associated with *Facies C1* and reflect transport of ooid bank material into palaeotopographic lows. These lows could be shelfal depocentres or may be associated with deeper basinal deposits of the Wandilla Formation to the east as described by Murray & others (1987). More channelised deposits in the association probably represent sedimentary bypasses. Similar deposits described by Mutti (1992) and Shanmugam & others (1995), occur in the Pyrenees and North Sea, respectively. These deposits are interpreted as forming part of a turbidite fan in a shelf margin to deeper basinal environment.

Coastal Gravelly Channel Deposits — Association 2

Description

Association 2 comprises polymictic conglomerate and sandstone of *Facies C2*. Massive to lensoid polymictic conglomerates are overlain by, and grade into, cross-stratified coarse to very coarse sandstone. The sandstone and conglomerate in *Facies C2* contain oolitic grainstone, reefal boundstone, rhyolite, andesite, and lithic siltstone clasts in a sandy matrix of ooids, skeletal debris, and volcanolithic material. No granitic clasts were observed in the rocks of this facies. The association has a fining upward trend and grades into feldspatholithic sandstone and siltstone (*Facies S1* and *S2*) of the *Sandy Shelf Association* and limestone (*Facies L1*) of the *Carbonate Bank Association*.

Interpretation

As discussed in Table 3, massive to lensoid conglomeratic facies (*C2*) were deposited in channels and followed incision of oolitic hardgrounds which provided a source for the rounded oolitic limestone pebbles in the conglomerate. Contemporaneous ooids with skeletal and volcanolithic nuclei were also included in these facies and are found as individual grains within the matrix indicating rapid incision of unconsolidated ooid banks. Volcanolithic pebbles in the conglomerates are commonly large, fresh and contain embayed volcanic quartz grains and were sourced from a nearby volcanic terrane. The lack of granitic clasts in the rocks suggests that the volcanic source was immature and was not unroofed by erosion. Many of the clasts are oxidised, substantiating the terrestrial provenance. Given the nature of the surrounding marine shelfal deposits, these coarse tractional sediments may have been deposited in submarine channels that reworked terrestrial and shelfal material. Similar Quaternary sediments have been described as an alluvial fan-delta system in the Andean forearc by Flint & others (1991). These Andean deposits include conglomerates with well-rounded clasts in a matrix of shell debris interbedded with planar and trough cross-stratified sandstones similar to those described in *Facies C2* and were interpreted as reworked fan delta deposits. Miller (1989) also described conglomeratic deposits with planar cross-stratification in the Klamath terrane, California, and suggested they formed in a braid delta environment. However, the lack of associated fine-grained facies and fining upward parasequences suggests that braid deltas were not the final environment of deposition for the sediments of this association. The composition of the rocks in Association 2 and the abundance of material derived from unconsolidated shelfal sediments indicate that the formation of these deposits involved shelfal erosion and reworking. They may have formed by the reworking of fan delta or braid delta deposits similar to those described by Flint & others (1991) or Miller (1989). Both of those examples formed in volcanic environments and were associated with periods of tectonic instability.

Sandy Shelf Deposits — Association 3

Description

Association 3 contains interbedded sandstone and siltstone of *Facies S1* and thick, massive sandstone of *Facies S2*. It also contains minor lenses of very coarse sandstone to conglomerate (*Facies C2*). Symmetrical to asymmetrical wave and current ripple lamination and graded bedding are the most characteristic sedimentary structures, but soft sediment deformation structures and convolute bedding occur where the association overlies muddier units of the muddy shelf association. The sediments contain a large proportion of feldspar and volcanic quartz crystals and are invariably clast supported. A cyclic fining upward trend in some of the deposits accompanies a change from graded bedding to current-ripple lamination and flat lamination.

Interpretation

Sedimentary structures in the association indicate that tractional processes were operating during deposition of the facies. Cyclic changes in the structures associated with changes in grainsize reflect changing energy levels in the sedimentary environment. The deposits are typical of tempestite deposits formed on shallow marine shelves (Walker, 1984) and are similar to classical turbidite facies (Mutti, 1992). Small scale density currents may be involved in their formation (Lowe, 1982; Greenwood & Davis, 1984; Walker, 1984; Boggs, 1987). The presence of wave ripples in *Sandy Shelf Deposits* supports a tempestite interpretation. The abundance of feldspar and quartz crystals is the result of concentration of ash fall material during settling through the water column. Allochthonous ooids in *Facies S1* and *S2* were transported by tractional currents and deposited in ripples as shown in Figure 85. The influence of wave, tide, and storm currents and the deposition of ash fall material are compatible with sedimentation on a sandy shelf.

Muddy Shelf Deposits — Association 4

Description

Siltstone and mudstone of *Facies F1* and *F2* dominate Association 4 with interbeds of graded very fine to fine feldspatholithic sandstone. Bedding is thin and rhythmic with rare trace fossils. Trace fossils are inter-woven unbranched structures interpreted as feeding traces on the basis of the fine nature of sediment deposited within them. These facies are not associated with matrix supported conglomerates (*Facies C1*), but are commonly associated with oolitic grainstones of *Facies L2*. Current ripple lamination occurs but is not as common as in the *Gravity Flow Association.* The association displays an overall fining upward trend.

Interpretation

The sparse fossils in the association indicate that it was deposited in a marine shelf environment, at or below storm wave base. Graded beds (Figure 81) were formed by settling from suspension in a low to moderate energy environment as discussed in the analysis of *Facies F1* and *F2*. These characteristics are consistent with a distal shelf environment similar to that described by Walker (1984) and Reineck & Singh (1980).

Fossiliferous Muddy Shelf Deposits — Association 5

Description

Facies in Association 5 are similar to those in Association 4 and include siltstone, sandstone, and mudstone of *Facies F2*, *F3*, and *S1*. The major difference, however, is that the siltstone and mudstone facies are commonly fossiliferous. Ichnofossils found in the area were identified as *Cruziana* facies based on the identification of *Phycodes*, *Crossopodia*, and *Cruziana* trace fossils. Coarse-grained sandstones of *Facies S1* are common and contain macrofossils including trilobites, gastropods, and brachiopods. The abundance of fossils varies throughout the association. Association 5 strata rarely contain ooids, and then only in minor amounts; sandstones are typically feldspathic to feldspatholithic in composition. The facies association has a characteristic fining upward nature. All facies occur throughout the project area and vary in their lateral extent.

Interpretation

Association 5 represents a marine shelfal environment at or below storm wave base where the fossil fauna is common and abundant. The abundance of fossils, their good preservation, and *Cruziana* assemblage trace fossils support this environmental interpretation (Heckel, 1972; Ekdale & others, 1984; Miller & Knox, 1985). Deposition occurred mainly by tractional shelf processes and sediment was sourced from ooid banks, bottom level communities, and volcanic ash falls as evidenced by the composition of sandstones in the association. Each of these sources dominated during different stages of deposition producing gradational variations in composition. Ooid banks commonly overlie the association and probably formed a barrier to wave and current activity, resulting in a change from high to low energy processes in a landward direction.

Carbonate Bank Deposits — Association 6

Description

Oolitic/pisolitic (*Facies L1*) and, less commonly, bioclastic (*Facies L4*) grainstones are the dominant and characteristic facies of this association. The dominant grainstones may interfinger with, and grade laterally into, trough to planar cross-stratified oolitic grainstones and calcareous sandstones of *Facies L3* and interbedded siltstone, sandstone, and mudstone of *Facies S1* and *F2*. Conglomeratic strata (*Facies C2* and *L7*) are uncommon. The association may include only a single type of grainstone, or may include a succession from one type (oolitic or bioclastic) to the other, or *Facies L1* and *L4* may intergrade to some extent. The association may consist almost entirely of carbonate facies, or may contain relatively large intervals of sandstone or siltstone (*e.g.* Cannindah Limestone, Old Cannindah; Gudman Oolite, Stanwell). The facies succession within a given bank was highly variable even over relatively short distances, and vertical and lateral facies changes may be numerous and abrupt (e.g. Cannindah Limestone, Jull, 1968). Siliciclastic material within the carbonates is generally minimal, and ooid nuclei typically represent skeletal debris or peloids. However, tuffaceous intervals occur in some otherwise pure oolitic grainstone sequences (e.g. Cannindah Limestone at Old Cannindah) suggesting the direct introduction of volcanic ejecta. Large rip-up clasts and sand- and granule-sized siliciclastic sediments occur most commonly near the base of carbonate successions. Finer sediments are uncommon. Individual carbonate intervals locally coarsen upwards (e.g. lower part of the Gudman Oolite).

Small patch reefs (*Facies L5* and *L6*) may occur in the succession within *Facies L1* and *L4* (Webb, 1989,1998, 1999, 2005) with particularly large, amalgamated reef complexes in the Cannindah Limestone (Shen & Webb, 2005). Reef distribution is irregular although some horizons appear to have abundant reefs over relatively wide areas (e.g. Lion Creek Limestone, Stanwell; upper part of Cannindah Limestone, Old Cannindah). The abundant coral faunas from the Rockhampton Group (Webb, 1990, 2000) are largely confined to the patch reef facies.

Interpretation

Association 6 is interpreted as representing more or less isolated, shallow marine carbonate sandbanks occupying topographic highs on a more extensive siliciclastic dominated shelf. The dominance of large ooids, pisoids and abraded skeletal sands suggests shallow depths well within fair weather wave base. Planar, trough, and bidirectional cross-stratification also suggest very shallow, tidally affected deposition. The occurrence of halos of reef-derived intraclasts and reefal debris around patch reefs is a further indication of the high-energy environment. Relative changes in terrigenous content of the carbonate intervals probably reflect their proximity to the landward source area and the amount of ash fall, and gradation into *Facies F3* may reflect the up-dip junction of the bank with the broader siliciclastic shelf. Abundant rip-up clasts and coarsening upward sediments in some carbonate intervals (e.g. lower Gudman Oolite, Stanwell) suggest the progradation and shallowing of individual, mobile sand banks over the shelf. However, the lateral discontinuity of most banks at larger scales suggests that carbonates were generally isolated on local topographic highs within the broader Yarrol Shelf.

Geochemistry

The Rockhampton Group contains sparse volcanic rocks, mainly of felsic composition. Two rhyolitic ignimbrites from Youlambie Creek 12km north of Monto were analysed. The samples have a large subduction component, being comparatively low in high field strength elements (HFSE), particularly Nb. The $(Th\text{-}Nb)_N$ gap is large, with values of 31.5 and 40.8 (Figure 93). On the granitoid discrimination diagrams of Pearce & others (1984), the analyses plot in the volcanic arc granite field (Figure 58). Pearce (1996) has pointed out that Rb, Nb and Y concentrations in volcanic arc felsic rocks are linked strongly to source, which is interpreted to be a depleted mantle and a crustal component from the subduction process. Because the average composition of continental crust lies within the volcanic arc field, crustal assimilation does not have much effect. Thus the felsic volcanics within the Rockhampton Group are interpreted to have been related to an active subduction zone.

Andesites have been reported from the Rockhampton Group in the Mungungo and Cania areas near Monto (Dear, 1963, 1968; Webb, 1973, 1977; Cummings, 1998). A sample analysed from Cania may be a flow or a dyke. The analysis is similar to those of samples from volcanic arcs built on relatively thin crust, like the central part of the Southern Volcanic Zone of the Andes, where the crustal thickness beneath the Llaima

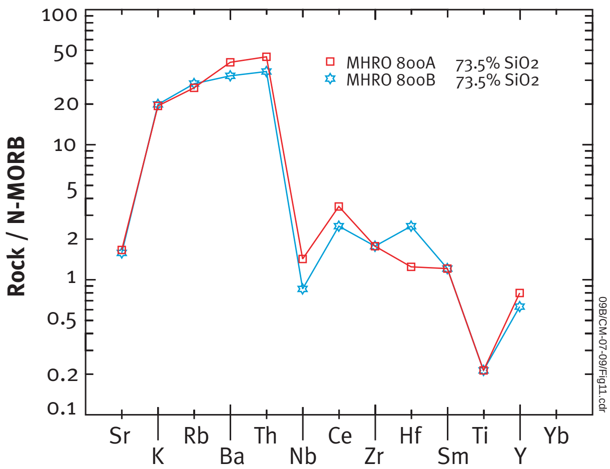


Figure 93. Spidergram plot of two rhyolite samples. N-MORB values are from Pearce (1983).

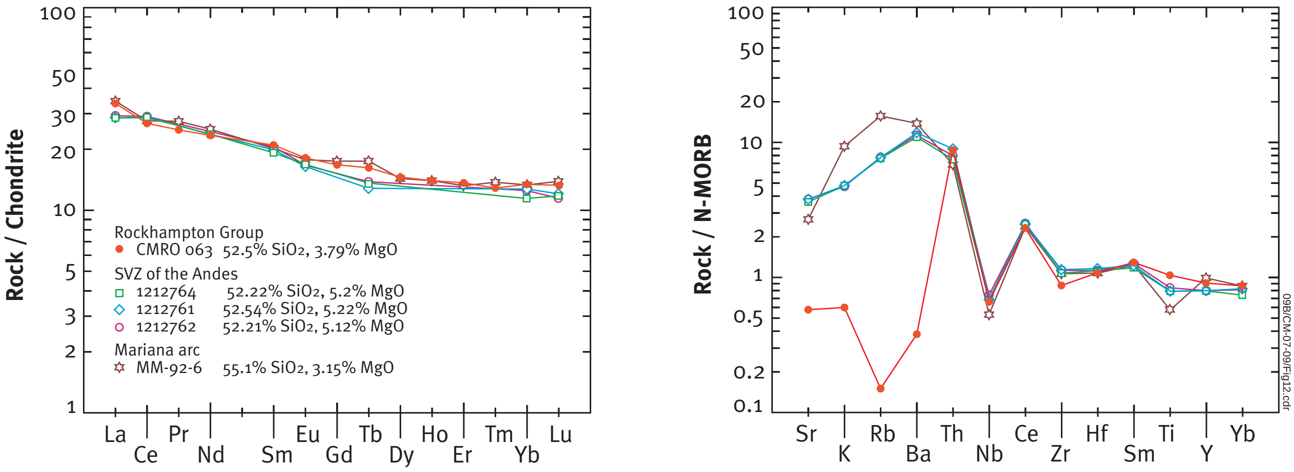


Figure 94. (a) REE and (b) spidergram plots comparing basaltic andesite with basaltic andesites from Llaima volcano in the southern part of the Southern Volcanic Zone of the Andes (Hickey & others, 1986; Tormey & others, 1991) and from the Mariana arc (Elliott & others, 1997). Chondritic REE values are from Sun & McDonough (1989), and N-MORB from Pearce (1983).

volcano is ~30km, or island arcs like the Marianas (Figure 94). The Ce/Y ratio of 0.85 suggests that crustal thickness was ~20km (Mantle & Collins, 2008), considerably less than that below geochemically equivalent rocks from the Andes. The $(La/Yb)_N$ value of 2.5 for the Cania sample is less than that typical of Triassic volcanics in the Yarrol area, which fall in the range 3.13–7.23 with an average of 5.06. Thus the sample is unlikely to be a later Triassic dyke. The analysis is very similar to andesite from the Yarrol type area of the mid-Permian Owl Gully volcanics, and could be a Permian dyke. However, a closer outcrop of Owl Gully Volcanics in the Spring Creek Syncline north of Monto has different geochemistry. It is tentatively concluded that the sample from Cania was coeval with deposition of the Rockhampton Group. If so, it reinforces the interpretation of the chemistry of the rhyolitic ignimbrites.

Stratigraphic Sequences

The Rockhampton Group is a unit bounded at the top and bottom by disconformities and their correlative conformities. Lithologically, the upper boundary is marked by a compositional change that can be identified both in the field and on images of airborne radiometric data. The lower boundary is marked by a change in the abundance of ooids in both the clastics and carbonates, a change from deeper shelf and slope facies (Mount Alma Formation) to shallow shelf facies (Rockhampton Group), and a subtle change in the composition of siltstones and mudstones either side of the boundary. Locally, the boundary forms an onlapping surface with sandstone, siltstone, and limestone beds terminating at this surface. In the Cannindah Creek area, the contact between the Mount Alma Formation and the Rockhampton Group marks the change from mudstone and wackestone to an oolitic grainstone. This change is interpreted to represent a distinct shallowing event near the base of the Carboniferous. In the Bindawalla area on the western side of the Calliope 1:100 000 Sheet area, the contact between these two units is marked by a change from siltstone to fossiliferous calcareous sandstone and limestone. The fossiliferous beds contain abundant gastropods and sparse to common brachiopods. This change in facies also indicates that a shallowing event occurred near the base of the Carboniferous and prior to deposition of shallow marine shelf faunas of the Rockhampton Group.

Simpson (1995) identified six unconformity-bounded sequences in the Rockhampton Group in an area 55km west of Rockhampton (Mount Salmon area) based on the recognition of key surfaces, aggradational, retrogradational, and progradational stacking patterns, and proximal to distal trends in lithofacies and facies associations. Similar facies architecture was observed in the Rockhampton Group throughout the Yarrol Project area although sequences could not be reliably correlated on a regional basis. However, sections in the western part of the Yarrol Project area (eg. Mount Salmon and Old Cannindah areas) were more slowly deposited than those to the east (e.g. Gracemere and Cannindah Creek areas). This, plus the overall fining of sediments to the east, supports previous interpretations (McKellar, 1967; Roberts & Engel, 1980; Murray & others, 1987), that more easterly facies represent deposition in more distal areas in the marine basin.

Environment of deposition

Kirkegaard & others (1970) described the depositional environment of the Group as shallow marine shelf but suggested that part of the thick succession of sandstones and shales within the Group may have been deposited in a relatively deep water marine environment.

Roberts & Engel (1980) discussed the Carboniferous palaeogeography of the New England Orogen and recognised the major depositional features of the Yarrol Basin including a western volcanic chain, marine shelf, and eastern slope and basin. They identified tectonism and global changes in sea level as the two major controls on sedimentation in the basin. Roberts & Engel proposed a eustatic lowering of sea level between the early and late Carboniferous which influenced deposition on the Yarrol Shelf. Roberts (1981, 1985a) proposed a transgressive-regressive cycle controlled by eustacy in the late Tournaisian to early Visean.

Webb (1990) compiled measured sections across the limestone-rich intervals in the Rockhampton Group in the Stanwell and Mount Salmon areas and interpreted the depositional environment of these rocks in terms of six biofacies. These facies included biohermal, biostromal, forebank slope, shallow shelf mobile sand, nearshore mobile sand, and offshore silt biofacies. He noted that the thickest limestone intervals generally represent autochthonous accumulations of ooids formed in shallow, high-energy conditions and that the remaining limestones represented allochthonous accumulations transported from shallower shoaling areas.

Webb (1990) also noted differences in the types of lithologies between the Mount Salmon area and the type area at Stanwell to the east, supporting a more proximal shelf environment for the Mount Salmon area. He found little evidence for the emergence of the Craigilee (Mount Salmon) area during the Tournaisian and suggested that the unconformity at the base of the Rockhampton Group might partly represent a regressive-transgressive cycle in late Tournaisian to early Visean times, similar to that proposed as a eustatic sea level event by Roberts (1981). Simpson (1995) developed a depositional model for the unit based on a sequence stratigraphic framework for the Mount Salmon area, utilising the conodont biostratigraphic framework proposed by B.G. Fordham (personal communication), Jenkins & others (1993), Mory & Crane (1982), and Druce (1969c). The following discussion summarises and partly revises previous work on the depositional environment of the Rockhampton Group.

The Rockhampton Group was deposited on a marine shelf to slope in a volcanically and tectonically active environment. Sediment was primarily sourced from intermediate to felsic volcanics and less commonly from basic volcanics. Some of this volcanic material was transported by tractional process and some, particularly that in the west, was delivered by explosive airborne events. Most volcanic-derived material was reworked by submarine channel, gravity flow, tidal, and wave processes. Primary volcanics include ignimbrites and andesite or basalt in the south-eastern part of the unit. The marine shelf environment in the early Carboniferous experienced fluctuations in siliciclastic influx and carbonate production possibly associated with tectonic events. These fluctuations are difficult to relate to eustatic sea level changes. Instability on the shelf is inferred from the presence of gravity flow deposits throughout the successions that sourced carbonate bank material from topographic highs on the marine shelf. The dominant palaeocurrent direction measured in the unit was directed toward the north-east and this direction is compatible with an increase in more distal fine-grained facies and a decrease in the abundance of unconsolidated volcanic material and carbonate facies. It is also compatible with previous interpretations by Fleming & others (1975) and Murray & others (1987) that ooids in the more easterly Wandilla Formation were derived from the Carbonate Bank Facies of the Rockhampton Group. Simpson (1995) identified a locally disconformable lower contact for the unit in the Mount Salmon area, marked by the onlap of overlying sedimentary rocks. The absence of such a hiatus throughout the Rockhampton Group suggests that the Mount Salmon area was a region of higher relief during the Tournaisian and was close to the western margin of the marine shelf at this time. To the east in the type area at Stanwell, Webb (2005) interpreted the lowermost Tournaisian Gudman Oolite at the base of the Rockhampton Group as a stranded lowstand deposit, possibly reflecting the same major regressive phase. The development of shallow marine facies in the lower section of the Rockhampton Group in contrast to deeper water facies of the underlying Mount Alma Formation indicates that a significant shallowing event occurred at the beginning of the Carboniferous. This shallowing event is coincident with a change in composition of the sediments and volcanics that formed the Late Devonian to early Carboniferous Yarrol Province succession.

Thickness

The thickness of the Rockhampton Group increases from west to east across the outcrop belt, consistent with deepening of the basin in this direction (see figure 8 of Kirkegaard & others, 1970). Complete sections (overlying the Mount Alma Formation and overlain by the Lorray Formation) are exposed only in the type section west of Rockhampton, near Mount Salmon to the north-west, and in the western limb of the Yarrol Syncline east of Monto.

In the type area, the unit was measured at 1980m thick (Fleming, 1960, 1967). This thickness has been increased by recent mapping that positions the top of the Rockhampton Group approximately 25m above the top as mapped by Fleming (1967). A limestone containing Visean corals was found at the top of this unit and the Rockhampton Group may, therefore, extend into the *Texanus-bilineatus* zone of the Visean Stage (late early Carboniferous). In the Mount Salmon area, the unit ranges from 1300 to 1700m in thickness (Kirkegaard & others, 1970; Simpson, 1995). Less complete sections along the western edge are thinner, as at Mount

Morgan (610m; Kirkegaard & others, 1970), Cania (550m; Dear, 1968), and Old Cannindah. Visean limestones at Old Cannindah, which are approximately 300m thick, appear to have been deposited more slowly than other successions in the Rockhampton Group. The Old Cannindah succession may represent a condensed section, although no obvious unconformities have been observed. Some of the succession may have been lost during periods of erosion or the limestone may have formed in an area of low sediment supply.

The Rockhampton Group is considerably thicker in the western limb of the Yarrol Syncline, comprising 2740m from the first appearance of oolitic limestone (McKellar 1966, 1967). Further east, it is difficult to ascertain the total thickness because of the restricted development of oolitic limestones. Jell (1961) measured a sequence 2775m thick at Mount Larcom with lepidodendroid plants at the base and ooids throughout the section, suggesting correlation with the Rockhampton Group. Just to the south of this locality, Kirkegaard & others (1970) estimated a thickness of ~3050m, and Donchak & Holmes (1991) gave a figure of 4000m. Willmott & others (1986) reported that the Rockhampton Group in the Glenmore area was much thicker than in the type area to the west.

It has not been demonstrated that any of the sequences along the eastern margin of the Rockhampton Group represents a complete section, and it is concluded that the unit thickens consistently from west to east. The decrease in grain size in this direction precludes an explanation as a clastic wedge derived from the east.

Structure

Outcrop is never continuous across the entire belt of the Rockhampton Group because of folding, faulting and subsequent erosion (Figure 82). Bedding is usually parallel or subparallel to the overall trend of the unit. Fold axes trend mainly north-north-west and are generally very gently plunging to horizontal.

Morand (1993b) concluded that the Rockhampton Group was deformed as part of a west-vergent fold and thrust belt in the Hunter–Bowen Orogeny (mid- to late Permian), but the thrusting was neither severe nor thin-skinned. Outcropping folds are symmetrical, and cleavage is very poorly developed, but is steep. South-east plunging folds in the area north of Mount Larcom (Morand, 1993b, figure 5, Darts Creek Anticline) and the Mount Salmon area may reflect proximity to major faults or indicate a later episode of folding.

The fold-thrust belt model is supported by structural studies by Holcombe & others (1997a) at the eastern edge of the Gogango Overfolded Zone. Here, moderately to steeply west-dipping rocks of the Rockhampton Group and the overlying Lorray Formation are interpreted as the western limb of a breached fault-propagation fold (Holcombe & others, 1997a, figure 4, section C–D). However, the Rockhampton Group does not exhibit the classic structures of a thin-skinned fold-thrust belt like the Gogango Overfolded Zone (Fergusson, 1991). Little of the sequence appears to be missing or repeated due to the relatively small displacements on faults, and dips are generally shallow to moderate (see cross section in figure 2 of Bryan & others, 2001). Moderate to steep dips are only regionally developed along the eastern margin, close to the Yarrol Fault. The Rockhampton Group appears to have been part of a relatively rigid block that moved westward as a whole, deforming the rocks of the Gogango Overfolded Zone (Fergusson, 1991). The Tiverton Fault in the Yarrol Syncline east of Monto is interpreted as a late, out-of-sequence thrust. Subsequently, the Rockhampton Group has been affected by brittle deformation resulting in reverse and normal faults.

No clear evidence of growth faulting during the Carboniferous has been observed.

Biostratigraphy and age

Fossils are abundant in the western, shallow marine facies characterised by repeated oolitic limestone development (Maxwell, 1952, 1954; Fleming, 1960; Dear, 1963, 1968; Jull, 1965; Kirkegaard & others, 1970; Dear & others, 1971; Radke, 1971; Webb, 1973, 1977; Webb, 1988, 1990; Cummings, 1998; Bryan & others, 2001), and are still prolific along the western limb of the Yarrol Syncline (Maxwell, 1964; McKellar 1966, 1967; Batchelor, 1970). However, there is a marked decrease in the abundance and diversity of faunas across the Yarrol Syncline, and no faunas older than middle Visean have been found on the eastern flank (McKellar, 1966, 1967). At other localities along the eastern margin of the Rockhampton Group, macrofaunas are either completely absent (Willmott & others, 1986), or are restricted to the late Visean (Jell, 1961). This clearly indicates that, for most of the time, the eastern part of the Rockhampton Group was deposited in water depths too great for faunal development.

McKellar (1966, 1967) identified a sequence of Carboniferous faunas which has been incorporated into the standard biostratigraphic scheme for eastern Australia (Campbell & McKellar, 1969; Roberts, 1975; Roberts & others, 1993). Druce (1969c, 1974) analysed conodonts of the Tournaisian stage in the Rockhampton Group.

More recently, work on coral faunas by Webb (1990, 1994, 2000) and conodont faunas (Mory & Crane, 1982; Jenkins & others, 1993) has provided a more tightly constrained age framework for the unit. Webb (1989, 1990, 2005) analysed carbonate facies in the Gracemere area as part of his coral biostratigraphic work in the Rockhampton Group. Simpson (1995) constructed a new sequence stratigraphic framework for the unit in the Mount Salmon area using conodont age determinations by B.G. Fordham (Appendix 2), Mory & Crane (1982), and Jenkins & others (1993).

Zircon U-Pb ages from volcanic rocks are consistent with an early Carboniferous age. Bryan & others (2004) obtained ages from two ignimbritic units within the Rockhampton Group, one of which was mistakenly attributed to the Mount Alma Formation. Ages from these ignimbrites are 361.1±5.2Ma (Dan Dan Creek) and 350.9±2.7Ma (upper Monal Creek). An ignimbrite collected from Youlambie Creek north of Monto gave 356.8±3.8Ma (Appendix 1). These results show that felsic volcanism was active very early in Rockhampton Group time, because the currently accepted age of the Devonian–Carboniferous boundary is 359.2±2.5Ma (Gradstein & others, 2004).

Stratigraphic relationships

At the northern end of the outcrop belt, the westernmost outcrops overlie older rocks; elsewhere, the western limit of the Rockhampton Group is overlain by or faulted against younger rocks.

Regionally, the Rockhampton Group conformably overlies Upper Devonian to lowermost Carboniferous strata, comprising the very extensive Mount Alma Formation and the more restricted Three Moon Conglomerate, Balaclava Formation, and Lochenbar Formation, although many contacts are faulted. Local disconformities are present in the Mount Salmon or Craigilee area, west of Rockhampton, and in the Old Cannindah area, east of Monto. The Rockhampton Group is faulted against older Devonian units including the Marble Waterhole beds, Calliope beds, Craigilee beds, and Channer Creek beds. The Upper Carboniferous to lowermost Permian Lorray Formation conformably overlies the Rockhampton Group in most areas, but the contact is disconformable in the Craigilee and Stanwell areas on the Rockhampton 1:250 000 Sheet, and locally is faulted. The Lower Permian Youlambie Conglomerate overlies the Rockhampton Group in many areas, the contact appearing to be either disconformable or unconformable. The Rockhampton Group is unconformably overlain by several younger stratigraphic units, and is intruded by granite plutons of Permian to Cretaceous age.

In the north-west of the Yarrol Project area, previous mapping by Malone & others (1969) on the Duaringa 1:250 000 geological map sheet did not differentiate the Rockhampton Group from the underlying undifferentiated unit identified by Kirkegaard & others (1970) on the Rockhampton 1:250 000 sheet. Mapping by the Yarrol Project team has delineated the distribution of these two units on a lithological basis.

The area of Rockhampton Group on the western limb of the Bindawalla Anticline has been altered in extent based on field mapping and the identification of fossil and ooid-bearing sandstone in areas formerly mapped as Crana beds (now Mount Alma Formation). This mapping is also based on radiometric and magnetic images.

In two regions, one near Glenmore homestead in the northern suburbs of Rockhampton and the other near Yarwun township north of Gladstone, the Rockhampton Group has not been differentiated from a second unit. In the Glenmore area, this was due to the poor outcrop and structural complexity of the area. In this area the Rockhampton Group has been folded and faulted with the Mount Alma Formation. This structural complexity is further exacerbated by the lack of distinctive limestone and sandstone outcrops that usually clearly identify the Rockhampton Group. The area was previously mapped as Rockhampton Group but outcrop of fissile micaceous mudstone and muddy limestone characteristic of the Mount Alma Formation was found in the area. Furthermore, one of the muddy limestones returned Late Devonian conodonts, which is more consistent with the bed forming part of the Mount Alma Formation.

In the Yarwun area, a thin carapace of Berserker Group covers the Rockhampton Group. As in the Glenmore area, the lack of distinctive outcrop has made differentiation of these two units difficult and they have been grouped together in a single lithological unit. This area was formerly mapped as Berserker beds apart from a small fault sliver containing lycopod fossils.

Correlation with other units

The Rockhampton Group is correlated with sequences west of the Peel Fault in the Tamworth Belt of New South Wales (Roberts & Engel, 1980). McKellar (1967) was the first to correlate supposed early Palaeozoic rocks east of the Yarrol Fault with the Rockhampton Group. This correlation was supported by Fleming & others (1975) and has been confirmed by radiolarian age determinations from similar sequences further south and by dating of detrital zircons (Korsch & others, 2009a). The Wandilla Formation of the Curtis Island Group is now regarded as a deeper-water equivalent of the Rockhampton Group.

Dear (1994) recognised cycles of volcanism in the Connors Arch, west of Marlborough, and correlated part of this volcanic succession with the Rockhampton Group. Recent work by Hutton & others (1999) suggests that this correlation does not fit age determinations from either succession.

Economic significance

Krosch & Kay (1977) and Krosch (1981) mapped limestone outcrops within the unit in the Rockhampton and Gladstone areas as part of an industrial resource potential assessment of limestone-bearing units. In general, the limestones are of lesser purity than those of Devonian age.

Curtis Island Group
(P.R. Blake)

Within the project area, the Wandilla Province is comprised of the Curtis Island Group, and occurs to the east of the Yarrol Province (Figure 95) being separated from it by a major fault marked by serpentinite lenses (Yarrol Fault). The Curtis Island Group is dominated by sedimentary rocks which are divided into three units, the Doonside, Wandilla and Shoalwater Formations. The sequence is structurally complex, with moderate to steep dips, tight folding and development of slaty and crenulation cleavage. Metamorphic grade is up to amphibolite facies. It should be recognised that the Doonside, Wandilla and Shoalwater Formations are not valid stratigraphic units. The accretionary wedge is a tectonostratigraphic assemblage in which stratigraphic contacts are only preserved on a local scale. However, the nomenclature has been a useful means of describing the broad lithological changes across the wedge (Murray, 1994).

Kirkegaard & others (1970) concluded that the Curtis Island Group was the oldest unit in the Rockhampton–Gladstone Yarrol/Coastal area, and assigned it an ?Ordovician–Silurian age. The Curtis Island Group is now interpreted as an accretionary wedge assemblage (see, for example, Fergusson & others, 1988, 1990a,b, 1993), and is often referred to as the Wandilla terrane or the Wandilla–Shoalwater accretionary

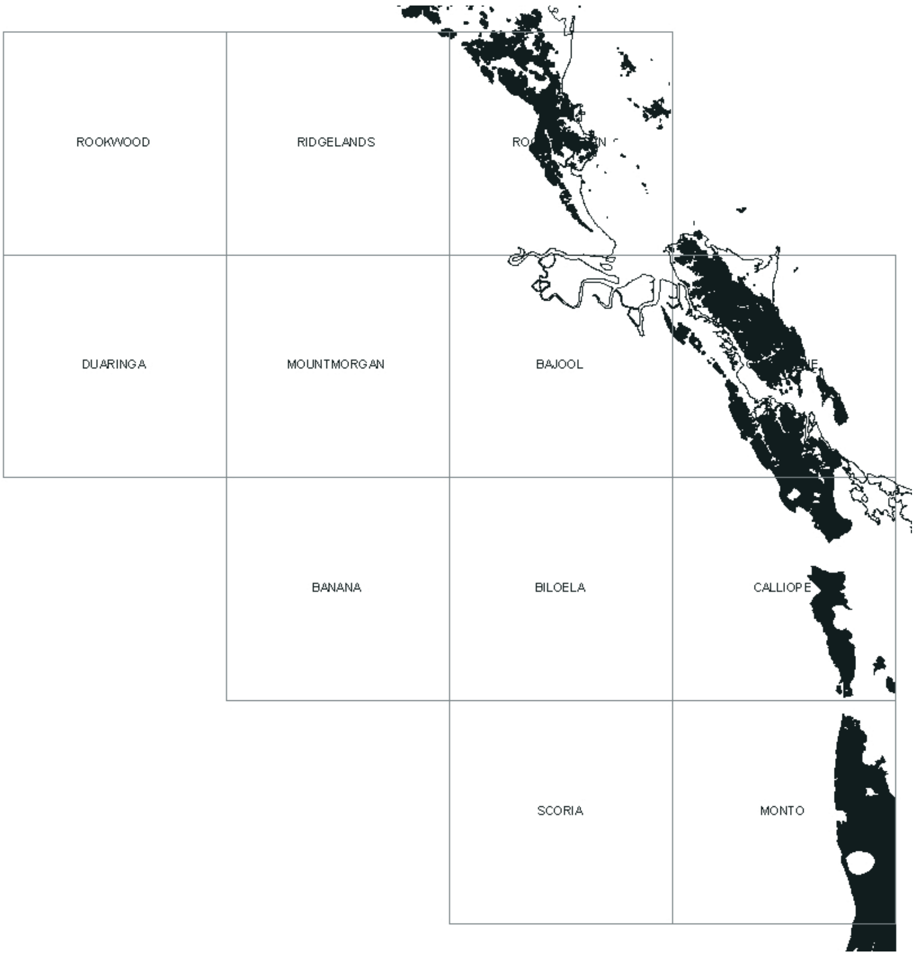


Figure 95. Distribution of the Curtis Island Group

wedge. Although the Doonside Formation contains rocks as old as Early Devonian, the actual growth of the Curtis Island Group as an accretionary wedge occurred throughout Carboniferous time.

The Curtis Island Group is assumed to consist of a stack of imbricate thrust slices, although this has not yet been conclusively demonstrated within the Yarrol Project area.

Throughout the whole New England Orogen, the accretionary wedge rocks are virtually unfossiliferous. Roberts (1985b, 1987) found that one type of elliptical crinoid stem recovered from the Wandilla Formation was confined to the Early Carboniferous *Schellwienella* cf. *burlingtonensis* brachiopod zone in the Tamworth Belt (Murray, 1997). This age is consistent with ages from radiolarian cherts within the rest of the accretionary wedge rocks along eastern Australia. Recent detrital zircon ages from the Wandilla and Shoalwater Formations (Korsch & others, 2009a) are consistent with the sparse fossil data.

The Curtis Island Group was not examined in detail during the Yarrol Project, and descriptions are based on previous literature and available data.

Doonside Formation (DCcd)

(P.R. Blake)

Introduction

Kirkegaard & others (1970) first defined the Doonside Formation as the uppermost unit of the Curtis Island Group. However, it is now regarded as the oldest unit within the Group and is considered to be Devonian to Early Carboniferous in age. The unit consists of ribbon chert and jasper with argillite and minor tuff and limestone, and is interpreted to represent an ocean floor association (Murray, 1994).

Distribution

The Doonside Formation is the westernmost unit of the Curtis Island Group, and forms a discontinuous belt immediately east of the Yarrol Fault. It extends from The Pointers in the Port Clinton 1:250 000 Sheet area in the north to Mundubbera in the south (Figure 96).

Topographic expression

In most places the Doonside Formation forms a prominent range of hills, including strike ridges of chert and jasper.

Geophysical expression

On images derived from airborne radiometric data, the Doonside Formation gives a very low response in all three channels, reflecting its association of ocean floor rock types. The unit also displays a low magnetic response. A linear magnetic high along its western margin is thought to be related to serpentinites along the Yarrol Fault. A linear magnetic feature in the eastern part of the belt on the Gladstone 1:100 000 Sheet area does not appear to be associated with a fault, and may represent a mafic intrusion at depth.

Lithology

Rhythmically interbedded radiolarian chert and mudstone dominate the Doonside Formation. Many of the jasper and chert units are thin-bedded and locally exhibit slumping, but other beds are very thick and massive. The mudstones are locally sandy, and some are silicified and display thin bedding. Minor greywacke, fine-grained volcaniclastic sandstone, and limestone are also present, and spilitic volcanics of the Balnagowan volcanic member occur sporadically in the western part of the unit. The fine-grained sandstones commonly contain minor amounts of quartz (Donchak & Holmes, 1991). Rare rock types include lithic sandstones containing clasts of chert, and mudstone conglomerates with angular siltstone fragments. Accessory sulphides are present in some beds.

Environment of deposition

The Doonside Formation is interpreted to have been deposited on the sea floor in a deep marine (abyssal) environment, and represents a very slow rate of deposition.

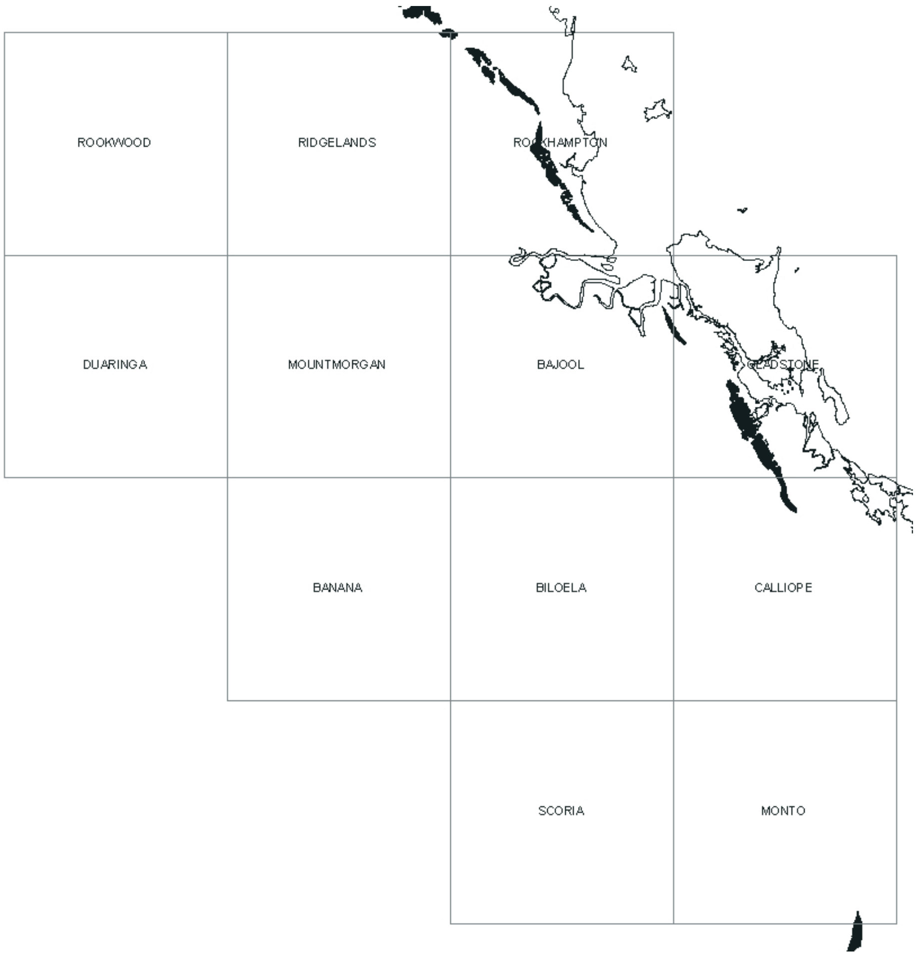


Figure 96. Distribution of the Doonside Formation

Thickness

Probable repetition by faulting in an accretionary wedge makes it impossible to estimate the thickness of this unit. Certainly the figure of 2100–3600m given by Kirkegaard & others (1970) is a gross overestimate.

Structure

The most detailed structural studies by Fergusson & others (1988, 1990a, 1990b, 1993) combined the Doonside and Wandilla Formations as the Wandilla terrane, and are summarised in the description of the Wandilla Formation. Thin-bedded chert within the Doonside Formation locally displays slumping.

Biostratigraphy and age

Bedded chert collected by Fergusson & others (1993) from Devils Elbow on the Calliope River 10km south-west of Gladstone yielded a conodont fauna that was assigned to the Late Silurian to Middle Devonian, but which may be confined to the Early Devonian (B.G. Fordham, personal communication). Radiolarians from this locality are poorly preserved and not age diagnostic. Other conodont and radiolarian faunas from equivalent rocks of the New England Orogen near Gympie and Brisbane are of Late Devonian and Early Carboniferous age (Aitchison, 1988; Ishiga, 1990; B.G. Fordham & H. Ishiga, personal communication). It is possible that deposition of the Doonside Formation extended through much of the Devonian and continued into the Early Carboniferous (Mississippian).

Stratigraphic relationships

The Doonside Formation is separated from the Yarrol Province by the Yarrol Fault. The contact with the Wandilla Formation to the east is both a stratigraphic and structural boundary.

Correlation with other units

The age determined from conodonts (Fergusson & others, 1993) suggests that part of the Doonside Formation may be a time equivalent of the Calliope, Erebus and Craigilee beds. However, there are no similarities in rock types since the Calliope, Erebus and Craigilee beds are dominantly volcaniclastic sediments whereas the Doonside Formation is dominated by chert.

If the Doonside Formation ranged into the Late Devonian and Early Carboniferous, then it must also include time equivalents of the Mount Alma Formation. One similarity between these units is the abundance of fine-grained sedimentary rocks that contain radiolarians.

Economic significance

Manganiferous mineralisation occurs in the Doonside Formation, and is more common than in the Wandilla Formation. A belt containing lenses of manganese ore is present between Mount Miller and Mount Beecher west of Gladstone. Psilomelane is the dominant ore, but some pyrolusite and probably braunite are also present.

Balnagowan Volcanic Member (DCcd$_b$)

(P.R. Blake)

Introduction

The Balnagowan volcanic member was described as an informal member of the Doonside Formation by Willmott & others (1986). Donchak & Holmes (1991) correlated mafic volcanics interbedded with cherts of the Doonside Formation on the Gladstone 1:100 000 Sheet area with the Balnagowan volcanic member. Only limited mapping was carried out during the present investigation, and this description is largely taken from previous studies.

This usage of the name Balnagowan volcanic member appears to refer to two distinctly different sequences. Willmott & others (1986) took the name from Balnagowan homestead, where a prominent hill is composed of shallowly-dipping felsic tuff and agglomerate. They suggested that this sequence could be a correlative of Lower to Middle Devonian volcanic arc rocks west of the Yarrol Fault, in which case they should be separated from the Doonside Formation. Previously, on the Rockhampton 1:250 000 geological map, this sequence was mapped as part of the Lower Permian Berserker beds. The most recent mapping has supported this latter interpretation, and placed these rocks in the Chalmers Formation of the Berserker Group.

In this report, the term Balnagowan volcanic member is restricted to mafic volcanic rocks and associated lithologies that locally are interbedded with cherts of the Doonside Formation immediately east of the Yarrol Fault.

Distribution

The main outcrop of the Balnagowan volcanic member forms a north-north-west trending belt 20km long and up to 3km wide from Balnagowan homestead to Cawarral, east of Rockhampton. A smaller outcrop is located south-east of Targinie (Figure 97).

Topographic expression

The member generally produces a low topography.

Geophysical expression

The radiometric response of the unit is generally very low, with only local patches slightly enriched in potassium. The unit gives no distinctive signature on images derived from airborne magnetic data.

Lithology

From the description by Willmott & others (1986), the sequence now included in the revised Balnagowan volcanic member consists of basic to intermediate lavas and tuffs, with some mudstone, chert and limestone. Volcanics correlated with the member by Donchak & Holmes (1991) were described as glassy to fine-grained basaltic tuffs and flows, containing hornblende, and in places, scattered clinopyroxene, with epidote, chlorite

Figure 97. Distribution of the Balnagowan volcanic member

and minor calcite occurring as common secondary minerals. The volcanics were massive and cut by numerous felsic veins. Flow margin breccias and amygdaloidal textures are present locally.

Environment of deposition

Excluding the sequence at Balnagowan homestead mapped as Chalmers Formation, the depositional environment of the Balnagowan volcanic member is interpreted to be deep marine, consistent with its position as a member of the Doonside Formation.

Thickness

The thickness of the member is not known, as it probably consists of a number of imbricate fault slices.

Structure

The structure is assumed to be similar to that of the remainder of the Doonside Formation, but the massive nature of most of the mafic volcanic rocks prevents useful observations.

Biostratigraphy and age

No fossils have been collected from this unit. No conodonts were recovered from limestone lenses near Cawarral and Tungamull.

Stratigraphic relationships

The Balnagowan volcanic member is at least locally interbedded with typical cherts of the Doonside Formation.

Correlation with other units

The Balnagowan volcanic member, like the Doonside Formation, could range in age through much of the Devonian into the Early Carboniferous. Geochemical studies on the mafic volcanics would help to establish relationships with volcanic-bearing sequences of this age west of the Yarrol Fault, although they probably include oceanic seamounts. It is possible that some parts of the Balnagowan volcanic member are actually Berserker Group, similar to the outcrop near Balnagowan homestead.

Economic significance

Limestone lenses within the Balnagowan volcanic member are small, but of high grade.

Wandilla Formation (DCcw)

(P.R. Blake)

Introduction

The Wandilla Formation was defined by Kirkegaard & others (1970) as the middle unit of the Curtis Island Group.

Distribution

The Wandilla Formation covers a large area, and forms a discontinuous belt along the entire eastern edge of the mapped area, over a total length of 240km and a width of up to 20km (Figure 98).

Figure 98. Distribution of the Wandilla Formation

Figure 99. Convolute bedding in steeply-dipping, deformed turbidites in Middle Creek east of Kalpowar, Wandilla Formation. Detrital zircons in a sandstone sample from this locality have been dated (Korsch & others, 2009a; WAN40).

Topographic expression

The topography varies from steep, irregular ridges to rounded hills. Some massive chert lenses form prominent strike ridges.

Geophysical expression

On false colour images derived from airborne radiometric data, the Wandilla Formation mostly has a moderate signature, with pink to red tones. Some areas have a dull blue-green colour indicating low levels of potassium and relative enrichment of uranium and thorium. Dark bands with a very low response east of Monto are ridges formed by large chert lenses, some of which are silicified.

The formation has a very low magnetic response.

Lithology

The Wandilla Formation consists dominantly of rhythmically interbedded mudstone and sandstone. Chert occurs as large lenses that form prominent ridges, particularly on Curtis Island, and is either thin-bedded or massive. The sandstones are dominantly sourced from intermediate to felsic volcanic rocks. Quartz content is variable, and ranges from minor to ~65%. Many of the sandstones are graded and interpreted as turbidites. Some show soft sediment deformation (Figure 99). In the middle of the unit is a sandstone horizon containing calcareous or silicified ooids that were sourced from the oolitic carbonate shoals in the Yarrol terrane to the west (Fleming & others, 1975). Fergusson & others (1988, 1993) reported the presence of massive silicic tuff on Picnic and Witt Islands in Gladstone Harbour.

Environment of deposition

Kirkegaard & others (1970) considered the Wandilla Formation to have been deposited in a deep-water (abyssal) environment. They suggested that sedimentation was rhythmic, with sandstone deposition during periods of crustal instability or erosion, and chert formation during periods of quiescence. However, subsequent interpretations of these rocks as part of an accretionary wedge are more compatible with a dual provenance, the volcaniclastic sandstones being deposited from turbidites sourced mainly from a continental margin arc to the west, and the cherts from the ocean floor to the east.

Thickness

The total sequence represented by the Wandilla Formation was estimated by Kirkegaard & others (1970) to be between 10 000–16 000m thick, but because the unit consists of an indeterminate number of fault slices within an accretionary wedge, the true thickness must be much less.

Structure

Fergusson & others (1988, 1990a, 1990b, 1993) have carried out the most detailed structural studies of the Curtis Island Group, and this summary is based on their work. They divided the accretionary wedge assemblage into the Wandilla terrane (Doonside and Wandilla Formations) and the Shoalwater terrane (Shoalwater Formation).

The earliest deformation in the Wandilla terrane (D_1) formed widespread mélange of lenticular character in which discoidal fragments of greywacke are set in a mudstone matrix with a steeply dipping scaly fabric (S_f) defined by anastomosing shear fractures. The mélange developed by layer-parallel extension along fault zones. In areas of coherent strata, D_1 formed upright tight to isoclinal folds with an axial plane slaty cleavage S_1, and probably imbricated the succession along steeply dipping faults. D_1 is inferred to have occurred soon after the deposition of the interbedded greywacke and mudstone in the Early Carboniferous as these were disrupted, before lithification, to form mud-seam mélange.

In the Emu Park – Yeppoon area, and in the eastern part of the unit south of Gladstone, the dominant structure is an intense crenulation cleavage that dips at moderate to shallow angles to the east, or is subhorizontal. This is interpreted as a D_2 structure, although it has not been seen to overprint steeply dipping S_f or S_1. The cleavage is parallel to or shallower than the transposed lithological layering, demonstrating an eastward vergence and suggesting the existence of an antiform to the east. These D_2 structures are also well developed in the Shoalwater terrane. Fergusson & others (1990, 1993) attributed them to the Late Permian to mid-Triassic Hunter–Bowen Orogeny on the basis of available metamorphic age determinations (Leitch & others, 1993), but their geometry suggests that they could be older structures related to the Early Permian extensional event recorded by sequences west of the Yarrol Fault.

A third deformation, D_3, produced local north-north-west striking kinks and broad upright folds with an axial plane crenulation cleavage, and a more widely distributed conjugate system of strike-slip faults (Fergusson & others, 1993). The D_3 structures are considered to be related to the Late Permian to mid-Triassic Hunter–Bowen Orogeny.

Biostratigraphy and age

Elliptical crinoid stems have been recovered from the Wandilla Formation (Fleming & others, 1975), and investigations by Roberts (1985b, 1987) found that they were confined to the Early Carboniferous *Schellwienella* cf. *burlingtonensis* zone. This, combined with the relatively common occurrence of calcareous ooids in the sandstones, has led to the interpretation that the Wandilla Formation was coeval with the Rockhampton Group in the Yarrol Province.

Detrital zircons have been dated from several localities in the Wandilla Formation at Gladstone, Emu Park, and Middle Creek 35km north-east of Monto (Korsch & others, 2009a). The results support an Early Carboniferous (Mississippian) age.

Stratigraphic relationships

The Wandilla Formation is in structural and locally stratigraphic contact with the Doonside Formation to the west and the Shoalwater Formation to the east. Transitional lithologies occur within the Wandilla Formation near its contacts with both of these units.

Correlation with other units

McKellar (1967) noted the similarity of oolitic sandstones of the Wandilla Formation to rocks in the eastern part of the Rockhampton Group. Oolitic limestones along the western edge of the Rockhampton Group were the source of calcareous ooids that were transported eastwards into the deeper parts of the forearc basin and down the continental slope in turbidites.

Economic significance

The formation contains minor phosphatic and to a lesser extent manganiferous mineralisation. Veins of variscite and metavariscite are best developed at Keppel Rocks, a small islet north of Curtis Island (Murray, 1968).

Shoalwater Formation (Ccs)

(P.R. Blake)

Introduction

The Shoalwater Formation is the easternmost unit of the Curtis Island Group. Kirkegaard & others (1970) regarded the Shoalwater Formation as the oldest unit in the group, but it is now recognised as the youngest unit of an accretionary wedge assemblage. Its eastern boundary is everywhere concealed beneath the Pacific Ocean.

Distribution

Within the area covered by this report, the Shoalwater Formation forms a chain of islands in Keppel Bay, continues south as a relatively thin strip along the east coast of Curtis Island to Facing Island, and reaches the mainland south of Gladstone (Figure 100). Much larger exposures, including the type area, occur on the Port Clinton 1:250 000 sheet to the north.

Topographic expression

This unit produces subdued topography with a closely spaced dendritic drainage pattern.

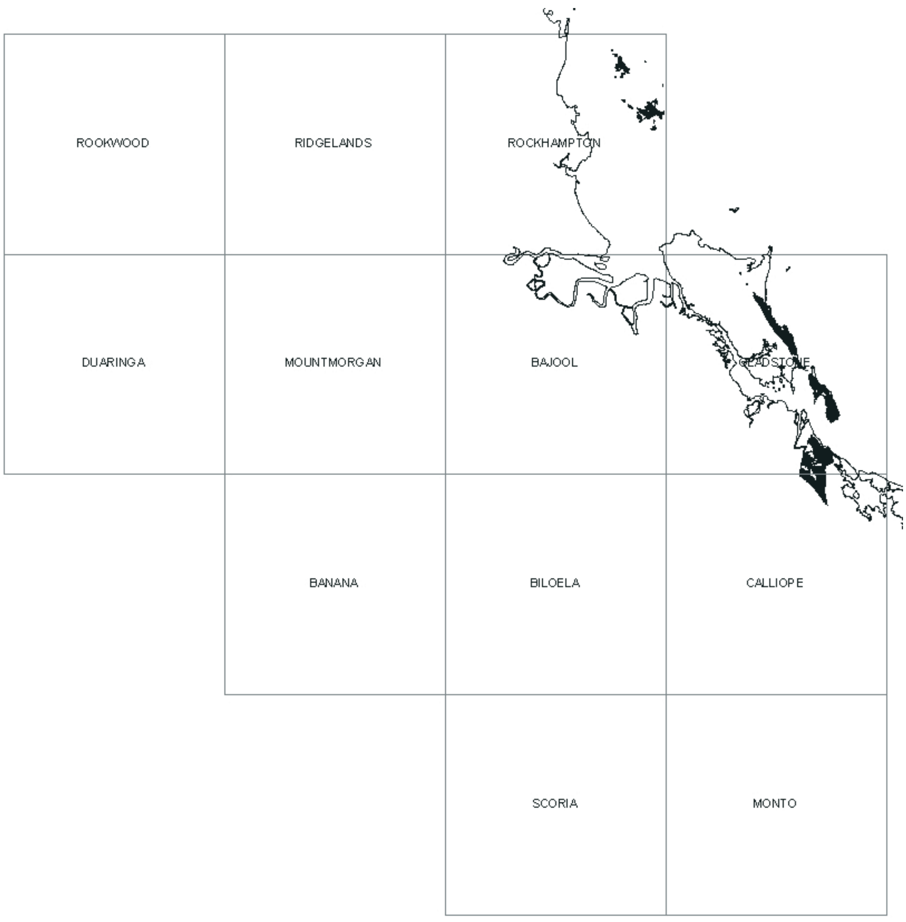


Figure 100. Distribution of the Shoalwater Formation

Geophysical expression

Only a very small area of the Shoalwater Formation crops out in the area covered by airborne geophysics, and it is difficult to determine its overall geophysical signature. However, the unit appears to have a low radiometric and magnetic signature.

Lithology

The Shoalwater Formation comprises thick quartzose sandstone beds which alternate with thinner mudstone units. The sandstones are mainly medium to fine-grained, poorly sorted, and have a restricted compositional range, averaging 79% quartz (Leitch & others, 2003). Beds are usually massive, although some are graded, and can be up to 3m thick. Mudstone is usually medium to dark grey to black, and is very thin-bedded to laminate. In the type area, the formation contains rare beds of chert or fine felsic tuff that produce strike ridges.

In some locations there is an increase in metamorphic grade and the sandstones become strongly foliated quartz-muscovite-biotite schists and gneiss, notably on Hummocky Island, the south end of Facing Island, and the Tannum Sands area south of Gladstone.

Environment of deposition

The Shoalwater Formation, like the remainder of the Curtis Island Group, is interpreted to have been deposited in a deep marine environment and incorporated into an accretionary wedge. Its quartz-rich nature has led to suggestions that it represents an exotic terrane, which collided with the remainder of the accretionary wedge in post-Carboniferous time (Korsch & Harrington, 1985, 1987). However, the contact between the Wandilla Formation and Shoalwater Formation in the type area has been described as transitional (Kirkegaard & others, 1970). The marked contrast in composition between sandstones of the Wandilla and Shoalwater Formations has been ascribed to provenance switching (Leitch & others, 2003), a concept that is strongly supported by recent dating of detrital zircon grains.

Thickness

Kirkegaard & others (1970) estimated the Shoalwater Formation to be over 10 000m thick. However, this is a gross overestimate that does not take account of its imbricate fault structure as part of an accretionary wedge (Fergusson & others, 1990b). The true thickness is not known.

Structure

Fergusson & others (1988, 1990a, 1990b, 1993) have carried out the most detailed structural studies of the Curtis Island Group, and this summary is based on their work. They divided the accretionary wedge assemblage into the Wandilla terrane (Doonside and Wandilla Formations) and the Shoalwater terrane (Shoalwater Formation).

Along the coast and adjacent islands from Yeppoon to south of Gladstone, the dominant structure in the Shoalwater terrane is an intense crenulation cleavage, interpreted as S_2, that ranges from shallowly east-dipping to subhorizontal, and dips gently to the west on Great Keppel Island. It is axial plane to isoclinal folds with an eastern to north-eastern vergence. Most of the strata are overturned. S_1 is preserved as a weak bedding parallel cleavage. Fergusson & others (1990, 1993) attributed the D_2 structures to the Late Permian to mid-Triassic Hunter–Bowen Orogeny on the basis of available metamorphic age determinations (Leitch & others, 1993), but their geometry suggests that they could be older structures related to the Early Permian extensional event recorded by sequences west of the Yarrol Fault.

On Great Keppel Island, Smith (1998) described strongly boudinaged sandstone beds that are folded by F_2 and overprinted by the S_2 foliation, indicating that boudinage occurred either during the D_1 accretion event, or during a previously unrecognised deformational event.

D_3 structures are similar to those in the Wandilla terrane.

Biostratigraphy and age

The only fossils known from the Shoalwater Formation are unidentifiable plant fragments in black shales. The unit is assumed to be younger than the Wandilla Formation based on its position within the accretionary wedge complex, and this is consistent with data from detrital zircons (Carson & others, 2006; Korsch & others, 2009a).

Stratigraphic relationships

The contact with the Wandilla Formation is structural and locally stratigraphic, and the boundary is transitional.

Correlation with other units

The Shoalwater Formation has no obvious correlatives within the Curtis Island Group or the Yarrol Province. The high percentage of quartz compared to the Doonside and Wandilla Formations broadly coincides with the increasing proportion of felsic volcanic and plutonic detritus in the Lorray Formation compared to the units that underlie it, but it is probable that the Shoalwater Formation was not locally sourced (Leitch & others, 2003; Korsch & others, 2009a).

LATE CARBONIFEROUS – EARLIEST PERMIAN

Lorray Formation (CPl)

(G.A. Simpson & P.R. Blake)

Introduction

The Lorray Formation is an Upper Carboniferous (Pennsylvanian) to lowermost Permian succession dominated by fine to coarse-grained sandstone and bryozoan-rich mudstone. The succession is also characterised by ubiquitous granitic detritus and quartz in conglomerate and sandstone.

The Lorray Formation combines a number of previously defined units, including: the Neerkol Formation in the Stanwell area (Fleming, 1960, 1967); the Branch Creek Formation, Poperima Formation and Rands Formation in the Yarrol and Cannindah areas (Maxwell, 1959, 1960a, 1964), subsequently combined into the Boiling Creek Group by McKellar (1967); and the Burnett Formation in the Yarrol area (Maxwell, 1959, 1960a,b, 1964; Dear & others, 1971). All these units are similar in constituent rock types, geophysical response, and faunas. Dear & others (1971, page 33) noted that the Boiling Creek Group "can be correlated almost in its entirety with the Neerkol Formation", and (page 36) that "beds in the upper part of the Rands Formation (of the Boiling Creek Group) are very similar to those in the lower part of the Burnett Formation", the boundary being impossible to recognise in many areas. The Rands and Burnett Formations were combined by Brown (1970) and Edwards (1974), although they did not introduce a new stratigraphic name. Subsequent mapping has confirmed that all of these units are difficult to distinguish other than by their location, and that previous local subdivisions have been based more on faunal successions than on changes in rock type.

Derivation of name

The name is taken from Lorray homestead, which is located 28km south-east of Monto at GR333600 7332900 on the Eidsvold 1:100 000 Sheet area.

Distribution

The Lorray Formation is largely restricted to localities in the north-west and south-east of the mapped area (Figure 101). The main outcrops are north-west and east of Mount Salmon west of Ridgelands, near Stanwell west of Rockhampton, and in the Yarrol Syncline east of Monto. Smaller areas are near Mount Morgan and west of Yarwun.

Dear & others (1971, page 30) reported fossils indicative of the Lorray Formation in boulders in Marble Creek north-west of Nagoorin. No outcrops have been found, and radiometric images of the catchment area give no indication of the occurrence of this unit, which typically gives a very strong and easily recognised response.

Type area

The type area for the unit is located in the Yarrol Syncline, ~25km south-east of Monto. Type sections for the Burnett Formation and the constituent formations of the Boiling Creek Group were nominated by Maxwell (1959), with subsequent minor changes by Maxwell (1964).

Fleming (1967) defined a type section for the Neerkol Formation west of Mount Lion, along part of Lion Creek and streams to the north.

Figure 101. Distribution of the Lorray Formation

Topographic expression

The unit forms low to moderately hilly topography with prominent strike ridges in most areas.

Geophysical expression

On images derived from airborne radiometric data, the Lorray Formation has a characteristically high response in all three channels. This response is similar to that for the Youlambie Conglomerate (a partial time equivalent) but much higher than that of the immediately underlying units.

The unit has a uniformly low magnetic signature, comparable to that of the underlying Rockhampton Group.

Lithology

Within the Lorray Formation, there is a significant difference in the proportion of rock types in the Yarrol Syncline area compared to the Stanwell and Mount Salmon areas. In the Yarrol Syncline, the unit comprises conglomerate, sandstone, siltstone, mudstone, and minor limestone. The sandstone and conglomerate are usually characterised by widely distributed detrital granite fragments and relatively high quartz content compared to the underlying Rockhampton Group and overlying Yarrol Formation.

Conglomerates have a granule to cobble grainsize, are grey-green or grey-brown in colour, and are composed dominantly of intermediate to felsic volcanics. They commonly contain sparse granitoid clasts. The sandy matrix consists of lithic grains, feldspar and minor quartz.

Sandstones are usually medium to coarse-grained, commonly cross-bedded, and lithic to feldspatholithic in composition. They usually consist of intermediate to felsic, aphyric to porphyritic volcanic clasts, granophyre, quartz and feldspar.

Minor rock types include fossiliferous fine-grained lithic sandstone, siltstone, dark calcareous mudstone, detrital limestone and minor fossiliferous limestone. Macrofossils include bryozoa, brachiopods and crinoids.

Dear & others (1971) reported some andesite near the top of the unit, but were uncertain whether this was part of the sequence or intrusive.

In the Stanwell and Mount Salmon areas, the dominant rock types are massive mudstone and fine-grained silty sandstone. Coarse sandstone, conglomerate and bioclastic limestone are only minor constituents (Kirkegaard & others, 1970). Fleming (1960) reported that feldspar is the main clastic component, averaging ~40% of sandstones. Clasts of felsic to intermediate volcanic rocks are abundant, and are dominant in some specimens. Quartz locally comprises as much as 50% of sandstones. Relatively large granitic clasts are a widespread but minor constituent.

Environment of deposition

The Lorray Formation was deposited in a marine shelf environment as indicated by the abundance of marine macrofossils. The scarcity of cross bedding and the lack of ripple marks suggest deposition below the tidal zone. The widespread distribution of granitoid detritus in sandstones and conglomerates indicates deposition during a period of extensive denudation of granite plutons, as well as felsic to intermediate volcanic rocks.

The much more restricted distribution of the Lorray Formation compared with that of the underlying Rockhampton Group (compare Figures 82 and 101) probably reflects both limited depositional extent and subsequent erosion.

Thickness

The Lorray Formation is from 3500–4100m thick in the type area of the Yarrol Syncline near Lorray homestead (Maxwell, 1964). Further north in the Cannindah Creek area, where the top of the unit is not preserved, it is only 1540m thick (McKellar, 1967). The thickness in the Stanwell area is also variable due to erosion, and ranges from 610–2130m (Fleming, 1967). Near Mount Salmon, the thickness is at least 1520m (Kirkegaard & others, 1970), but the upper part of the unit is concealed beneath younger rocks.

Structure

Folds are generally open, with moderate to shallow dips. The Yarrol Syncline is tighter, with limb angles up to 75°. Folding is attributed to west-directed thrusting, but the faulting is widely spaced, and most sections through the unit are continuous and undisturbed.

Late out-of-sequence thrusting is evident in the western limb of the Yarrol Syncline, where the Tiverton Fault, dipping to the east and south-east at ~30°, has emplaced younger strata over older.

Biostratigraphy and age

Fossil faunas in the Lorray Formation were described by Fleming (1960), Maxwell (1964) and McKellar (1965). Dear & others (1971) summarised the faunal succession of the Boiling Creek Group in the Cannindah Creek area as listed by McKellar (1967).

Specimens listed by Dear & others (1971) include *Lissochonetes montinis* McKellar, *Inflatia dakielis* McKellar, *Neospirifer campbelli* Maxwell, *Alispirifer laminosus* Campbell, *Punctospirifer hirsutus* McKeller, *Punctospirifer ambiguus* Maxwell, *Evactinopora* sp., *Levipustula levis* Maxwell, *Neospirifer pristinus* Maxwell, *Spiriferellina neerkolensis* Maxwell, *Spinuliplica spinulosa* Campbell, *Punctospirifer ambiguus* Maxwell, *Beecheria* sp. nov., *Limipecten burnettensis* Maxwell, *Schizodus elliotti* Maxwell, and *Planikeenia burnettensis* Maxwell. The fauna from the former Burnett Formation includes *Cancrinella levis* Maxwell, *Megadesmus cf. gryphoides* (de Koninck), *Eurydesma cf. playfordi* Dickins, *Promytilus cancellatus* Maxwell, and *Deltopecten squamuliferus* (Morris). Many of these fossils also occur in the former Neerkol Formation, including *Cancrinella levis* near the top of the unit, but the faunas are dominated by bryozoans.

The fossils indicate that most of the unit is of Late Carboniferous (Pennsylvanian) age, but the upper part in the Yarrol Syncline, formerly mapped as the Burnett Formation, is early Permian (Dear & others, 1971).

Stratigraphic relationships

The Lorray Formation appears to overlie the Rockhampton Group with structural conformity. Locally there is a hiatus at the boundary, and the contact is actually a disconformity, as in the Stanwell area (Jenkins & others, 1993). Here, the missing sequence, representing one conodont zone, is actually from the top of the Rockhampton Group.

The unit is conformably overlain by the Yarrol Formation in the Yarrol Syncline, the contact being marked by a distinct change of rock types. In the Stanwell area, the Dinner Creek Conglomerate disconformably overlies the Lorray Formation, with evidence of erosion of the upper part of the Lorray Formation as described by Fleming (1967) and Kirkegaard & others (1970). Near Mount Salmon, the Lorray Formation is unconformably overlain by the Cretaceous Mount Salmon Volcanics. West of Mount Morgan, the Lorray Formation is conformably overlain by the Youlambie Conglomerate.

Correlation with other units

In the Monto region, the upper part of the Lorray Formation, the former Burnett Formation, is a marine sequence equivalent in age to the dominantly terrestrial Youlambie Conglomerate. West of Mount Morgan, these two units have been mapped as a continuous stratigraphic succession, the Youlambie Conglomerate overlying the Lorray Formation. The upper part of the Lorray Formation is presumably missing at this locality.

A phase of granitoid intrusion in the Auburn Arch to the west and south-west, including the Glandore Granodiorite (323.4±3.9Ma), Boam Creek Granite (315±4.0Ma) and Evandale Tonalite (possibly 319±7Ma) (Withnall & others, 2009), coincided with the deposition of the Lorray Formation. Possibly unroofing of these granites provided the widespread granitic detritus seen in the Lorray Formation. Unroofing of granite plutons can be extremely rapid in this tectonic environment (Kimbrough & others, 2001).

Deposition of the Lorray Formation also continued through two phases of dominantly felsic volcanism in the Auburn Arch, named the Torsdale and Camboon Igneous Cycles by Hutton & others (1999), which could have provided a source of the abundant volcanic material in the unit.

LATEST CARBONIFEROUS – EARLY PERMIAN

Youlambie Conglomerate (CPy)

(G.A. Simpson & M.A. Hayward)

Introduction

The Youlambie Conglomerate was named by Dear (1968) following detailed mapping in the Cania area north of Monto. The unit was extended over the Monto, Rockhampton and Duaringa 1:250 000 Sheet areas by regional mapping programs in the 1960s (Malone & others, 1969; Kirkegaard & others, 1970; Dear & others, 1971).

Distribution

The unit forms a discontinuous, north-west trending belt 250km long and up to 25km wide from Monto in the south to Develin homestead in the north (Figure 102). It crops out in five main areas. The type area is in the western limb of the Spring Creek Syncline north of Monto. A thin belt runs along the eastern edge of the Kroombit Tops from Cedarvale homestead to Mount Seaview. The most extensive outcrop extends north-west from Kroombit Tops through the Callide area to Manton Creek. West of Mount Morgan, the Youlambie Conglomerate separates the western part of the Bouldercombe Igneous Complex from the Kyle Mohr Igneous Complex. A structurally complex area straddling the Fitzroy River from Moah Creek in the south almost as far as Develin homestead in the north includes faulted and folded outcrops of the formation.

Type area

The type section comprises two overlapping parts on either side of Youlambie Creek 13km north of Monto, and was described in detail by Dear (1968). His description has been modified to produce a generalised stratigraphic section (Figure 103).

Topographic expression

The unit is resistant to erosion and forms hilly, lightly timbered country with deeply incised streams. The most rugged topography is in the ranges in the Callide Creek area east-north-east of Biloela.

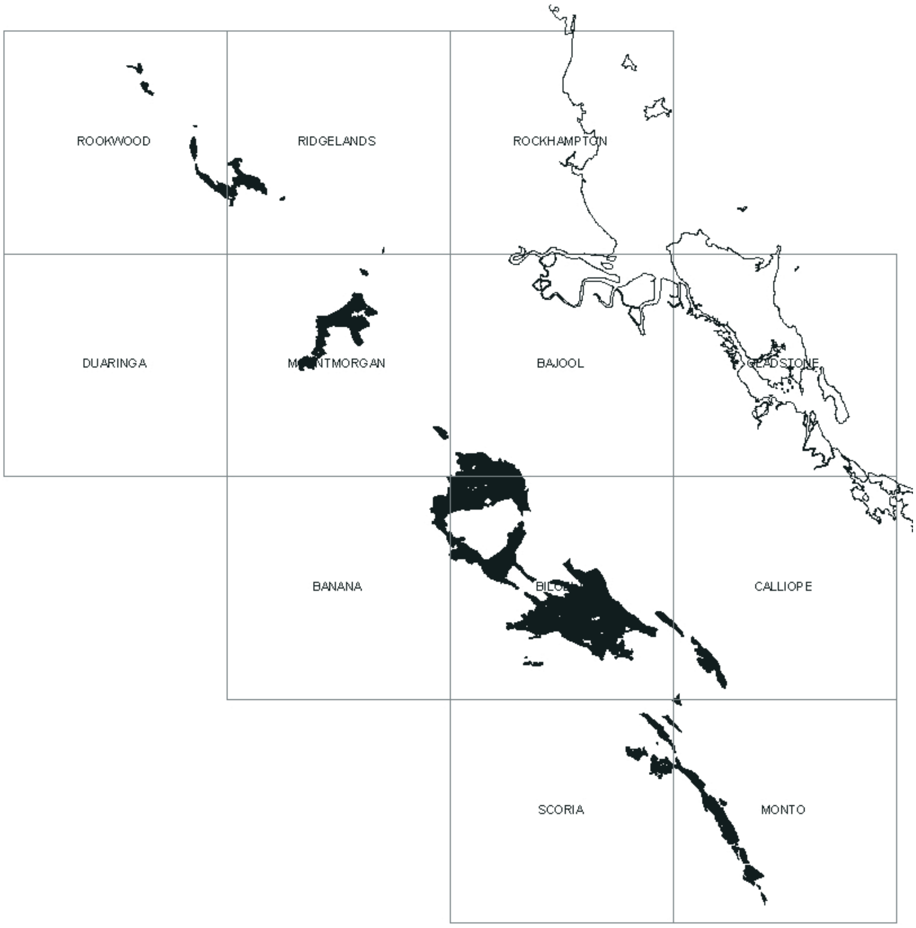


Figure 102. Distribution of the Youlambie Conglomerate

Geophysical expression

The Youlambie Conglomerate is characterised by a high radiometric response in all three channels, similar to the Lorray Formation, and distinctive compared to older geological units. It has a very low magnetic signature.

Lithology

Conglomerate, though not dominant, is characteristic of the unit (Figure 104). Rock types include light brown, commonly polymictic, granite and rhyolite-bearing conglomerate; light brown to pinkish brown lithofeldspathic to lithic quartz-bearing sandstone; dark grey to bluish grey thinly bedded to laminated siltstone and mudstone, locally carbonaceous; and minor greyish purple, rhyolitic ignimbrite. Many of the granitic clasts in the conglomerates and sandstones are granophyric (Figure 105).

A conglomerate along the Dawson Highway north-east of Biloela consists of rounded clasts of rhyolitic ignimbrite (Figures 106 and 107). In the spillway of Cania Dam, finely laminated mudstone-siltstone with possible dropstones (Figure 108) has been interpreted as varved deposits related to glaciation (Dear & others, 1971, page 38; Jones & Fielding, 2004).

Nine facies and six facies associations have been identified in the Youlambie Conglomerate, and are described in Tables 4 and 5.

Geochemistry

Numerous samples of rhyolite and rhyolitic ignimbrite from the Youlambie Conglomerate have been analysed, either from primary volcanics or from clasts in conglomerate. These have given very consistent results. The samples have a large subduction component, being comparatively low in high field strength elements (HFSE), particularly Nb and Zr. The $(Th/Nb)_N$ gap is large, with values ranging from 39.4–56 and averaging 47 (Figure 109). On the granitoid discrimination diagrams of Pearce & others (1984), the analyses plot in the

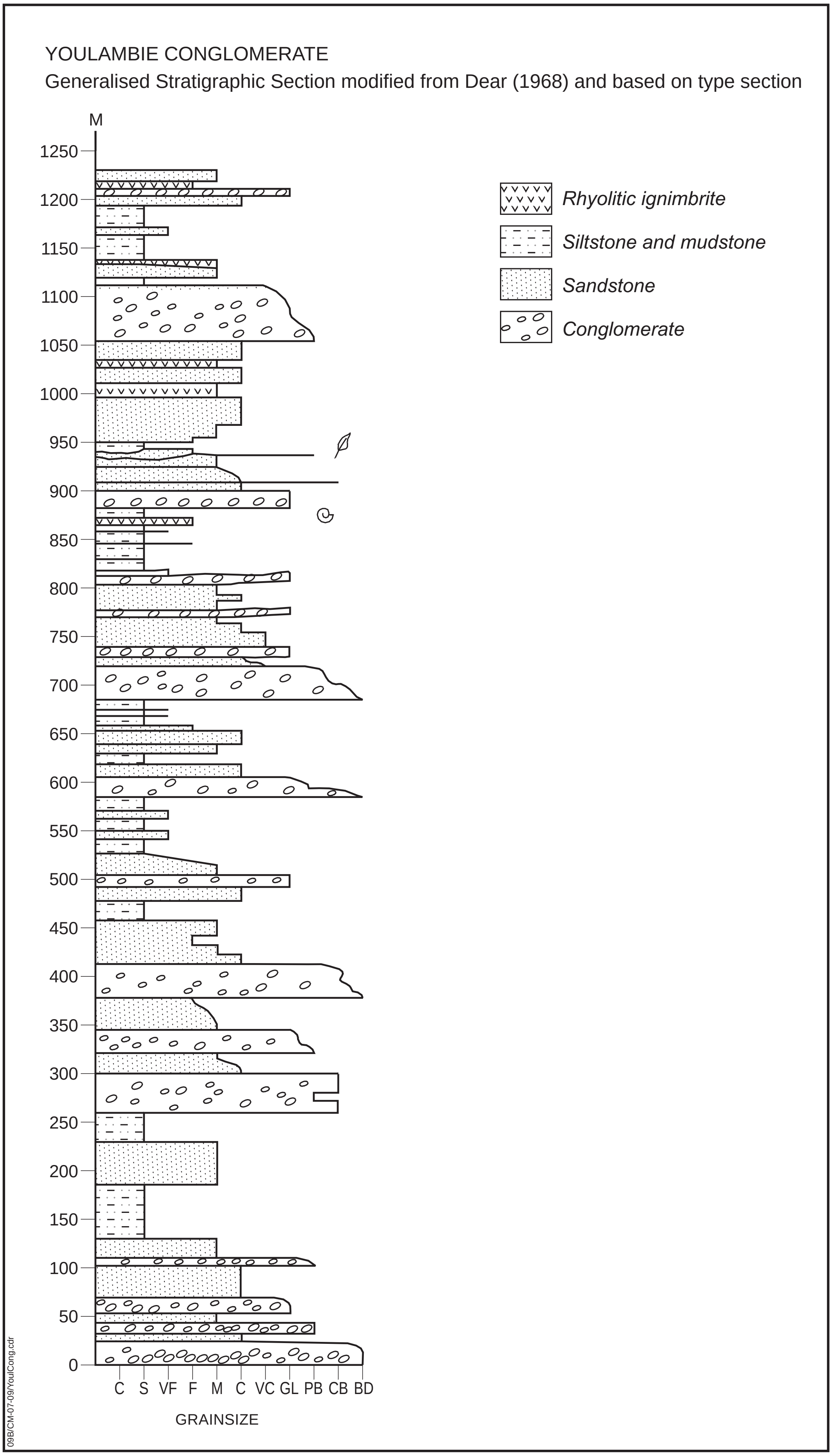


Figure 103. Generalised stratigraphic section of the Youlambie Conglomerate

Figure 104. Clast supported, pebble to cobble conglomerate composed of well rounded felsic volcanic and granophyre clasts, Youlambie Conglomerate. Platy clasts show some alignment with the bedding.

volcanic arc granite field (Figure 58). Pearce (1996) has pointed out that Rb, Nb and Y concentrations in volcanic arc felsic rocks are linked strongly to source, which is interpreted to be a depleted mantle and a crustal component from the subduction process. Because the average composition of continental crust lies within the volcanic arc field, crustal assimilation does not have much effect. Thus the felsic volcanics within the Youlambie Conglomerate, whether primary or as conglomeratic clasts, are interpreted to have been related to an active subduction zone.

Environment of deposition

The formation was deposited in a dominantly terrestrial environment and contains abundant plant fossils including both stems and leaves. Short-lived marine incursions are recorded by the presence of thin beds containing sparse but widely distributed marine fossils.

The terrestrial environment was dominated by fluvial processes and deposition in a braided stream environment, as evidenced by the presence of cross-bedding and imbrication in the conglomerate and sandstone facies, and the lateral extent of the coarse-grained facies in the unit. Varved siltstones with dropstones were interpreted by Dear & others (1971) and Jones & Fielding (2004) to have been deposited in a proglacial lacustrine environment. Dear & others (1971) observed faceting of pebbles and cobbles in the conglomerates of the formation, suggesting that ice flows or glaciers were involved in the transport of this material. They also suggested that the matrix-supported conglomerates of facies C3 were associated with the varved siltstones and represented fluvioglacial deposits. The presence of cobble sized rip-up clasts, the poor sorting, and mixed provenance of these conglomerates combined with their association with fine-grained deposits of a much lower energy regime supports the interpretation that these deposits were transported by ice flows or glaciers. Jones & Fielding (2004) concluded that the glaciation was confined to local valley or mountain glaciers, a finding that is consistent with the absence of evidence for glaciation in rocks of the Youlambie Conglomerate north from Cania (Kirkegaard & others, 1970).

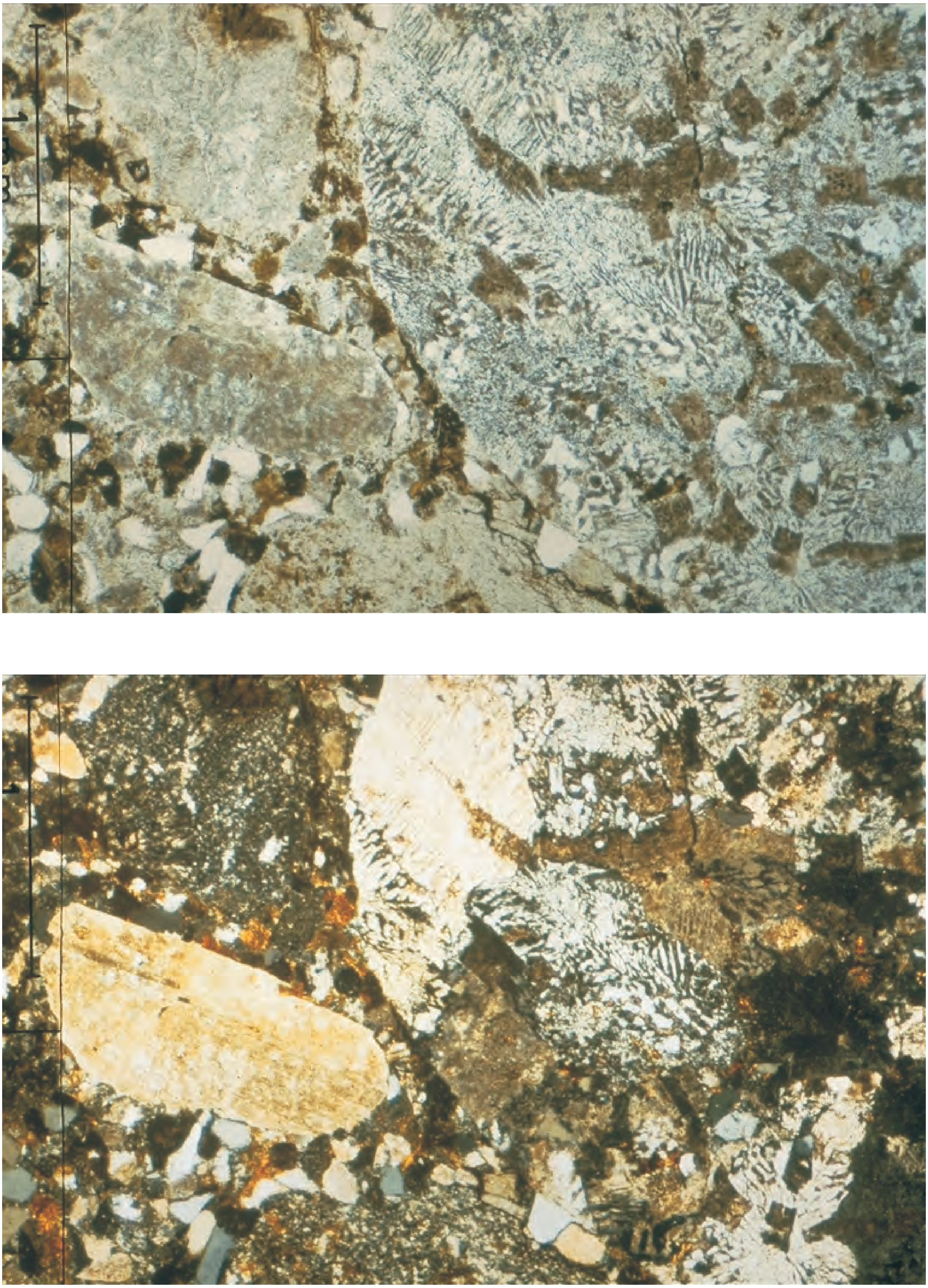


Figure 105. Photomicrographs of conglomerate showing granophyre, felsic volcanic and plagioclase clasts, Youlambie Conglomerate. Plane polarised light and crossed nicols. Scale bar is 1mm long.

Figure 106. Oligomictic pebble to cobble conglomerate composed of well rounded clasts of purple rhyolitic ignimbrite in a matrix of lithic fragments, quartz and feldspar, Youlambie Conglomerate. Field point MHRO 971 on Dawson Highway. Clasts from this locality have been dated (Appendix 1).

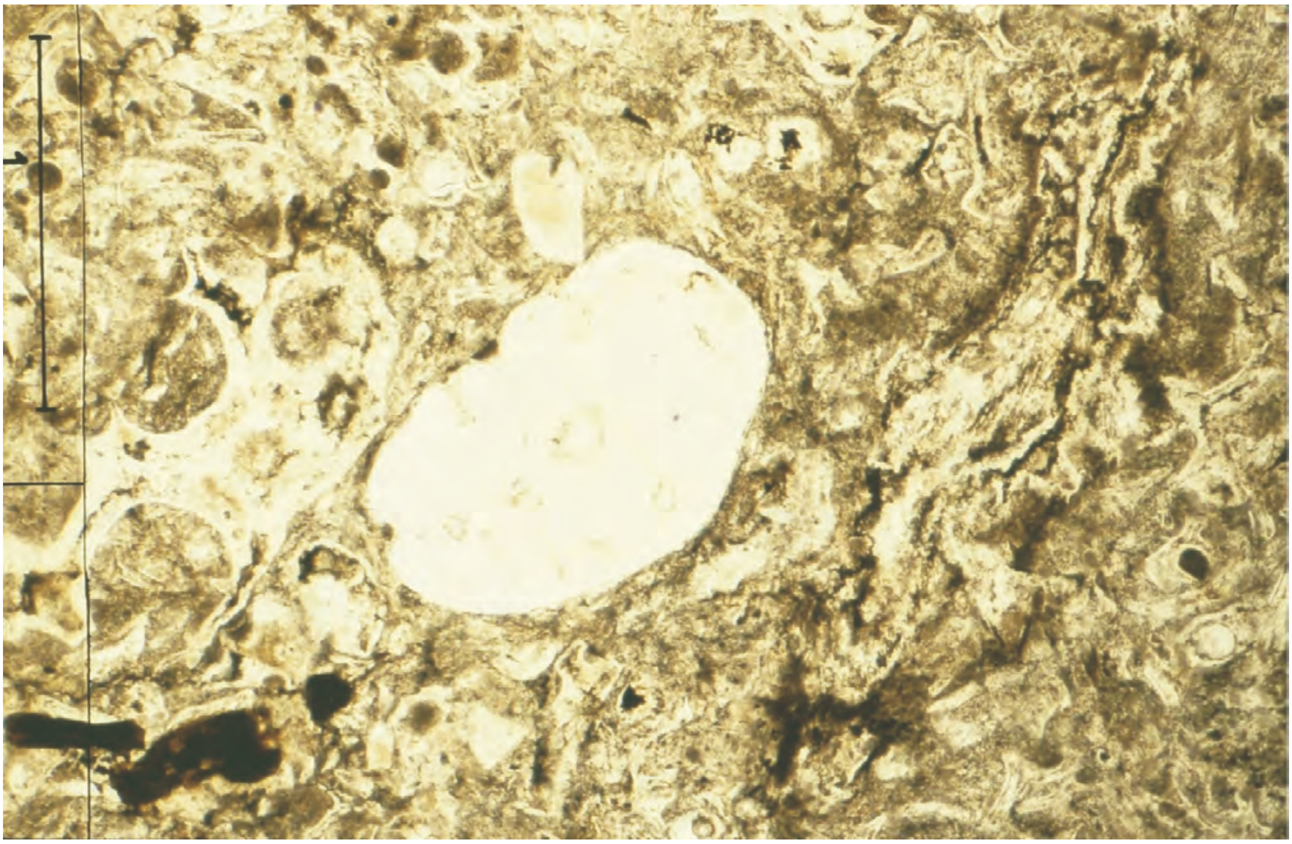


Figure 107. Photomicrograph of rhyolitic ignimbrite clast, MHRO971, Youlambie Conglomerate. Plane polarised light. Scale bar is 1mm long.

Figure 108. Varved siltstone and dropstones, spillway of Cania Dam, Youlambie Conglomerate

Table 4. Facies Description and Interpretation

Facies	Description	Transport Process
C1	Rounded, poorly sorted, pebble to boulder grain-sized, polymictic conglomerate with common granite and dacitic to rhyolitic volcanic clasts. The conglomerates are rarely crossbedded and sometimes imbricated. Facies C1 occurs both as massive, thick beds or as a thickly interbedded sequence with facies S1, S2, and rarely M1. It commonly extends laterally for over 500m.	Strong tractional currents
C2	Subrounded to rounded, pebble to boulder grain-sized, moderately to poorly sorted conglomerate with dominantly dark grey to purple dacitic to rhyolitic volcanic clasts. The facies is usually massive and laterally extensive (>500m).	Strong tractional currents
C3	Subangular to rounded, pebble to boulder grain-sized, matrix supported, poorly sorted polymictic conglomerate with granite and dark grey to purple dacitic-rhyolitic volcanic clasts, and thinly bedded siltstone rip-up clasts. The matrix is fine sand to mud. The facies is commonly massive at its base, becoming interbedded with facies S1 towards its top.	Mud or ice flow
S1	Medium to coarse lithofeldspathic to lithic sandstone with 5– 20% quartz, up to 20% granitic clasts, and up to 50% dark grey to purple dacitic to rhyolitic volcanic clasts. The sandstones commonly exhibit cross stratification and rarely contain fossil plant stems and leaves.	Moderate flow regime currents
S2	Fine to medium calcareous feldspatholithic to lithic sandstone with 5–20% quartz, up to 20% granitic clasts, and up to 50% dark grey to purple dacitic to rhyolitic volcanic clasts. The sandstones contain sparse marine fossils including gastropods, bivalves, and brachiopods.	Wave action or tidal currents with low to moderate flow regime
M1	Laminated thin to medium bedded siltstone and mudstone with rare cross-lamination, commonly containing fossil plant stems and leaves. The facies is commonly interbedded with facies C1 and S1.	Settling from suspension or low flow regime currents
M2	Thin to medium bedded siltstone and mudstone, commonly interbedded with facies C1 and S1.	Settling from suspension or low flow regime current
M3	Thin, rhythmically bedded graded siltstone and mudstone with rare pebble dropstones. There are 12 to 16 graded beds per 2.5cm (Dear & others, 1971; Jones & Fielding, 2004).	Suspension settling from ice flow
V1	Purple rhyolitic crystal-lithic ignimbrites	Pyroclastic flow

Table 5. Facies Association Description and Interpretation

Association	Description	Environmental Interpretation
C1/S1/V1	Thickly bedded conglomerate interbedded with thin to medium bedded sandstone and purple rhyolitic ignimbrite, commonly laterally extensive	Fluvial channels, possibly braided stream
C1/S1/M2	Medium to thinly interbedded conglomerate, sandstone and siltstone with sandstone and siltstone beds becoming thinner and more common towards the top of the sequence; laterally extensive.	Fluvial channel and floodplain
S1/M1/M2	Interbedded sandstone and siltstone with thin beds of plant fossil-bearing mudstone	Interchannel floodplain
C3/S1/M3	Thickly interbedded conglomerate and sandstone overlain by siltstone and mudstone beds	Terrestrial mass flow or glacial till
S1/M3	Thin to medium interbedded siltstone, sandstone, and mudstone	Proglacial lacustrine
S2/M2	Thinly interbedded sandstone, siltstone, and mudstone with marine macrofossils	Shallow marine shelf

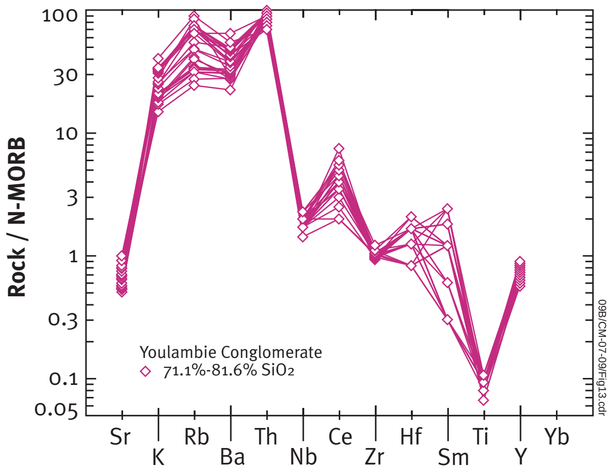


Figure 109. Spidergram plot of analyses of felsic volcanic and granophyre clasts. N-MORB values are from Pearce (1983).

Provenance

The conglomerates and sandstones in the unit were dominantly sourced from granitic and rhyolitic rocks (Dear, 1968). The presence of widespread and relatively abundant granitic and rhyolitic clasts in sedimentary rocks within the unit reflects extensive unroofing of granitic plutons and erosion of rhyolitic volcanics in the provenance area. Rhyolitic ignimbrites in the succession are similar in both colour and geochemical signature to rhyolitic clasts in conglomerates, indicating that volcanism in the source area was still active at the time of deposition.

Thickness

The unit was estimated by Dear (1968) to be 1320m thick in the type area, where both the bottom and top are exposed. Kirkegaard & others (1970, Table 1) suggested thicknesses of 750–1800m in the Rockhampton 1:250 000 sheet area.

Structure

The structure of the Youlambie Conglomerate is quite variable over its outcrop length.

In the type area on the western limb of the Spring Creek Anticline, dips are consistently moderate to steep to the east-north-east. The largest outcrop areas east and north of Biloela are characterised by gently folded, moderately dipping beds. North of Biloela, dips are suggestive of a broad basin centred on the overlying Smoky beds.

West of Mount Morgan, the Youlambie Conglomerate dips shallowly to the west-south-west, conformable with the underlying sequences. The only notable fold structure is a moderately tight curvilinear east–west trending syncline between the Kyle Mohr Igneous Complex and the western part of the Bouldercombe Igneous Complex. This fold parallels the contact of the Kyle Mohr Igneous Complex, and has been attributed to deformation during emplacement of these two plutons (Kirkegaard & others, 1970, pages 59–60). The age of folding must therefore have been late Permian, as the felsic phase of the Kyle Mohr Igneous Complex has been dated at 270.6±3.7Ma (Appendix 1), and the outer unit of the Bouldercombe Igneous Complex, the Flaggy Quartz Monzodiorite, has given latest Permian K-Ar ages.

The discontinuous outcrops of Youlambie Conglomerate that extend across the Fitzroy River from Scrub Creek to Craigilee homestead were regarded by Malone & others (1969) and Kirkegaard & others (1970) as a structurally simple belt that dipped to the south-west and south off a ridge of Silurian–Devonian Craigilee beds. However, detailed structural studies by O'Connell (1995) and Holcombe & others (1997a) have interpreted it as part of the western limb of a large antiformal structure that is a breached propagation fold related to westward thrusting. A relatively tight north-plunging anticlinal structure south of Scrub Creek at the western edge of the Ridgelands 1:100 000 sheet area is superimposed on the overall curvilinear trend of this belt.

Biostratigraphy and age

The most abundant fossils in the Youlambie Conglomerate are plants, including *Cardiopteris polymorpha*, *Noeggerathiopsis hislopi* Bunbury, *Gangamopteris* and *Glossopteris*. These indicate a Permian age, as the *Cardiopteris* overlies a marine fauna including *Eurydesma* (Dear, 1968).

Marine faunas occur at several localities midway through the unit in the type area of the Spring Creek Syncline. The faunas include *Eurydesma* cf. *playfordi*, *Eurydesma burnettensis*, *Eurydesma cordatum*, *Astarila* sp., *Schizodus minutus*, *Schizodus australis*, *Megadesmus gryphoides*, *Megadesmus pristinus*, *Promytilus cancellatus*, and *Deltopecten illawarrensis* (Dear, 1968; Webb, 1977; Briggs, 1998). These faunal assemblages also occur near the top of the Lorray Formation, in rocks previously mapped as Burnett Formation. They cover the *Strophalosia concentrica* to *Bandoproductus walkomi* zones of Briggs (1993, 1998), indicating a Sakmarian (early Permian) age, although the entire unit may extend down into the late Asselian and upwards into the early Artinskian.

A different marine assemblage in Stag Creek 23km east of the Callide opencut mine, at GR279500 7312000 on the Biloela 1:100 000 Sheet area, is believed to occur towards the top of the unit. This fauna has been described by Beeston (1975), who recorded *Montospira youlambiensis*, *Pyramus laevis*, *Stutchburia compressa*, *Myonia* sp., and crinoid stems. Correlation with the upper Lorray Formation (former Burnett Formation) is indicated by the presence of *Stutchburia compressa* in both units, and the similarity of *Montospira youlambiensis* with *Montospira montoensis* which is restricted to the Burnett Formation. Therefore the Stag Creek fauna is assigned the same age as those from the Spring Creek Syncline.

Malone & others (1969, page 18) recorded a poorly preserved brachiopod fauna of probable early Permian age from the top of the Youlambie Conglomerate just east of the eastern boundary of the Duaringa 1:250 000 Sheet area. This appears to be in the same stratigraphic position as the marine fauna reported by Fielding & others (1997a) from the Youlambie Conglomerate at Scrub Creek 30km south-west of Ridgelands. This fauna is significantly younger than those from the Spring Creek Syncline and Stag Creek, being referable to the *Echinalosia warwicki* or lower *Echinalosia preovalis* zones of Briggs (1993, 1998) and therefore of Artinskian age. In this area, the upper part of the sequence mapped as Youlambie Conglomerate by Fielding & others (1997a, figure 6) has been included in the Rookwood Volcanics on the Ridgelands 1:100 000 geological map because of the presence of extensive interbedded pillow basalts that are unknown from the Youlambie Conglomerate. The fossil locality falls near the contact between the two units, and could come from either the top of the Youlambie Conglomerate or the base of the Rookwood Volcanics. The latter alternative is supported by the occurrence of Artinskian foraminifers in the Rookwood Volcanics to the west (O'Connell, 1995; Fielding & others, 1997a).

Although fossils in the Youlambie Conglomerate are confined to the early Permian, deposition may have begun at the end of the Carboniferous. An ignimbrite interbedded with the conglomerate as well as three ignimbrite clasts in conglomerate were collected from along the Dawson Highway north-east of Biloela and submitted for zircon dating by SHRIMP. All ages are in error of each other (Appendix 1). The ignimbrite (MHRO970 from GR262661 7319645) gave an age of 300.8±3.4Ma (MSWD 1.7). The clasts from site MHRO971 at GR258242 7318366 gave ages from 300.0±3.8Ma to 303.6±3.7Ma. These ages lie on the Carboniferous–Permian boundary in the ISC timescale (Gradstein & others, 2004)

Stratigraphic relationships

The Youlambie Conglomerate unconformably overlies a number of older units, including the Marble Waterhole beds and Lochenbar Formation east of Biloela, the Mount Hoopbound and Balaclava Formations in the Don River area south of Mount Morgan, the Three Moon Conglomerate in the Callide – Dawson Highway area, and the Rockhampton Group at several localities from the Spring Creek Syncline near Monto to west of Mount Morgan. These underlying units range in age from Middle Devonian to Early Carboniferous, indicating a substantial interval of non-deposition or erosion (Yarrol Project Team, 1997). Some contacts show moderate to strong angular discordance, and at others the units appear to be concordant. The variation in relationships at this basal contact can be attributed to different degrees of tilting of basement blocks during latest Carboniferous to early Permian extension. This regional event changed the paleogeography and formed extensional basins that received coarse clastic sediments and rhyolitic volcanics (Holcombe & others, 1997b; Fielding & others, 1997a). The sediments include widely distributed granite clasts that suggest accelerated erosion and unroofing of plutons in the hinterland, probably located to the west in the now inactive Connors-Auburn Arc (Hutton & others, 1999).

The Youlambie Conglomerate overlaps the Lorray Formation in age. East of Monto, these units are separated by as little as 10km, but have significant differences. The Lorray Formation in the Yarrol Syncline is part of a continuous marine sequence from the Early Carboniferous to the Early Permian, but the Youlambie Conglomerate on the western limb of the Spring Creek Syncline is mainly terrestrial, and directly overlies the early Carboniferous Rockhampton Group. In the Spring Creek Syncline, the Youlambie Conglomerate is overlain conformably by the Yarrol Formation. The same relationship exists at Bell Creek north of Biloela, but here the Yarrol Formation pinches out, and the overlying Smoky beds lie directly on the Youlambie Conglomerate to the north of Bell Creek. In the Kroombit Tops area east of Biloela, the Youlambie Conglomerate is unconformably overlain by the Triassic Winterbourne Volcanics.

North from Mount Morgan, the Lorray Formation (formerly the Neerkol Formation) is finer-grained than in the Yarrol Syncline, and the Youlambie Conglomerate is readily distinguished as a conglomeratic facies. In this area, the Youlambie Conglomerate is mapped as a discontinuous belt overlying the Lorray Formation from Mount Morgan to west of the Fitzroy River at Craigilee homestead. No marine fossils are known from this belt apart from the fauna reported by Fielding & others (1997a) from the top of the Youlambie Conglomerate or bottom of the Rookwood Volcanics at Scrub Creek. West and north of Mount Morgan, the Youlambie Conglomerate is conformably overlain by the Lower Permian Rookwood Volcanics. Fielding & others (1997a) suggest that outcrop patterns near Scrub Creek “demonstrate interfingering between the Youlambie Conglomerate and the Rookwood Volcanics”. Their figure 6 shows extensive belts of pillow basalts within the Youlambie Conglomerate. Pillow basalts are not known from the Youlambie Conglomerate, but are characteristic of the Rookwood Volcanics. Therefore, the upper section of the unit mapped as Youlambie Conglomerate by Fielding & others (1997a, figure 6), with pillow basalts, is included in the Rookwood Volcanics on the Ridgelands 1:100 000 geological map. In this same area, the Youlambie Conglomerate is unconformably overlain by the Cretaceous Mount Salmon Volcanics. In the Mount Morgan 1:100 000 Sheet area near Stanwell (GR229400 7399800), the Youlambie Conglomerate is disconformably or unconformably overlain by the Dinner Creek Conglomerate.

Correlation with other units

In the Monto area, the dominantly terrestrial Youlambie Conglomerate is a lateral equivalent of the upper part of the marine Lorray Formation. North from Mount Morgan, the Youlambie Conglomerate is mapped as a conglomeratic facies overlying the much finer-grained marine sediments of the Lorray Formation. The Youlambie Conglomerate probably correlates at least in part with the Mount Bulgi Conglomerate that overlies the Torsdale Volcanics at the base of the Camboon Volcanics. Rhyolitic ignimbrite and comagmatic granites in the Torsdale Volcanics probably contributed to the felsic detritus in the Youlambie Conglomerate. SHRIMP dating of ignimbrite from the top of Torsdale Volcanics near Biloela gave an age of 298.2±5.1Ma (Withnall & others, 2009), which is within error of the ages reported from the Youlambie Conglomerate. North of the Marlborough Block, the Glenprairie beds may be a correlative of the Youlambie Conglomerate.

MID-PERMIAN

Yarrol Formation (Pa)

(G.A. Simpson & S.B.S. Crouch)

Introduction

The Yarrol Formation was defined in the Yarrol Syncline east of Monto by Maxwell (1959). The unit was extended to the Spring Creek Syncline north of Monto by Dear (1968). Dear & others (1971) also applied the name to Permian strata overlying the Youlambie Conglomerate along Bell Creek north of Biloela. Descriptions of the Yarrol Formation are taken from Dear (1968) and Dear & others (1971).

Figure 110. Distribution of the Yarrol Formation

Distribution

The Yarrol Formation crops out on both limbs of the Yarrol Syncline and on the western limb of the Spring Creek Syncline, east and north of Monto respectively. A small, isolated outcrop occurs at Bell Creek north of Biloela. The outcrop north of Monto is the largest, and is 14km long and 1km wide (Figure 110).

Type section

Maxwell (1959, 1960b, 1964) nominated a type area in the vicinity of Yarrol homestead on the Burnett River on the eastern limb of the Yarrol Syncline.

Topographic expression

The Yarrol Formation forms strike ridges that stand out on aerial photographs, but was described as forming “moderately broken hills and ridges” by Dear & others (1971).

Geophysical expression

The Yarrol Formation has a low radiometric response in all three channels that contrasts very strongly with and clearly distinguishes it from the underlying units, the Lorray Formation and the Youlambie Conglomerate. It also has a low magnetic signature.

Lithology

In the Yarrol Syncline, the unit consists of over 500m of limestone, siltstone, and lithic sandstone. In the Spring Creek Syncline, limestones are restricted mostly to the basal 30m, and the formation consists of a monotonous sequence of siltstone, fine-grained sandstone, and greyish-green chert. The rocks along Bell Creek are hornfelsed by a small granodiorite intrusion.

The limestones are grey and detrital, containing abundant brachiopods, bryozoa, crinoid stems, pelecypods, and gastropods, and grade into richly fossiliferous coquinitic silty limestone and calcareous siltstone. The siltstones and fine-grained sandstones are brownish black, black, and dark grey, and are usually calcareous. Many are graded, but others are poorly bedded with angular clasts of siltstone in a lighter matrix. Interbedded medium-grained sandstones are poorly sorted and contain clasts of andesitic volcanics and plagioclase in a dark brown calcareous, tuffaceous matrix. Quartz is only a minor constituent of the medium-grained sandstones but is more abundant as fine sub-angular grains in some of the siltstones and fine-grained sandstones.

Andesitic lavas similar to those in the overlying Owl Gully Volcanics are interbedded towards the top of the Yarrol Formation in the Spring Creek area.

Environment of deposition

The presence of a wide variety of marine fossils, and the type and character of the sediments, suggest a shallow marine, shelfal environment.

Thickness

The unit is 515m thick in the type area (Maxwell, 1960b), and 480m in the Spring Creek Syncline (Dear, 1968).

Structure

The Yarrol Formation dips at moderate angles in both the Yarrol and Spring Creek Synclines. These structures are thought to have formed during the episodic Hunter–Bowen Orogeny, and to reflect the development of a fold-thrust belt in response to compression from the east. At Bell Creek, bedding dips shallowly to the north.

Biostratigraphy and age

Prolific brachiopod-bryozoan assemblages occur in the limestones and calcareous siltstones in the Yarrol and Spring Creek Synclines. The following species were listed by Dear & others (1971): *Lissochonetes yarrolensis* Maxwell, *Taeniothaerus subquadratus* var. *cracowensis* Hill, *Strophalosia preovalis* Maxwell, *Linoproductus* sp., *Cancrinella* cf. *farleyensis*, (Etheridge and Dun), *Anidanthus springsurensis* (Booker), *Megousia* sp., *Terrakea pollex* Hill, *Grantonia* cf. *hobartense* Brown, *Trigonotreta* sp., *Ingelarella ovata* Campbell, *Ingelarella profunda* Campbell, *Cancellospirifer* sp., *Psilocamara* sp., *Deltopecten limaeformis* Morris, *Stutchburia costata* (Morris), *Parallelodon* sp., *Myonia* sp., *Conocardium* sp., *Astartila* sp., *Uraloceras cancellatum* Dear, *Polypora* spp. and *Fenestella* spp.

The fauna is diagnostic of the *Echinalosia warwicki* zone of Briggs (1993, 1998), and is Artinskian (early Permian). The goniatite *Uraloceras cancellatum* is particularly useful in age determination and indicates a probable lower Artinskian (Aktastinian) age.

Stratigraphic relationships

The Yarrol Formation conformably overlies the Lorray Formation in the Yarrol Syncline, and the Youlambie Conglomerate in the Spring Creek Syncline. At both localities it is overlain conformably by the Owl Gully Volcanics, also of Early Permian age. An abrupt lithological change occurs at the base of the formation, but there is a transition upwards into the overlying Owl Gully Volcanics (Dear & others, 1971), an observation that is confirmed by airborne geophysical survey data. At Bell Creek, the unit overlies the Youlambie Conglomerate, and is overlain by the Smoky beds which are probable correlatives of the Owl Gully Volcanics.

Correlation with other units

The Yarrol Formation is correlated in age with the lower part of the Rookwood Volcanics, but rock types in these units are very different.

Figure 111. Distribution of the Owl Gully Volcanics

Owl Gully Volcanics (Pvo)

(G.A. Simpson & S.B.S. Crouch)

Introduction

The Lower Permian Owl Gully Volcanics, defined by Maxwell (1959), are characterised by basic to intermediate volcanic flows, although volcaniclastic sediments make up much of the sequence. This description is largely taken from Dear (1968) and Dear & others (1971).

Distribution

The Owl Gully Volcanics crop out in two areas east and north of Monto. They form the core and uppermost unit in both the Yarrol and Spring Creek Synclines. The latter outcrop is the most extensive, forming a wedge-shaped area 20km long and up to 5km wide tapering to the south (Figure 111).

Two additional areas included in the Owl Gully Volcanics by Dear & others (1971) have been assigned to other units. A north-south belt west of Many Peaks is now mapped as undivided Permian volcanics (**Pv**), although much of this area appears to be occupied by Triassic dykes, and an outcrop in the Bell Creek – Back Creek area north of Biloela has been assigned to a new unit, the Smoky beds, with a much greater extent.

Type section

The type area of the unit is located on the eastern limb of the Yarrol Syncline west of Yarrol homestead (Maxwell, 1959). The unit is also well exposed in Apple Tree Creek, Scrubby Mountain Creek, and Tulloch Gully, which cut the Spring Creek Syncline 30km north-north-west of Monto (Dear, 1968).

Figure 112. Basaltic andesite with augite microphenocrysts, from the type area near Yarrol homestead, Owl Gully Volcanics

Topographic expression

The unit forms rugged hilly topography with few strike ridges, and the best exposure is along creeks and gullies.

Geophysical expression

The radiometric response of the Owl Gully Volcanics is low for all three channels, similar to that of the underlying Yarrol Formation. The magnetic signature is weak to moderate, but is higher than that of the surrounding sedimentary rocks, consistent with its mafic to intermediate composition.

Lithology

The unit is reported by Dear & others (1971) to consist of "…andesitic flows and pyroclastics, interbedded siltstone and minor lithic sandstones."

Andesitic lava is the characteristic rock type. The lavas are black, greyish green, or dark green, and are commonly vesicular. Very few are purple or reddish brown where weathered. Plagioclase and augite phenocrysts are common (Figure 112). In thin section, they are porphyritic, holocrystalline fine-grained rocks composed of plagioclase, augite and rare olivine in a groundmass of plagioclase microlites and intergranular augite, opaque iron oxide, calcite and chlorite. In samples from the Spring Creek Syncline, augite microphenocrysts are completely altered to calcite, some with an opaque rim. Chlorite appears to replace a different ferromagnesian mineral, either orthopyroxene or hornblende. Vesicles are infilled by chlorite, epidote, zeolite, prehnite, and calcite. No felsic volcanics are known in the unit.

Interbedded siltstones are black or brownish black and, like those in the Yarrol Formation, are soft rocks which have an "ashy" appearance when weathered. Many are graded bedded, and others are poorly bedded and contain angular clasts of black siltstone in a lighter coloured siltstone matrix. The siltstones are calcareous when fresh. Associated lithic sandstones are commonly graded bedded and consist of altered feldspar grains, commonly euhedral, and subangular andesitic lithic fragments in a green or dark brown calcareous matrix.

The rocks observed by the Yarrol Project Team on the western side of the Yarrol Syncline consisted of strongly weathered green porphyry that, in hand specimen, appeared to be similar to rocks found in the Rookwood Volcanics in the Fitzroy area. Breccia and fossil-bearing siltstone and sandstone were also observed in the Yarrol Syncline area.

Geochemistry

Samples of andesite have been analysed from the type area at Yarrol and the Spring Creek Syncline north of Monto. These two areas have different geochemistry, but all samples have a large subduction component, with enrichment of large ion lithophile elements (LILE) relative to high field strength elements (HFSE). In particular, Nb is strongly depleted. The SiO_2-poor andesites at Yarrol have similar patterns to those of basalts and andesites from the southern part of the Southern Volcanic Zone of the Andes, and from the Marianas and Lesser Antilles island arcs (Figure 113). $(La/Yb)_N$ ratios are 2.94 and 5.46, and Ce/Y values suggest a crustal thickness of about 25km (Mantle & Collins, 2008).

The andesites from the Spring Creek Syncline have very flat REE patterns, with $(La/Yb)_N$ ratios of 1.4, 1.5 and 1.51. Despite their andesitic compositions and relatively high SiO_2 contents, the rocks are significantly depleted in HFSE, particularly Nb, Ta and Ti, consistent with active subduction. The geochemical patterns are similar to less-fractionated samples from modern island arcs such as the Lesser Antilles and Marianas, and from Kamchatka (Figure 114). The Ce/Y ratios indicate that the crust was thin (15km or less) when the Owl Gully Volcanics accumulated in this area (Mantle & Collins, 2008).

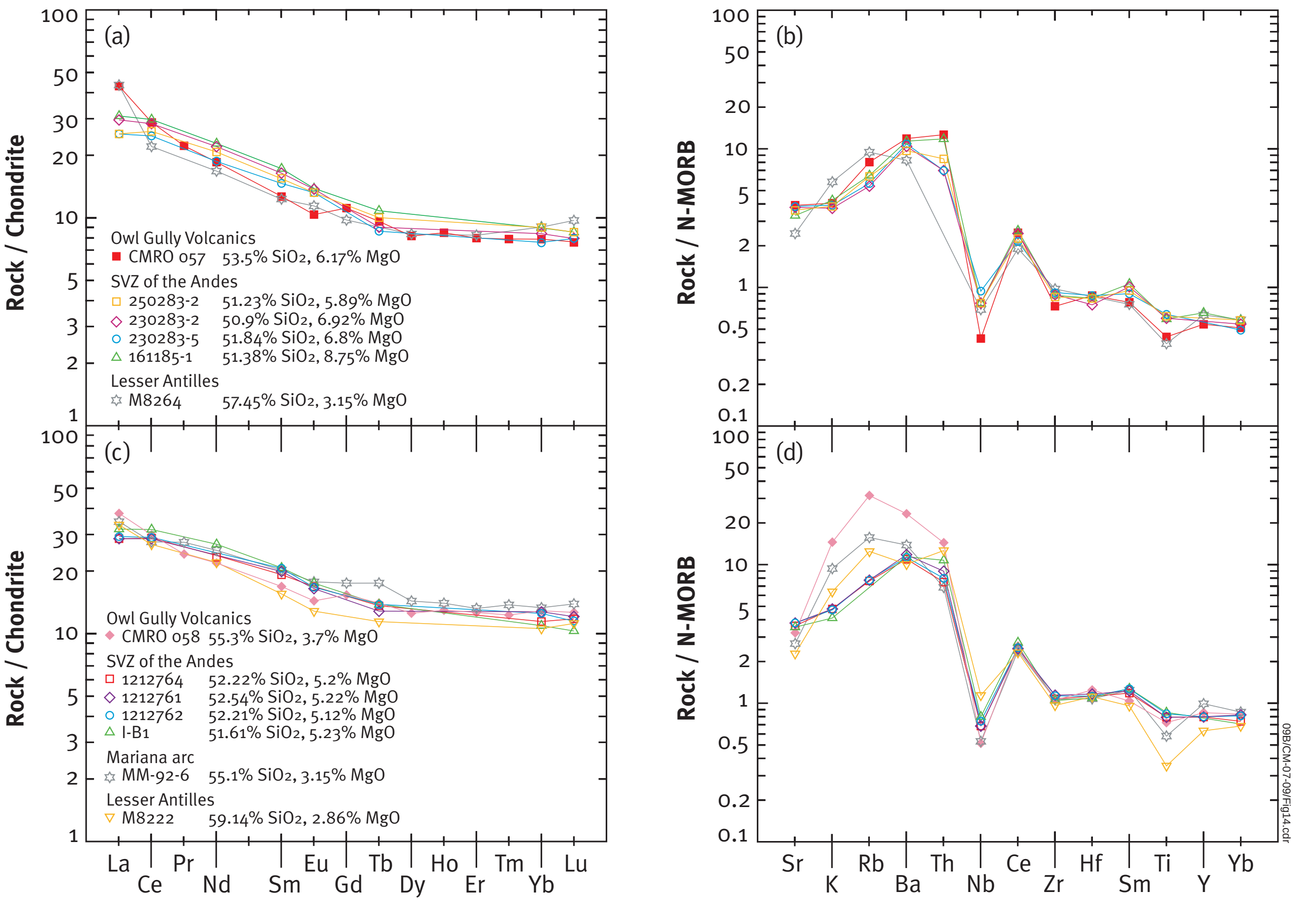


Figure 113. (a,c) REE and (b,d) spidergram plots comparing from the type area with basalts, basaltic andesites and andesites from Llaima, Puyehue and Reloncavi volcanoes in the southern part of the Southern Volcanic Zone of the Andes (Hickey & others, 1986; Tormey & others, 1991; Lopez-Escobar & others, 1995), the Lesser Antilles (Davidson, 1986; Thirlwall & others, 1994; Turner & others, 1996), and the Mariana arc (Elliott & others, 1997). Chondritic REE values are from Sun & McDonough (1989), and N-MORB from Pearce (1983).

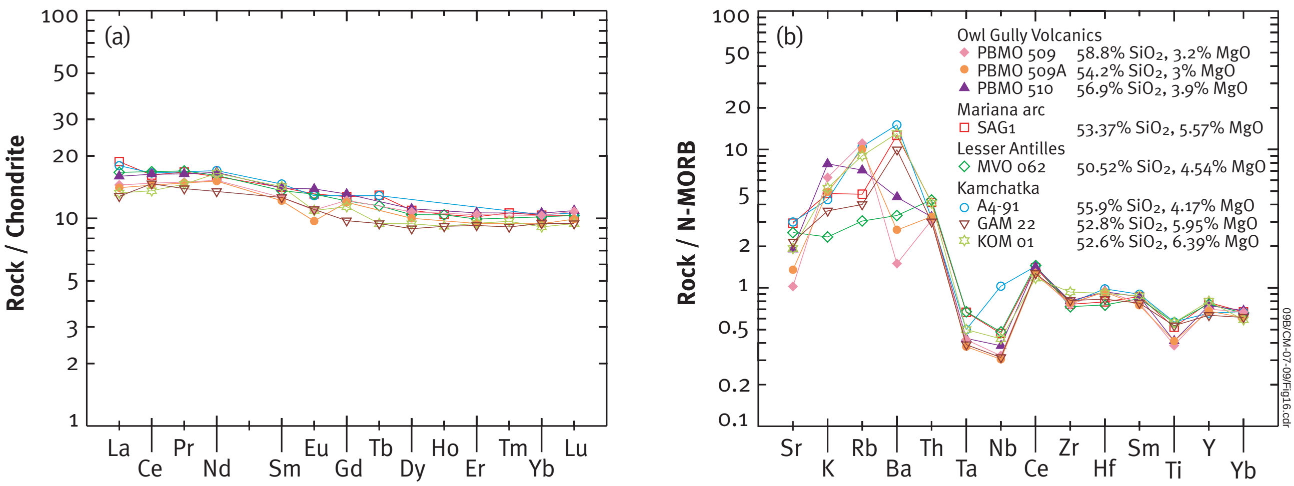


Figure 114. (a) REE and (b) spidergram plots comparing andesites from the Spring Creek Syncline north of Monto with basalts and basaltic andesites from the Mariana arc (Elliott & others, 1997), the Lesser Antilles (Zellmer & others, 2003), and Kamchatka (Turner & others, 1998; Churikova & others, 2001). Chondritic REE values are from Sun & McDonough (1989), and N-MORB from Pearce (1983).

Environment of deposition

The presence of marine macrofossils in a succession of lavas and volcaniclastic rocks indicates that the Owl Gully Volcanics were deposited in a marine environment influenced by volcanic activity.

Provenance

The Owl Gully Volcanics form part of a suite of Early Permian rock units that have a distinctly lower radiometric response than older units. This suite includes the Rookwood Volcanics and parts of the Camboon Volcanics. This change in response seems to reflect an overall change in the provenance from a mixed felsic volcanic and granitoid provenance in the late Carboniferous to earliest Permian to a basic to intermediate source in the Early Permian.

In contrast, the Lower Permian Berserker Group is a time equivalent of the Camboon Volcanics and Owl Gully Volcanics, but is composed of sediments and volcanics whose provenance represents predominantly felsic rock-types, suggesting they are sourced from a different area compared to those contemporaneous units found further west.

Thickness

Dear (1968) reported a thickness of 1500m in the type area, and somewhat more than that in the Spring Creek Syncline, but the top of the unit is not exposed at either locality, and is either erosional or faulted.

Structure

The unit forms the core of both the Yarrol and Spring Creek Synclines, and is gently to moderately folded with shallow dips. Dear (1968) reported average dips of 30° on both limbs of the Spring Creek Syncline.

Biostratigraphy and age

Dear & others (1971) recorded that marine fossils are rare and known from only two localities in the Spring Creek Syncline. *Atomodesma* sp., crinoid stems, and a straight nautiloid occur in siltstone in Apple Tree Creek, and *Astartila* sp. is found in siltstone 5km north of Scrubby Mountain. These faunas indicate a Permian age. Based on this evidence and the stratigraphic relationships, Dear & others (1971) proposed that the Owl Gully Volcanics were deposited during the Artinskian, possibly ranging upward into the Kungurian. This age was supported by Briggs (1998), who placed the Owl Gully Volcanics in his *Echinalosia preovalis* zone of Artinskian (early to mid-Permian) age.

Stratigraphic relationships

The unit conformably overlies the Yarrol Formation in the Yarrol and Spring Creek Synclines. Tertiary residual deposits unconformably overlie the unit.

Correlation with other units

The Owl Gully Volcanics may be partial time equivalents of the Rookwood Volcanics, Narayen beds, Nogo beds, and Buffel Formation based on recent stratigraphic revisions by Hutton & others (1999). The Narayen beds include rhyolitic to dominantly andesitic breccia and lava flows (Hutton & others, 1999). The Rookwood Volcanics include mafic pillow lavas and high-level intrusives with minor dacitic to rhyolitic lava (Yarrol Project Team, 1997). The Owl Gully Volcanics are also time equivalents of the Berserker Group, but the latter contain predominantly felsic volcanics, and sediments derived from felsic to intermediate rocks (Crouch & Parfrey, 1998).

Unnamed Permian (Pu^c)

(C.G. Murray)

Introduction

Permian rocks along the western side of the Boyne River valley were first recorded by Dear & others (1971), although workings associated with the Bompa silver-lead mines had been inspected as early as 1900.

Distribution

The Permian rocks form a north-south belt 16km long and up to ~3km wide, extending almost from Ubobo in the north to south-west of Many Peaks (Figure 115).

Topographic expression

The rocks form rugged country on the eastern slope of Mount Robert, passing into lower ridges and valleys to the east.

Figure 115. Distribution of the unnamed Permian (Pu^c)

Geophysical expression

On radiometric images, the sediments give a strong signature, typical of rocks of this age. The topographic relief causes some variation.

The magnetic response is low, although the unit partly coincides with the strong magnetic ring around the Robert Granite. The sediments are bordered to the east by the moderate magnetic signal from mafic volcanic rocks.

Lithology

The main lithologies are grey sandstone and dark grey to black siltstone. The sandstone is medium- to coarse-grained, and thick to thin bedded. Feldspar and quartz dominate the medium-grained fraction, with volcanic lithic fragments (mainly intermediate or mafic) forming granule-sized grains. The siltstone is also thick to thin bedded, and contains fossil remains, largely crinoids and bryozoans.

Amygdaloidal basalt is interbedded with the sediments locally. It is fine-grained, with microphenocrysts of plagioclase in a glassy groundmass. Only the mineralogical constituents of the amygdales could be identified in the field, comprising chlorite, calcite and epidote.

Small calcareous bodies appear to be confined to the western part of the unit, in very rugged country adjacent to the contact with the Robert Granite. They have been metamorphosed, and now consist largely of calcsilicates intruded by monzonite dykes.

Environment of deposition

The unnamed sediments were deposited under marine conditions, shallow enough to allow the growth of marine fossils.

Thickness

In the absence of any data on the structure, the thickness must remain unknown, but it could be considerable.

Structure

The rocks generally have close to meridional trends, and steep to moderate dips. Dips are mostly to the east, but some readings to the west suggest that the sequence is folded, and not just tilted due to faulting. No mesoscopic folds were seen in outcrop.

Biostratigraphy and age

Dear & others (1971) recorded Permian fossils from the unit, and the siltstones contain bryozoans and crinoid stems.

Stratigraphic relationships

The unnamed Permian sediments are faulted against the uppermost Silurian to Lower Devonian Calliope beds and the Lower Carboniferous Rockhampton Group to the north-east. Their relationship to the unnamed volcanic unit (**Pv**) to the east is unknown. Dear & others (1971) regarded these volcanics as part of the Owl Gully Volcanics, and proposed that they overlie the sedimentary unit conformably. The unnamed Permian sediments are interpreted to be partly overlain or faulted against presumed Upper Triassic felsic to intermediate volcanic units (**Rvf** and **Rva**) to the south-west. To the west, the Permian rocks are intruded by the Upper Triassic Littlemore Granodiorite, Robert Granite, Rule Gabbro and Deception Quartz Monzonite.

Correlation with other units

On the basis of the fossils they collected, Dear & others (1971) suggested that the unit was partly equivalent to the Yarrol Formation.

Economic significance

Small silver-lead workings west of Bompa, in the headwaters of Sugarbag Creek, were described by Rands (1901), Marks (1911, 1912), and Saint-Smith (1914), but production was minimal.

Unnamed Permian volcanics (Pv)

(C.G.Murray)

Introduction

Dear & others (1971) mapped these volcanics and referred them to the Owl Gully Volcanics. They were interpreted to overlie unnamed Permian sedimentary rocks to the west conformably. While this relationship is possible, chemical analyses are typical of Triassic volcanics rather than the Owl Gully Volcanics, and the unit is simply categorised as an unnamed volcanic unit on the Calliope and Monto 1:100 000 sheet areas.

Distribution

The volcanics have been mapped as a thin discontinuous belt more than 16km long and ~1km wide extending from near Ubobo in the north southwards to south-west of Many Peaks (Figure 116).

Topographic expression

The unnamed volcanic unit forms prominent ridges, most of which trend north-south.

Geophysical expression

The volcanic rocks have a moderate geophysical response, both for radiometrics and magnetics. The only marked departure from this pattern is west of Littlemore, where two smaller outcrops have a strong magnetic response. The response is similar to that of unnamed Triassic volcanics to the west and south.

Lithology

The volcanics were described as andesites by Dear & others (1971). The rocks all have microphenocrysts or more rarely phenocrysts of plagioclase. Accompanying these are varying proportions of clinopyroxene and

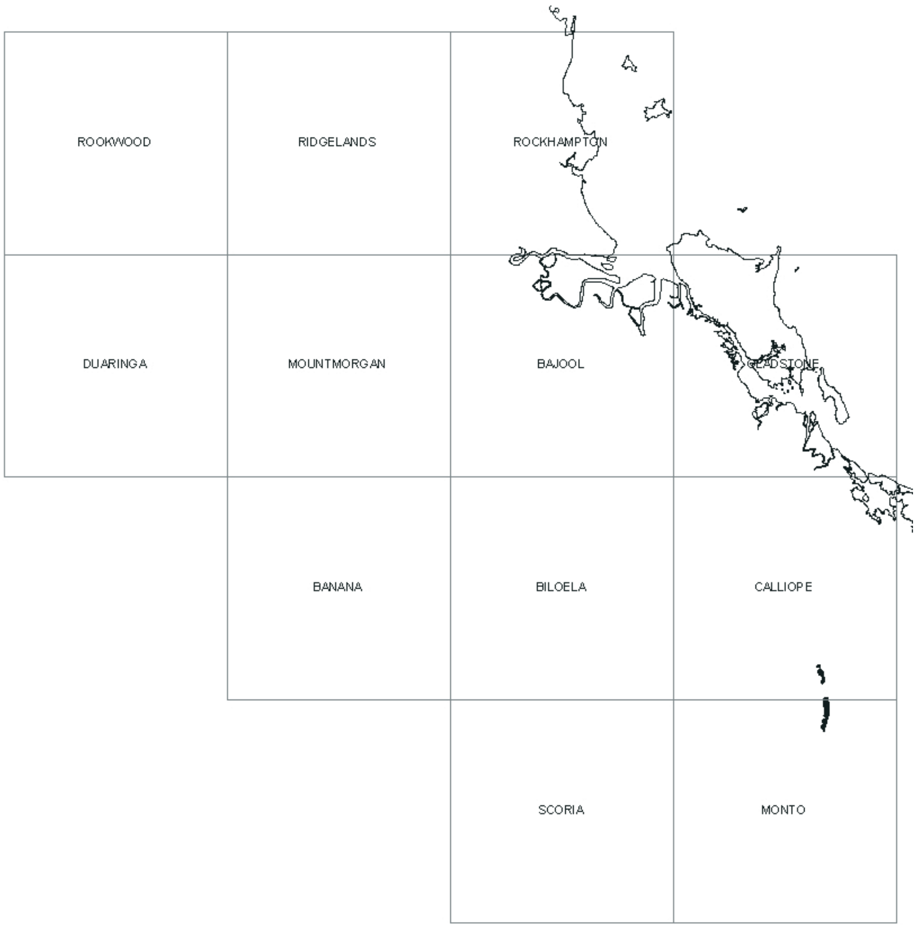


Figure 116. Distribution of the unnamed Permian volcanics (Pv)

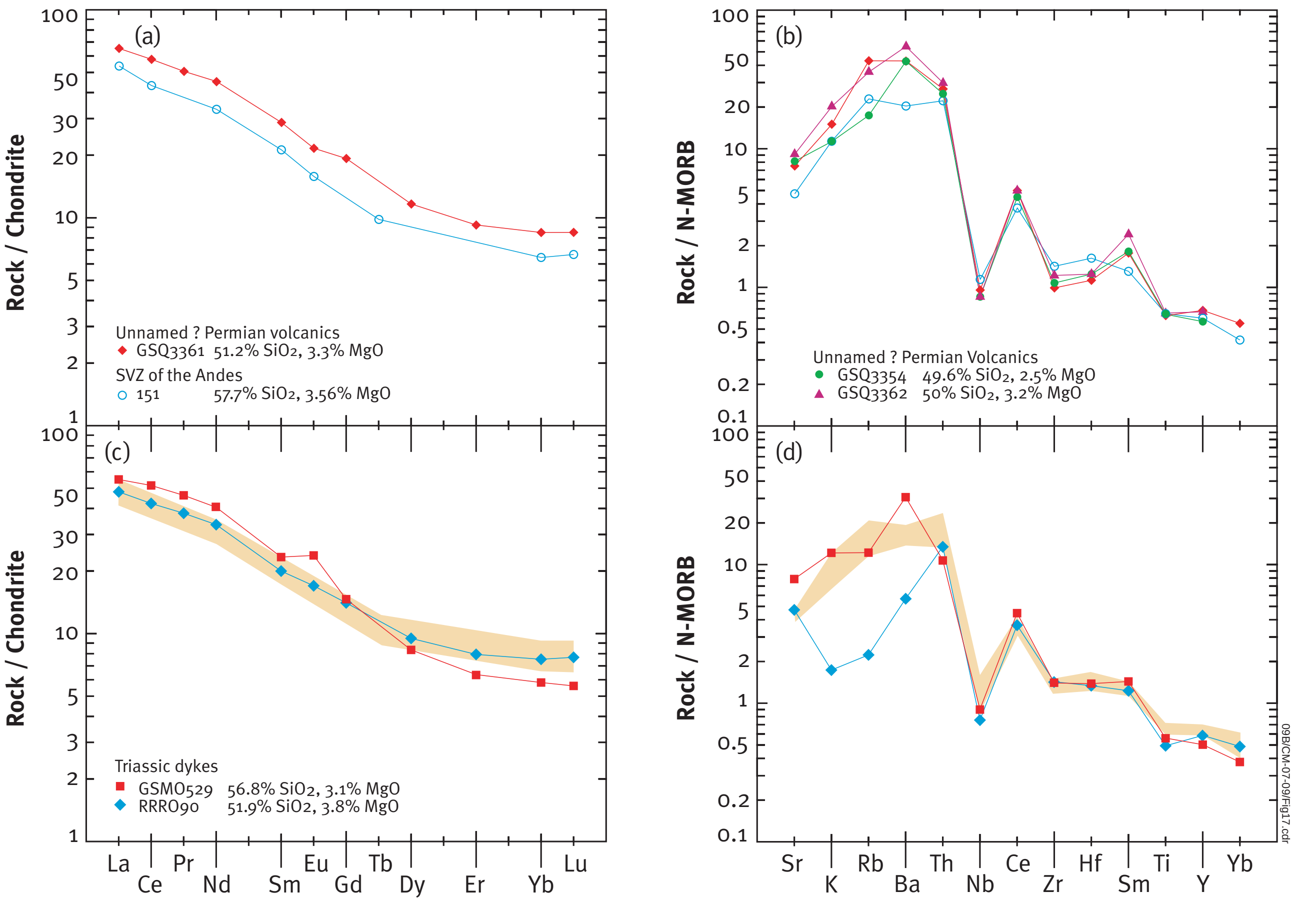


Figure 117. (a) REE and (b) spidergram plots comparing basalts from the unnamed Permian volcanics with andesite from Tatara volcano in the Southern Volcanic Zone of the Andes (Ferguson & others, 1992). (c) REE and (d) spidergram plots comparing Triassic basalt and andesite dykes with 10 basaltic andesites and andesites from Planchon, Azufre and Tatara volcanoes in the Southern Volcanic Zone of the Andes shown as a shaded pattern (Ferguson & others, 1992; Tormey & others, 1995). Chondritic REE values are from Sun & McDonough (1989), and N-MORB from Pearce (1983).

olivine microphenocrysts, altered to different degrees. The groundmass consists of plagioclase laths, clinopyroxene (in some cases altered to chlorite), and opaque grains. Amygdales are present in most samples, and consist of chlorite ± calcite ± epidote, in some cases zoned.

Geochemistry

Chemical analyses show that the rocks are basalts, and are most like Triassic volcanics, particularly with regard to likely crustal thickness. The $(La/Yb)_N$ ratio for one sample analysed by ICP-MS is 7.69. This is similar to ratios of 6.36–7.23 for the Muncon Volcanics, and much higher than values for known Permian volcanics such as the Owl Gully Volcanics (1.4–5.46, averaging 2.56). The three analysed samples have similar trace and REE patterns to andesite from the northern part of the Southern Volcanic Zone of the Andes, where the crustal thickness is ~45–50km. Ce/Y values indicate a thinner crust of about 40km (Mantle & Collins, 2008). The geochemical patterns are similar to those of Triassic dykes from the Yarrol Province which also can be compared to analyses from the northern part of the Southern Volcanic Zone of the Andes (Figure 117). On standard discriminant diagrams, the analyses fall in continental arc or volcanic arc fields (Figure 118).

Environment of deposition

The volcanics appear to have been erupted subaerially.

Thickness

The thickness is unknown, and depends on the attitude of the volcanics. The maximum topographic relief in the area is ~260m, so this would be a minimum thickness assuming the volcanics are flat lying.

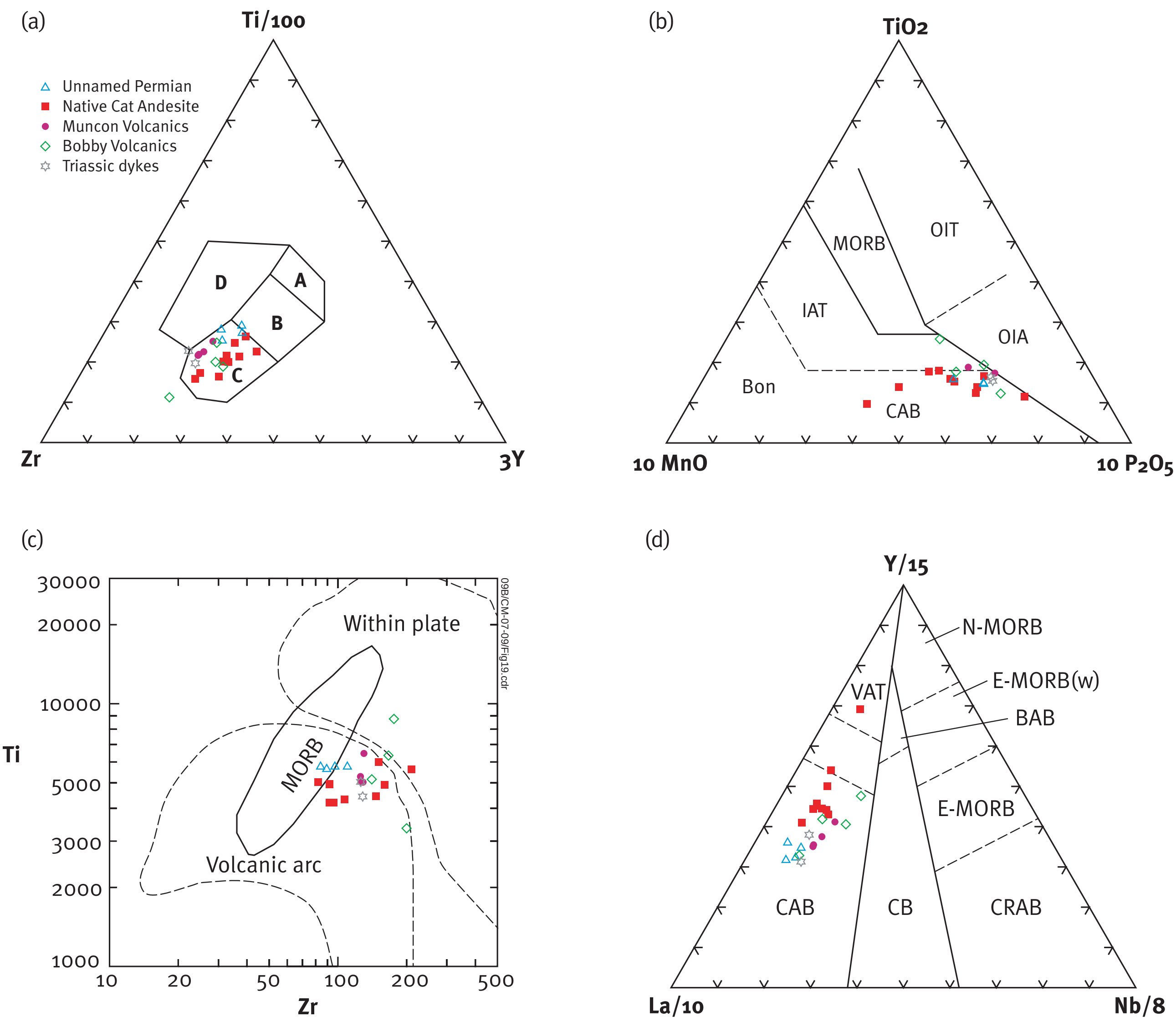


Figure 118. Analyses of basalts and andesites from the unnamed ?Permian volcanics, Triassic dykes, Native Cat Andesite, Muncon Volcanics, and Bobby Volcanics plotted on discriminant diagrams of (a) Pearce & Cann (1973): A = island arc tholeiites, B = island arc tholeiites, volcanic arc basalts and mid-ocean ridge basalts or MORB, C = calc-alkali basalts, and D = within plate basalts; (b) Mullen (1983): IAT = island arc tholeiite, CAB = island arc calc-alkaline basalt; (c) Pearce (1982); and (d) Cabanis & Lecolle (1989): CAB = calc-alkali basalts, VAT = volcanic arc tholeiites (overlap between these fields is indicated by the dashed lines), CB = continental basalts, CRAB = alkali basalts from continental rifts. The analyses are essentially confined to the calc-alkali and volcanic arc basalt fields on these diagrams.

Structure

The structure of the volcanics is unknown. Dear & others (1971) showed a single moderate to steep dip to the east (60°), similar to the sedimentary beds to the west, but this may represent the trend of a dyke.

Age

The unnamed volcanic unit has been assumed to be of Permian age, equivalent to the Owl Gully Volcanics (Dear & others, 1971). However, the geochemistry and geophysics is most like that of Triassic volcanic rocks.

Stratigraphic relationships

The unnamed volcanics are shown in contact with Permian sediments to the west, and almost certainly overlie them, but whether conformably or unconformably is not known. They are faulted against the uppermost Silurian to Lower Devonian Calliope beds and the Lower Carboniferous Rockhampton Group to the east, and against unnamed Triassic volcanics to the south. The unit is intruded by the Upper Triassic Littlemore Granodiorite in the north, and by a Tertiary basalt plug south of Station Creek.

Correlation with other units

The chemistry and geophysics suggests that the unit is of Triassic age, similar to unnamed volcanics to the south and in the Bobby Range.

Smoky beds (Ps)

(C.G. Murray)

Introduction

Dear & others (1971) mapped a large east-west trending mafic intrusion, the Mount Gerard Complex, centred 35km north of Biloela. Intrusive rocks of the Craiglands Quartz Monzodiorite make up only a small part of this area, and are more felsic than the diorite and gabbro described by Dear & others (1971). Most of the central part of what was mapped as the Mount Gerard Complex is in fact a sequence of volcaniclastic sedimentary and volcanic rocks here defined and described as the Smoky beds. This unit also includes rocks mapped as Owl Gully Volcanics by Dear & others (1971). The volcaniclastic sequence forms a thin roof which partly covers and is intruded by the Craiglands Quartz Monzodiorite.

Distribution

The Smoky beds form an ovoid outcrop area 13km long and 8km across, elongated in a north-westerly direction, and centred ~30km north of Biloela, north from Bell Creek (Figure 119).

Name

The name is from Smoky Creek which flows from east to west through the central part of the unit.

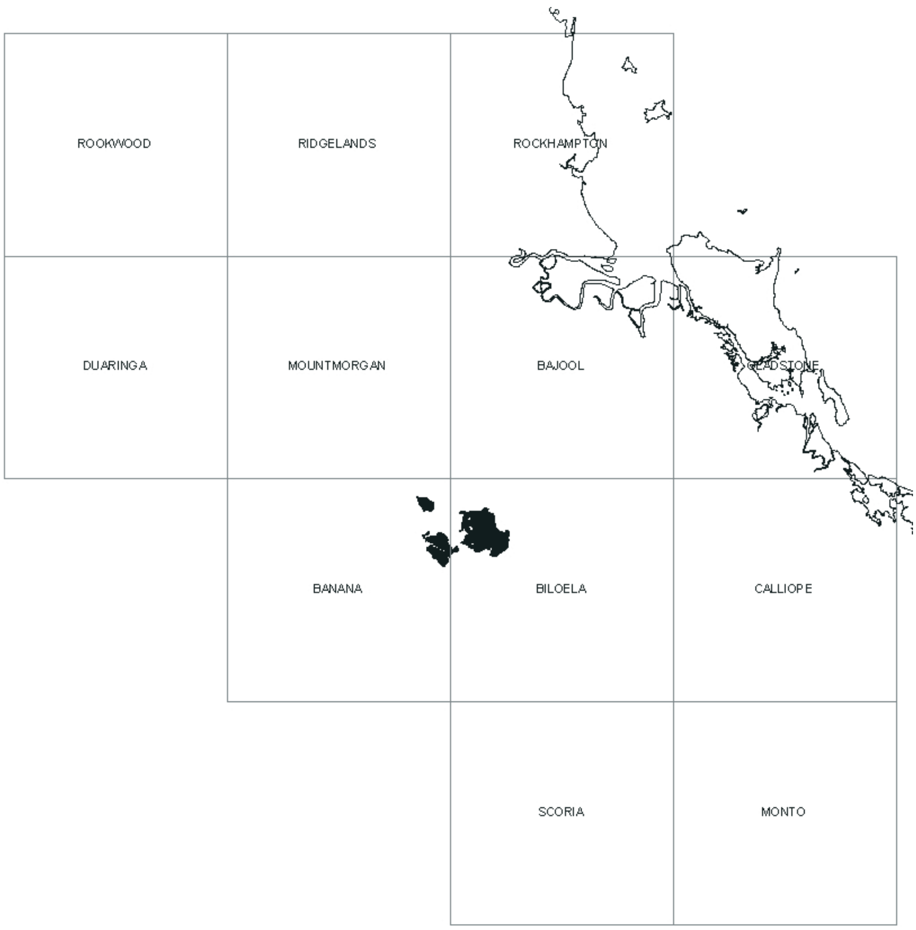


Figure 119. Distribution of the Smoky beds

Type section

A type section is nominated along Smoky Creek from GR251000E, 7331500N to GR255900E, 7333100N.

Topographic expression

The unit forms hilly topography with some development of benches and flat-topped hills.

Geophysical expression

The Smoky beds produce dark red tones on images derived from airborne radiometric data, suggesting an intermediate, relatively potassium-rich composition. The magnetic response appears to be very low, consistent with low magnetic susceptibility readings on outcrops, but is overwhelmed by the strong signature from the Craiglands Quartz Monzodiorite that underlies much of the Smoky beds.

Lithology

The unit is dominated by volcaniclastic conglomerate and breccia with clasts of andesitic to dacitic composition. The coarsest rocks contain clasts up to boulder size (Figure 120), with maximum grain size of 1m. Grey to green intermediate volcanics, both aphyric and with plagioclase phenocrysts, are the main component. Some samples are almost oligomictic, but polymictic conglomerates also occur, locally containing rare fragments of pink aplite or microgranite (Figure 121). Generally the degree of rounding increases with increasing grain size. Conglomerates are typically matrix supported, and the sandy matrix contains abundant feldspar. Secondary epidote and chlorite are common in both clasts and matrix. Medium and fine-grained sediments make up only a small proportion of the unit, and comprise khaki-brown feldspathic sandstone, siltstone and shale.

Andesitic rocks are concentrated in the south of the unit near Bell Creek, but occur sporadically in other areas as well. Lithologically they are similar to clasts in conglomerate and breccia, and locally contain pyrite associated with patchy alteration close to intrusive contacts with the Craiglands Quartz Monzodiorite. Many of these andesitic outcrops can be seen to be intrusions, but some form benches and terraces consistent with flat-lying flows or tuffs, and the abundance of volcanic clasts in the sequence clearly indicates that flows

Figure 120. Oligomictic boulder conglomerate consisting of sub-angular to rounded andesite clasts in a matrix of lithic fragments and plagioclase, 4.5km north of Craiglands homestead, Smoky beds

Figure 121. Poorly sorted pebble to cobble conglomerate with clasts mainly of andesitic composition, but including felsic volcanics, 5km north of Craiglands homestead, Smoky beds.

must have been present in the area. Some larger andesitic intrusions have been brecciated. The andesites typically contain abundant small plagioclase phenocrysts and lesser clinopyroxene or altered hornblende in an altered, fine-grained groundmass with plagioclase microlites.

In the east, near the Last Chance mine, the Smoky beds consist of grey volcanicastic breccia with subangular clasts of andesitic to dacitic composition.

Environment of deposition

The Smoky beds appear to have been deposited in an entirely continental environment. The dominantly volcanic composition, coarse-grain size, generally poor sorting and angular nature of clasts suggests rapid deposition close to an active volcanic centre, with only minimal reworking by fluvial processes.

Provenance

The unit was mainly derived from an intermediate volcanic source, represented by intrusive and probable extrusive andesitic rocks within the outcrop area.

Thickness

Assuming that the unit is essentially flat-lying, the thickness is at least 150m, but the top of the unit has been removed by erosion.

Structure

Outcrops are massive, and the only indicators of bedding attitudes are benches along some hillslopes. These suggest that bedding is essentially flat-lying or only gently dipping, and the sequence relatively undeformed.

Biostratigraphy and age

No fossils have been found in the Smoky beds. Its stratigraphic position above the Youlambie Conglomerate and Yarrol Formation suggests an Early Permian (Artinskian) age.

Stratigraphic relationships

Along their southern contact, the Smoky beds overlie the Yarrol Formation with apparent conformity. Elsewhere, the Yarrol Formation is absent, and the Smoky beds lie directly on the Youlambie Conglomerate.

The Smoky beds are intruded by the Late Permian Craiglands Quartz Monzodiorite and by an unnamed Permian to Triassic granodiorite straddling Bell Creek. The Inverness Volcanics are considered to be co-magmatic with the Craiglands Quartz Monzodiorite, and are therefore assumed to overlie the Smoky beds, although the nature of the contact cannot be determined in the field. A small outlier of Tertiary basalt overlies the eastern extremity of the Smoky beds.

Correlation with other units

The Smoky beds are similar to the Owl Gully Volcanics in their stratigraphic position and intermediate volcanic provenance. The southern part of the unit, where primary andesitic intrusives and possible flows are most abundant, was mapped as Owl Gully Volcanics by Dear & others (1971). No chemical analyses have been carried out on either volcanic clasts or primary magmatic rocks from the Smoky beds for comparison with the Owl Gully Volcanics.

Economic significance

A number of small gold deposits, including the Last Chance mine, occur in the eastern part of the unit, and anomalous gold values have been reported from stream sediment samples.

Rookwood Volcanics (Pvr)

(I.W. Withnall, R.M. Barker, J. Domagala, & S.B.S. Crouch)

Introduction

The lower to mid-Permian Rookwood Volcanics consist predominantly of basalt and high-level mafic intrusives, with minor dacitic to rhyolitic lava, volcaniclastic breccia and sandstone, siltstone, and mudstone. The unit hosts the Cu–(Zn) volcanic-associated massive sulphide deposit at Develin Creek.

The Rookwood Volcanics were first defined by Malone & others (1969) for spilitic pillow lavas in the Duaringa 1:250 000 Sheet area, with subsequent work on the Rockhampton and Monto 1:250 000 Sheet areas presented in Kirkegaard & others (1970), and Dear & others (1971) respectively. Details of the Rookwood Volcanics are presented in Clare (1993), Messenger (1994b), Fielding & others (1994, 1997a), O'Connell (1995), Bendall (1996), Holcombe & others (1997b), Withnall & others (2009), and the Yarrol Project Team (1997).

Distribution

The lower to mid-Permian Rookwood Volcanics occur towards the north-west margin of the project area as a number of discrete blocks. The main outcrop areas are in the Rookwood–Ohio area (where a type area was designated in Melaleuca Creek), Glenroy – Scrub Creek area (straddling the Fitzroy River) in the Rookwood, Duaringa and Ridgelands 1:100 000 Sheet areas, Westwood–Gogango in the Mount Morgan 1:100 000 Sheet area, and Lakeview near Goovigen in the Banana 1:100 000 Sheet area (Figure 122). Fielding & others (1994) proposed an extent of 650km^2 for the unit.

Derivation of name

The unit name was derived from the Rookwood homestead in the Rookwood 1:100 000 Sheet area (Malone & others, 1969).

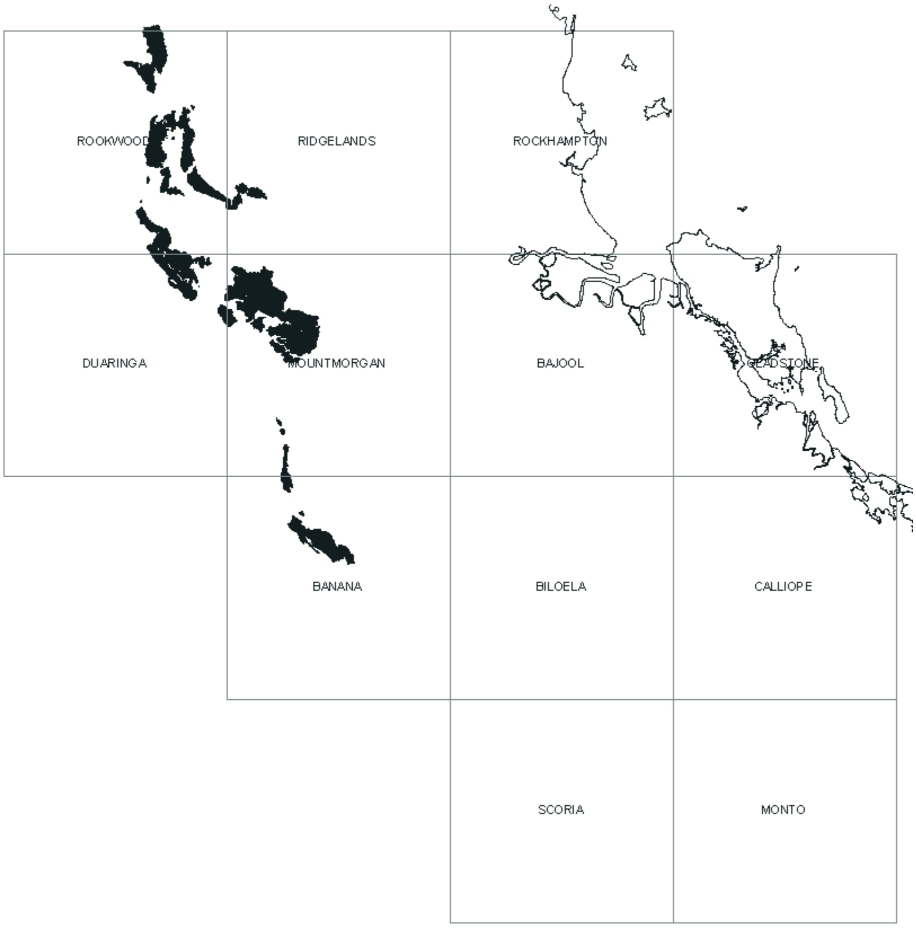


Figure 122. Distribution of the Rookwood Volcanics

Type section

No type section has been measured for the Rookwood Volcanics, but the type area is along Melaleuca Creek (Malone & others, 1969). It consists mainly of spilitic lavas, with minor agglomerate, volcanic breccia, and chert; tuffaceous sandstone and siltstone are also present.

Topographic expression

The volcanics are resistant to erosion and are well exposed (Malone & others, 1969), and form a low to moderate topographic expression.

Geophysical expression

The Rookwood Volcanics generally give a low response for potassium, thorium, and uranium. The magnetic response is low to moderate.

Lithology

The unit consists dominantly of basalt and high-level mafic intrusives, with minor dacitic to rhyolitic lava, volcaniclastic breccias, sandstones, siltstones, and mudstones. The basalt ranges from porphyritic to aphyric. Pillows, autobreccias, and peperites are relatively common in the northern blocks whereas the southern exposures are dominantly massive. Geochemical analyses indicate that the basalt has MORB-like affinities (O'Connell, 1995; Stephens & others, 1996a).

The lavas are mostly aphyric, but porphyritic basalt crops out in places. Phenocrysts are mainly plagioclase up to 7mm and minor augite. Where least altered, the aphyric basalt usually consists of randomly orientated laths of plagioclase to 0.25mm, with interstitial augite and skeletal opaque grains. The augite is commonly titaniferous, and in some samples, it forms aggregates of needle-like crystals, possibly a quench texture. However, alteration is generally extensive, the plagioclase being partly saussuritised, the augite replaced by chlorite and actinolite, and the opaque grains replaced by sphene.

Figure 123. Basalt peperite in the Fitzroy River east of Glenroy homestead, Rookwood Volcanics

At Develin Creek, Bendall (1996) stated that the Rookwood Volcanics are represented by pillow and massive facies of basalts, and that the basalt has undergone lower greenschist metamorphism. He observed no sedimentary rocks during the study, although Messenger (1994b) stated that they constitute <1% of the unit. Messenger (1994b) described the massive basalts as aphyric to porphyritic, with amygdales (<5–20%) in the former, which are filled with calcite, chlorite, hematite, jasper, and epidote. Veins of jasper are crosscut by veins of epidote and chlorite. Secondary calcite is also present in veins, fractures and cavities. Quartz veins exist at the contact with massive sulphide mineralisation. The pillow basalts dominate the sequence, and are similar mineralogically to the massive basalts. Bendall (1996) described the pillows as 0.2–1m in size, rarely 2m, with an altered plagioclase groundmass, and phenocrysts of plagioclase up to 2mm, commonly replaced by sericite, epidote, and chlorite. He observed a dark chloritised "rind" with a perlitic texture that surrounds the pillows and appears to flake off and be incorporated into the yellow-green, crystalline matrix. Veins of epidote, quartz, jasper, chlorite or calcite fill the tensional fractures of the pillows. Breccias of massive and pillow basalts are commonly found near the mineralisation and are characteristically monomict, and have jigsaw fit fractures. Messenger (1994b) also described a peperitic contact between the sedimentary units and the pillow basalts and hyaloclastites. In the Fitzroy River east of Glenroy homestead, basalt has intruded a mixture of comminuted basalt and mudstone to form peperite (Figure 123).

At Aeroview, O'Connell (1995) stated that the mafic units (pillow basalts and subordinate dolerite sills) constitute 65% to >95% of the stratigraphy, and that pillow basalts are a characteristic feature of the succession. Dolerite sills are prominent in the sedimentary packages. He described the pillow basalts as greenish grey, with textures ranging from aphanitic and aphyric, to feldspar-phyric, with medium to coarse feldspar phenocryts and minor augite. The pillows range from 0.1–1.5m, with green-grey inter-pillow material. The pillows are similar to those described in the Develin Creek area, although amygdales of quartz, calcite or epidote are minor components. The dolerite sills are hard, dark green to black, equigranular to porphyritic, and are composed of plagioclase and pyroxene.

The rocks examined in the Westwood, Lakeview and Goovigen areas are similar to those farther north, and consist mainly of altered dark greenish-grey aphyric basalt, although some porphyritic basalt or andesite, containing greenish epidotised plagioclase phenocrysts, occurs locally. Alteration is common, and most rocks are altered to assemblages of epidote, amphibole and chlorite. Epidotisation is commonly extensive, and is both pervasive and present as networks of veinlets. Local spectacular pillows and hyaloclastic breccias confirm a subaqueous emplacement (Figures 124 and 125). The pillows are ovoid and range from 15cm to 2m across. Interstices are generally filled with epidote and silica (chert), but in some cases, hyaloclastic material that has a silica-epidote matrix is present. The hyaloclastite consists of clasts of basalt up to 10cm with a

Figure 124. Shear zone in closely packed pillow basalts, Rookwood Volcanics

Figure 125. Close-up of pillow basalts, Rookwood Volcanics

Figure 126. Basalt peperite south of Westwood, Rookwood Volcanics

weak jigsaw fit, in a matrix of finer fragments (<5mm) and cherty material. Radial fractures in some pillows are also filled with epidote and silica. Amygdales are not common. Peperites are also present (Figure 126). According to Clare (1993), the upper part of the section in the Goovigen area is dominated by sheet flows that have brecciated or chilled tops. No sedimentary rocks were observed intercalated with the lavas in either area. Dear & others (1971) stated that the rocks in the Goovigen area are intruded by dolerite. Some possible dolerite outcrops were observed in the Lakeview area, but except for a slightly coarser-grainsize, they are difficult to distinguish from the basalt.

Slices of massive rhyolite occur with pillow basalt near Comanche homestead, and belts of felsic rocks also occur in the Scrub Creek area and near Westwood, particularly in a large outcrop centred 8km to the south (Figure 127).

Sedimentary rocks are more common in the north and locally form intervals up to 300m thick. They include volcaniclastic breccias, sandstones and siltstones, and thin-bedded to laminated carbonaceous mudstones, locally containing trace fossils (Figure 128). They are interpreted as density current deposits that accumulated below storm wave-base. The carbonaceous mudstones, which are indicative of shelfal depths, appear to be hemipelagic and low concentration turbidite deposits. A detailed description of sediments in the Develin Creek-Aeroview area can be found in Messenger (1994b) and O’Connell (1995).

Geochemistry

Geochemical analyses of the basalts by Messenger (1994b), O’Connell (1995), Bruce (1999), and Withnall & others (2009) indicate MORB-like affinities (Stephens & others, 1996a) (Figure 129). It is difficult to establish what the character of the crust was at time of formation, although current evidence suggests that it was continental. The compositions suggest that the crust must have been thin, and that eruption took place in an extensional setting. However, this has to be reconciled with the large quantity of basaltic lava extruded without evidence for crustal attenuation or the presence of deeply tapping faults. The lack of sedimentary fill within the Rookwood Volcanics suggests no substantial rift-related subsidence in the area during formation.

Figure 127. Hydrothermally altered rhyolite breccia south of Westwood, Rookwood Volcanics

Figure 128. Sequence of medium bedded sandstone grading upward into mudstone couplets with pencil cleavage, 25km south-south-west of Marlborough, Rookwood Volcanics.

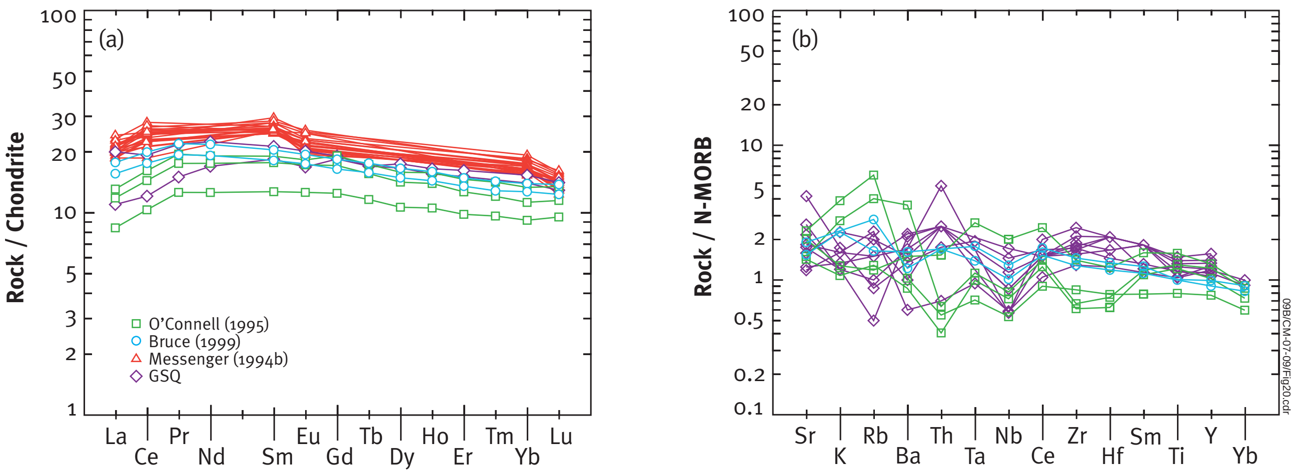


Figure 129. (a) REE and (b) spidergram plots of basalts. Analyses from Messenger (1994b), O'Connell (1995), Bruce (1999), and Withnall & others (2009). Chondritic REE values are from Sun & McDonough (1989), and N-MORB from Pearce (1983).

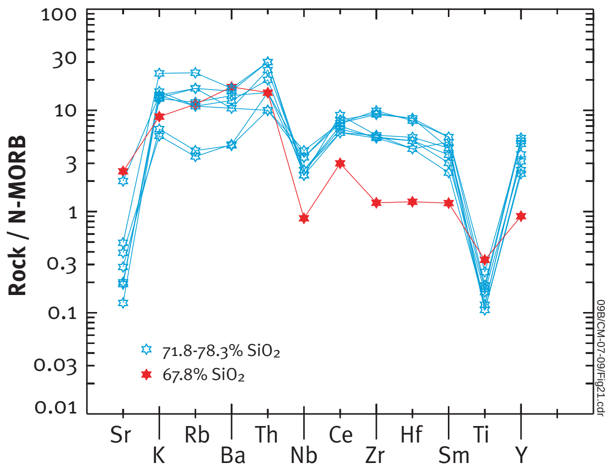


Figure 130. Spidergram plot of rhyolites. N-MORB values are from Pearce (1983)

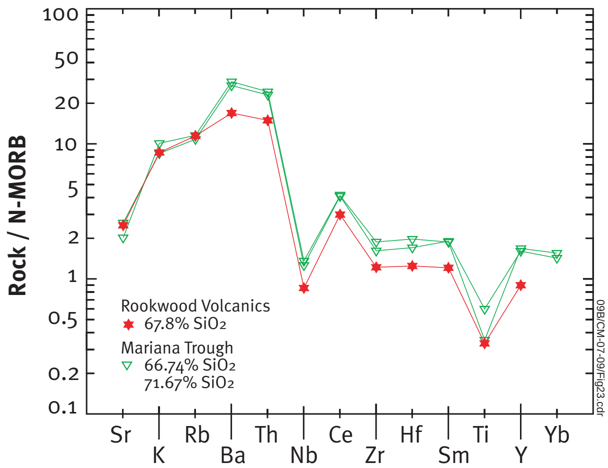


Figure 131. Spidergram plot comparing dacite from the Rookwood Volcanics with dacite and rhyolite from the northern Mariana Trough (Gribble & others, 1998; J. Pearce and R. Stern, unpublished data). N-MORB values are from Pearce (1983).

Several samples from the relatively large area of felsic volcanics south of Westwood have been analysed. They do not show typical subduction-related geochemical signatures, in that large ion lithophile elements (LILE) are not markedly enriched relative to high field strength elements (HFSE). This is reinforced by the small value of $(Th/Nb)_N$, which (excluding two obviously altered rocks) ranges from 2.5–17.5 and averages 8.5 (Figure 130). This indicates a significantly lower subduction component than in other mid-Permian volcanics (Owl Gully Volcanics and Berserker Group), or in felsic Triassic volcanics of the Native Cat Andesite. A sample of dacite from the felsic volcanics has similar geochemistry to dacite and rhyolite from the northern Mariana Trough (Gribble & others, 1998) (Figure 131), suggesting that it may be a differentiate of basaltic magma. The central felsic phase of the Kyle Mohr Igneous Complex has similar chemistry to the rhyolites (Figure 132), and appears to be comagmatic with them. This is clearly indicated on the granite discrimination plots of Pearce & others (1984).

In the Y versus Nb discriminant diagram of Pearce & others (1984), the rhyolites of the Rookwood Volcanics plot mainly in the Ocean Ridge Granite field, consistent with the basalt geochemistry. Analyses of granite from the Kyle Mohr Igneous Complex fall within the trend of the rhyolites, as do two granite samples from the southern margin of an unnamed gabbroic pluton (**PRg$_b$**) intruding the Rookwood Volcanics 7km south-west of Westwood (Figure 133a). In the Y+Nb versus Rb plot, unaltered volcanics form a group including the granites of the Kyle Mohr Igneous Complex, and fall mainly in the Within Plate Granite field, indicating that some continental crust was probably present (Figure 133b). The geochemical similarity of the felsic rocks is supported by a Ti/100 versus V plot (Shervais, 1982) comparing the chemistry of a gabbro sample from the mafic rim of the Kyle Mohr Igneous Complex with basalts of the Rookwood Volcanics. These plot together in the back-arc basin field (Figure 134a), unlike the Permo-Triassic gabbros which are clearly

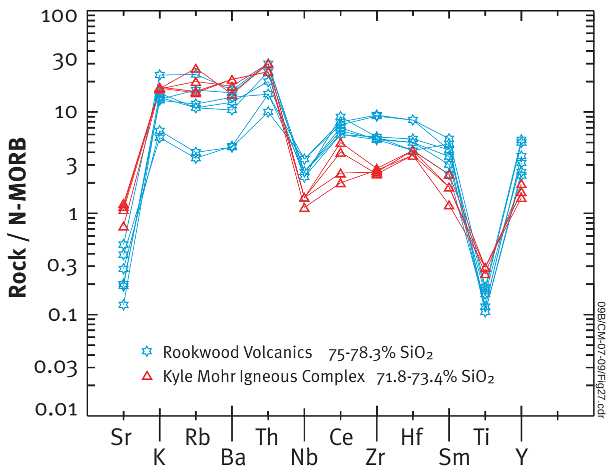


Figure 132. Spidergram plot comparing felsic volcanics of the Rookwood Volcanics with granite of the Kyle Mohr Igneous Complex. N-MORB values are from Pearce (1983).

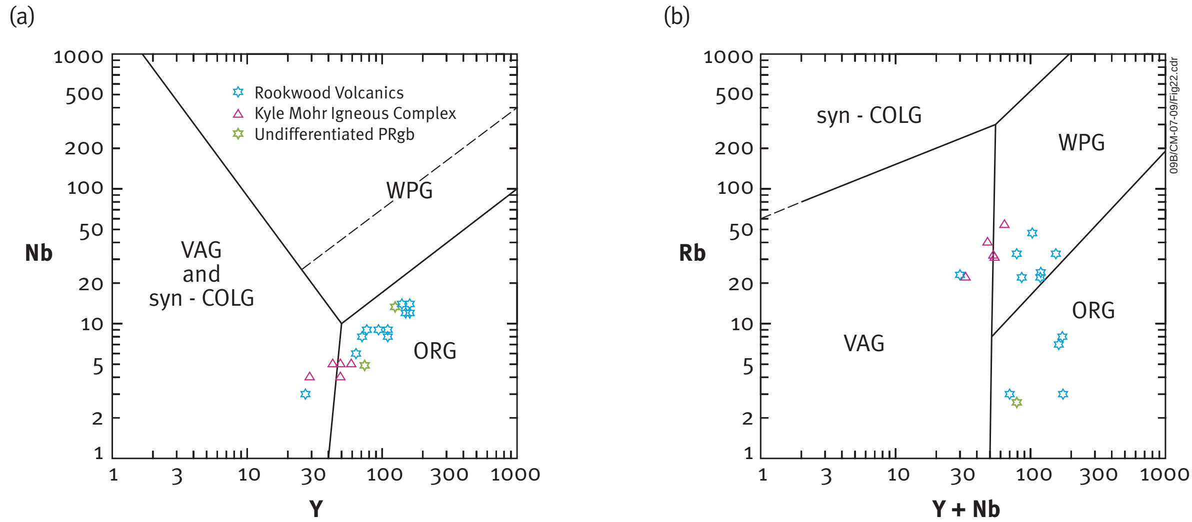


Figure 133. (a) Nb-Y and (b) Rb-Y+Nb discriminant plots of Pearce & others (1984), showing felsic volcanics from the Rookwood Volcanics, granites of the Kyle Mohr Igneous Complex, and granites from an unnamed gabbroic pluton south-west of Westwood. VAG = volcanic arc granites, ORG = ocean ridge granites, WPG = within plate granites, and syn-COLG = syn-collisional granites.

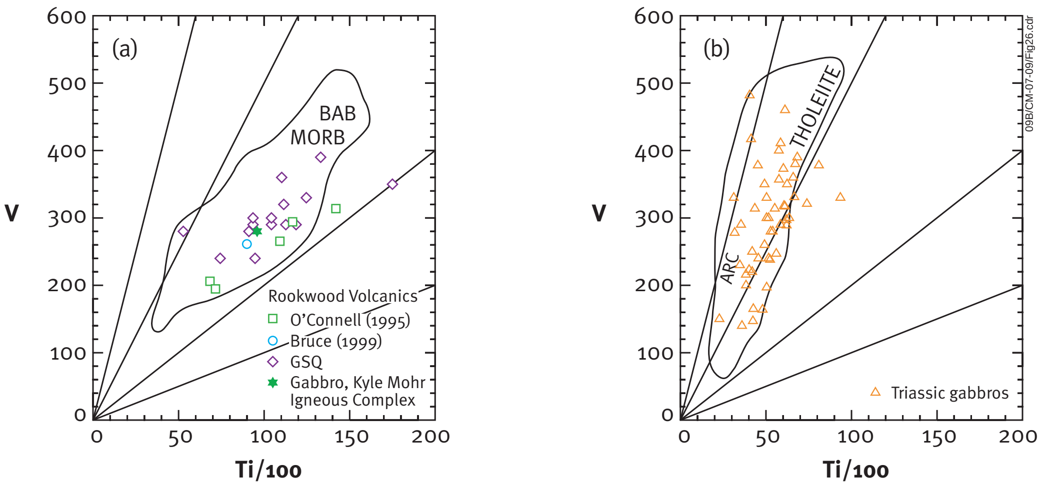


Figure 134. Ti/100 versus V plot (Shervais, 1982) of (a) basalts of the Rookwood Volcanics and a gabbro from the Kyle Mohr Igneous Complex, which fall in the back-arc basin basalt-MORB field (BABand MORB), and (b) Permo-Triassic gabbros, which fall in the arc tholeiite field. Although this diagram was designed specifically to discriminate basalts in ophiolite complexes, it also distinguishes volcanic arc and back-arc samples in continental environments.

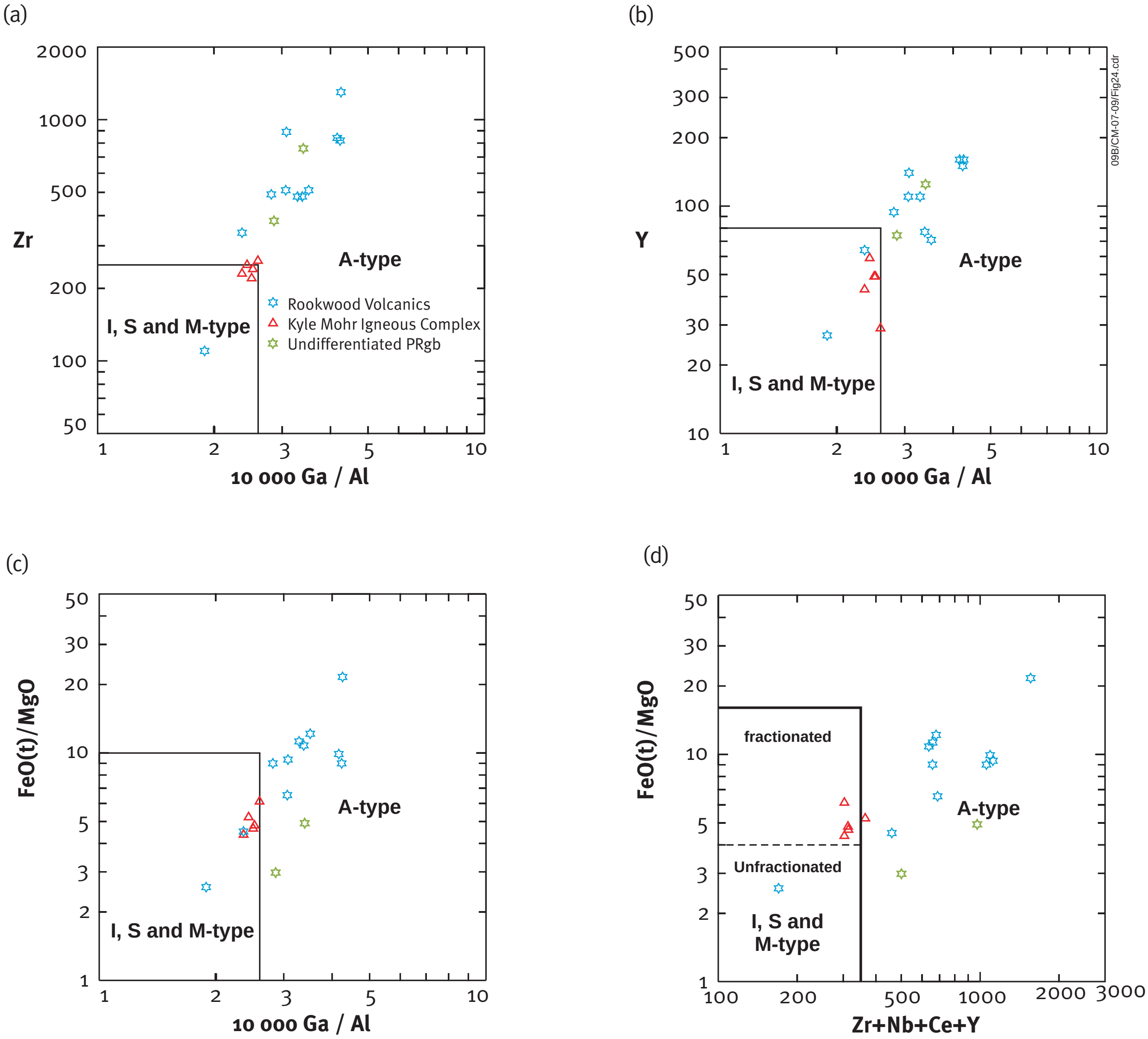


Figure 135(a–d). Discrimination diagrams of Whalen & others (1987) showing that the rhyolites of the Rookwood Volcanics mainly have A-type chemistry, and define trends that include granite samples from the Kyle Mohr Igneous Complex and an unnamed pluton.

arc-related (Figure 134b). The rhyolites and granites have many geochemical features of A-type granites (Figure 135), except for the lack of high Nb.

One proposed origin for A-type granites is by melting of relatively anhydrous lower crust due to intrusion of mantle-derived basalts (Whalen & others, 1987). This is certainly consistent with the geology of the Rookwood Volcanics, which are mostly basalts with no or minimal crustal signature (Stephens & others, 1996a), and the Kyle Mohr Igneous Complex, which has an almost continuous rim of gabbro surrounding and intruded by a granite core.

Environment of deposition

Holcombe & others (1997b) and the Yarrol Project Team (1997) discussed the environment of deposition and concluded that a marine origin for the sequence was likely, but with water depths no deeper than shelfal (<200m), but below storm wave base. This was based on the locally abundant trace fossils (*Planolites*, *Rhizocorallium* and *Chondrites* of the *Cruziana* ichnofacies), macro-invertebrate fragments and the one foraminifera locality. The volcaniclastic deposits are interpreted as density current deposits. The thin to laminated carbonaceous mudstones are interpreted as hemipelagic or low-concentration turbidites. The shelfal environment disagrees with previous interpretations of deep marine environments by Malone & others (1969), Clare (1993), and Messenger (1994b). Malone & others (1969) interpreted the petrographic characteristics of the Rookwood Volcanics as being due to spilitisation in a deep marine environment, but Holcombe & others (1997b) pointed out that similar mineralogy and textures are present in subaerially-erupted lavas in the region, such as in the Lizzie Creek Volcanics and Camboon Volcanics, and that these are due to burial metamorphism during the Bowen Basin evolution.

The Rookwood Volcanics have been interpreted by Holcombe & others (1997b), and Fielding & others (1994, 1997a) as representing the extensional phase during the early stages of development of the Bowen Basin. Although the geochemistry supports this interpretation, the lack of sedimentary fill within the Rookwood Volcanics suggests no substantial rifting in the area during formation. Also, the mid-Permian Rookwood Volcanics were erupted later than the initial rift phase in the Bowen Basin, and correlate with the sag phase (Korsch & Totterdell, 2009). The sag phase was followed by thrust loading and foreland basin formation and deposition of the bulk of the Bowen Basin fill. The north-western outcrops of the Rookwood Volcanics were incorporated into the Gogango Overfolded Zone. Holcombe & others (1997a) proposed a thrust model whereby sequences including parts of the Rookwood Volcanics were thrust from the east into their present position, locally overlying stratigraphically higher Bowen Basin sediments.

Previous suggestions that the Rookwood Volcanics were deposited in the deep Grantleigh Trough, which was separated from the Bowen Basin by the Camboon Volcanic Arc (Kirkegaard & others, 1970; Day & others, 1983), are not supported by sedimentological evidence from the Gogango Overfolded Zone (Fielding & others, 1994).

Thickness

The thickness of the Rookwood Volcanics is uncertain, although Clare (1993) estimated thicknesses of 3km and 4.5km for the Westwood and Goovigen areas respectively. In the latter area, this assumed dips of 45° to 85° to the north-east, and no repetition by folding or faulting. However, this estimate is disputed by Fielding & others (1994), and the unit is likely to be much thinner than the figures given by Clare (1993).

Structure

The Rookwood Volcanics are essentially massive, and except for the rare sedimentary intervals, dips are difficult to discern (although approximate dips and overall younging direction can be determined from the pillow lavas). Bedding trends are only poorly developed on aerial photographs. Messenger (1994b) states that north of the Develin Creek prospect, the sequence dips regularly to the north-west at ~30–40°, whereas at the prospect the area is strongly faulted and complex.

Unlike the rocks to the west in the Gogango Overfolded Zone, including the Camboon Volcanics, the Rookwood Volcanics are not pervasively foliated, although cleavage is well developed in some sedimentary beds (Figure 129).

Biostratigraphy and age

An early to mid-Permian age has been assigned to the Rookwood Volcanics. Artinskian foraminiferids that belong to the *Pseudohyperammina radiostoma* Zone were found in the upper part of the Rookwood Volcanics in the Scrub Creek area (V. Palmieri, written communication, 1995; O’Connell, 1995, appendix 2).

The Youlambie Conglomerate, which underlies and partly interfingers with the Rookwood Volcanics in the Scrub Creek area (Fielding & others, 1997a), contains early Permian (Artinskian) fossils in the upper part of the unit (O’Connell, 1995). They are mainly *Tomiopsis ovata,* which is referable to the *Echinalosia warwicki* and *Echinalosia preovalis* Biozones of Briggs (1998).

The Rookwood Volcanics may be comagmatic with the Kyle Mohr Igneous Complex, which has given a mid-Permian U-Pb zircon date of 270.6±3.7Ma. This is younger than the age assigned to the Rookwood Volcanics by Briggs (1998), but overlaps with dates from rhyolites in the correlative upper Berserker Group.

Stratigraphic relationships

Exposed contacts are rarely observed, and consequently, relationships with adjacent units are difficult to establish in the field. However, the Rookwood Volcanics are inferred to be mostly in fault contact with adjacent units.

In particular, the western contact of the belts appears to be a major thrust of the Rookwood Volcanics over the Back Creek Group. As discussed earlier, unlike the rocks to the west in the Gogango Overfolded Zone, including the Camboon Volcanics, the Rookwood Volcanics are not pervasively foliated. The contrast across contacts is quite marked, consistent with the presence of a major thrust. East of this structure, the Bowen Basin and Yarrol terrane rocks are much less deformed, with relatively shallow dips and weak to absent cleavage. The thrust thus marks the eastern extent of the Gogango Overfolded Zone.

In the Scrub Creek (or Aeroview) area, Fielding & others (1997a) and O’Connell (1995) reported that the Rookwood Volcanics both overlie and partly interfinger with the Youlambie Conglomerate, although farther west O'Connell (1995) interpreted a thrust contact dipping to the north-east. The geophysical data also indicate that along part of the contact, the Youlambie Conglomerate is thrust over the Rookwood Volcanics. In the Westwood area, the Rookwood Volcanics overlie and are faulted against the Yolambie Conglomerate.

The relationship of the Rookwood Volcanics to the Back Creek Group rocks (formerly Rannes beds) that lie to the east of the belt near Rookwood and Ohio homesteads is probably mainly faulted, although stratigraphic contacts may occur in places.

The Back Creek Group rocks overlie the Rookwood Volcanics, based on regional lithological correlation and biostratigraphic evidence. The Moah Creek beds in the north-western part of the Mount Morgan 1:100 000 Sheet area overlie the Rookwood Volcanics and they are probably correlatives of the Back Creek Group and Boomer Formation. Malone & others (1969) and Dear & others (1971) both stated that the Rookwood Volcanics overlie the Rannes beds, now included in the Back Creek Group. A ferruginous bed containing angular blocks of sedimentary rock and some quartz pebbles at the inferred base of the Rookwood Volcanics was cited by Malone & others as evidence of an unconformity. Neither group of authors considered the possibility that the contacts may be thrusts.

Drilling by QMC Exploration Pty Ltd (1995) at their Sweet Caroline prospect in the Lakeview area provided evidence for an easterly-dipping thrust contact. Holes sited in Rookwood Volcanics near their western boundary drilled through the thrust into carbonaceous sedimentary rocks.

Relationships in the area west of Goovigen are uncertain. The Rookwood Volcanics are in contact with an assemblage of strongly foliated, locally mylonitic phyllitic mudstone, lithic sandstone and chert or siliceous siltstone, tentatively assigned to the Back Creek Group and Woolein beds. As observed elsewhere, the Rookwood Volcanics are not foliated, and the contact is probably a thrust. An area of cleaved mudstone, lithic sandstone and siliceous siltstone or chert with the typical "hot" radiometric signature of the Back Creek Group is enclosed by the Rookwood Volcanics. It may be a "window" of Back Creek Group exposed as an antiformal core, beneath the thrust. However, exposure is poor, and bedding is poorly developed within the deformed sedimentary rocks, and this folding model could not be proved.

The contact relationships between the felsic and mafic rocks, such as the slices of massive rhyolite and pillow basalt near Comanche homestead, are unclear. However, the geochemistry of the large outcrop south of Westwood suggests that they are related and presumably coeval. Also, in relation to the surrounding stratigraphy it is difficult to know whether the dolerite sills are intrusions or lavas (O'Connell, 1995).

The Rookwood Volcanics are considered to be roughly coeval with the early to mid-Permian Berserker Group (Crouch, 1999a,b), and both host volcanogenic massive sulphide deposits (Develin Creek and Mount Chalmers respectively). However, they differ significantly in the composition of the volcanic rocks. The Rookwood Volcanics are predominantly basaltic, whereas the rocks of the Berserker Group are predominantly intermediate to felsic (Crouch & Parfrey, 1998). In addition, geochemical analyses from the Rookwood Volcanics are MORB-like, whereas analyses from the Berserker Group show evidence of mantle depletion, particularly of Nb.

Correlation with other units

According to the correlation of Briggs (1993, 1998), the Rookwood Volcanics are equivalent in age to at least the upper part of the Berserker Group, but are different in composition since the Berserker Group is dominated by felsic volcanics. They are also at least partial time equivalents of the Owl Gully Volcanics, which have a strong subduction component.

Economic significance

The Develin Creek Cu-(Zn) mineralisation is hosted within the sequence of undeformed massive and pillow basalts and minor sediments of the Rookwood Volcanics (Horton & others, 1994; Messenger, 1994b; O'Connell, 1995). Mineralisation of the Develin Creek deposits consists of massive sulphide breccias, polymict breccias, layered sulphides, and stockwork zone mineralisation (Bendall, 1996). Although the mineralisation in undeformed mafic volcanics is characteristic of Cyprus-type volcanic-associated massive sulphides, there is no evidence for the existence of an ophiolite, and the Rookwood Volcanics appear to have been erupted in a continental environment.

The depositional environment of the volcanic-hosted massive sulphide deposit at Develin Creek remains enigmatic. The sulphide mineralogy of this deposit suggests formation at water depths greater than 700m, but the sediments suggest shelfal depths.

The use of radiometrics to locate mineralisation in the Rookwood Volcanics was discussed by Jones (2000b).

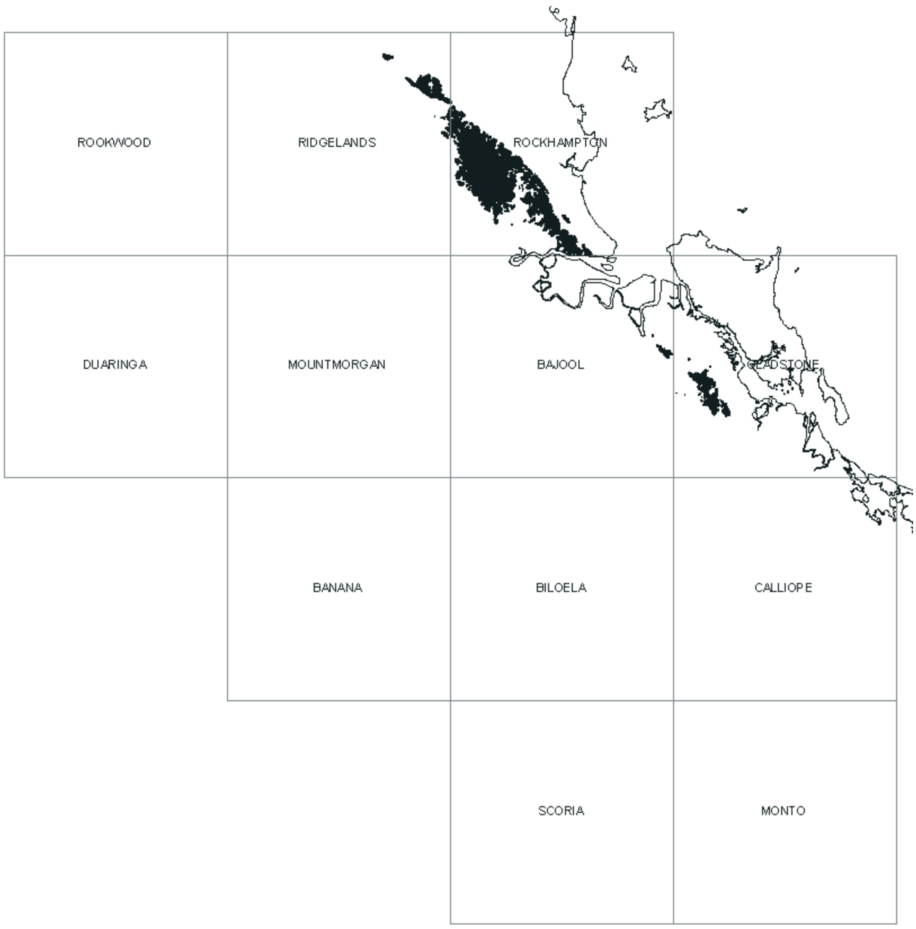


Figure 136. Distribution of the Berserker Group

Berserker Group (Pk)

(S.B.S. Crouch)

These rocks were first named the Berserker Series by Whitehouse (1930), who regarded them as Devonian, although recognising the local presence of Permian fossils at Lakes Creek (Whitehouse, 1928). Mapping by Hawkins & Whitcher (1962) was the first to establish that Permian faunas were widespread. Kirkegaard & others (1970) applied the name Berserker beds to the Permian sedimentary and volcanic succession which straddles the Fitzroy River. The name was derived from the Berserker Range, in which constituent rock types are well exposed. The distribution of the Berserker Group is shown in Figure 136.

The Berserker Group crops out in a partly fault bounded belt ~90km long and up to 15km wide, running from the Mount Etna area in the north to the Mount Larcom area in the south. Kirkegaard & others (1970) interpreted the Berserker beds to be a graben fill, and ~3000m thick. However, present mapping suggests that the Berserker Group extended beyond its existing structural margins, was not deposited in a graben, and is considerably thinner than the figure given by Kirkegaard & others (1970).

A number of different rock units, the Lakes Creek Formation, Chalmers Formation (including the Sleipner Member), and Ellrott Rhyolite, have now been mapped within the Berserker beds, and the name is revised to Berserker Group (Crouch & Parfrey, 1998). In addition, a small unit of Upper Permian sediments, the Warminster Formation, has been separated from the group. Relationships between these units are depicted in Figure 137. Permian rocks designated Berserker beds on the Rockhampton 1:250 000 Sheet west of Canoona (Kirkegaard & others, 1970) were not mapped in this project; these rocks are now classed as undifferentiated Permian until they are studied more closely and compared to the rocks of the Berserker Group.

The structure of the Berserker Group is simple. North of the Fitzroy River, the unit has a shallow to moderate regional dip to the west which is reversed in some areas due to local folding, and complicated by large intrusive rhyolite and dacite bodies. South of the river, the structure is less clear due to the difficulty of separating the Berserker Group from the Rockhampton Group. Here, Crouch & Parfrey (1998) and Crouch (1999b) postulated the existence of a syncline which was displaced by dextral faulting.

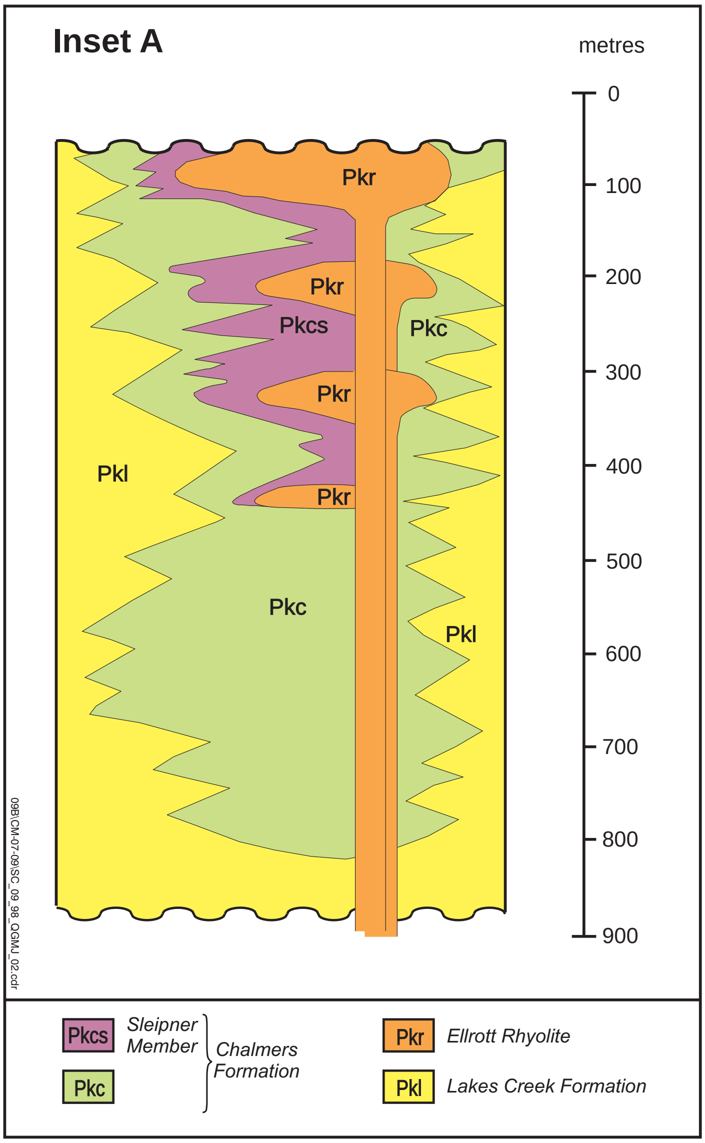


Figure 137. Relationships and thicknesses of units within the Berserker Group

Permian marine fossils found in the middle of the sequence, near Rockhampton and Mount Chalmers, and at its base in the Mount Larcom area, as well as U-Pb zircon dates obtained from the middle of the sequence and from the Ellrott Rhyolite, constrain the age of the top half of the Berserker Group.

The Berserker Group unconformably overlies the Upper Devonian to Lower Carboniferous Mount Alma Formation in the Mount Etna area, and the Lower Carboniferous Rockhampton Group in the Mount Larcom region. The Berserker Group is unconformably overlain by the Upper Permian Mount Warminster Formation 1km north-west of Mount Chalmers.

Lakes Creek Formation (Pkl)

(S.B.S. Crouch)

Introduction

The Lakes Creek Formation is the lowest unit of the Berserker Group, although it is at least partly equivalent to the Chalmers Formation.

A detailed study of the Lakes Creek quarry area covering ~3km^2 was conducted by East (1946), who defined an ~800m thick sequence comprising (from top to bottom) dark clay siltstone (250–300m), highly fossiliferous siltstone (100m), massive quartzose sandstone (60m), and interbedded sandy siltstone and

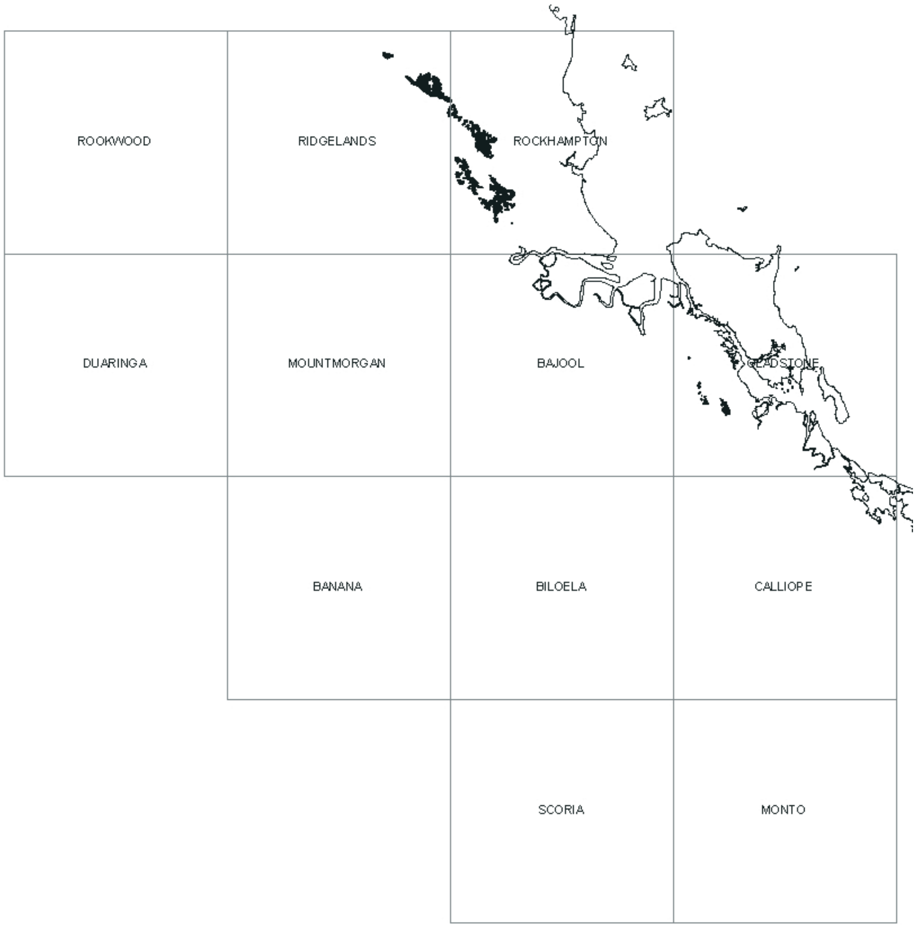


Figure 138. Distribution of the Lakes Creek Formation

sandstone (400m). Intrusive bodies of diorite, feldspar porphyry and quartz-feldspar porphyry were also found.

Derivation of name

The name is from Lakes Creek, a tributary of the Fitzroy River. The first use of this name was by Smith (1887), detailing fossils within indurated, pyritic shale of the Lakes Creek beds.

Type area

The type area of the Lakes Creek Formation is designated here as a 1km^2 area bounded by the Lakes Creek quarry on the west and Nerimbera settlement to the east. In the type area, the unit is dominated by interbedded siltstone and sandstone with local fossiliferous beds. Calcareous sandstone to limestone is restricted to Lakes Creek quarry.

Distribution

The Lakes Creek Formation forms a discontinuous belt about 45km long and up to 10km wide extending from Alligator Creek in the north to the Fitzroy River in the south. Much of this area is covered by younger units of the Berserker Group, the Chalmers Formation and the Ellrott Rhyolite. The main outcrops are north of Mount Etna, north of the Rockhampton–Yeppoon Road, and east of Rockhampton encompassing the type area near Lakes Creek. South of the Fitzroy River, the Lakes Creek Formation forms small outcrops in a belt about 12km long west of the peak of Mount Larcom, and a larger outcrop at Yarwun south-east of the Targinie Quartz Monzonite (Figure 138).

Topographic expression

The Lakes Creek Formation occupies generally flat, low areas although steeper terrain is common adjacent to the large rhyolite to dacite domes especially on the southern margin of the Berserker Range. The terrain is characterised by a widely spaced dendritic drainage pattern.

Figure 139. Interbedded sandstone and siltstone, Lakes Creek Formation

Geophysical expression

The Lakes Creek Formation has a moderate to strong radiometric response, and appears in red to white tones on images produced from airborne data. It has a low magnetic signature, and the main features are north-trending dykes that are most prominent in images of the first vertical derivative.

Lithology

The Lakes Creek Formation is a sequence of grey, massive, indurated siltstones and lithofeldspathic to quartzofeldspathic sandstones derived from felsic to intermediate volcanics. This sequence is composed of generally thin to medium beds of siltstone, fine to medium-grained, well sorted sandstone, and chert, giving the rock a banded appearance (Figure 139). Less common moderate to poorly sorted, granule to pebbly sandstones are found locally, and contain angular to rounded clasts of fine-grained sediments and volcanics.

Sedimentary structures include local occurrences of soft sediment deformation features, and rhythmic bedding which was found near Nankin, consisting of thin planar siltstone beds and thin, planar to disrupted sandstone beds.

Fossiliferous calcareous sandstone to limestone exists at Lakes Creek quarry but is not present elsewhere within the sequence. Similar rocks have been found further south near Mount MacDonald, and 2km north-west of Mount Kilner, but are associated with volcanic beds indicative of the Chalmers Formation.

Silicification resulting in induration is common throughout the area. Local hematitic and limonitic alteration is also present in association with felsic to intermediate intrusives.

Environment of deposition

The presence of a wide variety of marine fossils, and the type of sediments and sedimentary structures suggest that the Lakes Creek Formation formed in a shelf environment. Deposition below wave base is suggested by the common occurrence of fossils lacking the effects of abrasion, including the presence of fine spines on brachiopods and of entire blastoid cups. The fauna of brachiopods, molluscs, and bryozoans is considered to be typical of that found in water depths of around 30–50m, and no deeper than 200m.

Figure 140. Gentle folding in interbedded siltstone and sandstone, Rockhampton – Emu Park road, Lakes Creek Formation.

Thickness

East (1946) interpreted the Lakes Creek Formation to be ~800m thick at Lakes Creek based on a measured section and structural evidence. However, west of the peak of Mount Larcom the unit may be substantially thinner.

Structure

The Lakes Creek Formation generally dips at shallow to moderate angles. Small scale gentle to open folds with a wavelength of ~5m can be seen locally (Figure 140).

East (1946) described a north-east trending fault that bisects the Lakes Creek quarry area, the south block defined by westerly dipping beds, and the north block by a series of antiforms and synforms.

Biostratigraphy and age

Marine fossils correlative with faunas of the Lower Permian Buffel Formation of the Bowen Basin have been collected from Lakes Creek quarry, at Lakes Creek, Artillery Road, 8km north-north-west of Cabbage Tree Hill and 6km north-north-west of the peak of Mount Larcom. These include abundant fenestellid bryozoans, *Echinalosia warwicki* (Maxwell), *Anidanthus* sp., *Taeniothaerus* n. sp, *Costatumulus farleyensis* (Etheridge & Dun), and *Sulciplica* sp. The Lakes Creek quarry fauna has been assigned to the mid-Artinskian *Echinalosia warwicki* zone by Briggs (1998).

Stratigraphic relationships

The Lakes Creek Formation is probably laterally equivalent to the Chalmers Formation with the latter containing sediment similar to the Lakes Creek Formation interbedded with volcaniclastic rock (Figure 137).

The Lakes Creek Formation is intruded by rhyolites and dacites of the Ellrott Rhyolite, andesite dykes and gabbro bodies. The contacts are sharp and steep, and metamorphism is weak.

Shallowly dipping siltstone and sandstone beds of the Lakes Creek Formation overlie more steeply dipping volcaniclastic conglomerate of the Upper Devonian to Lower Carboniferous Mount Alma Formation 4km east of Mount Etna. Structural measurements and radiometric evidence also suggests that gently to moderately folded Berserker Group, including the Lakes Creek Formation, overlies steeply dipping siltstone and mudstone beds of the Lower Carboniferous Rockhampton Group west of the peak of Mount Larcom. An angular unconformity is inferred in these two areas. The possible angular unconformity between the Lower Carboniferous Rockhampton Group and the Lower Permian Berserker Group at the second location contrasts with a previous structural interpretation which defined a narrow fault bounded sliver of Carboniferous rocks 10km north-west of the peak of Mount Larcom (Donchak & Holmes, 1991). The unconformity between Chalmers Formation and the Rockhampton Group south of Mount Larcom suggests that the basement surface was higher in this area, impeding deposition of either all or a substantial part of the Lakes Creek Formation.

Correlation with other units

The Lakes Creek Formation is a time correlative of the Yarrol Formation based on similar brachiopod faunas (Briggs, 1998).

Chalmers Formation (Pkc)

(S.B.S. Crouch)

Introduction

The Chalmers Formation is a new geological unit encompassing sediments similar to those in the Lakes Creek Formation interbedded with rhyolitic to andesitic volcaniclastics. It also contains fossil-rich beds slightly younger than those in the Lakes Creek Formation, situated in the middle to upper part of the unit.

Derivation of name

This formation is named after Mount Chalmers, which is both a topographic feature and a settlement 15km north-east of Rockhampton.

Distribution

The Chalmers Formation is the most extensive unit in the Berserker Group. It extends from Mount Belmont in the north to the Mount Larcom area in the south (Figure 141).

Type area

The type area is located 300m south of Mount Archer, from GR252267 7416546 to GR251747 7415695, and consists of interbedded thick to very thick beds of grey to olive siltstone, fine sandstone, and grey to green-grey, volcaniclastic, feldspatholithic sandstone and breccia.

Topographic expression

Topographic expression is generally low to moderate for the Chalmers Formation with higher terrain found adjacent to rhyolite to dacite domes. The drainage ranges from a close to widely spaced dendritic pattern.

Geophysical expression

The radiometric response of the Chalmers Formation is moderate, and it gives red to pink hues on images derived from airborne data. Like other units in the Berserker Group, it has a weak magnetic signature, and the only features recognisable are probable north-trending dykes and a possible north-west trending fault just over 3km west of and parallel to the Yarrol Fault, most obvious on the first vertical derivative.

Lithology

The Chalmers Formation is composed of sediments similar to the Lakes Creek Formation as well as pyroclastic rocks and granule to pebble volcanic breccia containing siltstone to sandstone clasts and felsic to intermediate volcanic clasts. Disrupted contacts between underlying siltstone-sandstone and overlying volcanic breccia can be seen at a road cutting 1km south-south-east of Mount Archer, and also in Nankin Creek, 2km south-west of Mount Chalmers. The fossiliferous beds consist of grey-green, medium to thick-bedded sandstone and granule to pebble breccia.

The volcaniclastic rocks are generally grey to green-grey, medium to very thick bedded, matrix-supported to clast-supported, and moderately to poorly sorted (Figure 142). These rocks contain crystals, lithics and pumice. The crystals are mostly feldspar and to a lesser extent quartz with some rocks containing rare hornblende, pyroxene or biotite. Feldspar is fine to coarse-grained (up to 6mm), quartz is fine to medium-grained, and mafic minerals are mostly fine-grained. Lithics are sedimentary and volcanic, although

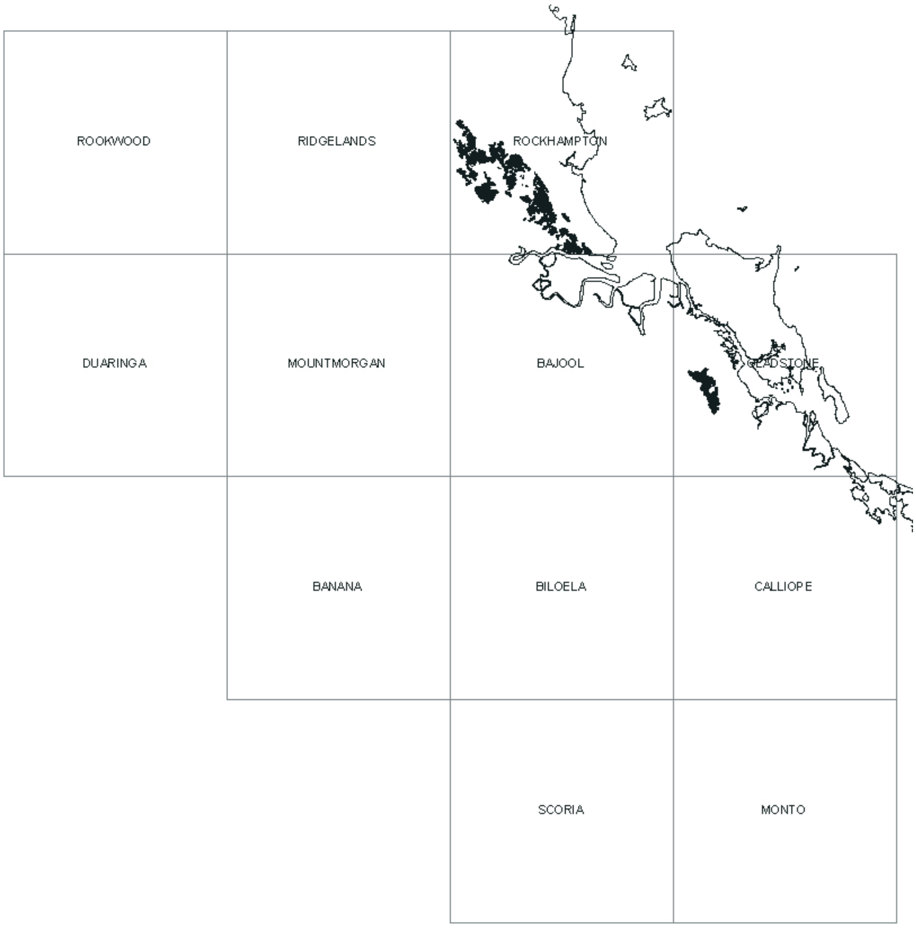


Figure 141. Distribution of the Chalmers Formation

in many cases they are difficult to distinguish from each other. Volcanic clasts dominate, and are granule to cobble in size, usually angular to subangular, and range from rhyolitic to andesitic in composition. The pyroclastic rocks are generally unwelded with rare fiamme. Pumice fragments are up to 50mm in length, angular, either elongate or irregular in shape, and are flattened due to compaction.

A thick-bedded calcareous sandstone containing Lower Permian fossils was mapped 1km south of Mount MacDonald. A similar looking horizon of interbedded calcareous and non-calcareous sandstone found 5km further south, and along strike, was unfossiliferous and contained well rounded, pebble to cobble sized sedimentary or volcanic clasts (Figure 143).

Approximately 1.5km north-north-west of Broadmount there is a relatively small area of schist and tectonite which is adjacent to the Yarrol Fault. The origin of these rocks is uncertain, as is their relationship with the Chalmers Formation. The most likely explanation is that these rocks are Chalmers Formation which were deformed during the westerly compressional event in the latest Permian. Their orientation is slightly south of west.

One kilometre north of the peak of Mount Larcom, a massive, possibly pyroclastic, volcanic rock contains green, rounded accretionary lapilli or infill features. On the margin of a large intrusive at the northern boundary of Mount Archer National Park, 2km west of Cabbage Tree Hill, there is a possible peperite defined by minor amounts of siltstone to fine sandstone anastomosing through the dacite intrusive. Peperite was also observed at Mount Chalmers mine 500m north of Mount Chalmers (Hunns & McPhie, 1999).

In thin section the volcaniclastic rocks are all devitrified, and have been silicified, chloritised, or saussuritised to some degree. Clay alteration is pervasive. Quartz veins are localised in areas of strong alteration.

Geochemistry

One sample of rhyolitic ignimbrite in a breccia from the Chalmers Formation was analysed. The result is similar to rhyolite and dacite from the Ellrott Rhyolite, and the geochemistry is discussed under that unit.

Figure 142. Thick bedded sediments and volcaniclastics dipping to the south-west, Mount Archer, Chalmers Formation

Figure 143. Interbedded calcareous and non-calcareous sandstone containing well rounded pebble to cobble-sized sedimentary and volcanic clasts, Mount MacDonald, Chalmers Formation

Environment of deposition

The Chalmers Formation is considered to represent incursion of volcaniclastic sands and breccias into a sedimentary basin receiving silts and sands equivalent to the Lakes Creek Formation. The fauna of brachiopods and molluscs indicates water depths of around 30–50m, and no deeper than 200m. Good preservation of the fossils suggests that the sediments were deposited below wave base.

Sainty (1992) observed trace fossils throughout sediments now assigned to the Chalmers Formation. The burrows include *Teichichnus*, *Planolites*, and abundant vertically orientated, upward-branched traces, with scattered *Rhizocorallium*- and *Zoophycus*-type burrows. These trace fossils belong to the *Cruziana* ichnofacies. Sainty (1992) suggested that both the macrofossils and trace fossils indicate a shallow-marine shelf setting.

Thickness

The Chalmers Formation is considered to be ~600m thick. This is based on two criteria. First, although the unit is gently to moderately folded, its regional orientation is relatively flat lying; and second, the unit has been identified from sea level to close to the top of many of the peaks in the region, including the highest, Mount Archer at 604m.

Structure

The Chalmers Formation shows gentle to moderate dips. Folding is best developed in the Mount Chalmers area, where there is a broad north-north-west trending syncline. A north-north-west trending anticline with gently dipping limbs is centred on the Mount Chalmers mine, with the main suphide horizon on its eastern limb (Fletcher, 1975; Taube, 1980b, figure 15; Taube & van der Helder, 1983, figure 11). Fletcher (1975), Taube (1979, 1980b) and Taube & van der Helder (1983) reported that a slaty cleavage or schistosity is locally well developed in the area covered by the main mineral deposits in the unit, but no foliation is generally present in the remainder of the Chalmers Formation.

Biostratigraphy and age

Marine fossils correlative with faunas of the Lower Permian Buffel Formation of the Bowen Basin have been collected 2km south-west of Mount Chalmers, 1km south-east of Mount MacDonald, and 2km north-west of Mount Kilner. These include *Echinalosia warwicki* (Maxwell), *Terrakea geniculata* (Waterhouse), *Eurydesma* sp., relatively few fenestellid byrozoans, and lack *Taeniothaerus* n. sp. There is also rare *Euryphyllum* sp. The ranges of the fossils found in the Chalmers Formation are slightly younger than those found in the Lakes Creek Formation.

A U-Pb zircon age of 277.1±3.5Ma (mid-Permian) was obtained from a volcaniclastic bed interpreted to be in the middle of the Chalmers Formation (Appendix 1). This result is consistent with the biostratigraphic scheme of Briggs (1998).

Stratigraphic relationships

The Chalmers Formation is interpreted to be laterally equivalent to the Lakes Creek Formation, although mapping in the Mount Archer area suggests that the Chalmers Formation locally rests on Lakes Creek Formation. West of the peak of Mount Larcom, at the base of the Mount Larcom Range, mapping suggests that the unit unconformably overlies Lower Carboniferous Rockhampton Group. An angular unconformity is inferred.

Five kilometres north of Broadmount, in an area immediately east of the Yarrol Fault, is a large outcrop of rock similar to the Chalmers Formation. Its relationship with the Devonian–Carboniferous Coastal Block is uncertain but the Permian unit would presumably overlie the older rock. This area was defined as Berserker beds on the geological map of the 1:250 000 Rockhampton Sheet area, but was interpreted as part of the Balnagowan Volcanic Member of the Devonian–Carboniferous Doonside Formation on the 1:100 000 Rockhampton Region geological special (Willmott & others, 1986).

The Chalmers Formation is intruded by rhyolites and dacites, andesite dykes and gabbro and granodiorite bodies forming sharp, shallow to steep contacts. The U-Pb zircon dates obtained from the Ellrott Rhyolite suggests the latter was synchronous with the deposition of at least the upper half of the Chalmers Formation.

Figure 144. Open cut at Mount Chalmers, showing remains of old underground workings, Chalmers Formation

Economic significance

The Mount Chalmers area is the site of typical Kuroko-style base metal and gold deposits, the largest of which is at Mount Chalmers itself (Figure 144), but with smaller examples at Woods shaft, Mount Warminster, and Boto's area (Large & Both, 1980; Taube & van der Helder, 1983). About 4km to the south are numerous small gold workings in the New Zealand Gully area. Production from these was both alluvial and from narrow lodes.

Sleipner Member (Pkcs)

(S.B.S. Crouch)

Introduction

The Sleipner Member was previously defined as the Sleipner Andesite Tuff by Taube (1979), and the name was published by Hunns (1994) as the Sleipner andesitic breccia. Taube described the unit as consisting of dark green lithic tuff with green to purple andesitic fragments 0.2–2cm across, set in a lighter green chlorite-epidote rich matrix. At some localities the fragments are both andesitic and rhyolitic in character, the latter generally being larger, commonly 20–30cm across.

The Sleipner Member is found in the upper half of the Chalmers Formation (Figure 137), and extends down to no more than 100m below the fossiliferous beds. It is characterised by monomictic to polymictic breccia and lesser sandstone and siltstone.

Derivation of name

The name was derived from the railway siding of Sleipner, ~15km east of Rockhampton.

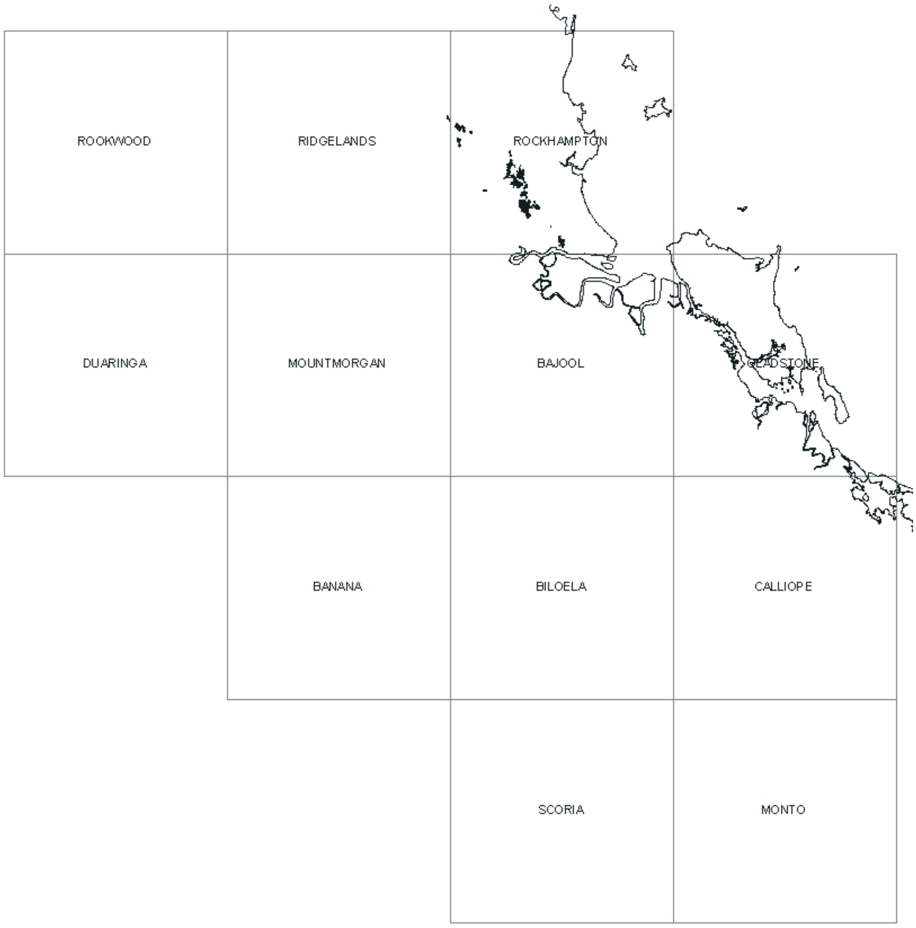


Figure 145. Distribution of the Sleipner Member

Distribution

The Sleipner Member is found at Mount Belmont, Mount Chalmers, Mount MacDonald, and Mount Kilner. It has not been recognised south of the Fitzroy River (Figure 145).

Type area

The type area of the Sleipner Member is situated on the southern side of Mount MacDonald. The area is dominated by relatively flat lying, very thick bedded, dark green-grey breccia and minor conglomerate. Interbedded with these rocks are thick-bedded, grey, feldspatholithic sandstone and minor siltstone.

Topographic expression

The Sleipner Member forms moderate to steep terrain as exemplified at Mount Chalmers and Mount MacDonald, with generally rocky terrain and a wide to closely spaced dendritic pattern of drainage.

Geophysical expression

On images derived from the airborne data, the Sleipner Member has a moderate radiometric response, and shows dominantly as red tones. Like the rest of the Berserker Group, it has a low magnetic signature.

Lithology

The Sleipner Member is dominated by dark green-grey, massive, poorly sorted, matrix supported to clast supported, monomictic and lesser polymictic, granule to boulder breccia and minor conglomerate (Figure 146). Interbedded with the breccia and conglomerate are grey to green-grey, fine to coarse-grained, well-sorted, feldspatholithic sandstone and minor siltstone. Porphyritic andesite to rhyolite, either massive or flow-banded, are the most common clasts and are characteristic components of a distinctive green-grey, monomictic rock within the unit. Grey to dark grey, massive, featureless clasts are also found in this rock and

Figure 146. Very thick bedded pebble to boulder breccia, Sleipner Member

in polymictic varieties. These massive clasts could be an aphyric volcanic, siltstone, or chert. Taube (1979) reported west-dipping andesitic lavas and their feeder intrusions ~2km west of the Mount Chalmers mine.

Silicification, chloritisation and clay alteration are observed in hand specimen and thin section. Saussurite alteration is seen in thin section.

Environment of deposition

The Sleipner Member is interpreted to be a marine mass flow deposit which is poorly sorted, massive and has a sharp contact with underlying beds. Near Mount Chalmers and Mount MacDonald it immediately overlies sediments rich in marine macrofossils. Because the Sleipner Member is spatially restricted, it suggests local areas of instability, probably proximal to volcanic centres defined by the Ellrott Rhyolite.

Thickness

The thickest section observed for the Sleipner Member is 360m at Mount MacDonald.

Structure

The Sleipner Member is folded into a broad, gentle synform approximately 5km in wavelength and trending north to north-north-west in the Mount Chalmers area.

Biostratigraphy and age

Several fossil collections fall within the mapped outcrops of this Member. They are all mid-Permian, like the remainder of the Berserker Group.

Stratigraphic relationships

The Sleipner Member is part of the Chalmers Formation. The contact is preserved in the Mount Chalmers area, both in outcrop and drill holes. Stratigraphic contacts with the Ellrott Rhyolite have not been observed but all occurrences are close to the Rhyolite. Four kilometres south of Mount Chalmers, the Ellrott Rhyolite intrudes the Sleipner Member.

The three dimensional extent of the Sleipner Member is unclear and must be gleaned from scattered outcrops and subsurface intersections. Both outcrop and drill hole sections contain more than one interval of the Member and it appears that the member consists of multiple bodies. These bodies are considered to represent aprons formed by erosion of volcanic domes of the Ellrott Rhyolite. The U-Pb zircon dates (within error) obtained from the Ellrott Rhyolite coincide with the appearance of the Sleipner Member within the Chalmers Formation.

The Sleipner Member appears to be restricted to the upper half of the Chalmers Formation (Figure 137), above and below the fossiliferous sandstone bed situated in the middle to upper part of the formation. Locally the Member overlies the fossiliferous sandstone bed, and drillholes have intersected it no deeper than 100m below this bed.

In the North Star area, 4km south of the Mount Chalmers mine, the geology is represented by large blocks of Sleipner Member within pyroclastic rocks. The blocks are predominantly massive although some do show bedding. This bedding is randomly orientated, suggesting that the area was probably a sequence of Chalmers Formation dominated by Sleipner Member that was brecciated on a large scale during an eruption of the Ellrott Rhyolite.

Ellrott Rhyolite (Pkr)

(S.B.S. Crouch)

Introduction

The term Ellrott Rhyolites was previously defined by Taube (1979), and the name Ellrott Rhyolite was published by Hunns (1994). It comprises rhyolite and dacite volcanic domes, some of which may be extrusive. These are widespread within the Berserker Group, scattered throughout the Rockhampton and Mount Larcom regions and up to several square kilometres in area. The rocks are typically dark grey, massive, and porphyritic in feldspar. Visible quartz is rare.

Derivation of name

The name was introduced by Taube (1979).

Distribution

The volcanic domes have been mapped from Mount Belmont south to an area 10km north of Mount Larcom. Large outcrops of rhyolite to dacite have been mapped at Mount Belmont, the Mount Archer – Cabbage Tree Hill – Nankin area, and also between Mount MacDonald and Mount Kilner. There is also a large body of rhyolite between Mount Chalmers and Mount MacDonald (Figure 147).

Type area

Cabbage Tree Hill is the type area. It represents a massive, dark grey, porphyritic rhyolite-dacite with visible feldspar and quartz. Mafic xenoliths are also present.

Topographic expression

The rhyolite to dacite domes form topographic highs because of their resistant nature relative to the surrounding country rock, and are probably responsible for the elevated country from Mount Belmont to Mount Kilner.

Geophysical expression

The Ellrott Rhyolite has moderate to strong radiometric expression, and gives red to white tones on images from the airborne data. It is weakly magnetic.

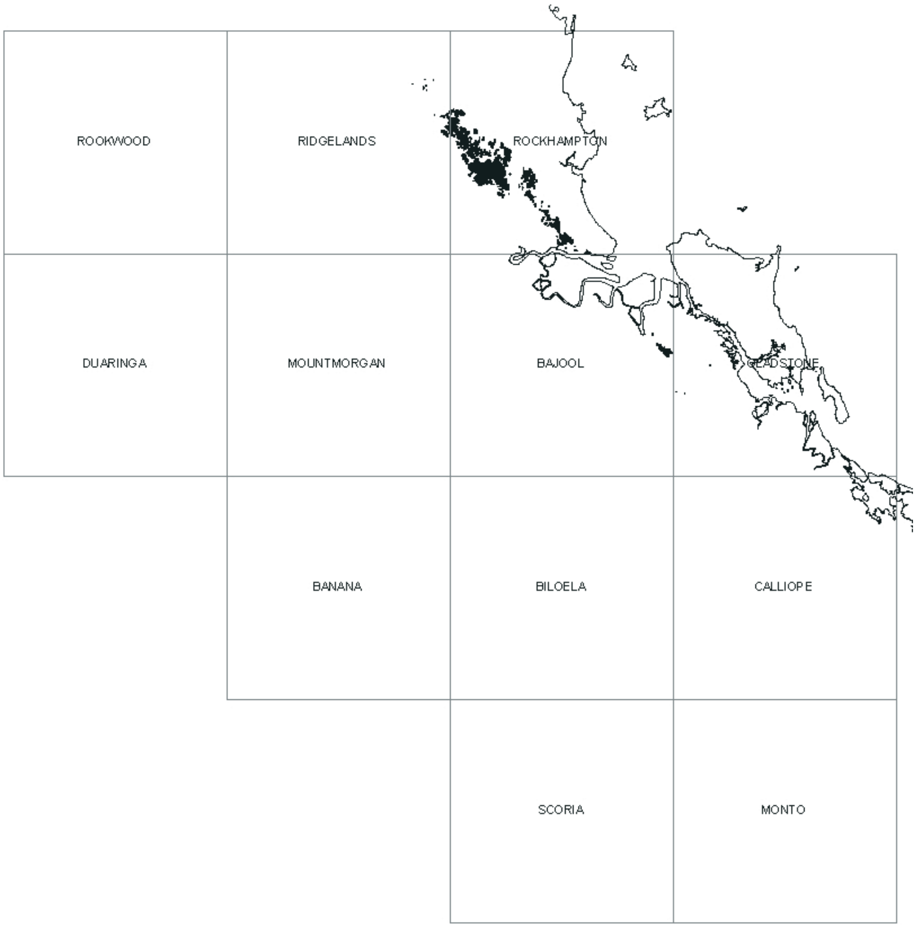


Figure 147. Distribution of the Ellrott Rhyolite

Lithology

The domes range from rhyolite to dacite in composition although some of the rocks with pyroxene may be andesitic. They are either massive or flow-banded (Figure 148), and display porphyritic texture, although aphyric rocks are found locally. Some of the flow-banded rocks are also autobrecciated. Amygdales filled with silica or calcite are common, as are ferromagnesian clots. Xenoliths are not common and seem to be abundant only in the Cabbage Tree Hill area. They are up to 20cm in size and are massive or trachytic-textured, porphyritic andesite.

The groundmass is microcrystalline and exhibits devitrification. Feldspar crystals are commonly fine to coarse-grained (usually up to 6mm although up to 15mm in some cases), and are anhedral to euhedral. Andesine to albite plagioclase and orthoclase are observed in thin section. The plagioclase is zoned; some crystals are deformed and broken, and alteration to clay, sericite and in some cases epidote is usual. Orthoclase forms either aggregates or is interstitial, and is weakly to strongly altered to clay. Rarely it occurs in myrmekitic intergrowths. Quartz is generally sparse, usually fine to medium-grained, (although some rocks have quartz up to 10mm), and is anhedral in shape. Most crystals display undulose character, and some are skeletal in texture. Hornblende is the most common mafic mineral. It is fine to medium-grained, anhedral to euhedral, and in most cases is weakly to strongly chloritised. Many grains are poikilitic, and some have pyroxene cores. Pyroxene is rare and is usually altered to amphibole or removed leaving a void. Biotite is also rare, and is generally associated with the more felsic intrusives. It is fine to medium-grained (up to 3mm), anhedral to euhedral, and is typically altered to chlorite. Secondary calcite is present in the intermediate compositions.

Many of the massive domes are adjacent to monomict breccia, which tends to suggest brecciation of the margins. On the margin of a large body at the northern boundary of Mount Archer National Park, 2km west of Cabbage Tree Hill, a possible peperite is defined by minor amounts of siltstone to fine sandstone anastomosing through the dacite.

Figure 148. Outcrop of flow-banded rhyolite, Ellrott Rhyolite

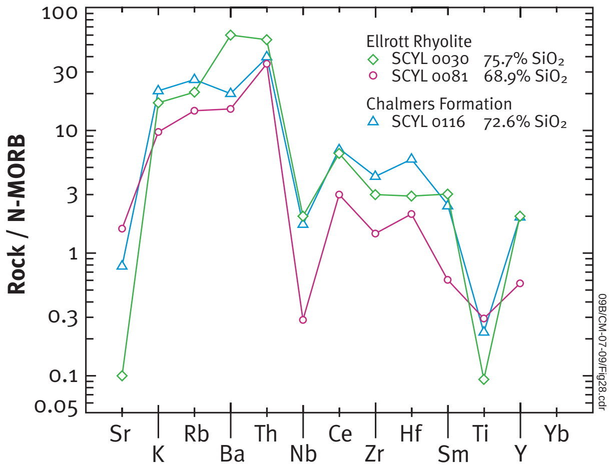


Figure 149. Spidergram plot of rhyolite and ignimbrite from the Chalmers Formation and Ellrott Rhyolite. N-MORB values are from Pearce (1983).

Geochemistry

Analysed samples of the Ellrott Rhyolite and ignimbrite from a breccia from the Chalmers Formation are medium-K dacites and rhyolites. Their geochemistry supports a moderate amount of subduction-related influence with some enrichment in large ion lithophile elements (LILE) relative to high field strength elements (HFSE). In particular, they have comparatively low Nb and medium to extremely large values of $(Th/Nb)_N$, ranging from 23.3–122.3 (Figure 149). On the Y versus Nb discriminant plot of Pearce & others (1984), two analyses fall in the ocean ridge granite field, and they fall in the within plate field on the Y+Nb versus Rb diagram (Figure 58). The geochemistry favours a back-arc basin origin for the Berserker Group.

Structure

The contacts of the intrusive rhyolite to dacite domes with the Lakes Creek Formation and Chalmers Formation are generally steep. However, some contacts are shallow, as seen 400m south-west of Mount Archer (along Pilbeam Road). Here, and 3km north-west of Mount Kilner, these bodies disrupt adjacent sediments and volcaniclastics to form steeply-dipping beds. The extrusive rhyolite to dacite domes have both shallow and steep contacts with the Chalmers Formation in the Mount Chalmers area and north of Mount MacDonald.

Age

A mid-Permian age is assigned to the Ellrott Rhyolite based on two U-Pb zircon dates of 276±3.9Ma and 268.2±3.9Ma (Appendix 1). Intrusion and extrusion probably covered a range of ages, since the rhyolite is regarded as the source for the Sleipner Member, and also locally intrudes that unit.

Stratigraphic relationships

The Ellrott Rhyolite is synchronous with deposition of the Chalmers Formation, and is a likely source of the Sleipner Member. The monomict breccia adjacent to the margins of some of the domes is similar in appearance to the Sleipner Member. The Ellrott Rhyolite is intruded by andesite to basalt dykes, rhyolite dykes, gabbros, and granodiorites. The relationship with the Upper Permian Mount Warminster Formation is unknown.

Unnamed Permian (Pu)

(C.G. Murray)

Introduction

Kirkegaard & others (1970) included Permian rocks west of Canoona in the Berserker beds (now Berserker Group). This sequence was examined only briefly during the Yarrol Project mapping, and it is separated as unnamed Permian because it cannot be correlated with any specific formation within the Berserker Group.

Distribution

The Permian beds form a roughly east-west trending belt 25km long and up to 5km wide from Canoona homestead to Princhester Creek, turning more to the north-west at its western end (Figure 150).

Topographic expression

The topographic expression of the unit is varied. Most of the outcrop area is low, undulating country, but there are rugged, north-west trending ridges in the western part which consist of fine-grained siliceous tuff or intrusive dacite.

Geophysical expression

The prominent ridges of fine-grained siliceous tuff or dacite give a strong radiometric response. Elsewhere the radiometric signature is moderate, with pink tones indicating a high level of potassium.

The magnetic signature is low, except on the first vertical derivative, which gives a more varied response, in part moderate.

Lithology

Rock types listed by Kirkegaard & others (1970) are dominant grey siltstone with interbedded lithic sandstone and conglomerate, fine-grained siliceous tuff and crystal tuff. The prominent ridges are composed of a grey, glassy rock that locally displays faint flow banding and consists of a mosaic of fine-grained plagioclase (which forms some microspherulites) and quartz, which could be an ash-rich tuff or an intrusive dacite.

Thickness

Based on the dips and outcrop thickness, the sequence may be as much as 3000m thick, but is probably much less.

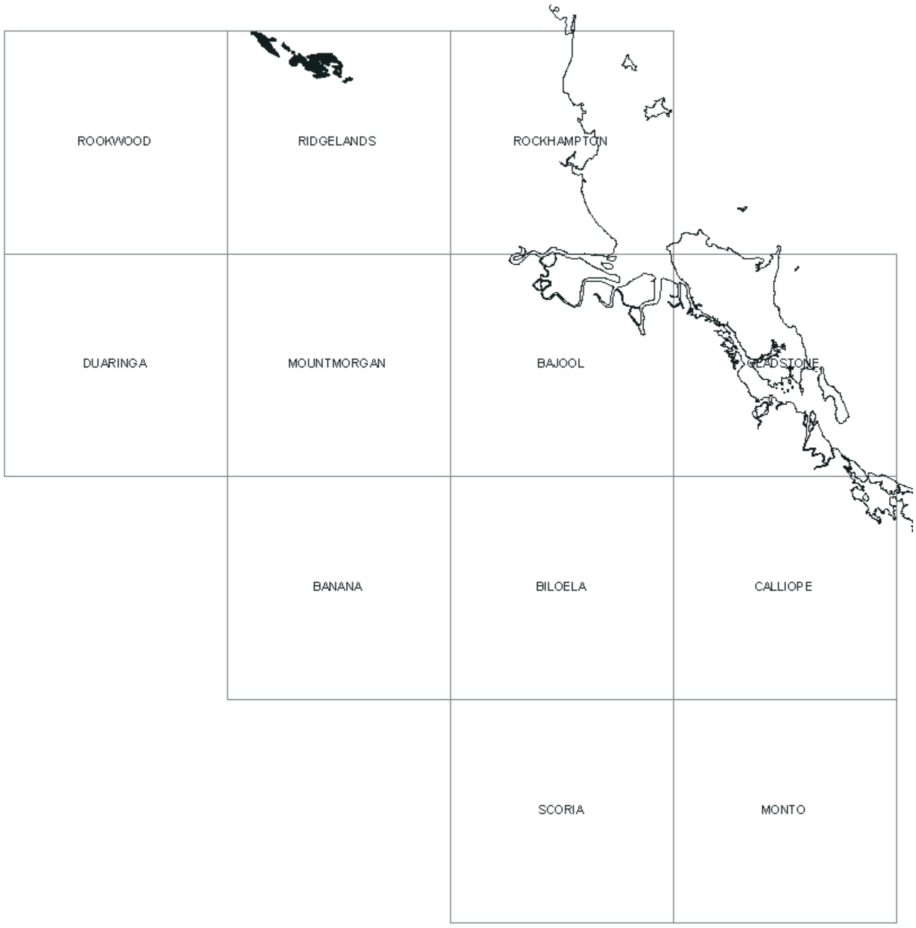


Figure 150. Distribution of the unnamed Permian unit (Pu)

Structure

Dips in the unit are moderate to steep, generally trend north-westerly, and are both to the north-east and south-west. No mesoscopic folding has been observed, but cleavage is locally developed, suggesting some deformation.

Biostratigraphy and age

Kirkegaard & others (1970) reported a sparse fauna of early to mid-Permian age near Mona Vale homestead (late Sakmarian – early Artinskian). U-Pb dating of the correlative Berserker Group is more indicative of a mid-Permian age.

Stratigraphic relations

The unnamed Permian sequence is faulted against the Marlborough Metamorphics. Small lenses of Princhester Serpentinite have been structurally emplaced within the unit, and contacts with larger masses are faulted. The Permian strata are intruded by the Permian Wattlebank Granodiorite and by unnamed Permian gabbro. They are overlain by Quaternary sediments of the Fitzroy River.

Correlation with other units

The rocks were included in the Berserker beds (now Group) by Kirkegaard & others (1970). They are of similar age and lithology.

Economic significance

Small gold deposits are known from the vicinity of Mona Vale homestead. The style of mineralisation is not known.

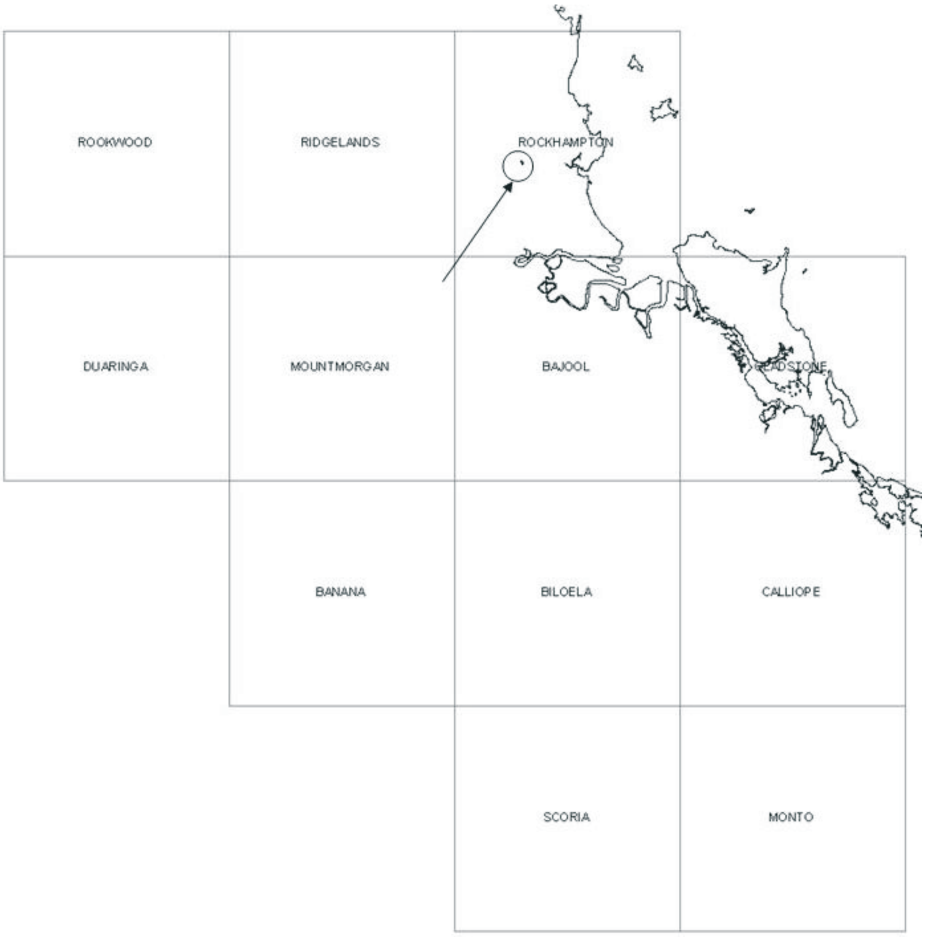


Figure 151. Distribution of the Warminster Formation

LATE PERMIAN

Warminster Formation (Pw)

(S.B.S. Crouch)

Introduction

The Warminster Formation is a unit covering only $300m^2$. It was named and described by Crouch & Parfrey (1998).

Derivation of name

The name Warminster comes from Mount Warminster (Ball, 1910), situated 1.5km north-west of Mount Chalmers mine.

Distribution

The Warminster Formation is located 1km north-west of Mount Chalmers mine, and covers an area of $300m^2$. Outcrop is scarce due to clearing and establishment of pineapple plantations in the area (Figure 151).

Type area

The type area constitutes the $300m^2$ area mapped.

Topographic expression

Because the unit is so restricted, its topographic expression cannot be determined.

Geophysical expression

The Warminster Formation is too small to resolve on airborne geophysics.

Lithology

The sandstone of the Warminster Formation is dark green-grey, fine to coarse, and well sorted. The other rock type is dark green-grey, massive, poorly sorted, matrix supported, granule to pebble, polymictic breccia. The breccia has a siliceous, fine to coarse sandstone matrix with volcanic clasts, siltstone clasts and cherty-looking clasts. The clasts are up to 30mm in diameter and are angular in shape, suggesting little reworking.

The rock is strongly silicified and chloritised.

Environment of deposition

The molluscs and gastropods found in the Warminster Formation are marine and indicate a shallow water environment. The presence of breccia tends to suggest minimal reworking.

Thickness

The Warminster Formation is estimated to be <100m thick.

Structure

The structure of the Warminster Formation is unknown, but is assumed to be flat-lying or gently dipping like the underlying Berserker Group.

Biostratigraphy and age

The unit contains the bivalve *Glyptoleda*, an indeterminate gastropod and brachiopods (Crouch & Parfrey, 1989). The macrofossils found in the Warminster Formation are the same age as those of the upper Barfield Formation, lower Flat Top Formation, Ingelara Formation and lower Peawaddy Formation in the Bowen Basin. The age of the unit is Late Permian.

Stratigraphic relationships

A contact cannot be seen between Warminster Formation and underlying Berserker Group. Fossil evidence suggests that a significant age gap and an unconformable relationship. The relationship between the Warminster Formation and the intrusive units of the region is unknown.

Correlation with other units

The Warminster Formation is equivalent in age to the Moah Creek beds.

Moah Creek beds (Pm)

(S.B.S. Crouch)

The Moah Creek beds are a thick sequence of siltstones, sandstones, and conglomerates which contain Late Permian marine fossils. This unit contains a relatively fine-grained succession of marine sediments indicating a period of passive, thermal subsidence, as well as coarse-grained intervals reflecting the commencement of the foreland thrust loading event associated with the Hunter–Bowen Orogeny.

Derivation of name

The name was introduced by Kirkegaard & others (1970) and is derived from Moah Creek.

Type section

The section from the head of Emu Creek to its junction with the Fitzroy River, and thence west along the river, is designated as the type section.

Figure 152. Distribution of the Moah Creek beds

Distribution

The Moah Creek beds crop out in the area drained by Moah, Emu, and Breakfast Creeks on the common boundaries of the Ridgelands, Mount Morgan, and Rookwood 1:100 000 Sheet areas (Figure 152).

Topographic expression

The Moah Creek beds form low rounded hills, but strike ridges are developed where the sediments are well bedded. Much of the unit is concealed by thin Cainozoic deposits, but it is well exposed in streams.

Geophysical expression

The Moah Creek beds generally give a moderate to high radiometric response from potassium, thorium, and uranium. The magnetic response is low. This unit has a similar response to the adjacent Dinner Creek Conglomerate.

Lithology

Kirkegaard & others (1970) stated that the Moah Creek beds consist dominantly of conglomeratic mudstone and mudstone, with minor conglomerate, coarse-grained lithic sandstone, and fine-grained to medium-grained sandstone. They identified the characteristic rock type of the lower part of the unit as conglomeratic mudstone. This consists of pebbles, cobbles, and boulders of sandstone, mudstone, and limestone in a sandy mudstone matrix. The cobbles and boulders are generally angular and constitute up to 40% of the rock. They include fragments of oolitic limestone and fossiliferous mudstone, indicating that the Rockhampton Group and Lorray Formation were exposed when the sediments were deposited. One block of oolitic limestone in the Fitzroy River is ~45m across, but most are only 0.1–1m across. The mudstone also contains round, tough, fine-grained concretions, many of which contain abundant fine pyrite. Carbonised plant debris occurs in the matrix. The conglomeratic mudstone grades into sandy and in places pebbly mudstone which contains poorly defined sandy streaks and thin sandstone interbeds. Interbeds of coarse-grained sandstone and pebble to cobble conglomerate are also present. The pebbles and cobbles are rounded, up to 15cm in diameter, and

Figure 153. Regularly interbedded sandstone and mudstone, bank of Fitzroy River near Moah Creek, Moah Creek beds

consist mainly of fine-grained sandstone, with some cobbles of vitric volcanics. The pebbles and cobbles are set in a sandy matrix.

The upper part of the unit consists dominantly of mudstone, which is generally finer than that in the lower part. It is blue grey when fresh, greenish grey when weathered, generally massive, slightly indurated, and crumbles into small pieces. Some is sandy, and contains ill-defined sandy bands and sandstone interbeds.

In places in the middle and upper parts of the section, mudstone and fine-grained to medium-grained sandstone are regularly interbedded in beds 5–8cm thick (Figure 153). These sections are identical to the Boomer Formation.

Detailed work at the type section by Fielding & others (1997b) established two alternating facies associations, A and B, containing nine facies. Association A consists dominantly of fine-grained clastic sediments containing intact and articulated bivalves, brachiopods and crinoids, indicating a relatively quiet depositional environment. Sandstone beds are sharp-based, typically graded and display flat laminations and ripple cross-laminations in their upper parts. They were probably deposited from turbidity currents. The upper part of sandstone beds and interbedded siltstones contain trace fossils, mainly *Zoophycos* and *Planolites*. Bioturbation can be ascribed to the *Cruziana* ichnofacies. Plant debris is also abundant, ranging from macerated detritus to large stalks and branches.

Association B comprises conglomerate and diamictite, with one distinctive sandstone facies. The coarse-grained facies have either sand or silt matrix and are transitional into one another. They are clast to matrix-supported and in many beds these also grade laterally into one another. Two populations of clasts are evident: (1) well-rounded pebbles and cobbles of basement rock types (mainly sandstone, chert, volcanic rocks and quartz), and (2) angular to rounded pebbles, boulders and large rafts of internally coherent, intraformational sandstone and interbedded sandstone and siltstone. In some cases these allochthonous masses are folded into cylindrical forms and overturned slump folds. Plant debris is abundant throughout the facies association in the finer-grained beds.

Structure

The Moah Creek beds are exposed in the western limb of the Mount Candlelight Syncline, but are not exposed in the eastern limb. Dips are generally <30°.

Environment of deposition

Fielding & others (1997b) interpreted the entire succession to have formed in an offshore marine environment and stated that the sediments of Associations A and B of the Moah Creek beds exposed at the Fitzroy River crossing were deposited on a gently sloping submarine surface that became periodically unstable. They noted that Association A is typical of mid-Permian offshore marine shelf facies across the greater part of the Bowen Basin, with facies representing suspension fallout and current deposits. The preservation of articulated fossils and absence of any wave- or combined-flow-generated structures suggest a relatively quiet environment. Sharp-bounded sandstones were deposited by tractional currents that flowed westwards. Some of these currents may have been related to turbidity flows. Dropstones may have been introduced from floating ice. The accompanying Association B contains facies indicative of slide, slump or debris flow deposits, debris surge, or high-concentration turbidity flow, all interpreted to be products of an unstable, submarine slope. The relatively fine-grained succession of marine sediments is indicative of a period of passive, thermal subsidence, during which the provenance was from the western (cratonic) margin. The coarse-grained intervals reflect the onset of a foreland thrust loading event to the east, associated with the Hunter–Bowen Orogeny (Baker & others, 1993; Fielding & others, 1997a,b).

Thickness

Kirkegaard & others (1970) stated that the maximum thickness of the unit is ~2100m. Fielding & others (1997b) estimated that the unit was >1500m thick. The thickness is difficult to determine because of the shallow dips and extensive cover. The section studied by Fielding & others (1997b) in the Fitzroy River is ~350m thick, but this only represents part of the unit.

Fossils and age

The upper part of the unit is sparsely fossiliferous in places. Several small *Atomodesma* (*Aphania*) sp. and rare poorly preserved *Strophalosia* have been collected (see McKellar & others, 1970). *Atomodesma* is restricted to Fauna IV of the Bowen Basin, and indicates an early Late Permian age. *Glossopteris* was noted in the field, but not collected.

Stratigraphic relationships

The unit unconformably overlies the Rookwood Volcanics and is overlain by the Dinner Creek Conglomerate. In places the Dinner Creek Conglomerate appears to fill slight erosion channels in the Moah Creek beds, although the relationship between the two is uncertain. Since the unit is not present in the eastern limb of the Mount Candlelight Syncline, it probably onlaps the Carboniferous units subsurface. The Moah Creek beds appear to be equivalent, at least in part, to the Boomer Formation. The Moah Creek beds are intruded by andesitic and trachytic dykes and sills.

Dinner Creek Conglomerate (Pd)

(S.B.S. Crouch)

Introduction

The Dinner Creek Conglomerate consists dominantly of conglomerate with minor sandstone and mudstone interbeds, is interpreted to be Late Permian in age, and is thought to have been deposited in a high energy alluvial environment after uplift of the source area. Its provenance is uncertain.

This description is largely taken from Fleming (1960, 1967) and Kirkegaard & others (1970).

Derivation of name

The name is derived from Dinner Creek, a tributary of Lion Creek which flows through the eastern part of the unit parallel to the strike.

Type section

Fleming (1967) designated the section along a line between the summits of Mount Lion and Native Cat as the type section. This section is in the eastern limb of the Mount Candlelight Syncline.

Figure 154. Distribution of the Dinner Creek Conglomerate

Distribution

The Dinner Creek Conglomerate crops out as a semicircular belt in the flanks of the south-plunging Mount Candlelight Syncline to the north and west of Stanwell (Figure 154).

Topographic expression

The unit forms bare hills cut by a closely spaced dendritic drainage pattern. Its topographic expression is distinct compared to that of the underlying Lorray Formation and Moah Creek beds.

Geophysical expression

The Dinner Creek Conglomerate generally gives a moderate to high response on images of airborne radiometric data. Its magnetic response is low, similar to that of the adjacent Moah Creek beds.

Lithology

The Dinner Creek Conglomerate consists dominantly of conglomerate with minor sandstone and mudstone interbeds. The conglomerate is grey or brown, occurs in massive beds, and consists mainly of well-rounded, ellipsoidal cobbles up to 20cm in diameter, set in a sandy matrix. The clasts are almost wholly fine-grained to medium-grained sandstone, with minor silicified siltstone and very rare quartz. The sandstone cobbles are quartzose to feldspathic in composition, and silicified. The cobbles generally show clast imbrication (Fielding & others, 1997a; Figure 155). Lenticular interbeds of lithic sandstone and carbonaceous mudstone occur in the conglomerate. The basal beds in places consist of interbedded quartz-lithic sandstone and carbonaceous mudstone, and interlaminated carbonaceous mudstone and fine-grained sandstone.

Figure 155. Cobble to boulder conglomerate with rounded, imbricated clasts of sedimentary rocks, 8km north-west of Stanwell, Dinner Creek Conglomerate.

Structure

The Dinner Creek Conglomerate crops out in both limbs of the south-plunging Mount Candlelight Syncline, which has a steeply dipping east limb and more gently dipping west limb. The structure is consistent with deformation in a west-verging fold-thrust belt during the episodic Hunter–Bowen Orogeny. The Dinner Creek Conglomerate is the youngest unit involved in this orogenic event.

Environment of deposition

The abundance of conglomerate in the unit, and the presence of interbedded mudstone with plant fossils, suggests that it was deposited in an alluvial fan environment marginal to a freshwater lake. The overall coarse-grain size points to an uplifted source area, and the excellent rounding of the clasts indicates significant reworking. Although no fossiliferous cobbles have been found, they are similar to the fine-grained to medium-grained sandstones in the Neerkol Formation, which was probably exposed in the source area. Palaeocurrent measurements indicate a source area to the east-north-east (Fielding & others, 1997a).

Thickness

The estimated thickness of the formation, based on dips measured on interbedded sandstone and mudstone, is 1050m.

Fossils and age

Several species of *Glossopteris* have been collected from the unit (see McKellar & others, 1970). These indicate a Permian age for the unit, and its stratigraphic position above the Moah Creek beds indicates a Late Permian age.

Stratigraphic relationships

Although there is no marked angular break between the Dinner Creek Conglomerate and the underlying Lorray Formation, there is an erosional disconformity between them. Small erosion gullies at the top of the Lorray Formation are filled with conglomerate, and varying amounts of this unit have been removed due to faulting and erosion before the Dinner Creek Conglomerate was deposited. The Dinner Creek Conglomerate also overlies the Moah Creek beds with what may be a slight erosional disconformity, marked by gully scour and fill. Fielding & others (1997a) suggested that the Dinner Creek Conglomerate may be a time equivalent of the Upper Permian coal measures of the Bowen Basin to the west.

The Dinner Creek Conglomerate is overlain unconformably by the Triassic Native Cat Andesite, which appears to be essentially flat-lying, although previous workers have suggested that the contact is conformable (Reid & Morton, 1928). This unconformity provides the best constraint from sequences in the Yarrol Province on the age of the Hunter–Bowen Orogeny, but neither of the units above and below has been precisely dated.

Inverness Volcanics (Pvi)

(C.G. Murray)

Introduction

The easternmost part of the Mount Gerard Complex mapped by Dear & others (1971) north of Biloela is a sequence of intermediate volcanics here defined and described as the Inverness Volcanics. Geochemistry suggests that this volcanic assemblage may be co-magmatic with the Late Permian Craiglands Quartz Monzodiorite.

Distribution

The Inverness Volcanics form an equant area ~5km square forming the divide between the upper reaches of Smoky and Back Creeks, 35km north-north-east of Biloela (Figure 156).

Name

The name is from the property Inverness, along Back Creek, on which typical outcrops are exposed.

Type section

The type section is along the upper part of Smoky Creek east from GR255900E 7333100N.

Figure 156. Distribution of the Inverness Volcanics

Topographic expression

The unit forms hilly country, much of it thickly vegetated.

Geophysical expression

The Inverness Volcanics give a moderate radiometric response and are pink on images generated from airborne data, similar to the Craiglands Quartz Monzodiorite and distinct from the much darker tones of the adjacent Smoky beds. The unit has a moderate and somewhat patchy magnetic signature, consistent with magnetic susceptibility readings from the type area. This moderate magnetic signature contrasts markedly with that of the strongly magnetic Craiglands Quartz Monzodiorite to the west, but is similar to an outcrop of this intrusion to the north of the Inverness Volcanics. The prominent arcuate boundary separating the two different magnetic domains is taken as the limit of the Inverness Volcanics.

Lithology

The unit consists solely of intermediate to felsic volcanics and hypabyssal intrusives. The volcanics are dark grey, and locally exhibit fragmental textures and contain abundant small plagioclase phenocrysts. Dark green hornblende and clinopyroxene are less common phenocryst phases, with rare orthopyroxene. Typically the groundmass is very fine-grained, and devitrification textures indicate its original glassy nature. Plagioclase microlites are present in the groundmass in some samples. Compositionally the rocks are close to rhyodacite. Some samples consist of aphanitic andesite-dacite clasts up to 5cm across enclosed in a matrix of porphyritic andesite with plagioclase phenocrysts up to 8mm long.

Small outcrops of hypabyssal intrusive rocks are common and widespread through the Inverness Volcanics. They are uniform in composition, and are quartz monzodiorite similar to the Craiglands Quartz Monzodiorite. They consist of relatively large zoned plagioclase laths and hornblende grains (replacing clinopyroxene) surrounded by a micrographic intergrowth of quartz and K-feldspar, with accessory iron oxides.

Geochemistry

The Inverness Volcanics plot along the same trends as the Craiglands Quartz Monzodiorite for many elements. The similarity is shown in a spidergram (Figure 157), and suggests a co-magmatic relationship, although the intrusive rocks are lower in SiO_2 than the volcanics. The two analyses have moderate $(Th/Nb)_N$ gaps of 26.3 and 30.6, indicating a substantial subduction component, similar to andesites from the Native Cat Andesite. On discrimination diagrams of Pearce & others (1984), the Inverness Volcanics plot in the Volcanic Arc Granite field (Figure 58).

Environment of deposition

The Inverness Volcanics are considered to be subaerial volcanics.

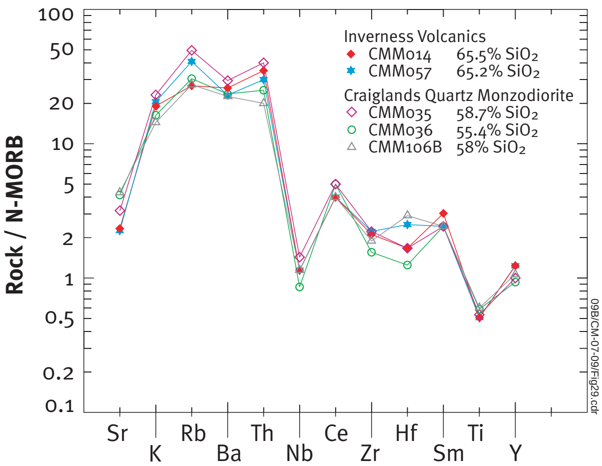


Figure 157. Spidergram plot comparing analyses from the Inverness Volcanics and Craiglands Quartz Monzodiorite. N-MORB values are from Pearce (1983).

Provenance

The restricted outcrop area suggests a local source, presumably related to the Craiglands Quartz Monzodiorite.

Thickness

The thickness is unknown. Total topographic relief is almost 300m, which may represent a minimum thickness assuming the unit is approximately flat-lying.

Structure

The unit consists entirely of massive volcanics. The magnetic pattern suggests that it forms a down-faulted block bounded by an arcuate fault.

Age

The unit is assigned a Late Permian age by correlation with the Craiglands Quartz Monzodiorite.

Stratigraphic relationships

The Inverness Volcanics overlie and/or are faulted against the Youlambie Conglomerate and Smoky beds. They are intruded by the eastern outcrop of the Craiglands Quartz Monzodiorite, although the close association of volcanic and hypabyssal rocks points to a coeval relationship. The unit is also intruded by unnamed granite and gabbro masses of Permian to Triassic age.

Correlation with other units

The only known unit which may correlate with the Inverness Volcanics is the unnamed Permo-Triassic volcanic unit at the Wasp prospect north of Yeppoon, but this could be Triassic. The Inverness Volcanics are interpreted as extrusive equivalents of the Craiglands Quartz Monzodiorite that have been preserved by down-faulting.

PERMO-TRIASSIC

Unnamed Permian-Triassic volcanics (PRv)

(C.G. Murray)

Introduction

Volcanics and associated intrusives in the northern part of the Rockhampton 1:100 000 sheet area at the Wasp prospect were investigated during company exploration, and a summary of their geology was prepared by Willmott & others (1986). They were not mapped during the Yarrol Project.

Distribution

The volcanics crop out in two areas. They occur adjacent to a small granitic intrusion in the headwaters of Cabby Creek 15km north of Yeppoon, and south of Limestone Creek ~10km north-west of Yeppoon. Both outcrops are small, with maximum dimensions of >2km (Figure 158).

Topographic expression

The volcanic rocks form low hills.

Geophysical expression

The unit was not covered by the semi-detailed airborne geophysical survey, and in any case is too small to have a recognisable geophysical expression.

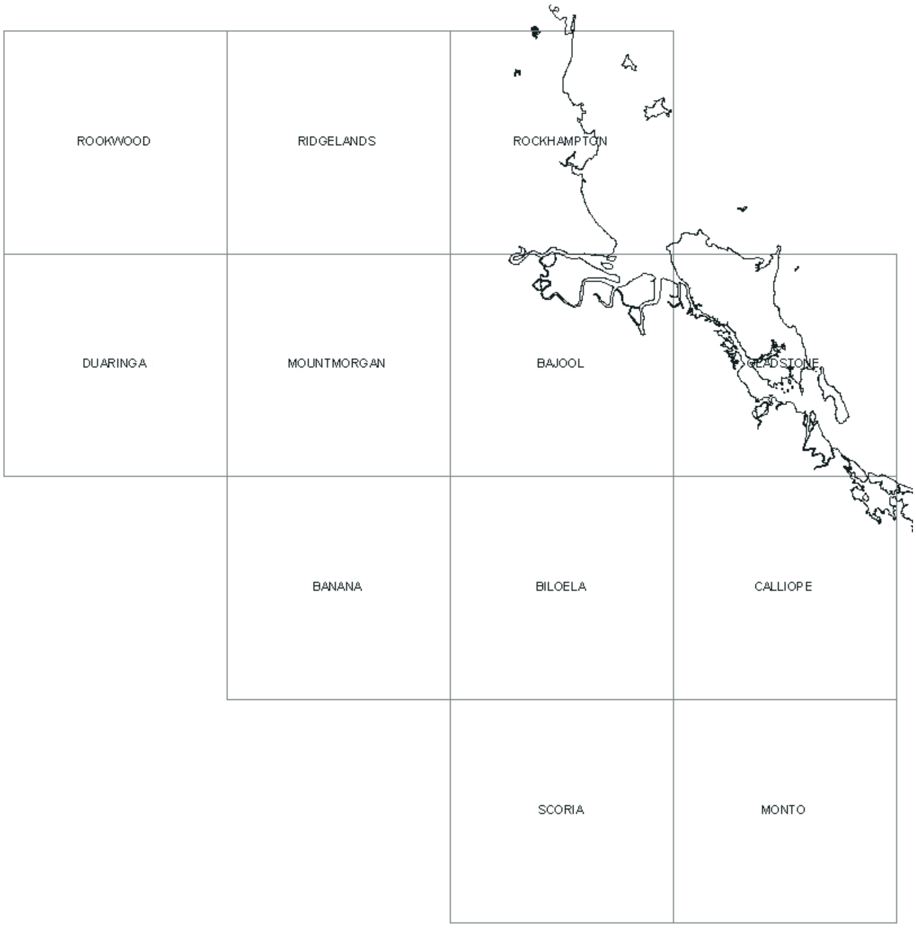


Figure 158. Distribution of the unnamed Permian–Triassic volcanics

Lithology

The rock types at the Wasp prospect are quartz and feldspar-phyric tuffs overlain by felsic tuffs or flows (Ballantyne, 1973). The lowest horizons consist of crystal tuffs with minor interbeds rich in ash and lapilli. These rocks are composed of broken crystals of quartz and plagioclase up to 1cm across in a matrix of fine crystal fragments and ash. This basal sequence is overlain by altered, leucocratic rocks made up of quartz and sericite, which may represent flows or a dyke. A mottled green and grey crystal tuff with broken crystals of quartz and plagioclase as well as rock fragments occurs in low hills to the south and south-east of the granitic intrusive, and a distinctive banded vitrophyre consisting of sanidine spherulites in a chalcedony matrix forms a hill to the south-west (Ballantyne, 1973).

The second occurrence south of Limestone Creek was described as acid to intermediate tuff by Willmott & others (1986).

Thickness

Drilling suggests that the sequence at the Wasp prospect is at least 200m thick.

Structure

The sequence at the Wasp prospect is an approximately flat-lying erosional remnant, with possible faulting to the north-west (Ballantyne, 1973).

Age

At the Wasp prospect, the volcanic rocks appear to be related to the granitic intrusion, which has given Late Triassic K-Ar dates (Ford & others, 1976). The age is therefore likely to be Triassic, but could be as old as Permian.

Stratigraphic relations

At both localities, the unnamed volcanic rocks overlie the Lower Carboniferous Wandilla Formation unconformably. They are intruded by and possibly gradational with the granitoid.

Correlation with other units

The composition of the volcanic sequence and its flat-lying attitude are similar to those of other Triassic volcanic units in the Yarrol Project area.

Economic significance

A widespread zone of alteration, centred over the granitoid, is associated with subeconomic pyrite, chalcopyrite and molybdenite mineralisation of porphyry style. The mineralisation is best developed in breccia zones around the margins of the granitoid.

TRIASSIC

Cynthia beds (Rvy)

(C.G. Murray)

Introduction

The name Cynthia beds was introduced by Whitaker & others (1974) for a sequence of sediments and volcanics in fault blocks along the eastern side of the Mulgildie Basin. The rocks were originally described by Dear & others (1971) as part of the Muncon Volcanics. Gray (1972a) named the Abercorn Trough which partly underlies the Mulgildie Basin, and de Jersey (1972) determined that the rocks were of Middle Triassic age and equivalent to the Esk Trough sequence.

Distribution

The Cynthia beds form two fault blocks up to 2km wide to the east of the Mulgildie Fault, extending for 16km south-south-east from Monto to the southern boundary of the Monto 1:100 000 sheet area (Figure 159).

Topographic expression

The unit forms subdued topography, as the rocks weather readily. On airphotos, prominent strike ridges reflect the steeply dipping nature of the rocks.

Geophysical expression

On images derived from radiometric data, the unit shows a moderate to strong response, characterised by reddish to white colours.

The Cynthia beds have a moderate magnetic signature which is especially evident on the first vertical derivative.

Lithology

The largest fault-bounded outcrop south-south-east of Monto is composed mainly of khaki-coloured conglomerate which varies in grain size and grades into feldspathic sandstone containing biotite flakes and some plant detritus. The clasts range in size up to ~30cm across, and are dominantly volcanic (porphyitic andesite and flow banded rhyolite) with a subordinate component of metasediments from the basement rocks. Sorting is poor, and most of the conglomerate is clast-supported. The sequence is quartz-poor.

Dear & others (1971) described the strata in the fault blocks as pink cobble and boulder conglomerate consisting of intermediate lava clasts in a tuffaceous matrix, overlain by volcaniclastic sandstone, in turn overlain by thick beds of agglomerate and lapilli tuff in which green and purple intermediate lava fragments are enclosed in a matrix of feldspar and lithic grains. Interbedded in the overall sequence are volcaniclastic sandstone and mudstone and minor andesitic flows.

Figure 159. Distribution of the Cynthia beds

The sequence in the APE Mulgildie 1 well is richer in andesitic lava and pyroclastics, although the proportion of primary volcanics may have been overestimated if andesitic boulder beds are present. In addition, sediments including sandstone and mudstone are present. Their composition suggests a mixed provenance, with derivation from both a granitic and an intermediate volcanic source.

The bottom 65m in GSQ Monto 1-2R are regarded as part of the Cynthia beds, which here consist of volcaniclastic, labile sandstone, in part calcareous and pebbly, carbonaceous shale, mudstone, siltstone and rare coal (Gray, 1972a). Leaching has affected the topmost 25m.

Further south in the Mundubbera 1:250 000 sheet area, the dominant lithology in outcrop is boulder conglomerate (Whitaker & others, 1974).

Thickness

The thickness of the Cynthia beds in APE Mulgildie 1 is 180m. In APE Abercorn 1 the unit is much thicker, almost 1500m (Whitaker & others, 1974). This reduction in the Yarrol Project area could be either the result of thinning to the north, or incomplete preservation.

Structure

In outcrop, dips are difficult to determine because of the conglomeratic lithology, but the presence of prominent strike ridges on airphotos indicates that the unit dips at moderate to steep angles to the west. A moderate dip is characteristic of strata in APE Mulgildie 1, where the rocks dip at about 40° (Dear & others, 1971). The sequence does not appear to be folded.

Biostratigraphy and age

De Jersey (1972) discussed spores recovered from the three wells, APE Abercorn 1 and Mulgildie 1 and GSQ Monto 1-2R. All indicated a general Triassic aspect, with the most precise indicator of a Middle Triassic age from GSQ Monto 1-2R. Because this represents only the top of the Triassic sequence, de Jersey (1972) suggested that lower strata in APE Abercorn 1 and Mulgildie 1 could be older.

Stratigraphic relationships

On the Monto 1:100 000 sheet area, the Cynthia beds are entirely fault-bounded. To the east they are faulted against the Lower Carboniferous Rockhampton Group and the uppermost Carboniferous to Lower Permian Youlambie Conglomerate. To the west they are faulted against the Jurassic Hutton Sandstone and Mulgildie Coal Measures of the Mulgildie Basin. In drillholes APE Mulgildie 1 and GSQ Monto 1-2R, the Jurassic sequence of the Mulgildie Basin is interpreted as unconformable on the Triassic Cynthia beds.

Correlation with other units

The Cynthia beds may correlate with other sedimentary sequences at the base of Triassic volcanics in the Yarrol Project area, although they are much thicker. They correlate best in lithology, structure and age with the Toogoolawah Group, the fill of the Early to Middle Triassic Esk Trough.

Native Cat Andesite (Rvn)

(A.D.C. Robertson & C.G. Murray)

Introduction

Native Cat Andesite is the name given to a sequence of dominantly andesitic volcanics by Kirkegaard & others (1970) that crops out north-west of Stanwell. They derived the name from the Native Cat Range, which they nominated as the type area. Reid & Morton (1928) considered these rocks to be equivalent to the “Lower Bowen Volcanics”, of Early Permian age. Although Kirkegaard & others (1970) did not date the unit, they assigned a Triassic age.

Distribution

The volcanics are exposed in an east-west belt 20km long and up to 10km across (Figure 160), extending from near Stanwell in the east to Black Mountain in the west.

Topographic expression

The Native Cat Andesite forms the dissected plateau of the Native Cat Range that is up to 250m above the surrounding country. The bounding scarps are densely vegetated. The plateau is only lightly to moderately vegetated but is progressively being populated by rubber vine, which has made much of the cleared area along watercourses now useless for agricultural pursuits.

Geophysical expression

Images derived from airborne radiometric data are deep red to pink in colour, indicating a relatively high potassium content for the Native Cat Andesite. The pink areas are thought to represent more felsic volcanics than usual. Several small areas along the Native Cat Range that give a strong radiometric response have been mapped as outcrops of Upper Cretaceous Mount Salmon Volcanics that unconformably overlie the Native Cat Andesite.

The unit has a moderately strong magnetic signature that contrasts with the very low response of the sedimentary sequences to east and west.

Lithology

The volcanics consist dominantly of andesitic flows, andesitic tuff, lapilli tuff, trachyandesite, andesitic to rhyolitic breccias, and minor rhyolitic flows. The andesitic flows are dark green to greenish grey to purplish, fine to medium grained and generally porphyritic, with phenocrysts dominantly of andesine, but also augite and sparse hornblende and iron oxide, set in a fine grained groundmass of small andesine laths, clinopyroxene grains, iron oxide and brown to green, interstitial, devitrified glass. The feldspar exhibits varying stages of alteration to clay minerals while the clinopyroxene alters to granular epidote and fibrous chlorite. Secondary calcite is widespread, locally forming lenses, and the rocks are cut by quartz veins.

Within a hornfelsed zone surrounding a granitic intrusion at Kalapa North, the clinopyroxene has been altered to a brown hornblende. At GR216414 7401499 on the Ridgelands 1:100 000 Sheet area, a very coarse

Figure 160. Distribution of the Native Cat Andesite

grained, purplish grey to green andesite has been exposed in a small roadside scrape. Altered anhedral to subhedral plagioclase (?andesine) phenocrysts range in size from 8–55mm. The accompanying augite phenocrysts rarely exceed 10mm in length, with the majority ranging from 4–6mm. Most of the large feldspar phenocrysts have been partly or wholly replaced by a mosaic of granular epidote and calcite. The augite phenocrysts exhibit strained extinction, and some crystals are fractured. Rare vesicles are filled with fibrous calcite. The groundmass contains feldspar laths, granular clinopyroxene and iron oxide as well as minor brown glass. The feldspar in the groundmass exhibits a weak flow alignment.

Breccia occurs throughout the unit but appears to be more common towards the top of the sequence. The most common breccia type consists of angular fragments of grey-black to green porphyritic andesite up to 8cm across, set in a fine-grained, reddish-brown matrix which contains devitrified volcanic glass and glass shards. Accompanying the andesitic clasts are minor amounts of quartzose sandstone, quartzite and rhyolite. Thin beds of rhyolitic breccia and rhyolitic ignimbrite occur towards the top of the unit. Clasts in this breccia are not matrix-supported as in the andesitic breccia, are dominantly rhyolite and andesite in the ratio of approximately 70:30, and are up to 5cm across. In places reddish-brown and purple lapilli tuff and fine-grained crystal tuff occur in beds 8–15cm thick (Kirkegaard & others, 1970).

Trachyandesites are reddish-brown to dark grey rocks. They are dominantly flows, but some may be vitric crystal tuff. They consist of ?oligoclase crystals set in a devitrified glassy matrix, which locally shows perlitic cracks and contains iron oxide rods. Some are vesicular with the vesicles filled with chlorite, calcite, and a zeolite.

The rhyolite is fine grained, white to grey in colour and usually strongly weathered with most of the feldspar altered to clay minerals. Flows vary in thickness from 0.5–3m and appear to be more prevalent towards the top of the sequence.

A small outcrop in Neerkol Creek near old clay pits south-west of Stanwell is a volcanic conglomerate with clasts ranging in composition from basalt to microgranite (Figure 161).

Figure 161. Volcanic conglomerate, Neerkol Creek, Native Cat Andesite

Geochemistry

In the Total Alkali versus Silica (TAS) classification, the composition of the rock types ranges from basalt to rhyolite. Basalts and andesites fall within the calc-alkali basalt field on standard discriminant diagrams of Pearce & Cann (1973), Mullen (1983), Pearce (1982) and Cabanis & Lecolle (1989) (Figure 118). Multi-element and rare earth element plots (Figure 162) show that analyses from the Native Cat Andesite are similar to basaltic andesite and andesite from volcanoes between 35.25° and 36°S within the Southern Volcanic Zone of the Andes, implying similar magmatic sources and crustal thickness. Their geochemistry is therefore consistent with a subduction-related origin at a continental margin where crustal thickness is about 45–50km. Ce/Y values suggest a much thinner crust of about 30km (Mantle & Collins, 2008). One sample from the Native Cat Andesite is almost identical to a sample from the Bobby Volcanics near Many Peaks.

Small outcrops of rhyolite on the Native Cat Range, previously included in the Mount Salmon Volcanics, are now regarded as part of the Native Cat Andesite. Samples have comparatively low Nb, and moderate to large values of $(Th/Nb)_N$ from 26.25 to 32.1 (Figure 163). On discriminant plots of Pearce & others (1984), these samples fall close to the volcanic arc boundary (Figure 58).

Thickness

Assuming that the unit is flat-lying, a thickness of about 300m has been estimated for the Native Cat Andesite.

Structure

The volcanics are considered to be essentially flat-lying, although they may have low dips towards the axis of the Mount Candlelight Syncline. They unconformably overlie the Dinner Creek Conglomerate.

Biostratigraphy and Age

No fossils have been found within the unit. Kirkegaard & others (1970) considered the Native Cat Andesite to be a volcanic pile that post-dated the Late Permian to Early Triassic Hunter-Bowen Orogeny. On the basis of

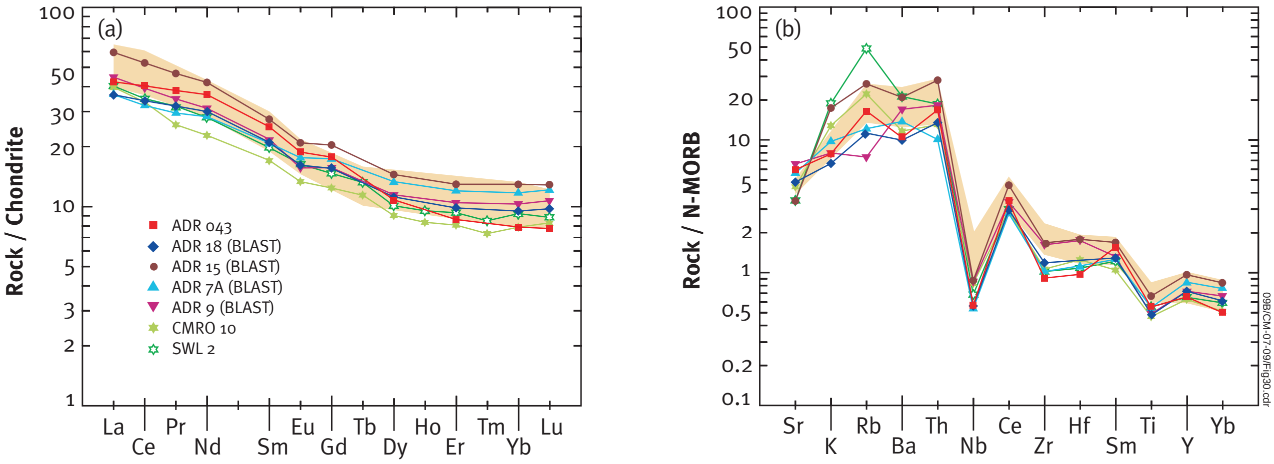


Figure 162. (a) REE and (b) spidergram plots comparing basalts and andesites of the Native Cat Andesite with 18 basaltic andesites and andesites from Planchon, Azufre, Peteroa and Tatara volcanoes in the Southern Volcanic Zone of the Andes shown as a shaded pattern (Ferguson & others, 1992; Tormey & others, 1995). Chondritic REE values are from Sun & McDonough (1989), and N-MORB from Pearce (1983).

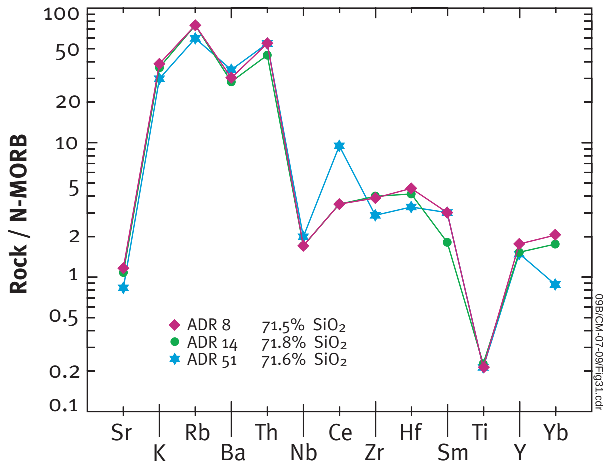


Figure 163. Spidergram plot of rhyolites. N-MORB values are from Pearce (1983)

its stratigraphic relationships, they suggested a Triassic age. This argument is reinforced by the similar geochemistry of this unit to the Bobby Volcanics which contain Lower to Middle Triassic plants (Dear & others, 1971).

Attempts have been made to radiometrically date the Native Cat Andesite, but with little success due to pervasive alteration.

Stratigraphic relationships

The Native Cat Andesite is interpreted to overlie the Dinner Creek Conglomerate with angular unconformity. To the north and south-west, it is overlain unconformably by the Mount Salmon Volcanics, Alton Downs Basalt, and unnamed basalt all of Late Cretaceous age. In the south-east, the Native Cat Andesite is separated from the Lower Cretaceous Stanwell Formation by the Stanwell Fault. Near Kalapa North, a composite granite-granodiorite body has intruded the andesite, forming a hornfelsed zone up to 400m from the contact.

Correlation with other units

The Native Cat Andesite is similar in geological setting and geochemistry to the Muncon Volcanics and unnamed Triassic volcanics in the Cania and Many Peaks areas (Dear, 1968; Dear & others, 1971).

Economic significance

Gold was first worked along Native Cat Creek, at Golden Spur (GR209200 7399250) and in Golden Spur Gully (GR210646 7401072) in 1864 (Bird, 1904), where rich alluvial and eluvial deposits were found. The source of this gold is attributed to the composite igneous body that intrudes the Native Cat Andesite at Kalapa

North. Around the margin of the intrusion, the andesite has been hornfelsed and hydrothermally altered. Along the intrusive contact, gold occurs in association with iron and copper sulphides in chloritised andesite. At the Native Cat mine (GR210458 7399054), the gold occurs in a pyritic quartz reef adjacent to the intrusive contact. In the vicinity of Golden Spur, quartz-sericite-pyrite alteration of the andesite is common. Supergene enrichment in poddy hematitic shear zones within the hornfelsed andesite is the most likely source of the rich gold deposits at Golden Spur and Golden Spur Gully.

Muncon Volcanics (Rvm)

(A.D.C. Robertson & C.G. Murray)

Introduction

The name Muncon Volcanics was established by Dear (1968) for a succession of siltstone, basic to intermediate lava, and "agglomerate" of Triassic age unconformably overlying Palaeozoic rocks near Mount Muncon, south of Cania. Previously the name had been used informally by McKellar (1967) for the basal part of a volcanic sequence 30km north-east of Monto (now mapped as unnamed Triassic volcanics).

Distribution

The Muncon Volcanics are exposed beneath escarpments of Lower Jurassic Precipice Sandstone to the south and south-west of Cania along the western side of Three Moon Creek, covering a total area 11km by 9km (Figure 164).

Topographic expression

A very subdued topography develops on the unit, as the rocks weather easily, and dips are mostly low. Dips can be moderately steep adjacent to faults, resulting in the development of low strike ridges. South of the wall

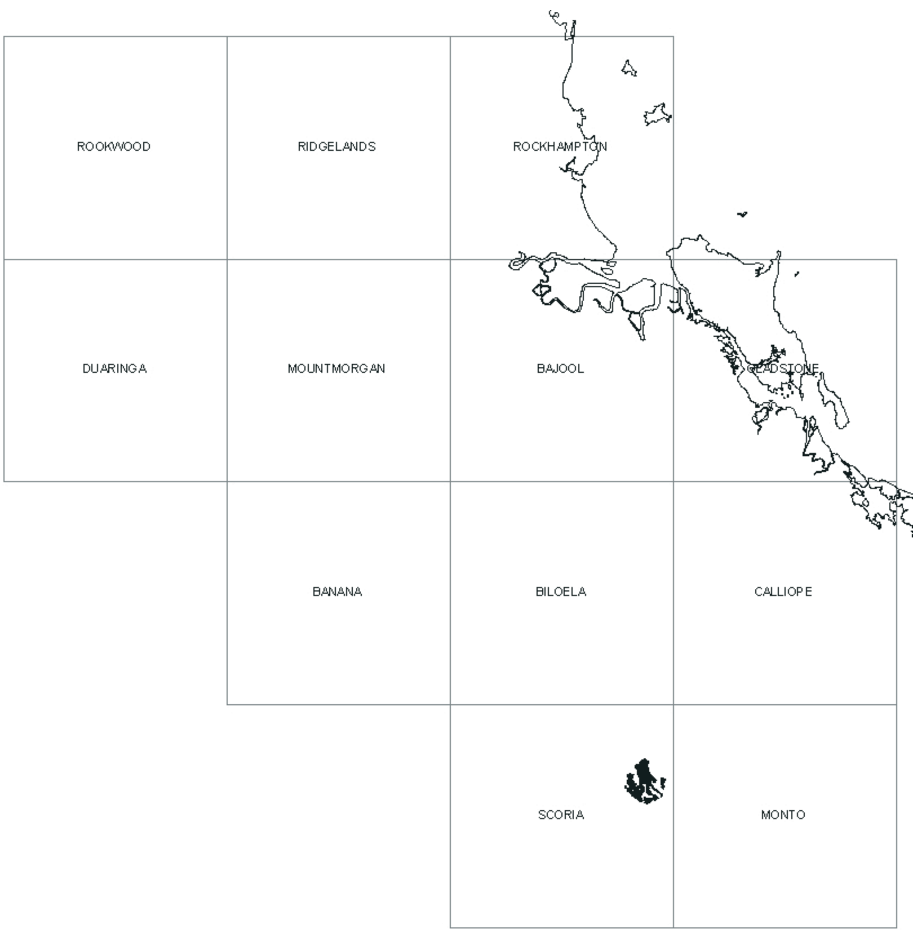


Figure 164. Distribution of the Muncon Volcanics

of Cania dam, the volcanic unit weathers to low rounded hills where it unconformably overlies granodiorite of the Wingfield Granite.

Geophysical expression

The magnetic signature of the Muncon Volcanics is masked, to some extent, by more strongly magnetised rock types such as diorite and granodiorite of the northern end of the Rawbelle Batholith that underlie the volcanics in the Cania area. The response is somewhat more 'noisy' than that of the batholith to the south, suggesting the presence of moderately thick basaltic flows. Magnetic susceptibility readings were not systematically made but one outcrop of andesite had values from 500 to 2200 x 10^{-5} SI units.

The radiometric response is low in all three channels, although the potassium response is slightly elevated, so that on composite images the unit is represented by a deep purplish hue. The response is moderately low compared to the much higher response shown by the Wingfield Granite that the volcanics partly overlie. Where debris slides from the overlying Precipice Sandstone cover the volcanics, the higher thorium response of the sandstone masks the radiometric response from the volcanics.

Lithology

The lower part of the unit comprises interbedded volcaniclastic pebble conglomerate, tuffaceous sandstone, siltstone and minor basaltic to andesitic lava. The volcaniclastic pebble conglomerate is composed of poorly sorted, subrounded andesitic volcanic clasts in a greenish grey volcaniclastic matrix. The tuffaceous sandstone is greenish grey, fine to medium grained and composed of quartz, feldspar, biotite and minor fragments of andesitic lava. Dark grey to off-white tuffaceous siltstone has a similar mineralogical make up to the sandstone, with the addition of muscovite and varying proportions of carbonaceous detritus. Groundmass cement includes silica, calcite and iron oxides. Iron oxide is also commonly concentrated along bedding planes and joints in the sediments and along joint planes. On a branch of Oaky Creek at GR292575 7270970, the sandstone and siltstone contain plant remains.

The lavas are dominantly of basaltic andesite composition. They are vesicular, fine-grained, and dark grey to black in colour. Some contain euhedral to subhedral olivine phenocrysts, now largely to entirely altered to iddingsite or bowlingite and opaque oxides, in a groundmass of plagioclase microlites, interstitial augite, and very small opaque grains. The groundmass is commonly extremely altered to a felted matt of clay minerals, calcite, hydrated iron oxides and chloritic minerals. Other samples contain augite phenocrysts as glomeroporphyritic aggregates, and rare altered orthopyroxene, enclosed by strongly aligned plagioclase laths, interstitial to prismatic augite grains, and opaques. The augite is typically fresh, but locally exhibits varying degrees of alteration to a green chloritic mineral, calcite and opaque oxides. These two types of basaltic andesite have the same chemical composition. Andesitic lavas are similar to the basaltic andesite in both texture and mineral assemblage with the exception that augitic pyroxene becomes the dominant phenocryst phase and vesiculation of the lava is not as common. At the road crossing on Oaky Creek (GR291265 7271845) south of The Nobbies, relatively fresh basaltic andesite has been exposed in a low road cutting.

Higher in the sequence basaltic to andesitic lavas dominate, and are accompanied by tuffs, volcanic breccia and minor pebble conglomerate. This conglomerate contains numerous clasts of white siltstone. The basaltic to andesitic lavas are similar in mineralogical composition to those previously described with the exception that some andesitic lava contains both biotite and rare hornblende as accessory minerals. The hornblende has been derived from the alteration of the augitic pyroxene during final cooling of the lava. Brown to purplish grey lithic tuff forms the majority of the tephra deposits. It contains angular to subrounded andesitic rock fragments along with minor angular grey siltstone fragments in an extremely fine grained matrix. The volcanic breccias appear to occur interspersed throughout the upper part of the sequence. At GR289250 7273700, clast supported breccia with clasts to 30cm crops out as part of a low knoll. Many of the clasts are basaltic to andesitic and exhibit varying degree of vesiculation and flow banding. Minor sedimentary clasts, possibly derived from the Youlambie Conglomerate, also occur. The matrix material ranges from lapilli to fine ash. Near the very top of the exposed sequence, thin grey trachytic flows occur. These are composed of varying proportions of plagioclase, potash feldspar and quartz with accompanying accessory biotite, iron oxide and minor hornblende.

Geochemistry

The Total Alkali versus Silica (TAS) chemical classification of the analysed lavas from the Muncon Volcanics indicates that they are basaltic andesites. They fall within the calc-alkali basalt field on standard discriminant diagrams of Pearce & Cann (1973), Mullen (1983), Pearce (1982) and Cabanis & Lecolle (1989) (Figure 118). Multi-element and rare earth element plots (Figure 165) show that analyses from the Muncon Volcanics are remarkably similar to basaltic andesite and andesite from Tatara volcano within the Southern Volcanic Zone

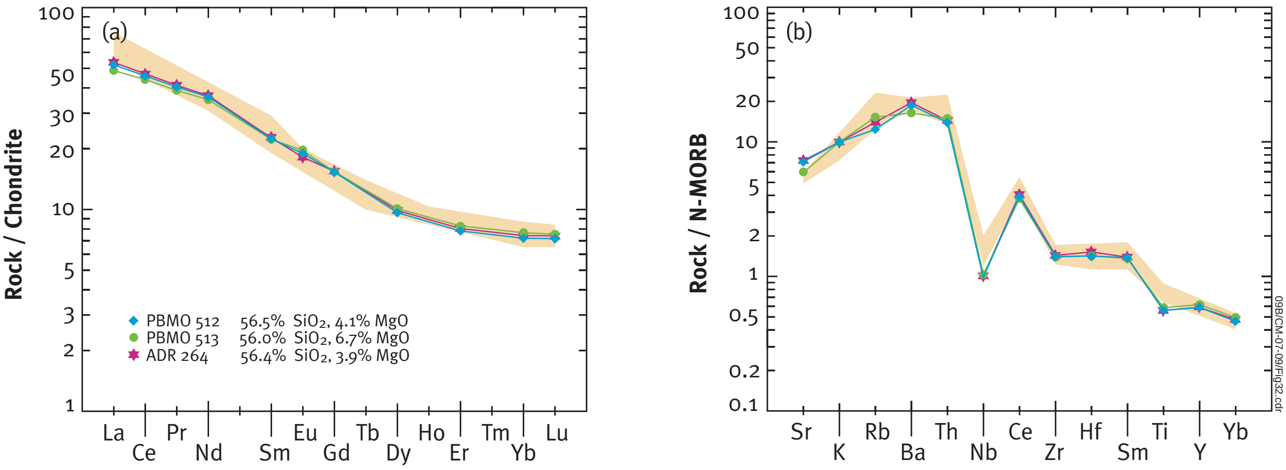


Figure 165. (a) REE and (b) spidergram plots comparing andesites of the Muncon Volcanics with 8 basaltic andesites and andesites from Laguna del Maule and Tatara volcanoes in the Southern Volcanic Zone of the Andes shown as a shaded pattern (Frey & others, 1984; Ferguson & others, 1992). Chondritic REE values are from Sun & McDonough (1989), and N-MORB from Pearce (1983).

of the Andes (Ferguson & others, 1992), implying similar magmatic sources and crustal thickness (about 45km). Ce/Y values suggest that the crust was thinner, closer to 35km (Mantle & Collins, 2008).

Thickness

The thickness is difficult to determine reliably due to variability of dip and the poor outcrop. Dear (1968) noted that at least 150m of sequence is exposed south-west of Cania, which is a minimum thickness given that the top of the unit is eroded.

Structure

Dips are generally shallow, although they can be as great as 40° in the vicinity of faults. Small scale kink folding has been observed in the vicinity of faults.

Biostratigraphy and age

A collection of plant remains was made from near the base of the formation on Oaky Creek south-south-west of Cania (Scoria GR292573 7270970). Included in the assemblage are *Cladophlebis* sp., *Dicroidium odontopteroides*, *Dicroidium odontopteroides* var. *lancifolium*, *Czekanowskia* cf. *tenuifolia*, *Linguifolium* sp., *Elatocladus* sp., and indeterminate equisetalean plant stems. Dear & others (1971) found similar material and suggested that the flora indicated an Early or Middle Triassic age.

Stratigraphic relationships

The Muncon Volcanics unconformably overlie and/or are faulted against the Lower Carboniferous Rockhampton Group and the Lower Permian Youlambie Conglomerate to the north and north-east. They also overlie the Wingfield Granite and small unnamed diorite bodies that intrude the Youlambie Conglomerate and Rockhampton Group. They are unconformably overlain by the Jurassic Precipice Sandstone to the south and west. Towards the headwaters of Paddys Gully, the Muncon Volcanics have been intruded by a xenolith-bearing Tertiary basalt dyke.

A sample collected from a road cutting 3.2km south-west of Cania dam (GR293365 7270400), 400m from a mapped contact with the Wingfield Granite, has been metamorphosed to a fine grained white mica-biotite-plagioclase hornfels. The biotite is greenish, and the white mica may be phengite or paragonite rather than muscovite. A blady opaque mineral (?ilmenite) is also present. The hornfels has largely retained the original basaltic andesite chemistry. K-Ar dates of 240±5Ma (biotite) and 247±5Ma (hornblende) from the southern part of the granite outcrop suggest that the volcanic sequence is younger, and the nature of the intrusive body causing the metamorphism is unknown.

Correlation with other units

Dear & others (1971) applied the name Muncon Volcanics to several outcrops of Triassic volcanics throughout the Monto 1:250 000 sheet area. These all display some variability in stratigraphic sequence, the proportion of different rock types, and geochemistry, and have been allocated to separate units. However, they are considered to be of similar age and tectonic setting, as is the Native Cat Andesite west of Rockhampton.

A volcanic interval in the APN Mulgildie 1 well near Mulgildie was included in the Muncon Volcanics by Dear & others (1971), but this is now regarded as part of the Cynthia beds, also of Triassic age (Whitaker & others, 1974).

Bobby Volcanics (Rvb)

(C.G. Murray)

Introduction

Dear & others (1971) included almost all the Triassic volcanic units on the Monto 1:250 000 sheet area in the Muncon Volcanics. However, the Muncon Volcanics within their type area south-west of Cania differ from sequences elsewhere. Intermediate to mafic volcanics in the Bobby Range extend eastwards into the Bundaberg 1:250 000 sheet area, where they have been named the Bobby Volcanics.

Distribution

The Bobby Volcanics lie east of Many Peaks, and extend from the Bobby Range in the north to the headwaters of the Kolan River in the south, and eastwards onto the Bundaberg 1:250 000 sheet area. In the Yarrol Project area, the main outcrop area is about 16km in a north-south direction, and about 11.5km across. A smaller outcrop area is interpreted to the north, along the crest of the Many Peaks Range, but was not field checked (Figure 166).

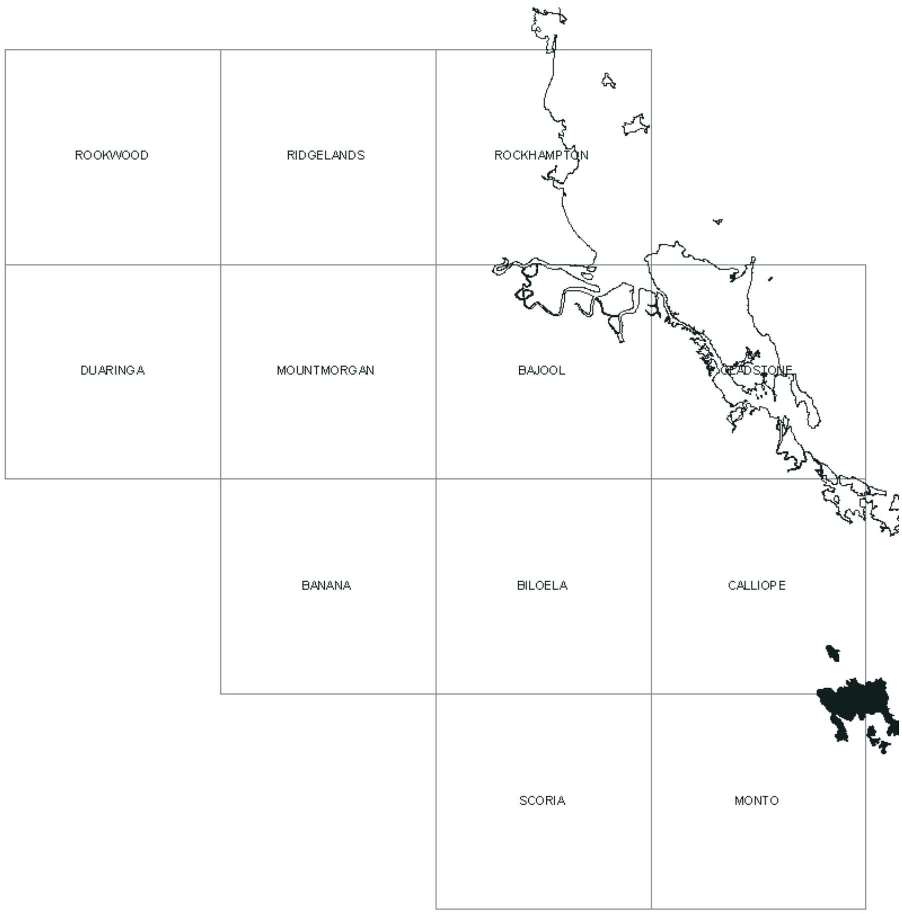


Figure 166. Distribution of the Bobby Volcanics

Topographic expression

Most of the volcanics form elevated areas that are thickly forested, with local steep slopes. A feature of the Bobby Range is the presence of deeply incised creeks controlled by linear joint patterns trending north-north-west to north-north-east.

Geophysical expression

As expected, the radiometric signature of the volcanics is controlled by the lithology. On images generated from airborne data, very dark to black tones are characteristic of areas where mafic volcanics are dominant.

Typically the volcanics show a weak to moderate magnetic signature dominated by short wavelengths, so that they are most readily distinguished on images displaying the first vertical derivative of the magnetic intensity.

Lithology

Dear & others (1971) describe the sequence in the Bobby Range as basal volcaniclastic sandstone overlain by massive agglomerate and thick tuff beds, with andesite (some of which is actually olivine basalt or basaltic andesite) and lapilli tuff at the top.

The basalt or basaltic andesite contains altered olivine microphenocrysts in a pilotaxitic groundmass of plagioclase laths, opaque grains, and secondary chlorite. Andesites mainly consist of plagioclase microphenocrysts in a felted groundmass of plagioclase microlites, altered glass, equant opaque grains, and small augite prisms, but some have microphenocrysts of euhedral augite and lesser hypersthene. The basalts and andesites are typically rich in amygdales (Figure 167). Clasts in the sedimentary rocks are dominantly of intermediate to mafic lavas like those of the associated primary volcanics, with fairly abundant plagioclase grains, and sparse quartz some of which is of plutonic origin (Figure 168). A secondary cement of calcite and green chlorite is present locally. Fine grained sediments are present locally (Figure 169).

The southernmost outcrop of Bobby Volcanics contains more felsic compositions, including rhyolite and rhyolitic ignimbrite, and has been mapped as a separate phase which may underlie the intermediate to mafic volcanics (Figure 170).

Geochemistry

Basaltic andesites and andesites of the Bobby Volcanics fall within the calc-alkali basalt field on standard discriminant diagrams of Pearce & Cann (1973), Mullen (1983), Pearce (1982) and Cabanis & Lecolle (1989) (Figure 118). Multi-element and rare earth element plots (Figure 171) show that analyses are similar to mafic rocks from volcanoes within the Southern Volcanic Zone of the Andes (Hickey-Vargas & others, 1989; Ferguson & others, 1992; Tormey & others, 1995), implying similar magmatic sources and processes. Their geochemistry is therefore consistent with a subduction-related origin at a continental margin where crustal thickness falls within the range 45 to 50km. Ce/Y values are similar to those of the Native Cat Andesite, and suggest a much thinner crust of about 30km (Mantle & Collins, 2008).

Environment of deposition

The volcanics and associated sedimentary rocks were deposited subaerially.

Thickness

Assuming that the strata are flat-lying or gently dipping, the maximum thickness estimated from topographic relationships may be as much as 400m, but 250m is probably more realistic.

Structure

The volcanics are considered to be essentially flat-lying, although the topography of the main outcrop area suggests the possibility that they dip gently towards the south.

Biostratigraphy and age

Plant fossils from the Bobby Range area were included in a list by Dear & others (1971), who assigned an Early or Middle Triassic age.

Figure 167. Amygdaloidal basalt or basaltic andesite, Bobby Volcanics. The smaller amygdales are composed of chlorite, whereas the larger ones consist of a calcite core and a rim of chlorite.

Figure 168. Conglomerate with clasts dominantly of mafic volcanics, Bobby Volcanics

Figure 169. Flat-lying fine grained sediments with thin air fall tuffs, Bobby Volcanics

Figure 170. Ignimbrite from the headwaters of the Kolan River, 10km south-east of Many Peaks, Bobby Volcanics

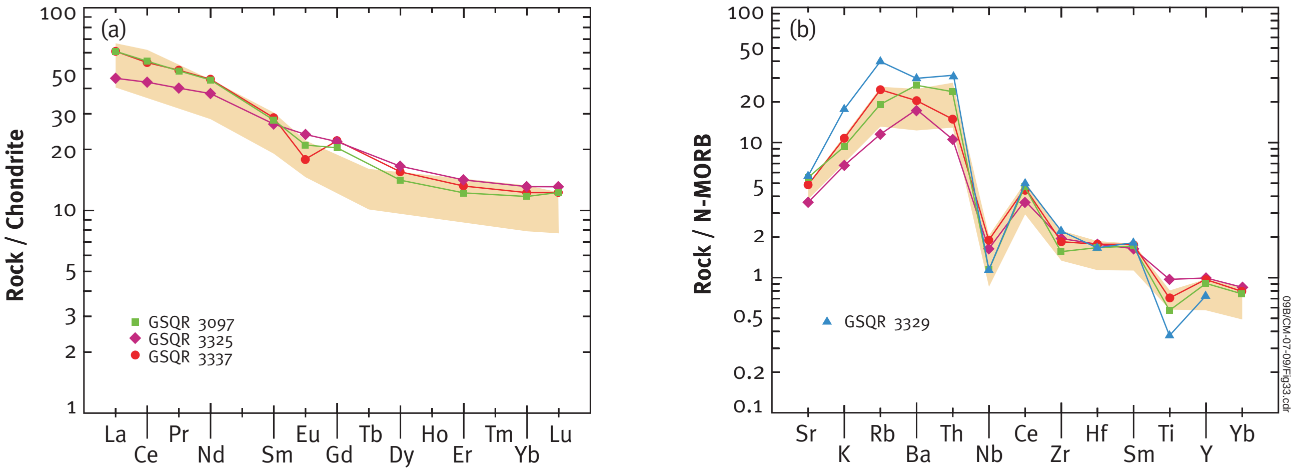


Figure 171. (a) REE and (b) spidergram plots comparing basalts and andesites of the Bobby Volcanics with 18 basaltic andesites and andesites from Planchon, Azufre, Peteroa and Tatara volcanoes in the Southern Volcanic Zone of the Andes shown as a shaded pattern (Ferguson & others, 1992; Tormey & others, 1995). Chondritic REE values are from Sun & McDonough (1989), and N-MORB from Pearce (1983).

Stratigraphic relationships

The Bobby Volcanics unconformably overlie the Lower Carboniferous Wandilla Formation. They also overlie or appear to overlie the Bootoo Ceek Granite, Miriam Vale Granodiorite, Castletower Granite, and unnamed gabbro. Their relationship with unnamed granite forming Mount Beatrice is unknown. They are faulted against the Tertiary Nagoorin beds of the Nagoorin Graben, although this relationship is obscured by Quaternary sediments.

Correlation with other units

The interpretation of the gross stratigraphic succession in the Triassic volcanics in the Many Peaks – Kalpowar district is based entirely on correlation of three separate sequences of felsic volcanics. Validation of the relationships obviously requires a targeted field mapping exercise.

Coulston Volcanics (Rvc)

(C.G. Murray)

Introduction

The Coulston Volcanics is a newly mapped unit of Triassic volcanics overlying the Miriam Vale Granodiorite.

Distribution

The volcanics form a small arcuate east-west trending area 3.5km long by up to 2km wide about 30km south-east of Calliope and 8km north-west of Bororen (Figure 172).

Type area

The type area is along the south side of Mount Coulston, north of Woodside homestead.

Topographic expression

Mount Coulston is rugged, reaching almost 500m, and thickly vegetated. The volcanics have only been examined along the lower southern slopes of the mountain.

Geophysical expression

The Coulston Volcanics display a low to moderate radiometric response, and are strongly magnetic. The magnetic pattern has been used to map the unit, but images of the first vertical derivative show a circular feature, whereas the magnetic high is arcuate in form.

Figure 172. Distribution of the Coulston Volcanics

Lithology

In the field the most widespread rocks were described as dacite, locally flow-banded and autobrecciated, with some welded crystal-poor ignimbrite. In places the volcanics appear to be more mafic, and to include intermediate or even basic compositions in coarse breccias. Chemical analyses confirm the felsic nature of most rocks, and show that they comprise trachyandesite to trachydacite compositions.

Geochemistry

Two analyses from the Coulston Volcanics range from 61 to 65% SiO_2, and have medium to high alkali contents. They have only a minor subduction component as estimated from the $(Th/Nb)_N$ ratios of 5 and 7 (Figure 173). Many of their geochemical features are similar to some analysed rhyolites of the Winterbourne Volcanics, and they plot together with them in the Within Plate Granite field on discriminant diagrams of Pearce & others (1984) (Figure 174).

Thickness

Assuming that they are flat-lying, the Coulston Volcanics could be as much as 400m thick.

Structure

The volcanic sequence is considered to be essentially flat-lying and undeformed. The brecciated nature of many outcrops, and the magnetic signature, suggests that the source of the volcanics was at or near Mount Coulston.

Age

The Coulston Volcanics are regarded as Triassic in age. Correlation with the Winterbourne Volcanics suggests that they are Late Triassic.

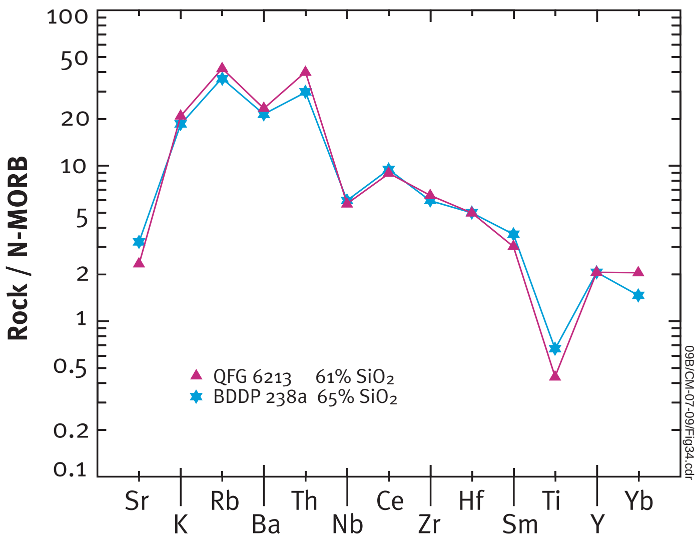


Figure 173. Spidergram plot of trachyandesite and trachydacite. N-MORB values are from Pearce (1983)

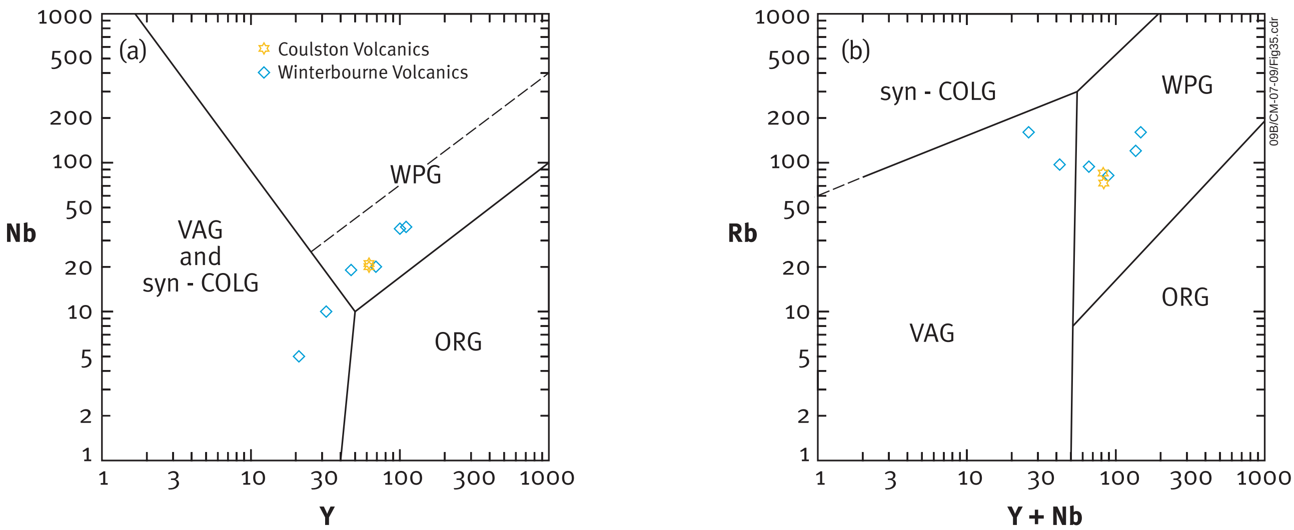


Figure 174. (a) Nb - Y and (b) Rb - Y+Nb discriminant plots of Pearce & others (1984), showing felsic volcanics from the Coulston Volcanics and Winterbourne Volcanics. VAG = volcanic arc granites, ORG = ocean ridge granites, WPG = within plate granites, and syn-COLG = syn-collisional granites.

Stratigraphic relations

The Coulston Volcanics overlie the Miriam Vale Granodiorite. Although they are geochemically similar to the Miriam Vale Granodiorite in many respects, the large age difference indicates that they are not comagmatic.

Correlation with other units

The unit is part of the extensive Triassic volcanics which occur as erosional remnants throughout the area. In particular it can be correlated with the Winterbourne Volcanics.

Economic significance

The Coulston Volcanics have no known relationship to any economic mineral deposit.

Winterbourne Volcanics (Rvw)

(A.D.C. Robertson)

Introduction

Dear (1968) mapped a sequence of andesitic volcanic rocks preserved in fault wedges to the east and south-east of Blue Hills homestead that he considered to be part of the Muncon Volcanics. Later, the volcanics that form a large part of the Kroombit Tops were included in this unit (Dear & others, 1971), although Murphy (1980) regarded them as different. The sequence of predominantly acid volcanic rocks forming the Kroombit Tops and the andesites to the east and south-east of Blue Hills homestead warrant definition as a separate unit, here named and defined as the Winterbourne Volcanics.

Distribution

The Winterbourne Volcanics form a large part of the Kroombit Tops, an equidimensional area of approximately 160km^2 centred approximately 45km east of Biloela and 80km south-west of Gladstone (Figure 175). The volcanics were derived from a caldera approximately 10km across, referred to as the Kroombit Tops caldera.

Dear & others (1971) mapped outliers of Muncon Volcanics as isolated cappings on hills to the east and north of the main outcrop area. The existence of many of these outliers, interpreted from aerial photographs, is doubtful.

Derivation of name

The name Winterbourne is from the Parish of Winterbourne in which typical rocks of the volcanic sequence crop out.

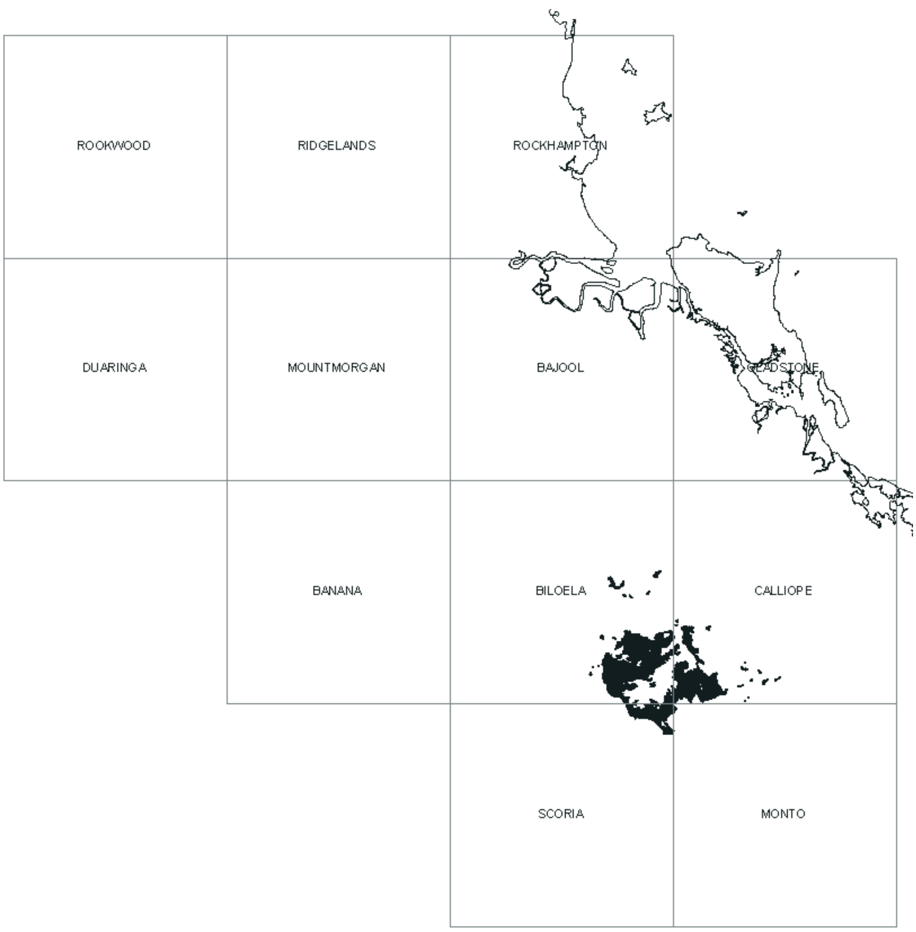


Figure 175. Distribution of the Winterbourne Volcanics

Type section

The type section nominated for the Winterbourne Volcanics is divided into two parts.The first is located along the bed of Annie Creek from its junction with Kroombit Creek (Biloela GR283850 7294650) to a position above the Kroombit Falls at Biloela GR288000 7293200. This section traverses the lower part of the eruption sequence.

The second is located towards the headwaters of Dry Creek and extends along the Creek from approximately Biloela GR290300 7202200 to GR294400 7304950. This part of the type section traverses the upper section of the Winterbourne Volcanics.

Topographic expression

The Kroombit Tops is a deeply dissected plateau with an elevation of approximately 900m. Deep gorges are common. The plateau forms the watershed between streams flowing into the Fitzroy, Burnett and Boyne River systems.

Geophysical expression

Aeromagnetic data show that most of the Winterbourne Volcanics has a low magnetic response, comparable to that of the surrounding dominantly sedimentary rock units. A stronger magnetic signature along the eastern side of the exposed volcanics, with a regular curvilinear boundary, is interpreted to be related to an intrusion at depth. On first vertical magnetic derivative maps, a circular magnetic low, surrounded by rocks of more intense magnetism, defines the position of the interpreted Kroombit Tops caldera. The approximate centre of this caldera is at GR28600 729700 on the Biloela 1:100 000 Sheet area. The caldera structure appears to cut across the moderately magnetic intrusion along the eastern margin.

The Winterbourne Volcanics give a moderate to strong radiometric response, dominated by potassium, similar to that of the Youlambie Conglomerate to the north. Localised more intense radiometric signatures occur within parts of the unit, especially around volcanic domes.

Overlying the volcanic sequence is the Precipice Sandstone that gives a radiometric response dominated by thorium. Its false colour pattern appears as green on images derived from airborne radiometric data. However, boundaries of the Precipice Sandstone interpreted from the radiometric data are not accurate. Locally, as in the north-western part of the Winterbourne Volcanics, weathering and loss of potassium has resulted in a similar radiometric response, giving the false impression that Precipice Sandstone still exists in the area.

Lithology

The Winterbourne Volcanics can be subdivided into lower and upper eruption sequences, both of which comprise numerous lava flows, interbedded breccia and ash deposits that range in composition from basalt to rhyolite.

Lower eruption sequence

The lower sequence ranges in composition from basalt to andesite, including both lavas and pyroclastic deposits. Breccia, tuff and minor ignimbrite form the pyroclastic assemblage.

The grain size of the basaltic to andesitic lavas varies considerably even within single flows. The upper surface may be glassy or fine grained (now devitrified), grading to coarse grained near the centre of the flow. The colour varies from dark grey to greenish-black when moderately fresh, and is grey on exposed surfaces. Most lavas are porphyritic with phenocrysts of plagioclase, pyroxene and less common opaque oxides (Figure 176). Olivine occurs as phenocrysts in basalts, and is widespread as a groundmass phase as well. Volcanic glass also appears in the groundmass of many of the basalts.

The accompanying breccia is dominantly andesitic and varies from matrix-supported to clast-supported. Clasts of underlying country rock are common in these early breccias. Poorly sorted conglomerate of similar composition is present (Figure 177). Ash, crystal and crystal-lithic tuffs are associated with the lava flows away from extrusive domes. Most of the tuff deposits are now extensively altered.

As eruptions continued, the ejected material became more silicic resulting in rhyolitic and trachytic lava and pyroclastic deposits. Fluidal rhyolite interbedded with fine to medium-grained rhyolitic breccia and fine-grained ash tuff dominated. The rhyolite is fine to medium-grained, grey to pink in colour and usually contains feldspar phenocrysts. Groundmass feldspar can exhibit strong flow alignment. Flow banding in

Figure 176. Hand specimen of porphyritic basalt or andesite, with phenocrysts of plagioclase and augite, Winterbourne Volcanics

extremely fine-grained, grey rhyolite is due to variations in grain size and mineral assemblages. This type of rhyolite can also exhibit hyalopilitic texture. Spherulitic texture has been noted in some rhyolite flows but is not strongly developed.

Trachytic flows appear to be restricted in extent. Most are fine-grained and grey in colour. They contain phenocrysts of an alkali feldspar and, more rarely, biotite. The dominant groundmass phase is feldspar in the form of microlites that exhibit a trachytic texture.

The interbedded breccia contains clasts that are mainly pebble to small cobble in size. Colour is usually grey to light pink depending on the type of rhyolite clast present. Clast shape varies from angular to subrounded even within a single breccia and the clasts are mainly matrix supported. Some breccias have clasts that are composed almost entirely of rhyolite while others have subordinate amounts of andesite, basalt and country rock fragments.

A vent breccia associated with the lower sequence rhyolite domes contains clasts of boulder dimensions that can exceed 10m in size. The majority of the clasts are either fluidal rhyolite (pink to grey in colour) or pink, porphyritic rhyolite. These clasts are set in a matrix that is itself a breccia with clasts of pebble to cobble size supported in a dacitic matrix. This breccia thins away from the domes.

Rhyolitic ash tuffs occur throughout the upper part of the lower sequence but are usually deeply weathered and poorly exposed. Rhyolitic ignimbrite also appears in the upper part of the lower sequence.

Upper eruption sequence

The lavas, breccias and tuffs that have been grouped into the upper eruption sequence were derived from a source on the north-eastern edge of the Kroombit Tops caldera (GR291350 7303700). Rhyolite, rhyolitic

Figure 177. Poorly sorted pebble to boulder conglomerate consisting mainly of clasts of mafic volcanics, Winterbourne Volcanics

ignimbrite, breccias and tuffs are dominant. Only in the early phases of this eruption sequence were more mafic rocks produced.

The rhyolite is texturally similar to that produced during the earlier eruptions from sources further to the south-west within the caldera. Fluidal rhyolite appears to be the dominant type and is interbedded with rhyolitic breccia that contains clasts of pebble to small cobble size. Some rhyolite is autobrecciated (Figure 178). Strongly welded rhyolitic ignimbrite exhibits columnar jointing (Figures 179 and 180). The noticeable difference between upper and lower eruption sequence is the higher proportion of breccia in the upper sequence where rhyolite and breccia occur in approximately equal proportions.

Feldspar porphyry dykes intrude both the lower and upper sequences as well as the surrounding rock units. These dykes vary in width from a few centimetres to 10m or more across. The porphyry contains white to pink feldspar phenocrysts set in a fine-grained grey to pink groundmass. Alteration due to weathering is usually extensive with most of the feldspar converted to clays.

Eruption history

During the emplacement of the lower sequence, the centre of activity was located adjacent to Kroombit Creek at approximately GR284125 7294795 where at least two lava domes occur. The initial material erupted was andesitic in composition accompanied by minor basalt and basaltic andesite. The composition of the lava became more felsic as eruption progressed. Rhyolite and trachyte flows and pyroclastics blanketed the earlier eruption products. Close to the domes, massive vent breccias are common with clasts exceeding 10m across. Away from the domes, andesitic to rhyolitic tuffs are interbedded with the lavas. All of this volcanic material was derived from a caldera approximately 10km across.

Figure 178. Autobrecciated flow banded rhyolite, Winterbourne Volcanics

The upper eruption sequence again commenced with andesitic to dacitic lavas and pyroclastic deposits but rapidly became silicic, producing rhyolite and trachyte flows breccias, tuffs and ignimbrites. The main point of eruption for this sequence is located towards the head of Dry Creek, at GR291356 7303686 on the Biloela 1:100 000 Sheet area, where a large rhyolitic lava dome occurs (Figure 181). This dome is located on the north-eastern edge of the caldera and contains spectacular columnar jointing that can be traced over a vertical height of approximately 220m on the side of the dome forming part of the Dry Creek gorge. The upper eruption sequence blanketed much of the lower sequence and the earlier formed caldera. Products of the upper eruption sequence are best seen in the cliffs along the northern side of the Kroombit Tops and along the steep sides of the Dry Creek gorge (Figure 182).

Geochemistry

Chemical analyses were only carried out on rhyolites of the Winterbourne Volcanics. The analyses show differing subduction components as indicated by low to moderate $(Th/Nb)_N$ ratios from 7 to 28 (Figure 183), entirely consistent with a tectonic environment changing from active subduction to extension. This transition is clearly shown on discriminant plots of Pearce & others (1984), in which the Winterbourne Volcanics straddle the Volcanic Arc Granite and Within Plate Granite fields (Figure 174), and is suggested by the strongly bimodal composition of the volcanic sequence.

Although generally not fluid-saturated before eruption, the presence of biotite in some of the rhyolitic lavas indicates that they were relatively hydrous during the eruption phase. The presence of Mg-rich biotite indicates that crystallisation occurred under relatively oxidising conditions.

Thickness

The thickness of the Winterbourne Volcanics varies throughout the outcrop area, and appears to be controlled by location inside or outside the interpreted caldera. In the headwaters of the South Kariboe Creek,

Figure 179. Fine grained, columnar jointed rhyolitic ignimbrite, Winterbourne Volcanics

Figure 180. Close up of the ignimbrite showing strong compaction of pumice fragments, Winterbourne Volcanics

Figure 181. Rhyolite dome with silicic volcanics to the left and a capping of Precipice Sandstone in the distance, Winterbourne Volcanics

Figure 182. Cliffs of rhyolite forming the northern escarpment of the Kroombit Tops, Winterbourne Volcanics

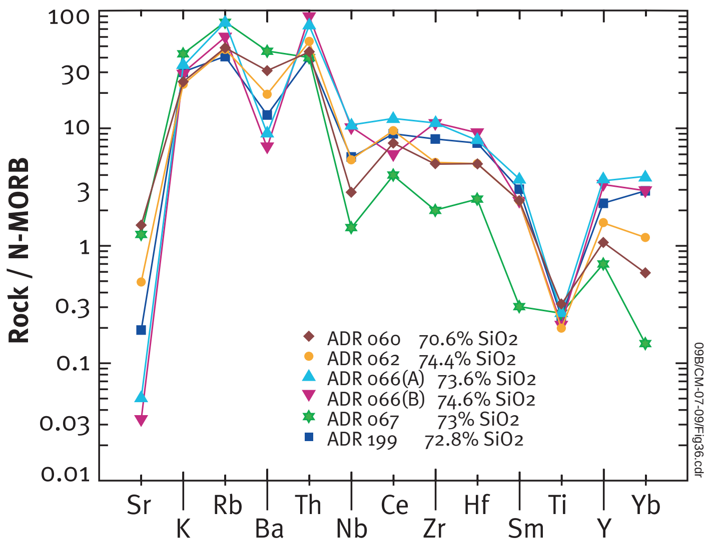


Figure 183. Spidergram plot of rhyolites of the Winterbourne Volcanics. N-MORB values are from Pearce (1983)

representing an outflow facies to the south of the caldera, the volcanic sequence does not exceed 30m in thickness. In the Kroombit Gorge, on Annie Creek, there is an estimated thickness of 270m of felsic flows and pyroclastic deposits. Towards the head of Dry Creek, in the vicinity of the upper lava dome, at least 220m of volcanics are exposed while near the northern end of the main outcrop area, north of the caldera, approximately 70m of the volcanic sequence is exposed in a cliff face.

Structure

The caldera and associated volcanics are an erosional remnant that preserve much of the original form of the volcano. The overlying Late Triassic to Early Jurassic Annie Creek beds and Early Jurassic Precipice Sandstone have not been deformed to any extent and show only a very shallow southerly regional dip. Within the caldera, there is evidence of slumping and faulting, but these structures cannot be traced over any distance.

Biostratigraphy and age

Dear & others (1971) considered that these volcanics were part of the Muncon Volcanics and suggested an Early to Middle Triassic age on the basis of plant fossil evidence in the type area. Rb-Sr age determinations on whole rock samples taken from both the lower and upper sequences of the Winterbourne Volcanics have given a poorly defined isochron of 213Ma, and similar ages have been obtained from zircons separated from a dyke near the western contact. White zircon gave ages of 215.1±0.7Ma ($^{206}Pb/^{238}U$) and 215.7±0.7Ma ($^{207}Pb/^{235}U$) and amber zircon gave ages of 223.2±1.1Ma ($^{206}Pb/^{238}U$) and 224.1±1.3Ma ($^{207}Pb/^{235}U$) (Appendix 1). All these dates are Late Triassic. A Late Triassic age is consistent with regional interpretations of the evolution of Late Permian to Late Triassic magmatism.

Stratigraphic relationships

The Winterbourne Volcanics unconformably overlie or are faulted against the Lochenbar Formation, Marble Waterhole beds, Three Moon Conglomerate, Mount Alma Formation, Rockhampton Group, and Youlambie Conglomerate, which collectively range in age from Middle Devonian to Early Permian.

The volcanic sequence is overlain by the Upper Triassic to Lower Jurassic Annie Creek Beds, Lower Jurassic Precipice Sandstone and Tertiary xenolith bearing basalt. There may be an angular unconformity between the Winterbourne Volcanics and the Precipice Sandstone at the north-eastern edge of the Kroombit Tops (Figure 184).

Correlation with other units

The Winterbourne Volcanics are younger than the more mafic Muncon Volcanics and Native Cat Andesite, and appear to represent an entirely separate eruptive event. They may be equivalent in age to the Dooloo Tops Volcanics. Their geochemistry is similar to that of the Coulston Volcanics, suggesting that they may be correlatives.

Figure 184. Possible angular unconformity between the Winterbourne Volcanics and the overlying Precipice Sandstone, north-eastern edge of the Kroombit Tops.

Economic significance

Hydrothermal alteration of rhyolitic rock types within the Winterbourne Volcanics takes several forms. The most common type is characterised by enrichment of breccias and the upper parts of flows with iron oxides, evidently related to degassing during eruption. This vapour phase alteration is not controlled by structural features and does not appear to represent a focussed or prolonged circulation of hydrothermal fluids.

A second form of alteration is the development of sericitic mica and the replacement of feldspar phenocrysts by cryptocrystalline silica in siliceous rhyolite. Alunite may also be associated with this form of alteration and the rock type can be also micro-veined by quartz.

The development of pyrite and a silica box-work within some rhyolites, breccias and to a lesser extent in tuffs, has been noted at several localities. In the vicinity of GR285147 7297594, a limonitic box work structure has developed in rhyolite as a result of the weathering of pyrite. In the same area, very weak copper staining was noted, suggesting that some of the sulphides originally present contained copper. Silica box work structures have been noted in the vicinity of GR288043 7301376 where they have developed in a rhyolitic breccia. Ferruginised breccias have also been noted in the vicinity of GR282897 7296570 and GR284689 7298385.

Traces of pyrite have been observed in a brecciated outcrop of quartz several metres across within rhyolite at GR282391 7294871.

Figure 185. Distribution of the Dooloo Tops Volcanics

Dooloo Tops Volcanics (Rvd)

(C.G. Murray)

Introduction

The Dooloo Tops Volcanics were mapped as part of the Muncon Volcanics by Dear & others (1971). However, they form a discrete outcrop with a different range and sequence of rock types, and are defined as a separate unit.

Distribution

The Dooloo Tops Volcanics form a belt 10km long and 4km wide along the crest of the Dawes Range between Dooloo Creek to the west and Glassford Creek to the east, about 30km north of Monto (Figure 185).

Derivation of name

The name is from the plateau of Dooloo Tops in the northern part of the outcrop area.

Type section

The type section is along Warrong Creek upstream from the base of the unit to the top of a prominent waterfall at GR31560E 72789N.

Topographic expression

The unit forms a broad plateau section of the Dawes Range with an average elevation of about 650m. It is fairly thickly forested in the south, and cleared to the north.

Figure 186. Autobrecciated rhyolite near the south-western edge of the Dooloo Tops, Dooloo Tops Volcanics

Geophysical expression

Most of the Dooloo Tops Volcanics have a strong radiometric response in all channels, and appear almost white on images generated from airborne data. An area in the south, just north of Pine Mountain, is an exception, displaying green colours. The unit largely coincides with a moderate magnetic feature, but this may represent a buried intrusion rather than the volcanics.

Lithology

The dominant rock type is rhyolitic ignimbrite. The prominent waterfall in Warrong Creek is formed by a thick sheet of purple ignimbrite with sparse lithic clasts and flattened pumice fragments that define an almost horizontal layering. The ignimbrite is cut by vertical andesitic dykes that have exerted some control on the course of the stream.

Volcanic breccia and conglomerate are also widespread. Clasts are exclusively volcanic lithic fragments of intermediate to felsic composition, in a groundmass of plagioclase feldspar, quartz and lithic grains. The relative abundance of intermediate volcanic clasts is surprising in view of the scarcity of primary volcanics of this composition.

Flow-banded rhyolite probably includes both extrusive lavas and dykes. Autobreccia is common (Figure 186).

Basalt occurs at the top of the sequence at both the northern and southern ends of the unit, corresponding to green areas on radiometric images. The rock is fresh, dark grey in colour, and sparsely porphyritic in olivine. These basalt outcrops have been mapped as Tertiary basalt, probably sourced from the adjacent basaltic plug at Pine Mountain, but the possibility exists that they form the topmost unit within the Dooloo Tops Volcanics.

Extending from both the northern and southern extremities of the volcanics are rhyolitic dyke swarms. These are well exposed in the gorge of Broad Creek. They are vertical, from 1 to 5m wide, and so closely spaced that little if any country rock remains (Figure 187). The wall rocks for most of the dykes are other dykes.

Figure 187. Dyke swarm in Broad Creek at the northern end of the Dooloo Tops Volcanics

Environment of deposition

The unit was erupted subaerially, apparently on a relatively subdued landscape.

Thickness

Assuming that the unit is essentially flat-lying, a thickness of about 200m is indicated by the change in elevation from base to top.

Structure

The volcanics appear to be a flat-lying erosional remnant. No significant faults were observed, but the presence of vertical dyke swarms up to 1km across indicates a substantial amount of east-north-east to west-south-west extension.

Biostratigraphy and age

In the absence of fossils and age determinations, the unit is assigned a Triassic age based on the similarity of its topographic, stratigraphic and structural relationships to those of other Triassic volcanic sequences.

Stratigraphic relationships

The Dooloo Tops Volcanics overlie the Three Moon Conglomerate and Rockhampton Group unconformably. They are intruded by the Warrong Teschenite and the Tertiary basalt plug at Pine Mountain. Their relationship with the Monal Granodiorite to the north may be either intrusive or unconformable. An unconformity is favoured by the fact that the northern dyke swarm cuts through the granodiorite.

Correlation with other units

If the topmost basalt is part of the Dooloo Tops Volcanics, the sequence of mafic volcanics overlying felsic rocks dominated by rhyolitic ignimbrite resembles that of unnamed Triassic volcanics east of Kalpowar, but is very different from the Muncon and Winterbourne Volcanics to the west.

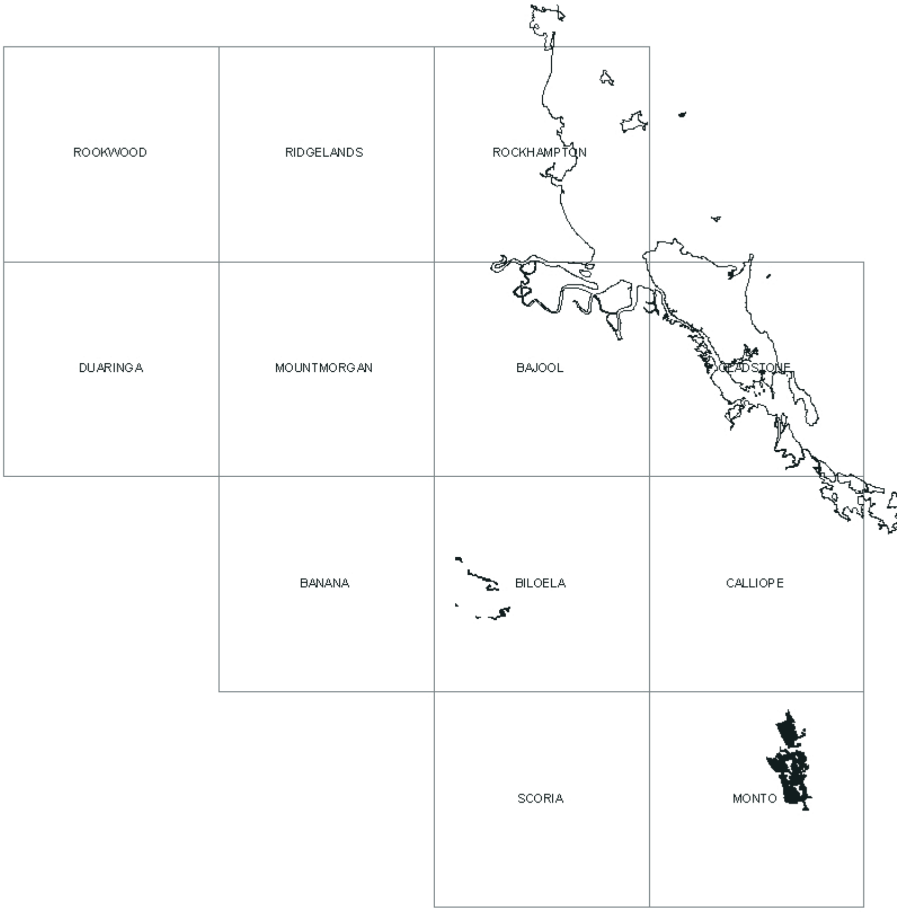


Figure 188. Distribution of the unnamed Triassic volcanics

Unnamed Triassic volcanics (Rva, Rvf, Rv?, Rvs)

(P.R. Blake & C.G. Murray)

Introduction

Dear & others (1971) included almost all the Triassic volcanic units on the Monto 1:250 000 sheet area in the Muncon Volcanics. However, the Muncon Volcanics within their type area south-west of Cania differ from sequences elsewhere. Where these different sequences have been studied during the Yarrol mapping project, they have been defined as new units, the Winterbourne Volcanics and Dooloo Tops Volcanics. The Bobby Volcanics were mapped and named in the the Miriam Vale 1:100 000 sheet area, and extend west into the Monto and Calliope 1:100 000 sheet areas. Areas that were not examined in detail have not been named or defined, although it has been possible to establish broad subdivisions. These broad subdivisions should be regarded as a starting point for future mapping investigations.

Distribution

Two extensive areas of unnamed Triassic volcanics occur in the Many Peaks – Kalpowar district. The larger covers an area 13km long and up to 9km wide between the Kalpowar and Yarrol Faults east of Kalpowar. The smaller is more than 6km long and up to 4km wide straddling Glassford Creek between its junctions with Lower Police Creek in the south and Deception Creek in the north. Smaller outliers of the volcanics occur between these outcrop areas, which are separated by only 1km (Figure 188).

Topographic expression

The larger southern outcrop east of Kalpowar is an elevated, locally rugged area at the northern end of the Burnett Range. Much of the natural vegetation has been replaced by pine plantations. The area along Glassford Creek is more subdued, with prominent north-south ridges reflecting a series of felsic dykes.

Figure 189. Flat-lying conglomerate and sandstone at the base of the Triassic volcanics 7km south-west of Many Peaks. Clasts are mainly derived from the underlying Rockhampton Group.

Geophysical expression

The radiometric signature of the volcanics is controlled by the lithology. On images generated from airborne data, very dark to black tones are characteristic of areas where mafic volcanics are dominant, as east of Kalpowar. In contrast, much of the small belt straddling Glassford Creek gives a strong, almost white response, reflecting its felsic composition.

Typically the volcanics show a weak to moderate magnetic signature dominated by short wavelengths, so that they are most readily distinguished on images displaying the first vertical derivative of the magnetic intensity.

Lithology

A distinct stratigraphic succession can be interpreted for the combined areas of volcanics in the Many Peaks – Kalpowar district, which differs somewhat from the description in Dear and others (1971). Locally the basal rocks are thin sediments derived from basement lithologies and containing sparse fossil leaves (Figure 189). The lower part of the volcanic sequence (**Rva**) is intermediate in composition, and consists mainly of purple volcaniclastic sandstone, conglomerate and breccia with abundant clasts of porphyritic and amygdaloidal andesite (Figure 190). Also within the sequence are some andesite lava flows, containing phenocrysts and microphenocrysts of plagioclase and rare clinopyroxene in a pilotaxitic groundmass of plagioclase laths, opaque grains, and altered interstitial glass.

This intermediate phase appears to be overlain by felsic volcanics (**Rvf**) in the outcrop straddling Glassford Creek. These are dominated by rhyolitic ignimbrites, flows and breccias, with lesser amounts of andesitic to dacitic breccias and minor volcaniclastic sandstones. A feature is the presence of numerous aplite dykes that trend north-north-west and dip at moderate angles to the west. Felsic volcanic sequences including rhyolitic ignimbrites occur in the headwaters of Cannindah Creek 6km south-south-east of Kalpowar. At the latter

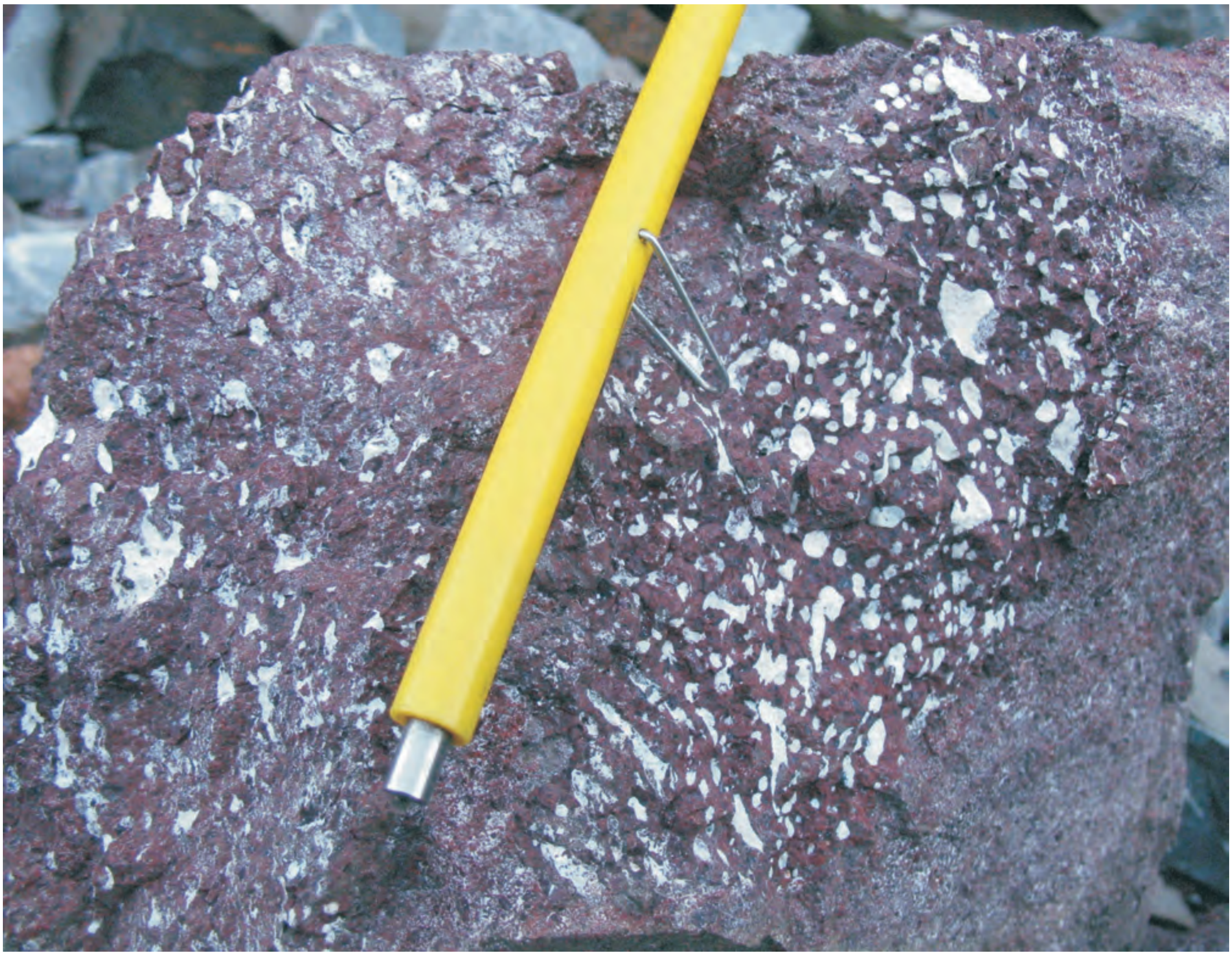


Figure 190. Purple amygdaloidal andesite clast in breccia, north of Kalpowar, unnamed Triassic Volcanics

location, the felsic volcanics appear to be overlain by a much more extensive mafic phase that continues to the north.

The area of mafic volcanics (**Rv?**) may be exaggerated by the inclusion of overlying Tertiary basalts. The Triassic sequence consists of basalt, andesite, and volcaniclastic sedimentary rocks. A basalt plug 3km north-west of Kalpowar (**Rib**) is interpreted to be related to the mafic Triassic volcanics in that area.

Along the Kalpowar-Many Peaks railway line at Barrimoon, Dear & others (1971) described about 15m of volcaniclastic sandstone, dark carbonaceous shale, and red -brown mudstone unconformably overlying the Rockhampton Group at the base of the volcanic rocks (Figure 191). This sequence includes a prominent bed of air fall material and has been mapped separately (**Rvs**).

Environment of deposition

The volcanics and associated sedimentary rocks were deposited subaerially. The variation in stratigraphic succession suggests either that the underlying topography was very uneven, or that the concept of a simple tripartite division of the volcanic units is incorrect.

Thickness

Dear and others (1971) reported a thickness of 150m for a sequence north of Kalpowar.

Structure

The volcanics are considered to be essentially flat-lying except adjacent to bounding faults, although the topography of the outcrop east of Kalpowar suggests the possibility that they dip gently towards the south.

Figure 191. Unconformity between the Rockhampton Group and the unnamed Triassic volcanics in cutting on the Calliope-Monto railway 5km north of Kalpowar. The base of the Triassic sequence is an air fall deposit.

Biostratigraphy and age

Plant fossils from the Barrimoon area were included in a list by Dear & others (1971), who assigned an Early or Middle Triassic age.

Stratigraphic relationships

The volcanics unconformably overlie the Calliope beds, Wandilla Formation, and Rockhampton Group, and are overlain by Tertiary basalts. They are faulted against the Lorray Formation along the Kalpowar Fault, and unnamed Permian strata west of Many Peaks. Their relationship with adjacent plutonic rocks is unclear.

Correlation with other units

The interpretation of the gross stratigraphic succession in the Triassic volcanics in the Many Peaks – Kalpowar district is based entirely on correlation of two widely separated sequences of felsic volcanics. Validation of the relationships obviously requires a targeted field mapping exercise.

Economic significance

Minor gold mineralisation is associated with an east-north-east trending shear zone at Barrimoon, north of Kalpowar.

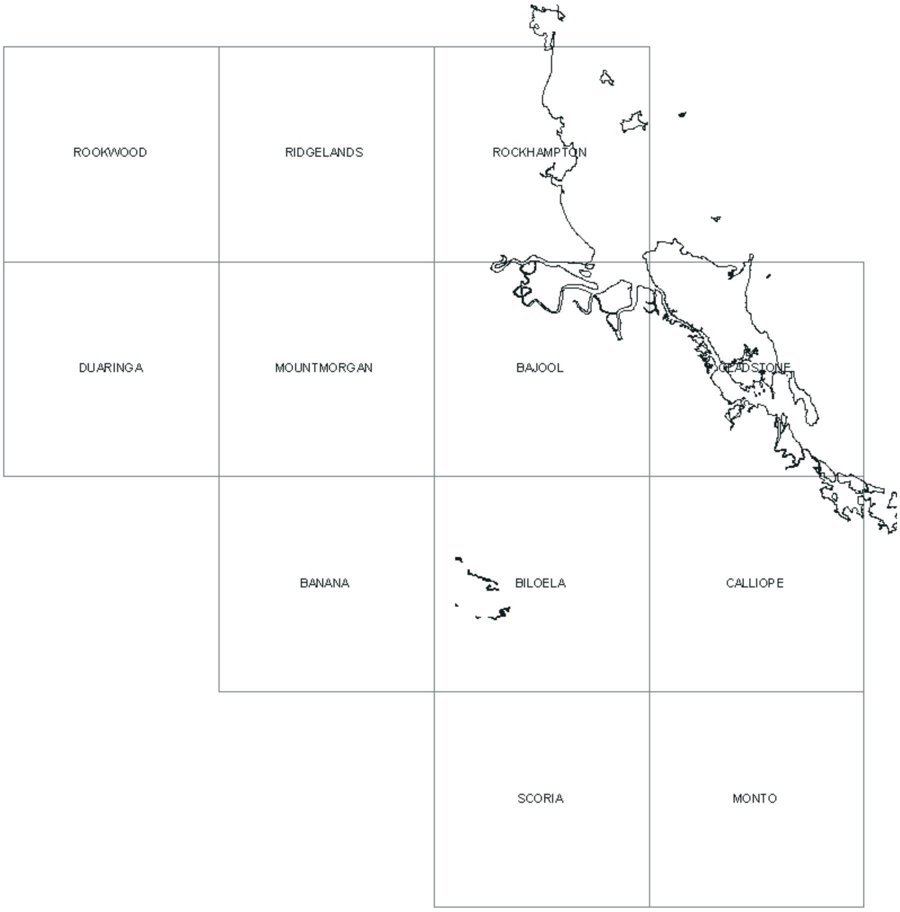


Figure 192. Distribution of the unnamed Triassic volcanics in the Callide Basin

Unnamed Triassic volcanics of the Callide Basin (Rv)

(C.G. Murray)

Introduction

Dear & others (1971) applied the name Muncon Volcanics to several areas of Triassic volcanic rocks in the Monto 1:250 000 sheet area, but the name is now restricted to the type area. They tentatively correlated a thin sequence of volcaniclastic sediments beneath the Callide Coal Measures with the Muncon Volcanics (Dear & others, 1971, page 70), but this unit differs in both rock types and stratigraphy. The following description is based on reports by Behets (1987), Biggs & others (1995), and Jorgensen (1997).

Distribution

The unit forms a narrow belt around the eastern and southern margins of the Callide Basin, centred about 15km north-east of Biloela, and also is exposed in some of the more deeply incised streams within the basin itself. It is exposed discontinuously as an arcuate belt more than 30km in length and up to 1km wide (Figure 192).

Topographic and geophysical expression

Because it forms only a narrow, discontinuous belt, the unit does not have any characteristic topographic or geophysical expression.

Lithology

The base of the unit consists of volcaniclastic pebble to cobble conglomerate interbedded with volcaniclastic sandstone. Clasts in the conglomerate are mainly andesitic to trachytic in composition. Some conglomerate has clasts of black and white siltstone (Behets, 1987).

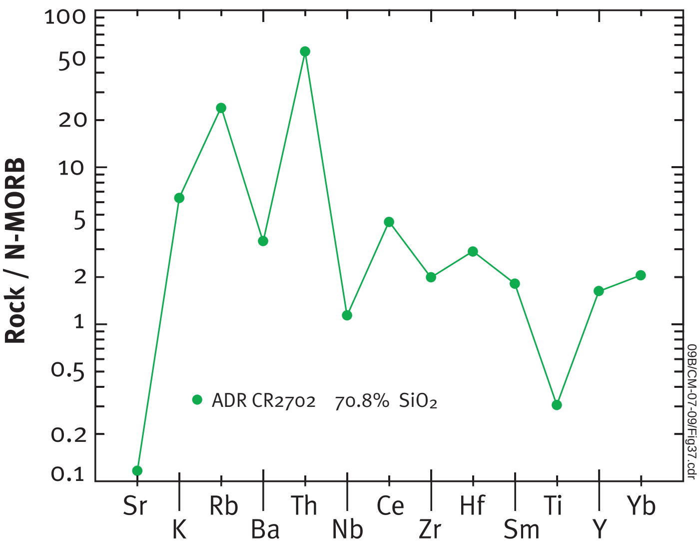


Figure 193. Spidergram plot of rhyolite. N-MORB values are from Pearce (1983)

The upper part is finer-grained overall, and consists of volcaniclastic to quartz-lithic sandstone and siltstone, some possible primary pyroclastic deposits, rare carbonaceous siltstone, and minor intermediate volcanic flows.

Geochemistry

A single sample of rhyolite from this unit was analysed. It probably came from a clast in conglomerate. The analysis has features typical of subduction-related volcanics, such as a low Nb content and high $(Th/Nb)_N$ ratio of 48.1 (Figure 193). However, it also shows some characteristics of within plate felsic volcanics, such as the moderate to extreme depletion of Sr, Ba and Ti. On discriminant diagrams of Pearce & others (1984), the analysis plots on the boundary between the Volcanic Arc Granite and Within Plate Granite or Ocean Ridge Granite fields (Figure 58).

Environment of deposition

Deposition was in a continental environment near an active volcanic source. It is uncertain whether the sequence is a remnant of a more extensive unit that was only preserved in the Callide Basin, or whether it was restricted to the vicinity of the basin. Jorgensen (1997) suggested that deposition took place in a fault-controlled basin or basins.

Thickness

Behets (1987) listed the maximum thickness as 85m, and Jorgensen (1997) gave a range of 2 to 75m.

Structure

The sequence dips to the north at angles of 5 to 30°, and displays small scale faulting and kink folding (Jorgensen, 1997).

Biostratigraphy and age

No fossils are known from the unit, and it is considered to be of Triassic age based on its stratigraphic relationships. Late Triassic volcaniclastic sediments dated by de Jersey (1974) from palynological studies were thought to be part of this sequence, but are now assigned to the overlying Callide Coal Measures (Jorgensen, 1997).

Stratigraphic relationships

The unnamed Triassic sequence unconformably overlies the Permian Youlambie Conglomerate along the eastern and southern margins of the Callide Basin. It is unconformably overlain by the Upper Triassic Callide Coal Measures, the Jurassic Precipice Sandstone, and Tertiary basalt flows and Biloela Formation.

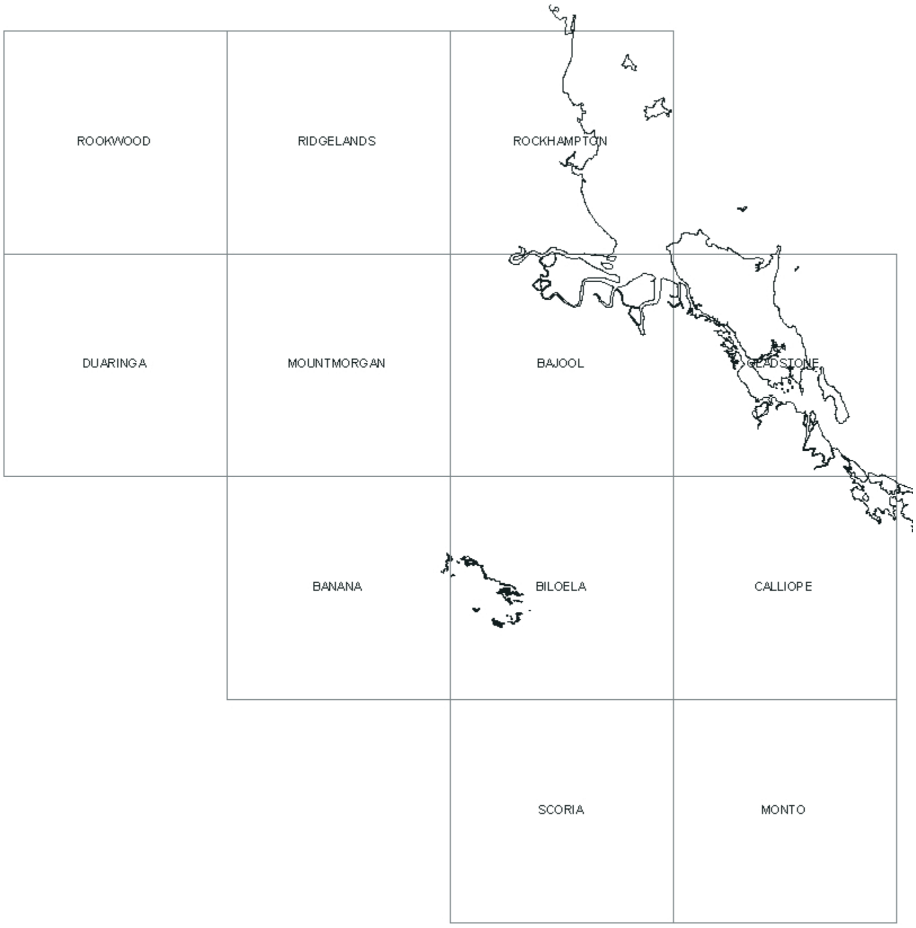


Figure 194. Distribution of the Callide Coal Measures

Callide Coal Measures (Rc)

(P.R. Blake & C.G. Murray)

Introduction

A coal-bearing unit north-east of Biloela was named the Callide Series by Dunstan (1915). Jensen (1923) changed the name to Callide Coal Measures, and correlated the strata with the Upper Triassic Ipswich Coal Measures. A brief description by Dear & others (1971) included rocks now mapped as Precipice Sandstone. They used the term Basin of Callide Coal Measures for the basin in which the coal measures were deposited, retaining the name Callide Basin proposed by Reid (1940) for the Tertiary basin to the west. The Tertiary basin was subsequently named the Biloela Basin, and Callide Basin is now used for the much smaller Late Triassic basin along its eastern edge (Allen, 1975).

Because of their economic importance, the Callide Basin and Callide Coal Measures have been extensively studied over the last few decades, and have been described in publications (Cameron, 1972; Svenson & Hayes, 1975; Stevens, 1981; Flood, 1985; Glikson & Fielding, 1991; Biggs & others, 1995; Jorgensen & Fielding, 1996), in several university theses (O'Sullivan, 1977; Stevens, 1978; Coates, 1982; Lumley, 1985; Behets, 1987; Waterfield, 1988; Sorbie, 1990; Biggs, 1996b; Jorgensen, 1997), and in numerous unpublished reports for exploration companies. The unit was not examined during regional mapping for the Yarrol Project, and the following summary is taken from the sources listed above.

Distribution

The Callide Coal Measures occupy a north-west trending, roughly rectangular basin 22.5 by 8km centred approximately 15km north-east of Biloela. However, the unit is mostly covered by the Jurassic Precipice Sandstone and Tertiary Biloela Formation, and crops out only around the periphery of the basin as a belt up to 1km wide, but generally much less, and in a few more deeply dissected streams (Figure 194).

Topographic expression

Because they form only a narrow belt, the Callide Coal Measures have no clear topographic expression, although Jorgensen (1997) states that they crop out strongly. Biggs & others (1995) show the preserved depositional basin, together with its cover of younger rocks, as an area of moderate erosion characterised by scarps and low plateaux. The 60m-high sandstone cliffs noted by Dear & others (1971) are in Precipice Sandstone.

Geophysical expression

The area of outcrop of the Callide Coal Measures is too small to produce a distinctive radiometric signature. Most of the Callide Basin is covered by Precipice Sandstone and Biloela Formation that are depleted in potassium and uranium, and give a relatively high response for thorium.

The coal measures and overlying strata give a low magnetic response. Moderately strong magnetic features are related to basement rocks, possibly intrusives.

Lithology

The Callide Coal Measures form an upward-fining sequence from a basal conglomerate facies to finer-grained sediments containing coal seams.

The base of the unit is marked by polymictic conglomerate with clasts ranging in size from pebbles to boulders. Clasts include volcanic rocks of intermediate composition, which are probably dominant, indurated sedimentary rocks (white siltstone, quartz sandstone, and chert), and subangular quartz. Sandstones and siltstones of similar composition form thin interbeds, and rare thin coal seams are also present. Volcanic-rich horizons, described as felsic tuffs and associated volcanic sediments (Jorgensen, 1997), occur in the eastern and central portions of the basin. These have been correlated with the underlying unnamed volcanic sequence (Cameron, 1972; Biggs & others, 1995), but are more felsic in composition, and are now regarded as part of the Callide Coal Measures (Jorgensen, 1997).

The top part of the unit is interbedded sandstone and siltstone with four main coal-bearing horizons. The sandstones are quartz-rich (65–95% quartz), with lesser amounts of rock fragments (mudstone, siltstone and volcanics) and plagioclase feldspar, and minor muscovite, tourmaline, rutile, and zircon. Grey and black carbonaceous siltstones are associated with the coal seams.

Recent studies have identified two types of debris flows within the Callide Coal Measures (Jorgensen & Fielding, 1996; Jorgensen, 1997). One type, which forms lobate bodies up to 1500m long, 600m wide and 15m thick, consists of diamictite, breccia and conglomerate, all with a distinctive grey-black argillaceous, sandy siltstone matrix. Pumiceous and other felsic volcaniclastic fragments are present locally, and can make up a significant proportion of the rocks. The second type, which is much less abundant and forms smaller bodies, consists of brecciated coal in a matrix of fine-grained coal detritus and some sedimentary clastic material. The debris flows occur in both the lower conglomeratic facies and the upper coal-bearing sequence of the Callide Coal Measures, and their location near faults and lineaments suggests that they were genetically related to contemporaneous faulting. Volcanic activity may also have been a contributing factor.

Environment of deposition

The Callide Coal Measures represent the preserved fill of an asymmetric basin considered by Jorgensen (1997) to be a half-graben. The basal conglomeratic facies was deposited within high energy, alluvial channel and fan environments. Subsequently, an alluvial plain environment was established over the entire basin, and sediment was distributed by low sinuosity alluvial channels forming braided channel belts. Palaeocurrent data indicate a consistent south to south-eastward direction of sediment transport throughout the history of the basin. Peat deposits accumulated in areas of the flood plain distant from the active channels (Biggs & others, 1995; Jorgensen, 1997).

Thickness

The Callide Coal Measures are up to 150m thick. The basal conglomeratic unit is up to 40m thick, but is absent in some parts of the basin. Maximum coal seam thickness (the Callide seam) is 23m.

Structure

The Callide Coal Measures are folded into an asymmetric north-west - south-east trending syncline with an axis near the south-western margin of the basin. Dips on the north-east flank of the syncline are gentle,

Figure 195. Boundary Hill open cut coal mine, Callide Basin, 1997

about 5°. Steeper dips on the south-western limb are at least partly related to faulting. The coal measures are deformed by gentle, open folds and by faults with displacements generally less than 15m.

Biostratigraphy and age

Palynological studies by de Jersey (1974) and Stevens (1978, 1981) established a Late Triassic age for the Callide Coal Measures. A disconformity separates the lower part of the unit (conglomeratic facies and lowest coal seam horizon), which is Late Carnian, from the upper part which is Early Rhaetian.

Stratigraphic relationships

Around the southern and eastern basin margins, the Callide Coal Measures unconformably overlie unnamed Triassic volcaniclastic sediments. Where this volcaniclastic unit is absent, the coal measures locally overlie the Youlambie Conglomerate and granitic rocks.

Over most of the Callide Basin the Callide Coal Measures are unconformably overlain and concealed by the Jurassic Precipice Sandstone and Tertiary Biloela Formation.

Economic geology

Coal was discovered in 1890, and first mined in 1945. Subsequent increase in production has been used for power generation in central Queensland. Annual production is now about 4 million tonnes. The coal is a sub-bituminous, non-coking coal with moderate to high ash content.

Over the last 25 years, exploration involving ground geophysical surveys, drilling, and downhole logging has been successful in delineating additional resources within the basin (Biggs & others, 1995; Nichols, 1995). The coal is mined by open cut methods (Figure 195).

TRIASSIC – JURASSIC

Bushley Dacite (RJvb)

(C.G. Murray)

Introduction

A belt of dacitic to trachytic volcanics just south of the Capricorn Highway is here defined as the Bushley Dacite. The belt was mapped as basaltic volcanics of Late Cretaceous age by Kirkegaard & others (1970). However, it gives a strong radiometric response which indicates that it is more felsic, and its stratigraphic relationship with typical sediments of the Precipice Sandstone show that it is older than the Cretaceous rocks of the area.

Distribution

The Bushley Dacite (Figure 196) extends from Bushley in the east to near Wycarbah in the west, and forms an east-north-east trending belt 5km long and up to 2.5km wide just south of the Capricorn Highway.

Derivation of name

The name is derived from the locality and railway siding of Bushley, at the eastern end of the outcrop.

Type area

The type area proposed for the Bushley Dacite is along the main central railway from Bushley to the railway crossing about 2km further west, where the unit is exposed in a number of cuttings. Some good exposures can also be seen in cuttings on the Capricorn Highway.

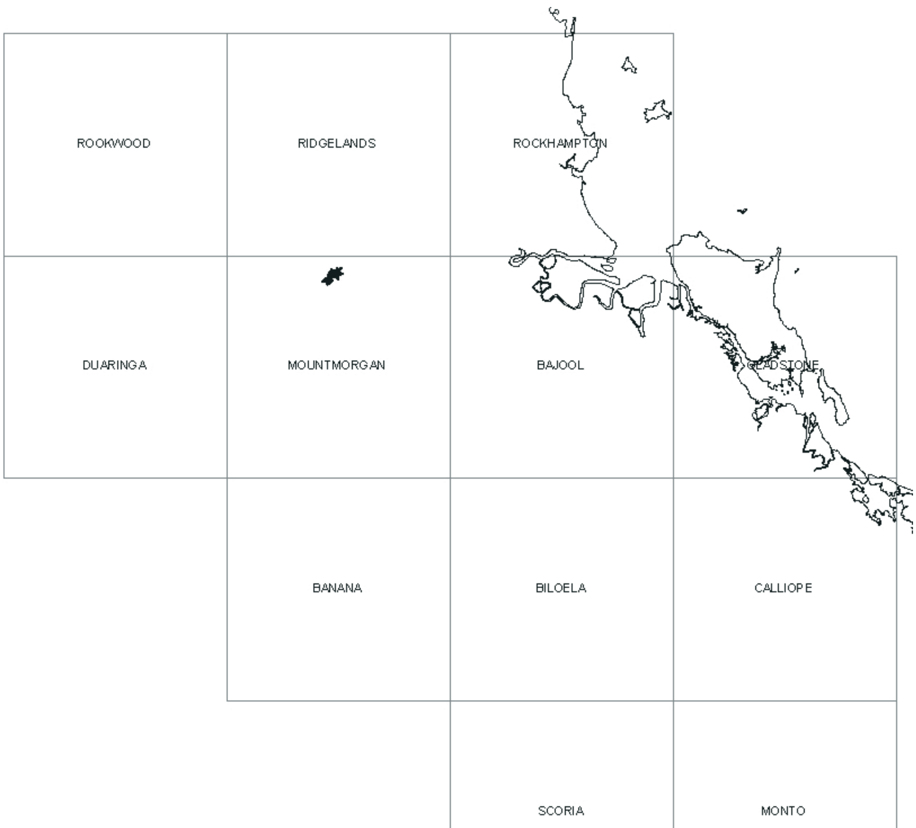


Figure 196. Distribution of the Bushley Dacite

Topographic expression

The Bushley Dacite forms low, dissected ridges which are relatively bare of vegetation.

Geophysical expression

The Bushley Dacite has a strong radiometric response, and appears white on images derived from airborne data, in contrast to the dark green to red colours of the surrounding units with the exception of the rhyolitic volcanics near Mount Hay.

The dacite has a low magnetic signature. On images derived from airborne data, the low magnetic response is hidden by the strong magnetism of the underlying units.

Lithology

The Bushley Dacite is mainly composed of dacite to trachyte or trachyandesite flows. The flows are purple to black in colour, and typically show flow banding. They were described as rhyolite in the field. They often have a patchy texture which may indicate devitrification, and the darker flows are glassy. Quartz is the only mineral recognisable in hand specimen, as slightly larger grains described as microphenocrysts in the field. Examination of thin sections leads to the conclusion that these grains are more likely to be inclusions from the basement rocks.

In thin section, the rocks are seen to be composed mainly of small, partly aligned feldspar laths which appear to consist largely of plagioclase set in a glassy matrix. Rare partly resorbed plagioclase phenocrysts have a sieve texture, and were obviously unstable with the erupted lava. The freshest rocks contain small euhedral to subhedral clinopyroxene grains. Clots of opaque and altered ferromagnesian minerals are present in some specimens. Quartz not only occurs as inclusions within particular flows, but also forms thin veins. Other rocks contain grains of a clear isotropic mineral with a low refractive index.

Geochemistry

Three samples analysed from the Bushley Dacite are very similar in their geochemistry. According to the TAS Chemical Classification, they fall close to the boundary between the dacite and trachyte fields. The geochemistry shows no trace of a subduction signature. On a spidergram, they show a strong negative slope from Th to Zr, with no minimum at Nb. The $(Th/Nb)_N$ ratio is low, from 10.5 to 12.1 and averaging 11.1 (Figure 197), despite the high Th content. They show moderately strong depletion of Ba, but only slight depletion of Ti. On discriminant diagrams of Pearce & others (1984), the Bushley Dacite falls in the Within Plate Granite fields (Figure 198).

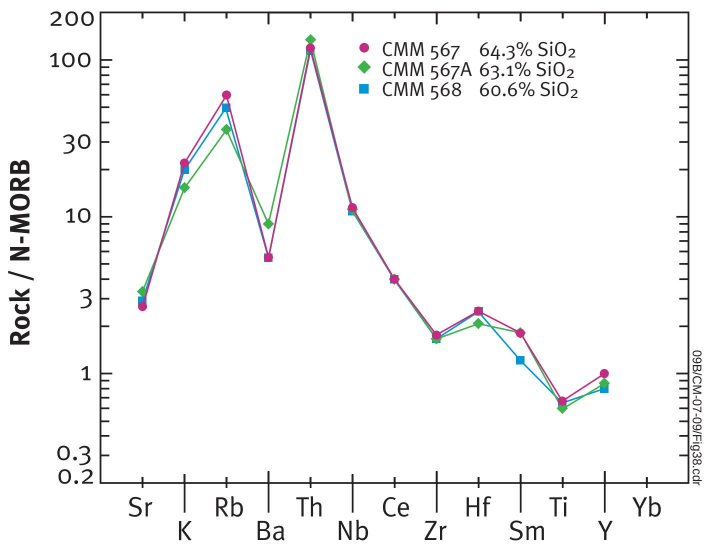


Figure 197. Spidergram plot of 3 dacites. N-MORB values are from Pearce (1983)

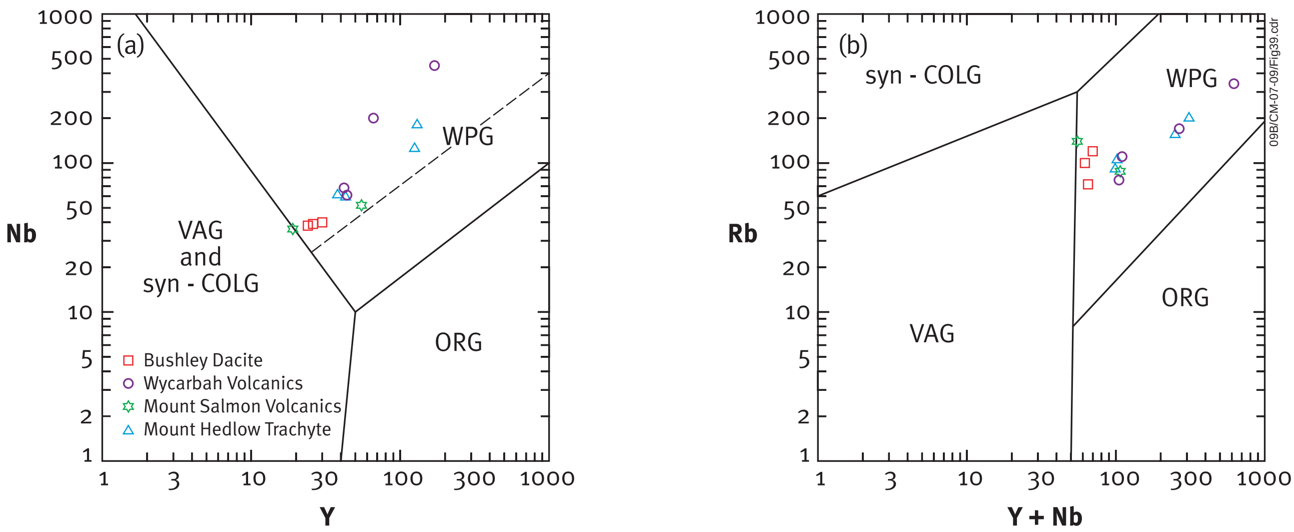


Figure 198. (a) Nb - Y and (b) Rb - Y+Nb discriminant plots of Pearce & others (1984), showing felsic volcanics from the Bushley Dacite, Wycarbah Volcanics, Mount Salmon Volcanics, and the Mount Hedlow Trachyte. VAG = volcanic arc granites, ORG = ocean ridge granites, WPG = within plate granites, and syn-COLG = syn-collisional granites.

Thickness

The maximum topographic height difference in the Bushley Dacite is about 100m. This gives a minimum thickness for the unit assuming it is flat-lying. Dips appear to be shallow to moderate, suggesting a greater thickness.

Structure

Judging by the attitude of the flow banding, the felsic volcanics are gently to moderately dipping, up to a maximum of 45°. The banding dips in opposite directions in one outcrop, and therefore may not be a reliable indication of dip. Otherwise the Bushley Dacite show no evidence of folding. It is not known whether the apparent gentle northerly dip is a primary feature or the result of earth movements or later erosion.

Age

The Bushley Dacite is overlain by the Precipice Sandstone, probably unconformably, and overlies the Umbrella Granodiorite of the Bouldercombe Igneous Complex. An age late in the Triassic is suggested by the lack of a subduction signature in the whole rock geochemistry, and the unit could even be earliest Jurassic.

Stratigraphic relationships

This volcanic unit unconformably overlies the Umbrella Granodiorite of the Bouldercombe Igneous Complex to the south, and is overlain by the Lower Jurassic Precipice Sandstone south of Bushley. To the west, it is overlain by the Lower Cretaceous Dalma Basalt, and to the north by Cainozoic alluvium. It must be older than the Lower Cretaceous Wycarbah Volcanics which form Mount Hay to the west, but the contact has not been established with certainty.

Correlation

The Bushley Dacite is the oldest unit in the Yarrol Project area to display an unequivocal within plate geochemical signature. It has no known correlatives. The limited distribution suggests a local source.

Associated mineralisation

No economic mineralisation has been recorded from the Bushley Dacite.

Annie Creek beds (RJa)

(A.D.C. Robertson)

Introduction

The Annie Creek beds is a new name proposed for a sequence of conglomerate, grey lithic sandstone, grey siltstone and carbonaceous shale that is lithologically distinct from the overlying Precipice Sandstone. This sedimentary unit had been previously mapped as part of the Jurassic Precipice Sandstone (Dear & others, 1971). In drillholes GSQ Monto 3 and 4, the base of the Precipice Sandstone included a relatively thin shale unit which may be equivalent to the Annie Creek beds (Murphy, 1980).

These sediments infilled a series of depressions on an elevated volcanic landform during the final phase of the eruption sequence. They display abrupt lateral variation in lithology and thickness, reflecting the irregular nature of the underlying volcanic surface.

Distribution

The Annie Creek beds are best exposed in the south-western part of the Kroombit Tops that lies approximately 45km east of Biloela and 80km south-west of Gladstone. Float derived from the unit has been recognised in the bed of Dry Creek on the north-western margin of the Kroombit Tops but the beds have not been mapped *in situ* in this area (Figure 199).

Derivation of name

The name of this sedimentary sequence has been derived from Annie Creek, a tributary of Kroombit Creek.

Type section

The reference section for the Annie Creek beds is near the junction of Annie Creek and Hut Creek (GR02896 72928) and extending upstream in Annie Creek for a distance of 100m.

Figure 199. Distribution of the Annie Creek beds

Geophysical expression

This sedimentary sequence has no magnetic signature that can be detected from aeromagnetic data. The outcrop areas of the beds in the headwaters of Annie Creek and South Kariboe Creek are so small that the surrounding rock types mask any radiometric signature. The underlying Winterbourne Volcanics give a strong potassium response while the overlying Precipice Sandstone gives a strong thorium response on the false colour radiometric plot of potassium-thorium- uranium (red-green-blue).

Lithology

Pebble and cobble conglomerates locally form thick beds. In the headwaters of the North Kariboe Creek (GR02882 72889) 4km north of Blue Hills, cobble conglomerate forms two beds from 8–10m thick. Other thick beds of pebble and cobble conglomerate are exposed in Annie Creek (GR02896 72928) near the junction with Hut Creek. Interbedded with these conglomerates are beds of fine-grained grey sandstone, black carbonaceous shale and grey siltstone. Dips on the shale beds at this location are between 8–10° with an azimuth of 62°. The mainly subangular clasts in the conglomerate are composed entirely of fine-grained, grey, cream or green rhyolite, ignimbrite and volcanic breccia derived locally from the underlying Winterbourne Volcanics. Matrix of the conglomerates is grey and tuffaceous in part.

Sandstone is composed almost entirely of grey, fine-grained, lithic volcanic fragments in an indurated, grey matrix. Little quartz or feldspar is visible in hand specimens.

Siltstone is dominant in the sequence away from the thicker conglomeratic lenses. It ranges from soft and fissile to resistant and blocky, with a strong tuffaceous component. Much of the siltstone, when associated with carbonaceous shale, contains abundant macerated plant detritus and also ‘rootlet’ zones marked by abundant sub-vertical carbonised rootlets. Some siltstone displays traces of an intraformational siltstone breccia texture. Black carbonaceous shale is common in the lower parts of the thicker sequences in the Annie Creek area and in the upper tributaries of North Kariboe Creek. Thin coal seams are associated with the soft, black, carbonaceous shale in Hut Creek. They occur over an interval of 0.15m with individual bands no more than 2.5cm thick and usually much thinner. A second thin interval of carbonaceous bands is developed higher in the sequence and is overlain by a granule conglomerate containing abundant angular clasts of coal, up to 6mm in diameter. Weathered coal bands occur also in a tributary of North Kariboe Creek at GR028820 7288900.

Grey lithic tuffs are interbedded in this sedimentary sequence, as are thin beds of fine-grained aphanatic cherty rocks, which may be silicified tuffs.

In GSQ Monto 3 and 4, the base of the Precipice Sandstone, which may equate to the Annie Ceek beds, consists of a thin bed of very coarse grained volcaniclastic sandstone overlain by siltstone and carbonaceous shale (Murphy, 1980). The sandstone comprises angular volcanic clasts and rare clasts of carbonaceous shale in a clay matrix. Murphy (1980) reported very thin coal bands up to 1cm thick, but it is not known whether they are present in the basal sequence, or interbedded with typical sandstone of the Precipice Sandstone higher in the sequence.

Environment of deposition

The Annie Creek beds were deposited in a shallow lake environment developed in depressions on the Upper Triassic Winterbourne Volcanics, prior to the fluviatile conditions represented by the Precipice Sandstone.

Thickness

Poor outcrop precludes an accurate measurement of total thickness. These beds were deposited in a shallow lake system developed on an irregular volcanic landform and now appear as a series of lensoidal deposits. The thickest section observed is in the headwaters of North Kariboe Creek (GR02882 728890) where an estimated thickness is of the order of 35–37m. In the lower part of Hut Creek, up to 22m formed mainly of siltstone with some carbonaceous shale, sandstone and granule conglomerate is exposed in a cliff section beneath the Precipice Sandstone. The beds are considered to be no more than 65m thick (based on a composite section) in the south-western part of the Kroombit Tops. The Winterbourne Volcanics are exposed in the beds of all the creeks to the east and south-east, precluding any significant thickening of the sequence in these directions. A slightly greater thickness may occur under the Precipice Sandstone to the north-east, but in GSQ Monto 3 and 4, the thickness of the basal sequence below the typical quartzose sandstone of the Precipice Sandstone is only 7m and 10m respectively (Murphy, 1980).

Biostratigraphy and age

Palynological examination of samples from the Annie Creek beds (J. Dear, personal communication) indicates an Early Jurassic (Classopollis classoides Sub-Zone) rather than Late Triassic age. This agrees with results from the basal sequence of GSQ Monto 3 and 4 (McKellar, 1980). Dating of the Winterbourne Volcanics using the Rb-Sr method produced an age of 218+3Ma (Late Triassic). The presence of tuff and tuffaceous sediment in the Annie Creek beds suggests a continuation of volcanic activity from the Late Triassic into the earliest Jurassic. The Annie Creek beds are therefore of Early Jurassic age, but could range down into the Late Triassic.

Structure

The Annie Creek beds are gently dipping with steepest dips of the order of 8–10° in areas where the sequence is thickest.

Stratigraphic relationships

Much of the Annie Creek beds is obscured by the overlying Lower Jurassic Precipice Sandstone and by irregular cappings of Tertiary basalt (in the headwaters of Hut Creek and North Kariboe Creek). Hence, the full distribution and maximum thickness of the unit is unknown. The unit disconformably overlies elements of the Upper Triassic Winterbourne Volcanics, being deposited in depressions within the volcanic sequence. Tertiary basalt and Precipice Sandstone cover the projected north-west extension of this sequence and to the south-west, Precipice Sandstone rests directly on deeply weathered volcanics.

Correlation with other units

A sequence of grey siltstone and quartzofeldspathic sandstone exposed in road cuttings along a forestry track 2km south-east of the forestry station on Kroombit Tops may represent equivalents of the Annie Creek beds. At this location these sediments directly overlie strongly weathered volcanics and are overlain by Precipice Sandstone. The Annie Creek beds have not been mapped elsewhere in the north-eastern part of the Kroombit Tops where exposures are poor and the Precipice Sandstone covers most of the area.

At the exposed base of the Precipice Sandstone, within the Cania Gorge, carbonaceous shale, siltstone and fine-grained tuffaceous? sandstone crop out. This material is similar to that occurring in the Annie Creek beds. The sandstone is not quartzose as is characteristic of the Precipice Sandstone, and at several localities strongly carbonised plant remains have been observed within the shale. This sequence appears to grade upwards into the Precipice Sandstone with increasing quartz content.

The lower third of the Precipice Sandstone in the Mount Morgan area (formerly the Razorback beds) consists of interbedded sandstone, mudstone, siltstone and conglomerate, which may be partly or entirely equivalent to the Annie Creek beds. The upper two-thirds comprises cross-bedded sandstone with thin silty and shaly bands (Playford & Cornelius, 1967).

JURASSIC

Precipice Sandstone (Jp)

(P.R. Blake & C.G. Murray)

Introduction

The name Precipice Sandstone was first used and defined by Whitehouse (1953, 1955), and was applied to the basal Jurassic sedimentary rocks on the Monto 1:250 000 Sheet area by Dear (1968) and Dear & others (1971). Jurassic sandstones in the Rockhampton 1:250 000 Sheet area, named the Razorback beds by Staines (1953) and Kirkegaard & others (1970), are now included in the Precipice Sandstone.

Distribution

The Precipice Sandstone is found in four main areas (Figure 200). From south to north these are: the Mulgildie Basin west of Monto; the Kroombit Tops east of Biloela; the Callide Basin north of Biloela; and the Mount Morgan – Stanwell area. A small isolated outcrop on Mackenzie Island near the mouth of the Fitzroy River (Dunstan, 1898a) is probably Precipice Sandstone.

Figure 200. Distribution of the Precipice Sandstone

Topographic expression

The upper part of the Precipice Sandstone almost always occurs as a ridge-capping, cliff-forming unit (Figure 201). The lower part typically forms steep slopes covered by scree.

Geophysical expression

On images produced from airborne radiometric data, the Precipice Sandstone usually has a strong bluish-green signature, indicating that the unit is depleted in potassium and relatively rich in thorium and uranium. This feature may be at least partly the result of prolonged weathering. The Precipice Sandstone has a very weak magnetic response, and on images derived from airborne data the magnetic pattern is usually that of the underlying basement rocks.

Lithology

In the Mulgildie Basin, the formation consists of quartzose sandstone, pebbly quartzose sandstone, and some siltstone. The sandstones are composed almost entirely of subangular to moderately rounded quartz grains, with rare white feldspar, muscovite, and dark lithic grains, and are thick-bedded and cross-bedded (Dear & others, 1971). In core from borehole GSQ Monto 1-2R, the Precipice Sandstone consists almost entirely of cream to white, fine to very coarse quartzose sandstone, with an argillaceous and siliceous matrix (Gray, 1972a). The only other rock type noted was conglomerate composed mainly of white quartz pebbles, which formed several beds. In the nearby APE Mulgildie No.1 well, a sequence of green siliceous sandstone, grey micaceous and carbonaceous siltstone, coarse-grained quartz sandstone and shale that was reported to underlie the Precipice Sandstone conformably (Dear & others, 1971) is probably of Triassic age.

The Precipice Sandstone that overlies the Callide Coal Measures north of Biloela is dominated by thick beds of cross-bedded white to buff quartz sandstone with minor interbedded quartz granule to pebble conglomerate and pale siltstone (Jorgensen, 1997).

Figure 201. Precipice Sandstone capping the ridge south-east of Cania Dam, looking from the site of the Cania goldfield

The Precipice Sandstone in the Mount Morgan area (formerly the Razorback beds) was described in detail by Playford & Cornelius (1967). The lower third of the unit consists of interbedded sandstone, mudstone, siltstone and conglomerate. The upper two-thirds comprises cross-bedded sandstone with thin silty and shaly bands. It is estimated that sandstone makes up 80% of this upper section. The sandstones are quartz-rich, with a cement composed of silica and minor carbonate. A secondary hematitic cement was introduced later, and hematitic concretions are common. The mudstones are mainly kaolinite with about 20% silt-sized quartz grains. They are generally carbonaceous and micaceous, and some are laminated. Conglomerate, which locally forms the base of the sequence, is polymictic with a secondary hematitic cement. Clast compositions reflect the adjacent basement rocks, and felsic volcanic clasts and quartz are most abundant. The nature of the quartz grains suggests that the sediments were derived mainly from granitic source rocks (Playford & Cornelius, 1967).

The small outcrop on Mackenzie Island consists mainly of fine-grained sandstone, usually massive but locally cross-bedded, similar to the Precipice Sandstone quarried at Stanwell (Dunstan, 1898a). Some coarser-grained and thin-bedded sandstones occur, as well as ironstone bands.

Environment of deposition

Consistent cross-bedding orientations show that the upper part of the Precipice Sandstone was deposited by a braided river system flowing in a general northward direction, ranging from north-east to north-west (Playford & Cornelius, 1967; Behets, 1987; Waterfield, 1988; Sorbie, 1990; Jorgensen, 1997) (Figure 202). The generally coarse grain size is consistent with fluvial deposition.

Figure 202. Cross-bedded sandstone at the Stanwell sandstone quarry, looking west, Precipice Sandstone

Thickness

The thickness of the Precipice Sandstone is reasonably constant considering its wide distribution. In GSQ Monto 1-2R, Gray (1972a) measured a thickness of 65m. The much greater thickness of 105m reported by Dear & others (1971) from APE Mulgildie 1 is probably incorrect. At Cania Gorge, the thickness is 60m (Dear & others, 1971). Where it overlies the Callide Coal Measures, the Precipice Sandstone may be as much as 100m thick (Jorgensen, 1997). The unit averages about 45m thick in the Mount Morgan area, with a maximum of 90m, although small erosional remnants are much thinner. Mudstone beds are up to 12m thick (Playford & Cornelius, 1967).

Structure

Over most of its outcrop area, the Precipice Sandstone is virtually flat-lying, or dips at very low angles. Some of the dips may reflect the topography of the pre-Jurassic land surface. In the Mount Morgan area, the base of the unit falls from over 400m in the Razorback Range to <100m at the southern edge of the Stanwell Basin. Here, and also at Flagstaff Hill near Stanwell, the sandstone dips beneath the Stanwell Formation at ~10° (Figure 203). Dunstan (1898a) recorded dips of up to 10° to the south on Mackenzie Island.

Steep dips at two localities have been produced by faulting. Along the eastern margin of the Mulgildie Basin north of Monto, the Precipice Sandstone has been tilted to almost vertical by movement on the Mulgildie Fault, and now forms a prominent, rocky ridge (Figure 204).

Biostratigraphy and age

The Precipice Sandstone is Early Jurassic in age. Playford & Cornelius (1967) reported spores of this age from mudstone in old gold workings on Mount Victoria, near Mount Morgan. De Jersey (1974) recovered spore assemblages from what is now mapped as Precipice Sandstone from the area covered by the Callide Coal Measures, and assigned an Early Jurassic age similar to that of the Precipice Sandstone in the Surat Basin and the equivalent Helidon Sandstone of the Moreton Basin. Spores of the same age were recovered by Helby & Partridge (1977) from a depth of 220m in core from a borehole through the Stanwell Formation near Stanwell.

Bartholomai (1966) identified bones collected by Cornelius as *Leptocleidus superstes*, a freshwater plesiosaur known from the Cretaceous of Europe and Africa.

Figure 203. Precipice Sandstone on Flagstaff Hill, near Stanwell, looking south

Figure 204. Vertical sandstone bed along the Mulgildie Fault at the eastern margin of the Mulgildie Basin, 20km north-north-west of Monto, Precipice Sandstone

Figure 205. Cappings of metamorphosed Precipice Sandstone overlying Flaggy Quartz Monzodiorite on two hills 6.5km east of Westwood

Stratigraphic relationships

The Precipice Sandstone unconformably overlies many Palaeozoic geological units, including intrusives. It also overlies several units assigned a Triassic age, including the Bushley Volcanics, Winterbourne Volcanics, Annie Creek beds, Callide Coal Measures, and Muncon Volcanics, and possibly the Native Cat Andesite. Some contacts with these units are interpreted as conformable or disconformable, whereas others are described as unconformable.

The Precipice Sandstone is overlain conformably by the Evergreen Formation in the Mulgildie Basin (Dear & others, 1971). In the Stanwell area, it is overlain by the Lower Cretaceous Dalma Basalt and Stanwell Formation. There is only local angular discordance between the Precipice Sandstone and these overlying units, but there is evidence for a substantial period of erosion.

Around Biloela, the Precipice Sandstone is overlain by the Tertiary Biloela Formation and by unnamed Tertiary basalt flows.

Metamorphism

In the Stanwell–Westwood area, the Precipice Sandstone locally exhibits well developed columnar jointing. The best examples are on four hills centred 6km east of Westwood, where the Precipice Sandstone overlies Permian granitic intrusives (Figure 205). The columnar jointing occurs at the top of a cross-bedded quartzose sandstone bed. The columns are vertical. They decrease in size, and become more regular in outline, downwards (Figures 206 and 207). The sandstone in the columns has been recrystallised so that the quartz grains form a continuous network. At one point they form a low cliff above an extensive scree slope of sandstone columns (Figures 208 and 209). Dunstan (1898b) described the same feature in sandstone along Quarry Creek 4km south of Stanwell.

The columnar jointing was undoubtedly the result of metamorphism by an overlying pile of basalts of the Dalma Basalt, probably including intrusive sills to produce the required heat. No trace of this overlying sequence remains at the localities exhibiting columnar jointing.

Correlation with other units

The upper, cliff-forming sandstone unit is characteristic and is everywhere mapped as the Precipice Sandstone. The lower part is covered by scree at many localities, is more difficult to delineate, and in some cases has been separated from the Precipice Sandstone, for example on the Kroombit Tops, where it is defined as the Annie Creek beds, and in the APE Mulgildie No.1 well.

Figure 206. Columnar jointing in cross-bedded sandstone, hill 6.5km east of Westwood, Precipice Sandstone

Figure 207. Lower limit of columnar jointing in cross-bedded sandstone, hill 6km east-north-east of Westwood, Precipice Sandstone

Figure 208. Cliff about 10m high above scree slope of sandstone columns, southern side of hill about 5.5km east of Westwood, Precipice Sandstone

Figure 209. Scree slope of sandstone columns, southern side of hill about 5.5km east of Westwood, Precipice Sandstone

Figure 210. Sandstone quarry, Stanwell, 1997, Precipice Sandstone

Economic significance

The Precipice Sandstone has been quarried for building stone at a site 2.5km south-east of Stanwell for more than a century (Figure 210). Some stone is used for renovation work on buildings in Sydney. The outcrop on Mackenzie Island was also considered as a possible source of building stone for Rockhampton (Bird, 1904).

Gold has been mined from the basal beds of the Precipice Sandstone at Mount Victoria west of Mount Morgan (Reid, 1939b). Material treated from this locality averaged about 30g/t gold.

Evergreen Formation (Je)

(P.R. Blake)

Introduction

The name Evergreen Formation was introduced by Jensen & others (1964) for the Jurassic sediments between the Precipice Sandstone and Hutton Sandstone in the Surat Basin. It included two previously named units, the Evergreen Shales (Whitehouse, 1953) and the Boxvale Sandstone (Reeves, 1947).

Distribution

In the Yarrol Project area, the Evergreen Formation occurs in a belt 35km long and up to 10km wide along the western side of the Mulgildie Basin, and continues south onto the Mundubbera 1:250 000 sheet area (Figure 211).

Topographic expression

This formation generally produces a recessive topography, and is mostly seen in valley areas, with a narrow but prominent ledge formed by a ferruginous oolite marker bed where it occurs on slopes beneath a capping of Hutton Sandstone.

Figure 211. Distribution of the Evergreen Formation

Geophysical expression

The Evergreen Formation gives a blue-green colour on the airborne radiometric images, indicating that it is either relatively enriched in thorium and uranium, or depleted in potassium. The rocks have undergone pervasive weathering, and are capped by laterite in places. The unit has a very low response on airborne magnetic images.

Lithology

The Evergreen Formation is divided into upper and lower sections by the ferruginous oolite marker bed. In the lower part of the formation, the only strata that crop out are white and grey, fine-grained flaggy sandstone and siltstone which overlie the Precipice Sandstone. The rocks are predominantly quartzose, but in addition contain weathered feldspar, light-coloured mica, and some lithic grains. They are soft sediments with a clay (?kaolin) matrix. Except in lateritised areas, the soil cover is grey and sandy, suggesting that poorly consolidated sandstones make up much of the interval (Dear & others, 1971). This lower sandstone-rich section may be transitional with the Precipice Sandstone.

The ferruginous oolite horizon is widespread around the basin, and its main development is about 30m below the Hutton Sandstone. It is yellow-brown to red-brown in outcrop, bedded, and the main bed is 3 to 4m thick. The rock is composed of light brown, concentrically zoned oolites, 0.5 to 0.6mm in diameter, cemented by red-brown iron oxides, silicates and calcite. The oolites consist of goethite, hematite, limonite and siderite in concentric shells.

Thin-bedded, flaggy sandstone is again the main rock type exposed towards the top of the formation. The sandstone is cream in colour, medium to coarse-grained, and friable due to its soft clay matrix. Quartz is the main clastic component, but 10–15% feldspar is also present (Dear & others, 1971).

In GSQ Monto 1-2R, the Evergreen Formation consists of interbedded and laminated, grey to dark grey and black, carbonaceous shale, mudstone, siltstone and minor coal (63%), and thick-bedded, pale grey to grey, fine to medium sublabile to labile sandstone (37%) (Gray, 1972a). The sandstone contains dark grey mud clasts, some coal lenses and calcite veins. The oolitic ironstone horizon occurs about 60 m below the top of the formation, and is almost 5m thick in total.

Environment of deposition

The formation is interpreted to have been deposited in fresh water fluvial and lacustrine deposits. It has been suggested that the oolitic ironstone horizon, which contains acritarchs and rare arenaceous foraminifera, represents quasi-marine conditions (Day & others, 1974), but ironstones of this type can be deposited in freshwater environments.

Thickness

The formation is estimated at between 105 and 120m thick on the western flank of the Mulgildie Basin (Dear & others, 1971). In both APE Mulgildie 1 and GSQ Monto 1-2R, it is 260m thick (Dear & others, 1971; Gray, 1972a).

Structure

The Evergreen Formation is folded into a very open synclinal structure. Dips nowhere exceed 5°.

Biostratigraphy and age

Microfloras from the Evergreen Formation in the Surat Basin are Early Jurassic (de Jersey & Paten, 1964), and this age is accepted for the unit in the Mulgildie Basin.

Stratigraphic relationships

The Evergreen Formation conformably overlies the Precipice Sandstone, and is overlain conformably by the Hutton Sandstone (Dear & others, 1971). Both upper and lower boundaries may be transitional.

Correlation with other units

The Evergreen Formation in the Mulgildie Basin is correlated with the Evergreen Formation in the Surat Basin (Gray, 1972a). Also, the oolitic ironstones are interpreted to be a time-concordant marker which can be traced from the Surat and Mulgildie Basins, through the Moreton Basin to the Landsborough Sandstone and Brighton beds of the Nambour Basin, and into the Tiaro Coal Measures of the Maryborough Basin (Day & others, 1974).

Economic significance

The oolitic ironstone has been investigated as a possible source of iron ore. In the Mulgildie Basin, the most promising prospects are north-west of Monto, in the Moonford and Coominglah Range areas, where average grades are more than 40% Fe (Berkman, 1962). However, the deposits are not considered to be economic because of the high overburden to ore ratio and the difficulty of upgrading the ore to produce an acceptable concentrate.

Hutton Sandstone (Jh)

(S.B.S. Crouch)

Introduction

The type section of the Hutton Sandstone is in the Surat Basin in the Eddystone 1:250 000 Sheet area. The unit extends into the Mulgildie Basin, where it overlies the Evergreen Formation and is overlain by the Mulgildie Coal Measures.

Distribution

The Hutton Sandstone forms a north-south trending belt along the central part of the Mulgildie Basin from Moonford to the southern boundary of the Monto 1:100 000 sheet area. This belt is 27km long and averages about 5km across (Figure 212). Smaller outcrops along the faulted eastern margin of the basin north-east and south-east of Monto have steep dips due to movement on the Mulgildie Fault.

Figure 212. Distribution of the Hutton Sandstone

Topographic expression

Typically, the sandstone gives rise to a broad belt of sandy plains with low rounded hills, and rarely, cliffs. The sandstone usually has a low bluff near its base, but weathers to form steep slopes rather than cliffs.

Geophysical expression

The Hutton Sandstone gives a blue-green colour on the airborne radiometric images, indicating that it is either relatively enriched in thorium and uranium, or depleted in potassium. This suggests that the rocks are weathered. The unit has a very low response on images of airborne magnetic data.

Lithology

Dear & others (1971) describe the formation as comprising well bedded, quartzose sandstone, minor granule conglomerate, and thin sandy siltstone beds. At the base of the section, the sandstones are well bedded, and commonly cross-bedded. They are cream to light brown, and of medium grain size. The average composition is quartz (70–80%), feldspar (10–20%), muscovite (1%), and clay matrix (10%). Lithic grains are rare or absent. Porosity and permeability are moderately good. Abundant clay-siltstone lumps, or clasts, are particularly characteristic of these lower sandstone beds. The clasts are usually less than 2.5cm across, but can be up to 22cm across. This unusual sandstone probably resulted from wave or current disruption of thin mud beds, and incorporation of the mud clasts in sandbanks. Very shallow water deposition seems likely. Mud clasts are less common in the upper part of the formation.

Sandstone from the upper part of the formation is well-bedded (15cm to 1m beds) and is white, or red-brown due to iron staining. The rock is quartzose, and grain size ranges from fine sand to granule conglomerate. A clay matrix is present, but surface samples have moderately good porosity and permeability. Grey-brown sandy siltstone is interbedded with the sandstone. Cross-beds in the basal sands at the northern end of the basin dip to the north-east or east, indicating current movement from the south-west or west.

Complete sections of the Hutton Sandstone were penetrated by the APE Mulgildie No. 1 well (Stewart & Gough, 1963) and by fully cored borehole GSQ Monto 1-2R (Gray, 1972a). In GSQ Monto 1-2R, the upper boundary of the Hutton Sandstone with the Mulgildie Coal Measures was placed at the top of the highest bed of white quartzose sandstone, which is 4m thick and easily recognised on the gamma-ray log. The basal contact with the Evergreen Formation was selected at the top of an 11m sequence of dark grey shale with minor sublabile sandstone beds. Gray (1972a) described the Hutton Sandstone as pale grey to white, fine to medium-grained, well sorted sublabile to quartzose sandstone (71%), thin-bedded to laminated grey to dark grey shale, mudstone and very minor coal (24%), and grey siltstone (5%). The sandstone contains minor shale beds and coal lenses, and shale clasts are particularly abundant in the bottom 15m of the unit, as noted by Dear & others (1971) from outcrops. Stewart & Gough (1963) did not attempt a formal subdivision of the Jurassic sequence in APE Mulgildie 1. Dear & others (1971) tentatively suggested that the upper contact of the Hutton Sandstone with the Mulgildie Coal Measures in this well is at 262m, but it must be much higher, and was placed at 55m in the nearby GSQ Monto 1-2R (Gray, 1972a). The only core from the Hutton Sandstone is a homogeneous section of pale grey quartzose sandstone with a few scattered, dark grey-brown micaceous mudstone clasts. The components in the sandstone are quartz (70%), cream angular feldspar (10%), black lithic grains (1-2%), and mica (1%). There is a grey clay (?kaolin) matrix, but some porosity and permeability. The remaining part of the formation appears to be sandstone and siltstone similar in composition to that in the cored interval, although siltstone beds were identified by Stewart & Gough (1963) from cuttings and logs in the interval 290–293m.

Environment of deposition

The Hutton Sandstone was deposited in a freshwater fluviatile environment.

Thickness

Along the western side of the Mulgildie Basin, the greatest thickness measured in outcrop by Dear & others (1971) was 24m. In GSQ Monto 1-2R, in the deepest part of the Mulgildie Basin, the Hutton Sandstone as defined by Gray (1972a) is 275m thick, from 55 to 330m. This thickness is comparable with that in GSQ Mundubbera 1, which cored a complete section of 200m through the Hutton Sandstone at the northern margin of the Surat Basin (Gray, 1972b). It is not known whether the formation was originally continuous across the Auburn Arch from the Surat to the Mulgildie Basin, but the presence of only a few tens of metres of section exposed in the western part of the Mulgildie Basin suggests that it was not.

Structure

With the other Jurassic formations, the sandstone is folded into an asymmetric syncline that has its axis close to the Mulgildie Fault. Dips on the western limb nowhere exceed 5°, but on the eastern limb the sandstone is vertical or nearly so due to movement on the Mulgildie Fault. The Anyarro Fault, first recognised by Reid (1927), is a normal fault along the central part of the Mulgildie Basin with a relatively small displacement at the surface.

Biostratigraphy and age

No fauna or flora has been obtained from the formation in the Yarrol Project area. In the Surat Basin, the age of the formation is late Lower and early Middle Jurassic (de Jersey & Paten, 1964).

Stratigraphic relationships

The Hutton Sandstone overlies the Evergreen Formation conformably, and is itself overlain conformably by the Mulgildie Coal Measures. Both upper and lower boundaries are gradational.

Correlation with other units

The Hutton Sandstone in the Yarrol Province is equivalent to the Hutton Sandstone in the Surat Basin (Gray, 1972a).

Mulgildie Coal Measures (Jm)

(S.B.S. Crouch & P.R. Blake)

Introduction

The Mulgildie Coal Measures were first described and named (as the Mulgeldie Coal Measures) by Reid (1927) from water bore records. He correlated them with the Walloon Coal Measures. Production by the Burnett Coal Company from the Selene mine south of Mulgildie totalled about 500 000t from 1949 to 1966, and supplied the Gladstone meat works. The Mulgildie Coal Measures were penetrated by petroleum exploration well APE Mulgildie 1, located about a kilometre south-south-east of Mulgildie (Stewart & Gough, 1963), and by the fully cored hole GSQ Monto 1-2R located a kilometre south of APE Mulgildie 1 (Gray, 1972a). The coal resources have been investigated by a series of exploration programs since 1969 (Svenson & Rayment, 1975).

Distribution

Reid (1927) mapped the coal measures along the synclinal axis of the Mulgildie Basin over a length of almost 20km and a width of about 4km, extending along the eastern side of Three Moon Creek from near Monto in the north to Kapaldo on the Eidsvold 1:100 000 sheet area in the south (Figure 213).

Topographic expression

The rocks of the Mulgildie Coal Measures are recessive, with very little surface outcrop, and produce a gentle topography typified by low undulating country with deep black and red-brown soils.

Geophysical expression

The Mulgildie Coal Measures have a blue-green colour on airborne radiometric images, indicating that the unit is either relatively enriched in thorium and uranium, or depleted in potassium. This probably shows that the rocks have undergone some weathering. The unit has a very low response on airborne magnetic images.

Figure 213. Distribution of the Mulgildie Coal Measures

Lithology

The formation consists of interbedded sandstone, siltstone and shale, with a number of lenticular coal seams. Little is known of the succession because it does not crop out. Dear & others (1971) divided the coal measure succession into four units based on cuttings and sparse cores from APE Mulgildie 1. However, the stratigraphy in this borehole was not established by Stewart & Gough (1963). The boundary of the Mulgildie Coal Measures with the underlying Hutton Sandstone is gradational, and it is probable that only the upper unit and possibly part of the next lower unit of Dear & others (1971) belong to the Mulgildie Coal Measures.

Based on cuttings from the APE Mulgildie 1 well, the upper unit recognised by Dear & others (1971) is a succession of soft light and dark grey mudstone and siltstone, and quartz sandstone beds, with coal and shaly coal at five main horizons, designated A to E by Reid (1927). The sandstone beds are white to brown, quartzose, very friable, and have a soft clay matrix. The mudstone is very soft, only slightly micaceous, but very carbonaceous. Limonite concretions and boulders are common in the soil derived from the beds. The description of the Mulgildie Coal Measures by Svenson & Rayment (1975) is very similar, although they do show minor limestone in their figure depicting a typical section of the unit. The coal seams are generally banded and most exhibit splitting or lateral gradation to carbonaceous shale (Svenson & Rayment, 1975).

Gray (1972a) reported that the Mulgildie Coal Measures in GSQ Monto 1-2R consist of grey to brown micaceous siltstone, grey, very fine to medium, well-sorted, sublabile sandstone with calcite veins, and grey to dark grey carbonaceous mudstone and shale, with common plant fragments. The hole was not cored above 30m, and the most prominent coal seams were not sampled.

Environment of deposition

Given the presence of coal and plant fossils, the Mulgildie Coal Measures are interpreted to have been deposited in a terrestrial environment that was presumably fluviolacustrine.

Thickness

Reid (1927) estimated the maximum thickness of the coal measures at about 150m. Dear & others (1971) gave a questionable figure of 365m, presumably based mainly on the section in APE Mulgildie 1, but this is much too great, and must include a significant thickness of Hutton Sandstone. In GSQ Monto 1-2R, located one kilometre south of APE Mulgildie 1, Gray (1972a) reported only 55m of Mulgildie Coal Measures, above a 4m thick bed of white quartzose sandstone that was distinct on the gamma-ray log and was interpreted as the top of the Hutton Sandstone. Svenson & Rayment (1975) gave a thickness of 150m.

Structure

The coal measures occur in the synclinal axis of the Mulgildie Basin. The syncline is asymmetric, the western limb dipping to the east at between 2° and 5°, and the eastern limb dipping more steeply to the west. Along the eastern margin of the basin, dips are locally as steep as 80° where Palaeozoic rocks have been thrust over the Mulgildie Coal Measures along the high-angle reverse Mulgildie Fault. Reid (1927) recognised a north-north-west trending normal fault, subsequently named the Anyarro Fault, along the western boundary of the coal measures, associated with steep dips. The Mulgildie Coal Measures appear to be downthrown to the east along this fault, which might be partly responsible for their preservation in the Mulgildie-Monto area. However, further south on the Eidsvold 1:100 000 sheet, the displacement is apparently reversed, and the Mulgildie Coal Measures have been mapped only west of the fault zone.

Biostratigraphy and Age

Small collections of plant fossils by Reid (1927) and Dear & others (1971) are consistent with the lithological correlation with the Walloon Coal Measures, and support a Middle Jurassic age. No microfloras have been described.

Stratigraphic Relationships

The Mulgildie Coal Measures conformably overlie the Hutton Sandstone, and are the topmost unit within the Mulgildie Basin. It is not known how much of the sequence has been removed by erosion. The only strata overlying them are unnamed Tertiary basalts and Quaternary alluvium.

Correlation with other units

There are no correlative units within the Yarrol Project study area. Regional equivalents of the Mulgildie Coal Measures are the lower section of the Injune Creek beds of the Surat Basin, the Walloon Coal Measures of the Moreton Basin, and the upper part of the Tiaro Coal Measures of the Maryborough Basin.

Economic significance

The succession of coal seams interpreted by Reid (1927) from water bore records has largely been confirmed by subsequent exploration drilling for coal (Svenson & Rayment, 1975). The coal seams are lenticular with marked thickness variations and seam splitting, typical of the coals of the Walloon Coal Measures. Coal development is greatest in the south-eastern part of the unit, in the region between the Selene mine and Mulgildie. Measured and indicated resources amenable to open-cut mining were reported as 110 million tonnes by Goscombe & Coxhead (1995). The coal is sub-bituminous and non-coking, and is suitable for power generation (Svenson & Rayment, 1975).

CRETACEOUS

Unnamed marine sediments (Kl)

(C.G. Murray)

Introduction

Marine fossils found by J.H. Reid adjacent to the Stanwell Fault were described by Whitehouse (1946) and Skwarko (1968) and assigned an earliest Cretaceous age. The rocks containing these fossils were included in the Stanwell Coal Measures (now the Stanwell Formation) by Kirkegaard & others (1970). However, the age difference between these quite thin units is so great that they cannot be considered to represent continuous deposition.

Distribution

The unnamed marine sequence occurs about 5km north-west of Bushley railway siding. Reid (1939a) traced the marine sequence for a distance of 4km west of the fossil locality (Figure 214). He also suggested that it extended to the east and was exposed in Stuart Creek 1.5km west of Stanwell. However, these siliceous sandstones containing a high proportion of volcaniclastic detritus and impressions of fossil wood are now believed to belong to unnamed sediments that underlie the Dalma Basalt.

Topographic expression

The beds form low but distinct strike ridges trending east-west along the base of the Native Cat Range.

Geophysical expression

The unit is too small to have a distinctive geophysical expression.

Lithology

The section comprises yellowish brown to reddish brown sandstone, calcareous in part, and a characteristic bed of coarse sandstone (grit) (Reid, 1939a; Whitehouse, 1946). The sandstone ranges from feldspathic to quartzose, the coarser beds having the higher quartz content. It displays both herringbone cross-bedding and hummocky cross stratification.

Environment of deposition

The sedimentary structures and the fossils indicate deposition in a shallow marine environment.

Thickness

The thickness of the marine sequence is at least 75m (Reid, 1939a).

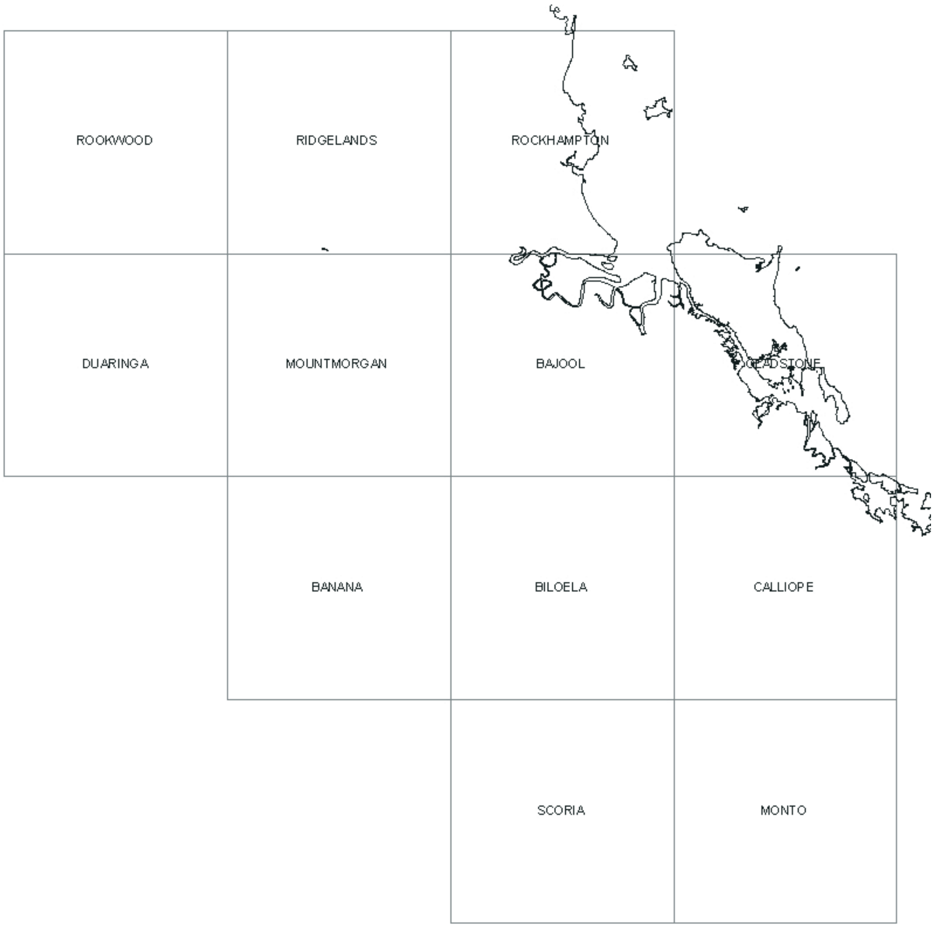


Figure 214. Distribution of the unnamed marine Cretaceous sediments (Kl)

Structure

The strata are steeply to vertically dipping, and are at least locally overturned. Their strike is close to east-west. These steep dips are attributed to south-block-down normal faulting along the east-trending Stanwell Fault.

Biostratigraphy and age

Whitehouse (1946) regarded the marine fauna as earliest Cretaceous (early Neocomian), whereas Skwarko (1968) considered it to be slightly younger (late Neocomian). Both ages are significantly older than the Stanwell Formation (early Albian).

Stratigraphic relationships

The unit is interpreted to be in fault contact with the older Native Cat Andesite to the north and the younger Stanwell Formation to the south.

Correlation with other units

The marine sequence is slightly older than the Early Cretaceous marine transgression in the Maryborough and Surat Basins (Whitehouse, 1946; Day & others, 1983), suggesting that the sea may have entered south-eastern Queensland through the Rockhampton region.

Unnamed sediments (Kn)
(C.G. Murray)

Introduction

Jack (in Jack & Etheridge, 1892), Rands (1892), and Dunstan (1898b) all described sediments beneath the basalts exposed in Stuart (Stewart) Creek about 5km north-west of Stanwell. These were not examined during the Yarrol Project fieldwork, but an extension of this unit, previously referred to as Cainozoic sediments (Kirkegaard & others, 1974), was inspected about 15km north-north-west of Stanwell.

Distribution

No outcrop of the unit has been mapped along Stuart Creek, where the sediments are almost entirely covered with basalt (Rands, 1892). The sediments at this location were described as being limited in extent and of no great depth (Dunstan, 1898b). An outcrop about 4km long and up to 1.5km wide straddles Deep Creek 15km north-north-west of Stanwell. The poorly consolidated sediments beneath basalt in Hunters Gully 10km west-south-west of Ridgelands (Reid, 1931b, 1934) may be equivalent, and there is a small outcrop containing fossil wood about 3km to the north-north-west that may belong to the unnamed Lower Cretaceous sequence. Another possible outlier is north of Mount Lion. Steeply dipping rocks containing fossil wood in the bed of Stuart Creek near its junction with Neerkol Creek (Reid,1939a) may represent the same unit (Figure 215).

Type area

No type area has been defined, but the unit is exposed on the ridge east of Deep Creek, about 15km north-north-west of Stanwell.

Topographic expression

The outcrop north of Stanwell forms a low, thickly wooded ridge that trends roughly north-south.

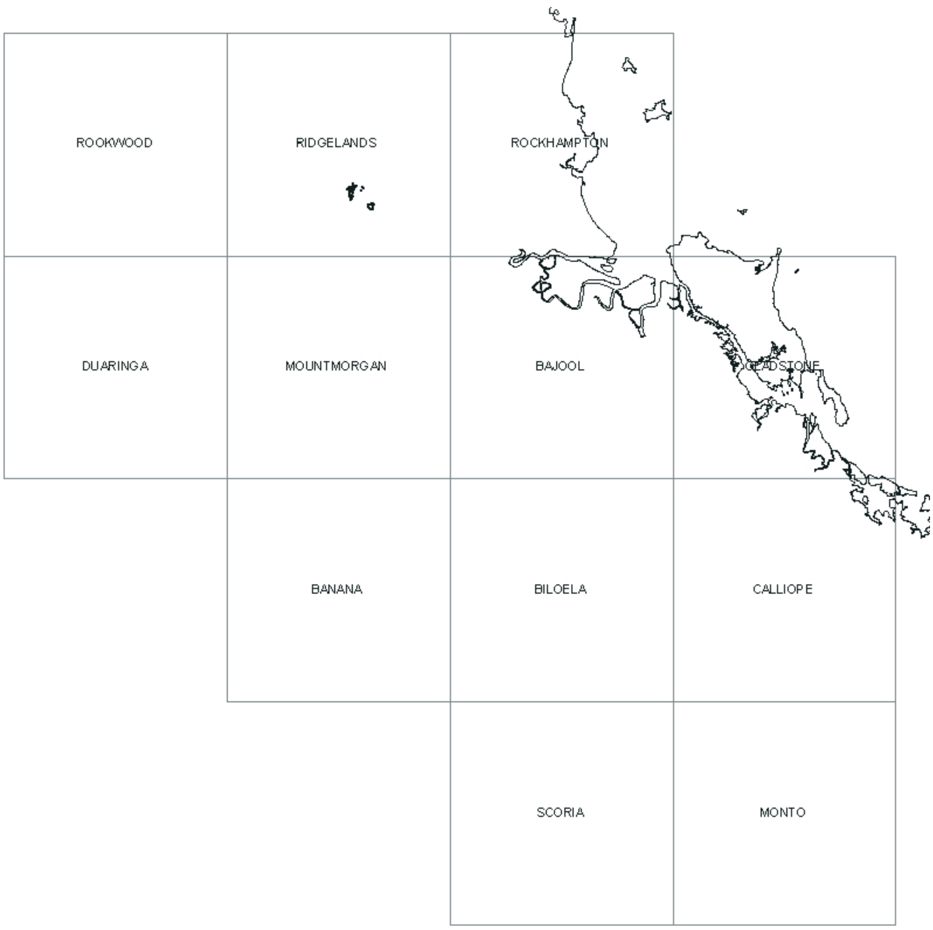


Figure 215. Distribution of the unnamed Cretaceous sediments

Geophysical expression

The outcrop north of Stanwell has a blue-green colour on images derived from airborne radiometric data, similar to the response from the Precipice Sandstone and reflecting the lack of potassium in these rocks. The magnetic signature is very low.

Lithology

The beds along Stuart Creek were described as sandstone and dark carbonaceous shale with thin, impure coal seams beneath and interbedded with basalt (Jack & Etheridge, 1892). Plant fossils were mainly found in ferruginous sandstone. Dunstan (1898b) also listed plant-bearing beds of white volcanic tuff, more argillaceous than sandy, and conglomerate at this locality.

The outcrop straddling Deep Creek is composed of poorly consolidated sedimentary rocks, dominantly sandstone. The sandstone is white, and consists of lithic fragments (?felsic volcanics), weathered feldspar, and quartz. The sequence contains quite a high proportion of ironstone concretions. Sorting is moderate to poor, and is better for finer-grained specimens. Shales are rare, and brown in colour. Conglomerate is sparse and fine-grained. Plant remains are common, and consist mainly of fossil wood and strap-shaped leaves.

Reid (1931b, 1934) described the sediments at Hunters Gully as mainly clays, with coarser bands of “wash”.

Thickness

The sequence on the ridge north-north-west of Stanwell is no more than 15m thick, and is absent from a central saddle where basement rocks are exposed. The sequence along Stuart Creek is probably even thinner, and that at Hunters Gully is only about 10m (Reid, 1934).

Structure

The rocks are flat-lying and not deformed, except for the possible equivalents in Stuart Creek near its junction with Neerkol Creek, which dip to the south at about 60° (Reid, 1939a), and the sediments below basalt at Hunters Gully, which dip to the north-east at 14° (Reid, 1934). There is no evidence of folding.

Biostratigraphy and age

Plant fossils collected from Portion 1288 along Stuart (Stewart) Creek were allocated a Jurassic age by Walkom (1915, 1917), who correlated them with the Walloon Coal Measures. Similar fossils from volcaniclastic strata of the Wycarbah Volcanics near Wycarbah are now regarded as Early Cretaceous, and the unit is now assigned this age.

Stratigraphic relationships

The unnamed Lower Cretaceous sediments unconformably overlie the Upper Devonian to Lower Carboniferous Mount Alma Formation, the Lower Carboniferous Rockhampton Group, and the Triassic Native Cat Andesite. They are overlain and partly interbedded with the Lower Cretaceous Dalma Basalt, and by Quaternary alluvium.

Correlation with other units

Lithological and stratigraphic similarities and the occurrence of the same floras indicate that the unnamed Lower Cretaceous sediments are equivalent to a volcaniclastic unit at the base of the Wycarbah Volcanics.

Economic significance

The “wash” in sediments overlain by basalt in Hunters Gully at Morinish is gold-bearing in part. Material treated from this locality averaged about 30g/t gold.

Figure 216. Distribution of the Wycarbah Volcanics

Wycarbah Volcanics (Kvw, Kvw_i)

(A.D.C. Robertson & C.G. Murray)

Introduction

The presence of acid volcanics containing "agate and chalcedony as pisolitic concretions" near Wycarbah was reported by Jack (1889). Dunstan (1898b, 1900) examined some of the volcanic rocks around Mount Hay and noted that the volcanic rocks contained several potential gemstone minerals. Kay (1981) carried out a detailed investigation of the gemstone-bearing rocks between Mount Hay and Wycarbah, and concluded that the sequence was of Early Cretaceous age.

Distribution

South of the Native Cat Range, groups of rhyolitic, trachytic and monzonite porphyry plugs are accompanied by rhyolitic and trachytic flows and pyroclastic deposits (Kirkegaard & others, 1970; Kay, 1981). The main area of volcanics extends from Wycarbah to north-west of Sugarloaf Mountain, over an area 15km by 6km. The largest group of plugs occurs south of Black Mountain at the western end of the Native Cat Range in the Fred Creek – Sandy Creek area. A smaller area of volcanism, containing two plugs, extends south of the Capricorn Highway from Mount Hay to Wycarbah. Isolated felsic intrusives occur near Sebastopol, in the headwaters of Gogango Creek and Sandy Creek (south-west of Westwood) and to the west of Black Mountain (Figure 216).

Derivation of name

The name Wycarbah is derived from the small settlement of the same name that is located approximately 35km west-south-west of Rockhampton on the main central railway.

Type area

The type area for the Wycarbah Volcanics is given as that section of rocks exposed between the Wycarbah plug and the plug forming Mount Hay but excluding igneous material of the Lower Permian Rookwood Volcanics?, Upper Permian to Lower Triassic Umbrella Granodiorite and Moonkan Granite of the Bouldercombe Igneous Complex and Lower Cretaceous Dalma Basalt (see Kay, 1981).

Topographic expression

The volcanic sequence has suffered extensive erosion with several plugs forming steep sided domes rising up to 200m above the surrounding countryside, and culminating in the prominent peak of Mount Hay at 376m. The remnant flows to the south of Black Mountain form flat topped ridges, indicating that the unit has not been subjected to major faulting and/or folding episodes. The native vegetation has been retained over much of the unit.

Geophysical expression

The Wycarbah Volcanics give a strong radiometric response, showing white on images derived from airborne data, and contrasting with most of the surrounding units. It cannot be distinguished from the Bushley Dacite to the east.

Although no magnetic susceptibility readings were taken, it is assumed that the magnetic response is low. This low magnetic signature is hidden by the moderate to strong magnetic response of the basement rocks.

Lithology

The Wycarbah Volcanics consist of number of rhyolite and trachyte plugs accompanied by rhyolitic and trachytic flows and pyroclastic deposits. Kay (1981) divided the eruption sequence in the Mount Hay area into three phases.

The rhyolite varies from fine- to medium-grained, greenish grey to green material, to a coarse grained, spherulitic, brecciated and flow banded phase. Groundmass can vary from very fine-grained equigranular to one containing flow orientated feldspar laths. Small clusters of sodic pyroxene and opaque oxides can accompany the groundmass minerals.

Spherulitic rhyolite varies significantly. The spherulitic rhyolite from the Mount Hay area is fine-grained, light red-brown to buff in colour, with sporadic patches of thunder egg development and agatised rhyolite (Kay, 1981). Brown, green and red coloured agatised rhyolite, containing numerous agate-filled spheruloids, occurs between the Mount Hay and Wycarbah plugs. The bases of some rhyolite flows are composed of perlite that can form layers up to 3m thick (Kay, 1981). The rhyolite derived from the Mount Hay vent ranges in colour from white to grey to buff, and comprises up to 70% spherulites (5mm in diameter) set in a fine-grained matrix. This rhyolite commonly has a thin rhyolitic tuff layer at the base of the flow and is vesicular near its upper surface. Kay (1981) reported that the spherulitic rhyolite from the Eggrock area differs markedly from the rhyolite from Mount Hay. The Eggrock material varies from white to dark red-brown in colour, is commonly brecciated and flow banded, and is strongly siliceous. Trachytic fragments are numerous and the spheruloids are very variable in size throughout the flow.

Trachyte is mainly fine- to medium-grained, light grey to greenish grey in colour, and contains sanidine feldspar as the dominant phenocryst phase. Occasionally, biotite develops as small phenocrysts. Riebeckite develops as a groundmass phase in some trachyte flows.

The pyroclastic deposits associated with the plugs vary from coarse vent breccias to fine ash tuffs. Around the Wycarbah plug, the coarse vent breccia is composed of dark reddish-brown, fine-grained siliceous spherulitic rhyolite similar in all respects to the material forming the plug. Grey to fawn coloured lapilli and ash tuffs crop out to the west of Bald Rock and south of the High Valley road. These tuffs contain the *in situ* fossil remains of a stand of Early Cretaceous vegetation. Numerous tree trunks, tree stumps and branches are preserved in the ash. Ash tuff containing plant remains also occurs in the vicinity of Wycarbah railway siding. Outcrops of coarse breccia with fossil tree stumps can be seen along the road up Sandy Creek about 2km south-east of Bushley (Figure 217), negating Kay's conclusion that the breccia becomes finer grained away from the plugs at Wycarbah and Mount Hay. The preservation of plant material suggests that this deposit was not formed at high temperatures, and probably should not be termed a tuff. The breccia contains fragments of finer-grained beds (Figure 218).

Buff coloured, porous lapilli tuff, and fine-grained, siliceous ash tuff overlies spherulitic lava north-west of the Wycarbah plug. The coarse-grained tuff commonly contains slaty lithic fragments and both the lapilli and

Figure 217. Tree trunk in volcanic breccia, Sandy Creek road, 2km south-east of Bushley, Wycarbah Volcanics

Figure 218. Coarse volcanic breccia containing fragments of finer-grained beds, Sandy Creek road, 2km south-east of Bushley, Wycarbah Volcanics

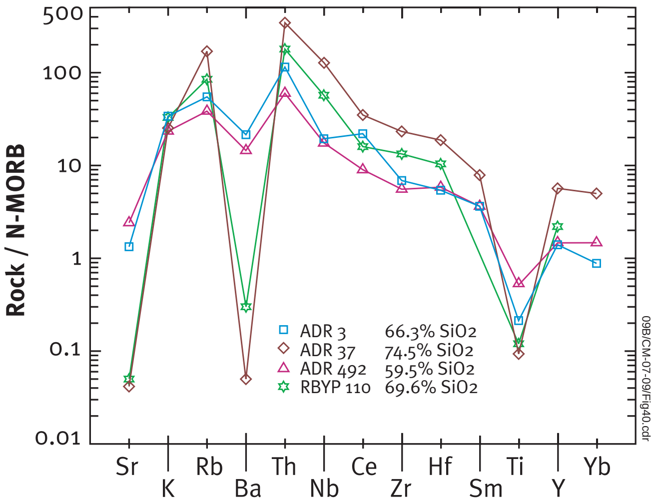


Figure 219. Spidergram plot of trachyandesite to rhyolite. N-MORB values are from Pearce (1983)

ash tuff exhibit graded bedding (Kay, 1981). Kay also reported that a fine-grained tuffaceous trachyrhyolite cut by siliceous rhyolitic ring dykes forms the summit of Mount Hay and extends a short distance down the northern and north-western slopes of the mountain, and found several small outcrops of basalt lying stratigraphically below the rhyolitic volcanics.

Geochemistry

Of four analyses from the Wycarbah Volcanics, three are from intrusive plugs, and one from a pyroclastic deposit. All are typical within plate analyses, with no depletion of Nb which has relatively high values. $(Th/Nb)_N$ is low, ranging from 2.7 to 12.15. Fractionation has obviously increased SiO_2 as it decreased Sr, Ba and Ti (Figure 219). The analyses resemble those of samples from the Mount Hedlow Trachyte reported by Sutherland & others (1996), implying a similar tectonic environment. In discriminant plots of Pearce & others (1984), the analyses fall in the Within Plate Granite field (Figure 198).

Thickness

In the Bald Rock area, north-west of Mount Hay, the estimated thickness is approximately 110m of pyroclastic deposits and rhyolite flows. In the vicinity of Mount Hay, an estimated maximum thickness of 80m is proposed for the volcanic deposits.

Structure

Apart from minor faulting, there is little evidence of earth movements affecting the volcanics.

Biostratigraphy and age

Plant fossils from fine-grained volcaniclastic sediments near Wycarbah railway siding were first reported by Tenison Woods (1883). They were grouped with fossils from sediments below basalt in Stuart Creek 5km north-west of Stanwell, and discussed by Jack & Etheridge (1892), Dunstan (1898b) and Walkom (1915, 1917). The fossils were regarded by Walkom (1915, 1917) as Jurassic, but Kay (1981) proposed an Early Cretaceous age for the volcanics at Wycarbah, where he found the fossil *Ptilophyllum pecten*.

An imprecise Rb-Sr isochron of 113Ma was obtained from trachyte of the Mount Sugarloaf plug and a sample of Dalma Basalt 4km to the north-north-west. Zircon recovered from the soil profile on top of Mount Sugarloaf provided a $^{206}Pb/^{238}U$ age of 128.0±0.4Ma and a $^{207}Pb/^{235}U$ age of 129.1±0.6Ma (Lower Cretaceous). Pyroclastic deposits (GR208083 7392259 on the Mount Morgan 1:100 000 Sheet area) north-west of Mount Sugarloaf yielded two populations of zircon. The colourless zircon gave a $^{206}Pb/^{238}U$ age of 120.2±0.3Ma and a $^{207}Pb/^{235}U$ age of 121.4±0.4Ma, similar to the SHRIMP age obtained from the Mount Salmon Volcanics, while the amber coloured zircon gave ages of 108.5±0.3Ma and 109.3±0.3Ma (Appendix 1).

A zircon sample collected from the intrusion to the west of Sebastopol provided a $^{206}Pb/^{238}U$ age of 237.8±0.7Ma and a $^{207}Pb/^{235}U$ age of 240.3±0.9Ma (Middle Triassic) (Appendix 1), and is considered to be a xenocryst derived from basement rocks. This age is very similar to the age derived from xenocrystic zircon

(242±5Ma) from Jim Crow Mountain (Mount Hedlow Trachyte), north-east of Rockhampton (Sutherland & others, 1996).

On the basis of a plant fossil age and ages derived from zircon recovered from elements of the Wycarbah Volcanics, an age of Early Cretaceous is proposed for this volcanic unit.

Stratigraphic relationships

The Wycarbah Volcanics either intrude and/or unconformably overlie Permian Moah Creek beds, Dinner Creek Conglomerate, and ?Rookwood Volcanics and the uppermost Carboniferous to Lower Permian Youlambie Conglomerate, as well as the Permian to Triassic Bouldercombe Igneous Complex, and Triassic Native Cat Andesite. They are assumed to overlie the latest Triassic to earliest Jurassic Bushley Dacite, but the contact has not been seen. The volcanics are themselves overlain by Lower Cretaceous Dalma Basalt and Quaternary alluvium.

Correlation with other units

The Wycarbah Volcanics have many similarities to the Mount Salmon Volcanics, but may be more rhyolitic than the latter unit. They are overlain by the Lower Cretaceous Dalma Basalt, which appears to form the base of the Mount Salmon Volcanics. Nevertheless, isotopic ages from the Wycarbah and Mount Salmon Volcanics overlap in part, and it is probable that they represent the same episode of within plate volcanism in the Early Cretaceous.

Economic significance

In the vicinity of Mount Hay, a number of mining leases have been taken out for gemstone and lapidary materials. The Central Queensland Gem and Mineral Company Pty Ltd is working a spherulitic rhyolite flow for thunder eggs and coloured agatised rhyolite and jasper, amethyst, smoky quartz and clear quartz crystals are also recovered. Three kilometres north-west of Mount Hay another small deposit of thunder eggs and agatised rhyolite has been placed under lease. Outside the lease areas, amethyst crystals have been recorded in vughs developed in trachyte flows in the vicinity of Bald Rock.

Dalma Basalt (Kvd)

(C.G. Murray)

Introduction

Kirkegaard & others (1970) mapped and briefly described basalt flows in the area between Westwood, Ridgelands, Rockhampton and Mount Morgan, and assigned them a Late Cretaceous age. This age appeared to be confirmed by Wellman (1978), who obtained a K-Ar date of 67.3±2.6Ma from basalt at Alton Downs north-west of Rockhampton. Recent mapping has shown that basalts overlying the Precipice Sandstone in the Mount Morgan – Westwood–Stanwell–Dalma area are older than the Upper Cretaceous Alton Downs Basalt, being overlain by the Lower Cretaceous Stanwell Formation. These older basalts are here named the Dalma Basalt.

Distribution

The Dalma Basalt crops out in three main areas. The furthest north is east of Mount Salmon, where it forms a north-north-west trending belt 6km long and 2km across. Around Dalma it occupies an area about 8km north-south and about 5km east-west, with a smaller offshoot along Stuart Creek. The third main area is west and south-west of Wycarbah, where it forms an equant area about 8km in a north-south direction and up to 6km wide. The Dalma Basalt also occurs as scattered small outliers overlying the Precipice Sandstone as far as Mount Morgan in the south-east (Figure 220).

Derivation of name

The name is taken from the Dalma district, 17km north-north-west of Stanwell, where typical rocks of the unit are exposed.

Figure 220. Distribution of the Dalma Basalt

Type section

The type section is along the Stanwell–Waroula road from 13km north-west of Stanwell north to the abandoned Dalma school. The unit is also well exposed as a series of flows forming terraces west of Wycarbah.

Topographic expression

North of Neerkol Creek and west of Wycarbah, the Dalma Basalt forms low plateaux and undulating hills covered with dark brown soil, and occupies stream valleys in this region. Much of this area is cultivated. To the south, it crops out as flow remnants forming small terraces overlying the eroded surface of the Precipice Sandstone that rises in elevation towards Mount Morgan.

Geophysical expression

The magnetic response of the Alton Downs Basalt is most obvious on images of the first vertical derivative, which exhibit the typical "mottled" effect produced by thin basaltic flows.

The basalt gives a low radiometric response.

Lithology

The basalt is dark grey, fine grained, and sparsely porphyritic in plagioclase and olivine.

Geochemistry

The Dalma Basalt is very uniform in composition. It consists mainly of high-K basalt, with some hawaiite and potassic trachybasalt (Figure 221). On spidergrams, it displays a typical within plate pattern, with $(Th/Nb)_N$ ratios from 1.12 to 1.75. The rocks are enriched in LREE compared to HREE, and the REE patterns show a

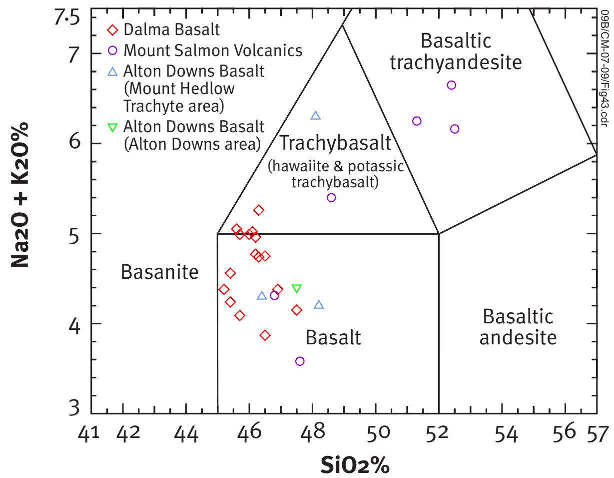


Figure 221. Total alkalis - SiO_2 plot showing analyses from the Dalma Basalt, Alton Downs Basalt (both from the type area and around the Mount Hedlow Trachyte), and basalt of the Mount Salmon Volcanics classified according to the scheme of Le Maitre (1989).

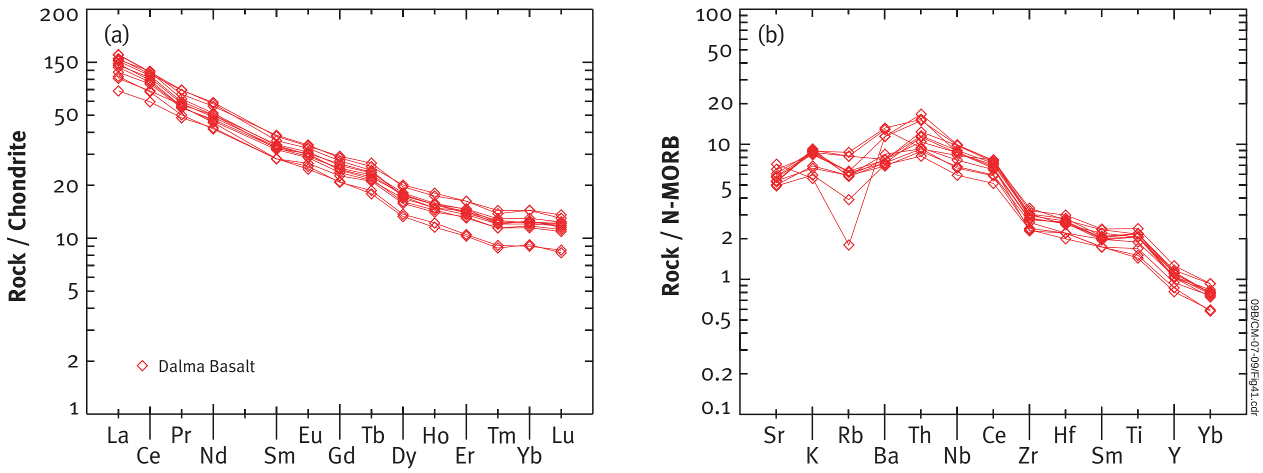


Figure 222. (a) REE and (b) spidergram plots of basalts from the Dalma Basalt. Chondritic REE values are from Sun & McDonough (1989), and N-MORB from Pearce (1983).

considerable slope, with $(La/Yb)_N$ ranging from 5.86 to 10.69 (Figure 222). On standard triangular discrimination diagrams (Pearce & Cann, 1973; Meschede (1986; Cabanis & Lecolle, 1989) and the Zr-Ti plot of Pearce (1982), the analyses fall in the Within Plate Basalt and Continental Basalt fields (Figure 223). A basalt sample from the type area of the Alton Downs Basalt (Sutherland & others, 1996) plots together with the Dalma Basalt on these diagrams, except for the La-Y-Nb diagram of Cabanis & Lecolle (1989).

Thickness

The maximum thickness of basalt reported from water bores in the Dalma Basalt is 67m from the region of the Dalma school (no basement reached). Exploratory drilling for coal in the area south-east of Mount Carpenter penetrated basalt to depths up to 180m, but the sequence could include Native Cat Andesite.

Structure

The basalt is essentially flat-lying. West of Wycarbah railway siding, the basalt flows dip easterly at a low angle, producing low strike ridges readily observed on air photography. Poor outcrop precludes a detailed examination of the structure. On the basis of the variability in mineralogical composition and the widespread distribution of the basalt, a number of vents are considered to have been active during the eruption cycle. The present outcrop limits are mainly erosional.

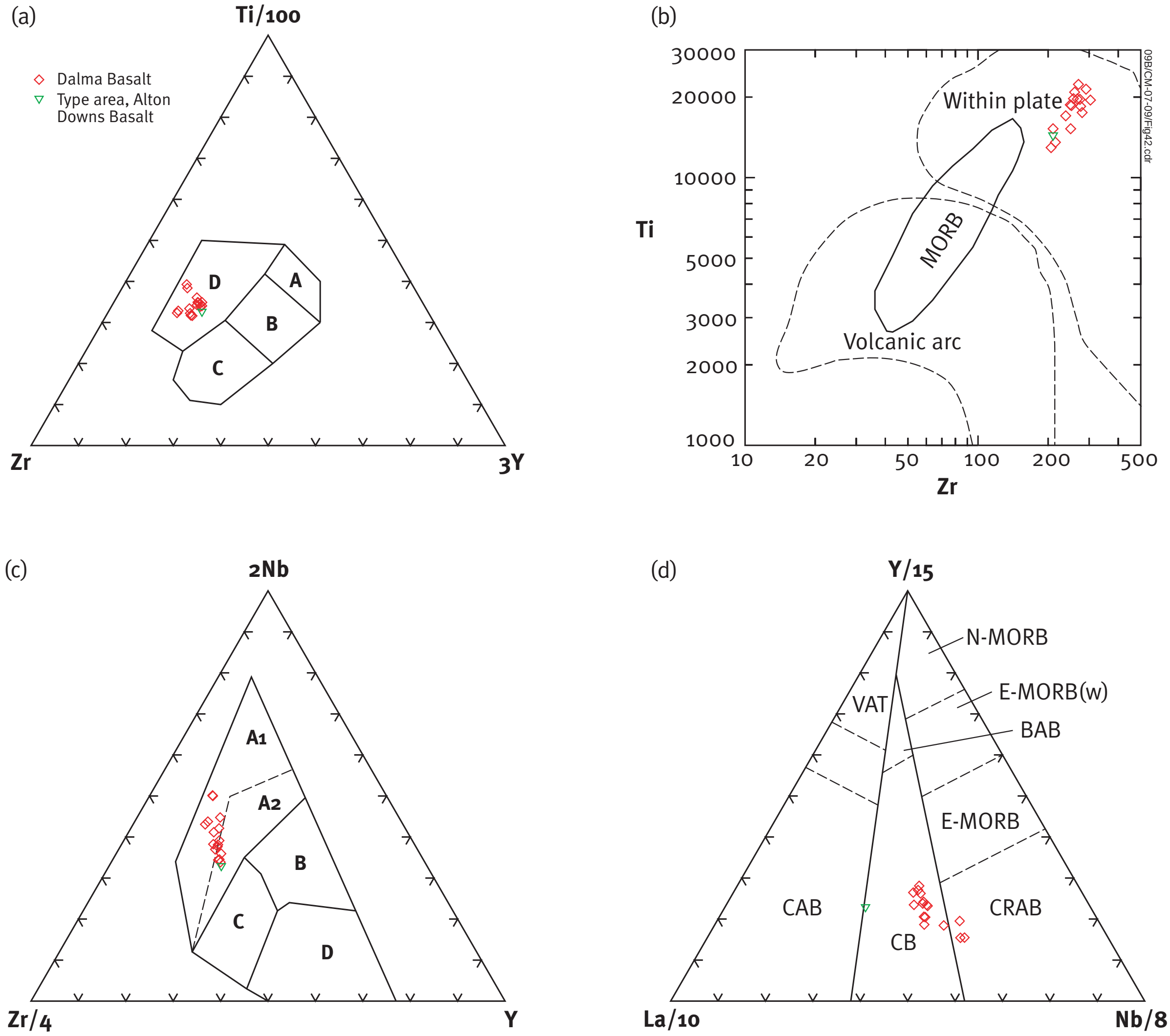


Figure 223. Analyses of basalts from the Dalma Basalt plotted on discriminant diagrams of (a) Pearce & Cann (1973): A = island arc tholeiites, B = island arc tholeiites, volcanic arc basalts and mid-ocean ridge basalts or MORB, C = calc-alkali basalts, D = within plate basalts; (b) Pearce (1982); (c) Meschede (1986): A1 = within plate alkali basalts, A2 = within plate alkali basalts and within plate tholeiites; and (d) Cabanis & Lecolle (1989): CB = continental basalts, CRAB = alkali basalts from continental rifts, CAB = calc-alkali basalts, VAT = volcanic arc tholeiites (overlap between these fields is indicated by the dashed lines). The analyses are essentially confined to the within plate basalt, within plate alkali basalt and continental basalt fields on these diagrams.

Age

Early Cretaceous volcanism covered an extensive area between Ridgelands, Westwood and Stanwell. The volcanism started with a dominantly trachytic phase (the Wycarbah Volcanics), then became mainly basaltic (the Dalma Basalt), and ended with more trachyte (the Mount Salmon Volcanics). These units are gradational and interbedding of the various flows is to be expected. However, on a regional scale, the Dalma Basalt overlies the Lower Cretaceous Wycarbah Volcanics, and is overlain by the Lower Cretaceous Mount Salmon Volcanics. It is also overlain by the Stanwell Formation. Its age must therefore be Early Cretaceous, about 120Ma or Aptian.

Stratigraphic relationships

The Dalma Basalt is unconformable on most underlying units, including the Mount Alma Formation, Rockhampton Group, Lorray Formation, Dinner Creek Conglomerate, Ridgelands Granodiorite, Native Cat Andesite, and Precipice Sandstone. An angular unconformity between the Precipice Sandstone and Dalma Basalt is exposed around the eastern side of Flagstaff Hill near Stanwell. At most locations, the relationship between these two units is a disconformity, with clear evidence of erosion of the Precipice Sandstone before the basalt was erupted. The relationship with unnamed Lower Cretaceous sediments that underlie the Dalma Basalt in the Stuart Creek and Dalma region north-west of Stanwell (eg. Dunstan, 1898b, figure 6) is unknown, but the two units may be conformable. So too may be the Dalma Basalt and the underlying Wycarbah Volcanics near Mount Hay (Kay, 1981).

The Dalma Basalt appears to form the base of the Mount Salmon Volcanics, and is overlain by the Lower Cretaceous Stanwell Formation at localities near Bushley, possibly conformably.

The Dalma Basalt comes within 2.5km of the westernmost outcrop of the Alton Downs Basalt 17km north of Stanwell. The two units are separated only by a thin belt of Upper Devonian to Lower Carboniferous metasediments, and appear identical in all respects. The actual limits of the two units may therefore differ from that shown on the Ridgelands 1:100 000 map sheet.

Economic significance

No mineralisation of economic significance has been reported from the Dalma Basalt. Sediments within or below the basalt flows are important aquifers.

Mount Salmon Volcanics (Kvs)

(A.D.C. Robertson & C.G. Murray)

Introduction

Kirkegaard & others (1970) described unnamed trachytic volcanics occurring between the Native Cat Range and Mount Salmon to the north as well as several isolated outcrops. At the time of their mapping, Kirkegaard & others (1970) indicated that the age of these felsic volcanics was unknown. They considered that the volcanics were probably related to the Mount Hedlow Trachyte and therefore of Late Cretaceous age. Simpson (1995) briefly described the felsic volcanics in the vicinity of Mount Salmon and proposed the name Mount Salmon Volcanics for the unit.

Distribution

The Mount Salmon Volcanics (Figure 224) extend from Mount Salmon in the north-west to near Stanwell at the south-eastern end of the Native Cat Range, a distance of 35km, and are up to 10km across.

Derivation of name

The name is derived from the prominent topographic feature of Mount Salmon located at the north-western end of the plateau.

Type area

The type area proposed for the Mount Salmon Volcanics is located towards the southern end of the plateau, between “The Lookout” and the headwaters of Bonley Creek. Typical exposures are also seen along the track up the plateau from the east beside Limestone Creek.

Topographic expression

The Mount Salmon Volcanics form a dissected plateau. For the most part, it is bounded by cliffs, except at the southern end where the plateau abuts the Native Cat Range. Trachyte and rhyolite flows within the volcanic sequence generally form benches and low cliffs (Figure 225). Dense vegetation covers a large part of the plateau.

Geophysical expression

The lack of magnetic material within the Mount Salmon Volcanics results in a poor magnetic response from the trachytic lava and pyroclastic deposits. First vertical derivative aeromagnetic images show a strong south-easterly trending linear magnetic feature in basement rocks along the western side of the volcanic sequence from south east of Mount Salmon to below “The Lookout”. Other linear magnetic features are present in the basement rocks and possibly represent basement faults.

The radiometric response of the felsic volcanics is strong with potassium being the dominant element. On false colour images, the Mount Salmon Volcanics appear as a white to pink coloured unit that stands out from most of the surrounding geological units, although they merge with the Upper Carboniferous Lorray Formation to the north-west. Where basalt is exposed within the acid volcanics, the spectral response is much more subdued. The radiometric response is also low in areas covered by sedimentary rocks and thick soil.

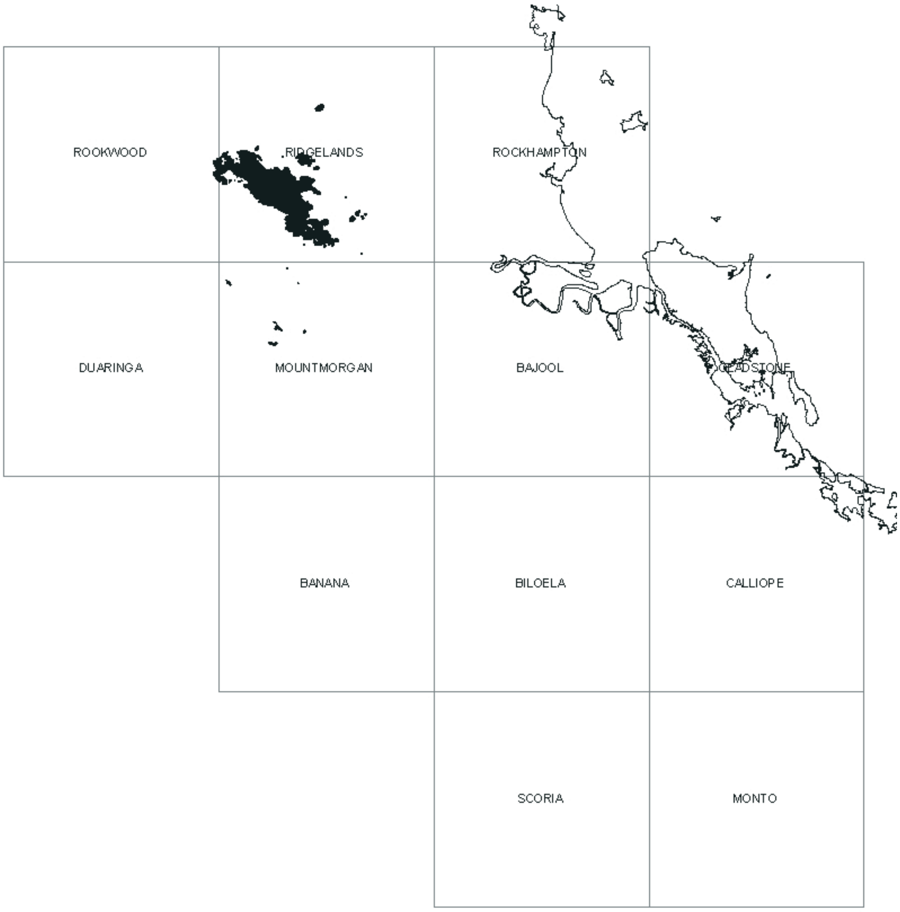


Figure 224. Distribution of the Mount Salmon Volcanics

Figure 225. Benches formed by trachytic volcanics north of Limestone Creek, Mount Salmon Volcanics

Lithology

The Mount Salmon Volcanics are composed of trachytic flows, tuffs, breccias and ignimbrite with accompanying rhyolite and basalt.

Trachyte flows are mostly fine-grained and grey in colour but weather to an off white, pink or purplish colour. Phenocryst feldspars include euhedral sanidine and sodic plagioclase set in a groundmass of feldspar, riebeckite, magnetite and cryptocrystalline quartz. In some of the trachyte flows the groundmass feldspar occurs as fine laths that exhibit a strong trachytic texture. In other flows, no defined orientation of the feldspar can be observed and the feldspar forms a felted mass. Magnetite, usually oxidising to hematite, occurs as dust-like particles scattered throughout the groundmass. When present, riebeckite forms as needle-like laths. Also, the riebeckite can be observed with glomeroporphyritic texture that imparts a spotted appearance to the trachyte. Small biotite laths occur in some trachyte samples and apatite occurs in trace amounts. Flows can be close to rhyolite in composition, with the addition of small amounts of quartz and biotite, and the presence of anorthoclase cores rimmed by sodic plagioclase. The main mafic is aegirine-augite that forms glomeroporphyritic clusters. The presence of arfvedsonite/riebeckite, and aegirine/aegirine-augite, ± aenigmatite in the groundmass of some of the trachytes and rhyolites characterise these volcanics as belonging to a peralkaline series.

Breccias are widespread throughout the sequence. Most are matrix supported with subrounded trachyte and rhyolite clasts having a mineralogical composition similar to the trachyte and rhyolite flows. Clast size varies. Some breccias in the southern part of the outcrop area have clast sizes ranging up to 3m in length. More often the clast size does not exceed 1m (Figure 226). Some finer grained breccias contain purple vesicular trachyte (Figure 227). Some breccias also contain subrounded to rounded clasts derived from the underlying Permian Dinner Creek Conglomerate (mainly quartz and quartzite) and Moah Creek beds. Breccia matrix is usually composed of material similar in composition to the dominant clast type and exhibiting a range of grain sizes from ash particles to small cobbles.

Trachytic ignimbrite was noted at some localities, and may approach rhyolitic compositions (Figure 228). Both trachytic and rhyolitic ash and lapilli tuffs are known to occur in the sequence.

Basalt occurs towards the base of the sequence, particularly in the vicinity of The Lookout. It is mainly feldspar-phyric. Augitic pyroxene and olivine also develop as phenocrysts but rarely do they become the dominant phenocryst phase. Feldspar-rich basalt contains euhedral to subhedral, twinned and normally zoned (labradorite to andesine) plagioclase phenocrysts, set in a groundmass of feldspar laths, altered olivine and

Figure 226. Breccia in trachytic to rhyolitic volcanics, south-east of The Lookout, Mount Salmon Volcanics

Figure 227. Purple trachytic to rhyolitic clasts in breccia, Limestone Creek, Mount Salmon Volcanics

Figure 228. Trachytic to rhyolitic ignimbrite, 2.5km north-east of Sugarloaf Mountain, Mount Salmon Volcanics

granular titanaugite and opaque oxides. Some basalt has feldspar phenocrysts in which the outer rims enclose granules of titanaugite.

Mineralogical variations within the basalt flows are usually confined to the varying percentages of feldspar, augite or olivine making up the phenocryst assemblage and/or the groundmass. Strongly vesiculated flows are not common. Vesiculation is usually restricted to the tops of flows and the vesicles may either be empty or filled with fibrous calcite, a green chloritic mineral or a greenish, waxy clay mineral. Most basalts are fine-grained and are usually dark grey to black in colour. Occasionally, feldspar-rich basalt can be coarse-grained with feldspar phenocrysts exceeding 10mm in length. This tends to give this rock type a grey and black mottled appearance.

Sedimentary rocks occur through the sequence and in an area of low radiometric response on top of the plateau near the headwaters of Limestone Creek. They range from cross-bedded sandstone to cobble conglomerate, and are sourced mainly from the volcanics themselves.

Geochemistry

Two samples of trachyte to rhyolite have been analysed from the Mount Salmon Volcanics. They have typical within plate chemistry, with no Nb depletion and $(Th/Nb)_N$ values of 4.04 and 12.15, and are very similar to samples from the Mount Hedlow Trachyte with equivalent SiO_2 contents (Figure 229). On discriminant diagrams of Pearce & others (1984), the analyses plot in the Within Plate Granite field (Figure 184). Outcrops which overlie the Native Cat Andesite along the Native Cat Range have quite different geochemistry, and are now regarded as part of that unit.

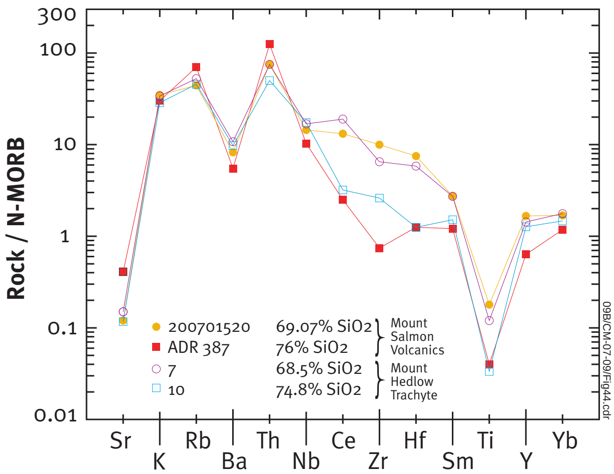


Figure 229. Spidergram plot comparing trachyte and rhyolite of the Mount Salmon Volcanics (unpublished analysis from Geoscience Australia) with rocks of similar SiO_2 content from the Mount Hedlow Trachyte (Sutherland & others, 1996). N-MORB values are from Pearce (1983).

The mafic rocks are basalt, basaltic trachyandesite and hawaiite (Figure 221). On spidergrams, they can be subdivided into three groups based on overall element contents and different $(Th/Nb)_N$ ratios of 1.21–1.25, 1.93–2.53 and 4.7–4.96 (Figure 230b). The range of $(La/Yb)_N$ is more restricted than that for the Dalma Basalt, from 6.03 to 8.51 (Figure 230a). They are within plate basalts, and plot in within plate, within plate alkali basalt, and continental basalt fields on standard discriminant diagrams (Figure 231).

Thickness

Throughout their outcrop area, the Mount Salmon Volcanics are variable in thickness. Kirkegaard & others (1970) proposed a maximum thickness of 120m. In the vicinity of Mount Salmon, approximately 140m of acid lava and pyroclastic deposits are exposed along the north-eastern side of the plateau.

Structure

For the most part, the felsic volcanics are flat lying and show little evidence of folding, especially in the north-western part of the plateau. Overall, they dip gently to the north-east. North of “The Lookout”, the Mount Salmon Volcanics exhibit a moderate north-easterly dip, the result of faulting during the Late Cretaceous to Early Cenozoic.

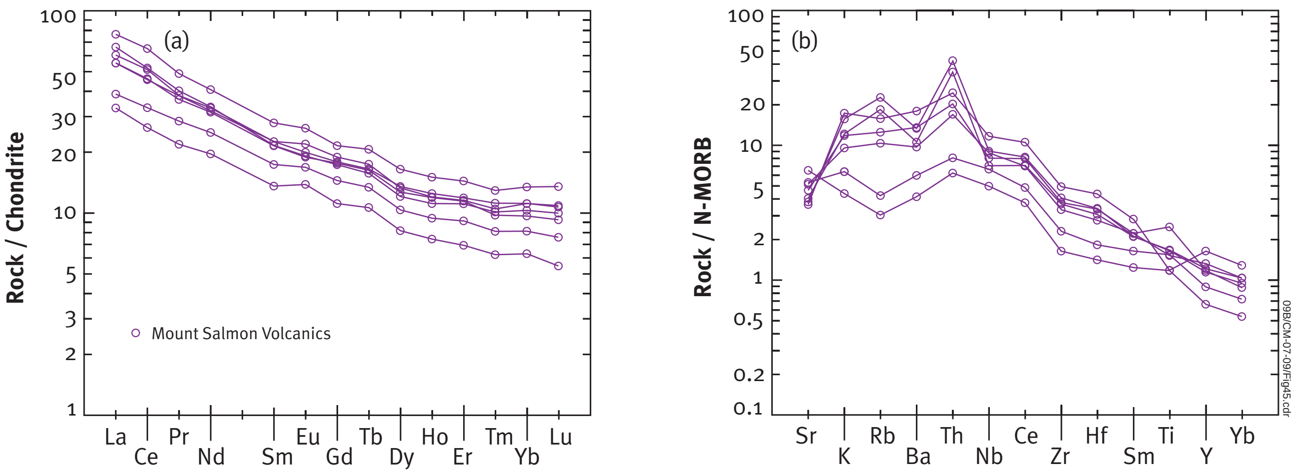


Figure 230. (a) REE and (b) spidergram plots of basalts from the Mount Salmon Volcanics. Chondritic REE values are from Sun & McDonough (1989), and N-MORB from Pearce (1983).

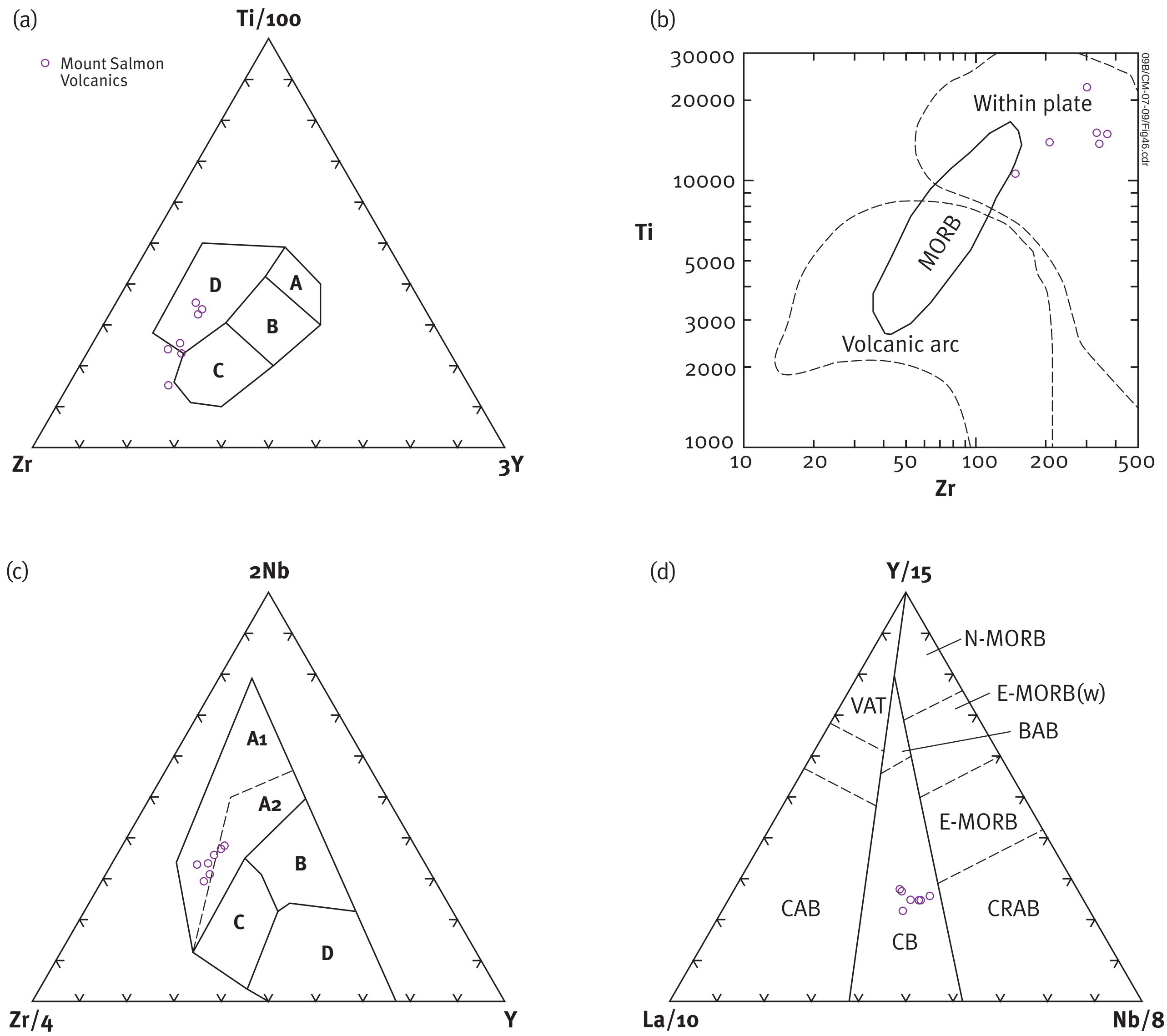


Figure 231. Analyses of basalts from the Mount Salmon Volcanics plotted on discriminant diagrams of (a) Pearce & Cann (1973): A = island arc tholeiites, B = island arc tholeiites, volcanic arc basalts and mid-ocean ridge basalts or MORB, C = calc-alkali basalts, D = within plate basalts; (b) Pearce (1982); (c) Meschede (1986): A1 = within plate alkali basalts, A2 = within plate alkali basalts and within plate tholeiites; and (d) Cabanis & Lecolle (1989): CB = continental basalts, CRAB = alkali basalts from continental rifts, CAB = calc-alkali basalts. The analyses are essentially confined to the within plate basalt, within plate alkali basalt and continental basalt fields on these diagrams.

Age

A SHRIMP U-Pb zircon age from trachyte east of The Lookout is 119.2±1.3Ma, or Early Cretaceous (Cross & others, 2009). This overlaps with results from the Wycarbah Volcanics. The whole of the Mount Salmon Volcanics were probably erupted in a relatively short time interval, and so the age of the Mount Salmon Volcanics is regarded as close to 120Ma, or Aptian in the Early Cretaceous.

Stratigraphic relationships

This volcanic unit unconformably overlies Silurian to Devonian Craigilee beds, Upper Devonian to Lower Carboniferous Mount Alma Formation, Lower Carboniferous Rockhampton Group, Upper Carboniferous to Lower Permian Lorray Formation and Youlambie Conglomerate, Upper Permian Dinner Creek Conglomerate and Moah Creek beds, and Permian to Triassic Ridgelands Granodiorite and Triassic Native Cat Andesite. It is conformable on the Dalma Basalt, which forms the base of the Mount Salmon Volcanics.

The Mount Salmon Volcanics are overlain by Tertiary to Quaternary sediments.

Economic significance

No economic mineralisation has been recorded from the Mount Salmon Volcanics.

Stanwell Formation (Ks)

(C.G. Murray)

Introduction

The Stanwell Coal Measures, first named by Dunstan (1912), were described briefly by Rands (1892), Jack & Etheridge (1892) and Maitland (1895), and in more detail by Dunstan (1898b), and were regarded as correlatives of the Ipswich Coal Measures. It was several decades before their true extent, age, and stratigraphic relations were determined.

The coal measures were originally thought to be more extensive than their present mapped extent, because both the Lower Cretaceous Stanwell Coal Measures and part of the Lower Jurassic Precipice Sandstone were included. Plant fossils collected from Wycarbah and along Stuart Creek during the initial investigations were described by Walkom (1915, 1917), who assigned a Jurassic age and correlated the Stanwell Coal Measures with the Walloon Series, but these fossils were from rocks now interpreted as unnamed Lower Cretaceous sediments and the Wycarbah Volcanics. It was not until much later that palynological studies on core samples from early exploration drilling for coal gave a definitive Early Cretaceous age for the Stanwell Coal Measures (Dettmann & Playford, 1969; Helby & Partridge, 1977). Earliest Cretaceous marine fossils described by Whitehouse (1946) and Skwarko (1963, 1968) are from a separate localised unit adjacent to the Stanwell Fault.

Rands (1892), Maitland (1895) and Dunstan (1898b) interpreted the upper sandstone-dominated part of the Precipice Sandstone (their Desert Sandstone) to overlie the Stanwell Coal Measures, an understandable error in view of the fact that the Precipice Sandstone is topographically higher than the coal measures. The correct stratigraphic relationship between the Stanwell Coal Measures and Precipice Sandstone (formerly Razorback beds) was recognised by Kirkegaard & others (1970) following the palynological results of Dettmann & Playford (1969), subsequently confirmed by Helby & Partridge (1977).

Despite more than a century of intermittent exploration, no coal has been found in the Stanwell Coal Measures. Thin seams intersected in boreholes and in shafts are mainly from other units. The name of the unit is therefore changed to Stanwell Formation.

Distribution

The Stanwell Formation forms an elongate, east-west trending belt of rocks 15km long and up to 3km wide centred roughly on Kalapa (Figure 232), and extending from Stanwell in the east to Kalapa North in the west.

Topographic expression

The unit produces subdued topography, and generally crops out only along some streams. The best outcrops are along the headwaters of Neerkol Creek. Over most of its extent the Stanwell Formation is covered by recent alluvial deposits.

Figure 232. Distribution of the Stanwell Formation

Geophysical expression

On images produced from airborne radiometric survey data, the Stanwell Formation exhibits a deep greenish-blue colour, indicating a low potassium content, typical of Mesozoic sedimentary sequences.

The rocks have very low magnetic susceptibility, and strong magnetic features within the area of outcrop are interpreted to be related to basement rocks.

The area covered by the Stanwell Formation coincides with an east-north-east trending gravity ridge, indicating that the preserved depositional basin is shallow, and that the basement rocks are dense.

Lithology

The sequence in Sanderson's bore No. 2 on Stony Creek was described by Dunstan (1898b) as white shaly sandstone and pale to bluish shale with some hard bands. The outcrop he described along Osborne Creek, a small tributary of Stony Creek, is at least partly in the Precipice Sandstone.

The most comprehensive descriptions of the unit are by Helby & Partridge (1977) from bores drilled during the search for coal. They presented logs of cores from three bores drilled in 1912 (Dunstan, 1912, 1913a). The Hobbs-MacNith bore, which bottomed in basalt, is thought to have been located close to Woodend railway station (see Dunstan, 1913a), and may have been the No.2 site shown by Dunstan (1912). Warrens bore of Helby & Partridge (1977) was probably sunk near Warren railway station (Dunstan, 1913a), close to the No.3 site of Dunstan (1912). Dunstan (1913a) stated that a third bore was located south-east of Warren railway station, and this was described by Helby & Partridge (1977) as the Stanwell bore. The logs show that the sequence consists of interbedded, finely laminated, fine grained green to grey sandstone and dark grey carbonaceous siltstone, with sparse coaly laminae. Bioturbation is common throughout.

Figure 233. Cliff outcrop of interbedded sandstone and siltstone, headwaters of Neerkol Creek, Stanwell Formation; note figure for scale

The best outcrops of the Stanwell Formation are in cliffs along the headwaters of Neerkol Creek (Figure 233). The sequence here is dominantly regularly bedded, pale weathering feldspathic sandstone with interbeds of grey siltstone and minor shale.

Environment of deposition

The occurrence of microplankton, the ubiquitous bioturbation throughout the unit (Figure 234), and the probable presence of glauconite all suggest deposition in a deltaic setting subject to periodic marine inundation (Helby & Partridge, 1977). Conditions were such that no significant accumulation of coal-forming material was possible.

Thickness

The maximum thickness recorded by drilling is about 170m in the Stanwell bore, located south-east of Warren railway station, in which the Stanwell Formation overlies the Precipice Sandstone. Dunstan (1898b) suggested a minimum thickness of 210m from outcrops along Stony and Osborne Creeks, but probably included part of the Precipice Sandstone in this estimate. The unit thins to the west to about 150m in Warrens bore, 90m in the Hobbs-MacNith bore, and from 45 to 20m in coal exploration boreholes west of Kalapa. A cliff outcrop 20m high in the upper reaches of Neerkol Creek (Figure 218) probably represents close to the full thickness of the unit in this area. Modelling of airborne magnetic data gives a thickness of less than 300m (Appendix 5).

Structure

Over most of its area of outcrop, the Stanwell Formation is gently dipping to flat-lying. The detailed mapping by Dunstan (1898b) along Osborne Creek shows an undulating, generally north-dipping sequence with dips of 2° to 7°. Steeper northerly dips mapped in the headwaters of Osborne Creek (12° to 25°) are in strata of the Precipice Sandstone. These steeper dips are the cause of the conflicting stratigraphic and topographic relations between the Precipice Sandstone and the Stanwell Formation. They suggest the presence of a north-dipping monocline in the Precipice Sandstone at its contact with the Stanwell Formation. Reid (1939a) noted a similar monoclinal fold (west-dipping) in the Precipice Sandstone at Flagstaff Hill south of Stanwell (Figure 189). This monoclinal folding with dips towards the centre of the Stanwell Basin accounts for the presence of Precipice Sandstone beneath the Stanwell Formation (Helby & Partridge, 1977) without the necessity of faulting.

Figure 234. Bioturbation in feldspathic sandstone, 2km east-south-east of Bushley Stanwell Formation

Steep dips were noted by Dunstan (1898b) and Reid & Morton (1928) in outcrops in Stuart Creek just west of Stanwell, produced by south block down normal faulting on the Stanwell Fault. The presence of fossil wood in these tuffaceous sandstones dipping at 60° to the south (Reid, 1939a) suggests that they could be unnamed Lower Cretaceous sediments.

Biostratigraphy and age

Plant fossils from Stuart (Stewart's) Creek and Wycarbah described by Walkom (1915, 1917) and figured in Hill & others (1966) from the Stanwell Coal Measures are from volcanogenic sediments of the Wycarbah Volcanics (**Kvw**) and unnamed sediments beneath the Dalma Basalt (**Kn**). Earliest Cretaceous (Neocomian) marine macrofossils described by Whitehouse (1946) and Skwarko (1968) are from a localised unnamed unit (**Kl**) older than and separate from the Stanwell Formation.

An age for the Stanwell Formation was only obtained by palynological studies. Dettmann & Playford (1969) determined an Early Cretaceous (latest Aptian to Albian) age (*Coptospora paradoxa* Zone) on core material. This result was confirmed by Helby & Partridge (1977), who suggested an early Albian age based on both microplankton and spore-pollen associations.

Stratigraphic relationships

The Stanwell Formation overlies the Lower–Middle Triassic Native Cat Andesite, the Lower Jurassic Precipice Sandstone, and the Lower Cretaceous Dalma Basalt. The Dalma Basalt is absent beneath the eastern part of the Stanwell Formation. This could represent its original depositional limits, or the effects of differential erosion. The Stanwell Formation is overlain by the extensive Quaternary alluvium developed along Neerkol Creek. The contact with the unnamed lowermost Cretaceous strata along the southern side of the Stanwell Fault is assumed to be faulted.

Correlation with other units

The Stanwell Formation is at least partly correlative with the Jim Crow beds north-east of Rockhampton, and is equivalent in age to the Burrum Coal Measures in the Maryborough Basin, and the Styx Coal Measures in the Styx Basin (Helby & Partridge, 1977).

Economic geology

Thin coal seams reported by past exploration almost all come from boreholes within other units including the Precipice Sandstone and unnamed Lower Cretaceous sediments at Stuart Creek north-west of Stanwell. The only occurrence definitely within the Stanwell Formation is shown on the log of the Stanwell bore (Helby & Paltridge, 1977) at a depth of about 160m. Prospects of finding a workable deposit appear to be non-existent.

Jim Crow beds

(S.B.S. Crouch, P.R. Blake & C.G. Murray)

Introduction

Exploration beneath the Mount Jim Crow flats north of Rockhampton by Pan D'Or Explorations (Queensland) Pty Ltd and Esso Exploration and Production Australia Inc. (Esso Exploration and Production Australia Inc., 1981) revealed a previously unknown basin of Lower Cretaceous sediments, here named the Jim Crow Basin and Jim Crow beds respectively.

Distribution

The Jim Crow beds do not outcrop, and are completely concealed. The basin is centred about 20km north-north-east of Rockhampton. From gravity and magnetic survey data, it appears to be a wedge-shaped basin widest in the north (possibly up to 7km across), and at least 10km in a north-south direction (Figure 235).

Topographic expression

The beds have no topographic expression, being concealed in a low-lying area.

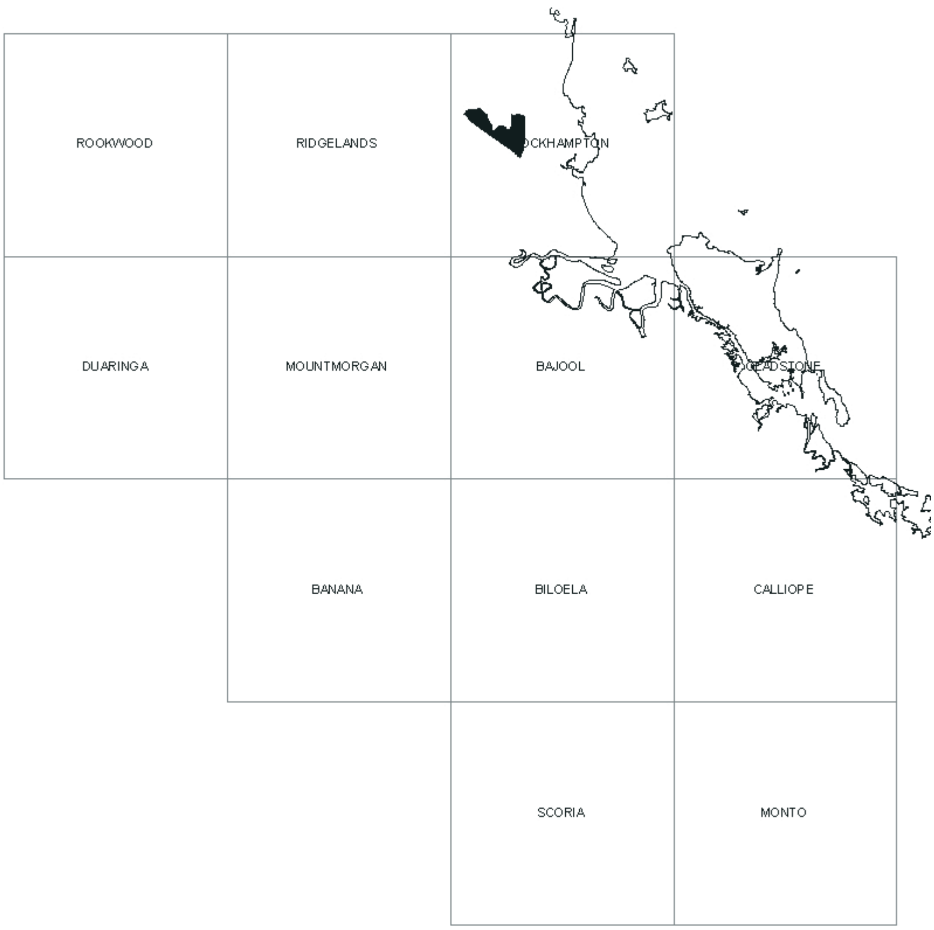


Figure 235. Location of the Jim Crow Basin

Geophysical expression

The beds have a low magnetic signature, and are confined to a basin which coincides with a negative gravity anomaly. The thickest sedimentary sequence in drillholes does not coincide with the gravity minimum, suggesting that granitic basement is partly responsible for the gravity low.

Lithology

Descriptions come entirely from examination of three drillcores from near Cawarral, west of the landing ground along Splitters Creek, and just east of Mount Jim Crow (Esso Exploration and Production Australia Inc., 1981). The Jim Crow beds are described as consisting mainly of greenish-grey, fine-grained, lithic sandstone with mudstone clasts, siltstone and mudstone containing pyrite aggregates. The sandstones are glauconitic. Minor intraformational conglomerate is present, with siltstone and mudstone clasts, and rare quartzose sandstone. The descriptions from the three drillholes indicate that the proportion of sandstone increases to the west. The sequence is cross bedded, and exhibits small scale slump structures and bioturbation. Some magnesite occurs towards the top of the beds.

Environment of deposition

A shallow marine environment is indicated by the faunas and bioturbation and the presence of glauconite. No evidence has been found that suggests a change to freshwater conditions.

Thickness

The drillhole north of Cawarral and another near Splitters Creek reached 181m and 153m, respectively, without intersecting basement. The third hole just east of Mount Jim Crow, which coincides with the gravity minimum, penetrated 93m of alluvium and sediments before intersecting granitic basement. Analysis of waveforms in the airborne magnetic data by means of a Fourier transform suggests a magnetic layer at 350m (Appendix 5).

Structure

The Jim Crow beds are flat-lying and undeformed. Some parts of the sequence are brecciated. The extent of the basin is interpreted from gravity and magnetic data. It is uncertain whether the margins of the basin are fault-controlled or merely onlap the older rocks. Some of the bounding faults may be younger than and therefore cut the basin sequence.

Biostratigraphy and age

Fossil pelecypods, dinoflagellates and foraminifera indicate an Early Cretaceous (middle Albian) age (Esso Exploration and Production Australia Inc., 1981). The sequence is assigned to the Endoceratium ludbrookii dinoflagellate zone, and possibly is slightly younger than the Stanwell Formation.

Stratigraphic relations

The Jim Crow beds are not exposed at the surface. Granitic basement was reached at 93m in the drillhole east of Mount Jim Crow, and this granite may be continuous with an outcrop near Bondoola. The basin is limited by serpentinite to the east and the mid-Permian Berserker Group to the south-west. The relationship between the Jim Crow beds and basalt mapped to the north-west and penetrated to a total depth of 44.2m by a drillhole just south of B.K. Creek is uncertain. This basalt has previously been regarded as Late Cretaceous in age (Kirkegaard & others, 1970; Willmott & others, 1986). However, it may be equivalent to the Lower Cretaceous Dalma Basalt and underlie the Jim Crow beds. Plugs of the Upper Cretaceous Mount Hedlow Trachyte must intrude the Jim Crow beds.

Gravity data indicate that the serpentinite is a thin sheet beneath which the basin may extend to the east.

Correlation with other units

The Jim Crow beds compare most closely in age, lithology and depositional setting with the Stanwell Formation west of Rockhampton.

Alton Downs Basalt (Kva)

(A.D.C. Robertson & C.G. Murray)

Introduction

Alton Downs Basalt is the name applied to the extensive sequence of basalt flows mapped as Late Cretaceous by Kirkegaard & others (1970) in the area between Westwood, Ridgelands, Rockhampton and Mount Morgan (Geological Survey of Queensland, 2002). The Late Cretaceous age was supported by a K-Ar date near the Cretaceous–Tertiary boundary (Wellman, 1978). However, basalt in the Stanwell–Westwood–Dalma area is overlain by the Stanwell Formation and Mount Salmon Volcanics, and has now been defined as the Lower Cretaceous Dalma Basalt. The Alton Downs and Dalma Basalts are identical in lithology, geochemistry, and stratigraphic relationships, and north of Stanwell are separated by only 2km of basement rocks. This raises the possibility that there is only a single basalt sequence, which would mean that the date reported by Wellman (1978) is either in error (it was regarded as a minimum age because the sample was slightly altered) or from a later basalt plug. Until this problem is resolved, the Alton Downs Basalt is described separately and assigned a Late Cretaceous age. Basalt flows surrounding the Mount Hedlow Trachyte plugs west of Yeppoon are also included in the Alton Downs Basalt, following the interpretation of Willmott & others (1986).

Distribution

The largest area of basalt is almost entirely covered by alluvium of the Fitzroy River, but its extent is clearly indicated by airborne magnetic survey data. It extends north-west from Rockhampton (where it was mapped by Ridgway, 1940) in a fault-bounded belt 35km long and 10km wide. The basalt associated with the Mount Hedlow Trachyte plugs forms a discontinuous north-west trending belt 20km long and 8km wide between Hedlow and Alligator Creeks. It is probable that these two sequences of basalt flows were originally continuous, and separated by subsequent faulting, erosion, and deposition of surficial alluvial sediments (Figure 236).

Derivation of Name

The name is derived from a land holding known as Alton Downs (GR023400 742100) adjacent to the Rockhampton–Ridgelands road. It was from this property that Wellman (1978) sampled the basalt for isotopic age determination.

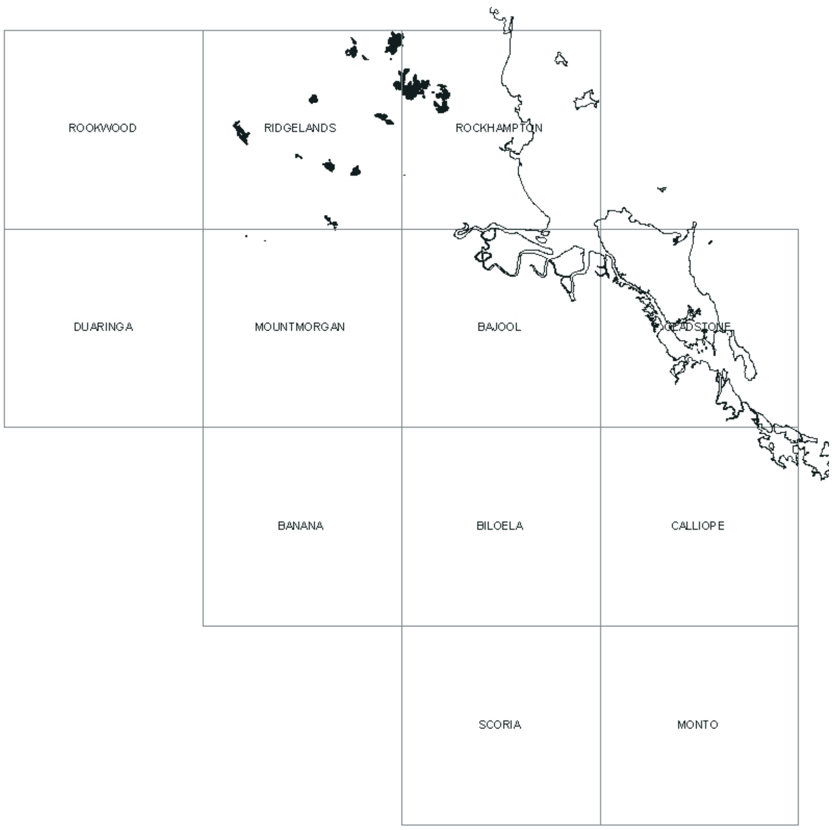


Figure 236. Distribution of the Alton Downs Basalt

Type area

The type area is 5km south of Alton Downs, where the basalt forms a low ridge trending north-east.

Topographic expression

The Alton Downs Basalt forms flat country or very low, undulating hills. Between Ridgelands and Rockhampton, much of the basalt is covered by alluvium associated with the Fitzroy River drainage system. Drilling results show that the thickness of alluvium along the river ranges from 15–35m (Mather, 1980). North of Rockhampton, the basalt forms the low-lying, swampy country of the Hedlow Creek plain, and a low plateau at Rossmoya covered by red lateritic soil. Outcrops of fresh basalt are scarce, and it occurs mainly as boulders in chocolate-brown soils.

Geophysical expression

The magnetic response of the Alton Downs Basalt is most obvious on images of the first vertical derivative, which exhibit the typical “mottled” effect produced by thin basaltic flows. The extent of the basalt beneath the alluvium of the Fitzroy River, and the existence of basement windows, is easily determined from this magnetic signature.

As expected, the basalt gives a very low radiometric response, and appears black on ternary coloured images.

Lithology

Kirkegaard & others (1970) described the Upper Cretaceous basalt as black, generally non-vesicular, and composed of labradorite, titanaugite, olivine and magnetite. Where present, phenocrysts are usually small. Sutherland & others (1996) listed samples from the Alton Downs Basalt as alkali basalt, hawaiite, and mugearite. Euhedral to subhedral olivine phenocrysts (commonly altered to iddingsite) are set in a finer grained groundmass composed of plagioclase (oligoclase to anorthite) laths, olivine (altering to iddingsite) and granular pyroxene and opaque minerals. In some samples, the groundmass feldspar exhibits flow alignment. Flow tops are locally vesiculated with some of the vesicles filled with a fibrous calcite. East (1945) recorded agglomerate from the outcrop east of Louisa Creek which might be part of the Mount Salmon Volcanics.

Geochemistry

Sutherland & others (1996) presented a single analysis from 5km west of Alton Downs, and three analyses from the area between Hedlow and Alligator Creeks. The rocks are basalts and one hawaiite (Figure 206). All samples have typical within-plate geochemistry (Figure 237), with $(Th/Nb)_N$ ranging from 1.88–3.89. On standard discriminant diagrams, the analyses plot in the within plate basalt field or its equivalent (Figure 238). On most of these diagrams, the sample from Alton Downs plots more closely with the Dalma Basalt (Figure 221).

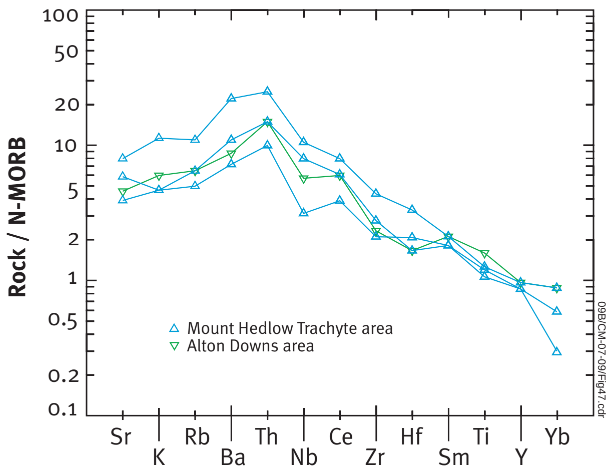


Figure 237. Spidergram plot of one basalt from the Alton Downs area, and three from around the Mount Hedlow Trachyte (Sutherland & others, 1996). N-MORB values are from Pearce (1983).

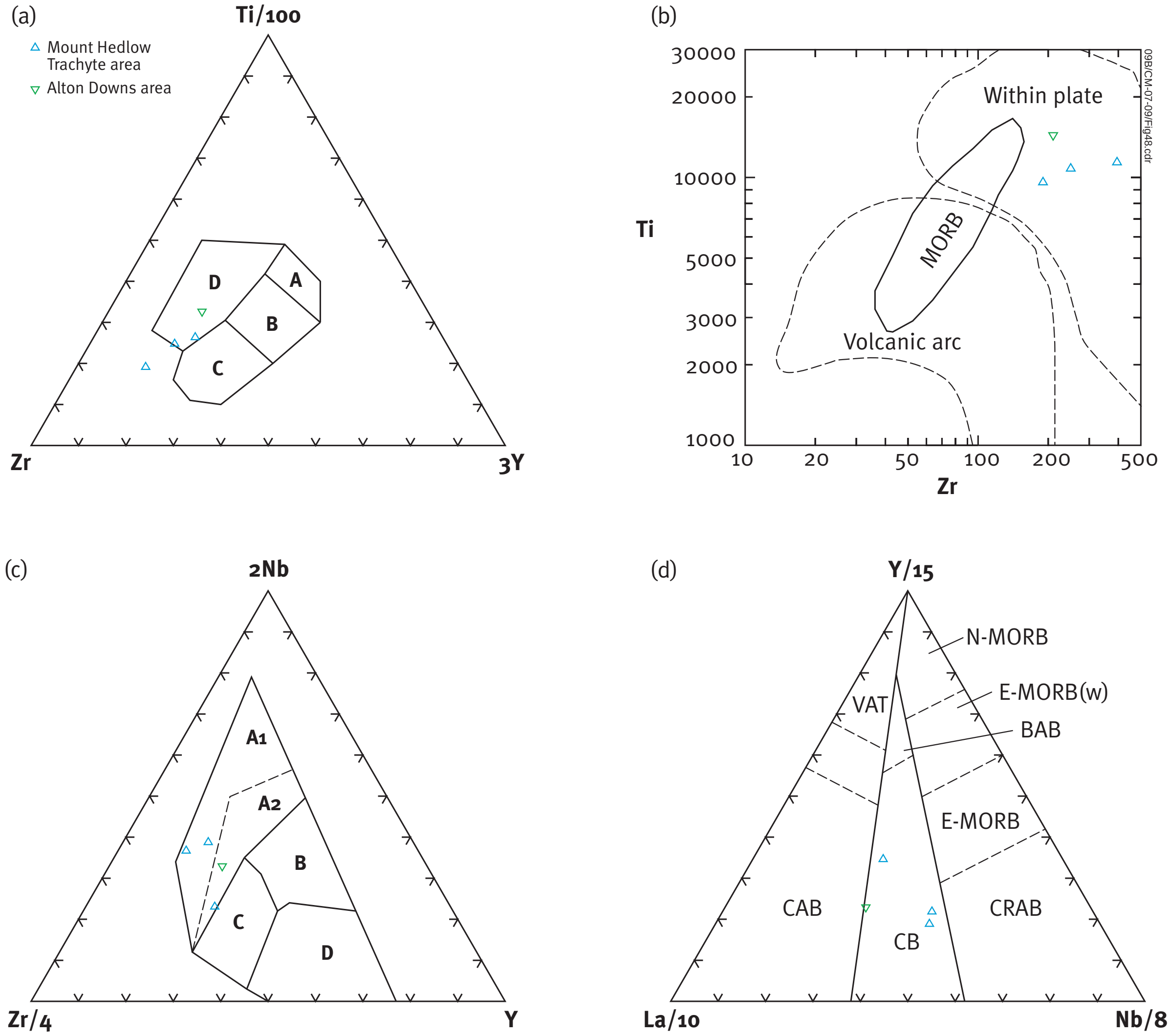


Figure 238. Analyses of basalts from the Alton Downs Basalt plotted on discriminant diagrams of (a) Pearce & Cann (1973): A = island arc tholeiites, B = island arc tholeiites, volcanic arc basalts and mid-ocean ridge basalts or MORB, C = calc-alkali basalts, D = within plate basalts; (b) Pearce (1982); (c) Meschede (1986): A1 = within plate alkali basalts, A2 = within plate alkali basalts and within plate tholeiites, C = within plate tholeiites and volcanic arc basalts; and (d) Cabanis & Lecolle (1989): CB = continental basalts, CRAB = alkali basalts from continental rifts, CAB = calc-alkali basalts. The analyses are essentially confined to the within plate basalt, within plate alkali basalt, within plate alkali basalt and tholeiite, and continental basalt fields on these diagrams.

Thickness

In the main outcrop area north-west of Rockhampton, the maximum thickness of Alton Downs Basalt is more than 200m. The MPY (Alton Downs) 1 petroleum exploration borehole bottomed in basalt at 213m; this is probably the bore referred to by Reid (1946). East (1945) reported about 60m of basalt in water bores through the alluvium along Louisa Creek 3km south-west of its junction with the Fitzroy River.

Structure

In a large part of the outcrop area, the Alton Downs Basalt is essentially flat lying. The belt extending north-west from Rockhampton is clearly fault-bounded on its eastern and southern boundaries. Mather (1980) suggested that it was deposited in a half graben that was downfaulted more on its eastern side, and this is supported by limited thickness data from bores.

Age

Wellman (1978) dated basalt, collected from Alton Downs, by the K-Ar method. The sample gave an age of 67.3±2.6Ma, which is considered to be a minimum age as the sample was slightly altered. However, the Alton Downs Basalt is separated by only 2km of basement from the identical Dalma Basalt, and its age may be Early Cretaceous.

Willmott & others (1986) report that plugs of the Mount Hedlow Trachyte intrude the basalt in the Jim Crow Basin. Samples from the Mount Hedlow Trachyte have given a K-Ar whole rock age of 71Ma (Harding, 1969), and U-Pb zircon ages of 73±3Ma to 80±3Ma (Sutherland & others, 1996). Assuming that the basaltic and trachytic magmatism was coeval, these results provide general support for the Late Cretaceous date obtained by Wellman (1978).

Stratigraphic relationships

In the Rockhampton region, the Alton Downs Basalt overlies the Lower Carboniferous Rockhampton Group, the Devonian-Carboniferous Doonside Formation, and elements of the Lower Permian Berserker Group. To the north-west of Rockhampton, it overlies the Rockhampton Group and the Devonian to Carboniferous Mount Alma Formation. In the Jim Crow Basin, the Mount Hedlow Trachyte intrudes the Alton Downs Basalt.

North of Stanwell, the Alton Downs Basalt is separated from the Dalma Basalt by only 2km of Mount Alma Formation. To east (Alton Downs Basalt) and west (Dalma Basalt) of this basement outcrop, the basalt flows appear identical, and there is no variation in their chemistry, and both sequences appear to overlie unnamed sediments.

Economic significance

No mineralisation of economic significance has been reported from the Alton Downs Basalt. Relatively thin sedimentary beds within and below the basalt flows are an important aquifer.

Unnamed Cretaceous rhyolite (Kr)

(C.G. Murray)

Introduction

Kirkegaard & others (1970) mapped rhyolitic rocks at White Rock about 25km north-west of Gladstone as an intrusive plug, but they were described as flows and breccias by Donchak & Holmes (1991).

Distribution

The rhyolitic rocks form an equant area about 1km across surrounding White Rock (Figure 239).

Topographic expression

White Rock is a prominent topographic feature rising to 252m.

Geophysical expression

The unit has a strong radiometric and magnetic signature. Both are considerably larger than the mapped outcrop of rhyolite. This suggests that detritus from the outcrop has been distributed around it, and that the rhyolite may be underlain by an intrusion.

Lithology

The unit consists of rhyolite flows and breccias (Donchak & Holmes, 1991).

Thickness

The thickness of the unit is unknown. Assuming it is flat-lying, the thickness is unlikely to exceed 250m.

Structure

The rhyolite is probably flat-lying and undeformed.

Stratigraphic relations

The unit lies unconformably on the Lower Carboniferous Wandilla Formation.

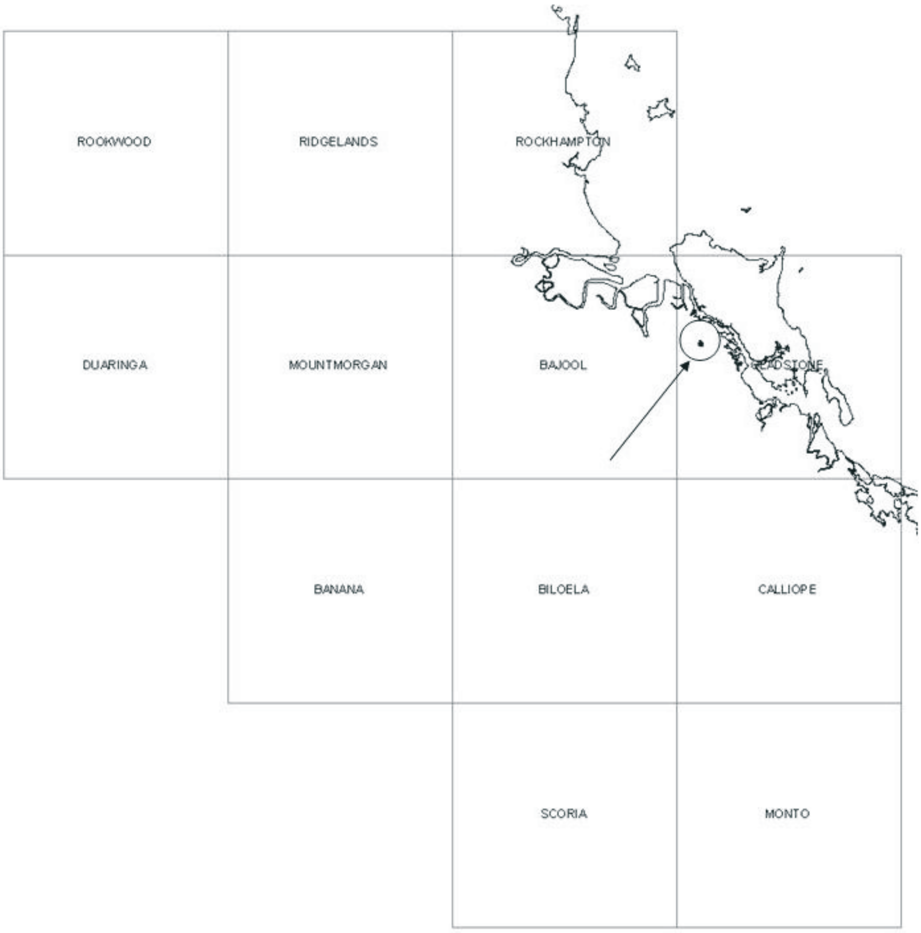


Figure 239. Distribution of the unnamed Cretaceous rhyolite (Kr)

Correlation with other units

The unit could be a correlative of Triassic or Cretaceous volcanics.

TERTIARY

The Narrows Group

(S.B.S. Crouch)

Introduction

The Tertiary sediments of The Narrows Graben were first called the Casuarina-Narrows Series by Dunstan (1913b), but were renamed The Narrows Beds by Kirkegaard & others (1970). The sediments were studied by Swarbrick (1974), Lindner & Dixon (1976) and Noon (1981).

Henstridge & Missen (1982) subdivided the sedimentary sequence of The Narrows Graben into three units — the upper Curlew Formation, the oil-shale bearing Rundle Formation, and the basal, informally named Worthington formation. The type section for the three formations is a composite section based on core from two boreholes: GSQ Rockhampton 2 in the south (GR313200 7363600) and Southern Pacific Petroleum's RD66 (GR299400 7382100) in the north. Using data from these holes as well as information from Esso Australia's ERD169 (GR301000 7381000), Henstridge & Coshell (1984) subdivided the Rundle Formation into eight members. Further studies were carried out by Finegan (1990).

Distribution

The Narrows Graben occurs beneath a north-north-west trending belt of coastal lowlands extending at least 30km north from Gladstone, bounded by the Rundle and Mount Larcom Ranges to the west and The Narrows

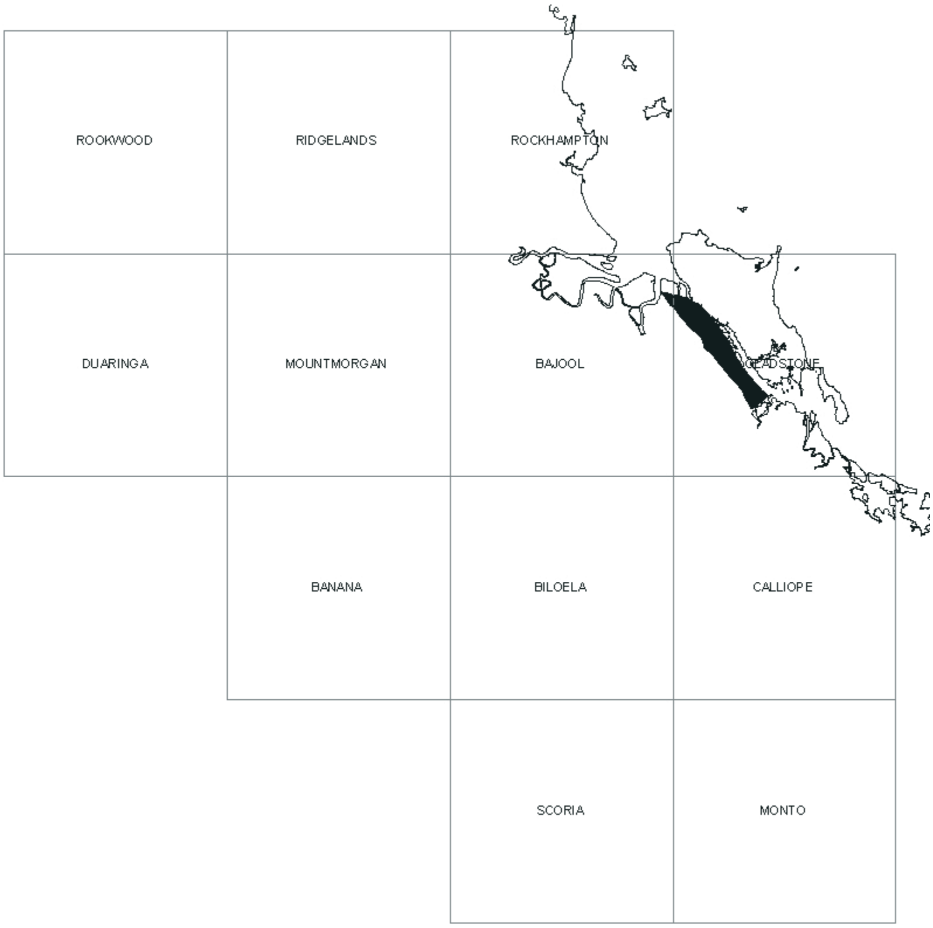


Figure 240. Location of The Narrows Graben

channel to the east. Two distinct depositional sub-basins are now recognised as containing significant oil shale deposits: Rundle in the north (between White Rock and Ramsay Crossing), and Stuart in the south (around Fishermans Landing) (Figure 240).

Topographic expression

The Narrows Group forms a very low area, partly below sea level.

Geophysical expression

The Narrows Graben falls partly outside the area covered by the semi-detailed airborne geophysical survey, but the sedimentary fill has a low radiometric and magnetic signature.

Lithology

Surface outcrops of the Tertiary sediments are mainly heavily lateritised and deeply weathered with a thick soil profile. Recognisable rocks can only be seen in some creek beds and intertidal zones. The stratigraphic sequence is known almost exclusively from borehole data and exploration trial cuts, which indicate the presence of three units — the Worthington formation (**Tw**) at the base, overlain by the Rundle Formation (**Tr**) and the Curlew Formation (**Tc**).

The Worthington formation consists of conglomerate and sandstone at the base grading upwards into sandy and silty claystone and siltstone, with massive claystone at the top.

The Rundle Formation is the major oil shale unit containing at least eight oil shale seams (locally pyrite bearing), interbedded with mudstone and minor sandstone, dolomite and lignite. Fossil ostracods, gastropods, turtle shells and fish and reptile fragments and coprolites are abundant in parts of the formation.

The overlying Curlew Formation occurs mainly in the northern part of the graben and consists of mudstone, carbonaceous siltstone, and minor calcareous sandstone and limestone.

Depositional environment

The basal coarse clastics of The Narrows Graben represent alluvial fans localised along fault scarps marginal to the developing basin. Progressively finer-grained sediments were deposited as the graben developed into a stable, shallow, freshwater, mainly lacustrine depression into which most of the Rundle and Curlew Formations were deposited. Swimming and bottom-dwelling flora and fauna flourished in this environment, and much algal-derived organic matter accumulated to form oil-shale.

Coshell & McIver (1989a,b) highlighted the cyclic nature of the sedimentation in the Rundle Formation, reflecting fluctuations in areal extent and relative dominance of coexisting subenvironments (shallow freshwater lacustrine, limnic marsh, variably anoxic limnic mudflat, pedogenic). The transition to the Curlew Formation marks a gradual filling of the basin, with fluvial conditions becoming increasingly dominant towards the top of the formation.

Thickness

The total thickness of the three units of The Narrows Group is in excess of 1000m.

Structure

The basin has an asymmetric half-graben morphology, with a complex series of normal faults marking the downthrown western margin. Most of the units are thick-bedded, and dip shallowly to the west. Anomalous steep dips are associated with local faulting, especially along the western margin of the graben. An alkali dolerite intrusion, recorded from drillhole data, has thermally metamorphosed some of the oil shale seams, destroying much of the kerogen content in the north central part of the Stuart deposit.

Biostratigraphy and age

Ostracode and microfloral assemblages indicate an early Tertiary, probably Eocene age for the Rundle and Curlew Formations. The microfossils have been described by Hekel (1972), Palmieri (1975), Foster (1979), and Foster & Harris (1981). Intrusive dolerite has been dated at 26.8Ma (Henstridge & Missen, 1981).

Stratigraphic relationships

The Tertiary basin is bounded by Palaeozoic rocks of the Wandilla Formation to the south and east, and the Wandilla and Doonside Formations to the west.

Economic significance

The oil shales are being evaluated for extraction as a fuel source.

Casuarina beds (Tc)

(S.B.S. Crouch)

Introduction

Dunstan (1913b) considered that the sedimentary sequence of the Casuarina Basin was similar to that in The Narrows Graben, and regarded the Casuarina-Narrows Series as continuous. Subsequently, a thick section of Tertiary sediments was postulated beneath the lower reaches of the Fitzroy River to account for a prominent negative gravity anomaly (Kirkegaard & others, 1970). In 1978, the Geological Survey of Queensland drilled a deep core hole (GSQ Rockhampton 1) to investigate the stratigraphic sequence of the basin (Noon, 1980). The informal name Casuarina beds was applied as there is insufficient information to allow formal definition of the unit. Further gravity work was carried out to define the basin more accurately (Huber, 1981).

Distribution

The Casuarina Basin is centred 30km south-east of Rockhampton, and the main depocentre defined from airborne magnetic data is 33 by 17km, oriented in a north-west to south-east direction (Figure 241). It seems that the shallow margins of the basin extend outside the main depocentre (Willmott & others, 1986). Near the Botanic Gardens in Rockhampton, the sandstone reported by Rands (1895) and Ridgway (1940) to overlie the

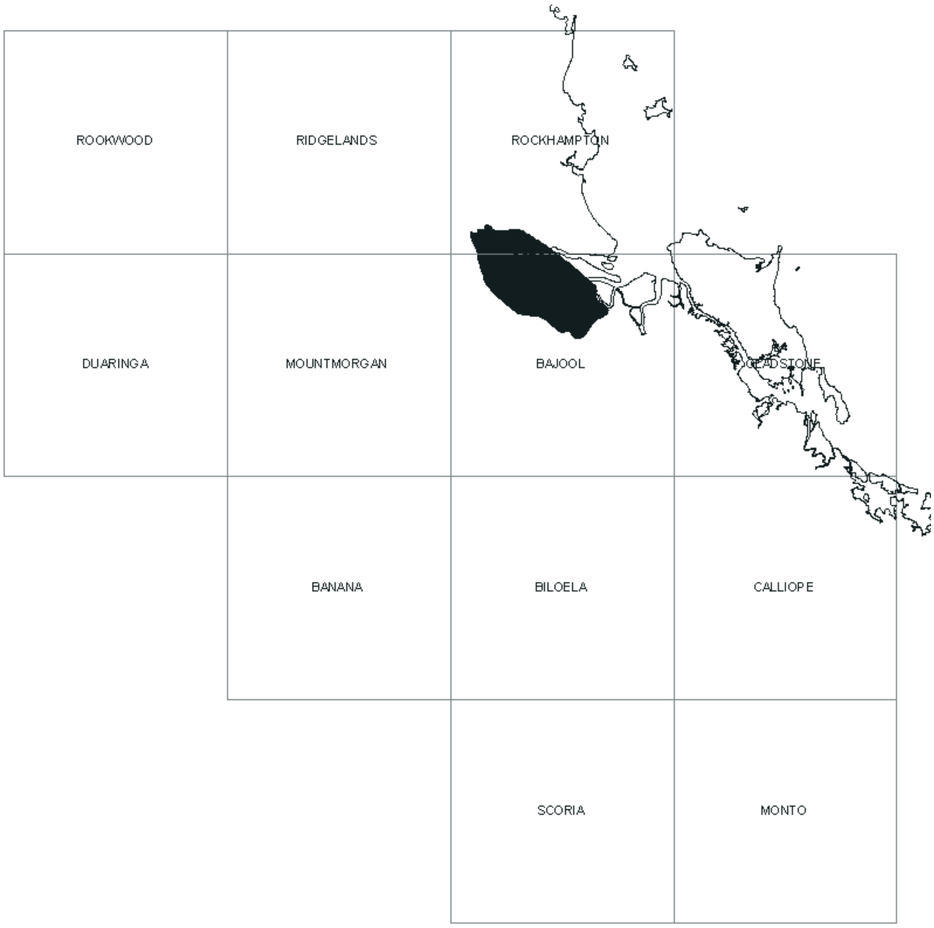


Figure 241. Location of the Casuarina Basin

Rockhampton Group probably represents basal sediments of the Casuarina beds. Bleached mudstone with worm burrows was encountered during the present mapping on the road between Nankin and Thompsons Point, where it apparently overlies Chalmers Formation of the Berserker Group.

Topographic expression

The Casuarina beds form a low area at the mouth of the Fitzroy River, and are covered by recent sediments.

Geophysical expression

The cover of recent sediments conceals the Casuarina beds, but the magnetic response is low, and probably the radiometric response as well.

Lithology

The sequence is best known from GSQ Rockhampton 1 (Noon, 1980), although Dunstan (1920) had earlier reported clay, sand, gravel and peat from holes that tested shallow level brines. The 561.5m of sediments of the Casuarina beds are predominantly mudstone, siltstone and shale. The base is marked by fining upwards sequences containing conglomerate and sandstone, and these are succeeded by multi-coloured mudstone and minor silty mudstone and oil shale. The next 123m are characterised by the occurrence of oil shale, and this is overlain by mudstone and siltstone with thin dolomite beds. The succeeding 125m of strata consist of interbedded siltstone, shale and mudstone with a high proportion of oil shale and a 1m-thick bed of lignite. The top of the Casuarina beds comprises interbedded siltstone and mudstone. Fossils include ostracodes, gastropods, fish scales and spines and plant material.

Depositional environment

Evidence from palaeontology, mineralogy and sedimentary structures indicates that the Casuarina beds were deposited in a freshwater lacustrine environment. Deposition in quiet water is supported by the thin bedding, the lack of current-related structures, and the overall fine grainsize. Reducing conditions are suggested by the preservation of organic material and by common fine-grained pyrite (Noon, 1980).

Thickness

GSQ Rockhampton 1 penetrated 561m of Tertiary sediments beneath 44m of alluvium before reaching basement. The detailed gravity study estimates 700m of Tertiary and Quaternary sediments to the south-east of Nerimbera. Under Casuarina Island, near the centre of the gravity low to the south, a thickness of at least 900m is indicated. A greater depth to basement of 1500–1600m is suggested by modelling and analysis of the airborne magnetic data (Appendix 5), but this may be too deep if the basement rocks are weakly magnetic.

Structure

The Tertiary sediments are flat-lying and undeformed. Gravity data give an indication of the shape of the basin in which they were deposited. They suggest that the basin is bilobate, with the deepest section to the south and a shallower section to the north-west.

Noon (1980) speculated that the basin was formed by damming of the ancestral Fitzroy River due to movement on the east-north-east trending Fitzroy River Fault (Kirkegaard & others, 1970), but the postulated fault appears to be too far north to justify this explanation.

Biostratigraphy and age

Examination of palynological samples is difficult, but suggests an Eocene age, and Noon (1980) tentatively correlated the lower part of the Casuarina beds with The Narrows beds. Dettmann & Clifford (2000) assigned a late Eocene to early Oligocene age to a palynoflora from 439.75m depth in GSQ Rockhampton 1.

Stratigraphic relations

The Casuarina beds are interpreted to unconformably overlie the surrounding units, which include the uppermost Silurian to Middle Devonian Erebus beds, the Upper Devonian to Lower Carboniferous Mount Alma Formation, the Lower Carboniferous Rockhampton Group, and the mid-Permian Berserker Group. The only contacts observed are to the north-east of the Casuarina Basin, where mudstone interpreted to be part of the Casuarina beds overlies the Berserker Group. GSQ Rockhampton 1 penetrated basement beneath 605m Casuarina beds, but the correlation of these rocks with exposed units in the vicinity is uncertain (Noon, 1980).

Economic significance

Oil shale is present in the intervals 100–225m and 295–425m (Noon, 1980). The upper part of the unit contains brines which have been a source of salt for many years (Laycock, 1980).

Yaamba beds (Ty)

(P.R. Blake & C.G. Murray)

Introduction

Exploration for oil shale over a negative gravity anomaly near Yaamba revealed a Tertiary oil shale basin (Olive & others, 1984). The rocks within the Yaamba Basin were named the Yaamba Group by Willmott & others (1986) in anticipation of the definition of constituent formations, but are here named the Yaamba beds because no units have been defined.

Distribution

The Yaamba beds are known only from the Yaamba Basin. This basin is centred just west of the small settlement of Yaamba on the Bruce Highway about 35km north-west of Rockhampton. Foster (1982) showed the basin as up to 15km long and 7km wide, and Noon (1984b) reported that it was about 12km long and up to 6km wide and covers an area of approximately 54km^2. However, airborne magnetic data suggest that it is more equant, with an approximate length of 9km and a width of 8km (Figure 242).

Topographic expression

The basin is situated beneath alluvium of the Fitzroy River and Alligator Creek, and forms low, flat topography.

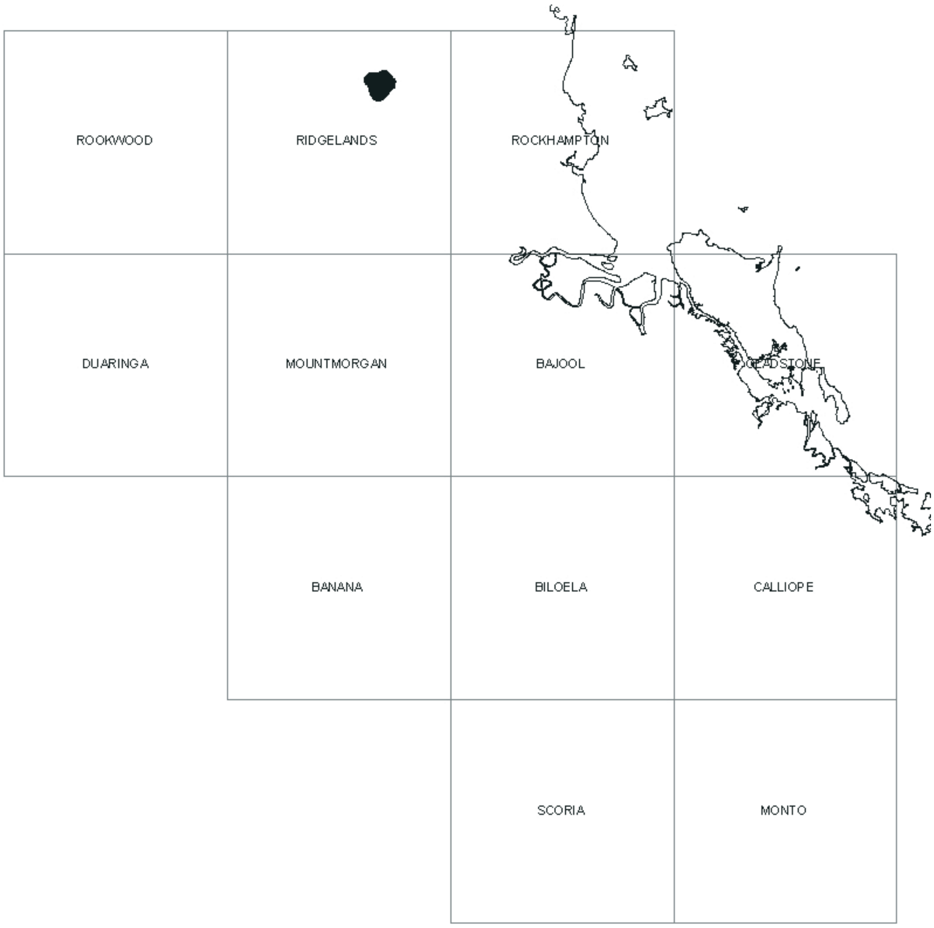


Figure 242. Location of the Yaamba Basin

Geophysical expression

The Yaamba beds have a very low magnetic response, and the basin is easily distinguished on images of airborne data. Their radiometric signature is unknown, as they are completely covered by alluvium of the Fitzroy River and Alligator Creek. The Yaamba Basin coincides with a strong negative gravity anomaly (Kirkegaard & others, 1970).

Lithology

The sedimentary succession in the Yaamba Basin is divided into five informal subunits, from A at the base to E at the top. Unit A consists of brown oil shale and green shale and mudstone. Units B to E are grouped as the upper sequence of carbonaceous and lignitic oil shale, black carbonaceous mudstone and shale, and green mudstone and sandstone. Oil shale predominantly occurs in the lower half of the sequence. The upper units towards the south-south-west consist of quartz sandstone and sandy mudstone, with some feldspathic sandstone possibly derived from granite. To the north-north-east is a thick wedge of basaltic conglomerate, pebbly mudstone and lithic sandstone (Dudgeon, 1985).

Depositional environment

In the central part of the Yaamba Basin, deposition took place in a quiet, freshwater environment like a lake or swamp (Foster, 1982). Non-marine dinoflagellates occur in all samples examined, and microspore massulae of the water fern *Azolla* are present in some samples. The typical fine grained rocks grade outwards into coarser marginal facies sedimentary rocks which were probably deltaic as well as fluvial in origin (Noon, 1984b; Dudgeon, 1985).

Thickness

The Yaamba beds are estimated to be 900m thick, with the upper subunits B to E about 500m (Foster, 1982). The average vertical thickness of oil shale is 162m (Olive & others, 1984). The central, deepest part of the basin has been penetrated to a depth of 750m without reaching basement. Faulting along the northern and eastern edges of the basin may be partly responsible for the comparatively thick sequence there (Dudgeon, 1983). Modelling of airborne magnetic data suggests a depth to magnetic basement of up to 1300m (Appendix 5).

Structure

The Yaamba Basin is a broad synclinal structure plunging north-west. The north and north-eastern margins appear to be faulted, and the sequence is cut by two north-west trending zones of small normal faults (Olive & others, 1984).

Biostratigraphy and age

Foster (1982) compared microfloras from the Yaamba beds with assemblages from the Gippsland Basin and concluded that they were middle to late Eocene in age. A similar result was obtained by Dudgeon (1983, 1985) assuming that correlations with the Gippsland Basin were valid.

Stratigraphic relations

The Yaamba beds overlie flat-lying basalt which may be Early or Late Cretaceous in age (see discussion of the Alton Downs Basalt). They are overlain by about 40m of Quaternary alluvium deposited by the Fitzroy River (Olive & others, 1984).

Correlation

The Yaamba beds are similar in geological setting, age and lithology to other Tertiary oil shale sequences in central Queensland.

Economic significance

Olive & others (1984) indicated a resource of more than 2800 million barrels of oil in seams at least 3m thick with an average oil content of 95 litres per tonne (dry). However, it would be necessary to divert the Fitzroy River before the oil shale could be extracted.

Nagoorin beds (Tn)

(P.R. Blake & C.G. Murray)

Introduction

The Nagoorin beds were named by Dear & others (1971). They are best known from exploration drilling for oil shale (Ivanac, 1983; Henstridge & Hutton, 1986; Patterson & others, 1987b; Patterson & Lindner, 1991).

Distribution

The Nagoorin beds occur in the Nagoorin Graben, a narrow north-south trending down-faulted basin 30km long and up to 4km wide along the Boyne River valley south-south-west of Gladstone (Figure 243). The beds are covered by Quaternary alluvial deposits. Outcrops have been recorded in Coal Creek (Rands, 1885), Ubobo Creek (Ivanac, 1983), and a tributary of Ubobo Creek (Dear & others, 1971).

Topographic expression

The Nagoorin beds form a north-trending topographic depression along the modern floodplain of the Boyne River, flanked to the east by the southern part of the Many Peaks Range and to the west by a discontinuous line of hills and mountains, of which Mount Robert is the highest at 1000m.

Geophysical expression

The radiometric signature is masked by the overlying unconsolidated alluvium, and the beds have a very low magnetic response.

An intense gravity low centred just to the east of the Nagoorin Graben may be in part related to the thick graben fill, but the anomaly is circular, and its origin is uncertain (Murray, 1994).

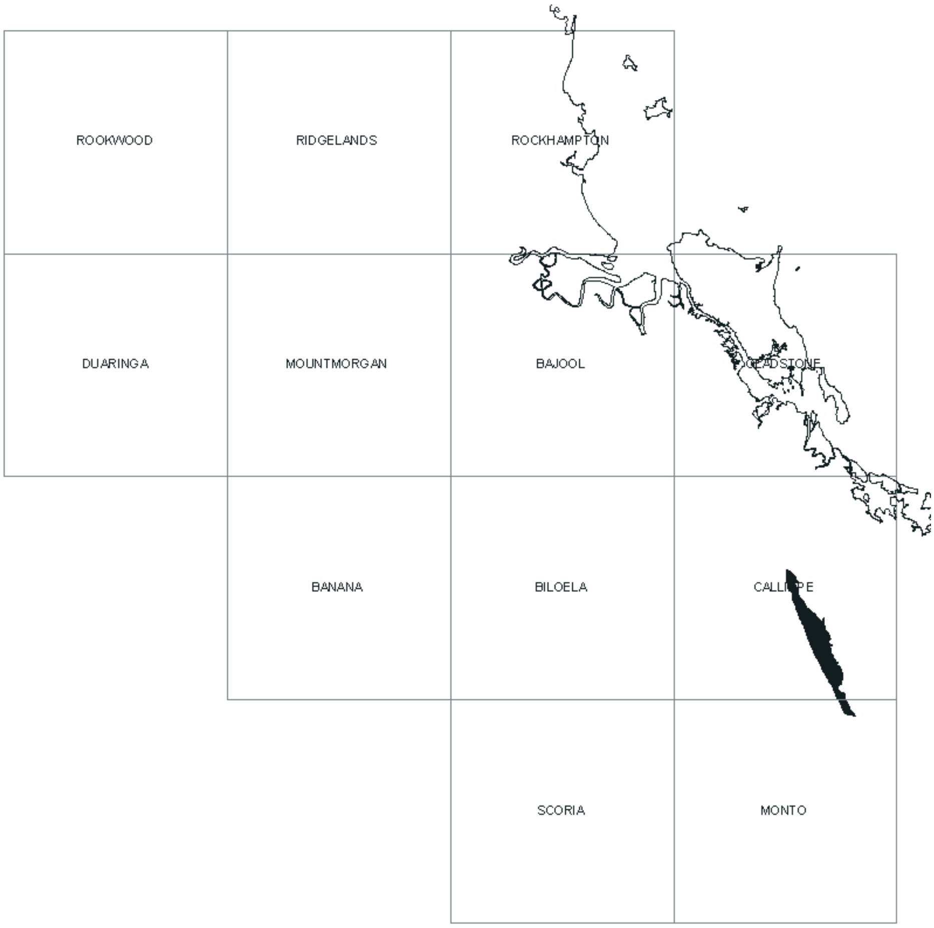


Figure 243. Location of the Nagoorin Graben

Lithology

The Nagoorin beds consist of sandstone with minor conglomerate and lignitic oil shale at the base overlain by a thick sequence of carbonaceous oil shale with intervals of green mudstone, siltstone and fine-grained sandstone. Some rock types contain abundant carbonate (Ivanac, 1983). The beds were divided by Henstridge & Hutton (1986) into five conformable sedimentary units. All except the basal unit contain significant oil shale deposits.

Environment of deposition

The depositional environment was dominantly lacustrine, with local fluviatile influence particularly towards the base of the sequence and around the margins of the basin.

Thickness

A drillhole 687m deep did not reach the base of the sequence. The minimum stratigraphic thickness is 870m (Ivanac, 1983; Henstridge & Hutton, 1986). Analysis of waveforms in the airborne magnetic data by means of a Fourier transform gives a depth of 600–650m near the northern end of the basin (Appendix 5).

Structure

The Nagoorin Graben was produced by early Tertiary crustal extension which has been related to the opening of the Coral Sea. Attempts at modelling have shown that the basin could not have been produced by pure strike-slip movements with a north-north-west orientation, and a component of east-west extension must have been involved.

The beds strike north-north-west, sub-parallel to the faulted basin margins, and dip at 15° to the west (Ivanac, 1983). Both strike and cross faults are present.

Biostratigraphy and age

By analogy with other oil shale sequences, the Nagoorin beds are assigned a mid- to late Eocene age.

Stratigraphic relationships

Contacts of the Nagoorin beds with older sequences are mainly faulted, but they overlap basement of Upper Silurian-Lower Devonian Calliope beds, Upper Devonian – Lower Carboniferous Mount Alma Formation, and Lower Carboniferous Wandilla Formation to the east. To the west they are faulted against the Calliope beds, the Mount Alma Formation, the Lower Carboniferous Rockhampton Group, and the Triassic Littlemore Granodiorite.

The Nagoorin beds are cut by Tertiary dolerite intrusions, and overlain by Quaternary alluvium up to 18m thick.

Economic geology

The oil shale resources of the Nagoorin beds are grouped into two deposits, Nagoorin and Nagoorin South, which together contain about 3 billion barrels of shale oil (Ivanac, 1983). The oil shale includes significant resources of both coal and coal seam gas.

Biloela Formation (To)

(I.W. Withnall)

Introduction

The Biloela Basin (Dickins & Malone, 1973) is one of the largest Tertiary basins in central Queensland. The sediments of the Biloela Basin were named Biloela Beds by Kirkegaard & others (1966). However, later reports mapped them as unnamed Tertiary sediments (Kirkegaard & others, 1970; Dear & others, 1971). Grimes (1980) revived the name Biloela Beds, and Noon (1982) defined them as the Biloela Formation (To).

Distribution

The Biloela Basin is about 125km long and up to 35km wide, extending from south-east of Biloela near Thangool to north of the Capricorn Highway (Figure 244).

Topographic expression

The Biloela Formation forms low topography along the valley of Callide Creek, a tributary of the Fitzroy River. The unit is largely exposed in the scarps of mesas which are topped by basalt flows, and as surrounding outwash deposits.

Geophysical expression

The radiometric response is low, and appears as a deep green colour on images from airborne data indicating relative enrichment in thorium, consistent with deep lateritic weathering. The Biloela Formation is essentially non-magnetic, and magnetic features reflect the underlying basement.

Lithology

Reid (1940) described the unit from water bores, because of the paucity of surface exposures. He found that the sedimentary sequence consists of 90–95% clay and 5–10% sandstone.

The sediments were only examined at a few localities in scarps of duricrust mesas in the northern part of the basin near Wowan and north-west of Gogango, where they are deeply weathered and capped by thin ferricrete. Kirkegaard & others (1970) reported silty and sandy claystone, and pale argillaceous quartz sandstone, locally pebbly to conglomeratic, from these localities. Grimes (1988) described several weathered exposures of claystone, siltstone, pebbly sandstone and conglomerate in quarries and road cuttings.

The type section of the Biloela Formation is in GSQ Monto 5 from 27.2m to 374m. The borehole is located near the eastern shore of Lake Victoria at GR224300 73392200. The section, which is described in detail by Noon (1982), comprises 335m of freshwater, lacustrine mudstone, siltstone, oil shale and sandstone, and minor lignite, carbonaceous mudstone and limestone. The bottom of the unit is shale that overlies basalt at

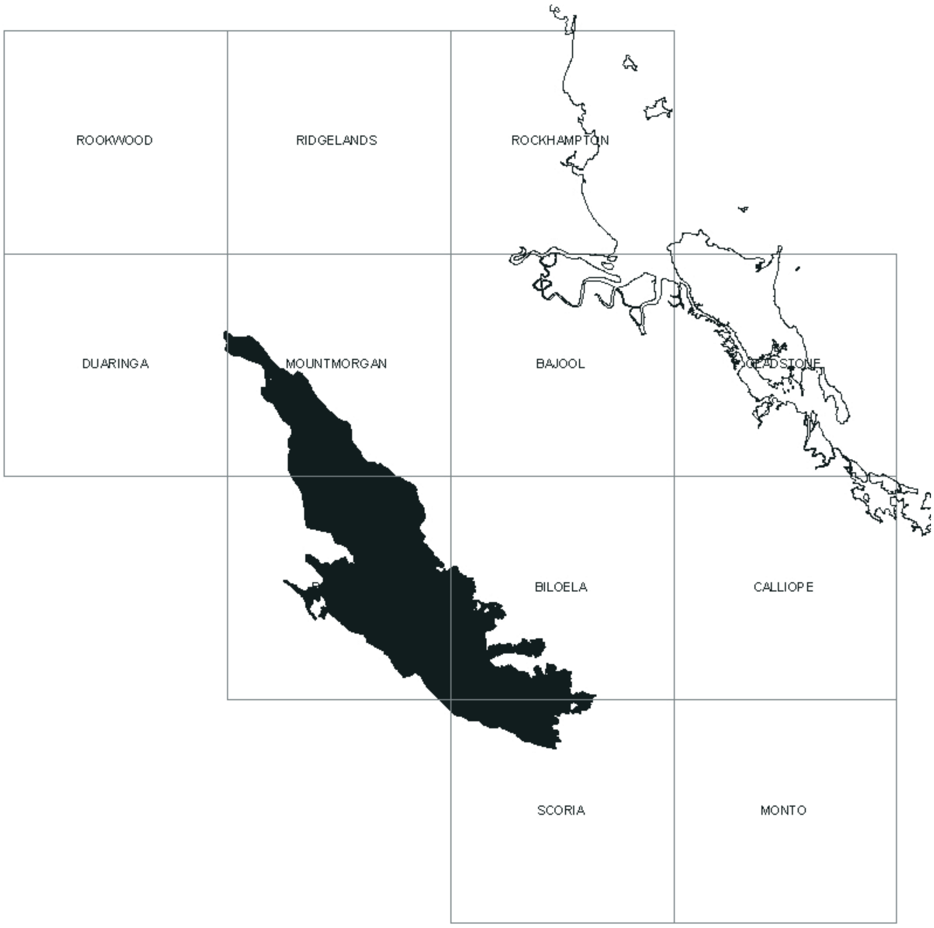


Figure 244. Location of the Biloela Basin

374m. The drillhole bottomed in the basalt at 390m and it is not known whether any older Tertiary sediments underlie it. The top of the section underlies poorly consolidated silty sand. A sill of basalt 11.7m thick occurs at 357.3m. The formation thins to the north and south and thickens to the west.

Depositional environment

The Biloela Formation was deposited mainly in a freshwater lacustrine environment, but probably also contained some fluvial sediments towards the basin margins. Noon (1982) suggested that initial ponding may have been in response to damming by the basalts that have been encountered by the drilling. Continued subsidence in response to tectonism accompanied the volcanism and allowed the development of a lake system. Bioturbation in the basal section, and presence of the freshwater gastropod *Planorbis* and ostracodes in the oil shale attest to shallow water conditions throughout the life of the lake. Carbonaceous and coaly sections indicate the development of localised swamps. Groundwater investigations (Pearce, 1971) indicate that the rocks are coarser grained in the southern part of the basin, suggesting that the rivers that fed the lake system flowed from the south.

Thickness

The thickness in GSQ Monto 5 is 335m (Noon, 1982). The basin is bilobate, and water bore information indicates that the deepest lobe is to the north (Pearce, 1971). A saddle just north of Jambin separates the lobes.

Structure

The basin is probably a half-graben like many of the other Tertiary basins, at least in its central part. Images of the airborne magnetic data show a well-defined western edge, where the magnetic response of the basement rocks is sharply truncated. However, the eastern margin is less well defined and the magnetic responses of the basement rocks fade gradually to the west. The western edge is also easily defined on aerial photographs, whereas the eastern margin is not.

Biostratigraphy and age

The Biloela Formation has not been studied palynologically, but is generally regarded as early Tertiary. The ostracode fauna is similar to that studied in The Narrows beds and assigned an early Tertiary age by Palmieri (1975). An upper limit is provided by a K-Ar age of 25±4Ma (mid-Oligocene) on basalt from the Dawes Range south-east of Biloela.

Stratigraphic relationships

Contacts of the Biloela Formation with older units are not exposed, but the unit is assumed to overlie adjacent Palaeozoic and Mesozoic rocks unconformably. Tertiary basalt, Tertiary to Quaternary outwash deposits, and Quaternary sediments overlie the Biloela Formation.

Economic significance

Several companies have explored the Biloela Basin for oil shale. EPMs 2634, 2635, and 2636 were explored by CRA Exploration Pty Ltd using core drilling to test the potential of the basin. Eight holes were drilled for a total depth of 1 300m. Assays from samples collected ranged from 2 to 180 litres oil per dry tonne. The best intersection was 3.2m of oil shale yielding 104.2 litres per dry tonne in drill hole Jambin 3C. The company considered that the potential for economic deposits was limited.

BP Australia Ltd explored the Biloela Basin in the 1980s under EPMs 2664 and 2665 (1980). Twenty-five non-cored holes and six partly-cored holes were drilled on a 10 by 5km grid. Some kerogenous seams were intersected in several holes but none of these intersections was considered economic.

Australian Consolidated Minerals Limited explored the southern part of the basin during 1981 under EPM 2918. They drilled two holes in the area (Thangool 1 and Thangool 2). The maximum oil yield for these holes was 8 litres per tonne, and it was concluded that the oil shale horizons were either too thin or contained only low concentrations of kerogen.

Also in 1981, Clifford Minerals under EPM 3098 analysed water bore records, previous exploration drilling records, and core from GSQ Monto 5. Rocks from GSQ Monto 5 were thought to be similar to sedimentary rocks in the tenement area and were tested for oil yields. These tests returned results of up to 165 litres per tonne and many samples yielded greater than 70 litres per tonne. Water bore records, however, were less promising and indicated that the Biloela Basin shallowed markedly to the south of Biloela. The tenement was deemed to have little potential for a major oil shale deposit.

Overall the oil shales are too thin and at depths too great to be developed by opencut methods. The sandstones contain sub-artesian groundwater in the Callide valley.

Rossmoya beds (Tr)

(C.G. Murray)

Introduction

The Tertiary sedimentary rocks in the Rossmoya Basin have been named the Rossmoya beds (**Tr**) since there is insufficient information for a formal definition (Willmott & others, 1986). They were discovered during exploration for oil shale.

Distribution

The Rossmoya Basin is centred 35km north of Rockhampton (Figure 245). Airborne magnetic data suggest that it has a rhomboid shape, about 7km by 5km. From drilling results, the basin was interpreted as elongate in a north-west direction, about 15km long and up to 4km wide, tapering to the south-east (Dear, 1980; Anonymous, 1993).

Topographic expression

Most of the basin lies beneath the alluvium of Hedlow Creek, and occupies a topographically low area. Lateritised sediments form low plateaux along the south-western edge of the basin.

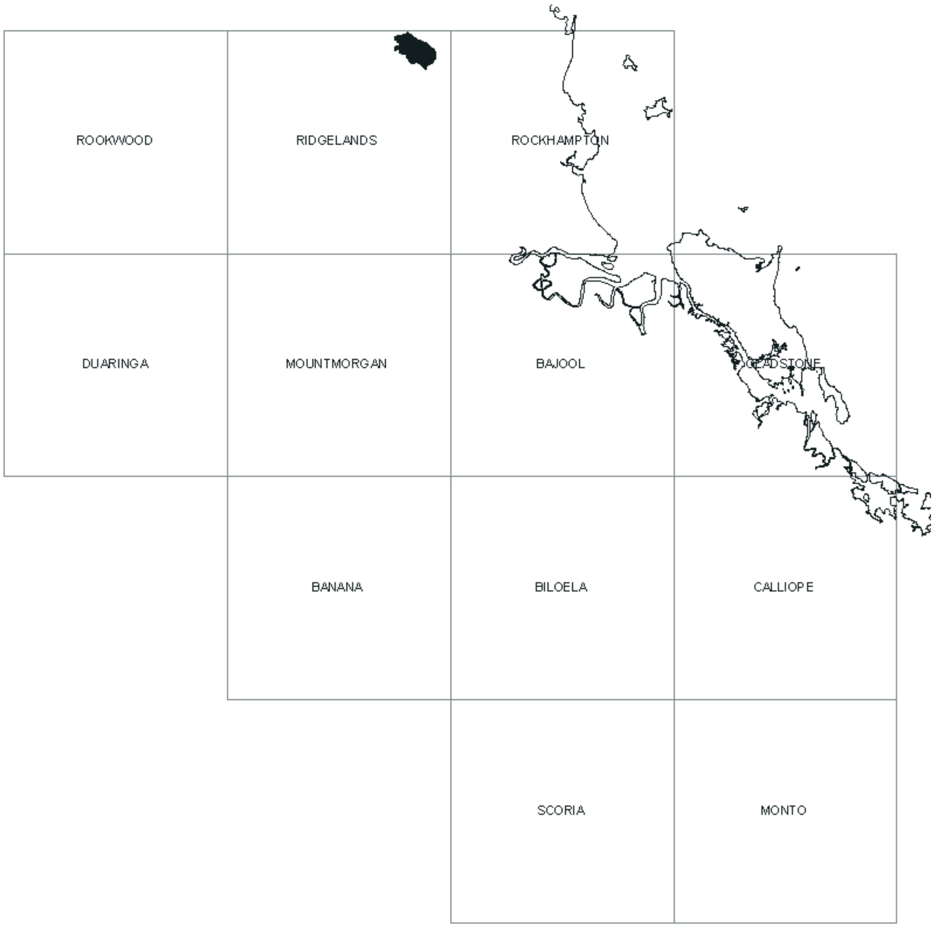


Figure 245. Location of the Rossmoya Basin

Geophysical expression

The radiometric expression of the Rossmoya beds is largely masked by alluvial cover, but appears to be low. The magnetic signature is also low, and the basin is marked by a “fuzzy” signal compared to the strong response from the surrounding basement rocks. The basin corresponds to a gravity low which is only partly separated from a more intense low over the Yaamba Basin.

Lithology

Drilling has shown that the Rossmoya beds consist of pale grey to pale greenish grey mudstone, sandy mudstone, medium and fine grained lithic quartz sandstone, and interbedded carbonaceous sequences of black lignitic oil shale and mudstone (Anonymous, 1993). No thick or high grade oil shale is present.

Depositional environment

The sedimentary rocks were deposited in a lacustrine environment.

Thickness

Gravity and magnetic data indicate maximum depths of about 200–300m and 300–400m for the Rossmoya Basin, respectively (see Appendix 5). These figures are consistent with drilling results. The deepest hole, which did not reach basement, is 210m, but a more typical thickness for the Tertiary sequence is 160m (Anonymous, 1993).

Structure

Examination of magnetic data suggests that the Rossmoya Basin is bounded by north-west trending faults to the north-east and south-west. These may have been strike-slip faults that controlled the basin formation. The basin fill is assumed to be little disturbed.

Age

The similarity with other Tertiary basins in the region suggests an early Tertiary (Eocene) age.

Stratigraphic relations

The Rossmoya beds overlie the Cretaceous Alton Downs Basalt, which crops out as strongly lateritised low plateaux along the western edge of the basin. They are overlain by alluvium of Hedlow Creek.

Correlation with other units

The Rossmoya Basin is similar to other Tertiary basins in the coastal area of central Queensland.

Economic significance

Drilling has failed to find any thick or high grade oil shale in the Rossmoya beds.

Unnamed Tertiary sediments west of Gladstone (Ts)

(C.G. Murray)

Introduction

Company drilling in the Larcom Creek valley west of the peak of Mount Larcom and near the Calliope River intersected Tertiary sediments beneath Quaternary cover (Layton & Associates, 1981).

Distribution

Tertiary strata are known from three separate basins which appear to be related to the same fault system. The northern basin is crossed by Munduran and Major Creeks, and is interpreted to be wedge-shaped broadening to the south, and to be 2.5km long and 1km wide. The central basin is similar, extending along Larcom Creek from Sneaker Gully to Cherry Creek, about 3.5km long and 1.5km wide. The southern basin straddles the Calliope River, and could be as much as 13km long and 2km wide (Figure 246).

Topographic expression

All the basins are in areas of alluvium along stream valleys, and therefore form topographically low country.

Geophysical expression

The Tertiary sediments have a negligible magnetic signature, but the small size of the basins means that responses from basement rocks may be dominant in airborne data. A basalt plug is interpreted to intrude the sequence in the southern basin.

The southern basin coincides with a significant negative gravity anomaly.

Lithology

The Tertiary sediments are not known to crop out, and drillholes provide the only information. The dominant lithology is described as multi-coloured clay, which either grades downwards into shale, or is interbedded with mudstone and siltstone and minor sand and gravel units (Layton & Associates, 1981). Basalt was reported from one drillhole in the northern basin.

Thickness

The maximum thickness penetrated was 54m in the northern basin (to basement), 211m in the central basin (basement not reached), and 375m in the southern basin (basement not reached) (Layton & Associates, 1981). Abrupt thickness variations were noted within the basins.

Structure

No structural disturbance was reported within the basin fill, and it is assumed to be close to horizontal. The basins are considered to have formed during the early Tertiary basin-forming episode, and therefore it is expected that they would deepen to the west. This appears to be the case for the two northern basins, which

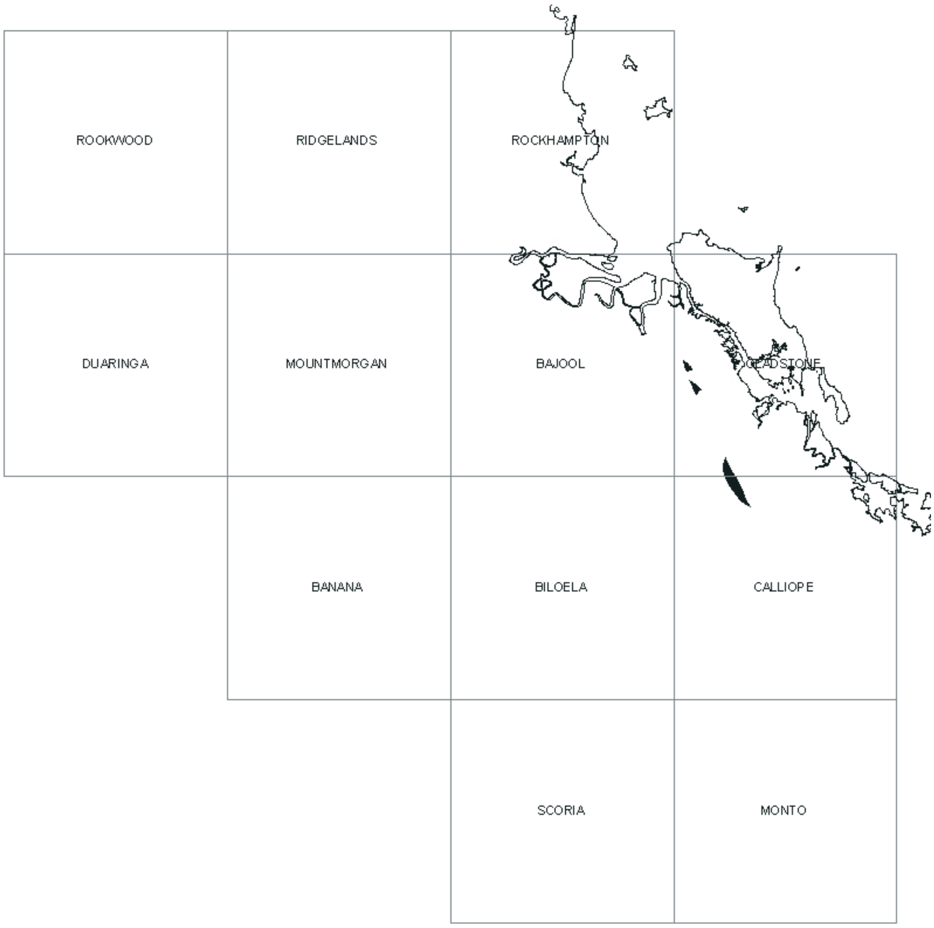


Figure 246. Location of the unnamed Tertiary basins

are located to the east of the controlling fault, but not for the southern basin, which is west of the interpreted southward extension of this fault.

Age

The age of these sediments has not been determined. They are assumed to be Tertiary because of their similarity to other sequences of this age.

Stratigraphic relationships

The Tertiary sequence overlies the Lower Carboniferous Rockhampton Group and the mid-Permian Berserker Group, and is overlain by Quaternary sediments.

Economic significance

Tests for oil yield from fine-grained samples from the three basins gave low or zero values (Dubow, 1979; Layton & Associates, 1981). Oil shale was present only in the central basin, and the potential of the basins is considered to be negligible.

Tertiary basalt (Tb, Tib and Tim)

(C.G. Murray)

Introduction

Remnants of Tertiary basalt lava flows (**Tb**) are widely distributed in the south-western part of the Yarrol Project area, sourced from numerous plugs (**Tib**). They all fall within the Monto province described by

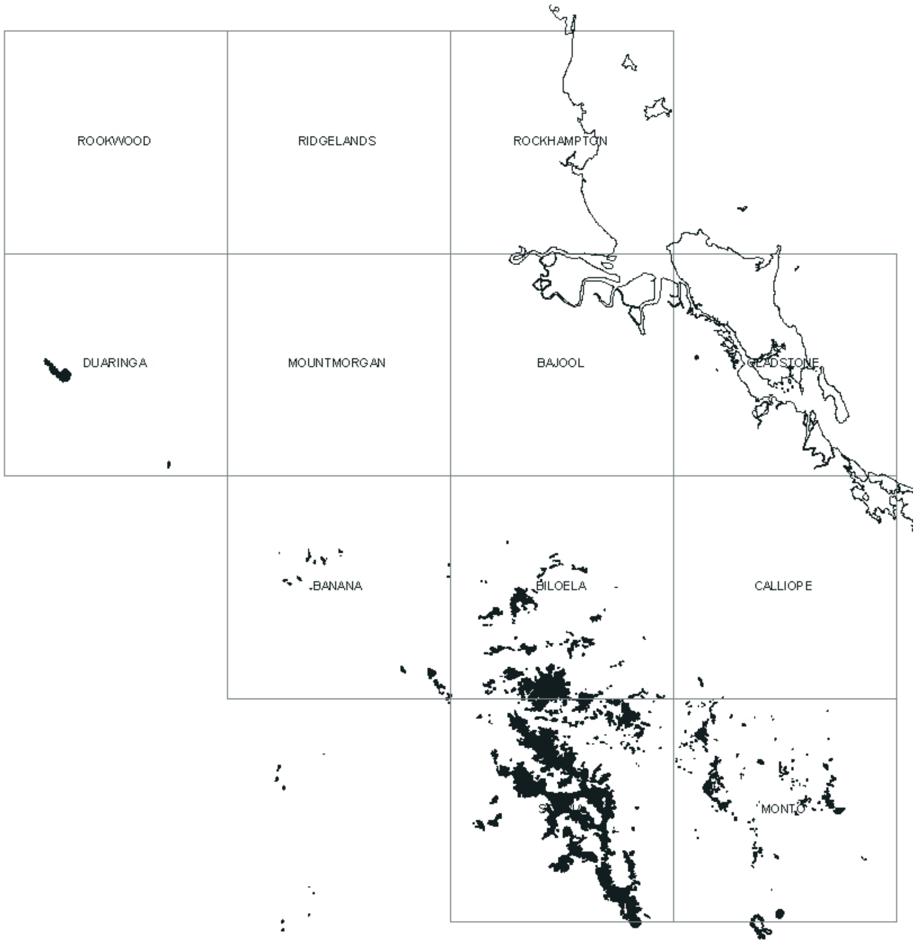


Figure 247. Distribution of Tertiary basalt

Sutherland & others (1989). On the Monto 1:100 000 Sheet area, at least one basaltic plug at Mount Coppin is known to contain phenocrysts of melilite and has been labelled as **Tim**.

Distribution

Definite Tertiary basalts occur only in the south-western and southern parts of the Yarrol Project area, in the Biloela, Scoria and Monto 1:100 000 sheet areas. In the north, basalts are mainly if not entirely of Cretaceous age. In the east (Bajool, Gladstone and Calliope 1:100 000 sheet areas), Tertiary basalts were either not erupted or are not preserved. However, numerous basaltic plugs of assumed Tertiary age have been interpreted from airborne magnetic data (Figure 247).

The Tertiary basalts are concentrated in the Biloela Basin and the Dawes Range to the south, with smaller outcrops extending to the east near Kalpowar. Isolated basalts also overlie and are interbedded with the Tertiary sediments of the Duaringa Basin.

Topographic expression

The basalts usually occur on top of small plateaux, and benches may indicate the presence of multiple flows (Figure 248). The basalt plugs form topographic prominences (Figure 249), which often display columnar jointing orientated at a high angle to the ground surface (Figure 250).

Geophysical expression

The radiometric signature of the basalts is low, and they have dark colours on images derived from the airborne survey data. Magnetic expression of the basalt flows is low to moderate, and is greatest in areas of maximum thickness in the Biloela Basin. Some basalt plugs are magnetic highs, but some were obviously intruded during a period of reversed magnetism, and form prominent magnetic lows.

Figure 248. Mount Kroombit 22km east of Biloela, capped by a remnant of a Tertiary basalt flow

Figure 249. Basalt plug 6km south-south-west of Many Peaks, Tertiary basalt

Figure 250. Columnar jointing dipping inwards, basalt plug 12km south-west of Many Peaks, Tertiary basalt

Lithology

In general the basalts are fine grained, with small phenocrysts dominated by olivine. In the thickest section in the southern Biloela Basin, basal flows of olivine and quartz tholeiite are overlain by alkali basalt and hawaiite (Sutherland & others, 1989). These authors also report tholeiitic basalts from the Monto area, and quartz tholeiite and K-rich alkali basalt and mugearite from further east around Kalpowar. The mugearite contains abundant upper mantle and lower crustal inclusions, as do several of the basaltic plugs in the region (Figure 251).

Some basalts appear to have undergone an extensive period of weathering, and are lateritised and associated with thick duricrust profiles, whereas others are unlateritised. The development of deep weathering profiles requires both time and minimal erosion (Grimes, 1988). Age is not the main criterion, because Cretaceous basalts in the area west of Rockhampton show little lateritic weathering.

Geochemistry

Geochemistry indicates that these are typical within plate basalts, similar to the Dalma Basalt (Figure 252). The analyses plot in the within plate basalt or equivalent field on standard discriminant diagrams (Figure 253). Both alkaline and subalkaline basalts were erupted, and compositions cover a wide field from nepheline-normative to quartz-normative. There was a general trend from initial alkali basalts to tholeiites and then to late-stage undersaturated alkaline rocks (Griffin & others, 1987). The main basaltic eruptions were Na-rich, and later plugs and small flows include K-rich, undersaturated rocks (Sutherland & others, 1989). Analyses mainly fall in the basalt field defined by Le Maitre (1989), but include basanite, hawaiite, potassic trachybasalt and basaltic andesite (Figure 254).

Thickness

The maximum thickness is in the Biloela Basin, where at least six flows have a thickness of up to 270m (Sutherland & others, 1989). In other areas, the Tertiary basalts are much smaller in volume and thinner.

Figure 251. Olivine and pyroxene inclusions in basalt plug 6km south-south-west of Many Peaks, Tertiary basalt

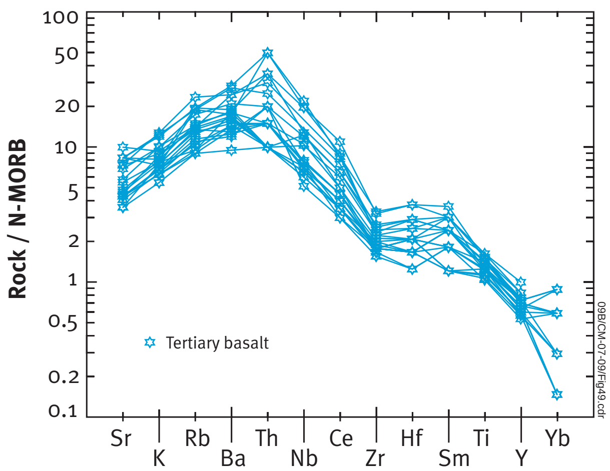


Figure 252. Spidergram plot of Tertiary basalts. N-MORB values are from Pearce (1983)

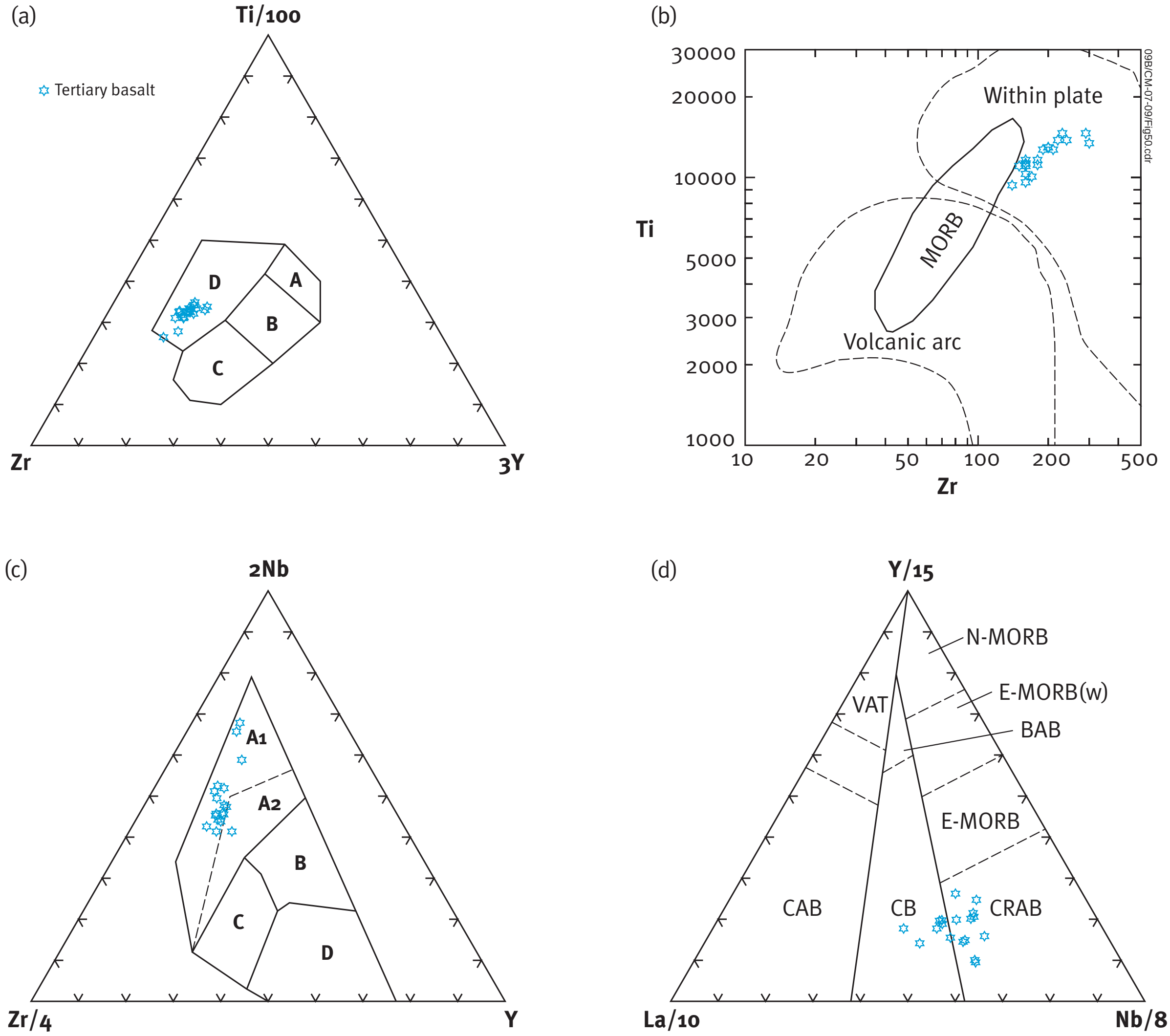


Figure 253. Analyses of Tertiary basalts plotted on discriminant diagrams of (a) Pearce & Cann (1973): A = island arc tholeiites, B = island arc tholeiites, volcanic arc basalts and mid-ocean ridge basalts or MORB, C = calc-alkali basalts, D = within plate basalts; (b) Pearce (1982); (c) Meschede (1986): A1 = within plate alkali basalts, A2 = within plate alkali basalts and within plate tholeiites; and (d) Cabanis & Lecolle (1989): CB = continental basalts, CRAB = alkali basalts from continental rifts. The analyses are essentially confined to the within plate basalt, within plate alkali basalt and continental basalt fields on these diagrams.

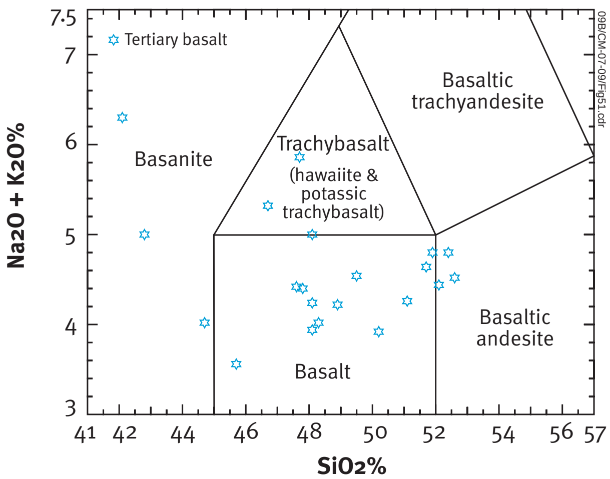


Figure 254. Total alkalis – SiO_2 plot showing Tertiary basalt analyses classified according to the scheme of Le Maitre (1989).

Structure

The Tertiary basalts are undeformed, and form flat-lying residuals.

Age

The main phase of basalt eruptions was about 25Ma. A K-Ar date of 25.4±0.5Ma was obtained from basalt at the southern end of the Biloela Basin (Grimes, 1988), and a mugearitic hawaiite flow east of Kalpowar gave an identical result of 25.2±0.3Ma (Sutherland, 1991). Hornblende from a sequence of basalt flows 3km east-south-east of Mount Bargogo east of the Cania dam yielded a similar age of 24.1±0.6Ma (F.L. Sutherland, unpublished data). Here, a coarse alkali basalt is overlain by K-rich nepheline mugearite with xenoliths including hydrous minerals (Griffin & others, 1987). The olivine nephelinite flow capping Mount Fort William east of Kalpowar is considerably younger, 20.3 and 20.7±0.4Ma (Sutherland, 1991).

All of the basaltic flows appear to have normal magnetism, but the plugs are both normal and reversed, suggesting that intrusion took place over a longer time span. However, limited dating results cover a similar time range. Pyroxene from a dolerite intruding the oil shale in The Narrows Graben has a K-Ar age of 26.8Ma, which is considered to be a reliable minimum age for crystallisation (Henstridge & Missen, 1981). A basanite plug near Glassford Creek west of the Lawyer Granite has been dated at 21.6±0.3Ma (F.L. Sutherland, unpublished data), and is reversely magnetised. All of these dates fall in the middle Tertiary, spanning the Oligocene–Miocene boundary. During this period the magnetic field reversed frequently, easily explaining the occurrence of normally and reversely magnetised rocks.

Basalt at the base of the Biloela Basin may be early Tertiary or Cretaceous, while that interbedded with the Tertiary oil shales of the Duaringa Basin must be of Eocene age.

Stratigraphic relations

The Tertiary basalt unconformably overlies numerous Phanerozoic stratified and intrusive units. Both the Tertiary basalt and the underlying Biloela Formation are essentially flat-lying, even though a large time gap exists between them.

Correlation with other units

The Tertiary basalts do not correlate with any other units in the Yarrol Project area, but are typical of the within plate basalts that extend around the eastern margin of Australia from Cape York to South Australia (Johnson, 1989).

Economic significance

Small gold deposits have been worked from sediments beneath Tertiary basalts in the Mount Rainbow goldfield 30km north-east of Biloela.

TERTIARY–QUATERNARY

Unnamed Tertiary to Quaternary sediments

(P.R. Blake)

Introduction

The unnamed Cainozoic sedimentary deposits have been treated as morphostratigraphic units, which differ from lithostratigraphic units in being defined on the basis of their preserved depositional landforms (Frye & Willman, 1962). Characteristic morphological types are grouped according to broad depositional setting (for example, alluvial plains, dunefields etc), and subdivided into individual types that have a distinctive morphology (Grimes, 1983). Age is determined by superposition and by the degree of preservation. The presence or absence of deep weathering profiles must be treated with caution, because correlations can rarely be made with confidence between different areas and units (Pain, 2005). From the early Cainozoic, an erosional phase affected the Yarrol project area (Jones, 2006), and deposition was essentially confined to relatively narrow belts along streams, at the base of steep slopes, and along the coastline.

These deposits were not studied during the Yarrol project, and were interpreted from airphotos. The main areas covered by Cainozoic units are along the Fitzroy River, south-west of Bajool, the Hedlow Creek alluvial

Figure 255. Distribution of TQa, Ta and Ts

plain, Callide Creek, and the present coastline. The following descriptions are based partly on previous work by Willmott & others (1986) and Donchak & Holmes (1991).

Alluvial deposits

Alluvial flood plain deposits of various ages occur throughout the Yarrol Province. The oldest recognised deposits are Tertiary in age (**Ta**). Sediment types include sandy claystone, clayey sandstone, siltstone and minor gravel beds. Despite their age, these are most commonly associated with the current river and stream valleys. However, locally, particularly in the southern part of the Monto 1:100 000 Sheet area, **Ta** forms flat, hill-capping units. Some sediments are mottled and appear to have been through a deep weathering cycle. Sediments mapped as **Ta** along the Fitzroy River as far north as Develin Creek may be outliers of Paleogene sediments deposited in the Biloela Basin. Younger alluvial deposits have been assigned a Tertiary to Quaternary age (**TQa**). **Ta** and **TQa** are usually semi-consolidated and smaller in extent than younger alluvial sediments. Small outcrops of quartzose sandstone and quartz-pebble conglomerate which overlie the Wandilla Formation and probably The Narrows beds along The Narrows in the Gladstone 1:100 000 Sheet area are labelled **Ts**. They are thought to have been deposited in a fluvial environment. The distribution of Ta, TQa and Ts is shown on Figure 255.

All of the Quaternary deposits are associated with modern fluvial systems and occur in river and stream valleys. Over most of the area, they comprise the modern flood plains, although some higher terraces may be included, and are simply shown as **Qa**. For example, in the Hedlow Creek plain, significant streams and terracing are absent and the alluvium has presumably been built up continuously. Silt and clay probably predominate in these areas. In the larger well developed river systems, the Quaternary alluvium can be subdivided on the basis of prominent terraces which represent different ages. From oldest to youngest, these are **Qpa**, **Qa_2**, **Qa_1**, and **Qha**. Terraces are assigned to **Qpa** if they are the highest and presumably oldest terraces or occur in places where it is impossible to determine if they are a **Qa_2** or a **Qa_1** terrace. **Qha** are the most recent deposits within the fluvial systems and mainly represent point bar deposits closely associated with the current channels.

Lateral migration of the main river channels has resulted in complex three-dimensional interfingering of the various types of deposits. For example, present day point bar deposits may overlie older channel sands, which may in turn overlie finer overbank deposits, these successive layers recording the relative position of the river channel, producing first the **Qa$_2$** terrace and then the **Qa$_1$** terrace. The down-cutting events were probably related to sea level changes. The preservation of fluvial depositional features (point bar deposits, levees, channel sands) becomes increasingly poorer with increasing terrace age, the oldest **Qpa** terrace being largely featureless. Little drill-hole information is available, but the alluvial deposits of the Fitzroy, Calliope and Boyne River systems may be up to 30m thick (Reid, 1946; Mather, 1980).

Fitzroy River

Deposition of the alluvium of the Fitzroy River valley has been a complex, ongoing process, involving migrations of the main river channel, deposition of point bar sediments, development of levees and overbank deposits, scouring of meander cut-offs, and gradual infilling of resulting oxbow lakes. Limited drilling by the Queensland Water Resources Commission has established no discernible pattern in the sediments, except that sand and gravel appear more common in the earlier, lower sediments, and silt and clay predominate in the uppermost layers (Reid, 1946; Willmott & others, 1986).

Qpa, **Qa$_2$**, **Qa$_1$**, and **Qha** are made up of silt, clay, sand and sandy gravel. **Qa$_2$** are the highest and apparently oldest terraces. **Qa$_1$** are lower terraces that cut into **Qa$_2$**. **Qha** are low terraces that cut both **Qa$_1$** and **Qa2**, and include the active river channels. Since the deposition of the oldest sediments (**Qa$_2$**) there appear to have been two major periods of downcutting by the river and infilling of lower, smaller terraces (**Qa$_1$** and **Qha**). In most areas the three major terrace levels are distinct, but in places there are intermediate terraces. The limited drilling information reveals no marked differences in the sediments beneath the **Qa$_2$** and **Qa$_1$** terraces, and the three subdivisions have been established essentially on geomorphological grounds. It could be expected, however, that coarse sediments will predominate in **Qha** areas, which are mainly point-bar deposits (Willmott & others, 1986).

A former major channel of the river, which is still evident west of Rockhampton, has cut into **Qa$_1$** and **Qa$_2$** deposits. The infilling sediments are shown as **Qha**, although they are probably somewhat older than the **Qha** terrace and point-bar deposits.

Along the Fitzroy River on the Rookwood and Ridgelands 1:100 000 Sheet areas there are some raised terraces that cannot be positively correlated with either **Qa$_2$** or **Qa$_1$** further downstream, and are labelled as **Qpa**.

No radiocarbon dates have been obtained from material in the alluvium, and relationships between the three terrace levels and changes in sea level cannot be determined.

Boyne and Calliope Rivers

In the lower reaches of the Boyne and Calliope Rivers, up to three periods of alluvial terrace development can be recognised, ranging from the highest and oldest flood plain terrace (**Qpa**), through an intermediate terrace (**Qa$_1$**) to the youngest terrace (**Qha**). The youngest terrace consists largely of point bar and channel sand and gravel deposits, whereas the older terraces are composed mainly of silt and clay. The distribution of Qa, Qha, Qa1, Qa2, and Qpa can be seen on Figure 256.

Fans and floodout sheets

Colluvial deposits, mainly of gravelly and bouldery clay, have accumulated around the base of the steeper topography in the Yarrol area. Their mode of transport is uncertain, but was probably by a combination of alluvial fan deposition where gullies exit from steeper country, sheetwash, and gravity (mainly in upper, steeper areas). In most areas, the deposits are no more than a few metres thick. The more consolidated deposits have been incised by modern gullies and streams. The boundaries with alluvium further downslope are approximate only.

Tertiary and Quaternary fans (**TQf** and **Qf**) have been mapped in several parts of the Yarrol Province (Figure 257). On the Scoria 1:100 000 Sheet area they were derived from the Wingfield Granite; on the Biloela 1:100 000 Sheet area they are shedding off Tertiary basalt and Precipice Sandstone; on the Ridgelands 1:100 000 Sheet area they are from the Princhester Serpentinite and Rookwood Volcanics; and on the Mount Morgan 1:100 000 Sheet area they shed off the Moah Creek beds, Rookwood Volcanics, Back Creek Group and the Camboon Volcanics.

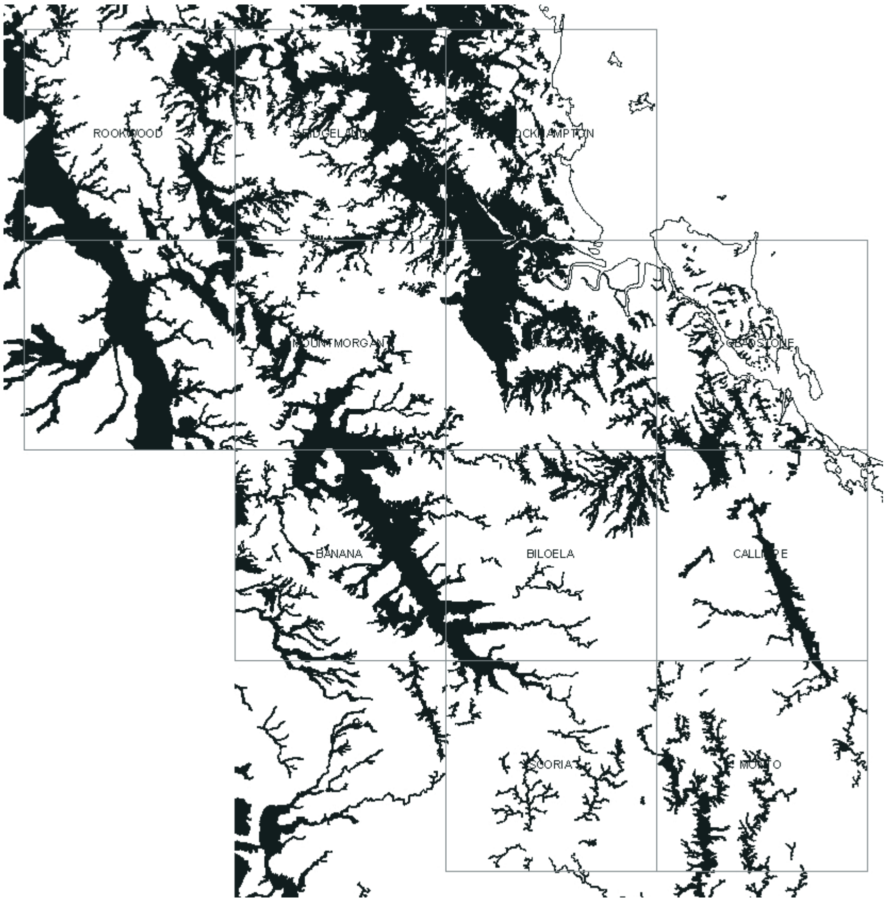


Figure 256. Distribution of Qa, Qha, Qa1, Qa2, and Qpa

Figure 257. Distribution of Qf and TQf

Figure 258. Distribution of TQcc, TQch, and TQgc

On the Rockhampton 1:100 000 Sheet area, two different deposit types have been recognised dependent on the parent rocks in the adjacent hills. **TQch** is composed of fragments of volcanics and forms on the flanks of Cretaceous plugs of the Mount Hedlow Trachyte. **TQcc** is a clayey gravel made up predominantly of chert fragments, and develops on the Curtis Island Group. **TQg$_c$** on the northern part of the Ridgelands 1:100 000 Sheet area is probably equivalent to **TQcc** (Figure 258).

The symbol **TQrh** is applied to colluvium and floodout sheets that have been shed off granites and their contact rocks on the Biloela 1:100 000 Sheet area (Figure 259). They occur mainly along the eastern side of the Calliope Range and south of Bell Creek, and consist of semi-consolidated sand and gravel a few metres thick incised by present day streams.

Residual deposits

Residual soil development is common on most rock types throughout the mapped area, but only occurs as a mappable unit in a few places. Tertiary and Quaternary residual deposits are a variety of sands, silts, soils and gravels (**TQr** and **Qr**) that occur locally over most of the Yarrol Project area, and are usually associated with the current drainage systems. They include a small proportion of transported material. The most extensive areas of **TQr** and **Qr** are in the Biloela Basin, where they form aprons of clay and sand deposited below the scarps of mesas of deeply weathered and duricrusted sediments. However, some remnants of **TQr** are found on the top of ridges and hills, as on the Mount Morgan Trondhjemite 10km south-east of Mount Morgan (Figure 260). Sands, silts, soils and gravels occurring near The Narrows Graben have been given the symbol **Qrs**. Apparently the low-lying, soft, easily weathered Tertiary sequence in the graben was a more favourable site for residual soil development than the more indurated and elevated Palaeozoic and Mesozoic rocks.

Subscripts are used to denote compositional variations, such as dominance of gravel and red or black soils derived from basalt and other mafic rocks. **TQr$_g$** comprising quartz and ironstone gravel, sand, soil, and conglomerate containing volcanic and sedimentary clasts occurs on the Biloela 1:100 000 Sheet area. Outcrops are proximal to Callide and Kroombit Creeks and usually overlie the Biloela Formation, but near Callide Dam they appear to overlie the Lochenbar Formation. One small outcrop surrounds a Tertiary basalt

Figure 259. Distribution of TQrh

Figure 260. Cemented residual gravelly granite wash on the Mount Morgan Trondhjemite, 10km south-east of Mount Morgan, TQr

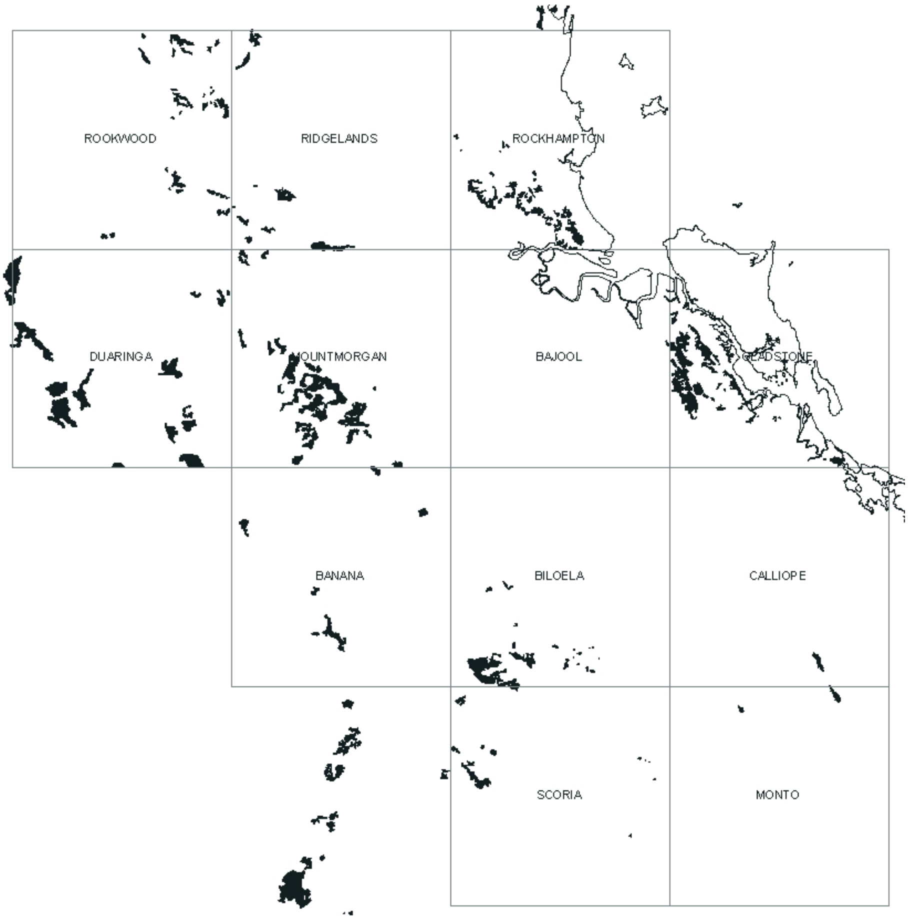


Figure 261. Distribution of TQr, TQrg, TQrr, Qr, and Qrs

plug. On the Gladstone 1:100 000 Sheet area, distinctive red soil developed on mafic intrusives has been labelled **TQr$_r$**. The distribution of TQr, TQrg, TQrr, Qr and Qrs is displayed on Figure 261.

Deep weathering profiles (duricrust, laterite, and silcrete)

Deep weathering profiles in the Yarrol Project area are mainly developed on Tertiary rocks, but smaller occurrences are known from older units. Deep weathering primarily depends on a long period of stability in the landscape (Pain, 2005). The deep weathering profiles are depicted in two ways on geological maps of the Yarrol Project area: they are either depicted as separate units, **Td** and **TQd$_q$**; or they are shown as a screen superimposed on the unit colour and pattern.

Pain (2005) has emphasised the problem of using the occurrence of deep weathering profiles and thick lateritic soils to assist in determining the age of Cainozoic rocks. Iron can cement any materials to form ferricretes (laterites) in any part of the landscape where it accumulates, so the presence or absence of ferricrete should not be used as the main indicator of age.

Lateritic weathering profiles (**Td**) are developed as mappable units over the Princhester Serpentinite at the northern edge of the Yarrol Project area. The weathering profiles are assigned to the Tertiary because they are faulted and tilted. They contain veins of chrysoprase and magnesite which have been or are being mined, and considerable resources of lateritic Ni-Co ore which has been mined to a limited extent in the past and continues to be of interest (Parianos, 1994).

Elsewhere **Td** has only been mapped on Tertiary sediments and basalt. The largest area is in the Biloela Basin, where a well developed deep weathering profile has formed on the Biloela Formation (**To**). Grimes (1988) described exposures of this profile, including one locality where weathered sediments are overlain by fresh basalt dated at 25.4±0.5Ma, giving a minimum age for the deep weathering in this area. Small outcrops of silcrete (**Td$_q$** and **TQd$_q$**) occur near the margins of the basin. Scattered occurrences of a deep weathering profile (**Td**) on Tertiary sediments (**Ta**) along the Fitzroy River on the Rookwood 1:100 000 Sheet area may represent a link between the profiles on the Princhester Serpentinite and the Biloela Formation, and all may have been produced by the same event.

Figure 262. Distribution of Td and TQdq

Near Monto, a deep weathering profile (**Td**) has been mapped on both Tertiary sediments (**Ta**) on a low plateau 20km east of Monto, and on Tertiary basalt (**Tb**) south of Monto. The proximity of these exposures suggests that they are related. If so, either the Tertiary basalt has a different age to that in the Biloela Basin, or they must have been the result of a different weathering event.

Deep weathering of other units in the Yarrol Project area is indicated by a screen on the geological maps. In the north-eastern part of the area, a lateritic profile is best developed on the colluvium and underlying mudstone and sandstone of the Wandilla Formation north of Yeppoon. There, the iron-rich upper zone generally occurs in the chert gravel of the colluvium (**TQcc**), and the mottled and pallid zones occur in mudstone of the Wandilla Formation. On the summits of the nearby chert ridges, the laterite is represented only by iron staining in the chert and in thin residual chert gravels. The colluvial chert gravels flanking the Doonside Formation are also commonly iron stained, but a complete laterite profile does not appear to have developed.

Also in the north-east, a bright red iron-rich zone of uncertain thickness is developed on the Rossmoya beds and Alton Downs Basalt in the Rossmoya and Milman North areas, and to a lesser extent on a small outcrop of Casuarina beds near the Rockhampton Botanic Gardens (Willmott & others, 1986).

In the Mulgildie Basin west of Monto, zones of deep weathering are developed on the Hutton Sandstone and Evergreen Formation.

The distribution of Td and TQdq is displayed in Figure 262.

Coastal deposits

The subdivision of coastal deposits in the Rockhampton 1:100 000 sheet area is based on previous studies by the Beach Protection Authority (1979), Hekel (1980) and Brooke & others (2006). Most of the subdivisions can also be recognised on the Gladstone 1:100 000 sheet area.

Figure 263. Distribution of Qpb and Qpe

Pleistocene beach ridges (Qpb) and estuarine – back barrier deposits (Qpe)

The oldest coastal deposits in the Yarrol Project area are Pleistocene beach ridges (**Qpb**) to the north of Farnborough on the Rockhampton 1:100 000 Sheet area. These were built up along a former shoreline and are composed of medium to fine sand. The age is interpreted from the degraded nature of the ridges. Estuarine and back barrier sediments adjacent to the beach ridges are grouped as **Qpe** (Figure 263). Sediments consist dominantly of mud and muddy sand. The position of the boundary between these Pleistocene deposits and Holocene estuarine sediments to the east is uncertain.

Frontal dunes (Qhf)

The symbol **Qhf** is used to designate foredunes and blow-out dunes behind modern beaches that are part of the present-day active beach zone. This dune system is mainly present on the Rockhampton 1:100 000 Sheet area, though it extends southward into the Bajool 1:100 000 Sheet area (Figure 264). South of Keppel Sands and north of Farnborough, small blowouts are developing within **Qhf**, thus broadening the unit. It consists of fine to medium sand.

Holocene beach ridges (Qhcb)

During the Holocene, north-north-west trending beach ridges (**Qhcb**) and associated back barrier (estuarine) deposits accumulated in the area between Gladstone and Hummock Hill Island of the Gladstone 1:100 000 Sheet area, as well as along the coast on the Rockhampton 1:100 000 Sheet area, and in the north-eastern part of the Calliope 1:100 000 Sheet area (Figure 265). Continual accretion of ridges onto coasts with shallow profiles and ample sediment supply has produced some wide areas made up of this unit, and led to an eastward advancement of parts of the coastline during the Holocene, particularly north of the Fitzroy River.

On the Gladstone 1:100 000 Sheet area these deposits have a maximum width of 1.2km. A broader series of beach ridges occurs at the northern end of Curtis Island. Here they trend north-west and are flanked by extensive areas of estuarine and back barrier deposits. The largest area is up to 2km wide and over 7km long.

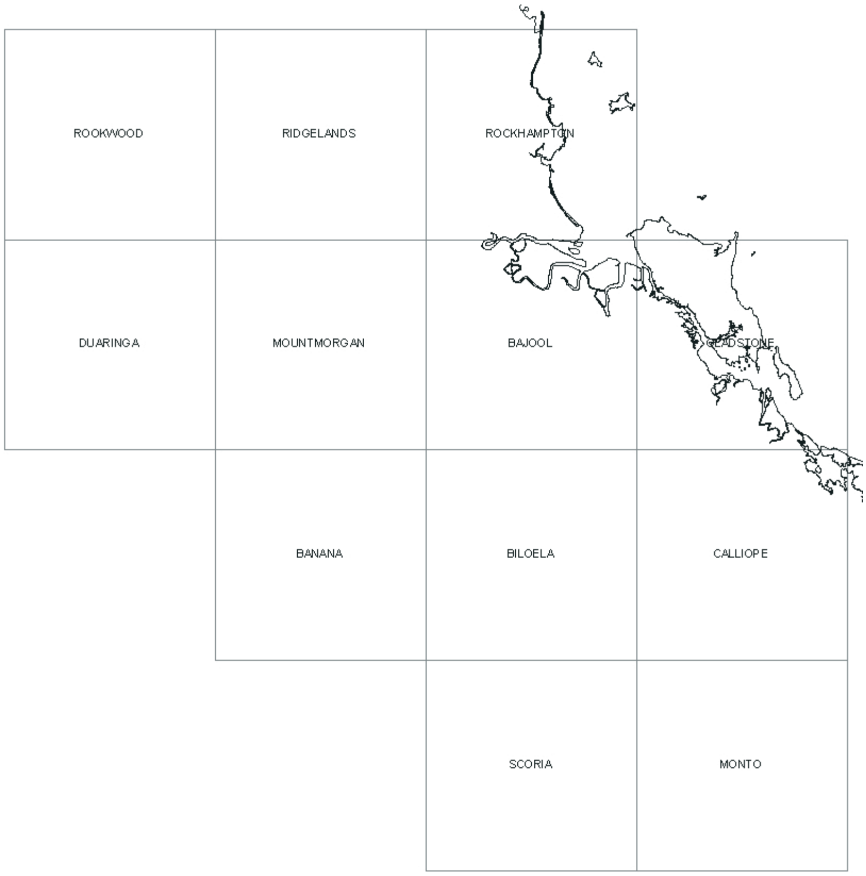


Figure 264. Distribution of Qhf

Figure 265. Distribution of Qhcb

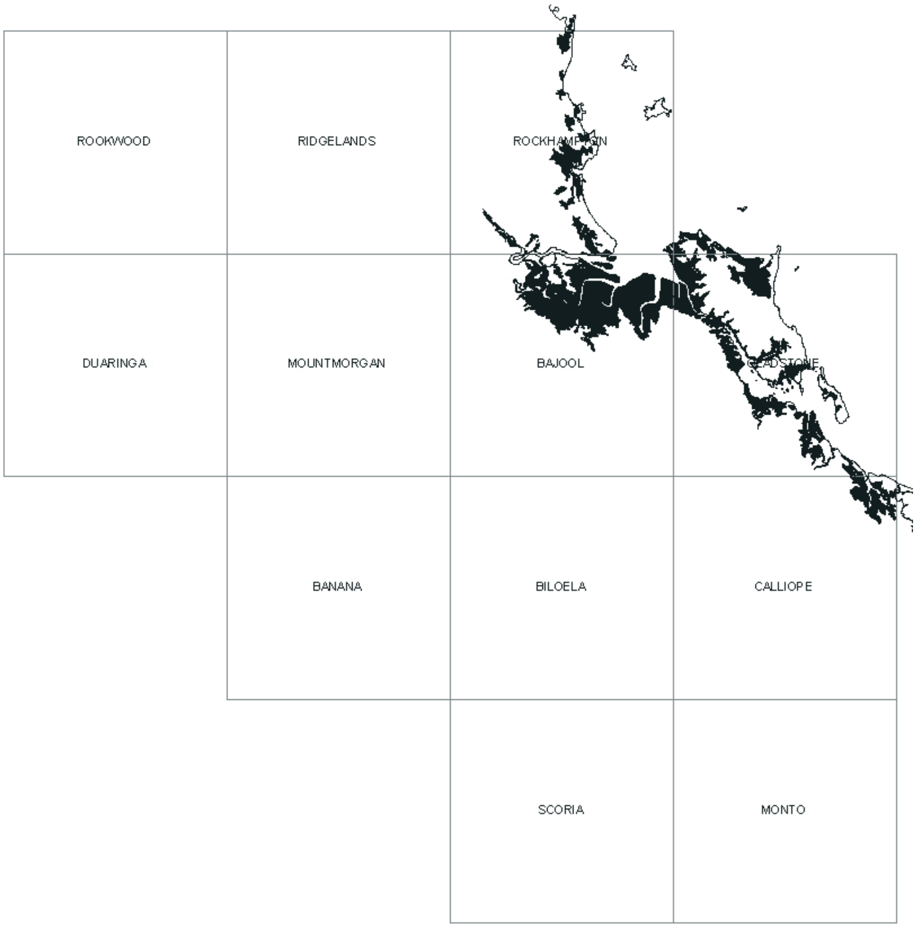


Figure 266. Distribution of Qhe

The beach ridges have only a low relief and are separated by parallel interdune depressions, in places occupied by swamps. The ridges consist of fine-grained quartz sand and minor organic-rich silt. Deposits of heavy minerals, chiefly ilmenite, are known to occur within the beach ridges, particularly on Wild Cattle and Hummock Hill Islands. A summary of exploration for heavy minerals on these islands can be found in Blake & others (1996).

On the Rockhampton 1:100 000 Sheet area up to 4km of **Qhcb** has been accreted near Cattle Point. Brooke & others (2006) documented that over the last 1500 years, ridges were emplaced in rapid episodes following periods of high sediment discharge from the Fitzroy River, approximately once every 500–200 years.

Holocene estuarine – back barrier deposits (Qhe)

Estuarine environments are associated with the coastal deposits and lower reaches of fluvial systems. Sediments consist dominantly of mud and muddy sand, and on the mangrove flats have a high organic content. Current estuarine deposits (**Qhe**) have been mapped on the Gladstone, Calliope, Rockhampton and Bajool 1:100 000 Sheet areas (Figure 266). These sediments are estuarine or back barrier deposits that accumulated between beach-ridge complexes, as well as adjacent to major creek mouths and in the lower reaches of the Fitzroy River. The estuarine environments include tidal and mangrove flats, supratidal flats and coastal grasslands.

Holocene high dune systems (Qhcd)

The unit **Qhcd** (Figure 267) comprises high blow-out dunes that are assumed to have formed during the transgression of the sea in the early Holocene, and consist of fine to medium sand.

On the Gladstone 1:100 000 Sheet area, large north-west-trending blow-out dunes up to 80m above sea level occur on the north-eastern part of Curtis Island. In places they form relatively thin surficial deposits of wind blown sand resting on shallow bedrock. Concentrations of heavy minerals occur along the active eastern

Figure 267. Distribution of Qhcd

beach but the limited information available suggests that grades in the high dunes themselves would be low. Smaller dune systems occur along the eastern side of Facing Island.

On the Rockhampton 1:100 000 Sheet area, limited areas of high blowout dunes are mainly present near Emu Park, Bluff Point, Farnborough, Great Keppel Island and North Keppel Island.

Tidal delta sands (Qhmt)

Marine tidal sand deltas (**Qhmt**) (Figure 268) occur on the Gladstone 1:100 000 Sheet area at the far northern end of Curtis Island and in three areas south of Facing Island. The Curtis Island, Boyne River and Colosseum Inlet deltas were produced by currents produced by tidal flushing of local inlets. The East Banks delta was formed by currents associated with tidal flushing of Gladstone Harbour. The deltas are composed of sand and are well displayed on aerial photographs. Their fluctuations can be documented using satellite imagery.

Swamp deposits

In the Fitzroy River valley in the north-eastern part of the Rookwood 1:100 000 Sheet area is a small area of clay and silt swamp deposits (**Qhw**) overlying a weathered granitoid (Figure 269).

Anthropogenomorphic deposits

In some parts of the Yarrol Project area, human activity has piled up dirt and rock to a scale where they are mappable entities. These are displayed as **Qhh** on the geological maps. Their main occurrence is as spoil piles around the large open cut Callide coal mine and the abandoned Mount Morgan gold-copper mine. They also occur at Gladstone where landfill has been deposited as part of a land reclamation program (Figure 270).

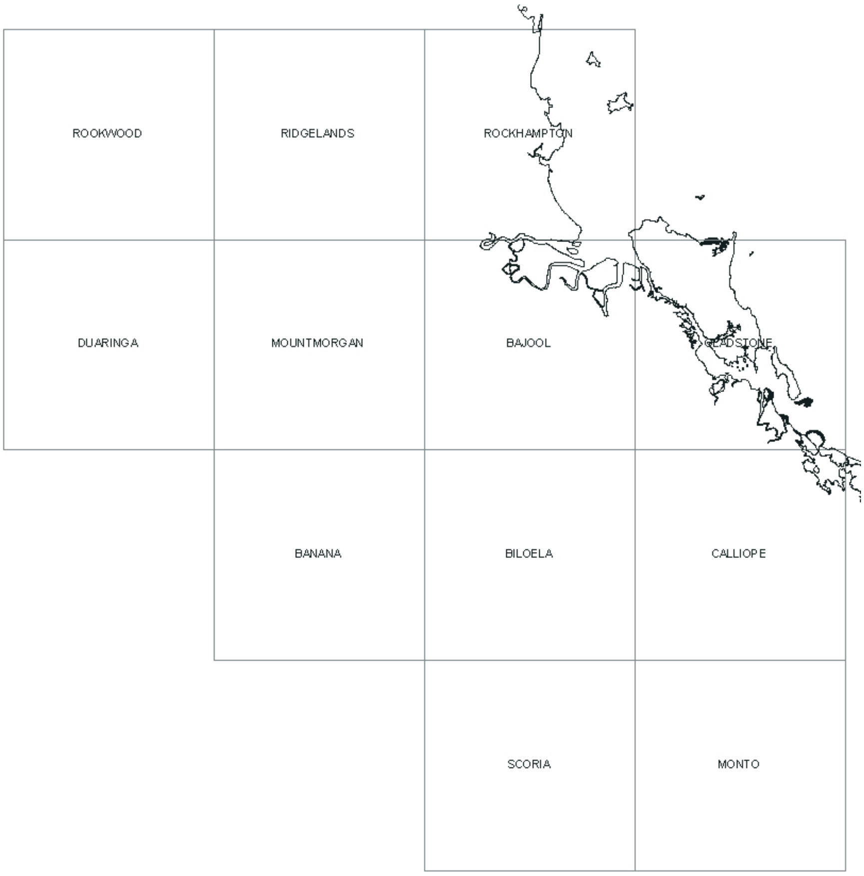


Figure 268. Distribution of Qhmt

Figure 269. Distribution of Qhw

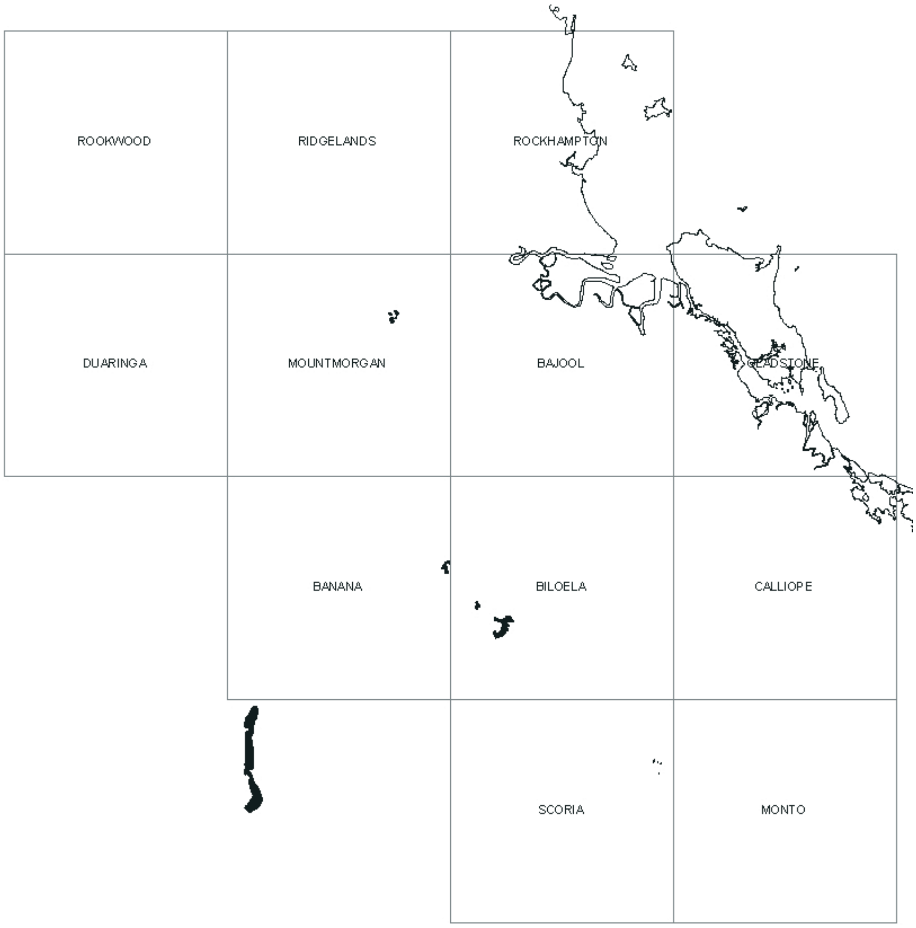


Figure 270. Distribution of Qhh

INTRUSIVES

NEOPROTEROZOIC AND DEVONIAN

Princhester Serpentinite and unnamed serpentinite (PPzsp and PPzx)

(I.W. Withnall, S.B.S. Crouch, & C.G. Murray)

Introduction

Murray (1969) described three large masses of ultramafic rocks which were mapped on the Port Clinton, Saint Lawrence, Duaringa and Rockhampton 1:250 000 Sheet areas. The name Princhester Serpentinite is applied to the ultramafic rocks, which are part of the Marlborough Block. The map by Harbort (2001) gave separate names to the serpentinite in each of the thrust sheets in this area, but because the map used the same geographic names for different units, as well as names that are unavailable because of prior usage, its nomenclature cannot be formalised. Therefore the name Princhester Serpentinite (one of the names used by Harbort) is used for simplicity for all of the serpentinite that forms multiple thrust sheets. The names ‘Marlborough ophiolite’ and ‘Marlborough ultramafics’ have also been used by various authors, but the name ‘Marlborough Metamorphics’ has prior usage.

Distribution

The Princhester Serpentinite crops out over about 580km^2 as several sinuous belts in a roughly rectangular area that extends from near Marlborough in the north-west to Canoona in the south-east. Most of this outcrop area lies north of the Yarrol Project area. The Princhester Serpentinite only occurs on the Ridgelands and Rookwood 1:100 000 geological maps, but it is probable that unnamed serpentinite bodies to the south-east have a similar origin. The distribution of serpentinites is illustrated in Figure 271.

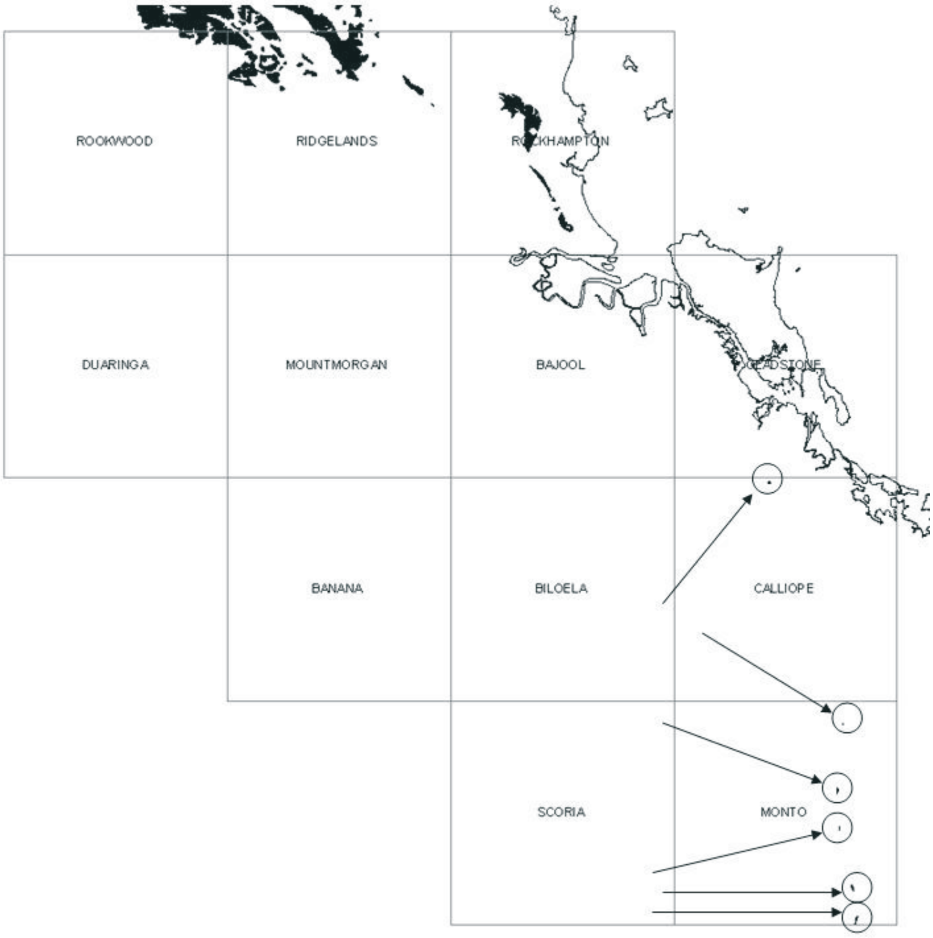


Figure 271. Distribution of the Princhester Serpentinite and unnamed serpentinite

The Balnagowan–Cawarral–Bondoola mass of unnamed serpentinite occurs east and north-east of Rockhampton, and has a length of about 40km and an average width of less than 2km.

Small lenses of unnamed serpentinite are also present along the Yarrol Fault. These are probably more numerous than those mapped east of Kalpowar and near Poperima Creek on the Monto 1:250 000 Sheet. Serpentinite also occurs within the Calliope beds west of the Yarrol Fault, most notably in the town of Calliope.

Topographic expression

The ultramafic rocks form high, sharp-crested ridges, with short dissected spurs, but the intervening flat areas are also generally underlain by ultramafic rocks. Lateritic and siliceous caps developed on the ultramafic rocks during the Tertiary, but late Cainozoic faulting and erosion has resulted in the removal of much of the laterite. Vegetation is dense where the weathered caps are preserved, but elsewhere it is moderate. Ultramafic rocks are indicated by numerous grass trees (*Xanthorrhoea* sp.) and spinifex (*Triodia hostilis*).

Geophysical expression

The ultramafic bodies have an extremely low radiometric response in all channels, and a very high magnetic signature. Two bodies with a high magnetic signature similar to that represented by the ultramafics are concealed by Quaternary sediments at the mouth of the Fitzroy River, and at Hedlow Creek near Iron Pot Mountain.

In general, the serpentinites coincide with gravity highs, but there are exceptions. The gravity pattern suggests that the serpentinite at Mount Wheeler east of the Jim Crow Basin is a very thin sheet.

Type locality

No type locality has been assigned to the Princhester Serpentinite.

Lithology

The Princhester Serpentinite has been regarded as an ophiolite by many authors (for example, Holcombe & others, 1997a; Bruce & others, 2000b; Spaggiari & others, 2003), although it should be pointed out that it represents only the basal component of an ophiolite, the harzburgite tectonite, and the higher level non-cumulate gabbros, diorites, plagiogranites, mafic sheeted dykes and mafic volcanic rocks are mostly absent, except as minor fault-bounded blocks and dykes.

According to Bruce & others (2000b), serpentinised mantle peridotite comprises about 95% of the Princhester Serpentinite. Primary mineralogy has mostly been entirely replaced or pseudomorphed by serpentine with only scattered relicts remaining, but a dominant tectonite harzburgite protolith is apparent. Typically it contained about 75% olivine (forsterite, Fo_{92-93}), 20% orthopyroxene (enstatite, En_{91}), 4% diopside and 1% chrome or magnesian spinel (Murray, 1969; Bruce & others, 2000b). The least serpentinised rocks consist of more than 50% olivine and pyroxene. Olivine forms equidimensional grains up to 7mm in diameter. Some crystals show kink bands parallel to (100), or undulatory extinction. Orthopyroxene occurs as irregular prisms with exsolution lamellae of diopside parallel to (100). Murray (1969) also described minor pyroxenite cumulate and dunite in the harzburgite.

Most serpentinised rocks are massive and retain the texture of the harzburgite. Dark green enstatite pseudomorphs can be distinguished from the lighter green serpentinised olivine. The original grains are outlined by fine magnetite grains (Murray, 1969). Schistose serpentinite is developed at some margins of the larger masses, and throughout the small lenses. It consists of fragments and blocks of massive serpentinite surrounded by thin zones of sheared serpentinite. The massive fragments retain the original harzburgite texture and are composed of irregularly orientated antigorite laths. The sheared zones exhibit a glossy lustre, commonly show slickensides, and consist of slip-fibre chrysotile. Some of the small lenses have brecciated margins. Bruce & others (2000b) divided the serpentinites into three classes on the basis of microscopic texture: pseudomorphic, transitional, and non-pseudomorphic (O'Hanley, 1995).

Small chromite bodies are scattered throughout the serpentinite. They are generally massive and podiform in texture with chromite grains interspersed with gangue material.

A variety of mafic rocks occur as inclusions and dykes within the serpentinite (Bruce & others, 2000b; Bruce & Niu, 2000a,b). These have been subdivided into groups on the basis of geochemistry and age. The oldest group (V1 of Bruce & others, 2000b) includes fault-bounded blocks of non-layered, medium to very coarse-grained metagabbro with small basaltic pockets interpreted to represent trapped melt, and fine- to medium-grained mafic dykes. In most of these rocks, metamorphism is of low grade, and higher grades up to amphibolite facies may have been caused by the intrusion of nearby granitoid stocks. Locally, anorthite veins crosscut the metagabbro, which grades into Ca-metasomatised rodingite with anorthite partially saussuritised and pseudomorphed by hydrogrossular (Bruce & others, 2000b). At least three younger phases of mafic intrusions are present, mainly dykes, and are described by Bruce & Niu (2000a,b) as the V2, V3 and V4 groups.

Geochemistry

Bruce (1999), Bruce & others (2000b) and Bruce & Niu (2000a,b) analysed both the ultramafic rocks and the mafic inclusions. They concluded that the ultramafic geochemistry was the result of two processes, initial partial melting and subsequent interaction with basaltic melts, and recognised three separate groups of serpentinite. Murray (2007) interpreted the data of Bruce (1999) in the same way, but proposed that subduction-related basaltic rocks of Late Devonian age within the Princhester Serpentinite (the V2 group of Bruce & Niu, 2000a) had reacted with the depleted peridotite to produce the full range of compositions reported by Bruce & others (2000b), including U-shaped rare earth element patterns. Tholeiitic basalts within the V2 group have similar geochemistry to Upper Devonian basalts in units south of Mount Morgan (Murray & Blake, 2005).

The geochemistry of the mafic blocks and dykes was interpreted by Bruce & others (2000b) and Bruce & Niu (2000a,b). The oldest group (V1) of Neoproterozoic age has N-MORB chemistry and was considered to be related to the original formation of the ultramafics at an oceanic spreading centre. The V2 group was dated as Late Devonian and is typical of rocks in modern island arcs. The V3 group gives a latest Carboniferous age and the chemistry suggests that these basalts are continental within plate basalts. The youngest group, V4, is typified by a calc-alkaline arc signature and was assigned to the Permian based on a K-Ar age from diorite in the type area in the Percy Islands. Murray (2007) proposed minor changes in terms of samples actually assigned to particular groups, but agreed with the interpretation of tectonic setting.

Structure

Within the Marlborough Block, the Princhester Serpentinite forms alternating slices with metamorphic rocks of the Marlborough Metamorphics. Both faulted contacts, metamorphic foliation and bedding (where visible) dip at moderate angles to the east (Henderson & others, 1993; Holcombe & others, 1997a; Harbort, 2001). Regional gravity data (Darby, 1969, plate 11) indicate that the Marlborough Block is a flat sheet less than 3km thick, and Murray (1974) postulated that it was emplaced to its present position by at least 60km of westward thrusting. The mechanical problem of emplacing a thin sheet over such a distance led Harbort (2001) and Harbort & others (2001) to propose that the thrust-imbricated slices of the Marlborough Block are a thrust duplex in an out-of-sequence nappe structure that was thrust over the rocks of the northern part of the Yarrol Province and adjacent Gogango Overfolded Zone. Holcombe & others (1997a) named the basal thrust the Marlborough Thrust, and noted that it “is a brittle structure with little deformation, even within metres of the contact”. Internal faults within the block are associated with kilometre-scale ductile deformation and although now steep, are thought to be rotated thrusts or remnants of extensional faults. The serpentinite is largely massive with localised cleaved zones, mainly in regions disrupted by late (Cretaceous and Tertiary) normal faulting, and strongly schistose zones near the contacts with surrounding units. The bases of the serpentinite sheets are metamorphosed to upper greenschist and amphibolite facies and generally display high angle, ductile thrusting with east-over-west shear sense indicators.

The mechanism of emplacement of the Princhester ophiolite is not certain. An early model put forward by Murray (1974), before the much older age was known, suggested a crude obduction mechanism driven by initiation of an Early Permian subduction zone dipping beneath the continent. Bruce & others (2000b) and Bruce & Niu (2000a) did not provide a detailed emplacement mechanism. They argued that if the V2 and V3 magmatic events represented island arc and continental intra-plate volcanism respectively, emplacement was constrained between 390Ma and 295Ma. The westward thrusting by the out-of-sequence thrust in the late Permian was a later event. This suggests a gap of at least 180 million years between the formation of the oceanic lithosphere represented by the Princhester Serpentinite at 560Ma and its emplacement onto the continental margin in the Late Devonian or later. Spaggiari & others (2003) produced a generalised tectonic reconstruction that linked the ophiolites of the New England Orogen with those of the Lachlan Orogen, reflecting a period of complex oceanic crust formation in an oceanic or back-arc position above an east-dipping subduction zone on the edge of Gondwana from 560–495Ma. No explanation was offered for the huge time-gap between the emplacement of the Lachlan ophiolites in the 515–440Ma interval and the New England Orogen ophiolites in the mid-Carboniferous. The models of Bruce & others (2000b) and Spaggiari & others (2003) present problems because it is difficult to envisage a tectonic scenario that would allow oceanic crust and upper mantle to avoid either subduction or renewed spreading over such extended periods of time. It would also have been difficult to detach a thin slice from an old lithosphere that had become cold and strong. Old dense oceanic lithosphere becomes denser and is easily subducted. Ophiolites can only form from young, hot and thin lithosphere and must be emplaced within a short time of their generation, typically 10 million years or less (Dewey, 2003).

Murray (2007) proposed that the Princhester Serpentinite was originally created at an oceanic spreading centre as the depleted mantle section of an ophiolite in latest Proterozoic time (V1 event), and subsequently was converted to a supra-subduction zone ophiolite by reaction with Late Devonian island arc basalts accompanied by a small degree of additional partial melting above an east-dipping subduction zone (V2 event). Blockage of the subduction zone by collision with the Australian continent during the Late Devonian led to slab break-off and the reversal of subduction direction, trapping the Late Devonian ophiolite in a forearc position. Its location, in a forearc setting above a growing accretionary wedge, conforms to the definition of a Cordilleran-type ophiolite (Moores, 1982). This interpretation is consistent with the geochemistry of the ultramafic rocks and with current views that most ophiolites are formed from young, hot and thin oceanic lithosphere at forearc, intra-arc and backarc spreading centres in a supra-subduction zone setting (Dewey, 2003; Hawkins, 2003; Pearce, 2003).

Age

Bruce & others (2000b) and Bruce & Niu (2000a) reported the results of Sm-Nd age determinations on mafic rocks in the Princhester Serpentinite. Five samples of the V1 group with smooth N-MORB-like rare earth element patterns gave a well-constrained whole rock Sm-Nd isochron that yielded an age of 562±22Ma (late Neoproterozoic). This was interpreted as the age of formation of the ultramafic rocks by partial melting at a sea-floor spreading centre in an extensive oceanic basin. Four dolerites from the V2 group gave a Late Devonian isochron of 380±19Ma, interpreted by Murray (2007) as the age of SSZ ophiolite generation. An age of 293±35Ma was obtained for three whole rocks and a plagioclase separate from the V3 magmatic episode.

The timing of final emplacement of the Marlborough duplex is provided by the age of mid-Permian rocks which underlie the basal thrust, and by the age of intrusive rocks that either are cut by the basal detachment or pierce both the detachment and the internal thrusts. Harbort & others (2001) referred to 20 previously published K-Ar ages and 3 Rb-Sr ages supplemented by 22 $^{40}Ar/^{39}Ar$ step-heating results that constrained emplacement of the Marlborough Block to ~253Ma (late Permian) over an interval of 8 million years. If correct, this means either that the K-Ar dates of 264 and 269Ma obtained by Webb (1969) from hornblende of the Ridgelands Granodiorite (now the Wattlebank Granodiorite) intruding serpentinite on the south bank of the Fitzroy River near Eden Bann homestead are incorrect, or that the granodiorite was intruded before the thrusting event and is part of the nappe sheet. Near Milman railway siding and Eden Bann, the Princhester Serpentinite has been thrust over the Berserker Group and correlatives. Therefore, the overlap of these ages with that obtained from U-Pb zircon dating of the Ellrott Rhyolite of the Berserker Group (268.2±3.9Ma; Appendix 1) suggests that the dates obtained by Webb (1969) must be older than the emplacement event for the Princhester Serpentinite.

The metamorphic rocks interleaved with the ultramafics of the Marlborough Block have been dated by Rb-Sr at 245±2Ma and 242±16Ma (Henderson & others, 1993) and by Ar-Ar at 248.8±0.5Ma and 242.8±0.4Ma (Holcombe & others, 1997a,b).

Stratigraphic relationships

Within the Marlborough Block, the Princhester Serpentinite forms thrust-imbricated slices interleaved with upper and lower greenschist facies metasedimentary units, and lower amphibolite facies S-type metagranite, meta-sedimentary rocks and metabasite of the Marlborough Metamorphics. The age of these rocks is unknown, although Holcombe & others (1997b) thought that some may be equivalents of the accretionary wedge to the east. The occurrence of limestone beds (now marble) in the lower grade rocks makes such a correlation unlikely.

Within the Yarrol Project area, the Princhester Serpentinite is thrust over the lower to mid-Permian Berserker Group and correlatives, as well as the uppermost Silurian to Middle Devonian Craigilee beds, the Upper Devonian to Lower Carboniferous Mount Alma Formation, and the Lower Carboniferous (Mississippian) Rockhampton Group.

The serpentinite is intruded by the mid- to late Permian Wattlebank Granodiorite and by unnamed and undated gabbro and granodiorite intrusions. Ages from intrusions outside the Yarrol Project area constrain thrust emplacement of the Marlborough Block at ~253Ma (latest Permian) or earlier.

Tertiary weathering produced lateritic cappings over the ultramafic rocks of the Marlborough Block and remnants of these are preserved as low plateaux, some of which are tilted. The Princhester Serpentinite is overlain by Quaternary sediments along stream courses.

Economic significance

Chromite occurs in the ultramafic rocks, but production has been minor due to the small size and low grade of deposits (Krosch, 1990c). Lateritic profiles on the Princhester Serpentinite are enriched in nickel and cobalt and have been the subject of exploration, mining, and mining proposals (Zeissink & Hewitt, 1974; INAL Staff, 1975; Brooks, 1979; Burger, 1989; Parianos, 1994). Chrysoprase is also mined from the laterite (Robertson, 1986; Grimes, 1988; Robertson & Krosch, 1989; Krosch, 1990b; Vasconcelos & Singh, 1996), but not currently within the Yarrol Project area. Magnesite occurs as veins in weathered serpentinite (Brooks, 1976; Cooper, 1989), but much larger deposits occur in old drainage channels in the vicinity of the ultramafics, as at Kunwarara and at the top of basins such as the Yaamba Basin and the Jim Crow Basin (Frost & others, 1988; Cooper, 1989; Burban, 1990; Milburn & Wilcock, 1994). The serpentinites contain thin veins of chrysotile asbestos (Simpson, 1976; Geological Survey of Queensland, 1978; Allen, 1980).

DEVONIAN

Mount Morgan Trondhjemite (Dgmo)

(C.G. Murray)

Introduction

The Mount Morgan Tonalite was defined by Kirkegaard & others (1970, page 86), who discussed earlier nomenclature. They described the dominant rock type as a tonalite according to the classification system of Morgan (1964), but also recognised the presence of more mafic phases.

Figure 272. Distribution of the Mount Morgan Trondhjemite

The intrusion shows most variation at its northern end, around Mount Morgan, where detailed studies have been carried out by Cornelius (1968) and Messenger (1996). Here, the dominant rock type is trondhjemite as classified by the IUGS (Le Maitre, 1989) and Barker (1979), and trondhjemite is virtually the only rock type exposed in the more extensive part of the intrusion south of Mount Morgan. The name of the unit is therefore changed to Mount Morgan Trondhjemite. Subdivisions comprising trondhjemite, tonalite, quartz diorite to quartz gabbro, and breccia have been mapped out.

Distribution

The Mount Morgan Trondhjemite forms an elongate belt about 25km long and averaging 3–4km wide which extends south-south-east from Mount Morgan (Figure 272). In the northern part of the intrusion around Mount Morgan, trondhjemite (**Dgmo$_r$**) occurs to the east of the belt of Capella Creek Group which hosts the Mount Morgan gold-copper orebody (the Mine Corridor), whereas tonalite (**Dgmo$_t$**) is largely confined to the west. These different phases were recognised by geologists working on the Mount Morgan mine, who named them the Town Granite and Mount Morgan Granite, respectively (Staines, 1952, 1953; The Staff, Mount Morgan Limited, 1965). Relatively small bodies of quartz diorite to quartz gabbro (**Dgmo$_d$**) have been mapped both east and west of the Mine Corridor. South of the Mine Corridor, the intrusion is entirely trondhjemite in composition, with the exception of breccia (**Dgmo$_c$**) along Bull Creek.

Topographic expression

The intrusion forms elevated country ranging from 300 to 500m in height. Well developed jointing results in surface exposures being mainly rubble up to boulder size, rather than large tors (Figure 273). Extensive pavement outcrops occur in deeply incised streams such as Nine Mile, Raspberry and Capella Creeks and the Dee River. A plateau area at the head of Bull Creek is covered by a layer of cemented granite wash at least one metre thick, interpreted as residual regolith (**TQr**) (Figure 261).

Figure 273. Closely spaced jointing in the bed of Raspberry Creek 13.5km south-south-east of Mount Morgan, Mount Morgan Trondhjemite

Geophysical expression

Overall, the Mount Morgan Trondhjemite is slightly more magnetic, and gives a higher radiometric response, than the surrounding Capella Creek Group and Mount Hoopbound Formation. However, the geophysical response is patchy, possibly reflecting variations in alteration. More mafic phases have a much stronger magnetic signature. An area along the Burnett Highway between Hamilton and Trotter Creeks gives a high magnetic and low radiometric response, and is interpreted as quartz diorite and quartz gabbro. Part of this area was mapped as gabbro by Nimmo (1992).

Lithology and petrography

The Mount Morgan Trondhjemite ranges in composition from trondhjemite through tonalite and quartz diorite to quartz gabbro, but trondhjemite is by far the dominant rock type. The intrusive suite is characterised by its quartz-rich nature and by the paucity of K-bearing minerals such as K-feldspar and biotite. The ubiquitous presence of micrographic quartz-plagioclase intergrowths, the medium-grain size with local porphyritic texture, and the occurrence of breccia pipes all suggest that it was a high-level pluton (epizonal in the sense of Buddington, 1959).

Trondhjemite ($Dgmo_r$) is a pale grey, leucocratic rock characterised by large quartz grains with a waxy appearance. It is very uniform in composition, consisting of 90% or more of plagioclase and quartz. Plagioclase (45–60% of rock) forms subhedral prisms up to 4mm long, and small grains intergrown with quartz. Measurement of extinction angles indicates that crystal cores are andesine in composition, and this is confirmed by microprobe analyses (Messenger, 1996). The plagioclase is normally zoned to thin, more sodic rims. In many specimens, the plagioclase is albitised, with a cloudy appearance due to replacement by epidote and clay, particularly in more calcic crystal cores. Quartz (35–45% of rock) occurs as rounded grains up to 6mm across, or as granular aggregates. Micrographic to myrmekitic intergrowths of plagioclase and quartz are present in almost all thin sections (Figures 274 and 275). Similar intergrowths have been described from

Figure 274. Photomicrograph showing granophyric intergrowth between plagioclase and quartz, Mount Morgan Trondhjemite. Crossed nicols. Scale bar is 1mm long.

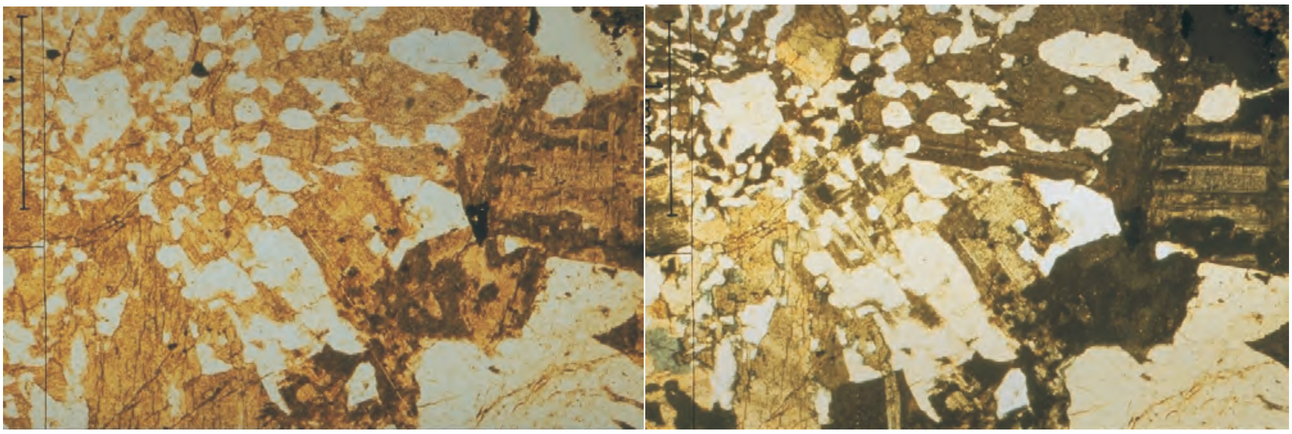


Figure 275. Photomicrograph showing granophyric intergrowth between plagioclase and quartz, and K-feldspar rimming and replacing plagioclase, Mount Morgan Trondhjemite. Plane polarised light and crossed nicols. Scale bar is 1mm long.

several trondhjemites formed in oceanic settings (Coleman & Donato, 1979, pages 153–154; Barker & others, 1979, page 535; Phelps, 1979, page 553; Gill & Stork, 1979, page 636). Most of the trondhjemite is totally devoid of K-feldspar. However, up to 10% K-feldspar occurs locally in specimens collected from Raspberry Creek, in which it rims and replaces plagioclase (Figure 276). This clearly indicates that the lack of K-feldspar is not the result of sodium metasomatism and albitisation, but is a primary feature related to the unusual potassium-poor nature of the magma.

Mafic minerals constitute 10% or less of rocks mapped as trondhjemite, except at one locality on the eastern outskirts of Mount Morgan, where they make up 15% of the rock. Brownish-green hornblende is dominant, with only minor biotite. Hornblende ranges in habit from elongate prisms to anhedral grains and aggregates. It is typically altered to actinolite, epidote, chlorite, and calcite. Biotite is altered to chlorite ± small quantities of rutile. Minor titaniferous magnetite (Messenger, 1996), and accessory apatite, sphene and zircon are present. In some specimens, sphene is partially replaced by rutile, calcite and quartz (Figure 277). Similar alteration assemblages have been reported by Corlett & McIlreath (1974; see also Deer & others, 1982,

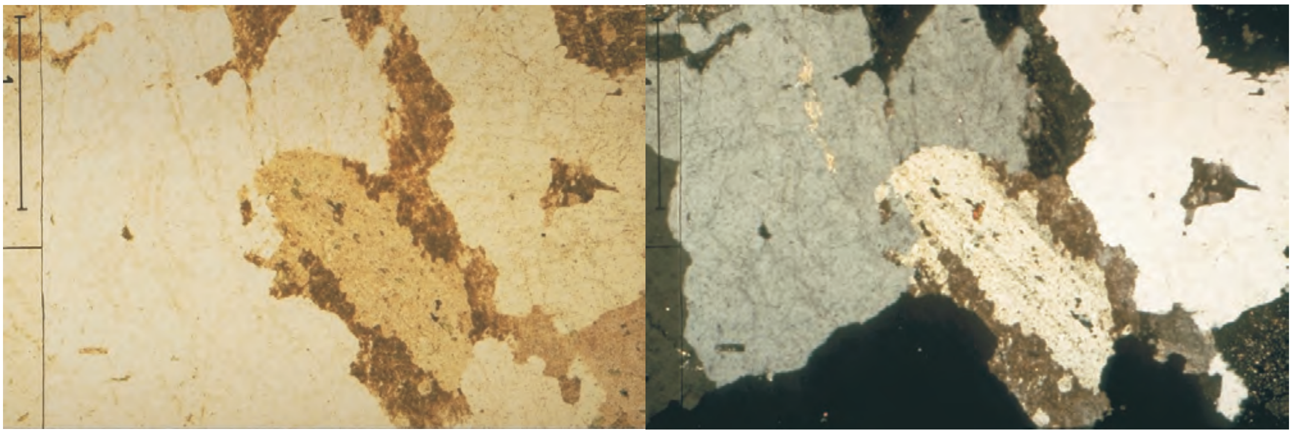


Figure 276. Photomicrograph of K-feldspar rimming and replacing plagioclase, Mount Morgan Trondhjemite. Plane polarised light and crossed nicols. Scale bar is 1mm long.

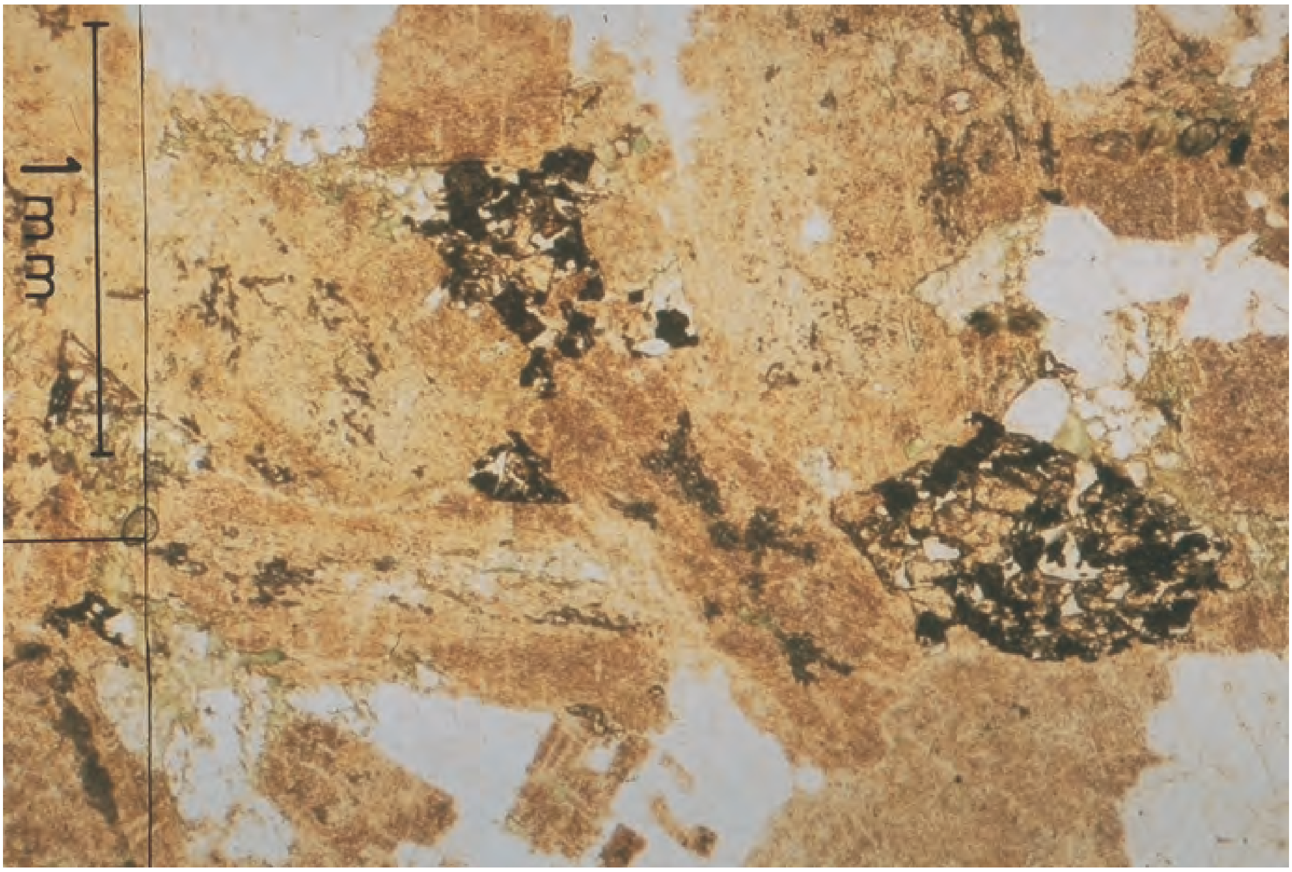


Figure 277. Photomicrograph of sphene crystal replaced by rutile, calcite and quartz, Mount Morgan Trondhjemite. Plane polarised light. Scale bar is 1mm long.

page 454). The presence of pyrite, generally accompanied by calcite, was noted by Kirkegaard & others (1970).

Commonly, joint planes in the trondhjemite have acted as pathways for fluids which have profoundly altered the rock. Alteration has formed intersecting dark grey chlorite-rich zones up to about 15cm wide, usually with a thin quartz vein in the centre (Figure 278). Small rutile crystals are associated with the altered zones (Figure 279).

In most areas, the trondhjemite contains sparse inclusions of more mafic composition. Locally, as in the headwaters of Bull Creek, these inclusions are much larger and more abundant, and some exhibit crenulate margins suggesting that they were in a partly fluid state when incorporated.

Tonalite ($Dgmo_t$) is similar mineralogically to the trondhjemite, but contains less quartz (22–33% of rock), a greater average content of plagioclase of slightly more calcic composition (52–60% of rock), and more deep

Figure 278. Thin quartz veins surrounded by chloritic alteration zones, Mount Morgan Trondhjemite. Bed of Raspberry Creek 13.5km south-south-east of Mount Morgan.

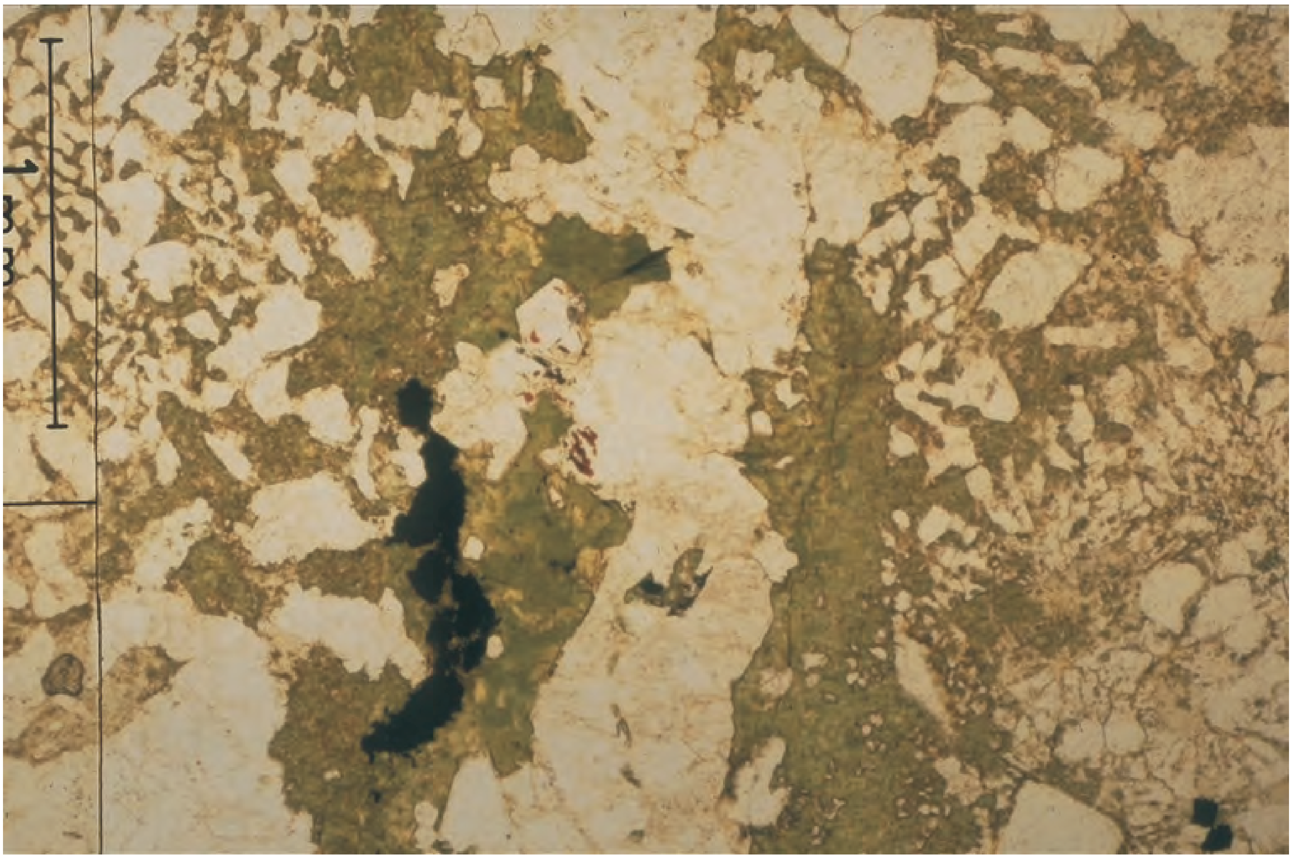


Figure 279. Photomicrograph of quartz vein and surrounding chloritic alteration zone, with secondary opaque mineral and rutile, Mount Morgan Trondhjemite. Plane polarised light. Scale bar is 1mm long.

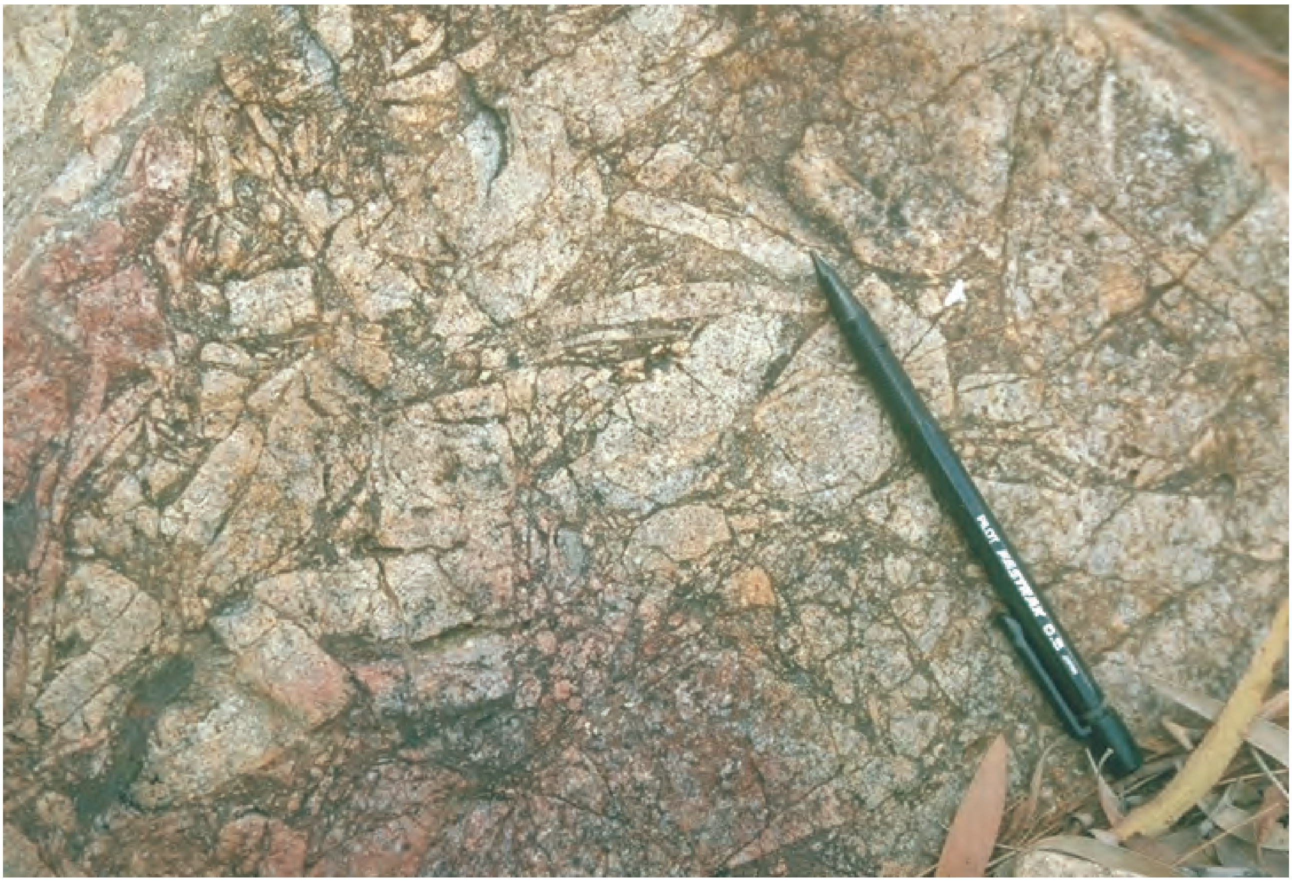


Figure 280. Breccia in the bed of Bull Creek 10km south-east of Mount Morgan, Mount Morgan Trondhjemite. Jigsaw fit textures are displayed in the bottom part of the photograph. Elongate shards of trondhjemite are products of hypogene exfoliation as described by Farmin (1937). The matrix is dominantly chlorite.

brownish-green hornblende (up to about 15% of rock) and biotite (up to 4%). Hornblende occurs as ophitic to subophitic crystals enclosing plagioclase laths, or as anhedral grains and aggregates interstitial to plagioclase. Relict augite cores are preserved in some grains. Plagioclase in the tonalite is not albitised, and the rocks are generally less altered than the trondhjemite. Fine- to medium-grained mafic inclusions are locally abundant (Messenger, 1996).

Quartz diorites and **quartz gabbros (Dgmo$_d$)** have been mapped together because of their small size. The quartz diorites include both hornblende-bearing and pyroxene-bearing varieties, whereas quartz gabbros contain augite and hypersthene and only secondary amphibole.

Breccia (Dgmo$_c$) has been mapped along Bull Creek, where it forms a relatively large body about 1km long and 400m wide (Randall, 1996). It seems probable that much of the breccia has previously been described as part of a granitoid-bearing conglomerate at the base of the Mount Hoopbound Formation (Brown & Wilson, 1965), but it is clearly different. The clasts are exclusively of the host trondhjemite; in general they are angular, with some jigsaw fit textures (Figure 280). Tabular clasts are common, and were produced by decompressive shock (Fletcher, 1977), also called hypogene exfoliation (Farmin, 1937), as described by Baker & others (1986). Maximum clast size is about 30cm. The breccia is clast supported, with a relatively small proportion of dark grey to black chloritic matrix in most outcrops.

Locally the clasts are more rounded, with a greater proportion of matrix, suggesting some degree of milling. In these rocks, a subsequent episode of brecciation is indicated by chloritic veins cutting the clasts (Figure 281).

Applying the model of Baker & others (1986), all these features suggest that the breccia in Bull Creek represents the lower portion of a hypabyssal breccia pipe related to magmatic and hydrothermal activity. The presence of smaller breccia pipes and widespread fracture-controlled chloritic alteration zones in the Mount Morgan Trondhjemite points to a late phase of the trondhjemite itself as the source of the magmatic and hydrothermal fluids. However, no breccia clasts have been found in the basal conglomerates of the Mount

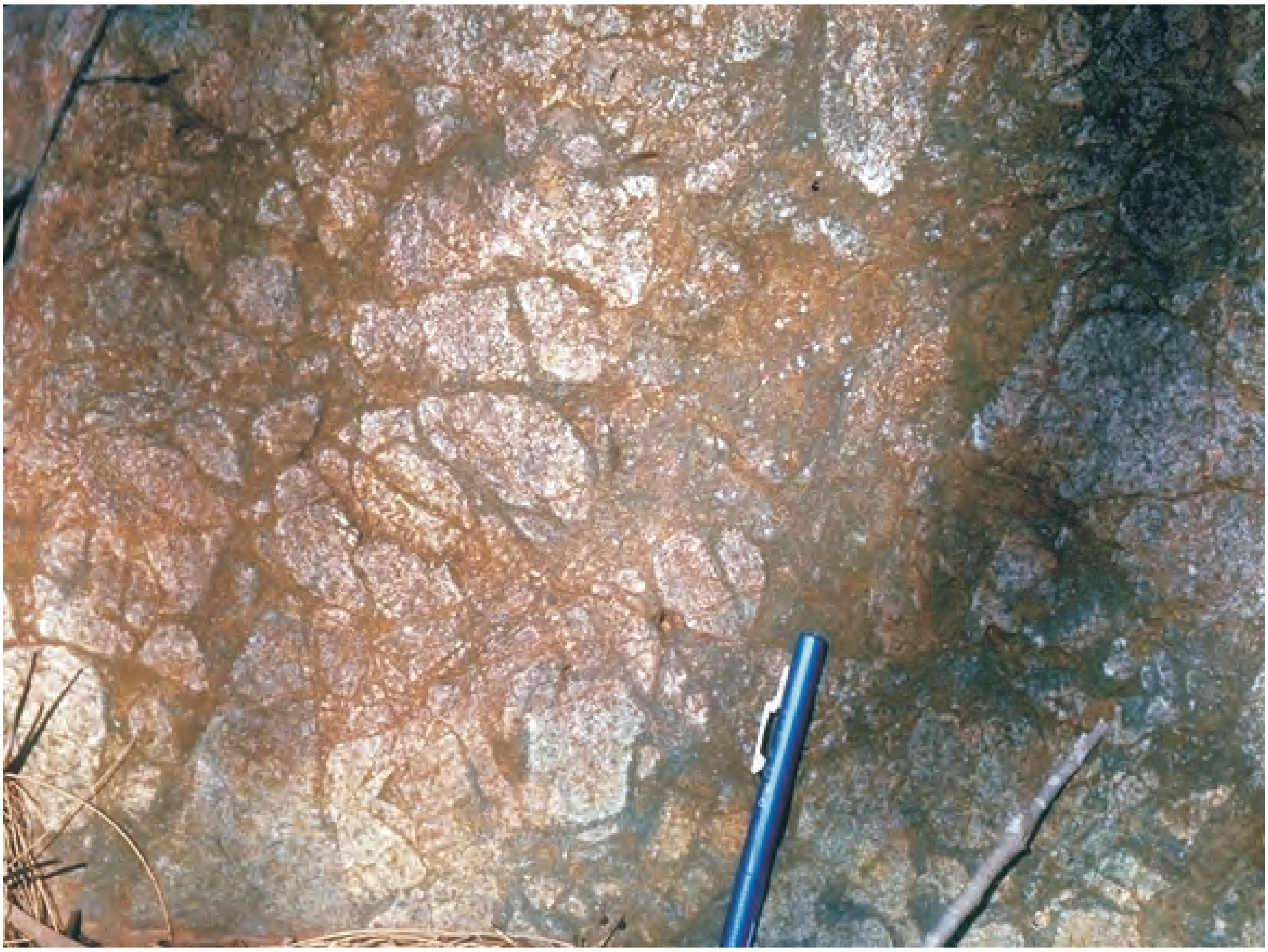


Figure 281. Rounded breccia clasts cut by a second phase of brecciation marked by chloritic veins, Bull Creek, Mount Morgan Trondhjemite

Hoopbound Formation, suggesting that brecciation post-dated the uplift and erosion of the trondhjemite. If so, it is possible that volcanism associated with the Mount Hoopbound Formation was the cause.

Cornelius (1968) mapped 5 small breccia pipes within the Mount Morgan Trondhjemite just south of the Mount Morgan mine, and Arnold & Sillitoe (1989, figure 1) showed 13 small pipes in and adjacent to the intrusion in this area. Descriptions by Arnold & Sillitoe (1989, page 1814) indicate similarities to the breccia along Bull Creek: “A range of brecciation features is present, including *in situ* shattering, sheeted contacts, transported clasts in rock flour matrix, mineral infill of interclast open space, and spheroidal exfoliated fragments. Tonalite is a widespread clast lithology.” They noted that the breccia pipes are preferentially located along the margins of the Mount Morgan Trondhjemite, and that some had been altered to quartz-sericite-pyrite or chlorite-magnetite±pyrite assemblages.

Geochemistry

Chemical analyses of the Mount Morgan Trondhjemite were carried out by Cornelius (1968), Hackman (1995), Randall (1996), Messenger (1996), P.L. Blevin (unpublished) and the present investigation. The Mount Morgan Trondhjemite is a high SiO_2 – low K_2O suite. Most of the rocks conform to the geochemical definition of trondhjemite given by Barker (1979).

Widespread albitisation within the Mount Morgan Trondhjemite, particularly in rocks actually classified as trondhjemite, has undoubtedly increased the Na/Ca ratio, although some of the Ca may have been retained in secondary epidote and possibly in calcite. However, there is no evidence to suggest that the low K_2O content is the result of post-magmatic alteration or metasomatic processes. Microprobe analyses by Messenger (1996) show that fresh plagioclase in trondhjemites has an average K_2O content of 0.24%, placing severe limitations on the amount of K_2O which could be lost by alteration. Some K_2O would have been lost by chloritisation of biotite, but the low modal proportion of biotite and its alteration products (averaging only about 1% for trondhjemite) requires that the actual amount was very small. Also, clear evidence of late stage addition of

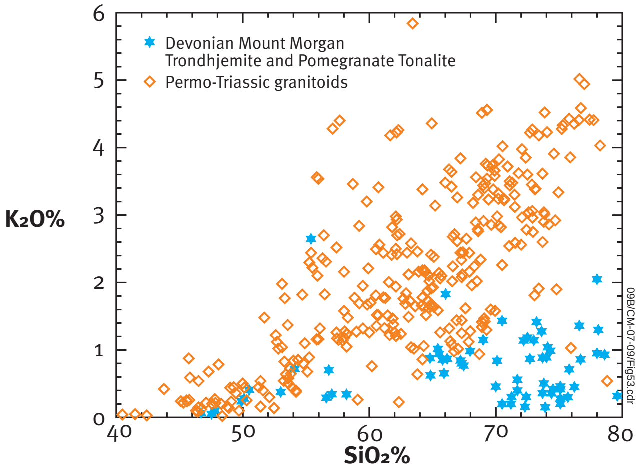


Figure 282. Harker diagram showing the low-K_2O nature of the Mount Morgan Trondhjemite and Pomegranate Tonalite compared with the Permian to Triassic granitoids.

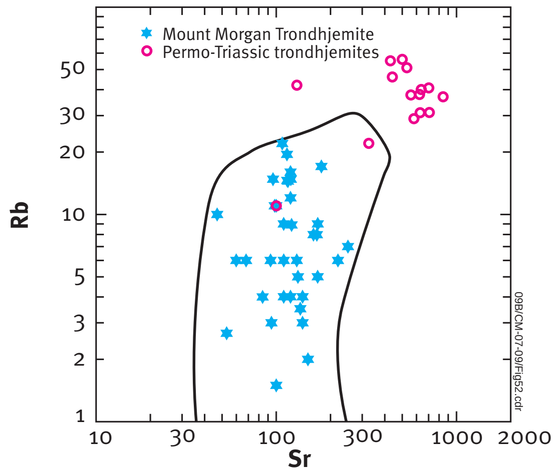


Figure 283. Rb-Sr plot comparing trondhjemites of the Mount Morgan Trondhjemite with trondhjemites from the Permian to Triassic granitoids. The definition of trondhjemite of Barker (1979) is used. Samples from the Mount Morgan Trondhjemite fall within a field defined by 87 samples of trondhjemitic composition from modern volcanic arcs (Bryan, 1979; Gill & Stork, 1979; Johnson & Chappell, 1979; Whalen, 1985; Tsvetkov & Gladkov, 1991; Wharton & others, 1994; Bloomer & others, 1994,1995; Kawate & Arima, 1998; Smith & others, 2003a,b; Haraguchi & others, 2003; Shukuno & others, 2006; Leat & others, 2007) and 41 samples from ancient island arcs (Malpas, 1979; Payne & Strong, 1979; Phelps, 1979; Brouxel & others, 1987).

K_2O, rather than removal, is provided by the local replacement of plagioclase by K-feldspar (Figure 276). There can be no doubt that the compositions give a true indication of the high SiO_2 and low K_2O content of the original magma, shown in Figure 282.

Barker (1979) reviewed the tectonic setting of trondhjemitic-tonalitic suites such as the Mount Morgan Trondhjemite. He concluded that most Phanerozoic trondhjemites were related to subduction, either at continental margins, where they typically occur as smaller masses oceanward of major calc-alkaline batholiths, or in the subvolcanic regions of island arcs.

Various geochemical discriminators have been proposed to differentiate between continental and oceanic settings. Coleman & Donato (1979) showed that oceanic plagiogranites associated with ophiolitic complexes and continental trondhjemites form two widely separated fields on a plot of Rb versus Sr. When trondhjemites and K-poor dacites from modern and ancient island arcs are plotted on this figure, they define a larger field that overlaps that of oceanic plagiogranites, but is separate from continental trondhjemites. Specimens from the Mount Morgan Trondhjemite fall within the island arc field, and are easily distinguished from Permo-Triassic intrusive rocks of trondhjemitic composition (as defined by Barker, 1979) (Figure 283). This plot is potentially unreliable because of the effect of alteration on Rb and Sr values. Arth (1979, page 124) pointed out that “the concentrations of Rb, Sr, and Ba cannot be considered to represent those of the original magmas unless the äO^{18} value of the rock is measured and is in the range 5.2 to 8.9 per mil”. Although samples from the Mount Morgan Trondhjemite do fall within this range (9 samples range from 5.0 to 8.1 per mil, Messenger, 1996, page 170), the plot in Figure 283, while suggestive, cannot be considered conclusive.

Plots of rare earth elements (REE) are more useful in determining tectonic setting because of the relative insensitivity of these elements to alteration. Rocks from oceanic settings, including both ophiolites and island arcs, have “flat or light rare-earth-depleted patterns showing whole rock/chondrite ratios greater than 10 and negative Eu anomalies” (Arth, 1979, page 124). Patterns for samples from the Mount Morgan Trondhjemite suite match analyses of intrusives from New Britain (Murray & Blake, 2005, figure 10), and are also similar to felsic intrusive and volcanic rocks from island arcs such as Kermadec, South Sandwich, Kuriles and Kyushu-Palau (Figure 284). Al_2O_3 content was considered a good discriminator by Arth (1979), because oceanic trondhjemites have less than 14.5% Al_2O_3, whereas most continental trondhjemites contain greater amounts. On his discrimination diagram, which plots Yb against Al_2O_3, silicic rocks of the Mount Morgan Trondhjemite fall in the low-Al_2O_3 oceanic field (Figure 285a). Comparison with Permo-Triassic trondhjemites is possible if Y is used as a proxy for the heavy REE, having the same charge and size as Ho (Smith, 1963, page 392; Whalen, 1985, page 617). Figure 285b shows that samples from the Mount Morgan

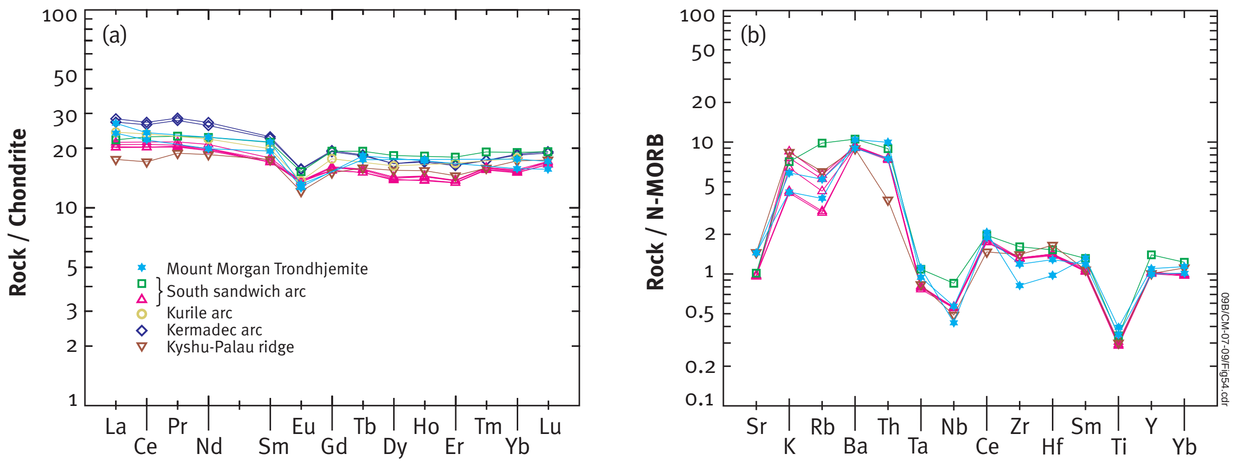


Figure 284. (a) REE and (b) spidergram plots comparing analyses from the Mount Morgan Trondhjemite (P. Blevin, unpublished data) with intrusive and extrusive rocks of trondhjemitic composition from the South Sandwich, Kurile, and Kermadec arcs and the Kyushu-Palau ridge (Tsvetkov & Gladkov, 1991; Leat & others, 2003, 2007; Smith & others, 2003a; Haraguchi & others, 2003). Chondritic REE values are from Sun & McDonough (1989), and N-MORB from Pearce (1983).

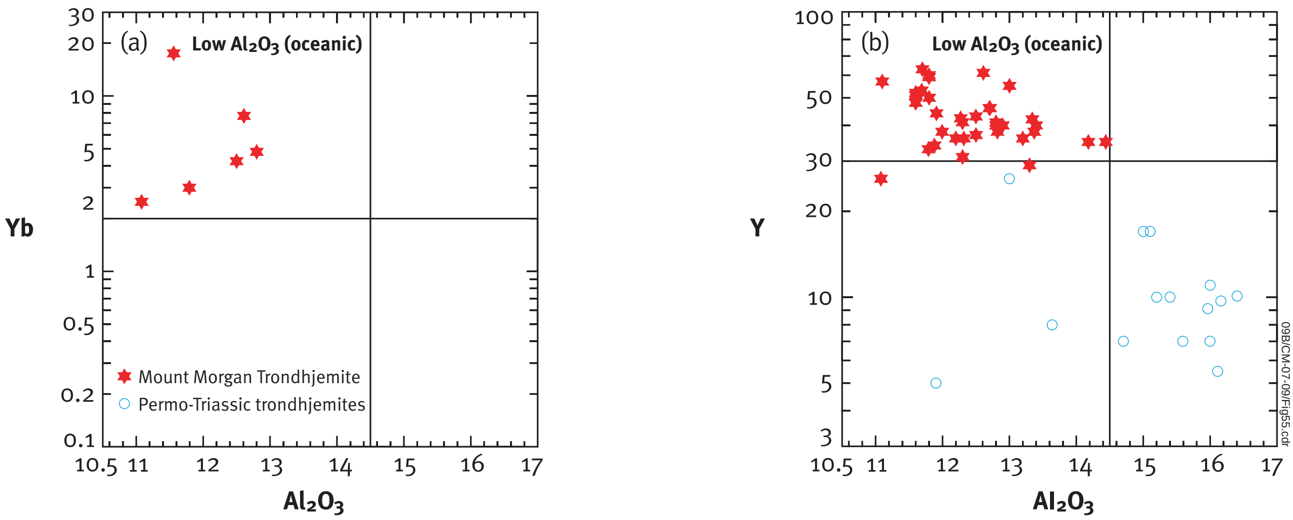


Figure 285. (a) Yb-Al_2O_3 plot of trondhjemites, after Arth (1979). Analyses are from Messenger (1996) and P. Blevin (unpublished data). (b) Y-Al_2O_3 plot of trondhjemites from the Mount Morgan Trondhjemite compared to Permian to Triassic granitoids of trondhjemitic composition. Analyses are from Messenger (1996), and GSQ, P. Blevin, B. Chappell & R. Bultitude (unpublished data). The definition of trondhjemite of Barker (1979) is used.

Trondhjemite all have Al_2O_3 contents less than 14.5%, whereas Permo-Triassic trondhjemites are mainly high in Al_2O_3 and have lower Y contents. These diagrams support the result from Figures 283 and 284. Blevin & others (1996, page 282) previously noted the high Yb/Al_2O_3 ratio of the Mount Morgan Trondhjemite, and the oceanic character of its chemical composition.

An oceanic arc setting is also supported by discriminant diagrams for granites (Pearce & others, 1984). On plots of Nb versus Y and Rb versus Y+Nb (Figure 286), felsic compositions from the Mount Morgan Trondhjemite fall across the boundary between volcanic arc granites and ocean-ridge granites.

It is concluded from this evaluation of the major and trace element chemistry that the Mount Morgan Trondhjemite formed in an island arc setting. Comparable high SiO_2 – low K_2O suites are found today in relatively immature, truly oceanic arcs such as Tonga-Kermadec, Izu-Bonin and South Sandwich. The most widely accepted model for trondhjemite genesis is by either fractional crystallisation or partial melting of a basaltic source (Barker, 1979). Both experimental and field studies suggest that dehydration partial melting is the dominant process (Beard & Lofgren, 1991; Beard, 1995; Nakajima & Arima, 1998; Malpas, 1979; Gill & Stork, 1979; Payne & Strong, 1979; Brouxel & others, 1987; Tamura & Tatsumi, 2002; Leat & others, 2003, 2007; Smith & others, 2003a,b; Shukuno & others, 2006). Modelling by Messenger (1996) indicated that the felsic rocks of the Mount Morgan Trondhjemite were produced by low pressure partial melting of basaltic

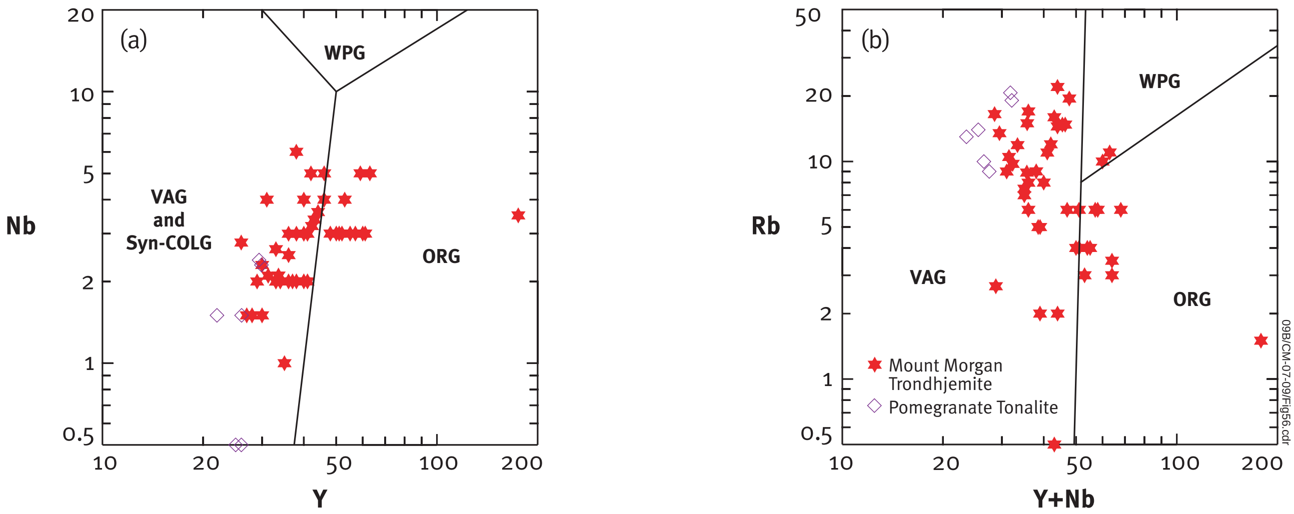


Figure 286. (a) Nb - Y and (b) Rb - Y+Nb discriminant plots of Pearce & others (1984) of samples from the Mount Morgan Trondhjemite and Pomegranate Tonalite. VAG = volcanic arc granites, ORG = ocean ridge granites, WPG = within plate granites, and syn-COLG = syn-collisional granites.

andesites, followed by fractional crystallisation to generate the most silicic melts. For oceanic trondhjemites in general, and the Mount Morgan Trondhjemite in particular, the low Al_2O_3 and Sr contents and negative Eu anomalies require plagioclase as a residual phase for either process, and the flat REE patterns suggest that garnet was not residual (Arth, 1979). Quartz gabbros of the Mount Morgan Trondhjemite have small positive Eu anomalies, suggesting that they may represent plagioclase cumulates (Messenger, 1996, page 159).

The geochemical similarity of the Mount Morgan Trondhjemite and volcanic rocks from the Capella Creek Group (Messenger, 1996) clearly indicates that the intrusion was emplaced into a volcanic pile derived from the same source region and processes.

Age

Webb & McDougall (1968) first determined the Devonian age of the Mount Morgan Trondhjemite by K-Ar dates on hornblendes from 5 samples in the vicinity of the Mount Morgan mine. Their results, recalculated to the revised constants of Steiger & Jäger (1977), are:

GA5372	Quartz diorite east of the Mine Corridor	365±7Ma
GA5339	Quartz diorite west of the Mine Corridor	367±7Ma
GA5370	Quartz diorite west of the Mine Corridor	372±7Ma
GA5369	Trondhjemite north-east of the Mine Corridor	364±7Ma
GA5371	Tonalite west of the Mine Corridor	377±7Ma

A Devonian age was confirmed by Green (1975), who dated 6 hornblende samples from one locality just south of the mine by K-Ar and Ar-Ar methods, obtaining the following results:

A/A49	Tonalite west of the Mine Corridor	352±18Ma
A/A50	Tonalite west of the Mine Corridor	375±22Ma
A/A61	Tonalite west of the Mine Corridor	369±20Ma
A/A88	Quartz diorite west of the Mine Corridor	336±34Ma
Q/A119	Tonalite west of the Mine Corridor	257±8Ma
Q/A120	Tonalite west of the Mine Corridor	267±9Ma

Golding & others (1994b) attempted more precise Ar-Ar dating of the Mount Morgan Trondhjemite using hornblende sample GA5371 of Webb & McDougall (1968), and obtained a plateau-like segment age of 372.1±1.0Ma. They also reported a U-Pb zircon date of 381.0±4.7Ma obtained by C. Perkins from a drillhole sample of tonalite west of the Mine Corridor near the mine site (Golding & others, 1994b; Messenger, 1996).

A sample of trondhjemite from Nine Mile Creek, about 8km south-south-east of Mount Morgan, gave a U-Pb zircon date of 379.9±4.27Ma (Appendix 1). This result, from a single magmatic zircon population, clearly indicates that all phases of the Mount Morgan Trondhjemite were emplaced over a very short time interval.

Relationships

The Mount Morgan Trondhjemite intrudes the Capella Creek Group, of late Middle Devonian (Givetian) age, and is overlain by the Upper Devonian (Frasnian to Famennian) Mount Hoopbound Formation. Coarse conglomerate with a large proportion of clasts derived from the trondhjemite occurs at the base of the Mount Hoopbound Formation at several localities (Brown & Wilson, 1965; Kirkegaard & others, 1970; Messenger & Golding, 1996; Randall, 1996) (Figures 41, 49, 50 and 51), consistent with rapid unroofing and exposure of a high-level intrusion.

Relationships between the different phases in the Mount Morgan Trondhjemite are difficult to determine due to lack of outcrop in critical areas. The presence of locally abundant mafic inclusions in trondhjemite and tonalite, and the lack of the reverse relationship, suggests that the felsic phases were younger. However, crenulate margins on some of the mafic inclusions in Bull Creek are evidence for incorporation before solidification was complete, and for essentially coeval intrusion. This is consistent with observations by Messenger (1996, pages 57–58), who recorded sharp contacts between tonalite and gabbro west of the Mine Corridor, but could not interpret relative ages because of the absence of chilled margins in either phase. Sillitoe (1987) described an intrusive contact between relatively fresh tonalite or quartz diorite and hydrothermally altered trondhjemite to the north of the Mount Morgan mine. This relationship was used by Messenger (1996) to argue that the trondhjemite was significantly older than the less altered tonalite. However, U-Pb zircon dates from trondhjemite and tonalite are indistinguishable.

The restricted range of isotopic dates constitutes strong evidence that the entire plutonic complex was emplaced over a relatively short time interval.

Economic significance

The world-class Mount Morgan Au-Cu deposit (Figure 20) is hosted by a roof pendant of Mount Warner Volcanics within the Mount Morgan Trondhjemite. Most early interpretations of the Mount Morgan gold-copper orebody considered that it was an epigenetic replacement deposit sourced from and related to the Mount Morgan Trondhjemite, particularly the tonalitic phase west of the Mine Corridor. This view was reiterated most recently by Arnold & Sillitoe (1989). Since the publication of articles by Paltridge (1967) and Lawrence (1967), however, the Mount Morgan deposit has generally been considered to be a pipe-like VHMS style deposit (Taube, 1986; Large, 1992).

Reviews by Golding & others (1994a,b) supported a sub-sea floor replacement origin involving fluids derived from a combination of seawater and magmatic sources. They also confirmed an earlier observation by Lawrence (1972) that the Au-Cu orebody had been recrystallised, probably by the Mount Morgan Trondhjemite which must therefore have post-dated the main mineralisation. From his detailed study of the deposit, Messenger (1996) concluded that large-scale hydrothermal circulation responsible for the Au-Cu mineralisation was associated with submarine low-K volcanism, and that the Mount Morgan Trondhjemite was intruded during the waning stages of this circulation. This is impossible given the time interval between eruption of the Mount Warner Volcanics and intrusion of the Mount Morgan Trondhjemite (about 15 million years according to the time scale of Gradstein & others, 2004), and the short-lived nature of hydrothermal systems (Cathles & others, 1997). However, there is clear evidence of ongoing magmatic activity of trondhjemitic composition during part of that time (Hayward & others, 1999).

Precise determinations of Pb isotopes by Ulrich & others (2001, 2002) suggest that the Mount Morgan Trondhjemite was the source of the ore metals, rather than the volcanics that host the orebody. In view of the age relations, it is more likely that the source was a similar but earlier magma chamber of trondhjemitic composition.

Murray & Blake (2005) drew an analogy between the Mount Morgan orebody and a gold-rich sulphide deposit currently forming on the floor of the Myojin Knoll caldera in the Izu-Bonin arc (Fiske & others, 2001). This idea shares some common features with the model proposed by Taube (1986) involving cauldron subsidence and resurgence.

Whatever model is accepted for Mount Morgan, either a composite model as suggested by Ulrich & others (2001, 2002) or a VHMS, the association with a long-lived felsic volcanic-plutonic centre in a submarine island arc environment seems indisputable. Such centres should therefore be regarded as exploration targets for Mount Morgan-style deposits.

Figure 287. Distribution of the Pomegranate Tonalite

Pomegranate Tonalite (Dgpo)
(P.R. Blake)

Introduction

The Pomegranate Tonalite is a newly defined unit that occurs towards the south-western corner of the Bajool 1:100 000 Sheet area.

Distribution

The Pomegranate Tonalite is exposed in only a small area and occurs about 1.5km south of the Pomegranate homestead (Figure 287).

Derivation of name

This intrusion is named after Pomegranate Creek, which flows through the intrusion.

Type area

The hill at GR249000 7355400 on the Bajool 1:100 000 Sheet area is the type area. A sample from this location has been geochemically analysed and isotopically dated.

Topographic expression

This intrusion forms low hills.

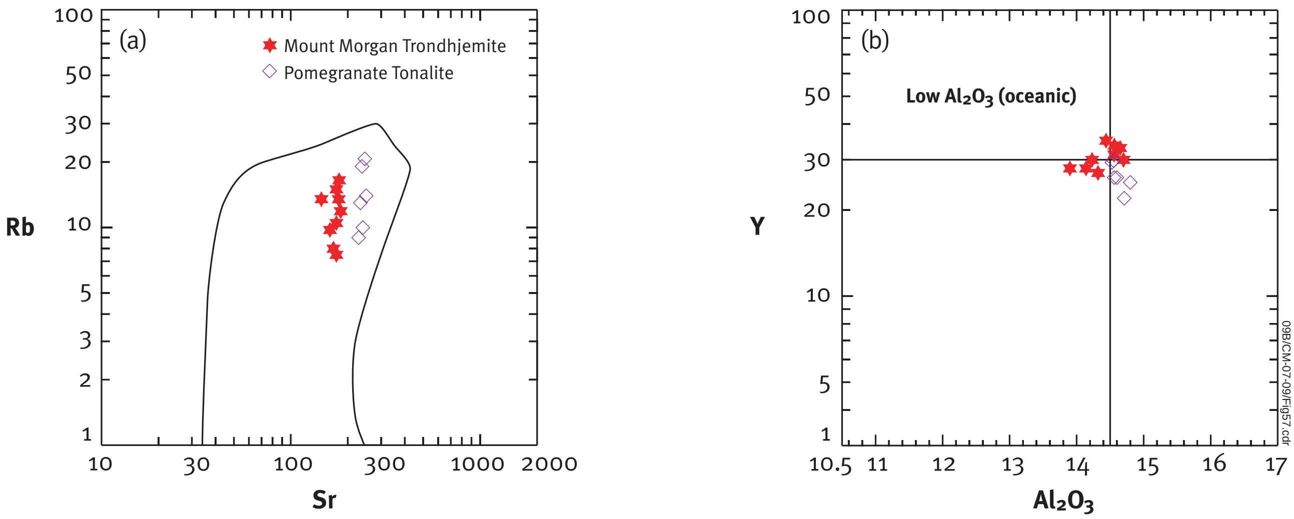


Figure 288. (a) Rb-Sr and (b) Y-Al_2O_3 plots comparing samples of the Mount Morgan Trondhjemite with samples from the Pomegranate Tonalite. SiO_2 contents range from 65.4 to 68%. Analyses from Messenger (1996) and GSQ and B. Chappell & R. Bultitude (unpublished data).

Geophysical expression

The Pomegranate Tonalite gives a very low response in all three channels of the airborne radiometric surveys. On airborne magnetic images the Pomegranate Tonalite possesses a low magnetic signature.

Lithology and petrography

The Pomegranate Tonalite is medium-grained and equigranular. The rock is dominantly composed of plagioclase (60–70%), with common quartz (~10%) and mafics (20–30%) (usually augite and hornblende, but rarely biotite). K-feldspar is an accessory mineral.

Geochemistry

The Pomegranate Tonalite is similar in chemistry to the Mount Morgan Trondhjemite. Six analysed samples have a very limited range of SiO_2 contents, from 65.4 to 68%. These are compared with rocks of similar SiO_2 content from the Mount Morgan Trondhjemite in Figure 288. The plots show that the Pomegranate Tonalite has slightly higher Al_2O_3, Sr and Rb and marginally lower Y. However, the Pomegranate Tonalite still falls well within the field of felsic rocks from modern and ancient island arcs on the graph of Sr versus Rb. It falls in the volcanic arc granite field on the discriminant diagrams of Pearce & others (1984), amongst samples from the Mount Morgan Trondhjemite (Figure 286).

Age

Hornblende from a sample of the intrusion gave K-Ar dates of 415±3 and 417±3Ma (Appendix 1). Because these dates are older than the host rocks, a U-Pb date was determined on zircon. This gave an acceptable age of 369.0±4.2Ma, or Late Devonian (Appendix 1).

Relationships

The Pomegranate Tonalite intrudes the Raspberry Creek, Mount Hoopbound, and Balaclava Formations. The Balaclava Formation includes some rocks with Early Carboniferous fossils, so the tonalite must have been intruded while the unit was being deposited.

Associated mineralisation

No mineralisation is known to be associated with the the Pomegranate Tonalite.

Figure 289. Distribution of the Kyle Mohr Igneous Complex

PERMIAN

Kyle Mohr Igneous Complex (Pgkm)

(C.G. Murray)

Introduction

This pluton was mapped and described as the Kyle Mohr Granodiorite by Kirkegaard & others (1970). They noted the presence of marginal zones with abundant mafic xenoliths, but did not attempt to subdivide the intrusion.

Distribution

The Kyle Mohr Igneous Complex is an oval-shaped body 12km long and 6km wide, orientated in a north-west direction. It is centred 15km west-south-west of Mount Morgan, and covers the area between Top Rock Creek in the north, Sandy Creek to the west, and Oaky Creek to the east (Figure 289). The intrusion consists of a central granitic core 8.5km long and up to 4km across surrounded by an almost complete ring of dioritic to gabbroic rocks up to 2km wide (Figure 289).

Derivation of name

The name is from Kyle Mohr holding in the south-western part of the pluton.

Type area

No type area was designated by Kirkegaard & others (1970). Typical outcrops of both main phases of the intrusion can be seen along the headwaters of Oaky Creek.

Topographic expression

The intrusion forms an elevated area of rounded hills, with streams flowing outwards from it in all directions. A set of east-north-east trending joints is prominent on aerial photographs as deeply incised valleys. Basalt dykes intruded along the joints are more readily eroded than the surrounding granitic rocks. One of the joints is co-linear with a fault to the east of the pluton.

Geophysical expression

Images derived from airborne geophysical data can be used to divide the Kyle Mohr Igneous Complex into two contrasting segments, one surrounding the other. The almost complete outer ring of dioritic to gabbroic rocks has a very low radiometric response, and a moderate magnetic signature. The central granitic core has a moderate, potassium-rich radiometric signature, and a lower magnetic response. The lower magnetic response is consistent with its measured magnetic susceptibility of 1130 x 10^{-5} SI units compared to 2020 x 10^{-5} SI units for the diorite and gabbro.

A small area of aplite at the north-west corner of the intrusion has extremely low magnetic susceptibility (9 x 10^{-5} SI units) and also a low radiometric response.

Lithology and petrography

The most common rock type in the outer ring is a grey to dark grey, fine to coarse-grained, mafic-rich gabbro that has been extensively altered and in some cases hornfelsed by the later granitic phase. Typically these rocks consist of subhedral to euhedral plagioclase laths (labradorite) enclosed by ophitic to subophitic augite crystals up to 8mm across that are rimmed and partly to completely replaced by brown or green-brown hornblende. Some augite grains are replaced by pale green secondary actinolitic amphibole. Biotite is common, but is largely altered to chlorite. Rare hypersthene occurs as small grains usually enclosed by augite. The form of the opaque minerals suggests that both magnetite and ilmenite are present. The proportion of mafic minerals ranges from 30 to 50%. Secondary minerals include prehnite, chlorite and epidote, and apatite is a ubiquitous accessory.

Many of the gabbroic rocks show evidence of hybridisation or hornfelsing by the later granite, including variations in grain size even within a single thin section. Some contain relatively large quartz grains that appear to be associated with the introduction of late stage interstitial sphene. The form of the augite changes from ophitic or subophitic to small stumpy prisms. Zircon is present in many of these hybrid rocks, and one specimen contains relatively large tourmaline crystals.

The most abundant rock type in the central core is a leucocratic biotite-hornblende granite. Plagioclase is always the dominant mineral, comprising from 40 to 60% of the rock. It forms subhedral laths to anhedral grains up to 5mm long, zoned from cores of andesine to rims of oligoclase. K-feldspar ranges from 10 to 15%, and occurs as rims on plagioclase, as interstitial grains, and as micrographic intergrowths with quartz. Much of it is microperthite. Anhedral quartz grains make up from 25 to 35% of the rock. Mafic minerals rarely constitute more than 10% of the rock. Green to brown hornblende is much more abundant than biotite, which is largely altered to chlorite. Accessory minerals include opaques, apatite, zircon, and late stage interstitial sphene usually associated with quartz.

Locally the K-feldspar content rises to more than 25%, the proportion of biotite increases, and the rock becomes a biotite-hornblende granite. Along Top Rock Creek the rocks are very leucocratic, and the percentage of K-feldspar decreases, producing trondhjemitic compositions.

Granitic dykes intrude the outer gabbroic phase around the southern margin of the pluton. These dykes are identical in composition to the central core.

Geochemistry

The Kyle Mohr Igneous Complex is bimodal, consisting of an outer gabbro ring and a granite core. Its chemistry is similar to that of the Rookwood Volcanics, and it is suggested that they were comagmatic (Figures 132 to 135). The geochemistry is distinctly different from the typical Permo-Triassic granitoids.

Only a single sample of gabbro was analysed. This falls within the range of variation of basalts from the Rookwood Volcanics (*cf.* Figure 129). The similarity is confirmed on a plot of Ti versus V (Shervais, 1982), on which the gabbro falls within the group of Rookwood Volcanics basalts, quite separate from Permo-Triassic gabbros (Figure 134).

The granite core has similar geochemistry to the rhyolites of the Rookwood Volcanics, except that the rhyolites are slightly more fractionated and, judging by the relative Zr and Nb values, contain more zircon (Figures 132 and 135). On the Y-Nb discriminant plot of Pearce & others (1984), the granitoids lie along a possible fractionation trend within the ocean ridge granite field that is co-linear with that of rhyolites from the Rookwood Volcanics (Figure 133a). On the Y+Nb versus Rb plot, they are grouped with less altered rhyolites in the volcanic arc granite to within plate granite field (Figure 133b). The granites and volcanics have some A-type characteristics, and fall along a possible fractionation trend on discriminant diagrams proposed by Whalen & others (1987) (Figure 135).

Age

A sample of hornblende granite from the south-western margin of the pluton was dated by U-Pb on zircon as 270.6±3.7Ma, or mid-Permian (Appendix 1).

Relationships

The Kyle Mohr Igneous Complex intrudes the Balaclava Formation in the south and the Youlambie Conglomerate in the north. A tight syncline in the Youlambie Conglomerate parallels the northern contact, and appears to be a rim syncline generated during emplacement.

The Kyle Mohr Igneous Complex conforms to the description of nested plutons, which are usually zoned with a more mafic outer rim that was still highly plastic or in a viscous state when intruded by the felsic core (Bouchez & Diot, 1990). In such a case, there could not have been a significant time gap between emplacement of mafic and felsic magmas. Contacts between the inner granite and gabbro are generally sharp (Figure 290), but locally are gradational and fuzzy (Figure 291), suggesting that the gabbro was not entirely solid when the felsic rocks were intruded. The relationship between mafic and felsic rocks therefore has some of the characteristics of net-veined intrusions. The pluton also can be classified as a ring complex, because on the southern side the gabbro is intruded by granodiorite ring dykes.

Figure 290. Gabbro intruded by granite veins and patches, showing sharp contacts, Kyle Mohr Igneous Complex. Headwaters of Oaky Creek 13km south-west of Mount Morgan.

Figure 291. Fuzzy contacts between gabbro and intrusive granite, headwaters of Oaky Creek, Kyle Mohr Igneous Complex.

Associated mineralisation

The Dee copper mine and associated workings mined a series of sulphide-bearing veins parallel and adjacent to the south-eastern margin of the pluton. The intrusion has been explored for gold.

Miriam Vale Granodiorite (PRgmv)

(C.G. Murray)

Introduction

The name Miriam Vale Granodiorite was introduced by Dear & others (1971) for a batholith exposed mainly in the Bundaberg 1:250 000 sheet area, but extending westwards to the foothills of the Many Peaks Range. They described the batholith as mainly of granodioritic composition, grading to tonalite and diorite. In fact, it is a composite batholith with individual plutons ranging from gabbro to granite. The Miriam Vale Granodiorite was not examined in detail during the present investigation, and it has not been subdivided on the Calliope 1:100 000 sheet area except for the geophysically distinct Bororen Tonalite. It extends to the east, where it has been mapped as the Jackass Gabbro on the Miriam Vale 1:100 000 sheet area.

Distribution

The Miriam Vale Granodiorite extends from Hummock Hill Island in the north to Colosseum Creek in the south, and forms a continuous belt 50km long and up to 15km wide to the eastern limit of the Calliope 1:100 000 sheet (Figure 292).

Derivation of name

The name is from the township of Miriam Vale on the Bruce Highway in the Bundaberg 1:250 000 sheet area.

Figure 292. Distribution of the Miriam Vale Granodiorite

Type area

Dear & others (1971) did not nominate a type area. Available information suggests that individual areas are unlikely to contain the full range of rock types within the batholith, and that there is no single representative lithology.

Topographic expression

In general the Miriam Vale Granodiorite forms low, gently undulating country that has been cleared for grazing, and outcrop is sparse. Local relief of up to almost 200m is provided by hills and ridges south and west of Iveragh as far as Inkerman Peak.

Geophysical expression

Overall, the radiometric response from the Miriam Vale Granodiorite is low, and it gives dark colours on images generated from airborne data. Local areas of moderate radiometric signature, ranging from pink to red, are present on Hummock Hill Island, south of Tannum Sands, along hills and ridges south-west from Iveragh, and along the eastern foothills of the Many Peaks Range. Some of these areas correspond to topographic highs, suggesting that depth of weathering may have an effect on the radiometric response.

Surprisingly, the magnetic signature of the intrusion as indicated by the airborne data is uniformly low, even from areas of gabbroic or dioritic composition.

Lithology and petrography

Examination of thin sections prepared during the initial mapping program carried out by Dear & others (1971) gives an indication of the range and distribution of rock types.

A sample from the Bruce Highway 3km south of the turnoff to Tannum Sands is a medium-grained hornblende-biotite syenogranite. Microperthitic K-feldspar is by far the dominant constituent, comprising about half the rock as relatively large poikilitic grains to 4mm across that enclose plagioclase, biotite and hornblende. Plagioclase and quartz are present in approximately equal proportions, each making up about 20% of the rock. Plagioclase forms small, strongly zoned subhedral to euhedral laths, whereas quartz occurs as anhedral grains. Mafic minerals total about 10%, with biotite more abundant than green hornblende. Accessories include opaques, sphene, apatite and zircon. This rock type may be typical of the area of moderate (pink) radiometric response south of Tannum Sands. A further 3km to the south, the rock is a hornblende quartz microdiorite, consisting of about 60% plagioclase laths, 30% green hornblende prisms, and 10% interstitial quartz grains. Minor opaques and interstitial sphene and accessory apatite are also present. Similar dioritic rocks probably occur in the area east of the Bruce Highway north of Iveragh, characterised by dark green to black colours on radiometric images.

In the area to the west of Mount Coulston, extending south from Inkerman Peak, mafic compositions ranging from gabbro to quartz diorite are most common. Subhedral to euhedral plagioclase laths are dominant, comprising from 60 to 75% of the rocks. Core compositions range from calcic labradorite to calcic andesine. The content of mafic minerals ranges from 25 to 40%. Augite is most abundant, as anhedral to sub-ophitic grains partly replaced in some specimens by aggregates and needles of green actinolitic amphibole. It is always rimmed by hornblende, mainly dark brown interstitial to poikilitic grains, but also green. Up to 5% olivine and minor biotite are present locally, and opaque grains are ubiquitous. The more dioritic samples contain about 5% quartz as small interstitial grains, as well as accessory apatite and secondary sphene. Felsic rocks in this area are tonalitic in composition, consisting of 55 to 60% subhedral to euhedral plagioclase laths with cores of sodic andesine, 25 to 30% interstitial quartz grains, 15% mafic minerals, and a little K-feldspar. Dark brown biotite is more abundant than green hornblende. Minor opaques and accessory zircon are present. A sample from the south-western side of Mount Coulston has been recrystallised, and biotite occurs as fine-grained aggregates. Similar tonalitic rocks, but with relict augite partly replaced by hornblende, and rare relatively large poikilitic K-feldspar grains, crop out just to the west of Iveragh.

Around the upper reaches of Colosseum Creek, biotite-hornblende granodiorite and hornblende quartz diorite appear to be the main rock types. The granodiorite consists of 45 to 50% plagioclase (subhedral zoned laths with cores of sodic andesine), 20 to 25% anhedral quartz grains, 15 to 20% small interstitial K-feldspar crystals, and 10 to 15% mafic minerals. Green to brown hornblende with extremely rare relict augite cores is more abundant than biotite. Minor opaque grains and accessory apatite and zircon are present. Metamorphism of one sample has produced aggregates of small green hornblende prisms, and thin sphene rims around opaque grains. Quartz diorite contains 55 to 60% plagioclase (zoned subhedral to euhedral laths with cores of calcic andesine), about 5% quartz, up to 5% interstitial K-feldspar, and 35% mafics. Brown to green-brown hornblende is by far the most abundant mafic mineral as ophitic grains enclosing plagioclase. Minor biotite and opaques are present, and accessory apatite and sphene.

Geochemistry

On Harker diagrams, the Miriam Vale Granodiorite, together with the Castletower Granite, is characterised by high TiO_2, total alkalis, Ce, Y and Zr, high K_2O and Rb at high SiO_2 contents, and low Sr (Figure 293). The group is transitional from the volcanic arc granite field to the within plate granite field on discriminant diagrams of Pearce & others (1984) (Figure 294). Silica-rich compositions tend towards A-type granites. Some geochemical features of the Miriam Vale Granodiorite and Castletower Granite are similar to the Upper Triassic Winterbourne and Coulston Volcanics, and to the Robert and Molangul Granites (*cf.* Figures 174, 294 and 301). On the Ti/100 versus V discriminant diagram of Shervais (1982), mafic rocks of the Miriam Vale Granodiorite plot separately from the Permo-Triassic samples (Figure 295), in a similar position to the single gabbro from the Kyle Mohr Igneous Complex (Figure 134).

Age

No samples from the Miriam Vale Granodiorite in the Calliope 1:100 000 sheet area have been dated. The unit is assigned a Permian age because of its relationship with the Castletower Granite.

Relationships

The Miriam Vale Granodiorite intrudes the Wandilla and Shoalwater Formations, and is overlain by the Coulston Volcanics. To the east on the Miriam Vale 1:100 000 sheet, the intrusive mass is poorly exposed and has been named the Jackass Gabbro. It is intruded by the Castletower Granite, the Norton and Bororen Tonalites, and by an unnamed intrusion along Oaky Creek.

Associated mineralisation

No mineralisation is known to be associated with the Miriam Vale Granodiorite in the Calliope 1:100 000 sheet area.

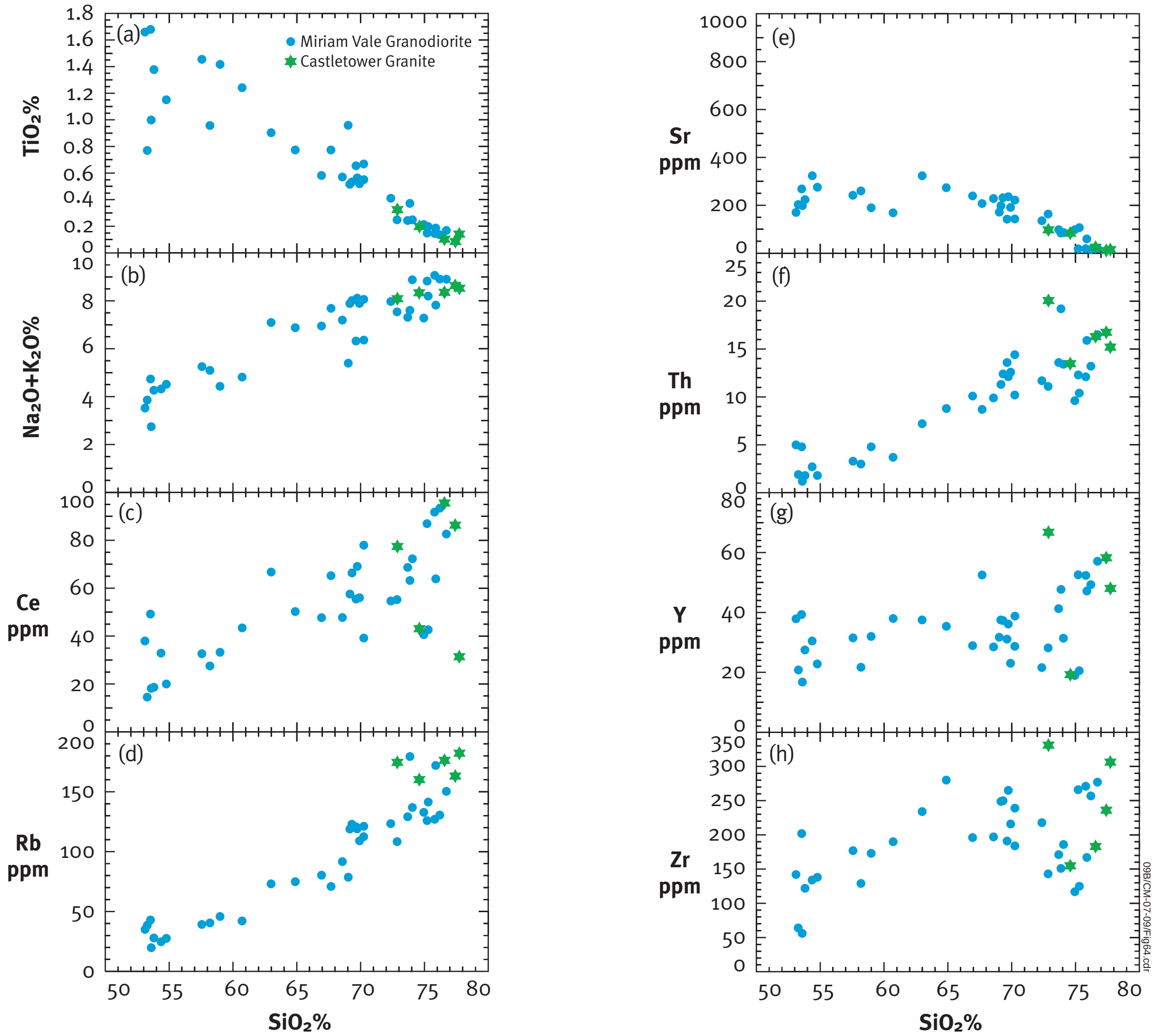


Figure 293. Harker diagrams of the Miriam Vale Granodiorite and Castletower Granite (HPGL files 71A-74A). Analyses from B. Chappell & R. Bultitude (unpublished data).

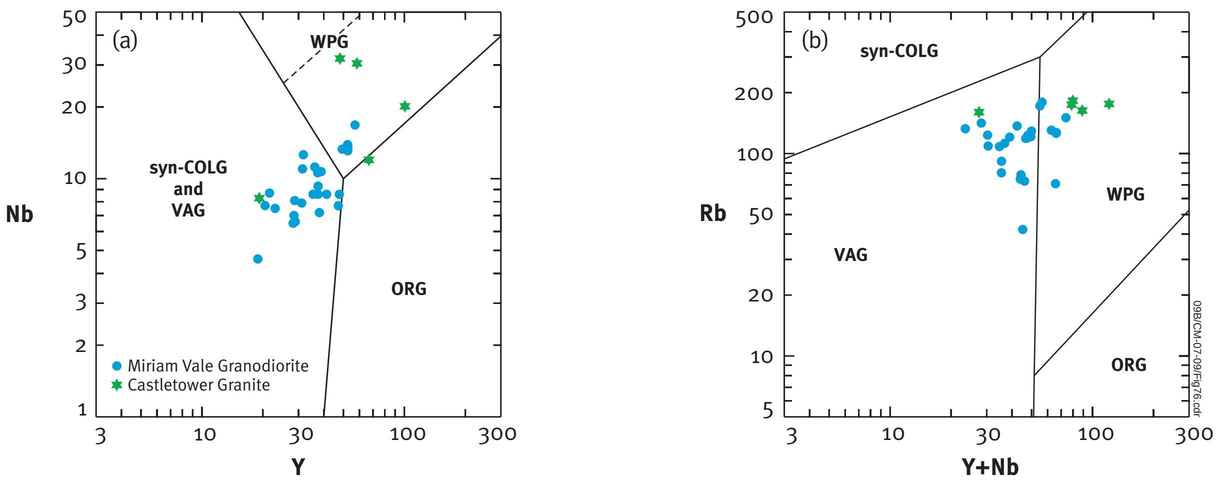


Figure 294. (a) Nb - Y and (b) Rb - Y+Nb discriminant plots of Pearce & others (1984) of felsic samples from the Miriam Vale Granodiorite and the Castletower Granite. VAG = volcanic arc granites, ORG = ocean ridge granites, WPG = within plate granites, and syn-COLG = syn-collisional granites. Analyses from B. Chappell & R. Bultitude (unpublished data).

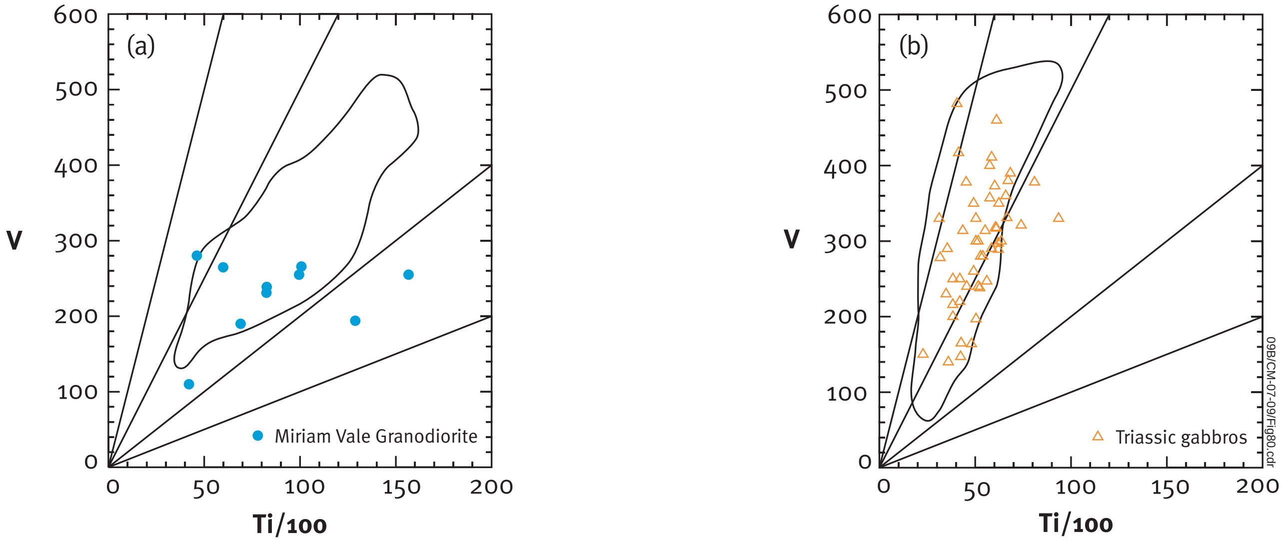


Figure 295. Ti/100 versus V plot (Shervais, 1982) of (a) mafic rocks of the Miriam Vale Granodiorite, which mainly fall in the back-arc basin basalt-MORB field (BAB and MORB), and (b) Permo-Triassic gabbros, which fall in the arc tholeiite field. All samples have less than 55% SiO_2. Although this diagram was designed specifically to discriminate basalts in ophiolite complexes, it also distinguishes volcanic arc and back-arc samples in continental environments.

Castletower Granite (PRgct)

(C.G. Murray)

Introduction

The Castletower Granite was mapped and briefly described by Dear & others (1971).

Distribution

The intrusion forms a number of separate plutons in a belt 45km long and up to 10km wide along the Many Peaks Range from Lake Awoonga in the north to the headwaters of the Boyne River in the south (Figure 296).

Derivation of name

The name is from Mount Castletower in the centre of the northernmost pluton.

Type area

Dear & others (1971) did not nominate a type area. Typical outcrops can be seen along the road through Blackmans Gap from Littlemore to Miriam Vale.

Topographic expression

The intrusion forms high, extremely rugged country, much of it forested, with bare rock faces common in the north around Mount Castletower and Mount Stanley. Locally, joint sets are marked by prominent linear valleys that are deeply incised.

Geophysical expression

The Castletower Granite gives a very strong radiometric response in all channels, and is almost all pure white on images generated from airborne data. Overall, it has a weak magnetic signature, similar to that of the surrounding sedimentary units. The northern pluton around Mount Castletower is slightly more magnetic than the outcrops to the south.

Lithology and petrography

The dominant and almost sole rock type is a pink to buff, medium-grained, equigranular, leucocratic syenogranite. Quartz and microperthitic K-feldspar are present in approximately equal proportions (each about 40% or more of the rock) as equant anhedral grains, and in micrographic intergrowths. Plagioclase

Figure 296. Distribution of the Castletower Granite

occurs in some specimens as small, slightly zoned subhedral laths with calcic oligoclase cores that are partly altered. It comprises up to 15% of some rocks, and contains myrmekitic intergrowths at contacts with K-feldspar grains. Mafics include green to very dark indigo alkali amphibole and/or dark brown biotite flakes, which make up 5% of the rock. Accessories include opaques, zircon, tourmaline, allanite and fluorite.

Geochemistry

The Castletower Granite is similar geochemically to the more siliceous compositions in the Miriam Vale Granodiorite (Figure 293). As shown in Figure 294, the Castletower Granite falls in the within plate field on discriminant diagrams of Pearce & others (1984). It tends towards A-type chemistry. Together with the more silicic granites of the Miriam Vale Granodiorite, it has similar geochemistry to the Late Triassic Robert and Molangul Granites and the Winterbourne and Coulston Volcanics (cf. Figures 174, 294 and 301).

Age

A sample from Blackmans Gap has given a mid-Permian U-Pb date on zircon, similar to that of the Kyle Mohr Igneous Complex (A. Cross, personal communication, 2009). However, it is possible that the Castletower Granite as mapped is a composite intrusion which may include phases of different age.

Stratigraphic relationships

The Castletower Granite intrudes the Calliope beds and Wandilla Formation. It is assumed to intrude the Miriam Vale Granodiorite to the east, although contact relationships have not been observed. The nature of its contact with the Bobby Volcanics is also uncertain, and may indicate that the mass mapped as the Castletower Granite is composite. The granite is intruded by the Norton Tonalite, by a similar unnamed pluton occupying the drainage area of Oaky Creek along the Blackmans Gap road, and by a sill-like gabbro east of Norton Creek.

Economic significance

The Castletower Granite is largely devoid of mineralisation, but it does host the deposits of the Mount Jacob Goldfield. The source of this mineralisation is considered to be the unnamed tonalitic pluton intruding the Castletower Granite along Oaky Creek.

PERMIAN–TRIASSIC

Permian–Triassic intrusions

(C.G. Murray)

Introduction

Most granitic intrusions in the Yarrol project area are of Late Permian to Late Triassic age. They range from multiphase batholiths to small plutons, and have very different topographic expression depending on their composition and texture. The dominant composition is granodiorite, but a full range from gabbro to syenogranite is present. A feature is the presence of several layered gabbros. Many of the plutons are zoned, typically with a mafic rim intruded by a felsic core, giving them a bimodal composition. A subduction-related origin is favoured, with a transition to extensional magmatism in the Late Triassic due to slab rollback. Associated mineralisation consists mainly of small vein-hosted gold and minor copper deposits at the margins of the intrusive bodies or in the adjacent contry rocks. Some aspects of the intrusives have been reviewed by Gust & others (1993, 1996) and Murray (1998, 2003).

Distribution

Granites of this age are part of a north-north-west trending belt which extends from Brisbane to north of Rockhampton. In the Yarrol project area, they mainly intrude Upper Devonian and Carboniferous sedimentary rocks west of the Yarrol Fault (Figure 297). They range from small composite batholiths to zoned plutons.

The batholiths are oriented north-north-east, at right angles to the dominant north-north-west structural trend. From north to south these are the Ridgelands Granodiorite, Bouldercombe Igneous Complex, Galloway Plains Igneous Complex, and Glassford Igneous Complex. Sizes range from about 25km long and 6km wide (Ridgelands) to 40km long and 15km wide (Bouldercombe). The batholiths are from 40 to 50kms apart. This spacing could reflect crustal thickness (Rickard & Ward, 1981), but Cruden (2005) suggested that the size and spacing of plutons is controlled by the degree of melting and the mechanics of melt withdrawal in the lower crust. The north-north-east orientation may reflect faults or other structures that controlled emplacement. Smaller plutons are irregularly distributed amongst the batholiths.

Shape

Cruden (2005) postulated that plutons in a continental magmatic arc are tabular in form, with parallel roofs and floors and steep sides. This appears to be the case in the Yarrol project area. Aeromagnetic patterns due to plutons do not extend outside their outcrop area, indicating that the walls are steep. Gravity suggests that the intrusions are a few kilometres in depth, with possible deeper regions associated with feeder zones. It has been suggested that uplift and erosion levels tend to stabilise close to the roof of plutons (Cruden, 2005), but the lack of any remnants of pluton roofs shows that erosion has continued to a greater depth.

Topographic expression

Biotite-rich granodiorites disintegrate on weathering due to expansion of biotite to hydrobiotite and vermiculite (biotite-induced grussification of Isherwood & Street, 1976), and erode rapidly to form topographic basins (*e.g.* the Bundaleer and Bocoolima Granodiorites). In cases where biotite granodiorite or tonalite forms hilly topography, this is considered to be the result of resistant cover rocks or intrusion by a more resistant felsic phase (*e.g.* the Rocky Point Granodiorite). In contrast, hornblende-bearing monzogranites or syenogranites with micrographic groundmass are resistant to weathering and form prominent peaks (*e.g.* the Redshirt and Robert Granites and the felsic core of the Mount Seaview Igneous Complex) (Figures 298 and 299).

Geophysical expression

The geophysical expression of the felsic intrusives is variable. On images derived from airborne radiometric data, they give red to white colours, reflecting the relative abundance of thorium, uranium and particularly

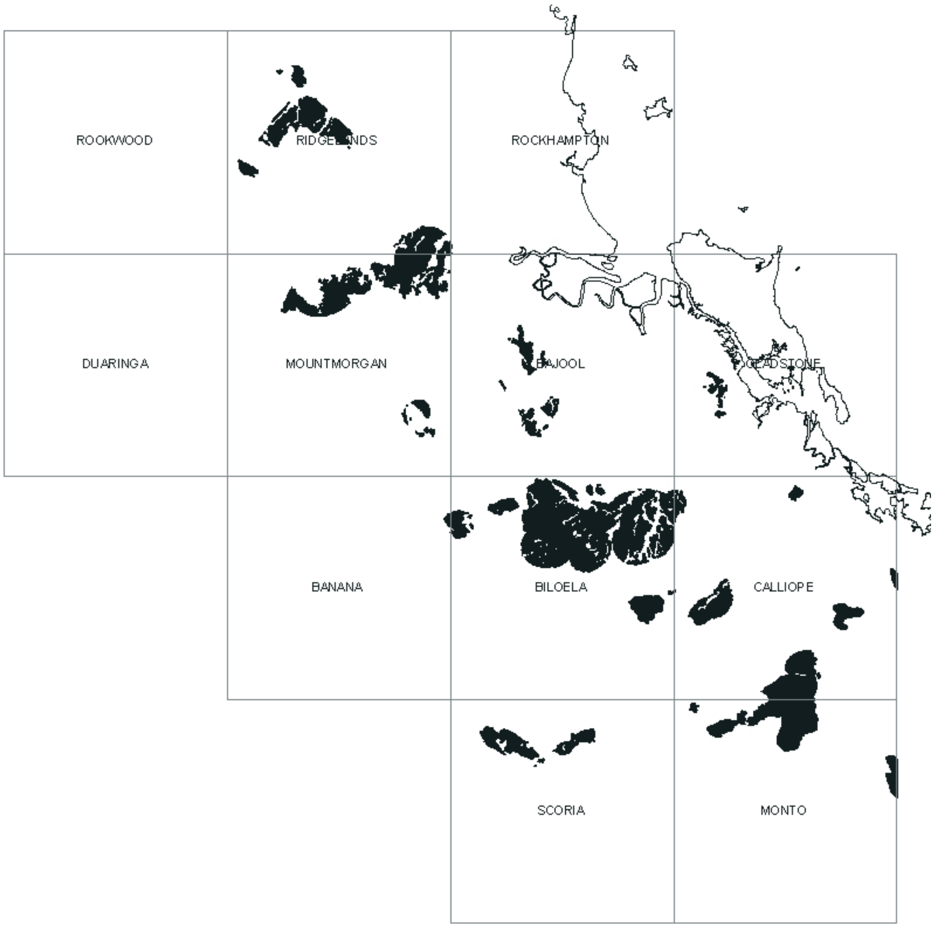


Figure 297. Distribution of the Permian to Triassic intrusives

Figure 298. Redshirt Granite forming the prominent peak of Mount Redshirt viewed from the west, rising above the Dumgree Tonalite and Rocky Point Granodiorite

Figure 299. Small cliffs viewed from a rocky knoll north of the summit of Mount Robert, composed of Robert Granite

potassium. Most have a moderate aeromagnetic response. Some are centred on deep gravity lows and must extend to depths of a few kilometres. The layered gabbros are marked by strong magnetic and positive gravity anomalies.

Lithology and petrography

The dominant lithology is biotite-rich granodiorite, but there is a complete range from layered gabbros to syenogranite. The intrusions have calc-alkaline chemistry, and typically contain clinopyroxene, hornblende and biotite. Muscovite is rare, and only is present as a primary mineral in the Pack Granite and possibly in the Voewood Granite.

Many of the plutons are zoned, typically with a mafic rim intruded by a felsic core, giving them a bimodal composition. The Moonmera Granodiorite is surrounded by concentric dyke swarms and exhibits some of the features of a ring complex. Other plutons are cut by a later series of parallel andesitic dykes trending north-north-west (eg. the Bocoolima Granodiorite). The scarcity of foliation, prevalence of micrographic textures, and narrow hornfels zones indicate emplacement at a high level in the crust. This is supported by the occurrence of Triassic volcanics of similar composition.

Most granites are I-type, but some of the later intrusions are alkaline with sodic amphibole and A-type chemistry. Apart from this change in the Late Triassic, the Permo-Triassic group as a whole shows no systematic geochemical trends.

Geochemistry

On discriminant diagrams of Pearce & others (1984), the Permo-Triassic granitoids fall entirely in the volcanic arc granite fields (Figure 300). The Robert and Molangul Granites, both of which contain alkali amphiboles, fall in the within plate field (Figure 301), and tend towards A-type chemistry. Of these, only the Robert Granite has been dated, and has given a Late Triassic age (Appendix 1).

Geochemically, the Permian-Triassic intrusives in the Yarrol Project area belong to the Clarence River Supersuite defined in the New England region of New South Wales by Shaw & Flood (1981) and Bryant &

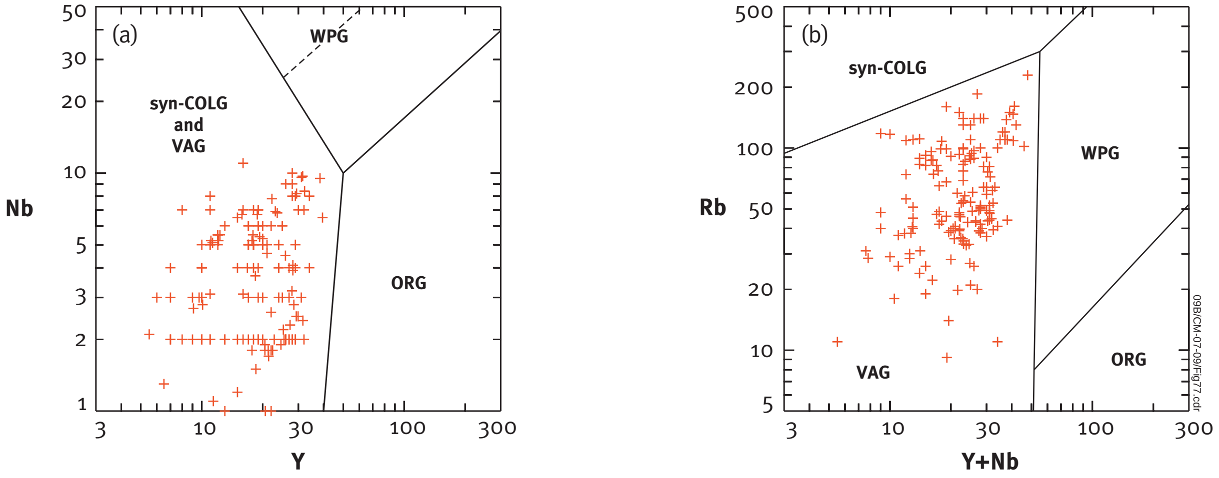


Figure 300. (a) Nb - Y and (b) Rb - Y+Nb discriminant plots of Pearce & others (1984) of felsic samples from the Permian to Triassic granitoids. VAG = volcanic arc granites, ORG = ocean ridge granites, WPG = within plate granites, and syn-COLG = syn-collisional granites. Analyses from GSQ and from B. Chappell & R. Bultitude (unpublished).

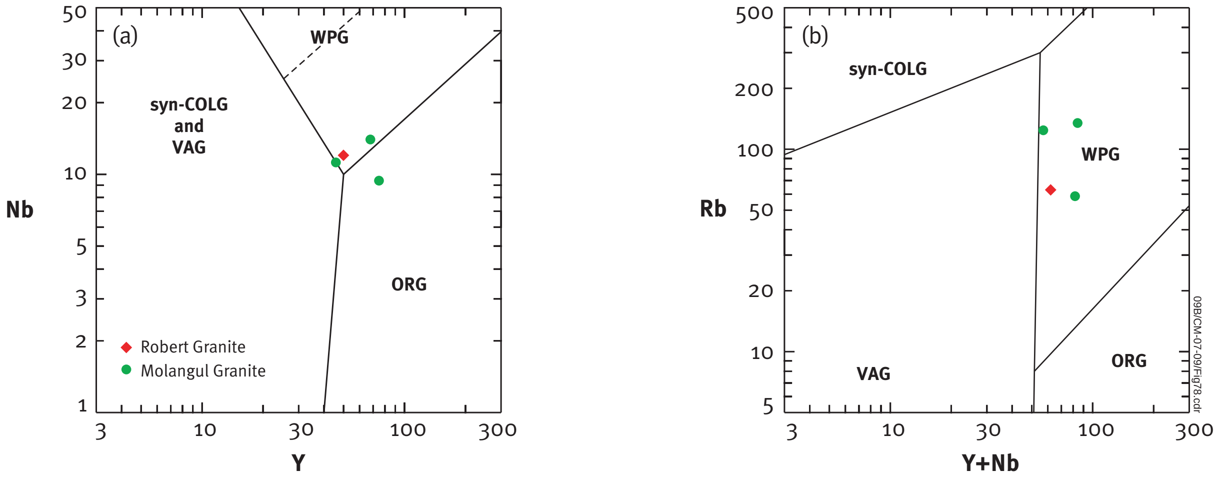


Figure 301. (a) Nb - Y and (b) Rb - Y+Nb discriminant plots of Pearce & others (1984) of samples from the Robert and Molangul Granites. VAG = volcanic arc granites, ORG = ocean ridge granites, WPG = within plate granites, and syn-COLG = syn-collisional granites. Analyses from GSQ and from B. Chappell & R. Bultitude (unpublished).

others (1997). A number of different groups can be recognised within this supersuite. For example, compositional similarities between the Duncans Creek Trondhjemite, Mount Ephraim intrusion, and the high-Si group within the Kaloe Granodiorite were noted by Bryant & others (1997, page 984). Collectively, these intrusions share a number of geochemical features such as high Al_2O_3 and Sr, and low Sc and Y. K_2O, Rb and Th are low and for the group as a whole tend to decrease with increasing SiO_2. The Bruxner Monzogranite, Koreelan Creek Granodiorite, Boswell and Gummi Plains intrusions, the high-K group within the Dumbudgery Creek Granodiorite, and the low-Si group within the Kaloe Granodiorite have high contents of total alkalis, K_2O, Rb, Th, and Ce, and low CaO. TiO_2 and Zr are also high at SiO_2 levels up to about 65%. The Towgon Grange and Jenny Lind Granodiorites, the Omadale Brook intrusion, and the low-K Dumbudgery Creek Granodiorite show compositional similarities (Bryant & others, 1997) such as high CaO and low total alkalis. They have low values of K_2O, Rb and Th that increase only slightly with increasing SiO_2.

Similar subdivisions can be delineated within the Permo-Triassic intrusives of the Yarrol Province. These are described by means of the informal term group because in almost all cases the chemistry of each individual pluton or intrusive phase differs from all others to the extent that definition of suites is not justified. Six groups have been defined, each representing two or more separate plutons or intrusive phases. However, some plutons have unique combinations of geochemical patterns that prevent their inclusion in any of the groups. Distinction between groups is blurred at low levels of SiO_2, and it is also difficult to classify plutons that have only high SiO_2 contents because only the Ridgelands group extends to high SiO_2 levels. For example, the Wyalla and Moonkan Granites are most like the Ridgelands group, but have Zr contents which are too high. The problem is particularly severe for composite plutons with markedly bimodal compositions, such as the Mount Seaview Igneous Complex.

The five groups recognised within the Permo-Triassic intrusives of the Yarrol Province are:

1. Dumgree group, comprising the Dumgree Tonalite, Rocky Point Granodiorite, Bundaleer Tonalite, Norton Tonalite, Umbrella Granodiorite, Mannersley Quartz Microdiorite, Zig Zag Tonalite and unnamed granodiorite at the Briggs or Riverhead prospect at the head of the Calliope River. On Harker diagrams, this group is characterised by high Al_2O_3, CaO, and Sr, and low TiO_2, MgO, and total alkalis. K_2O, Ce, Rb, Th, Y and Zr are all low and for the group as a whole tend to decrease with increasing SiO_2.

2. Littlemore group, comprising the Littlemore Granodiorite and Lawyer Granite. This group is high in total alkalis, K_2O, Rb, and Th. Ce and Sr are high and show a decreasing trend with increasing SiO_2. Y and particularly Zr exhibit curved trends on Harker diagrams, indicating fractionation of zircon.

3. Craiglands group, comprising the Craiglands, Flaggy and Kabra Quartz Monzodiorites. This group has lower SiO_2 contents than the Littlemore group, but shares some geochemical features such as high K_2O and Rb, and plots along trend from the Littlemore group for several elements on Harker diagrams. However, compared to the Littlemore group, the Craiglands group has lower MgO and Th. The Kabra Quartz Monzodiorite has low Rb contents which are at least partly due to alteration.

4. Monal group, comprising the Monal Granodiorite, Diglum Granodiorite, Bocoolima Granodiorite, and Munholme Quartz Diorite. This group is intermediate between the Dumgree and Littlemore groups on Harker plots for several elements, but overlaps these two groups, and reaches extreme values in some cases. The Littlemore group shows curved (fractionated) trends for Ce, Y and Zr. The trend for Sr is similar to that of the Littlemore group, with relatively high values that decrease with increasing SiO_2. Two samples from the Monal Granodiorite from Ridler Creek south of Mount Sugarloaf have low Ce, Y and Zr, suggesting that they are from a separate intrusion.

5. Ridgelands group, comprising the Ridgelands Granodiorite, Wattlebank Granodiorite, Lookerbie Igneous Complex, Voewood Granite, Pack Granite and an unnamed granite along Back Creek near Inverness homestead north of Biloela. This group is intermediate between or overlaps the Dumgree and Monal groups for several elements such as TiO_2, MgO, CaO, total alkalis, Ce and Zr. The main distinguishing feature is its low Sr content, which remains fairly constant at different SiO_2 values. The Sr contents of the Voewood and Pack Granites are slightly higher than the remainder of the group.

Available geochronological data suggest that there is no simple pattern of geochemical change with time except the change to A-type chemistry in the Late Triassic. Also, within batholiths consisting of plutons representing more than one geochemical suite, any one of the suites may be the earliest.

Selected Harker plots depicting these geochemical relationships are shown in Figures 302 to 306.

Age

Ages range from Late Permian to Late Triassic (Appendix 1; Webb & McDougall, 1968; Webb, 1969; Green, 1975; Ford & others, 1976), and the Bucknalla Gabbro may even extend into the basal Jurassic (Holcombe & others, 1997b). Many of the granitoids have ages close to or spanning the Permian-Triassic boundary, including K-Ar and Ar-Ar ages on biotite and hornblende, Rb-Sr ages on whole rocks and biotite, and U-Pb ages on zircon. Based on the radiometric ages, the peak of granite emplacement occurred at this time.

Genesis

A subduction-related origin is favoured, with a transition to extensional magmatism in the Late Triassic possibly due to slab rollback (Gust & others, 1993, 1996). The chemistry is consistent with development at a convergent continental margin (Figure 300). The close association of mafic and felsic magmas suggests an origin involving crustal melting by mantle-derived basalts, although details of this process await further study (Gust & others, 1996). Hildreth & Moorbath (1988) proposed that the geochemistry of volcanics in the southern Andes is established in the lower crust within zones of melting, assimilation, storage and homogenisation (MASH). The chemistry is controlled by the proportions of deep crustal and mantle components, the composition of the source rocks in the lower crust, and the depth at which the MASH process takes place. Similar controls can be envisaged for the intrusives of the Yarrol project area. One reason for the prevalence of granites in magmatic arcs, and therefore in a convergent margin setting, is that the thermal input into the base of the crust is high (Cruden, 2005).

The Sr isotopic ratios of granites are generally low, indicating that neither themselves nor their source had a long crustal residence time (Rogers & Greenberg, 1990). An extended period of melting could deplete the lower crustal source material leading to a decrease of $^{87}Sr/^{86}Sr$ values (Hildreth & Moorbath, 1988, page 83). The only Sr isotopic measurement specifically from the Yarrol project area is that for the Ridgelands

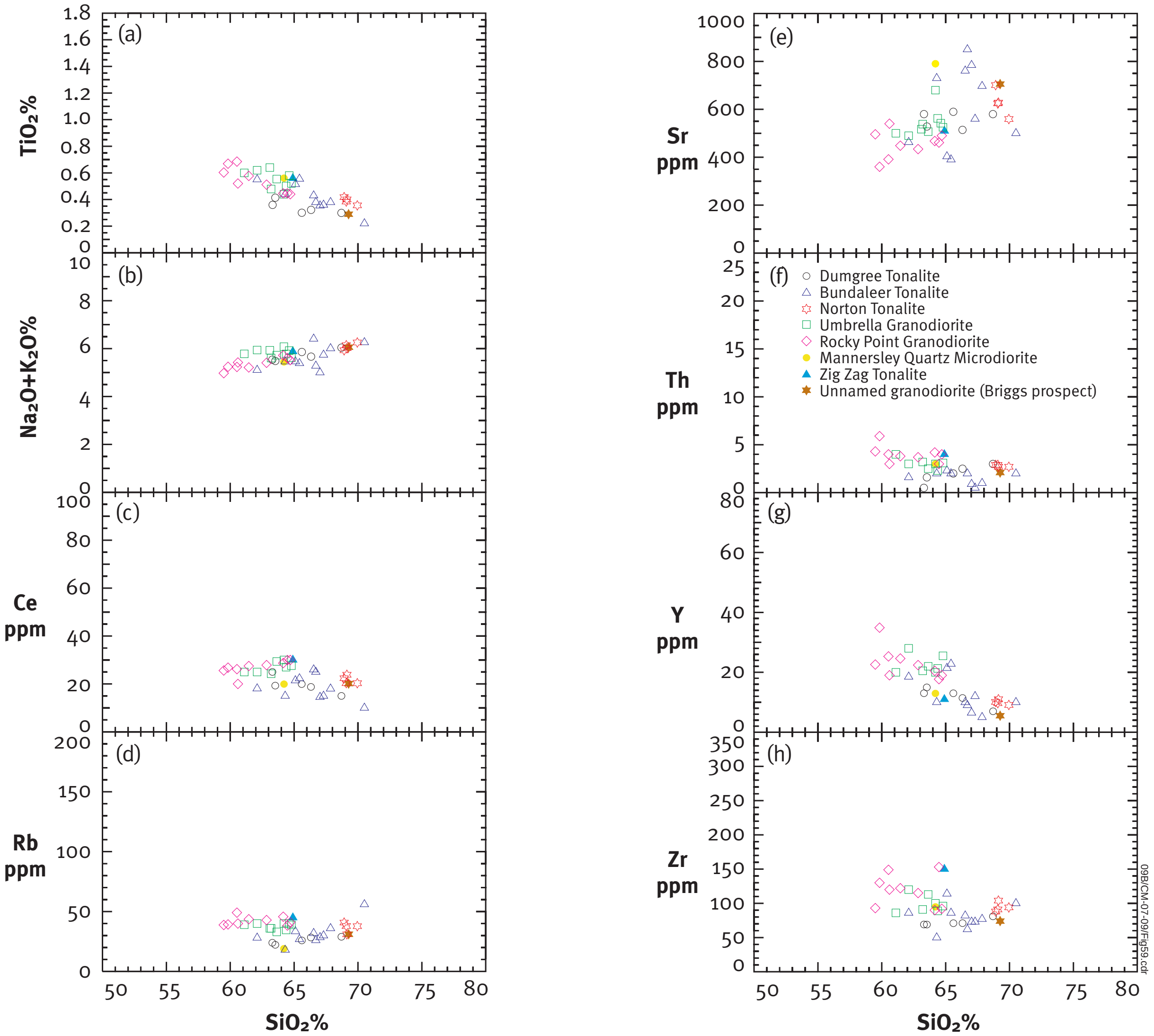


Figure 302. Selected Harker plots for the Dumgree group. Analyses from GSQ and B. Chappell & R. Bultitude (unpublished).

Granodiorite (Appendix 1), which gives an age of 251±2Ma close to the Permian–Triassic boundary and I_{Sr} of 0.70359±5, similar to values obtained from Late Triassic granites in the Maryborough–Gayndah area (Webb & McDougall, 1967; Stephens, 1991).

The position of the Norfolk Rise and Lord Howe Ridge in reconstructions pre-dating the opening of the Tasman Sea is a possible problem in assigning a subduction-related origin to the Permo-Triassic granites. Either these land masses came into contact with Australia after the Middle Triassic, or the trench associated with the postulated subduction zone must have been located several hundred kilometres east of the magmatic belt.

Emplacement

The age range of the Permo-Triassic granites completely overlaps that of the Hunter–Bowen Orogeny, which commenced at 267Ma and concluded with termination of sedimentation in the Bowen Basin at 230Ma (Fielding & others, 1997a,b). This agrees with the conclusion of Hutton (1996, page 114) that “granite is fundamentally syntectonic and occurs mainly in convergent orogenic settings”. However, most plutons are massive, showing no evidence of true syntectonic deformation at the same time as that in the wall rocks, or of solid state deformation (see discussion in Paterson & others, 1989). The only foliated plutons are the Flaggy Quartz Monzodiorite and the Gracemere Gabbro, and these were deformed during emplacement rather than by a regional tectonic event. This raises the question of how unstressed high level granites were emplaced in a contractional environment. However, the Hunter–Bowen Orogeny was an episodic event with relatively brief periods of thrusting alternating with stability and perhaps even extension.

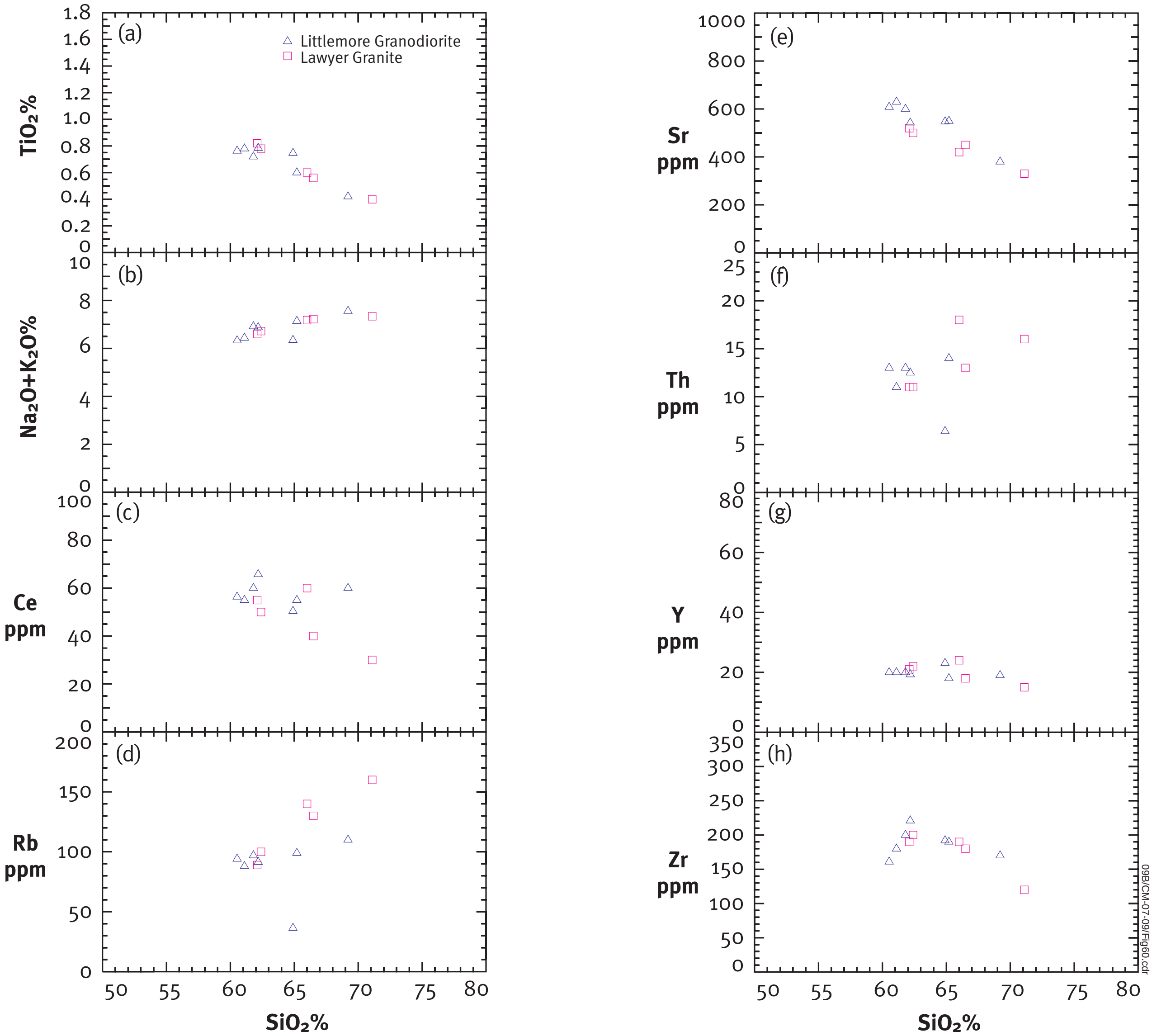


Figure 303. Selected Harker plots for the Littlemore group. Analyses from GSQ and B. Chappell & R. Bultitude (unpublished).

The space for the granitic intrusions is probably the result of vertical movements, either subsidence of the floor, or uplift of the roof. The Flaggy Quartz Monzodiorite has a narrow, foliated contact aureole with a very steep lineation, suggesting that the mechanism of emplacement may have been similar to that proposed by Paterson & Miller (1998). They argued that downward flow of host rocks in a narrow zone surrounding plutons is the main mechanism to create space in arcs being shortened and thickened.

Economic significance

Blevin & Chappell (1992), Blevin & others (1996) and Blevin (2003a,b), have studied the economic aspects of granitic rocks concentrating on eastern Australia, and Blevin & Chappell (1996) have summarised the Permo-Triassic granite metallogeny of the New England Orogen. The relatively oxidised and unfractionated nature of the Clarence River Supersuite, including the intrusives of the Yarrol project area, means that associated mineralisation will mainly be confined to gold and copper and to a lesser extent molybdenum. However, the oxidation state may be too low to form significant copper-gold porphyry deposits (Blevin, 2003b), and gold enrichment may be limited by low K_2O/Na_2O ratios (White, 2003). Known porphyry-style deposits are low grade and currently subeconomic (*e.g.* Wasp, Moonmera, Struck Oil, Briggs, Riverston, Cannindah). Horton (1978) showed that the porphyry Cu-Mo deposits form a narrow linear belt, consistent with a subduction-related origin. They are associated with small intrusions either within or near the margins of larger plutons.

Other mineralisation styles include base metal and magnetite skarns (Glassford, ?Many Peaks), and mesothermal Au-quartz veins (Bouldercombe). Some layered gabbros include magnetite-ilmenite bands, and have been evaluated for platinum group elements (Eulogie Park, Bucknalla).

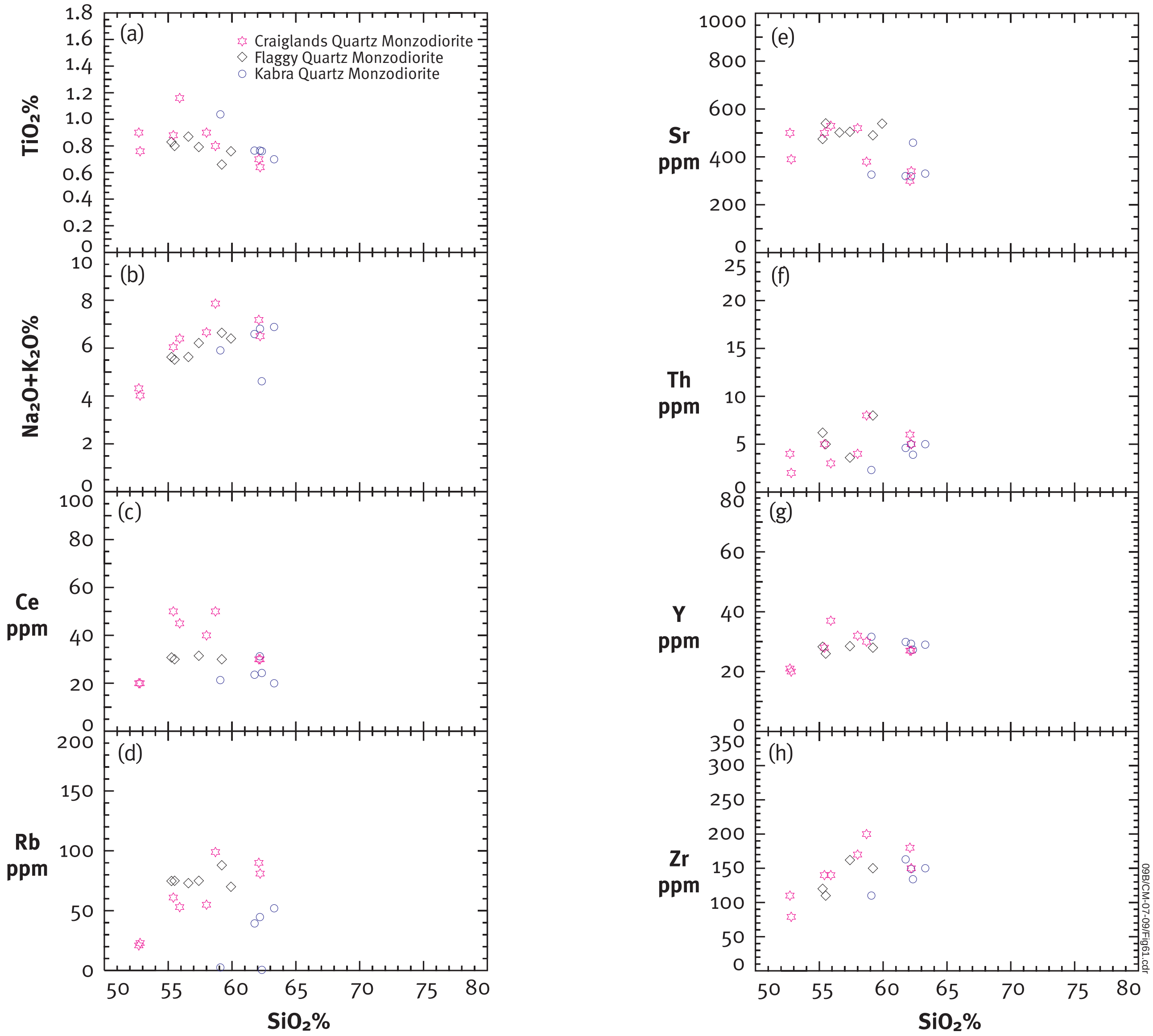


Figure 304. Selected Harker plots for the Craiglands group. Analyses from GSQ and B. Chappell & R. Bultitude (unpublished).

Wattlebank Granodiorite (PRgwa)

(M.A. Hayward)

Introduction

The Wattlebank Granodiorite was included as part of the Ridgelands Granodiorite by Kirkegaard & others (1970). However, it is an entirely separate pluton, with subtle differences in geophysical expression and geochemistry, and age determinations suggest that it is significantly older. It is here defined as a new unit.

Distribution

The pluton straddles the Fitzroy River 45km north-west of Rockhampton. The main outcrop covers an area 5km by 2.5km between Lake Learmouth (Figure 307) and Third Sugarloaf, with a smaller outcrop to the west at the site of the Eden Bann Weir. Another small occurrence is found at GR067 425 near a gauging station in the Fitzroy River, south-west of Wattlebank homestead, and an outcrop of granitic rocks has been reported in the Fitzroy River 1km east of Lake Learmouth. Airborne magnetic data reveal that all these outcrops are part of a single, much larger pluton 14km long and up to 4km wide, orientated east-west, that is partly concealed by alluvial deposits.

Derivation of name

The name is from the property Wattlebank that lies within the main outcrop area.

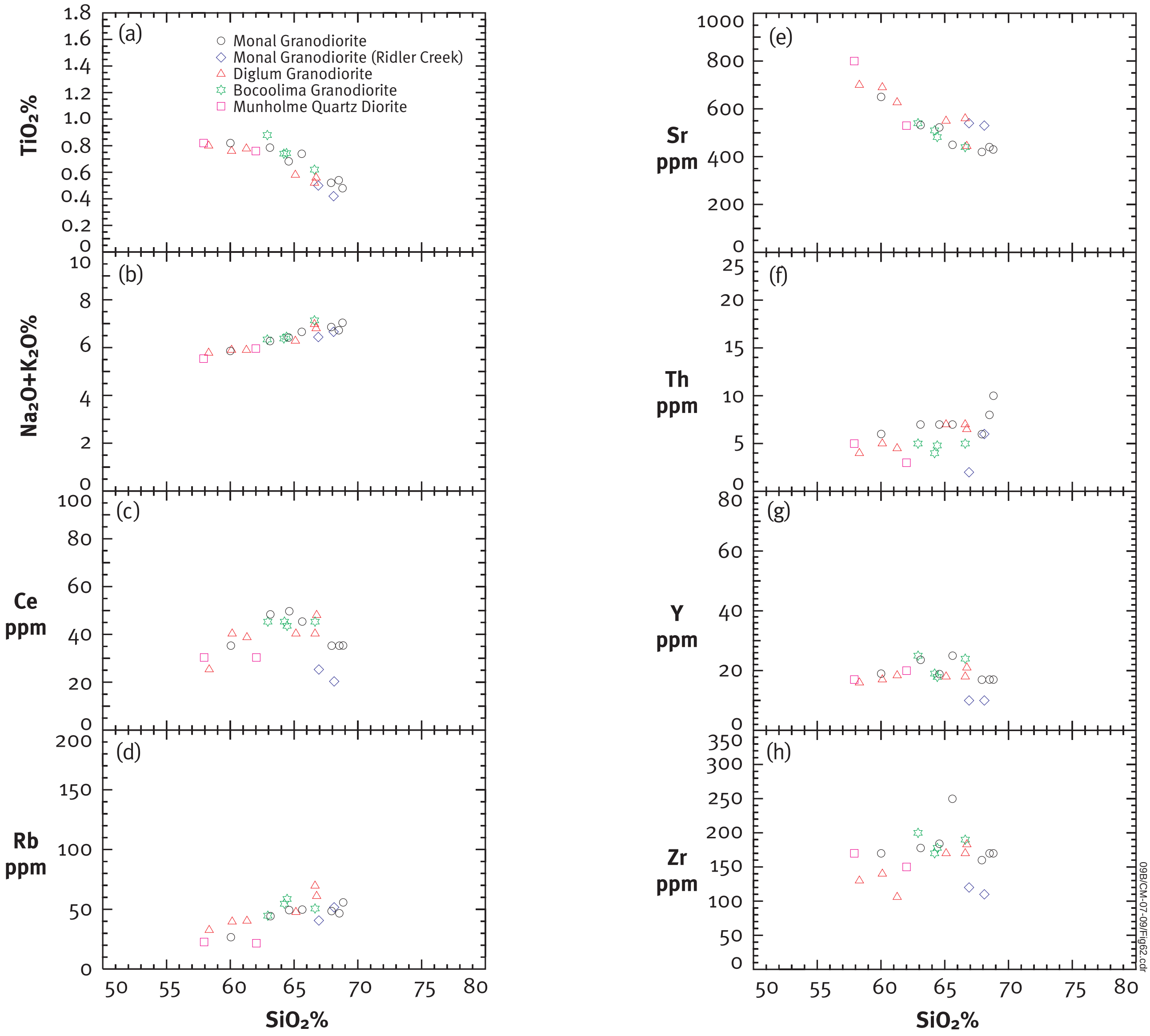


Figure 305. Selected Harker plots for the Monal group. Analyses from GSQ and B. Chappell & R. Bultitude (unpublished)

Type area

The type area is in the vicinity of Wattlebank homestead.

Topographic expression

The intrusion forms low, cleared country along the north bank of the Fitzroy River.

Geophysical expression

The Wattlebank Granodiorite gives dark red colours on images generated from airborne radiometric data, indicating a relatively high content of potassium. This response is similar to that of the Ridgelands Granodiorite. The magnetic signature of the Wattlebank Granodiorite, however, is considerably less than that of the Ridgelands Granodiorite.

Lithology and petrography

The central part of the intrusion adjacent to Wattlebank homestead is a fairly uniform hornblende-biotite granodiorite with sparse, grey, fine to medium-grained xenoliths to 10cm across. Subhedral plagioclase laths up to 4mm long zoned from andesine cores to oligoclase rims make up about 45% of the rock. Quartz (25 to 30%) occurs as large equant grains to 5mm across. K-feldspar constitutes from 10 to 20%, and forms interstitial to poikilitic grains up to 5mm across. Mafic minerals total from 10 to 15% of the rock, and occur

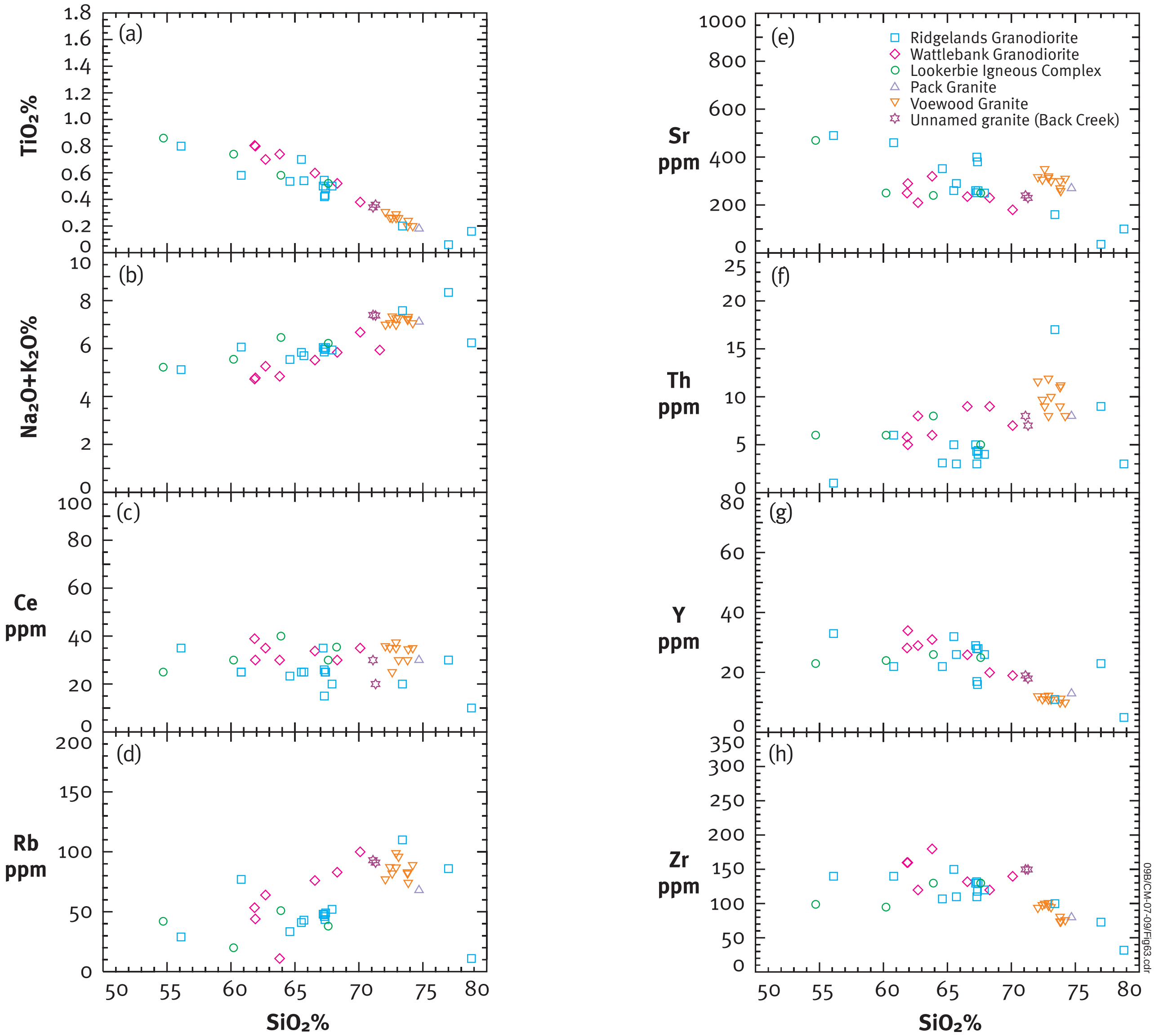


Figure 306. Selected Harker plots for the Ridgelands group. Analyses from GSQ and B. Chappell & R. Bultitude (unpublished).

in bunches that give the rock a glomeroporphyritic appearance. Hornblende and biotite occur in approximately equal proportions, although on average biotite (including secondary chlorite) is probably more abundant. The hornblende is green to pale brown and forms elongate euhedral to subhedral prisms to 6mm long. Accessory minerals include opaques, sphene, apatite and zircon.

Around the margins of the pluton, including the smaller western outcrop near Eden Bann homestead, the composition changes to tonalite. The proportion of plagioclase increases as that of K-feldspar declines to a maximum of 5%, and the percentage of mafic minerals increases to 20 to 25%. One specimen from near the northern contact of the main outcrop is an obviously contaminated rock that contains only about 10% quartz and little or no K-feldspar, and in which augite and very pale brown hornblende are the main mafic minerals (15% of rock). Sphene is particularly abundant, and commonly surrounds small grains of rutile. Other rock types in this area include a pink, medium- to coarse-grained, biotite-bearing pegmatite and a slightly foliated, hornblende-biotite microgranodiorite.

Medium-grained equigranular tonalite crops out in the bed of the Fitzroy River immediately below the Eden Bann Weir. It contains sparse metasedimentary xenoliths to 150cm, and abundant igneous xenoliths (microdiorite and equigranular granodiorite) to 50cm. Aligned mafic minerals impart a weakly to moderately well-developed foliation. Euhedral, twinned hornblende (approximately 20%) is dominant over biotite (5%). Subhedral to euhedral sphene comprises about 1% of the rock.

Figure 307. Distribution of the Wattlebank Granodiorite

Large boulders of microgranodiorite crop out in the bed of the Fitzroy River immediately downstream of the gauging station at GR067 424. Fine biotite is the principal mafic mineral. K-feldspar comprises about 10–15% of the rock.

A veneer of duricrust has formed on the granodiorite about 1.5km north-east of Wattlebank homestead.

Geochemistry

The Wattlebank Granodiorite is very similar in composition to the Ridgelands Granodiorite, and is placed in the Ridgelands group. The main differences from the Ridgelands Granodiorite are higher average contents of Ce, Rb and Th, and lower Ba and possibly Sr (Figure 306).

Age

A sample of tonalite intruding serpentinite on the southern bank of the Fitzroy River at Eden Bann homestead gave K-Ar dates on hornblende of 264 and 269±5Ma (Webb, 1969). This mid- to late Permian age is considerably older than that obtained for the Ridgelands Granodiorite. However, the geochemistry suggests that it is part of the Permian to Triassic suite rather than the mid-Permian intrusives.

Relationships

Airborne magnetic data show that the Wattlebank Granodiorite is a single pluton that is separate from the Ridgelands Granodiorite to the south. It intrudes the Princhester Serpentinite, Mount Alma Formation, and unnamed Permian strata, but its relationship to an unnamed gabbro mass south of the Fitzroy River is uncertain. A small area of Tertiary duricrust is developed on this unit 1.5km north-east of Wattlebank homestead (GR077 423).

Economic significance

No mineralisation is known to be associated with the Wattlebank Granodiorite.

Bouldercombe Igneous Complex

(C.G. Murray)

The name Bouldercombe Complex was introduced by Kirkegaard & others (1970) for an east-north-east-trending batholith of several distinct rock types between Westwood and Gracemere. Although compositions ranging from granite to gabbro were described, the distribution of individual phases and plutons within the batholith was not mapped out. Only general observations, such as the concentration of mafic rocks around the margin of the complex, were made.

Overall, the batholith has a distinctive bilobate shape (Figure 308), although much of the western part is covered by younger sediments and volcanics. The eastern and western lobes were referred to as the Gracemere Complex and Westwood Complex, respectively, by Wood (1974) and Ford & others (1976). Wood (1974) subdivided the western lobe (his Westwood Complex) into three major intrusive phases. He referred to these phases simply as the southern tonalite, the northern tonalite, and granodiorite transitional to granite. These three phases are formally defined as the Flaggy Quartz Monzodiorite, the Umbrella Granodiorite, and the Moonkan Granite respectively. Elsewhere, the only detailed mapping carried out within the batholith was around the Moonmera porphyry Cu-Mo prospect, where Whitcher (1975), Dummett (1978) and Male (1992) outlined a porphyritic granodiorite associated with the mineralisation.

The batholith has now been subdivided into nine named units whose areal extent is based largely on interpretation of airborne radiometric and magnetic data, and the name is changed to Bouldercombe Igneous Complex. The overall form of the complex, and the arcuate shape of constituent rock units, shows some similarity to ring complexes such as those of the Coastal Batholith of Peru (Cobbing & others, 1981).

Figure 308. Distribution of the Bouldercombe Igneous Complex

Figure 309. Distribution of the Flaggy Quartz Monzodiorite

Flaggy Quartz Monzodiorite (Pgbf)

Introduction

The Flaggy Quartz Monzodiorite was first recognised as a distinct phase of the Bouldercombe Igneous Complex by Wood (1974), who called it simply the southern tonalite, and mapped a boundary between it and rocks to the north which have similar mineralogy, but differ in texture and geochemistry.

Distribution

The Flaggy Quartz Monzodiorite forms an arcuate band about 2km wide and 12km long around the southern margin of the western lobe of the Bouldercombe Igneous Complex (Figure 309). Airborne magnetic data suggest that it continues as a marginal phase around the concealed northern part of this lobe.

Derivation of name

The name is derived from Flaggy Creek, which flows through the intrusion into Sandy Creek 9km east of Westwood.

Type area

The type area is along Flaggy Creek about 7km east of Westwood, where the intrusion is well exposed as large flat slabs, and on hillslopes around the headwaters of Flaggy Creek, where it forms tors and boulders.

Topographic expression

For most of its outcrop area, particularly the western part, the Flaggy Quartz Monzodiorite is deeply weathered, and is covered by a regolith layer composed of weathered granitic material or grus. This style of weathering, termed biotite-induced grussification (Isherwood & Street, 1976), is characterised by granular disintegration of the rock resulting from expansion of biotite during chemical alteration, and is typical of biotite-rich granitic rocks.

Apart from sparse boulder outcrops, exposures are confined to large flat slabs in stream beds. Outcrop, mainly in the form of boulders and tors, is more abundant on the slopes of hills covered by younger sedimentary and volcanic rocks, which have protected the intrusion from erosion.

Geophysical expression

The Flaggy Quartz Monzodiorite gives a moderately strong magnetic response and a moderate radiometric response, particularly in the potassium channel. Both responses are similar to those from other phases of the western lobe of the Bouldercombe Igneous Complex, but the Flaggy Quartz Monzodiorite is distinguished by its slightly stronger magnetic signature. This is consistent with its higher average magnetic susceptibility reading of 3480 x 10^{-5} SI units on outcrops, compared with 3015 x 10^{-5} for the adjacent Umbrella Granodiorite.

The contact with the Youlambie Conglomerate is very sharp on both magnetic and radiometric images, because of the very weak magnetic response and strong radiometric response of this sedimentary unit.

Lithology and petrography

The Flaggy Quartz Monzodiorite is a grey to pinkish grey, medium to coarse-grained foliated quartz monzodiorite.

Plagioclase is the dominant mineral, making up a little more than half the rock. It forms zoned subhedral laths up to 10mm long which are andesine in composition. In some outcrops, the laths are strongly aligned. Many crystals are bent and locally broken (Figure 310), supporting the interpretation that the monzodiorite was deformed during the final stages of emplacement, prior to final cooling.

K-feldspar and quartz both form interstitial grains, and each constitutes about 10% of the rock, although the content of K-feldspar shows more variation. Abundant myrmekite occurs at K-feldspar contacts with

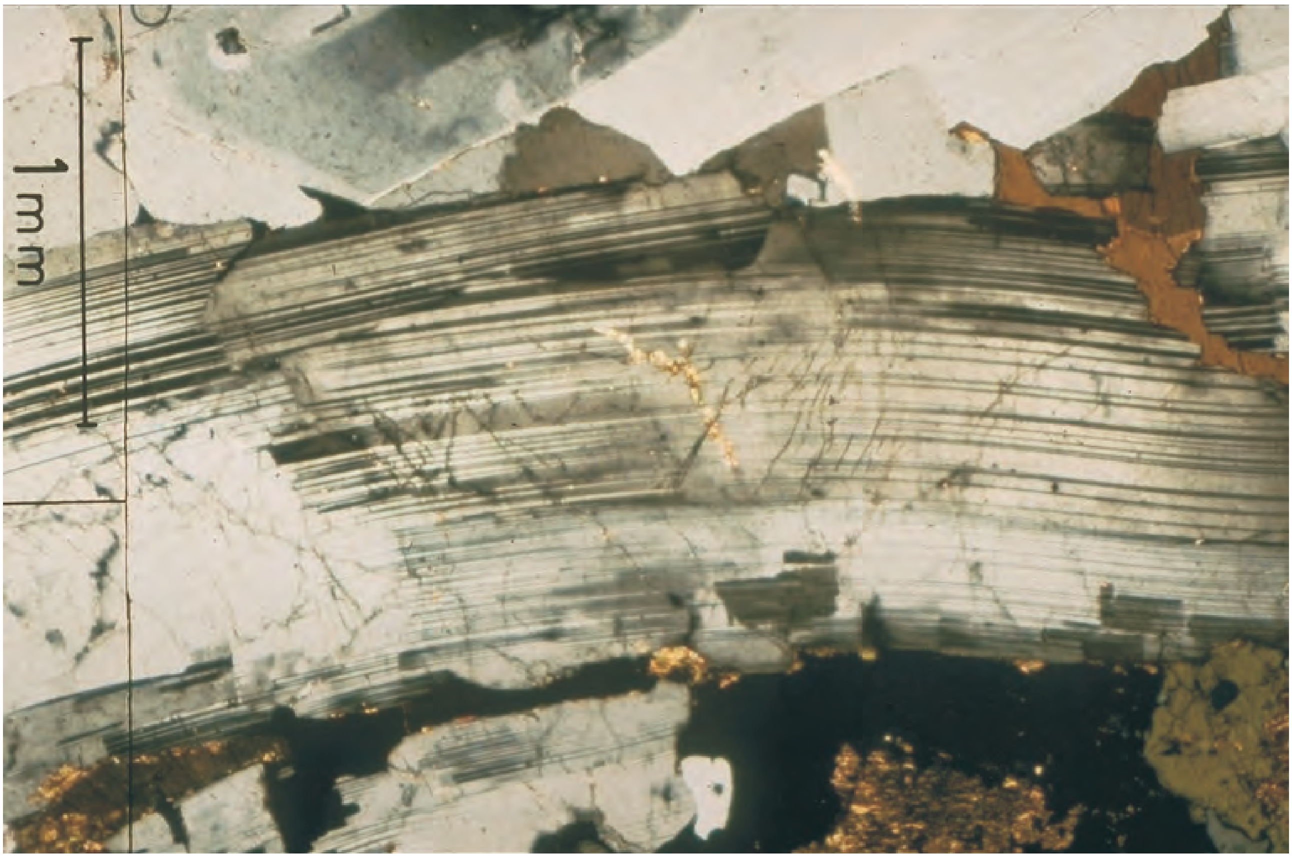


Figure 310. Photomicrograph showing bent and fractured plagioclase grain, Flaggy Quartz Monzodiorite. Crossed nicols. Scale bar is 1mm long.

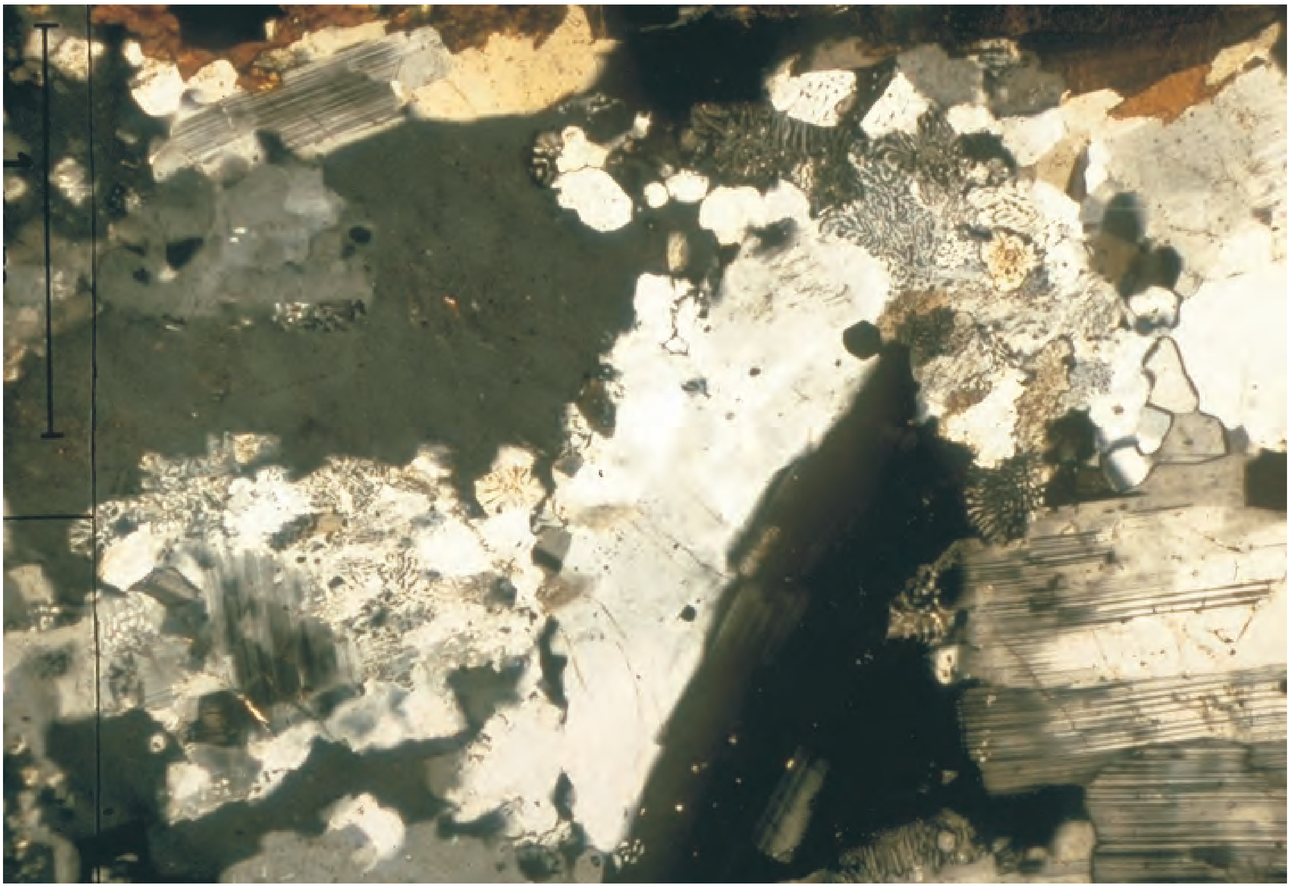


Figure 311. Photomicrograph showing myrmekite at plagioclase — K-feldspar contacts, Flaggy Quartz Monzodiorite. Crossed nicols. Scale bar is 1mm long

plagioclase (Figure 311), and is presumably related to the deformation (Vernon & others, 1983; Simpson, 1985; Simpson & Wintsch, 1989; Paterson & others, 1989).

Mafic minerals make up from 20 to 25% of the rock. All samples contain deep green to brown hornblende, commonly with cores of pale green amphibole probably derived from alteration of pyroxene, and biotite. Specimens from the type area along Flaggy Creek also contain augite and lesser hypersthene partially or totally rimmed by hornblende.

Minor and accessory minerals include opaque(s), apatite, sphene, and zircon. Wood (1974) noted the presence of secondary sericite, clay minerals, calcite, epidote, chlorite, and rutile. Local pegmatitic patches contain relatively coarse magnetite crystals. Tourmaline aggregates are widely distributed as float, but have not been observed in rock samples.

Structure

The Flaggy Quartz Monzodiorite displays a prominent foliation in all outcrops. The foliation is most obviously expressed by elongate lenses of more mafic material (Figure 312), although it is also marked by alignment of plagioclase laths which may cause the rock to split preferentially along the foliation surface. The foliation is generally vertical and parallel to the southern curvilinear intrusive boundary. A well exposed contact with the Youlambie Conglomerate 11km east of Westwood at GR220750 7385375 shows a vertical foliation trending about 50° (parallel to the contact) in both the intrusion and the country rocks. Inclusions of the country rock at this locality are elongated parallel to the foliation and the contact (Figure 313). A vertically plunging lineation suggests that the foliation was related to the emplacement of the quartz monzodiorite.

Geochemistry

The intrusion falls within the Craiglands group. Analyses of the Flaggy Quartz Monzodiorite cover only a narrow range of SiO_2 values (55.3 to 59.9%). The contents of several elements (Al_2O_3, total Fe, CaO, Na_2O+K_2O, K_2O, Rb, Th, and V) fall within the range or along the trend of the Littlemore suite, but TiO_2, MgO, Sr and Ce are lower, and Y is higher (Figures 303 and 304).

Figure 312. Compositional layering due to the concentration of mafic minerals, 4km east of Westwood, Flaggy Quartz Monzodiorite

Figure 313. Elongate inclusions close to the margin of the intrusion giving the rock a migmatitic appearance, 9km south-east of Wycarbah, Flaggy Quartz Monzodiorite

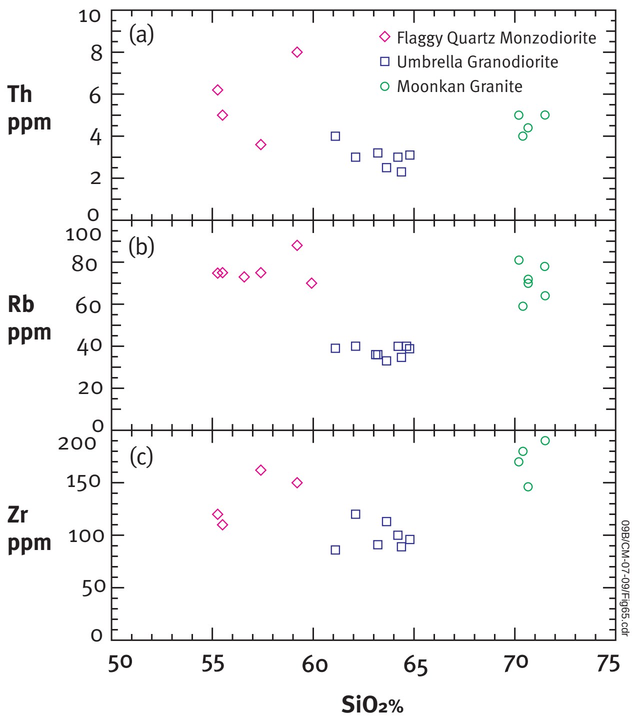


Figure 314. Harker diagram comparing the Flaggy Quartz Monzodiorite, Umbrella Granodiorite and Moonkan Granite. Each of these intrusions has a small and distinct range of SiO_2 values, and the Th, Rb and Zr contents suggest that there is no simple genetic relationship between them. Analyses from Wood (1974) and GSQ and B. Chappell & R. Bultitude (unpublished).

Geochemically, the Flaggy Quartz Monzodiorite is quite distinct from the Umbrella Granodiorite and the Moonkan Granite, the other phases of the western lobe of the Bouldercombe Igneous Complex. Each of the granitoids falls within a relatively small range of SiO_2 contents, and there is no overlap between them (Figure 314). The Flaggy Quartz Monzodiorite has the lowest SiO_2 values, but has significantly higher contents of incompatible elements than the more silicic Umbrella Granodiorite, and for Rb and Th even exceeds or equals the Moonkan Granite (Figure 314). The geochemical data suggest that the Flaggy Quartz Monzodiorite has no simple genetic relationship with either the Umbrella Granodiorite or the Moonkan Granite.

Age

Wood (1974) and Ford & others (1976) reported K-Ar dates of 257±5Ma on biotite and 258±8Ma on hornblende (recalculated to new constants) from a locality 8km east of Westwood. The Flaggy Quartz Monzodiorite is therefore assigned a Late Permian age.

Stratigraphic relationships

The marginal position, arcuate shape, and internal deformation of the Flaggy Quartz Monzodiorite suggest that it was the earliest phase of the western lobe of the Bouldercombe Igneous Complex. It is interpreted to be intruded by the Umbrella Granodiorite to the north, and by the Bucknalla Gabbro to the west, but no contacts have been observed in either case. The monzodiorite intrudes the Youlambie Conglomerate, and is overlain by flat-lying Mesozoic sedimentary and volcanic rocks.

Economic significance

The Flaggy Quartz Monzodiorite is not known to be associated with any mineralisation.

Umbrella Granodiorite (PRgbu)

Introduction

The Umbrella Granodiorite was first recognised as a separate phase of the Bouldercombe Igneous Complex by Wood (1974), who called it the northern tonalite.

Figure 315. Distribution of the Umbrella Granodiorite

Distribution

The Umbrella Granodiorite is the most extensive phase of the western lobe of the Bouldercombe Igneous Complex, covering an area about 12km long and 4km wide in the central part of the lobe (Figure 315).

Derivation of name

The name is derived from Umbrella Creek, which flows through the south-western extremity of the intrusion.

Type area

Typical exposures of the Umbrella Granodiorite, with boulder outcrop, form low ridges on both sides of Junction Creek just south of its confluence with Sandy Creek, 10km east-north-east of Westwood. The unit is also well exposed in boulder covered hills south of Mount Gordon, Mount Coombs and Browns Lookout between Sandy and Middle Creeks.

Topographic expression

The unit mainly forms low hills and ridges with scattered boulder outcrops and isolated nubbins (Campbell, 1997). Some of the ridges are joint-controlled and trend about 110°.

Geophysical expression

The Umbrella Granodiorite gives a moderate magnetic and radiometric response. Compared with other phases of the western lobe of the Bouldercombe Igneous Complex, its magnetic susceptibility of about 3015 x 10^{-5} SI units is slightly less than that of the Flaggy Quartz Monzodiorite, and greater than that of the Moonkan Granite. Despite its lower K content than the Flaggy Quartz Monzodiorite, the Umbrella Granodiorite gives a similar response in the airborne radiometric data.

Lithology and petrography

The Umbrella Granodiorite is a pale grey, medium-grained, massive biotite-hornblende granodiorite with ubiquitous, and locally abundant, inclusions of darker grey and finer-grained material.

The unit is remarkably uniform in composition. Plagioclase is the dominant mineral, comprising from 50 to 60% of the rock. It forms zoned subhedral to euhedral laths up to 4mm long. Composition ranges from calcic andesine cores to calcic oligoclase rims.

K-feldspar forms both small interstitial grains, and poikilitic crystals up to almost 10 mm across, and makes up from 10 to 15% of the rock. Locally there is minor, very small scale development of myrmekite at K-feldspar-plagioclase contacts. Quartz is slightly more abundant than K-feldspar, ranging from 15 to 20%. It is mainly interstitial, but also occurs as rare poikilitic grains up to 5mm across.

Hornblende and biotite occur in approximately equal proportions, and together comprise 10 to 20% of the rock. Hornblende is green, and forms subhedral prisms up to 3mm long. Pale green amphibole in the centre of some hornblende grains probably represents altered augite cores. Minor and accessory minerals include opaque(s), sphene, apatite, and zircon.

Geochemistry

The Umbrella Granodiorite falls within a fairly narrow range of SiO_2 from 60 to 65%, and is geochemically distinct from the Flaggy Quartz Monzodiorite and Moonkan Granite (Figure 314). It is placed within the Dumgree group (Figure 302), although for many discriminating elements it also falls within the Monal group.

Age

K-Ar dating of biotite from the Umbrella Granodiorite, corrected to new constants, gave an age of 243±5Ma, or Early Triassic (Wood, 1974; Ford & others, 1976). However, the Moonkan Granite, which intrudes the Umbrella Granodiorite, has given dates as old as latest Permian. Therefore, a Late Permian to Early Triassic age has been assigned to the Umbrella Granodiorite.

Stratigraphic relationships

The Umbrella Granodiorite probably intrudes the Flaggy Quartz Monzodiorite, and is in turn intruded by the Moonkan Granite. It is overlain by flat-lying Jurassic and Cretaceous sedimentary and volcanic rocks.

Economic significance

Localised alteration systems associated with brecciation and intrusion of small quartz porphyry plugs in the Sandy Creek area are considered to be a late stage of the magmatic event which produced the Umbrella Granodiorite and Moonkan Granite.

Moonkan Granite (PRgbm)

Introduction

Wood (1974) mapped a small pluton and several smaller outcrops, ranging in composition from granodiorite to granite, to the east of Mount Hay. This intrusion is defined here as the Moonkan Granite.

Distribution

The main pluton covers an area east of Mount Hay about 3 by 1.5km, with a narrow extension to the east across Sandy Creek. Other smaller outcrops assigned to the Moonkan Granite occur to the north-east along Sandy Creek, and south of Mount Gordon (Figure 316).

Derivation of name

The name is derived from Moonkan pastoral holding, centred about 3.5km south-south-east of Wycarbah in the Parish of Playfair. Moonkan holding covers most of the main outcrop area of the granite.

Type area

The type area is in the central part of Moonkan holding, 3.5km south-south-east of Wycarbah, and comprises a group of prominent rocky, forested hills with large outcrops, tors and boulders.

Figure 316. Distribution of the Moonkan Granite

Topographic expression

The Moonkan Granite is more resistant to erosion than adjacent units of the Bouldercombe Igneous Complex, and generally forms hills. The east-west trend of the most prominent group of hills, which form the type area, is probably controlled by jointing.

Geophysical expression

The Moonkan Granite gives a stronger radiometric response than the surrounding Umbrella Granodiorite, due not only to its higher content of radiogenic elements, but also to its fresher outcrop. Although the magnetic susceptibility of the granite is significantly less than that of the Umbrella Granodiorite (about 1740 x 10^{-5} SI units compared with 3015 x 10^{-5} SI units), it is not readily distinguished on aeromagnetic images, probably because of its small areal extent and irregular shape.

Lithology and petrography

The Moonkan Granite is a pale pinkish grey, massive, equigranular, medium-grained hornblende-biotite granite, locally containing sparse fine-grained grey xenoliths. Wood (1974) observed that the smaller outcrops east of the main pluton have finer-grain size.

Plagioclase is the dominant mineral, being slightly more abundant than K-feldspar and quartz. It occurs as zoned subhedral to euhedral laths up to 3mm long, and comprises from 30 to 40% of the rock. Although many crystals show oscillatory zoning on a small scale, they generally exhibit a sharp distinction between a relatively uniform core of andesine composition (as calcic as An_{40}) and a broad relatively uniform rim of calcic oligoclase. Cores of some of the larger grains are altered to chlorite+calcite. Myrmekite is developed in plagioclase rims at contacts with K-feldspar, and in some specimens is reasonably abundant.

K-feldspar and quartz each make up from 25 to 30% of the rock, and form interstitial grains and less common poikilitic crystals up to 4mm across.

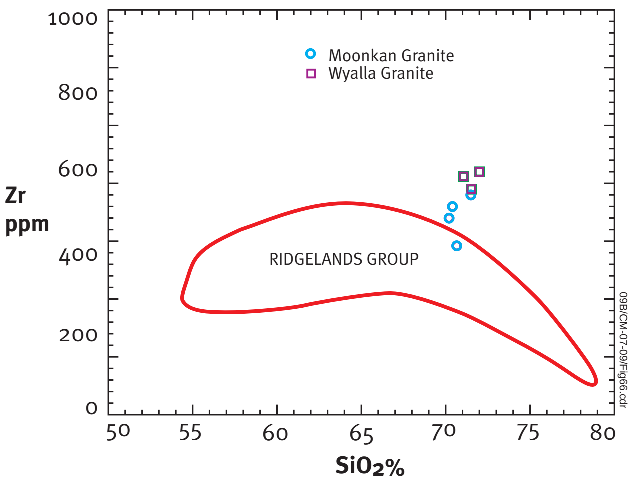


Figure 317. Zr contents of the Moonkan and Wyalla Granites compared to the Ridgelands group. Analyses from GSQ and B. Chappell & R. Bultitude (unpublished).

Biotite and subhedral to euhedral prisms of green hornblende together form about 10% of the rock, with biotite more abundant than hornblende. Minor and accessory minerals include opaque(s), sphene, apatite and zircon.

Geochemistry

Analyses of the Moonkan Granite show a very restricted range of SiO_2, from 70 to 72%. The Moonkan Granite is geochemically distinct from the Umbrella Granodiorite, and no direct genetic connection appears possible. It has similar levels of incompatible elements to the Flaggy Quartz Monzodiorite, despite being much higher in SiO_2 (Figure 314). It is similar to the Ridgelands group apart from Zr, which is considerably higher (Figure 317).

Age

Webb & McDougall (1968) dated biotite from a sample collected along Sandy Creek as 243±2Ma (corrected to new constants), which is the same as the date from the Umbrella Granodiorite. However, K-Ar ages of 251±5 and 257±5Ma were obtained on biotite from a sample from the main outcrop area (Wood, 1974; Ford & others, 1976). Therefore, the Moonkan Granite has been assigned a Late Permian to Early Triassic age.

Relationships

Wood (1974) recorded relationships which clearly indicate that the Moonkan Granite intrudes the Umbrella Granodiorite. At its western extremity, the Moonkan granite is intruded by the Cretaceous rhyolite plug which forms Mount Hay.

Economic significance

Small outcrops of the Moonkan Granite in the Sandy Creek area are spatially related to hydrothermal alteration and breccia pipe development associated with minor gold mineralisation (Wood, 1974).

Unnamed (PRgbe)

Introduction

The youngest intrusions in the western lobe of the Bouldercombe Igneous Complex are small stocks, plugs and dykes of quartz porphyry and diorite which are associated with breccia development and alteration, including greisenisation. Because of their small size and close spatial relationship, these are all grouped together. The following summary is based on a detailed description by Wood (1974).

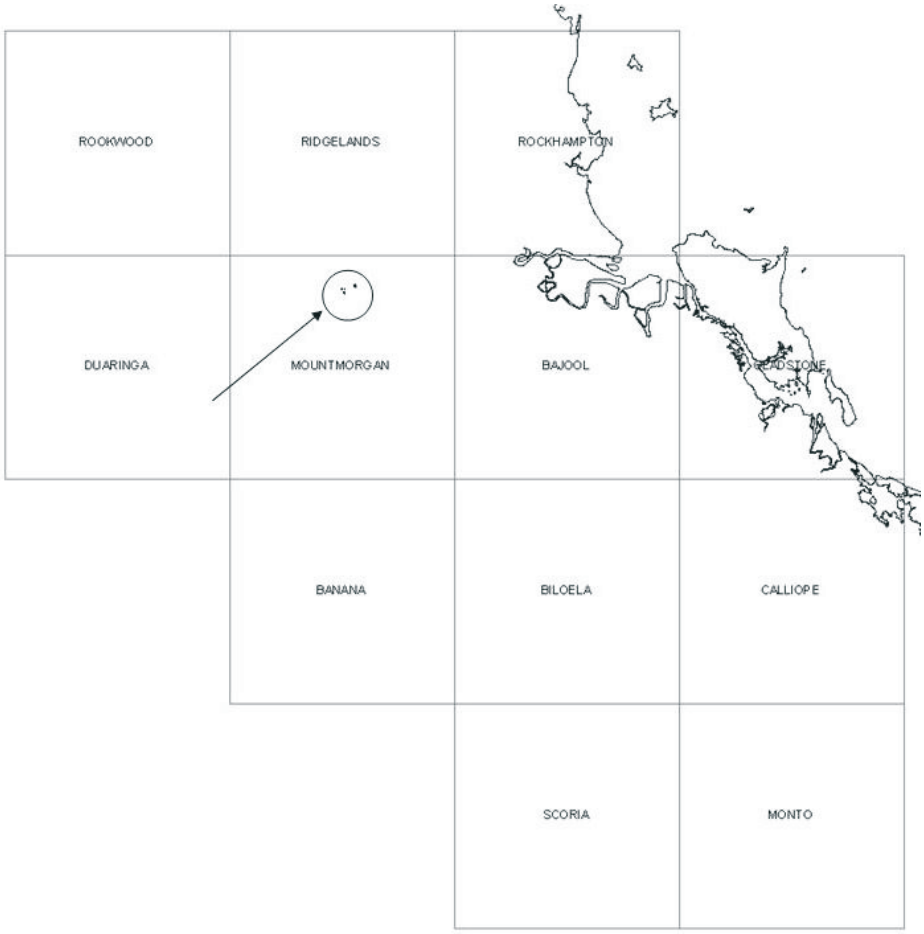


Figure 318. Distribution of unnamed phases of the Bouldercombe Igneous Complex (PRgbe)

Distribution

These unnamed intrusions are concentrated in an east-north-east trending belt about 5km long extending from Sandy Creek to Mount Gordon(Figure 318). The largest single composite intrusion and alteration system, at Mount Gordon, is less than 0.2km^2 in area.

Quartz porphyry forms Mount Gordon and Mount Coombs, and another small plug on the west bank of Sandy Creek centred at GR219000 7389825. A small mass of altered rock, similar in composition to the quartz porphyry, occurs on the eastern side of Pilkington Creek, centred at GR223150 7391375. The largest outcrop of diorite is just east of Sandy Creek at GR219625 7389000.

Topographic expression

Most of the intrusions and alteration zones are more resistant than the surrounding granodiorite and granite, and form low hills.

Geophysical expression

Because of their small size, the intrusions and alteration zones cannot be distinguished on images produced from airborne geophysical data.

Lithology and petrography

The porphyry forming Mount Gordon and Mount Coombs consists of subhedral quartz phenocrysts generally less than 1mm across, and sparse phenocrysts of altered subhedral plagioclase laths, in a groundmass of quartz, plagioclase and K-feldspar (Wood, 1974). The altered porphyry plug on the western bank of Sandy Creek, termed quartz-latite porphyry by Wood (1974), consists of plagioclase, quartz, K-feldspar and hornblende in a quartzo-feldspathic groundmass. The small mass of altered rock east of Pilkington Creek is composed essentially of quartz and sericite (Wood, 1974). Hydrothermal alteration has affected the enclosing granodiorite and granite as well as the felsic intrusives themselves. Collapse breccias with roughly rectangular fragments and incompletely filled voids are associated with many of the alteration systems (Wood, 1974).

Equigranular, chloritised diorite forms the small intrusion just east of Sandy Creek, and dykes of porphyritic hornblende diorite intrude the Umbrella Granodiorite and Moonkan Granite in this area. Both are considered by Wood (1974) to post-date the main phase of hydrothermal alteration.

Geochemistry

Analyses of diorite samples by Wood (1974) fit reasonably well with the Monal group for most elements, but TiO_2 is consistently low, and Rb somewhat high.

Age

Coarse sericite from an altered collapse breccia pipe just east of Sandy Creek gave an Ar/Ar date of 251±5Ma, recalculated to new constants (Wood, 1974; Ford & others, 1976). The close spatial relationship between the hydrothermal alteration and the quartz porphyry intrusives suggests that this is close to their age of intrusion. If so, emplacement of the Umbrella Granodiorite and Moonkan Granite, intrusion of the quartz porphyries, hydrothermal alteration and brecciation may all represent different phases of the same Late Permian to Early Triassic magmatic event.

Hornblende from diorite intersected in a drillhole at Mount Gordon was dated at 234±10Ma and 237±10Ma, recalculated to new constants (Wood, 1974; Ford & others, 1976). These results support the view that the diorite intrusions post-date the alteration.

Economic significance

The close spatial relationship of the quartz porphyry intrusions to hydrothermal alteration and breccia pipe development associated with minor gold mineralisation suggests a direct genetic connection. Wood (1974) considered that the diorites post-dated the mineralising event.

Gavial Gabbro (PRgbg)

Introduction

Kirkegaard & others (1970) noted the presence of gabbro south and south-west of Bouldercombe. These mafic rocks are here defined as the Gavial Gabbro.

Distribution

The main outcrop of the Gavial Gabbro covers an irregularly shaped area of several square kilometres south of the township of Bouldercombe, at the south-eastern extremity of the Bouldercombe Igneous Complex. A smaller mass further west in the headwaters of Poison Creek is also mapped as Gavial Gabbro (Figure 319).

Derivation of name

The name is derived from Gavial Creek, which flows northwards through the intrusion, and which provides the best outcrops of the gabbro.

Type area

The type area is along a short section of Gavial Creek south of Bouldercombe township, where the intrusion is in contact with rocks of the Capella Creek Group.

Topographic expression

The Gavial Gabbro generally forms low, undulating hills with an almost total lack of fresh exposures. At its contact with the more resistant Capella Creek Group, the unit forms locally steep hillslopes.

Geophysical expression

The Gavial Gabbro has high magnetic susceptibility (4585 x 10^{-5} SI units), and stands out as a magnetic high on images produced from airborne geophysical data. The mapped extent of the unit is largely interpreted from the magnetics.

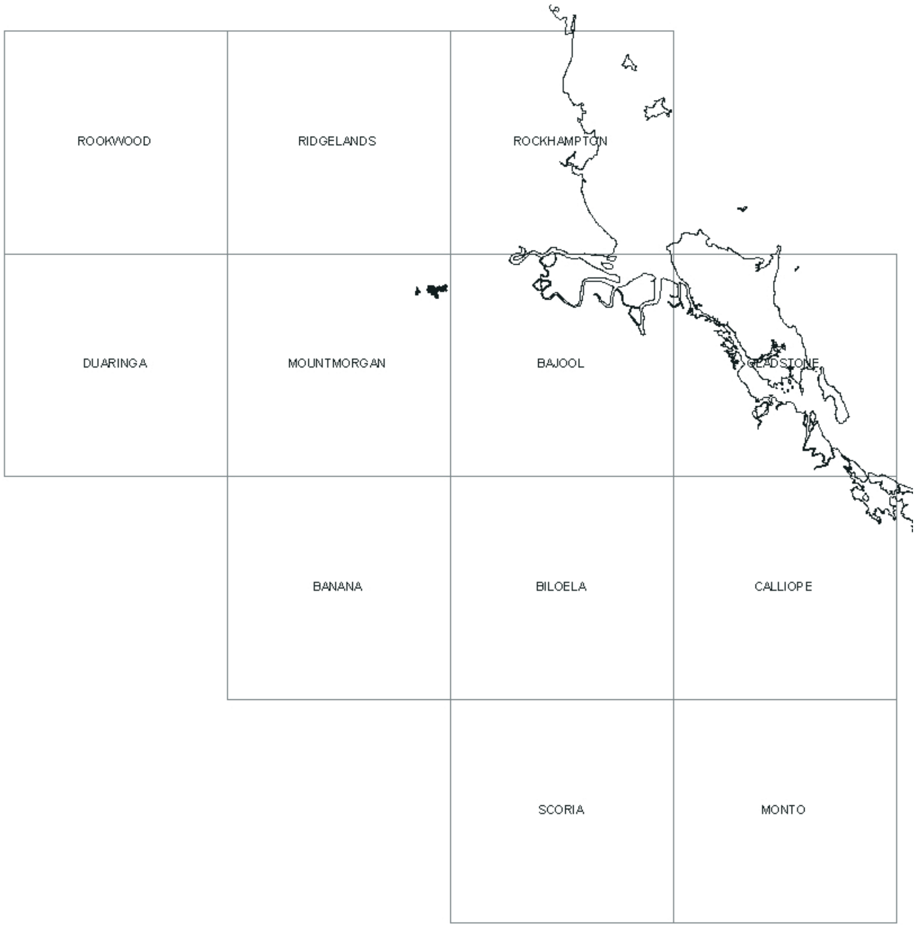


Figure 319. Distribution of the Gavial Gabbro

As expected, the unit gives a low radiometric response in all channels. A slightly higher response for the potassium channel in the easternmost part of the main outcrop area may indicate that the interpreted eastern boundary is incorrect, and should be located a few hundred metres further west.

Lithology and petrography

The Gavial Gabbro is a grey to dark grey medium-grained gabbro with local compositional banding (Figures 320 and 321).

Plagioclase laths up to 3mm long, showing a crude alignment, make up more than 60% of the rock. Compositionally, it falls at the bytownite-labradorite boundary.

The main mafic minerals are augite, hypersthene and hornblende, with very rare altered olivine. Augite is most abundant as subhedral prisms to 2.5mm long, and anhedral grains. Hypersthene forms much smaller stumpy prisms which commonly occur in aggregates. Brown to green hornblende partly replaces and rims augite, and also occurs as poikilitic grains up to 5mm across which enclose hypersthene prisms and plagioclase laths. Opaque grains are relatively abundant.

Geochemistry

Because of its mafic composition, the Gavial Gabbro cannot be definitely assigned to a geochemical suite.

Age

The Gavial Gabbro is assigned a Late Permian to Early Triassic age.

Relationships

The Gavial Gabbro intrudes the Devonian Capella Creek Group and Devonian to Carboniferous Mount Alma Formation. Its relationship with the Bundaleer Tonalite cannot be determined due to lack of outcrop. The

Figure 320. Compositional banding in the Gavial Gabbro

Figure 321. Compositional banding in the Gavial Gabbro

marginal position of the gabbro suggests that emplacement preceded that of the tonalite, but its irregular shape is more consistent with intrusion after the tonalite.

Associated mineralisation

Several gold workings in the Bouldercombe area, the largest of which is the Mount Usher mine, are hosted by the Gavial Gabbro, and there seems to be a definite genetic relationship between intrusion and mineralisation.

Gracemere Gabbro (PRgbr)

Introduction

The name Gracemere Quartz Diorite was introduced by Wolff (1957) for rocks quarried for building stone at a location about 3km south-east of Gracemere. Although the name Gracemere has subsequently been used for the entire eastern lobe of the Bouldercombe Igneous Complex (Gracemere Complex of Wood, 1974 and Ford & others, 1976; Gracemere Granite of Whitcher, 1975; Gracemere Granodiorite of Dummett, 1978), the rocks named by Wolff (1957) are part of a discrete mafic unit which forms the north-eastern margin of the Bouldercombe Igneous Complex.

Distribution

The Gracemere Gabbro forms an arcuate belt about 10km long and from 2–3km wide extending south-east from the township of Gracemere, and comprises the north-eastern extremity of the Bouldercombe Igneous Complex (Figure 322).

Derivation of name

The name is derived from the Parish of Gracemere, which includes most of the intrusion, and from the township of the same name at its northern edge.

Figure 322. Distribution of the Gracemere Gabbro

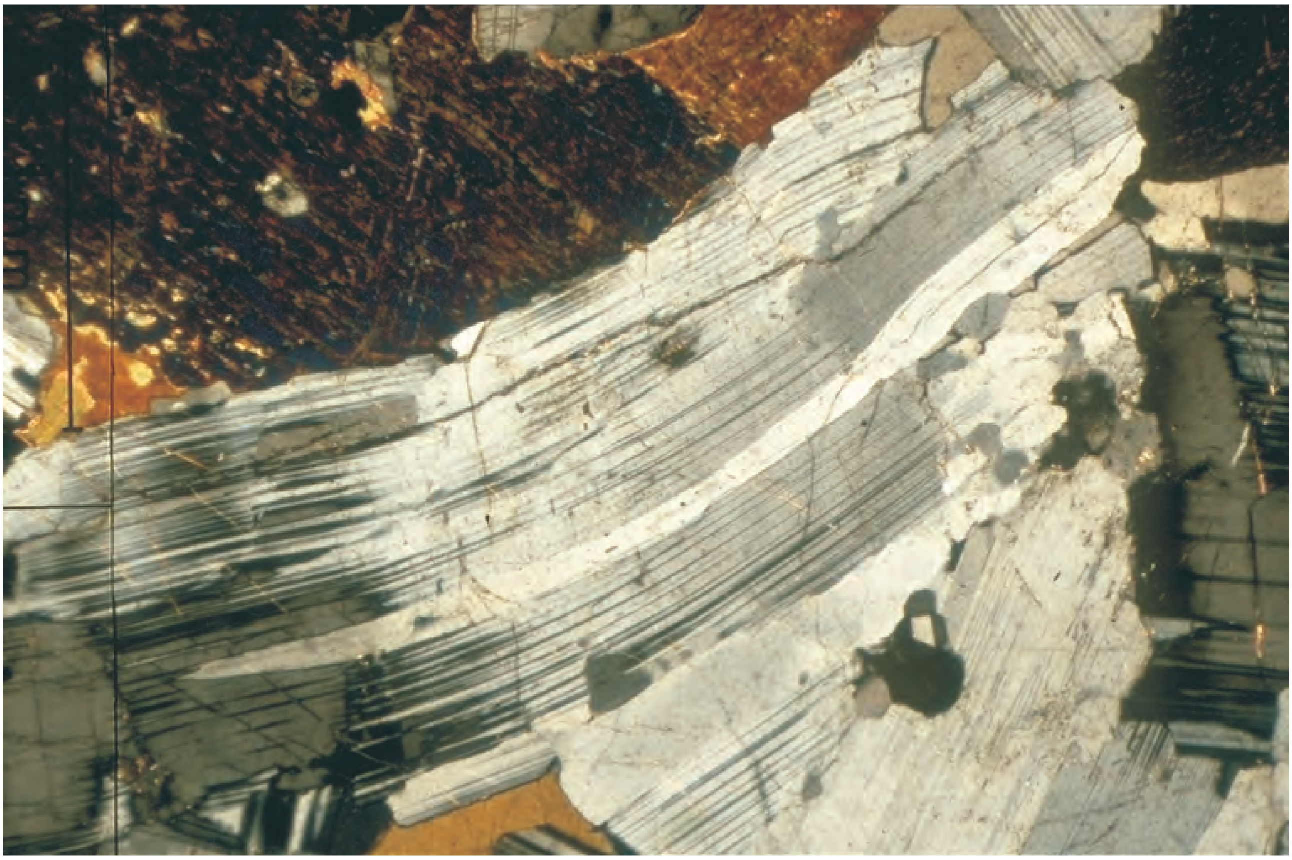


Figure 323. Photomicrograph showing bent plagioclase grain in the Gracemere Gabbro. Crossed nicols. Scale bar is 1mm long.

Type area

The type area is a series of decomposed granite pits about 3km south-east of Gracemere, the largest of which is now used as the local rubbish dump.

Topographic expression

The Gracemere Gabbro forms a low, gently undulating area completely covered by a thick regolith of decomposed granitic material or grus. Fresh rocks are only exposed as corestones in pits and shallow quarries. The style of weathering is typical of biotite-induced grussification (Isherwood & Street, 1976).

Geophysical expression

The unit gives a low radiometric response in all channels. Samples from the intrusion all have moderate magnetic susceptibility, averaging 2565 x 10^{-5} SI units. However, most of the unit gives an extremely low magnetic response on airborne geophysical survey data, consistent with reversed magnetisation. Only a very thin marginal strip around the periphery of the intrusion shows the positive magnetic response expected from the susceptibility readings. The reason for this variation in magnetic properties is unknown.

Lithology and petrography

The Gracemere Gabbro is a uniform, grey, medium to coarse grained gabbro with a prominent foliation due to alignment of plagioclase laths.

Plagioclase of labradorite (An_{57}) composition is the most abundant mineral, comprising about 60% of the rock. Plagioclase laths are up to 8mm long. Many are bent (Figure 323), indicating deformation prior to final cooling of the intrusion.

Mafic minerals, in increasing order of abundance, are hornblende, hypersthene, biotite and augite. Green-brown hornblende partially rims pyroxene grains, particularly augite. Hypersthene forms subhedral to euhedral prisms up to 4mm long. Biotite mainly occurs in aggregates with the other mafic minerals. Subhedral to anhedral augite grains are up to 6mm long, and locally are bent in similar fashion to the plagioclase laths. Subsolidus equilibration is indicated by the occurrence of exsolution lamellae in both pyroxenes.

Interstitial quartz is a minor component, and the rocks contain relatively large opaque grains and accessory apatite.

Geochemistry

Because of its mafic composition, the Gracemere Gabbro cannot be definitely assigned to a geochemical suite.

Age

A sample of the Gracemere Gabbro from the type area was dated by Webb & McDougall (1968). They obtained a result of 244±5Ma (Early Triassic) by K-Ar dating of biotite (corrected to new constants). However, the Bundaleer Tonalite, which is interpreted to intrude the Gracemere Gabbro, has given Late Permian ages. Therefore, a Late Permian to Early Triassic age is assigned to the Gracemere Gabbro.

Stratigraphic relationships

The Gracemere Gabbro intrudes the Carboniferous Rockhampton Group to the north and east. Its relationships with other units of the Bouldercombe Igneous Complex cannot be observed due to lack of outcrop, but can be inferred.

The marginal position and deformed nature of the Gracemere Gabbro suggests that it is intruded by the Bundaleer Tonalite. The boundary between them is placed at the transition from reverse to normal magnetisation interpreted from the airborne geophysical survey. However, if the pattern of magnetisation in the Gracemere Gabbro is symmetrical, the thin band of normally magnetised rocks along its north-eastern margin may be matched by a similar zone to the south-west, which would not be seen on the airborne data because of the similar magnetic susceptibility of the gabbro and the Bundaleer Tonalite (2565 versus 2735 x 10^{-5} SI units respectively).

The airborne magnetic data clearly show that the Gracemere Gabbro is cut off to the west by the Kabra Quartz Monzodiorite.

Economic significance

The Hector mine, reputed to be the first lode gold mine worked in Queensland, is located within the Gracemere Gabbro at its southern end. It is not known whether the mineralisation is associated with the gabbro itself, or with felsic dykes which are presumably related to the adjacent Bundaleer Tonalite. Smaller gold mines also occur within the sedimentary country rocks in this vicinity.

Quarry Gabbro (PRgbq)

Introduction

The Quarry Gabbro is a newly recognised unit within the Bouldercombe Igneous Complex.

Distribution

The gabbro is a relatively small east-west trending intrusion about 4.5km long centred about 2.5km south-east of the Stanwell power station (Figure 324).

Derivation of name

The name is derived from Quarry Creek, which flows through the western part of the intrusion.

Type area

The type area of the Quarry Gabbro is the semi-circle of hills surrounding the southern end of the water storage dam for the Stanwell power station. Excellent outcrops, mainly as large boulders, occur throughout these hills.

Topographic expression

Apart from the valley along Quarry Creek, the intrusion forms fairly low but quite steep hills.

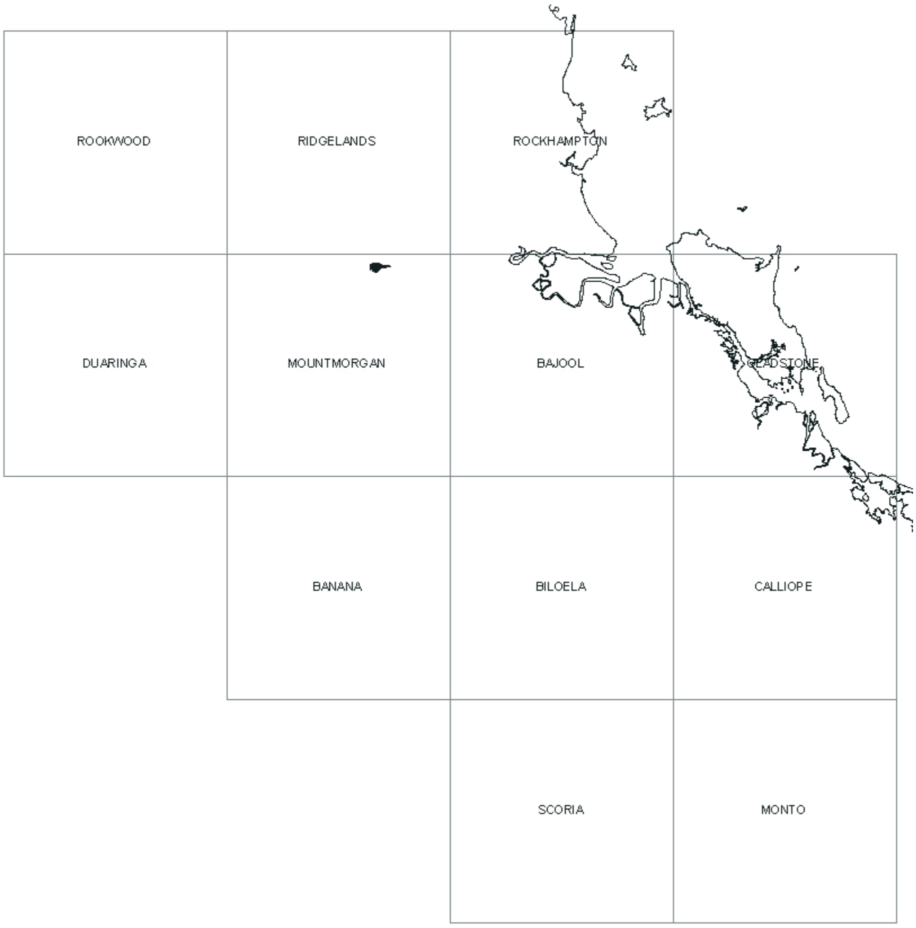


Figure 324. Distribution of the Quarry Gabbro

Geophysical expression

Surprisingly, in view of its overall mafic character and low K_2O content, the Quarry Gabbro gives a moderately strong response for the potassium channel of the airborne radiometric data.

The gabbro corresponds to a prominent magnetic low. Because of its moderate magnetic susceptibility (averaging about 2175 x 10^{-5} SI units), this is interpreted as indicating that the intrusion is reversely magnetised.

Lithology and petrography

The dominant rock type is a dark grey to black medium grained gabbro.

Plagioclase is most abundant, comprising about 60% of the rock. It forms subhedral laths up to 4mm long, and is sodic labradorite in composition.

Augite and hypersthene are the main mafic minerals, with lesser amounts of hornblende and biotite. Augite occurs as subhedral prisms up to 2mm long, and anhedral grains. A combination of twinning and exsolution gives some crystals a herringbone structure. Hypersthene is present as smaller, more regular prisms. Green-brown hornblende rims and partly replaces augite, and rims some hypersthene prisms. In some samples, the pyroxenes are largely replaced by pale green secondary tremolitic amphibole. Opaque grains are almost all enclosed by or associated with biotite, except for localised symplectic intergrowths with hypersthene aggregates.

Minor interstitial quartz is present, and accessory sphene and apatite.

Geochemistry

The Quarry Gabbro cannot be assigned to a geochemical group because of its mafic composition.

Age

The Quarry Gabbro is assigned a Late Permian to Early Triassic age.

Stratigraphic relationships

The Quarry Gabbro intrudes the Devonian Capella Creek Group and the Carboniferous to Permian Lorray Formation, and is overlain by the Jurassic Precipice Sandstone. Its relative age relationships with the Bundaleer Tonalite are unknown.

Economic significance

The Quarry Gabbro is not known to be associated with any significant mineralisation. Large boulders from a site 3km south-east of the Stanwell power station have been quarried as a source of black granite for the building stone industry.

Bundaleer Tonalite (PRgbb)

Introduction

This is a new name for deeply weathered granitic rocks which form most of the eastern lobe of the Bouldercombe Igneous Complex.

The name Gracemere has been applied to the whole of the eastern lobe of the Bouldercombe Igneous Complex (Gracemere Complex of Wood, 1974 and Ford & others, 1976; Gracemere Granite of Whitcher, 1975; Gracemere Granodiorite of Dummett, 1978). However, the name Gracemere Quartz Diorite was first used by Wolff (1957) for rocks which are part of a discrete mafic unit now defined as the Gracemere Gabbro.

The southern part of the Bundaleer Tonalite was referred to as the Moonmera Granite by Mount Morgan mine geologists (Staines, 1952, 1953; The Staff, Mount Morgan Limited, 1965). However, the name Moonmera has since been extensively used for a porphyry copper prospect associated with a younger porphyritic granodiorite intrusion at Moonmera 7km north of Mount Morgan (Whitcher, 1975; Ford & others, 1976; Dummett, 1978; Male, 1992), and it is considered more appropriate to restrict its use to this locality.

Distribution

The Bundaleer Tonalite is the most extensive unit within the Bouldercombe Igneous Complex, forming a north-east trending belt about 16km long and 8km wide from the Razorback Range almost to Gracemere (Figure 325).

Derivation of name

The name is derived from the old siding of Bundaleer on the abandoned Mount Morgan to Kabra railway line, in the western part of the intrusion, where typical outcrops occur.

Type area

The best exposures of the Bundaleer Tonalite are in cuttings along the abandoned Mount Morgan to Kabra railway line up the Razorback Range at Moonmera, and this is designated as the type area.

Topographic expression

The Bundaleer Tonalite, particularly the eastern part, is deeply weathered, and forms a low peneplain covered by a thick regolith layer composed of weathered granitic material or grus, with broad alluvial tracts along streams. The style of weathering is indicative of biotite-induced grussification (Isherwood & Street, 1976). The best exposures are on hill slopes at intrusive contacts with more resistant country rocks, and where the tonalite is being exhumed from a cover of younger sedimentary rocks.

Geophysical expression

The eastern part of the Bundaleer Tonalite, characterised by deep weathering and extensive alluvial cover, gives a very low radiometric response in all channels. The western part, an area of better outcrop, gives a low to moderate response for potassium.

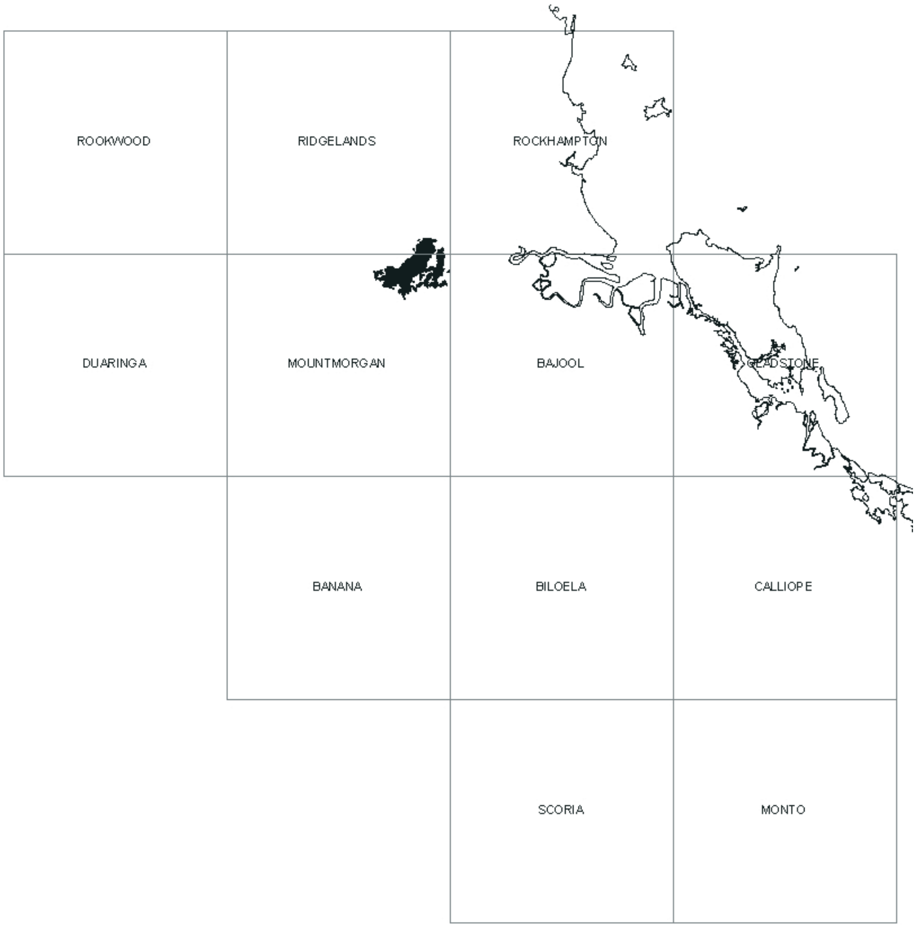


Figure 325. Distribution of the Bundaleer Tonalite

The unit gives a fairly uniform, moderate magnetic response consistent with its average magnetic susceptibility of 2735 x 10^{-5} SI units. A prominent Y shaped magnetic low extending in a north-north-east direction through the central part of the intrusion is interpreted as a demagnetised zone due to alteration. Altered samples collected from the eastern margin of this zone have magnetic susceptibility of only 7 x 10^{-5} SI units.

Lithology and petrography

The Bundaleer Tonalite is a pale grey medium grained leucocratic hornblende-biotite tonalite.

Subhedral laths of plagioclase up to 3mm long are the main constituent, ranging from 55 to 70% of the rocks and averaging about 60%. Crystals are zoned from calcic andesine cores to thin rims of oligoclase. In most samples, K-feldspar ranges from zero to 5%, and is only a minor component. Samples from the westernmost part of the intrusion, near Centre Creek, contain about 10% K-feldspar as poikilitic crystals up to 5mm across. In these rocks, myrmekite commonly occurs in the marginal zones of plagioclase laths at K-feldspar contacts. Quartz occurs as anhedral grains up to 3mm across, ranges from 15 to 30%, and averages about 24%.

The mafic minerals, biotite and hornblende, comprise a little more than 10% of the rocks. Overall, biotite is slightly more abundant than hornblende, and locally, as in the rocks from the western part of the intrusion containing poikilitic K-feldspar, it is the only mafic mineral present. The hornblende is green, and forms prisms up to 2mm and rarely 5mm long.

Other minerals include minor and accessory opaque grains, sphene, apatite, and zircon, and secondary chlorite, epidote, sericite and clays. Ayres (1974) reported the opaque oxides magnetite, hematite and ilmenite from rocks near the Moonmera porphyry copper-molybdenum prospect.

The altered rocks from the eastern margin of the prominent magnetic low are almost pure white in colour. As suggested by their low magnetic susceptibility readings, opaque grains have been completely destroyed. The original mafic minerals, which may have been dominantly hornblende, have been totally replaced by diallage, a very pale calcic monoclinic pyroxene with a single parting. This forms twinned prismatic crystals up to

5mm long, presumably pseudomorphing hornblende, and also smaller grains and aggregates. Rare grains are complex intergrowths of diallage and pale green secondary amphibole. The plagioclase laths are extensively altered, with irregular patches and bands along twin planes, presumably of more albitic composition.

Geochemistry

The Bundaleer Tonalite, together with the Dumgree Tonalite from the Galloway Plains Igneous Complex, is one of the main units which define the Dumgree group (Figure 302).

Age

Webb & McDougall (1968) dated a single sample of the Bundaleer Tonalite from a cutting along the abandoned Mount Morgan-Kabra railway line, and obtained a K-Ar biotite age of 242±5Ma (recalculated to new constants). Subsequent dating of samples from the same locality, reported by Green (1975) and Ford & others (1976), gave results of 251±5 and 264±5Ma from K-Ar dating of biotite (revised to new constants), and 250±8Ma from hornblende. On the basis of these results, the Bundaleer Tonalite is assigned a Late Permian to Early Triassic age.

Stratigraphic relationships

The Bundaleer Tonalite intrudes the Devonian Capella Creek Group and the Devonian to Carboniferous Mount Alma Formation, and is overlain by the Jurassic Precipice Sandstone. As far as relationships with other units of the Bouldercombe Igneous Complex are concerned, the Bundaleer Tonalite appears to intrude the Gracemere Gabbro, is intruded by the Moonmera Porphyritic Granodiorite, and is interpreted to be intruded by the Kabra Quartz Monzodiorite and by undivided rocks of the complex south of Table Mountain. Relative age relations with the Gavial Gabbro and Quarry Gabbro are unknown.

Economic significance

No mineralisation is known to be associated with the Bundaleer Tonalite.

Moonmera Porphyritic Granodiorite (PRgbm)

Introduction

The name Moonmera Granite was first used by Mount Morgan mine geologists (Staines, 1952, 1953; The Staff, Mount Morgan Limited, 1965) for all the intrusive rocks north of the Razorback Range. However, the name Moonmera has since been extensively used for a porphyry copper prospect associated with a younger porphyritic granodiorite intrusion at Moonmera north of Mount Morgan (Whitcher, 1975; Ford & others, 1976; Dummett, 1978; Male, 1992), and it is considered more appropriate to restrict its use to this locality.

Distribution

The Moonmera Porphyritic Granodiorite is a small composite intrusion covering less than 1km^2 on the slopes of the Razorback Range 7km north of Mount Morgan (Figure 326).

Derivation of name

The name is derived from the district of Moonmera and its former railway station on the original rack railway up the Razorback Range.

Type area

Because of its small size and the range of rock types represented, the type area for the Moonmera Porphyritic Granodiorite covers the entire intrusion.

Topographic expression

The Moonmera Porphyritic Granodiorite crops out on a steep east-facing hill slope of the Razorback Range beneath a capping of cliff-forming Precipice Sandstone.

Geophysical expression

The intrusion corresponds with an area of low magnetic and high radiometric response relative to the surrounding Bundaleer Tonalite, caused by demagnetisation and potassic alteration respectively.

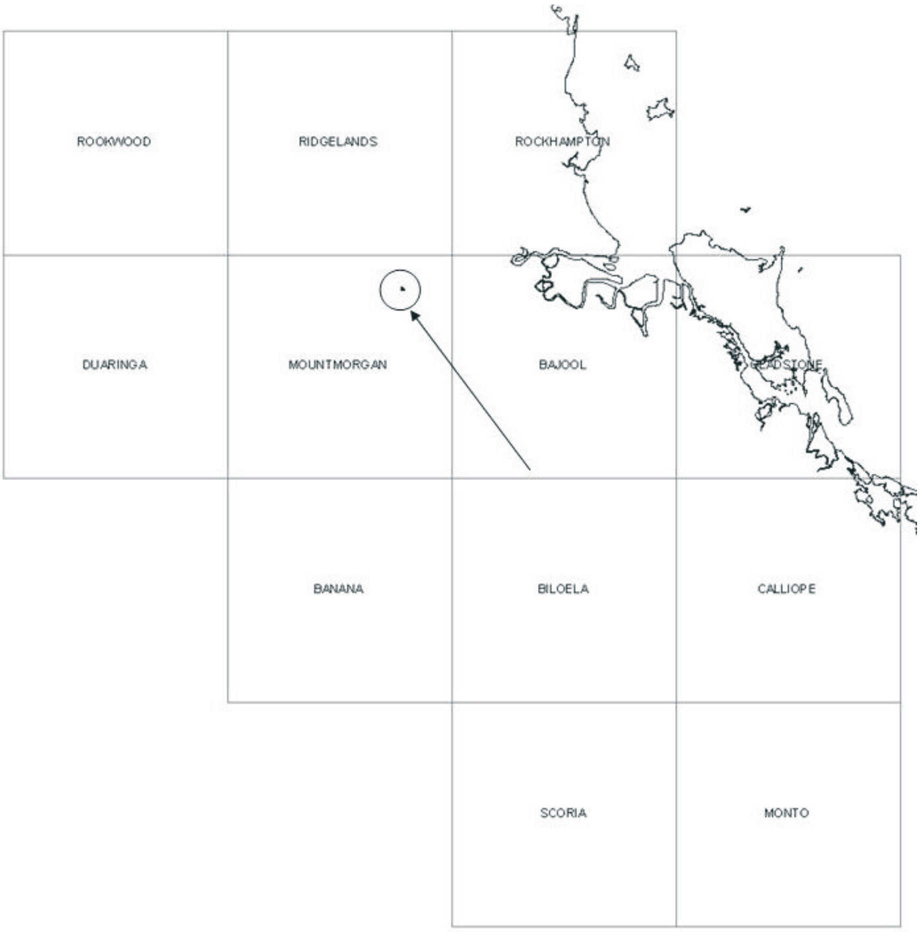


Figure 326. Distribution of the Moonmera Porphyritic Granodiorite

Lithology and petrography

The Moonmera Porphyritic Granodiorite has been described by Ayres (1974), Whitcher (1975), Dummett (1978) and Male (1992). The following summary is compiled from their work.

The intrusion consists of at least two phases of altered porphyritic biotite granodiorite and a hypabyssal breccia pipe. Numerous dykes, some radial and some forming incomplete ring structures, surround the intrusive complex.

The oldest porphyritic granodiorite occurs as two separate bodies in the centre and north. It comprises a high proportion of phenocrysts of plagioclase, quartz and biotite (totalling from 70 to 75% of the rock) in a quartzofeldspathic groundmass. A younger porphyritic granodiorite phase, exposed in 3 separate outcrops, contains a similar mineral assemblage, but with a lower proportion of quartz phenocrysts. Emplacement of this younger porphyry was associated with brecciation of the surrounding rocks. Male (1992) recognised very small outcrops of an even younger porphyry with a glassy groundmass.

The breccia pipe, termed tuffisite by Dummett (1978), includes clasts of both older and younger porphyritic granodiorite phases, of the Bundaleer Tonalite, and of some rock types which are not exposed in the area.

The dykes include a range of compositions. One of the larger arcuate dykes, well exposed in a cutting along the abandoned railway north of the main intrusive centre, is an extremely felsic rock composed very largely of zoned plagioclase (sodic andesine) with minor interstitial quartz and K-feldspar and virtually no mafic minerals. Secondary minerals are calcite and sericite (both of which replace plagioclase cores), prominent rutile, and almost colourless chlorite. A parallel dyke to the west contains euhedral phenocrysts of zoned plagioclase and lesser green to pale brown hornblende in a groundmass of plagioclase and fine micrographic intergrowths of K-feldspar and quartz, with accessory apatite, zircon and tourmaline and secondary chlorite and calcite. Other dykes contain phenocrysts of quartz and biotite in addition to plagioclase (Ayres, 1974; Dummett, 1978).

Geochemistry

Extensive and locally intensive alteration of the porphyritic granodiorite bodies is a problem for interpretation of their primary geochemistry. The range of SiO_2 values for each of the two main phases of porphyritic granodiorite is quite restricted, suggesting that Si is relatively constant during alteration, and that these phases represent distinctly different primary compositions. Dykes show a much wider range of SiO_2 contents. Some correspond to the two main porphyritic granodiorite phases, whereas others are clearly not related. Harker plots demonstrate that the elements Al, Ti, Zr, Y, V, La, Nd and Ce are least affected by alteration, and Ca, Na, K, Rb, Sr, Nb and Th are severely affected. The different intrusive phases cannot be placed in any geochemical group.

Age

No samples of the Moonmera Porphyritic Granodiorite have been directly dated. However, K-Ar dates of 257±5Ma and 258±5Ma were obtained for hydrothermal biotites from veins in altered Bundaleer Tonalite adjacent to the granodiorite by Ford & others (1976). These dates cannot be distinguished from ages obtained from primary biotite of the Bundaleer Tonalite, so the Moonmera Porphyritic Granodiorite is also regarded as Late Permian to Early Triassic.

Stratigraphic relationships

The Moonmera Porphyritic Granodiorite intrudes the Bundaleer Tonalite, and is overlain by the Jurassic Precipice Sandstone to the west. Within the porphyritic granodiorite complex, it is clear that emplacement, brecciation, and associated mineralisation took place over a geologically short time span. This may explain the different interpretations that have been proposed for the age relations between the porphyritic granodiorite phases and the breccia pipe (Whitcher, 1975; Dummett, 1978; Male, 1992).

Economic significance

The main mineralisation at Moonmera forms a semi-circular zone in altered and fractured rocks of the Bundaleer Tonalite surrounding the porphyritic granodiorite. Chalcopyrite, pyrite and molybdenite occur mainly along fractures, but are also disseminated. An inner zone of chalcopyrite-molybdenite is surrounded by successive zones dominated by chalcopyrite-pyrite and pyrite (Dummett, 1978). The mineralisation is associated with quartz-sericite-kaolinite alteration (Ayres, 1974). Secondary biotite is abundant, with K-feldspar and anhydrite at depth. A different style of mineralisation, worked by a small open cut, consists of coarse aggregates of chalcopyrite and pyrite filling voids in breccia formed during emplacement of the younger granodiorite porphyry (Figure 327). Intrusion of the porphyritic granodiorite-breccia complex was responsible for the alteration, fracturing and mineralisation at Moonmera.

Undivided Bouldercombe Igneous Complex (PRgb)

A small area to the south of the Kabra Quartz Monzodiorite is distinguished from the Bundaleer Tonalite because of its higher response on images derived from airborne radiometric and magnetic survey data. This area was not visited during the Yarrol project fieldwork, and therefore has been mapped as undivided Bouldercombe Igneous Complex. It appears to be intruded by the Kabra Quartz Monzodiorite, but the relationship with the Bundaleer Tonalite is not clear (Figure 328).

Kabra Quartz Monzodiorite (Rgbk)

Introduction

The Kabra Quartz Monzodiorite is clearly defined as a distinct phase in the eastern lobe of the Bouldercombe Igneous Complex by airborne magnetic data.

Distribution

The intrusion forms an elongate mass 10km long and 4km wide extending from the cattle saleyards at Gracemere in the north to Table Mountain in the south. It is centred about 15km south-west of Rockhampton (Figure 329).

Derivation of name

The name is taken from the township and railway siding of Kabra in the north-western part of the intrusion.

Figure 327. Chalcopyrite and pyrite filling voids in breccia, small open cut near old railway line from Kabra to Mount Morgan, Moonmera Porphyritic Granodiorite

Type area

The type area is Table Mountain, where good exposures can be seen as boulders and in cuttings along the road to the tower on the summit.

Topographic expression

The deeply weathered northern end of the intrusion forms low country traversed by the Capricorn Highway. In contrast, the southern portion is hilly and includes the steep sided Table Mountain.

Geophysical expression

The radiometric response is varied, and can be correlated with the topography. The northern, deeply weathered section gives a low response, whereas the southern hilly part gives a moderate response dominated by potassium.

The magnetic pattern of the Kabra Quartz Monzodiorite on images derived from aerial data is unusual. Most of the intrusion has a moderate to high magnetic signature, consistent with magnetic susceptibility readings averaging 2840 x 10^{-5} SI units. However, a band of very low magnetism runs down the centre of the body. Readings taken within this zone have higher than average susceptibility, so the anomaly must be due to reversed remanent magnetism, suggesting either that the intrusion is composite, or that a reversal occurred during crystallisation.

Lithology and petrography

Table Mountain, the best exposed part of the pluton, is composed of hornblende quartz monzodiorite.

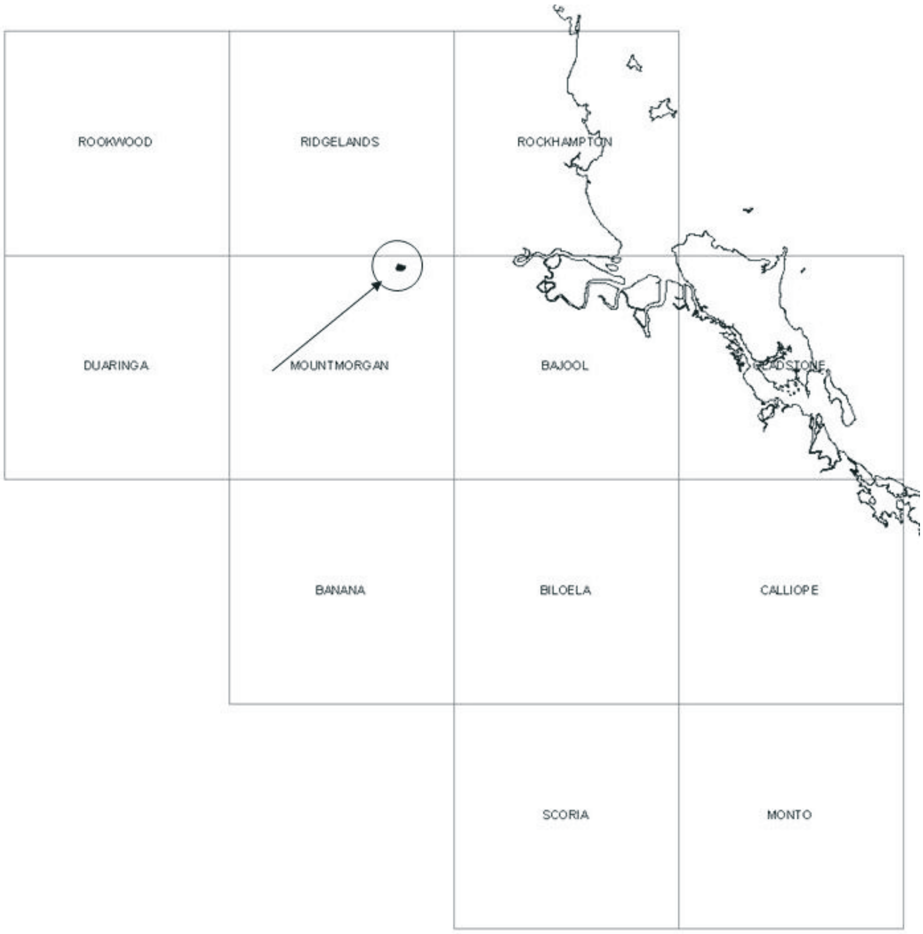


Figure 328. Distribution of the undivided Bouldercombe Igneous Complex (PRgb)

Subhedral zoned laths of calcic andesine are the dominant constituent, making up from 65 to 70% of the rock. Cores are cloudy and altered to epidote. Green hornblende is the main mafic mineral, comprising 10 to 15% of the rock. It replaces augite, which forms relict cores. A micrographic intergrowth of quartz and K-feldspar (15 to 20%) fills interstices between plagioclase laths. In many cases, the K-feldspar is optically continuous with the outermost most sodic zones of the plagioclase grains. Minor and accessory minerals include opaque grains, locally abundant sphene, apatite and zircon.

On the southern side of Table Mountain, the rocks are more mafic. Plagioclase is labradorite with dusty cores crowded with small opaque inclusions. The proportions of quartz and particularly K-feldspar are lower. Some specimens contain stumpy prisms of augite and elongate prisms of hypersthene rimmed by hornblende, together with relatively abundant biotite.

Samples from the reversely magnetised band in the centre of the intrusion are also quartz gabbro in composition. Subhedral to euhedral laths of sodic labradorite are the main constituent, with augite as the chief mafic mineral. Fine grained aggregates of pale secondary amphibole fill the interstices between plagioclase laths. Quartz makes up less than 5% of the rock, and locally is intergrown with minor K-feldspar at the margins of plagioclase grains. Opaques are more abundant than in the typical quartz monzodiorite.

Geochemistry

Analyses of the Kabra Quartz Monzodiorite cover a very restricted range of SiO_2 contents, from 59.1 to 63.3%. Alteration is responsible for low contents of K_2O and Rb in some samples, and for non-linear variation. The intrusion is placed in the Craiglands group, but is lower in Rb (Figure 304). In the case of two samples with very low Rb, this reflects alteration. Whether the lower Rb content of the remaining three samples can also be attributed to alteration is less certain.

Age

No age determinations have been carried out on the Kabra Quartz Monzodiorite. It has been assigned a Permian to Triassic age on the basis of its relationships with other intrusive units.

Figure 329. Distribution of the Kabra Quartz Monzodiorite

Stratigraphic relationships

The Kabra Quartz Monzodiorite intrudes the Raspberry Creek Formation, Mount Alma Formation, Rockhampton Group, and Lorray Formation. The magnetic pattern indicates that the intrusion cuts both the Bundaleer Tonalite and the Gracemere Gabbro, and probably a small area of undivided Bouldercombe Igneous Complex to the south.

Economic significance

Two small magnetite-rich skarns are developed in the Rockhampton Group at the contact with the Kabra Quartz Monzodiorite.

Craiglands Quartz Monzodiorite (Pgcr)

(C.G. Murray)

Introduction

The Mount Gerard Complex, about 35km north of Biloela, was mapped by Dear & others (1971) as a single, continuous intrusion more than 20km long and up to 10km across, and described simply as diorite and gabbro cut by syenite dykes. Interpretation of airborne magnetics and radiometrics showed that: (1) the mapped boundary as shown on the Monto 1:250 000 Sheet is inaccurate; (2) there are obvious subdivisions within the complex; and (3) the complex gives too great a radiometric response, particularly for potassium, to be entirely or even partly of gabbroic composition.

The present mapping program has confirmed these interpretations. Plutonic rocks in fact make up only a relatively small proportion of the Mount Gerard Complex as originally mapped, and are quartz monzodiorite (here defined as the Craiglands Quartz Monzodiorite) rather than diorite or gabbro. The central part of the complex is a sedimentary–volcanic unit, the Smoky beds, which divides the quartz monzodiorite into western

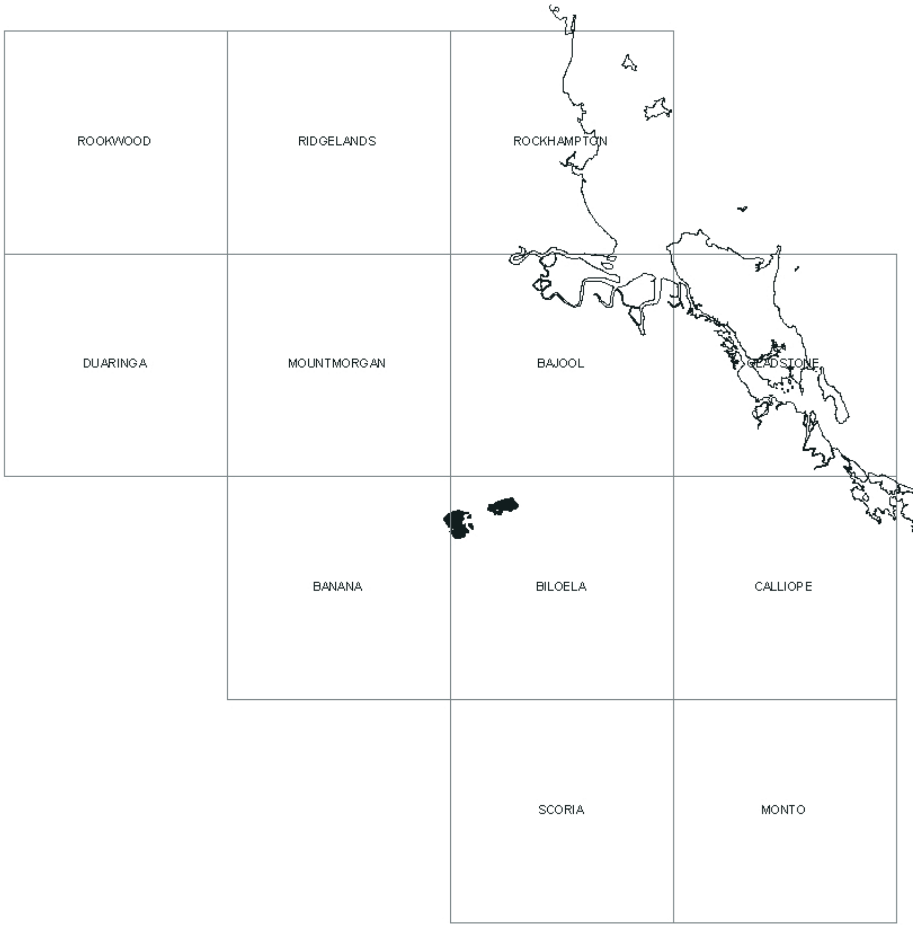


Figure 330. Distribution of the Craiglands Quartz Monzodiorite

and eastern sections. The only rock of mafic composition is a small plug of foliated gabbro which straddles Smoky Creek at the western contact of the main monzodiorite body.

Distribution

The Craiglands Quartz Monzodiorite is located 30 to 35km north of Biloela. The larger western segment extends from Smoky Creek in the south to Pumpkin Creek in the north, and covers an area 7km long and up to 4.5km wide. The slightly smaller eastern part extends from the headwaters of Gerard Creek in the west to Cattle Creek, a tributary of the Don River, in the east, and covers an area 7km by 3km. Small isolated outcrops are also exposed along Smoky and Pumpkin Creeks (Figure 330).

Derivation of name

The name is derived from the property Craiglands, which includes almost the entire outcrop area of the unit.

Type area

The type area is along Spring Creek upstream for 2km from the point where it is crossed by the station track leading to Craiglands homestead.

Topographic expression

The topographic expression ranges from low, open valleys to steep, forested hills. In the eastern part of the intrusion, deeply incised creeks follow the trend of andesite dykes that are obviously more easily eroded than the quartz monzodiorite.

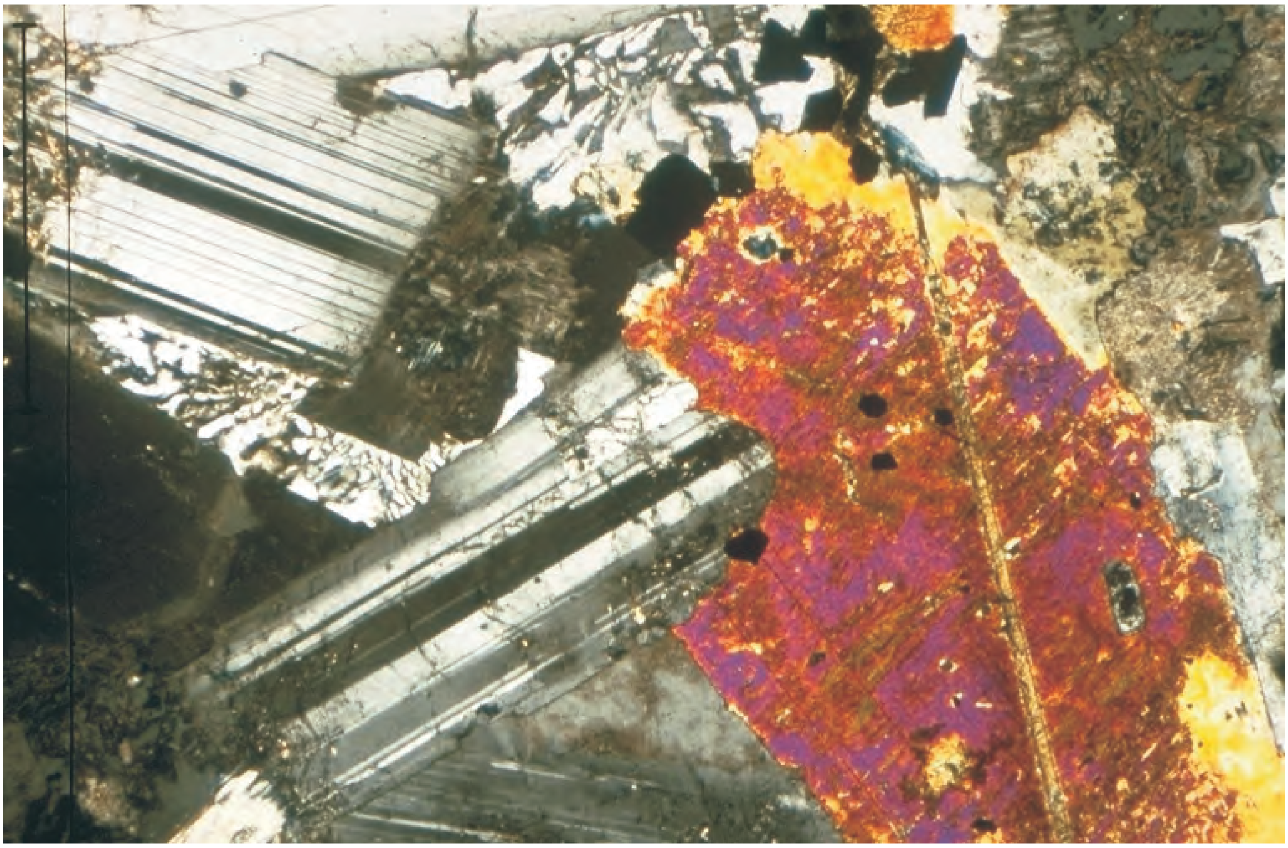


Figure 331. Photomicrograph showing micrographic intergrowths of K-feldspar and quartz rimming plagioclase laths, Craiglands Quartz Monzodiorite. Crossed nicols. Scale bar is 1mm long.

Geophysical expression

The Craiglands Quartz Monzodiorite gives a uniform moderate to strong radiometric response, and appears in pink tones on images generated from airborne data. It contrasts with the dark red shades of the Smoky beds, and the almost white colours of the surrounding Youlambie Conglomerate.

In contrast, the magnetic signature is variable. The western part of the monzodiorite coincides with a very strong positive magnetic anomaly that continues to the east beneath a thickening cover of Smoky beds (see Appendix 5). This pattern is consistent with magnetic susceptibility measurements in the field, which range from 1690 to 7240 x 10^{-5} SI units and average 4045 x 10^{-5}. Most of the eastern segment is much more weakly magnetic, and together with the Inverness Volcanics stands out as a distinct feature on images generated from the aeromagnetic data. This significant difference in magnetic signature is greater than would be expected from field measurements of magnetic susceptibility, which range from 1950 to 4515 x 10^{-5} SI units and average 2990 x 10^{-5} for the eastern segment of the Craiglands Quartz Monzodiorite.

The main exposure of the monzodiorite coincides with a positive anomaly along the prominent Marlborough Gravity Ridge of Lonsdale (1965). Other positive culminations along this ridge correspond to layered gabbros (Bucknalla and Eulogie Park Gabbros), suggesting that the Craiglands Quartz Monzodiorite may have a mafic root.

Lithology and petrography

The rocks are medium grained and grey to pink in colour. Compositions range from quartz gabbro to quartz monzonite, with biotite-augite-hornblende quartz monzodiorite and biotite-augite quartz monzogabbro most common. There appears to be no significant difference in composition between the two main segments of the intrusion.

Except in one sample of quartz monzonite, plagioclase is the main mineral. It forms subhedral to euhedral laths to 5mm long with large, uniform cores of calcic andesine or sodic labradorite and narrow zoned rims. Typically it comprises from 45 to 65% of the rock. K-feldspar occurs as rims on plagioclase, as interstitial grains, and in micrographic intergrowths with quartz (Figure 331). In the most mafic samples, only minor amounts of K-feldspar are present, and the rocks are quartz gabbros, but proportions of 5 to 20% are more usual, and the quartz monzonite contains about 40% as rims on plagioclase. Quartz is always present, and forms separate interstitial and anhedral grains, commonly associated with K-feldspar, as well as intergrowths. It ranges from minor amounts to at least 15% in some samples.

Mafic minerals constitute from 15 to 30% of the rocks. Overall, augite and hornblende are most abundant, and occur in approximately equal amounts, but their relative proportions in individual samples are quite variable. Augite is commonly replaced and rimmed by green hornblende, or in some cases by pale green actinolitic amphibole. In one sample, actinolitic amphibole that appears to have replaced augite is itself rimmed by brown hornblende. Biotite is rarely absent, but typically ranges from only minor amounts to 5%, and is largely altered to chlorite. It increases to 10%, and is equal to hornblende, in rocks from the western margin of the intrusion. Possible altered olivine is present locally. Opaque grains are relatively abundant, and comprise up to 5% of the rocks. Accessories include ubiquitous apatite, late stage sphene, and sparse zircon. Secondary minerals include epidote, sericite, chlorite, and prehnite.

Geochemistry

The Craiglands Quartz Monzodiorite is placed within the Craiglands group (Figure 304).

Age

U-Pb zircon dating of a sample from the western margin of the intrusion gave a result of 256.8±2.6Ma, or latest Permian (Appendix 1).

Stratigraphic relationships

The Craiglands Quartz Monzodiorite intrudes the Youlambie Conglomerate to the west and north. It intrudes and is overlain by the Smoky beds, airborne magnetic data indicating that the western part of the intrusion becomes more deeply buried to the east (see Appendix 5). It is intruded by a small gabbro plug that straddles Smoky Creek at the western contact. Its relationship to the Inverness Volcanics is unclear because of poor access and outcrop due to thick vegetation, but the similarity of chemical compositions suggests that the two units may be comagmatic.

Economic significance

Because of its initial description as diorite and gabbro, the intrusion has been explored (unsuccessfully) for platinum group metals. Several small gold deposits occur around the eastern end of the Craiglands Quartz Monzodiorite, and some stream sediment samples have given anomalous gold values. In view of this, the area may have some potential for gold mineralisation.

Mount Seaview Igneous Complex (PRgsv)

(C.G. Murray)

Introduction

The Mount Seaview Igneous Complex was mapped, but not named or described, by Dear & others (1971). It consists of an outer dioritic phase completely surrounding a more felsic core.

Distribution

The pluton is a triangular body with sides about 7km long, situated 40km south-west of Calliope, and forming part of the Calliope Range (Figure 332). The felsic core is ovoid in shape, elongate east-west, 4km long and 2.5km across.

Derivation of name

The name is from Mount Seaview, which is located within the central felsic phase of the intrusion.

Type area

The type area of the dioritic phase is along the main access road up the eastern escarpment of The Tableland. Good boulder outcrops of this phase can also be seen along the track that branches off westwards from this road and runs down Callide Creek towards Biloela. The type area of the felsic phase is Mount Seaview itself, and the lower ridge to the south-west known locally as Griffiths Hill.

Figure 332. Distribution of the Mount Seaview Igneous Complex

Topographic expression

The dioritic phase forms the plateau named The Tableland, with an average elevation of 500m and a steep eastern escarpment to the headwaters of Dan Dan Creek. This plateau is gently undulating, and has largely been cleared for grazing. The more resistant felsic phase forms steep and rugged slopes culminating in the prominent 815m peak of Mount Seaview (also known as Futters Hat) on the Calliope Range.

Geophysical expression

The two phases of the complex are clearly distinguished on images derived from airborne geophysical data. The felsic core has a high radiometric response in all channels, and appears white, in contrast to the low signature and dark tones of the dioritic phase. Conversely, the dioritic phase has a higher magnetic response than the granite, in accord with field measurements of magnetic susceptibility of 1805 x 10^{-5} SI units and 995 x 10^{-5} SI units respectively. The much stronger magnetic response of the southern section of the dioritic rim, part of which is concealed by volcaniclastic contact rocks, may reflect compositional variations. Troensegaard (1969) noted that the diorite in this area was more mafic-rich than elsewhere. No consistent differences were observed during the current mapping program.

Lithology and petrography

The mafic rim of the pluton consists mainly of grey, medium to coarse grained, augite-biotite-hornblende quartz diorite. Subhedral plagioclase laths make up about 65% of the rock, and have large cores of sodic labradorite or andesine composition with narrow zoned rims. Anhedral, interstitial quartz grains comprise up to 10% of the rock, and rare interstitial to poikilitic K-feldspar is present. The proportion of mafic minerals is fairly uniform at about 25%. Green hornblende is most abundant, replacing and rimming augite. The former presence of augite cores is also indicated by pale secondary amphibole. Biotite is as abundant as or even exceeds hornblende in some samples. Minor hypersthene is present. Accessory minerals include zircon (surprisingly abundant in one specimen), opaques, apatite, and sphene. Secondary epidote, calcite and sericite are locally developed. One sample from the eastern edge of the intrusion is a dark grey, medium grained,

Figure 333. Andesitic dykes cutting decomposed diorite, road cutting on the east slope of the Calliope Range, Mount Seaview Igneous Complex

hornblende-augite-biotite-hypersthene quartz gabbro in which hypersthene is the main mafic mineral. The diorite is cut by andesitic dykes (Figure 333).

The felsic core is a pink, medium-grained, altered, leucocratic biotite granite to granodiorite. Plagioclase comprises about half the rock as subhedral to euhedral laths of sodic andesine up to 5mm long, extensively altered to sericite and calcite. Quartz ranges from 20 to 30%, and has a bimodal size distribution in some samples, with one population of equant grains averaging about 1mm in diameter, and another about 0.1mm. The proportion of K-feldspar is difficult to estimate, but averages about 15%. Biotite never makes up more than 10% of the rock, and most grains are at least partially altered to chlorite and epidote. Accessories include opaques, apatite, sphene and zircon. Pyrite occurs locally along joint planes. A zone of coarse breccia is developed on Griffiths Hill.

Geochemistry

The Mount Seaview Igneous Complex is similar to the Miriam Vale Granodiorite except for higher levels of Sr, particularly in the more mafic rocks (Figure 334).

Age

U-Pb dating of zircon from a sample of the felsic phase gave a bimodal age distribution (Appendix 1). The older population gives a result of 258±4.3Ma, and the younger population 238.3±6.9Ma. Because zircon is relatively abundant in some samples of quartz diorite, it is possible that these dates represent the times of crystallisation of the mafic and felsic phases of the complex, respectively. The combined populations give a result of 252.6±5.3Ma, close to the Permian–Triassic boundary, and a late Permian to Early Triassic age is assigned to the Mount Seaview Igneous Complex.

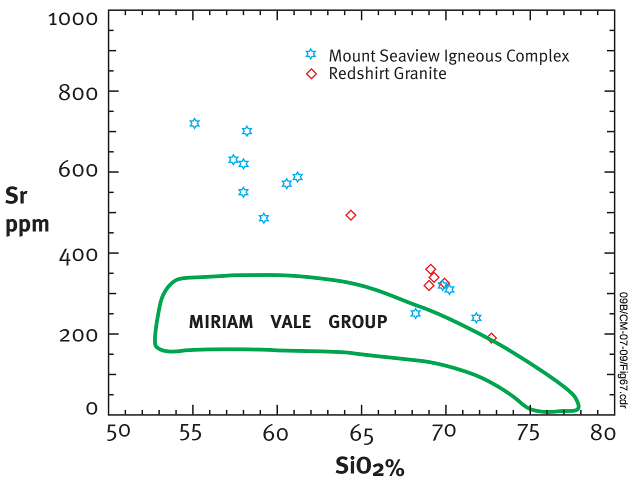


Figure 334. Sr contents of the Mount Seaview Igneous Complex and the Redshirt Granite compared to the Miriam Vale Granodiorite. Analyses from GSQ and B. Chappell & R. Bultitude (unpublished).

Stratigraphic relationships

The felsic core appears to be completely encircled by the mafic rim. Although no contacts were seen, this arrangement suggests that the quartz diorite was emplaced first, and was intruded by the granite to granodiorite phase.

The quartz diorite intrudes the Three Moon Conglomerate, Rockhampton Group, and Youlambie Conglomerate. A small outlier of Jurassic Precipice Sandstone overlies granite of the felsic phase on Griffiths Hill (Denmead, 1932), indicating that the intrusion had been unroofed by the end of the Triassic. Poorly consolidated gravel sourced from Mount Seaview overlies quartz diorite near the base of the eastern escarpment of the Calliope Range, and produces a radiometric anomaly. This gravel is being eroded by present day streams, and is considered to be of Tertiary to Quaternary age (**TQrh**).

Economic significance

Most of the recorded production from the Barmundoo Goldfield comes from the area around Mount Seaview, particularly Griffiths Hill. The gold deposits are mainly confined to the felsic phase of the intrusion, and include stockworks of thin auriferous quartz veins, and veins associated with dykes.

Galloway Plains Igneous Complex

(C.G. Murray)

The name Galloway Plains Tonalite was applied by Dear & others (1971) to a small batholith straddling the Dawson Highway and the Gladstone–Moura railway about 45km north-east of Biloela (Figure 335). They described it as consisting solely of grey tonalite except for a small outlier of Muncon Volcanics forming the summit of Mount Redshirt. Interpretation of airborne geophysical data and field studies show that the batholith comprises a number of plutons ranging in composition from gabbro to granite. One unnamed and seven named units are defined within the Galloway Plains Igneous Complex. In common with other composite intrusions in the project area, mafic rocks are restricted to the margins of the batholith, and granites occur in the centre.

Most of the mass gives a low radiometric response, as would be expected from a tonalite. The glaring exception is an area centred on Mount Redshirt, which is not Muncon Volcanics, but a biotite granite, locally tourmaline bearing, which is differentiated as the Redshirt Granite. Aeromagnetic data show that the Galloway Plains Igneous Complex consists of two contrasting parts, the western more strongly magnetic than the eastern. There is reasonable outcrop over much of the western part which is divided into the Rocky Point Granodiorite and a central mass of Dumgree Tonalite. Locally these intrusions are cut by north-east trending rhyolite dyke swarms related to the Redshirt Granite. The Rocky Point Granodiorite is also intruded by the Wyalla Granite. The eastern part of the Galloway Plains Igneous Complex is low lying and deeply weathered, and is largely devoid of fresh outcrop. This part, named the Bocoolima Granodiorite, appears to consist of a series of nested plutons indicated by arcuate to elliptical magnetic rims. It is cut by prominent north-west to

Figure 335. Distribution of the Galloway Plains Igneous Complex

north-north-west trending dykes of generally andesitic composition, and is also intruded by the Voewood Granite, which has closely spaced joints filled with pyrite-bearing quartz veins.

Strongly magnetic diorite and gabbro locally occur around the periphery of the Dumgree Tonalite and Rocky Point Granodiorite, and may represent marginal phases. The prominent magnetic feature associated with the northernmost diorite–gabbro phase (the Sawnee Gabbro) continues to the south-east into a linear magnetic high which is related to metamorphosed contact rocks including limestone.

Bocoolima Granodiorite (PRgab)

Introduction

The eastern part of the Galloway Plains Igneous Complex is very deeply weathered, and exposures of fresh rock are lacking. It is likely that the rock type is a biotite-rich granodiorite or tonalite, and that the deep weathering is a reflection of biotite-induced grussification (Isherwood & Street, 1976). Although variations may be masked by the lack of outcrop, this part of the complex has been mapped as a single unit, the Bocoolima Granodiorite, intruded by the relatively small Voewood Granite.

Distribution

The Bocoolima Granodiorite is a large ovoid pluton 25km long and 15km wide centred about 30km west-south-west of Calliope (Figure 336). The Dawson Highway and Gladstone-Moura railway run through the northern half of the intrusion.

Derivation of name

The name comes from the property Bocoolima, located in the southern part of the granodiorite.

Figure 336. Distribution of the Bocoolima Granodiorite

Type area

The type area is along the headwaters and tributaries of Bell Creek south of Blackfellow Mountain on the Calliope Range. Other typical outcrops occur at the base of the eastern side of a prominent ridge of metamorphic rocks 2km south-south-west of Rockfield homestead.

Topographic expression

Most of the Bocoolima Granodiorite forms low, flat to gently undulating country that has largely been cleared for grazing. The southern contact is marked by the steep escarpment of the Boyne Range, representing a rim of hornfelsed metasediments, and is partly obscured by poorly cemented gravels and coarse sands representing colluvial floodout sheets of Tertiary to Quaternary age (**TQrh**).

The south-western extremity of the intrusion forms the Calliope Range, and is characterised by rugged hills with a prominent joint-controlled rectilinear drainage pattern. The steep eastern scarp of the Calliope Range has sourced extensive sheets of poorly cemented granitic colluvium (**TQrh**) incised by present day stream channels.

Geophysical expression

The extensive, topographically subdued part of the Bocoolima Granodiorite gives a low radiometric response, partly due to the deep weathering, and it also has a relatively low magnetic signature, consistent with magnetic susceptibility readings on fresh material averaging 1125 x 10^{-5} SI units. In this part of the intrusion, a prominent arcuate pattern with a more magnetic rim suggests that the intrusion may constitute at least two nested plutons. This feature forms a semicircle surrounding the western section of the Voewood Granite and at a distance of about 2km from it. An almost circular positive magnetic anomaly at the western boundary just south of the Calliope River was identified as a potential target (the White Cane prospect) and drilled by an exploration company, but only normal granodiorite to tonalite was encountered. Cross-cutting north-west to north-north-west trending andesitic dykes are strongly magnetic and intrude both the Bocoolima Granodiorite and its contact rocks.

The south-western part of the intrusion forming the Calliope Range has a higher response for both radiometrics and magnetics, with magnetic susceptibility readings averaging 1855 x 10^{-5} SI units. On images, a south-east trending arcuate belt parallel to the contact of the intrusion with the Youlambie Conglomerate south-west of Blackfellow Mountain gives a very low magnetic response. The fact that *in situ* magnetic susceptibility readings from an outcrop within this belt are similar to those from adjoining magnetic highs suggests that this magnetic low may reflect a significant reversely polarised component.

The intrusion corresponds to a negative gravity anomaly that forms the southern end of the Bajool Gravity Low of Lonsdale (1965). There is a steep gradient between this low and the gravity high associated with the Craiglands Quartz Monzodiorite to the west, part of the Marlborough Gravity Ridge.

Lithology and petrography

Fresh exposures of the Bocoolima Granodiorite are confined to the south-western corner of the intrusion, and occur along the crest and western slopes of the Calliope Range, and to small outcrops along the eastern side of a ridge of metamorphic rocks south-west of Rockfields homestead. Hornblende-biotite granodiorite is virtually the only rock type.

Subhedral to euhedral plagioclase laths make up 50 to 60% of the rock, with local development of myrmekitic intergrowths at contacts with K-feldspar. Anhedral quartz grains constitute 15 to 20%, and interstitial to poikilitic K-feldspar from 10 to 15%. Green hornblende with relict augite cores is less abundant than biotite; together they comprise from 10 to 20% of the rock. Minor and accessory minerals are opaque grains, apatite, sphene and zircon, and a little secondary muscovite occurs locally.

Geochemistry

The Bocoolima Granodiorite is one of the defining units of the Monal group (Figure 305). It is clearly distinguished from the Rocky Point Granodiorite and Dumgree Tonalite by its higher content of TiO_2, total alkalis, K_2O, Ce, Rb and Zr, and its lower CaO.

Age

No age determinations have been carried out. The Bocoolima Granodiorite is assigned a Permian to Triassic age based on its relationships with other phases of the Galloway Plains Igneous Complex.

Stratigraphic relationships

The Bocoolima Granodiorite intrudes the Three Moon Conglomerate, Mount Alma Formation, Rockhampton Group and Youlambie Conglomerate. It is intruded by the Voewood Granite, which has been dated as Middle to Late Triassic (233.6±6.8Ma; Appendix 1). The shape of the contact suggests that it is intruded by the Rocky Point Granodiorite, although complex magnetic patterns across the boundary are difficult to interpret. A small lens of garnet-clinopyroxene-plagioclase hornfels along the contact is interpreted as a screen of country rock partially separating the two intrusions.

Economic significance

The Maxwelton Goldfield worked auriferous quartz veins in the Three Moon Conglomerate near its contact with the Bocoolima Granodiorite.

Unnamed gabbro (PRgau)

Introduction

A small mafic intrusion forms the western extremity of the Galloway Plains Igneous Complex near the headwaters of Back Creek.

Distribution

The intrusion is centred about 6.5km west of Mount Redshirt, and is 4km from north to south with an average width of 1.5km (Figure 337).

Topographic expression

The unit forms low hills with sparse bouldery outcrops and abundant surface rubble.

Figure 337. Distribution of the unnamed gabbro (PRgau)

Geophysical expression

The mafic intrusion is expressed as a very strong and sharply defined magnetic high on images generated from airborne data. This is consistent with high *in situ* magnetic susceptibility readings that average 5165 x 10^{-5} SI units, more than twice the adjacent Dumgree Tonalite. The magnetic high extends further to the north-east than the mapped outline of the mafic body, which is based mainly on the distribution of the characteristic low radiometric signature.

Lithology and petrography

The intrusion consists of medium-grained quartz-augite-hypersthene-hornblende gabbro. Plagioclase forms from 65 to 70% of the rock and occurs as subhedral laths with dusty cores of labradorite composition (An_{62}) and clear rims. Pyroxene makes up 15% of the rock, and consists of roughly equal proportions of augite and hypersthene rimmed and replaced by subophitic to interstitial hornblende (10%). Opaques constitute up to 5% of the rock, and quartz less than 5%. Minor biotite is present locally.

Geochemistry

No samples from this intrusion have been analysed.

Age

No samples have been dated. A Permian to Triassic age is assigned based on results from other units in the Galloway Plains Igneous Complex.

Stratigraphic relationships

The unit intrudes the Youlambie Conglomerate, and is interpreted to intrude the Inverness Volcanics. No contacts were seen with the Dumgree Tonalite or Rocky Point Granodiorite, and the relationship with these two units is unknown.

Economic significance

No mineralisation is known to be associated with the intrusion.

Sawnee Gabbro (PRgas)

Introduction

The Sawnee Gabbro is a marginal mafic unit that forms a narrow belt along the north-western extremity of the Galloway Plains Igneous Complex. Small separate mafic intrusions just to the north may be related to the Sawnee Gabbro.

Distribution

The Sawnee Gabbro forms a continuous east-west belt about 8km long and up to 1km wide between the headwaters of the Don and Calliope Rivers (Figure 338).

Derivation of name

The name is derived from Mount Sawnee, a local name for a peak in the central part of the outcrop area.

Type area

The type area comprises the ridges west of the Calliope River leading up to Mount Sawnee, where good bouldery outcrops can be seen.

Topographic expression

The gabbro forms hilly country with steep, sharp ridges near the Calliope River, and more subdued topography further west.

Figure 338. Distribution of the Sawnee Gabbro

Geophysical expression

The Sawnee Gabbro coincides with a strong positive magnetic anomaly, consistent with susceptibility readings on outcrops that average 5750 x 10^{-5} SI units, and gives a subdued radiometric response.

Lithology and petrography

The most common rock type is a hornblende gabbro, with lesser amounts of hornblende-hypersthene-augite gabbro.

Plagioclase is the dominant mineral, making up from 55 to 65% of the rocks as subhedral laths of labradorite up to 4mm long, aligned in some specimens.

Hornblende comprises up to more than 40% of the rocks in the more mafic samples, and occurs as green subophitic to poikilitic grains to 10mm across. In the pyroxene-rich rocks, containing up to 30% combined augite and hypersthene, hornblende merely forms rims on pyroxene crystals.

Opaque minerals comprise up to 5% of the rocks, and minor sphene, apatite and secondary epidote are also present.

Geochemistry

Because of its mafic composition, it is not possible to assign this intrusion to a geochemical group.

Age

The Sawnee Gabbro is assumed to represent one of the earliest phases of the Galloway Plains Igneous Complex, and is assigned a Late Permian to Early Triassic age.

Stratigraphic relationships

The gabbro intrudes the Balaclava Formation to the north and the Youlambie Conglomerate to the west. Spatial relations suggest that it is intruded by the Rocky Point Granodiorite, but no contacts were seen to confirm this.

Economic significance

The Sawnee Gabbro has been investigated as a source of magnetite, but no concentrations of economic significance were found. It is associated with small garnetiferous skarn deposits in the Balaclava Formation, some of which contain minor base metals.

Rocky Point Granodiorite (PRgar)

Introduction

On a broad scale, the western part of the Galloway Plains Igneous Complex appears to be a single ovoid pluton oriented north-south that ranges in composition from tonalite to granodiorite, and is intruded by the prominent Redshirt Granite. However, geophysical images show some distinct variations, such as a stronger radiometric response in the southern part of the pluton, and a higher magnetic signature in the south-west. Chemical analyses also form two significantly different trends. On this basis, the pluton has been subdivided into the Dumgree Tonalite in the centre, surrounded by the Rocky Point Granodiorite. Although this subdivision reflects broad compositional trends, it is almost certainly too simplistic. For example, there are areas of quartz diorite to tonalite within the Rocky Point Granodiorite. It is anticipated that detailed mapping and geochemical sampling may reveal more complex variations in composition and intrusive history.

Distribution

The Rocky Point Granodiorite forms an ovoid pluton 20 by 13km surrounding Mount Redshirt, 40km north-north-east of Biloela (Figure 339). The Dawson Highway cuts across the southern part of the intrusion.

Derivation of name

The name is derived from Rocky Point Mountain on the Calliope Range.

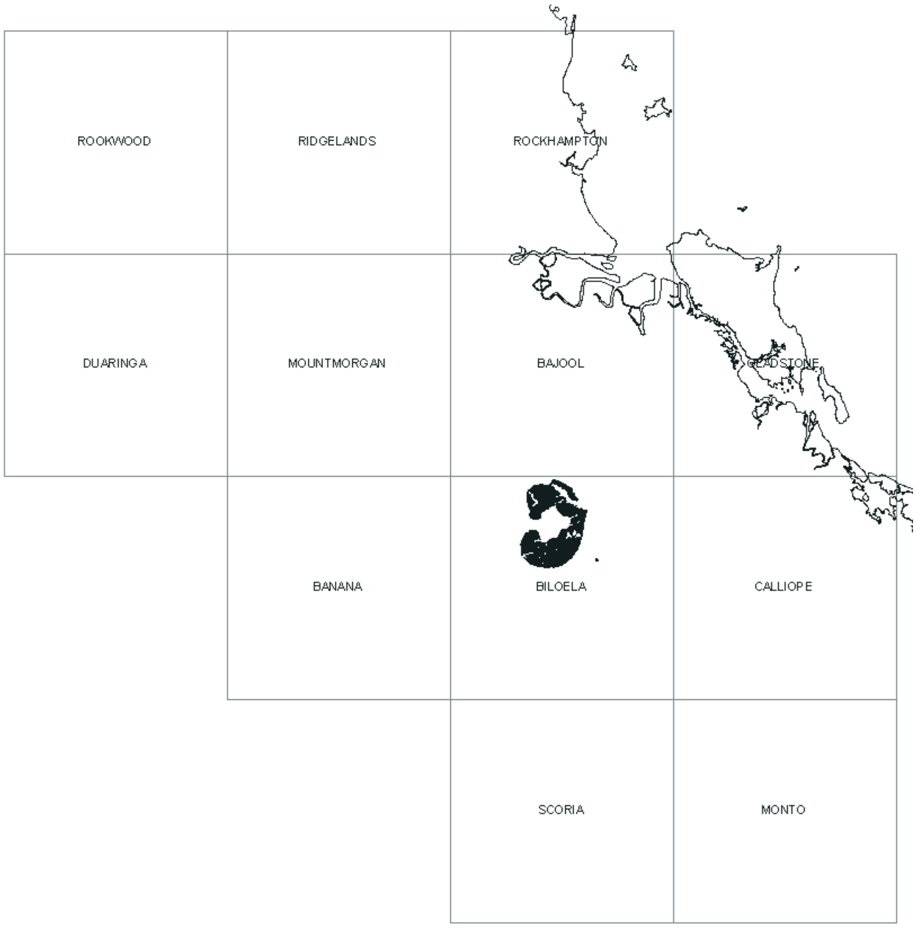


Figure 339. Distribution of the Rocky Point Granodiorite

Type area

The type area is along Bell Creek and its tributaries north and north-east of Mount Buckland.

Topographic expression

The topography of the Rocky Point Granodiorite is quite varied. Much of the southern part of the intrusion is undulating hills with sparse outcrop as tors and boulders. Slopes increase towards the southern contact with more resistant metasedimentary rocks. Sheets of poorly cemented granitic colluvium (**TQrh**) have accumulated at the base of some of the steeper slopes, and are incised by present day stream channels. In this area, a capping of Tertiary basalt has protected the granodiorite from erosion (Figure 340). Only remnants of this capping remain, but the presence of an outlier at the head of a tributary of Back Creek 10km to the west suggests that it was originally very extensive.

The northern part of the Rocky Point Granodiorite is divided into two sections by the steep east-facing scarp of the Calliope Range. The low-lying eastern section is partly covered by the alluvial flats of the Calliope River, and has very sparse outcrop. The pronounced difference in elevation, about 200m, results from the greater erosive power of eastward flowing streams and the resistant nature of the Redshirt Granite and its associated dyke swarm.

Geophysical expression

Overall, the Rocky Point Granodiorite gives a moderately strong magnetic response, consistent with *in situ* magnetic susceptibility measurements averaging 2380 x 10^{-5} SI units, a little higher than the Dumgree Tonalite. On images, a broad belt along the south-western margin of the intrusion gives the strongest response, and a small area south-west of Rocky Point Mountain is very weakly magnetised, possibly due to a significant reversely polarised remanent component.

Figure 340. Tertiary basalt overlying decomposed Rocky Point Granodiorite, 3km south-west of Mount Buckland

A prominent linear feature marking the eastern limit of the typical magnetic signature of the Rocky Point Granodiorite is taken as the eastern boundary with the Bocoolima Granodiorite, which generally has lower magnetic susceptibility. However, an arcuate magnetic feature, concave to the east, just west of the boundary suggests the possibility of overlapping plutons at depth.

The Rocky Point Granodiorite, particularly the southern part, has a stronger radiometric response than the Dumgree Tonalite.

The gravity effect of the intrusion is masked by its location on a steep gravity gradient between the high centred on the Craiglands Quartz Monzodiorite to the west and the low over the Bocoolima Granodiorite and Voewood Granite to the east.

Lithology and petrography

The main rock type is biotite-hornblende granodiorite, with subordinate augite-biotite-hornblende tonalite. An area of pyroxene-biotite-hornblende quartz diorite about 2km by 1km extends south-south-west from Rocky Point Mountain.

Granodiorites and tonalites contain from 40 to 60% plagioclase as subhedral to euhedral laths zoned from uniform labradorite-andesine cores to narrow oligoclase rims. Some myrmekite is developed at contacts with K-feldspar.

Quartz (10 to 25% of the rocks) and K-feldspar (5 to 20% of the rocks, averaging about 15%) both form interstitial to poikilitic crystals that are up to 10mm across.

Green-brown hornblende is the main mafic mineral, comprising about 7 to 15% of the rocks. In some specimens it rims augite or aggregates of secondary pale green amphibole replacing augite. Biotite ranges from 5 to 10%.

Minor and accessory minerals include opaque grains, apatite, sphene and zircon.

The quartz diorite and minor quartz gabbro near Rocky Point Mountain are fairly uniform in composition, consisting of from 65 to 75% plagioclase and about 10% quartz. The plagioclase forms subhedral laths with narrow, clear sodic rims on dusty cores that are labradorite or labradorite-andesine in composition. Some specimens exhibit strong alignment of plagioclase laths giving the rocks a pronounced foliation. The main mafic minerals are green-brown hornblende (10 to 15% of the rocks) and biotite (5 to 15%). Augite occurs both as cores in hornblende and discrete grains, and hypersthene accompanies augite in some specimens.

Geochemistry

The Rocky Point Granodiorite, along with the Dumgree Tonalite, is one of the defining units of the Dumgree group, characterised by low total alkalis, TiO_2, MgO and high Al_2O_3, CaO and Sr. K_2O, Ce, Rb, Th, Y and Zr are all low and for the group as a whole tend to decrease with increasing SiO_2 (Figure 302). For most elements, the Rocky Point Granodiorite does not form well defined linear trends on Harker diagrams. It is distinguished geochemically from the Dumgree Tonalite by higher TiO_2, MgO, K_2O, Ce, Rb, Th, V, Y and Zr and lower Al_2O_3, Na_2O and Sr. Contents of most elements are very similar to those of the Umbrella Granodiorite of the Bouldercombe Igneous Complex.

Age

Webb & McDougall (1968) dated two samples from the western part of the Galloway Plains Igneous Complex. Sample GA1166 from the southern part of the Rocky Point Granodiorite in the headwaters of Collards Creek gave K-Ar dates of 244±5Ma (biotite) and 247±5Ma (hornblende), and sample GA1167 located on the Dawson Highway near the interpreted boundary between the Dumgree Tonalite and Rocky Point Granodiorite gave 244±5Ma (biotite) and 240±5Ma (hornblende), suggesting an Early Triassic age and supporting the interpretation that emplacement of these two units was essentially synchronous. The Redshirt Granite, which intrudes both, gave a U-Pb zircon date of 251±4Ma, and therefore all three have been assigned a Late Permian to Early Triassic age.

Stratigraphic relationships

The Rocky Point Granodiorite intrudes the Youlambie Conglomerate, Rockhampton Group, Three Moon Conglomerate, Balaclava Formation, Mount Alma Formation and Raspberry Creek Formation, and is overlain by Tertiary basalt. It is intruded by the Redshirt and Wyalla Granites. No intrusive relationships have been observed between the Rocky Point Granodiorite, Dumgree Tonalite and Bocoolima Granodiorite. As presently interpreted, spatial relationships suggest that the Rocky Point Granodiorite intrudes the Bocoolima Granodiorite and is intruded by the Dumgree Tonalite. A small lens of garnet-clinopyroxene-plagioclase hornfels occurs along the contact with the Bocoolima Granodiorite and is considered to be a screen of country rock partially separating the two intrusions.

Relationships between phases in the Rocky Point Granodiorite are clearly displayed at an outcrop in Bell Creek at the margin of the quartz diorite at GR274100 7325050. Dark grey medium-grained hypersthene-augite-biotite-hornblende quartz gabbro is intruded by and occurs as blocks in foliated grey biotite-hornblende quartz diorite (Figure 341). Both gabbro and diorite are cut by dykes and small masses of pink to grey locally leucocratic biotite granodiorite.

Economic significance

The Mount Rainbow Goldfield lies within the southern part of the Rocky Point Granodiorite. Most of the production came from deep leads in sediments at the base of Tertiary basalt flows, but the gold was sourced from quartz veins in the granodiorite itself or in the adjacent country rocks. Some basalt flows lie directly on weathered granodiorite (Figure 340).

The size of workings in the Fig Tree Provisional Goldfield north of Mount Redshirt suggests that some production took place, although none is recorded. In the Annual Report of the Department of Mines for 1895, it was reported that mines in this field were abandoned when the yield fell from 120g/t to 45g/t gold, the grade then being too low to pay for cartage. The gold mineralisation was probably related to the prominent north-east trending dyke swarm associated with the Redshirt Granite, rather than to the Rocky Point Granodiorite itself.

Dumgree Tonalite (PRgad)

Introduction

The Dumgree Tonalite occupies the central part of the ovoid pluton which forms the western third of the Galloway Plains Igneous Complex.

Distribution

The Dumgree Tonalite is an equidimensional mass about 8km across located south-west of Mount Redshirt 40km north-east of Biloela (Figure 342).

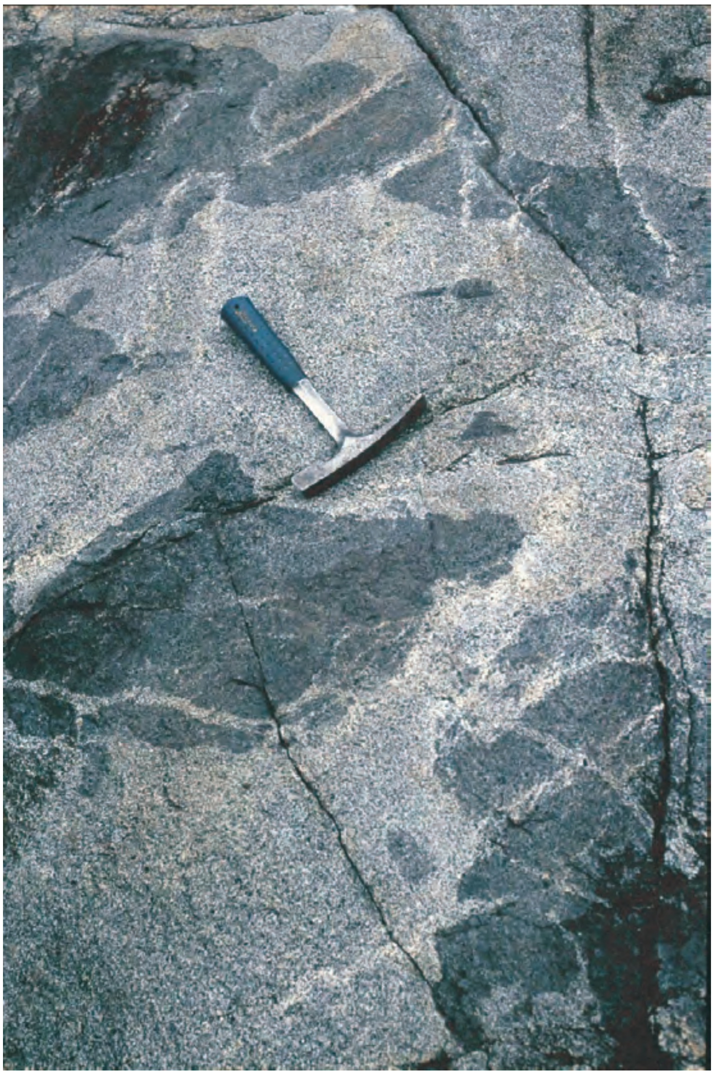


Figure 341. Inclusions of gabbro in foliated quartz diorite, Bell Creek 4km south-south-east of the Dawson Highway, Rocky Point Granodiorite

Derivation of name

The name is derived from the property of Dumgree that extends over much of the outcrop area of the unit. The intrusion also lies largely within the northern part of the Parish of Dumgree.

Type area

The type area is along the Gladstone–Moura railway line from 1.5km south of Mount Redshirt to 2km east-north-east of Mount Rainbow siding, where the tonalite is well exposed in numerous cuttings.

Topographic expression

The Dumgree Tonalite is divided into two sections by the steep east-facing scarp of the Calliope Range. The larger western part forms undulating hills with scattered outcrop as tors and large boulders. The low-lying eastern section is partly covered by the alluvial flats of Spring Creek, a tributary of the Calliope River, and has very sparse outcrop. The pronounced difference in elevation, about 200m, results from the greater erosive power of eastward flowing streams and the resistant nature of the Redshirt Granite.

Geophysical expression

The Dumgree Tonalite appears as a fairly uniform, moderately strong magnetic body on images derived from aeromagnetic data, consistent with *in situ* magnetic susceptibility readings averaging 2000 x 10^{-5} SI units. The cause of local strongly magnetic responses around the Redshirt Granite is unknown.

The Dumgree Tonalite gives a comparatively low radiometric response for a felsic intrusive unit, consistent with its composition.

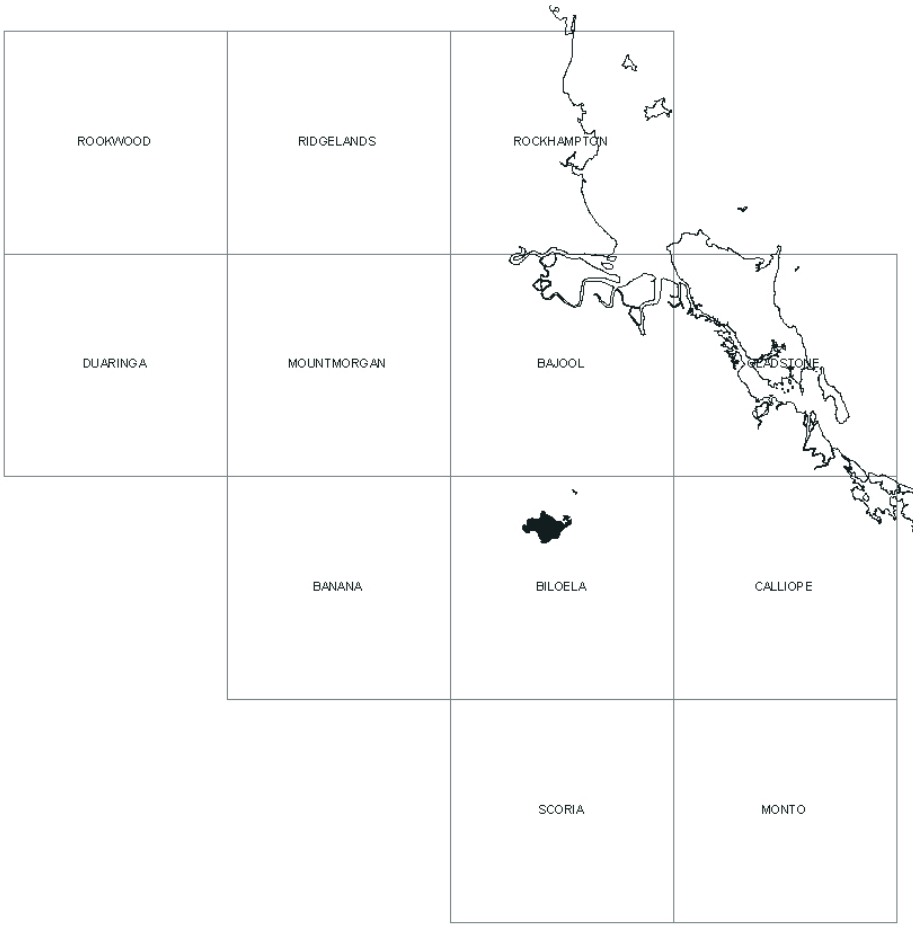


Figure 342. Distribution of the Dumgree Tonalite

Together with the Rocky Point Granodiorite, the Dumgree Tonalite lies on a steep gravity gradient between the high centred on the Craiglands Quartz Monzodiorite to the west and the low over the Bocoolima Granodiorite and Voewood Granite to the east.

Lithology and petrography

The Dumgree Tonalite ranges from quartz diorite to granodiorite in composition. Tonalite is dominant, and some samples are so low in mafics that they can be classed as trondhjemites. The more mafic varieties come from the western edge of the pluton, particularly around the unnamed gabbro intrusion. The rocks are an even medium grain size, and are grey to white depending on the proportion of mafic minerals present.

Plagioclase is always the most abundant mineral, comprising from 55 to 70% of the rocks. It mainly forms zoned subhedral laths, locally aligned near contacts, but ranges from euhedral to anhedral. Cores are andesine in composition, but approach labradorite in more melanocratic samples. Minor development of myrmekite occurs at contacts with K-feldspar.

Quartz makes up from 5 to 30% of the rocks, either interstitial to plagioclase laths or as anhedral grains which are locally aggregated together to form mosaics. It is most abundant in trondhjemites with low contents of mafic minerals.

K-feldspar is always very subordinate to plagioclase, ranging up to 10% of the rocks but usually being less than 5% if present at all. It occurs as interstitial to poikilitic grains up to 5mm across, or in some tonalities and quartz diorites as thin discontinuous rims on plagioclase laths.

Green-brown hornblende is the most abundant mafic mineral, comprising from 2 to 30% of the rocks. In more melanocratic samples it may rim and replace pyroxene, mainly augite (up to 5% of the rocks) but also hypersthene (minor). Locally it forms subophitic to poikilitic grains up to 5mm across partially or totally enclosing plagioclase laths. Biotite is ubiquitous but is less abundant (2 to 10% of the rocks) than hornblende.

Figure 343. Irregular ?Permian to Triassic andesite dykes and more regular ?Tertiary basalt dyke intruding Dumgree Tonalite, cutting on Gladstone–Moura railway 2km south-west of Mount Redshirt

Minor and accessory components include opaque grains, apatite, sphene, and zircon, and secondary muscovite and epidote.

The Dumgree Tonalite is intruded by andesitic and basaltic dykes (Figure 343). The andesitic dykes are assumed to be related to the Permo-Triassic magmatism, whereas the basaltic dykes are probably Tertiary. Locally, the basaltic dykes have acted as a barrier to circulation of fluids, and separate fresh from more weathered tonalite (Figure 344).

Geochemistry

The Dumgree Tonalite is one of the defining units of the Dumgree group, characterised by low total alkalis, TiO_2, MgO and high Al_2O_3, CaO and Sr. K_2O, Ce, Rb, Th, Y and Zr are all low and for the group as a whole tend to decrease with increasing SiO_2 (Figure 302). The Dumgree Tonalite has well defined linear trends on Harker diagrams that in some cases differ from those of the overall group, notably Rb, Th and Zr that all increase slightly with increasing SiO_2. It is distinguished geochemically from the Rocky Point Granodiorite by lower TiO_2, MgO, K_2O, Ce, Rb, Th, V, Y and Zr and higher Al_2O_3, Na_2O and Sr.

Age

Emplacement of the Dumgree Tonalite and Rocky Point Granodiorite is considered to have been essentially synchronous, as they appear to be part of a single large pluton. Webb & McDougall (1968) dated two samples from these intrusions and obtained Early Triassic K-Ar ages for both. However, the Redshirt Granite, which intrudes the Dumgree Tonalite, gave a U-Pb zircon date of 251±4Ma (Appendix 1), and therefore all three units have been assigned a Late Permian to Early Triassic age.

Stratigraphic relationships

The Dumgree Tonalite intrudes the Youlambie Conglomerate, and is intruded by the Redshirt Granite and part of its associated dyke swarm (Figure 345). It is assumed to intrude the Rocky Point Granodiorite because of their spatial relationship, and may also intrude the unnamed gabbro at the western extremity of the Galloway Plains Igneous Complex.

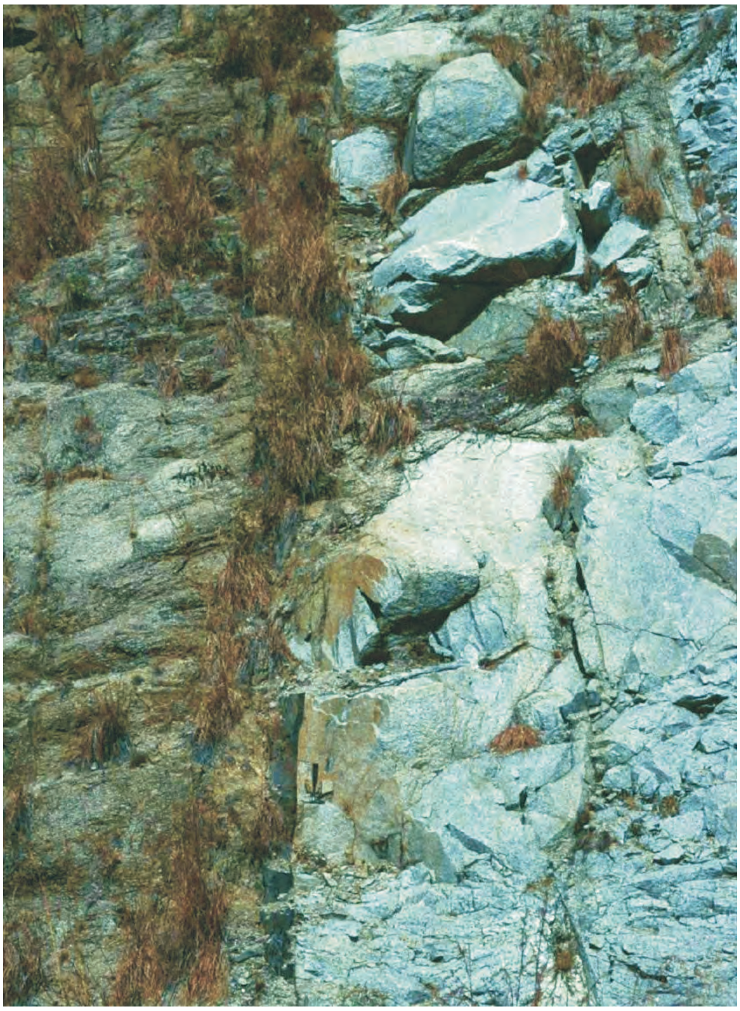


Figure 344. Thin vertical basalt dyke separating fresh and decomposed tonalite, cutting on Gladstone–Moura railway 4km south of Mount Redshirt, Dumgree Tonalite,

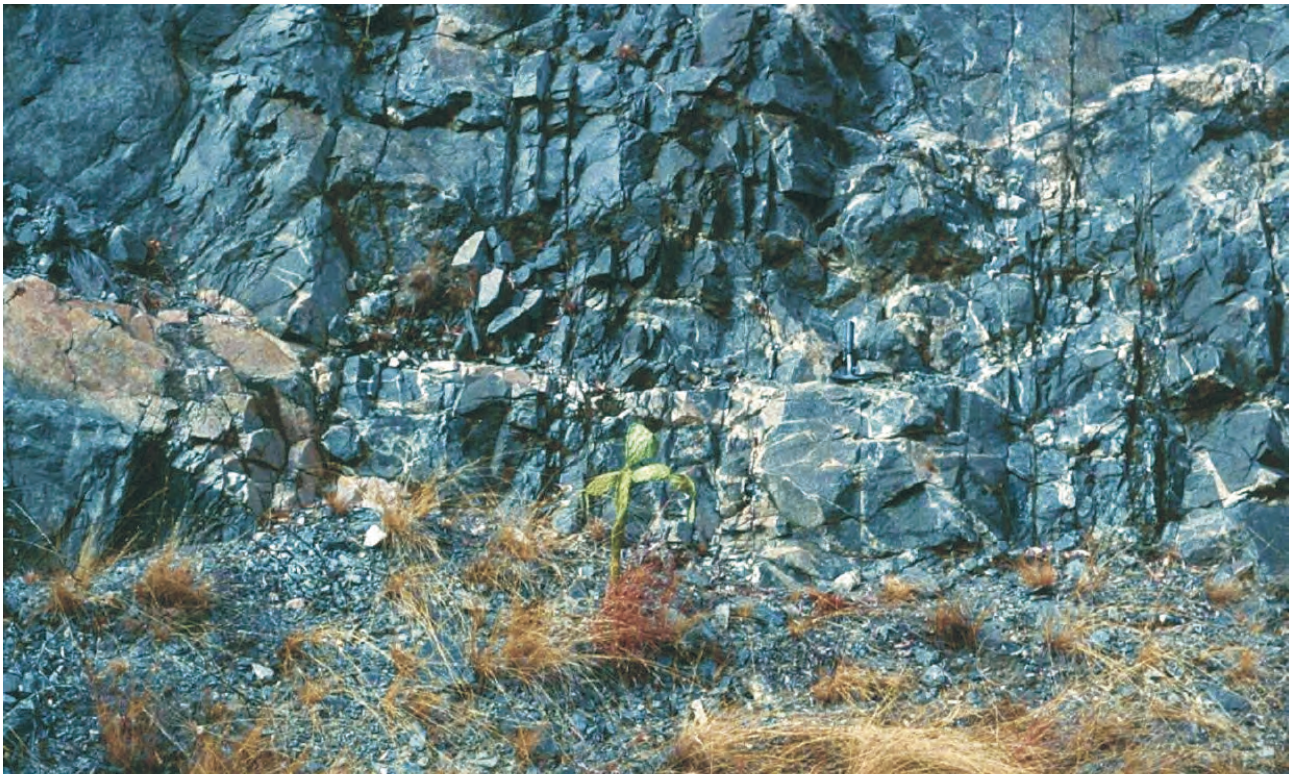


Figure 345. Irregular network of dykes from the Redshirt Granite intruding Dumgree Tonalite, cutting on Gladstone–Moura railway 1.5km south of Mount Redshirt

Economic significance

No mineralisation is known to be associated with the Dumgree Tonalite.

Wyalla Granite (PRgaw)

Introduction

The Wyalla Granite is a newly defined unit within the Galloway Plains Igneous Complex that intrudes the Rocky Point Granodiorite.

Distribution

The Wyalla Granite is located just south of the Dawson Highway about 40km north-east of Biloela, with dimensions of 2km by 1km (Figure 346).

Derivation of name

The name is derived from the property Wyalla that covers much of the outcrop area.

Type area

The type area is along a track to a powerline pylon 1.5km south-east of Wyalla homestead.

Topographic expression

The Wyalla Granite forms steep hills along the eastern escarpment of the Calliope Range.

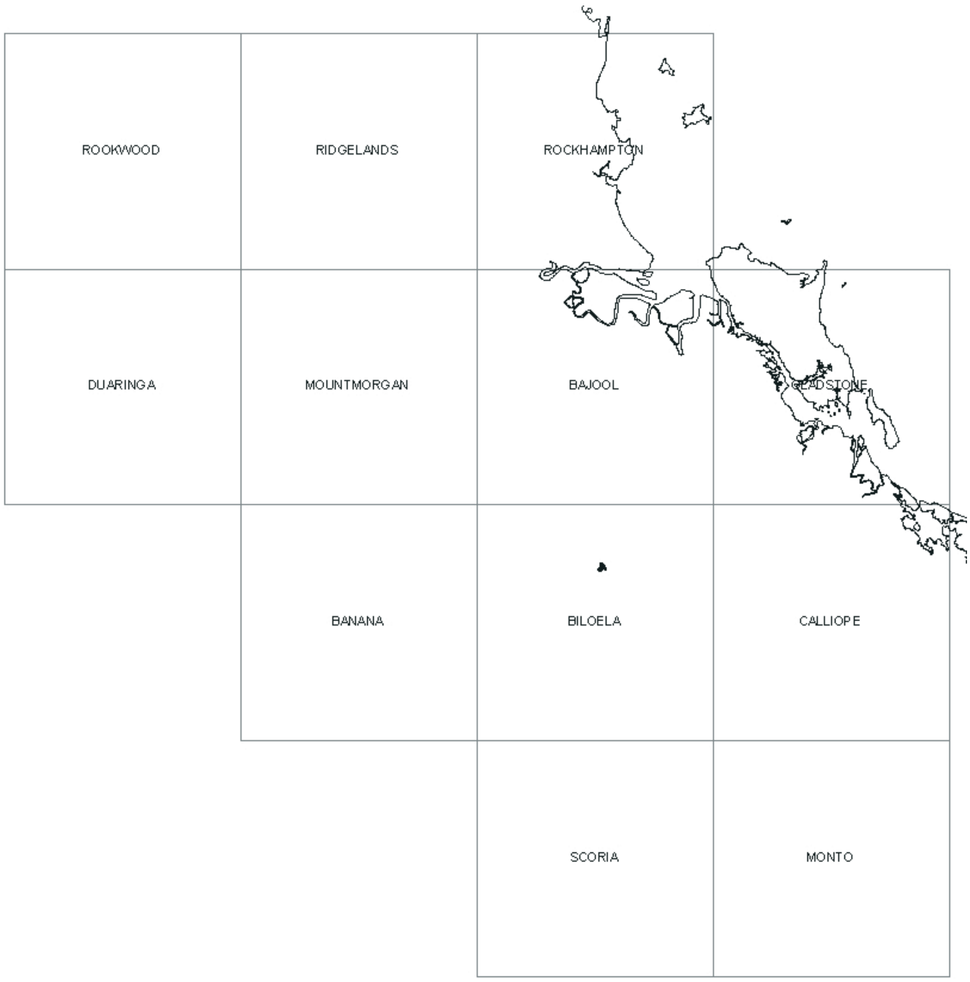


Figure 346. Distribution of the Wyalla Granite

Geophysical expression

The granite is easily distinguished from the surrounding Dumgree Tonalite and Rocky Point Granodiorite on images generated from aerial radiometric data by its orange to red colour, indicating a moderate to high potassium content.

It forms a relative magnetic low with a poorly developed rim, consistent with its magnetic susceptibility reading of 595 x 10^{-5} SI units that is much less than the surrounding units.

Lithology and petrography

The Wyalla Granite is a pale pink, massive, medium grained biotite granite.

About half the rock is made up of zoned, subhedral, sodic plagioclase laths. Anhedral quartz grains comprise about 25% of the rock, and are accompanied by small interstitial crystals of microcline microperthite (15 to 20%). Greenish-brown biotite makes up 10% of the rock, with accessory opaque grains, apatite, sphene, and zircon, possibly some tourmaline, and a little interstitial calcite.

Geochemistry

The geochemistry of the Wyalla Granite is similar to that of the Ridgelands group with the exception of Zr, which is significantly higher (Figure 317).

Age

No age determination has been carried out, and a Permian to Triassic age is assigned.

Stratigraphic relationships

The Wyalla Granite is interpreted to intrude the Rocky Point Granodiorite.

Economic significance

No mineralisation is known to be associated with the Wyalla Granite.

Redshirt Granite (PRgae)

Introduction

The Redshirt Granite is a newly defined unit within the Galloway Plains Igneous Complex that is readily distinguished from the surrounding Dumgree Tonalite by its topographic and geophysical expression. Dear & others (1971) mapped Triassic volcanics at the summit of Mount Redshirt, but none are present, with large outcrops of granite exposed right to the top.

Distribution

The Redshirt Granite is an oblong intrusion 4.5 by 2.5km located north of the Gladstone–Moura railway 40km north-north-east of Biloela (Figure 347).

Derivation of name

The name comes from Mount Redshirt, which is composed exclusively of granite of the intrusion.

Type area

Excellent exposures as large boulders and tors are present throughout the area covered by the Redshirt Granite. Relations with the surrounding Dumgree Tonalite are best seen in railway cuttings.

Topographic expression

The Redshirt Granite forms the prominent, steep peak of Mount Redshirt (597m), which rises almost 500m above the level of the Dumgree Tonalite and Bocoolima Granodiorite to the east (Figure 298).

Figure 347. Distribution of the Redshirt Granite

Geophysical expression

The granite has an extremely strong radiometric signature, showing white on images. The strong radiometric signature can be traced along stream channels draining the intrusion to the east. Its magnetic response is lower than that of the Dumgree Tonalite, consistent with its lower susceptibility averaging 1815 x 10^{-5} SI units.

Lithology and petrography

The Redshirt Granite is a medium to coarse grained, pink, hornblende-biotite monzogranite with patches of tourmaline-bearing pegmatite.

Plagioclase is the most abundant mineral, comprising from 35 to 40% of the rock as subhedral laths of zoned andesine-oligoclase up to 5mm long. K-feldspar (25 to 35%) locally forms micrographic intergrowths with quartz, but occurs mainly as interstitial to poikilitic grains up to 10mm across. Quartz makes up 20 to 30% of the rock as small equant grains.

Biotite is the main mafic mineral, comprising up to 10% of the rock. It is typically partly altered to chlorite. Subordinate hornblende is present in some specimens, with rare relict augite cores. Sphene is prominent as relatively large and abundant euhedral crystals, locally altered to rutile and ilmenite. Other accessory minerals include opaques, apatite, zircon, and zoned tourmaline. A little secondary muscovite, epidote and calcite are present.

Geochemistry

Like the felsic phase of the Mount Seaview Igneous Complex, the geochemistry of the Redshirt Granite is similar to that of more silicic samples from the Miriam Vale Granodiorite, with the exception of higher Sr, particularly at lower levels of SiO_2 (Figure 334).

Age

A U-Pb age determination on zircon gave an age of 251±4Ma, straddling the Permian–Triassic boundary (Appendix 1). This age is slightly greater than, but within the experimental error of, K-Ar dates from the Rocky Point Granodiorite and Dumgree Tonalite. A Late Permian to Early Triassic age has been assigned to all three units.

Stratigraphic relationships

The Redshirt Granite intrudes the Dumgree Tonalite, as seen in railway cuttings traversing the contact (Figure 342). It is the probable source of rhyolite dykes that form north-east trending swarms to the north and south-west of Mount Redshirt, and is locally overlain by cemented residual coarse sand deposits of Tertiary to Quaternary age (**TQrh**).

Economic significance

Some company exploration reports mention gold workings near the summit of Mount Redshirt, but none were observed during the mapping program. Small but rich gold deposits are associated with the rhyolite dyke swarm north of Mount Redshirt, including Plumtree Hill and Fig Tree Hill.

Voewood Granite (Rgav)

Introduction

The Voewood Granite is the only separate intrusion that can readily be mapped out within the large area covered by the Bocoolima Granodiorite.

Distribution

The intrusion covers a total area of 7.5 by 2km, centred about 45km north-east of Biloela. It is divided into western and eastern parts by alluvium along Maxwelton Creek (Figure 348).

Figure 348. Distribution of the Voewood Granite

Derivation of name

The name is taken from the property Voewood, on which typical exposures occur.

Type area

The type area is a hill west of Lost Spring Creek, 2km east-south-east of Voewood homestead.

Topographic expression

The Voewood Granite forms low undulating hills, and represents a distinct topographic high surrounded by the Bocoolima Granodiorite.

Geophysical expression

The intrusion has a moderately strong radiometric expression, with orange to red colours on images, contrasting markedly with the surrounding Bocoolima Granodiorite. The magnetic response is lower than that of the Bocoolima Granodiorite, consistent with susceptibility readings averaging 520 x 10^{-5} SI units. The western part of the intrusion has a well developed magnetic rim.

The Voewood Granite lies within the gravity low corresponding to the Bocoolima Granodiorite.

Lithology and petrography

The intrusion is quite uniform in composition, consisting of a medium grained pale pinkish grey biotite granite.

Plagioclase is the most abundant mineral (35 to 50% of rock), comprising zoned subhedral laths up to 3mm long with cores of sodic andesine that are partly to extensively altered to sericite. Microcline (20 to 30%) forms interstitial to poikilitic grains up to 3mm across, and is accompanied by anhedral quartz crystals up to 4mm across (25 to 30% of rock). Myrmekite is developed at contacts between K-feldspar and quartz. Biotite comprises about 5% of the rock, with minor and accessory opaque grains, sphene, apatite and zircon. Muscovite occurs as relatively large flakes in some specimens, and may be primary.

The western part of the intrusion is strongly jointed with pyrite-bearing quartz veins along the joints.

Geochemistry

The Voewood Granite is placed in the Ridgelands group (Figure 306).

Age

The Voewood Granite gave a U-Pb zircon age of 233.6±6.8Ma (Middle to Late Triassic) (Appendix 1).

Stratigraphic relationships

The Voewood Granite intrudes and is completely surrounded by the Bocoolima Granodiorite.

Economic significance

Apart from the pyrite-bearing quartz veins, no mineralisation is known to be associated with the Voewood Granite.

Ridgelands Granodiorite (PRgri)

(M.A. Hayward)

Introduction

The Ridgelands Granodiorite was first mapped by Kirkegaard & others (1970) and named after Ridgelands Creek which cuts the eastern part of the batholith.

Figure 349. Distribution of the Ridgelands Granodiorite

Distribution

The Ridgelands Granodiorite is centred 20km north-west of Rockhampton and comprises a main boomerang-shaped body and a smaller area to the south-west (Figure 349). The main body extends from Ridgelands township in the east, almost to Mount Salmon in the west, and to the Fitzroy River in the north, and covers approximately 115km^2. The smaller body crops out around and north-west of Sugarloaf Mountain (GR993 176) over an area of about 10km^2.

Type area

Outcrops and large tors of medium-grained granodiorite which crop out between Station Creek and Two Mile Creek (eg. at GR040 288) are typical of this unit. Dykes and a small intrusion of biotite microgranite are also exposed in this area.

Topographic expression

Although the granodiorite crops out poorly in most areas, and is characterised by low undulating country, in areas of better outcrop around Two Mile Creek it gives rise to moderately hilly country. Colour aerial photography over the Ridgelands Granodiorite commonly exhibits light brown to cream tones, particularly where sandy soil is exposed near gullies and creeks.

Geophysical expression

On a red-green-blue composite radiometric image, the main body of the Ridgelands Granodiorite is characterised by a moderate pink to red response. The small mass near Sugarloaf Mountain returns a slightly lower radiometric response. At the south-eastern edge of the main body, two roughly circular gabbro/diorite intrusions 1–2km in diameter, which probably intrude the main body, are represented by dark areas of low radiometric response. A small intrusive body of microgranite between Two Mile and Station Creeks produces a prominent radiometric high. Microgranite dykes in the surrounding granodiorite are too narrow to produce a noticeable radiometric signature.

Figure 350. Mafic inclusions in the Ridgelands Granodiorite

On a pseudocolour image of airborne magnetic data, the main body of Ridgelands Granodiorite shows a moderate magnetic response, and is surrounded by a rim of higher response. The main body also exhibits several north-west-trending ridges of higher magnetic response. The two gabbro/diorite bodies in the south-east are represented by two bullseyes of high magnetic response. These rocks have magnetic susceptibilities in the range 6000–12 000 x 10^{-5} SI units, accounting for their appearance as prominent magnetic highs. A smaller separate magnetic high to the south is also associated with a gabbro/diorite intrusion, and is assumed to be related to the larger bodies to the north. One area of low to very low response is located on the eastern limb of the main body.

The south-east trending lobe of the main outcrop coincides with a local negative gravity anomaly located on the steep gradient between the Marlborough Gravity Ridge to the west and the low related to the Yaamba Basin to the east.

Lithology and petrography

Although previously mapped as dominantly granodiorite, recent mapping reveals that there is considerable variation in the composition of this unit.

The main outcrop area is dominantly equigranular granodiorite and tonalite that have locally been intruded by biotite-bearing microgranite, diorite and gabbro. Biotite and hornblende are generally present in subequal amounts, and comprise a total of 10 to 20% of the rock by volume. K-feldspar is commonly present as large crystals (up to 2cm) which poikilitically enclose hornblende and plagioclase. Mafic xenoliths including pyrite-bearing andesite and microdiorite up to 40cm diameter are common (Figure 350).

Between Station and Two Mile Creeks, a small intrusive body, and numerous north-west-trending narrow dykes of biotite microgranite, intrude the coarser granodiorite. The microgranite is dominantly quartz and K-feldspar with only minor plagioclase and about 2% mafics. Biotite, which is the principal mafic phase, is ragged to elongate. Rare euhedral hornblende is commonly altered. Xenoliths were not observed in this phase.

The three small gabbro/diorite bodies that crop out around Ridgelands township are generally dark grey, equigranular to sparsely porphyritic, and locally pegmatitic. Small scale (1–2cm) layering was noted at one location. Gabbro at GR204 258 contains locally abundant black clots of mafic minerals, and sparse coarse gabbro/diorite xenoliths. The bullseye shape of these bodies suggests that they have intruded the main granodiorite body.

The smaller outcrop near Mount Sugarloaf is generally more mafic in composition than the main outcrop area. Rock types include biotite-hornblende quartz diorite, diorite, biotite-hornblende tonalite, and weathered granodiorite, all characterised by strongly zoned plagioclase laths. Hornblende is clearly the dominant mafic mineral. Weak to moderately well developed foliation is present at GR987 174. Rare, diffuse mafic xenoliths were observed at this site, but are more common and larger at GR993 181.

Geochemistry

The Ridgelands Granodiorite is the main unit defining the Ridgelands group, characterised chiefly by its low Sr content, which remains fairly constant over a range of SiO_2 values (Figure 306).

Age

Previously reported dates of 264 and 269Ma from the Ridgelands Granodiorite (Webb, 1969) are from outcrops now included in the Wattlebank Granodiorite. Two samples from the main body have been dated by Rb-Sr on biotite-whole rock pairs at 251±2Ma, close to the Permian–Triassic boundary (Appendix 1).

Stratigraphic relationships

The Ridgelands Granodiorite intrudes several Palaeozoic units including the Siluro-Devonian Craigilee Group, Devonian–Carboniferous Mount Alma Formation, Upper Carboniferous Neerkol Formation and Lower Permian Youlambie Conglomerate. The small isolated gabbro/diorite mass south-west of Ridgelands also intrudes the Lower Carboniferous Rockhampton Group. The Ridgelands Granodiorite is overlain by Cretaceous basalt and trachyte (Dalma Basalt and Mount Salmon Volcanics), and probably intruded by related plugs and dykes. A small area of Tertiary duricrust is developed on this unit 2km north of Mostowie homestead (GR174 263).

Economic significance

Three of the earliest goldfields worked in the Rockhampton district, Morinish, Ridgelands and Rosewood, are located within or adjacent to the outcrop area of the Ridgelands Granodiorite. It is possible that the Ridgelands Granodiorite was the source of at least some of the gold mineralisation. Most production from these goldfields was alluvial, and reef workings were of lesser importance.

Bajool Quartz Diorite (PRgaj)

(P.R. Blake)

Introduction

The presence of granitic rocks in the Bajool area was first noted by Reid (1943) while inspecting a silica deposit. Lonsdale (1965) and Darby (1969) postulated the existence of an extensive granite intrusion responsible for the Bajool Gravity Low, but the northern part of this gravity feature is due to the Tertiary Casuarina Basin. The intrusion south of Bajool was mapped as unnamed Permian granodiorite by Kirkegaard & others (1970), and is here named the Bajool Quartz Diorite.

Distribution

The Bajool Quartz Diorite extends southwards from Bajool to Mount Bomboola, covering an area 18km long by up to 15km wide (Figure 351). Most of the intrusion is deeply weathered, and outcrops are rare.

Derivation of name

The name is taken from the town of Bajool at the northern margin of the intrusion, and the previously named gravity feature, the Bajool Gravity Low.

Type area

Fresh rock is exposed in a small quarry excavating decomposed granitoid for road base at GR266300 7372800 on the Bajool 1:100 000 Sheet area, and this was the site sampled for geochemistry. Several weathered

Figure 351. Distribution of the Bajool Quartz Diorite

outcrops were seen in nearby creeks to the south of the pit. This pit and surrounding area are considered the type area.

Topographic expression

The Bajool Quartz Diorite typically forms very flat topography with only low hills.

Geophysical expression

Images produced from airborne radiometric data show a low response in all three channels. The airborne magnetic data generally show a moderate to low response, with some semicircular concentric zoning developed in the northern part of the intrusion. A small zone 10km south-south-west of Bajool has a relatively high magnetic response, and may represent a separate phase. Strong magnetic highs around the northern margin of the intrusion are assumed to be caused by hornfelsing of the surrounding country rocks.

The intrusion coincides with a negative gravity anomaly, part of the Bajool Gravity Low of Lonsdale (1965) and Darby (1969).

Lithology and petrography

The Bajool Quartz Diorite is medium grained and equigranular. Fresh rocks were only found at a single location, and consist of approximately 72% plagioclase, which is andesine in composition, 25% hornblende, 2% quartz, and 1% opaques.

Geochemistry

Analysed samples from the Bajool Quartz Diorite all come from a single locality, and give virtually identical results. Some aspects of the geochemistry are similar to the Dumgree group (high Al_2O_3 and CaO and low

TiO_2, total alkalis, K_2O and Rb). However, the Bajool Quartz Diorite has much lower total Fe, V and possibly Sc, and markedly higher Sr. Due to these differences, and the lack of variation of SiO_2, the intrusion has not been assigned to any geochemical group.

Age

Altered granodiorite from the Limonite Hill porphyry copper prospect 5km south-south-west of Bajool gave an age of 249±5Ma by K-Ar on sericite (Green, 1975; Ford & others, 1976). This is assumed to be close to the age of intrusion, and the Bajool Quartz Diorite is assigned a Permian to Triassic age.

Stratigraphic relationships

The Bajool Quartz Diorite intrudes the Mount Warner Volcanics, Raspberry Creek Formation and Ginger Creek beds of the Capella Creek Group, Erebus beds, Mount Alma Formation, Rockhampton Formation, and unnamed Devonian gabbros.

Economic significance

Quartz pipes near Bajool are potential sources of lump silica (Reid, 1943; Connah, 1976; Geological Survey of Queensland, 1978; Allen, 1980; Sawers & Cooper, 1985; Watkins, 1989). Three circular pipes up to 73m in diameter occur. Minor production has taken place for flux at Mount Morgan.

Limonite Hill, a leached capping 5km south-south-west of Bajool, has been explored as a potential porphyry-style deposit. Assays from drilling programs have shown elevated but uneconomic copper values, and only very low values for molybdenum (Horton, 1978, 1982; Blake & others, 1995).

Cecilwood Quartz Diorite (PRgce)

(C.G. Murray & R.J. Bultitude)

Introduction

An unnamed Permian intrusion at Langmorn Creek was mapped as unnamed Permian granodiorite by Kirkegaard & others (1970), and is here named the Cecilwood Quartz Diorite.

Distribution

The Cecilwood Quartz Diorite is located about 25km south of Bajool, and extends for about 10km in a north-south direction and up to 8km east-west (Figure 352). The intrusion is deeply weathered, and outcrops are rare.

Derivation of name

The name is taken from Cecilwood homestead, within the area covered by the intrusion.

Type area

No type area has been assigned. The only fresh rock comes from GR262380 7359347 near the western edge of the intrusion.

Topographic expression

The intrusion forms very flat topography along Langmorn Creek and its tributaries.

Geophysical expression

Images produced from airborne radiometric data show a low response in all three channels. The airborne magnetic data generally show a moderate to low response, with a rim of hornfelsed country rock having much stronger magnetic signature. This hornfelsed rim shows patches of high radiometric response that have been reported as due to thorium-rich skarns.

The intrusion coincides with a negative gravity anomaly, part of the Bajool Gravity Low of Lonsdale (1965) and Darby (1969).

Figure 352. Distribution of the Cecilwood Quartz Diorite

Lithology and petrography

The single fresh specimen collected from the Cecilwood Quartz Diorite is a greenish-grey, medium grained, equigranular quartz diorite. It consists mainly of plagioclase (75–80%) as tabular grains slightly altered to sericite. Quartz makes up about 10% of the rock, and is characterised by undulose extinction. Augite is the main mafic mineral as stubby euhedral to subhedral prisms, with subordinate pale green hornblende. Minor and accessory components include relatively abundant sphene, K-feldspar, and biotite. The K-feldspar forms myrmekitic intergrowths with quartz.

Geochemistry

The single sample analysed from the Cecilwood Quartz Diorite is most similar to the Dumgree and Monal groups, but differs from both of these in particular elements. It is higher than the Dumgree group in Ce, Sr and possibly in total alkalis, and is low in Rb. Compared with the Monal group, it is high in Sr and low in TiO_2 and Rb, and slightly low in Th, Y and Zr (Figure 353).

Age

In the absence of any data, the Cecilwood Quartz Diorite is assigned a Permian to Triassic age.

Stratigraphic relationships

The Cecilwood Quartz Diorite intrudes the Raspberry Creek Formation and presumably the Ginger Creek beds of the Capella Creek Group, and the Mount Alma Formation. It is overlain by the Quaternary alluvium (**Qa**) of Langmorn Creek and its tributaries.

Economic significance

Thorium-rich skarns have been reported from the contact rocks on the southern side of the intrusion.

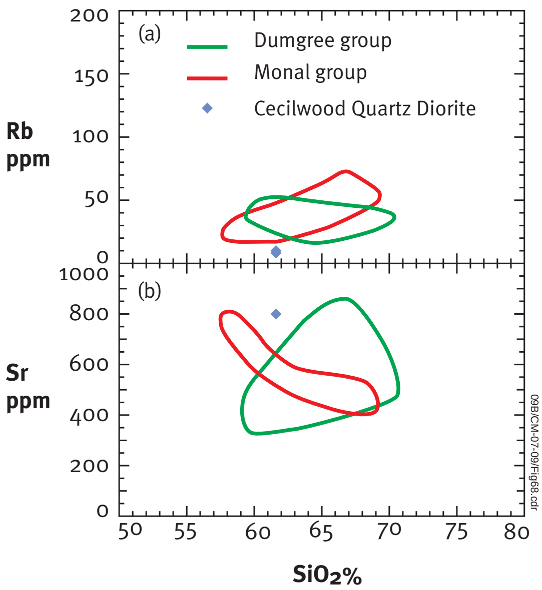


Figure 353. Comparison of Rb and Sr of the single analysis from the Cecilwood Quartz Diorite with the Dumgree and Monal groups. Analysis from B. Chappell & R. Bultitude (unpublished data).

Glassford Igneous Complex
(C.G. Murray)

The name Glassford Complex was introduced by Dear & others (1971) for a small granitic batholith located west of the townships of Ubobo and Many Peaks, about halfway between Gladstone and Monto (Figure 354). They mapped the Glassford Complex as a larger east-north-east trending northern section 30km long and averaging 10km across, and a smaller equant southern section, almost separated by a cover of Triassic Muncon Volcanics. The complex was described as comprising granodiorite, adamellite and tonalite, but it was not subdivided. Triassic ages were obtained from the complex by Webb & McDougall (1968).

Neale (1968) described a multiphase syenite intrusion within the area of the Glassford Complex, in the headwaters of Ridler Creek, which was dated as earliest Cretaceous (Green & Webb, 1974). This Cretaceous syenite was shown on the first edition Monto 1:250 000 geological map (Murphy, 1981).

Interpretations of airborne magnetic surveys suggest that the Glassford Complex consists of at least several discrete plutons, including one concealed beneath the Dooloo Tops Volcanics (R. Deakin, in Rees, 1997).

Field mapping has subdivided the Glassford Complex into 13 separate units, two of which are themselves composite, and it is renamed the Glassford Igneous Complex. It consists entirely of intrusive rocks, Triassic volcanics being absent from the vicinity of Mount Robert where they were mapped by Dear & others (1971).

Monal Granodiorite (Rggm)

Introduction

The western end of the Glassford Igneous Complex displays a ‘doughnut’ type magnetic pattern caused by a strongly magnetic hornfelsed rim almost completely surrounding an intrusion ranging in composition from quartz diorite to granite, which is defined here as the Monal Granodiorite. Although interpretation of magnetics suggests that the intrusion is a discrete pluton restricted to the centre of the magnetic doughnut (R. Deakin, *in* Rees, 1997), on the ground it appears to be continuous with granodioritic rocks of similar petrography extending to the east. All these rocks are therefore included in the Monal Granodiorite, although there are minor geochemical differences. The Monal Granodiorite is divided into a phase ranging from quartz diorite to granodiorite (**Rggm$_d$**) and a phase ranging from granodiorite to granite (**Rggm$_g$**).

Distribution

The Monal Granodiorite is located in the headwaters of Monal and Ridler Creeks, about 35km north of Monto, and forms an elongate intrusion about 15km long and 2km wide trending east-north-east (Figure 355).

Figure 354. Distribution of the Glassford Igneous Complex

The granodiorite to granite phase (**Rggm$_g$**) is confined to a small ovate area of about 4km^2 between Monal and Broad Creeks.

Derivation of name

The name is derived from Monal Creek, which flows through the intrusion in a north-south direction.

Type area

The type area for the Monal Granodiorite extends from Spring Creek in the west to Broad Creek in the east, and is centred on Monal Creek. All compositional variations are represented in this area. Excellent exposures of the quartz diorite to granodiorite phase (**Rggm$_d$**) can be seen in the bed of Ridler Creek south of Mount Sugarloaf.

Topographic expression

The Monal Granodiorite is hilly in the east, and forms part of the crest of the Dawes Range. In the west, between Spring and Monal Creeks, it forms a topographic basin rimmed by higher hornfels ridges. Here, rhyolite dykes crop out as small but prominent sharp ridges on low, undulating hills.

Geophysical expression

On images displaying airborne magnetic data, the Monal Granodiorite appears as an area of low magnetic response. However, this appearance is largely due to the strong contrast between the granodiorite, with average magnetic susceptibility of about 1700 x 10^{-5} SI units, and the surrounding hornfels with an average value of 7335 x 10^{-5} SI units. Modelling indicates that the central intrusion is essentially non-magnetic, and that the strongly magnetic hornfels rim is close to vertical along the north-west margin, and dips at moderate angles to the south-east along the south-eastern contact (R. Deakin, *in* Rees, 1997).

Figure 355. Distribution of the Monal Granodiorite

The intrusion gives a moderate radiometric response, largely due to potassium, except in the area between Monal and Broad Creeks. Here, the more felsic granite to granodiorite phase (**Rggm$_g$**) is clearly outlined by its stronger radiometric signature.

Geophysics is very effective in delineating boundaries between the Monal Granodiorite and other units of the Glassford Igneous Complex. The Littlemore Granodiorite gives a stronger response for both radiometrics, particularly in the thorium and uranium channels, and magnetics. The change in magnetic signature is less well defined than that in the radiometric response, reflecting the similarity in average magnetic susceptibility (1700 x 10^{-5} SI units for the Monal Granodiorite compared with an average of 1725 x 10^{-5} SI units for the Littlemore Granodiorite). The Ridler Monzonite, Radley Nepheline Syenite and Burns Spur Nepheline Monzonite all have a much stronger radiometric signature than the Monal Granodiorite.

The Monal Granodiorite has no obvious gravity expression.

Lithology and petrography

In the more mafic marginal phase (**Rggm$_d$**), the dominant rock type is a grey, medium grained equigranular biotite-hornblende granodiorite to tonalite, consisting of 55–60% plagioclase, 5–15% K-feldspar, 15–25% quartz, 10–15% mafics (hornblende and biotite), minor opaque minerals, and accessory apatite, sphene and zircon. Plagioclase typically forms subhedral laths up to 5mm long, and consists of large relatively uniform cores of andesine with thin to moderate rims zoned to more sodic compositions. K-feldspar occurs as interstitial grains. In some granodiorites there is local myrmekite development at plagioclase – K-feldspar contacts. Hornblende is usually more abundant than biotite, but in some samples their relative order is reversed. Hornblende forms euhedral to subhedral prisms in most specimens, and is usually pale green in colour. One sample from the western margin of the intrusion in Spring Creek is a biotite-hornblende quartz diorite with sparse relict cores of augite in hornblende, and rocks of this composition appear to be abundant in this area. Some specimens of Monal Granodiorite show obvious evidence of hornfelsing, and small shear zones occur in outcrops in Ridler Creek (Figure 356), possibly caused by intrusion of the adjacent Ridler Monzonite.

Figure 356. Small shear zone in Monal Granodiorite in Ridler Creek south of Mount Sugarloaf

The felsic phase (**Rggm$_g$**) ranges from medium grained equigranular grey biotite-hornblende granodiorite similar to that in phase **Rggm$_d$** to pink hornblende-biotite granite consisting of subhedral plagioclase laths (sodic andesine) and pale green hornblende prisms to 5mm long in a finer grained groundmass of quartz, K-feldspar and biotite. Quartz occurs as small equant grains either embedded in K-feldspar grains or forming mosaics.

Swarms of east-north-east trending dykes, dominantly rhyolite but locally of more intermediate composition, cut the western part of the granodiorite between Monal and Spring Creek.

Geochemistry

The Monal Granodiorite is the type intrusion for the Monal group (Figure 305). This group is intermediate between the Dumgree and Littlemore groups for many elements, eg. total alkalis.

Outcrops along Ridler Creek south of Mount Sugarloaf show minor geochemical differences compared to the Monal Granodiorite further west. Two analyses from this area have slightly lower Ce, Nb, Y and Zr (Figure 305). These differences may indicate that this area represents a separate intrusive phase.

Age

No age determinations have been carried out on the Monal Granodiorite, but it is considered to be Triassic because of its close relationship with the Littlemore Granodiorite.

Stratigraphic relationships

Within the Monal Granodiorite, the felsic phase (**Rggm$_g$**) appears to intrude the marginal granodiorite-tonalite phase (**Rggm$_d$**). Outcrops in Broad Creek show grey tonalite to quartz diorite cut by dykes and veins of pink hornblende granite.

The Monal Granodiorite intrudes and has hornfelsed the Devonian–Carboniferous Three Moon Conglomerate. It is intruded by the Burns Spur Nepheline Monzonite and the Ridler Monzonite, both of Cretaceous age, by a rhyolitic dyke swarm related to the Triassic Dooloo Tops Volcanics, and by Tertiary basalt plugs. The Dooloo

Tops Volcanics are interpreted to overlie the Monal Granodiorite, and the northern end of the plateau forming the Dooloo Tops appears to be an exhumed Triassic weathering profile developed on the granodiorite.

The relative age relationship between the Monal Granodiorite and the Littlemore Granodiorite is unknown. The common boundary between these two units has been interpreted from airborne geophysical data, and is consistent with geochemical results.

Economic significance

Numerous gold workings of the Monal Goldfield are located around the periphery of the Monal Granodiorite, confirming that mineralisation was related to the intrusion.

Littlemore Granodiorite (Rggi)

Introduction

Most of the northern part of the Glassford Complex, along the valley of Ridler Creek, is granodiorite with subsidiary quartz monzodiorite. The name Littlemore Granodiorite is proposed for this unit.

Distribution

The Littlemore Granodiorite forms the north-eastern part of the Glassford Igneous Complex, and is the largest unit within the complex. It extends in a west-south-west trending belt about 20km long and 6km wide along Ridler Creek from the township of Ubobo to Mount Sugarloaf (Figure 357).

Derivation of name

The name comes from the Littlemore scheelite mine, which occurs within the granodiorite. The railway siding of Littlemore is located 3km east of the intrusion.

Figure 357. Distribution of the Littlemore Granodiorite

Type area

The type area is along Connell Creek for about 4km upstream from its confluence with Ridler Creek. The intrusion exhibits considerable compositional variation in this section. The Littlemore Granodiorite also forms prominent outcrops in cliffs along the northern side of Mount Robert.

Topographic expression

The north-eastern part of the Littlemore Granodiorite, in the lower reaches of Ridler Creek west of Ubobo, forms low undulating hills with outcrop of sparse tors. To the south, its elevation increases up the northern flank of Mount Robert and the Table Range, where it forms a very steep escarpment 500m high. This abrupt change in topography almost certainly reflects the presence of rocks more resistant to erosion to the south (for example, the Robert Granite forming Mount Robert).

Geophysical expression

Despite its moderate magnetic susceptibility (averaging 1725 x 10^{-5} SI units), the Littlemore Granodiorite appears as a uniform, fairly well defined body on images of airborne magnetic data because of the strong contrast with virtually non-magnetic sedimentary units to the east, north and south-west. The magnetic pattern suggests the presence of at least two discrete plutons (R. Deakin, *in* Rees, 1997), but these could not be distinguished either in the field or by petrography and geochemistry. Modelling of north-south and east-west profiles across the northern part of the Littlemore Granodiorite shows a weakly magnetic central intrusive with a moderately magnetic northern rim (R. Deakin, *in* Rees, 1997).

Where outcrops of fresh rock are relatively abundant, the Littlemore Granodiorite gives a moderately strong radiometric response in all channels. However, deep weathering in the low lying north-eastern portion of the intrusion has reduced potassium and uranium levels to a much greater extent than thorium, and the response from the thorium channel is relatively enhanced.

The Littlemore Granodiorite occurs at the western edge of the most intense gravity low in the region, the cause of which is unknown, although it is undoubtedly due in part to the Tertiary Nagoorin Graben.

Lithology and petrography

The rocks are grey medium to coarse grained equigranular biotite-hornblende granodiorite and quartz monzodiorite. Plagioclase (45–55% of rocks) occurs as subhedral laths up to at least 6mm long with uniform cores of andesine composition and thin more sodic rims. Quartz and K-feldspar both form interstitial grains and comprise 10 to 25% of the rocks. In the higher parts of the intrusion on the Table Range and the northern slopes of Mount Robert, presumably closer to its original roof zone, K-feldspar forms micrographic intergrowths with quartz and also rims plagioclase laths. Hornblende and biotite together make up 10 to 25% of the rocks, with green to green-brown hornblende generally more abundant than biotite. In some samples, hornblende contains pyroxene cores which may be fresh (augite) or altered to pale green secondary amphibole. Minor opaque grains are ubiquitous, as are accessory sphene, apatite and zircon.

Locally, the rocks show significant alteration. Replacement of feldspars by clay minerals gives them a cloudy appearance in thin section, and changes the colour of the rock to pink in hand specimen. Biotite is completely converted to chlorite, epidote and sphene. One altered sample in the type area along Connell Creek contains rare flakes of molybdenite up to 10mm across, and sparse small grains of tourmaline.

Geochemistry

The Littlemore Granodiorite, together with the Lawyer Granite, is the type intrusion for the Littlemore group (Figure 303).

Age

Webb & McDougall (1968) obtained K-Ar dates of 221±4Ma (biotite) and 216±4Ma (hornblende) from a sample of Littlemore Granodiorite collected 4km west-south-west of Ubobo, indicating a Late Triassic age. However, the Littlemore Granodiorite appears to have been intruded successively by the Rule Gabbro and Robert Granite. A U-Pb zircon age of 230.2±3.7Ma from the Robert Granite (Appendix 1) suggests either that the interpreted intrusive relationship is incorrect, or that the K-Ar dates are too young due to Ar loss. The Littlemore Granodiorite is assigned a Triassic age.

Stratigraphic relationships

The Littlemore Granodiorite intrudes the Carboniferous Rockhampton Group and unnamed Permian sedimentary and volcanic rocks. Interpretation of magnetic data suggests that it is intruded successively by the Rule Gabbro and Robert Granite, and it is intruded by the Ridler Monzonite. Its relationship with the Monal Granodiorite is not known. The eastern edge of the intrusion is cut off by the fault which forms the western margin of the Tertiary Nagoorin Graben.

Economic significance

The Littlemore scheelite deposit occurs in altered rocks of the Littlemore Granodiorite associated with a porphyry dyke. Disseminated copper mineralisation has recently been discovered in outcrops on the northern flanks of Mount Robert (McGraths and Plumtree prospects). Copper also occurs in skarns formed in contact rocks of the Rockhampton Group both to the north (Mount Hector – Monument area) and south (Hourigans prospect) of the intrusion. Calc-silicates have replaced limestone lenses in unnamed Permian sedimentary rocks to the east of the Littlemore Granodiorite, but no sulphides were observed.

Lawyer Granite (Rggl)

Introduction

The southern part of the Glassford Igneous Complex south of Coppermine Creek is a zoned pluton ranging from granite in the core to quartz monzodiorite at the margin. This zonation was recognised by Crick (1963) and Hill (1963) in their detailed studies of the Glassford copper deposits. They described a range of compositions from adamellite in the core (with approximately equal proportions of plagioclase, K-feldspar and quartz), to granodiorite (locally foliated) at the margin, with a narrow outer contact zone of porphyritic trondhjemite. No intrusive contacts have been observed between the central granite phase ($\mathbf{Rggl_g}$) and the marginal quartz monzodiorite ($\mathbf{Rggl_r}$).

Distribution

The Lawyer Granite, centred 15km west-south-west of the town of Many Peaks, covers an area of about 30km^2 between Coppermine Creek in the north and Glassford Creek in the south (Figure 358). The granite core ($\mathbf{Rggl_g}$) makes up most of this area, and forms a roughly rectangular mass 5km from north to south and 4km from east to west. The marginal quartz monzodiorite ($\mathbf{Rggl_r}$) is thin and locally discontinuous around the eastern, southern and western sides of the granite, and more extensive along the northern edge.

Derivation of name

The name comes from Lawyer Creek, which flows from north to south through the southern part of the intrusion, and along which typical rocks of both phases are exposed.

Type area

The type area is along Little Glassford Creek from its junction with Coppermine Creek to the western contact of the intrusion at the Blue Bag copper mine. Excellent exposures of both phases can be seen in this creek section.

Topographic expression

The Lawyer Granite creates relatively hilly topography, particularly in the centre of the granite phase where resistant finer grained aplitic rocks (mapped as Triassic Muncon Volcanics on the 1st edition Monto 1:250 000 map) form a small plateau, and at contacts with more resistant units. At its north-western extremity, the intrusion reaches altitudes of 500m where it is intruded by the Robert Granite.

Geophysical expression

Like the Littlemore Granodiorite, the Lawyer Granite is clearly defined on images produced from airborne magnetic data, because its average susceptibility (1780 x 10^{-5} SI units for the quartz monzodiorite phase $\mathbf{Rggl_r}$ and 1550 x 10^{-5} SI units for the granite phase $\mathbf{Rggl_g}$) is much greater than that of the surrounding sedimentary rocks except for local zones of skarn development. The magnetic pattern suggests that it represents a single equant pluton, but it may originally have extended further to the north, as it appears to be cut off in this direction by the strongly magnetic Rule Gabbro. Modelling results for a transect across the Lawyer Granite show that it is a weakly magnetic equidimensional intrusion cut by slightly more magnetic dykes (R. Deakin, *in* Rees, 1997).

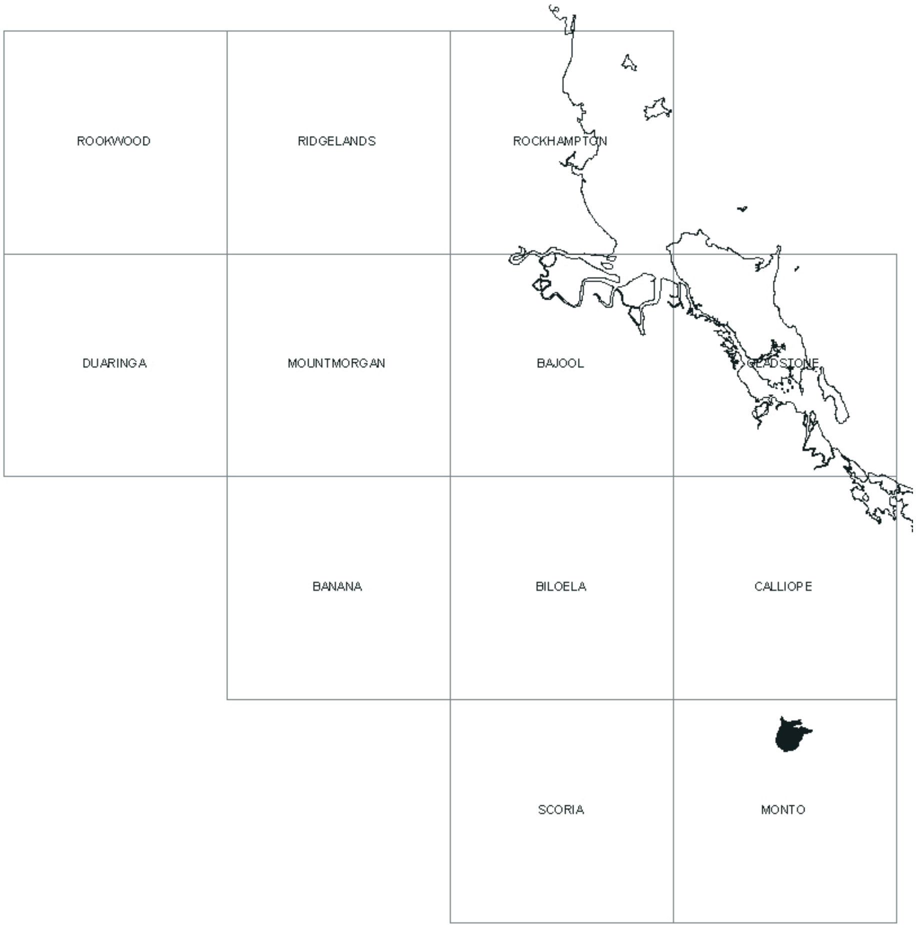


Figure 358. Distribution of the Lawyer Granite

The central granite phase (**Rggl$_g$**) gives a very strong, uniform response in all radiometric channels. The response from the marginal quartz monzodiorite (**Rggl$_r$**) is much lower for potassium, but still relatively uniform, whereas for uranium it is patchy, and for thorium very weak.

The Lawyer Granite has no obvious gravity effect.

Lithology and petrography

The granite phase (**Rggl$_g$**) is quite uniform in composition, and consists dominantly of pinkish grey, medium to coarse grained equigranular biotite-hornblende granite containing about 15% mafic minerals. Plagioclase (30–40% of rock) occurs as subhedral laths to more than 6mm long, with a relatively homogeneous andesine core surrounded by a zoned sodic rim. K-feldspar (20–35%) is either equal or subordinate to plagioclase as interstitial to poikilitic grains up to greater than 5mm across. There is minor myrmekite development at plagioclase – K-feldspar contacts. Quartz (20–30%) forms typical anhedral grains slightly smaller than the feldspars. The mafic content is fairly uniform at about 15%, with equal amounts of green to green-brown hornblende and biotite. Minor opaque grains and accessory sphene, apatite and zircon are ubiquitous.

Some rocks from the type area along Little Glassford Creek approach minimum melt compositions, with approximately equal proportions of plagioclase, K-feldspar and quartz, less mafics (5–10% of rock), and no hornblende. K-feldspar and quartz locally form micrographic intergrowths, and plagioclase cores are sodic andesine.

The marginal phase is composed dominantly of grey, medium grained equigranular augite-hornblende-biotite quartz monzodiorite with sparse dark grey, fine grained inclusions. Plagioclase is the main mineral (50–60% of rock) and forms zoned subhedral laths up to 5mm long with calcic andesine cores. K-feldspar comprises 10 to 15% of the rock as small interstitial grains, except along the western side of the intrusion where it forms poikilitic crystals up to more than 5mm across, with common myrmekite development in the marginal zones of included plagioclase laths. Quartz content is remarkably consistent at about 15%. Mafic minerals range from 15 to 25%, and consist of green-brown hornblende with relict augite cores, and slightly more abundant biotite. Minor opaque minerals and accessory sphene, apatite and zircon are present.

Foliated quartz gabbro occurs where Coppermine Creek crosses the eastern edge of the intrusion. The foliation is produced by alignment of the plagioclase laths, which make up from 60 to 65% of the rock and are labradorite in composition. Mafic minerals range from 25 to 30% and comprise augite and hypersthene, with or without rims of brown hornblende, and biotite. Quartz content is about 5%.

Magnetite-rich skarns around this intrusion are prominent on magnetic images.

Geochemistry

The Lawyer Granite and Littlemore Granodiorite are the type intrusions for the Littlemore group (Figure 303).

Age

Two samples from the Lawyer Granite dated by Webb & McDougall (1968) gave remarkably consistent results. Sample GA 1247, from the quartz monzodiorite phase, gave K-Ar dates of 226±4.5Ma on biotite and 224±4.5Ma on hornblende. The second sample, GA1248, was probably from the granite phase, although it is described simply as granodiorite. Biotite from this rock was dated at 226±4.5Ma. The results suggest that the two phases are coeval, and of Late Triassic age. However, the Lawyer Granite appears to have been intruded successively by the Rule Gabbro and Robert Granite. A U-Pb zircon age of 230.2±3.7Ma from the Robert Granite (Appendix 1) suggests either that the interpreted intrusive relationship is incorrect, or that the K-Ar dates are too young due to Ar loss. The Lawyer Granite is assigned a Triassic age.

Stratigraphic relationships

The Lawyer Granite intrudes the Carboniferous Rockhampton Group. Interpretation of magnetic data suggests that it is intruded successively by the Rule Gabbro and the Robert Granite, and therefore by the Deception Quartz Monzonite also. Tertiary basalt plugs intrude the unit along its south-eastern margin, forming topographic highs (Mount Weary and an unnamed hill to the north-east). The relationship of the Lawyer Granite to unnamed Triassic volcanic rocks to the east is unknown.

Economic significance

Garnet and magnetite bearing skarns along the western edge of the Lawyer Granite have produced copper from the Blue Bag and Lady Inez deposits of the Glassford Creek mine group. Further south, zinc-rich skarns occur at Mount Sperber. Skarns are also developed along the intrusive contact near Mount Weary.

The small dioritic intrusion associated with the Boggy Creek gold prospect is located just east of the Lawyer Granite and may be related to it.

Rule Gabbro (Rggu)

Introduction

The most prominent feature on images of the Glassford Igneous Complex produced from airborne magnetic data is an elliptical ‘doughnut’ style magnetic high between the Littlemore Granodiorite and the Lawyer Granite. The southern part of this anomaly coincides with gabbroic rocks that are here named the Rule Gabbro.

Distribution

The Rule Gabbro forms an arcuate mass extending 6km from Station Creek in the east to Coppermine Creek in the west, with an average width of about 2km (Figure 359). A separate small outcrop occurs 2.5km north-east of Mount Robert (N Cook, personal communication, 1998).

Derivation of name

The name is derived from the Parish of Rule, which includes all of the larger outcrop area of the gabbro.

Type area

The type area is along the southern branch of Deception Creek from 1km above its junction with the north branch for a further 1km upstream.

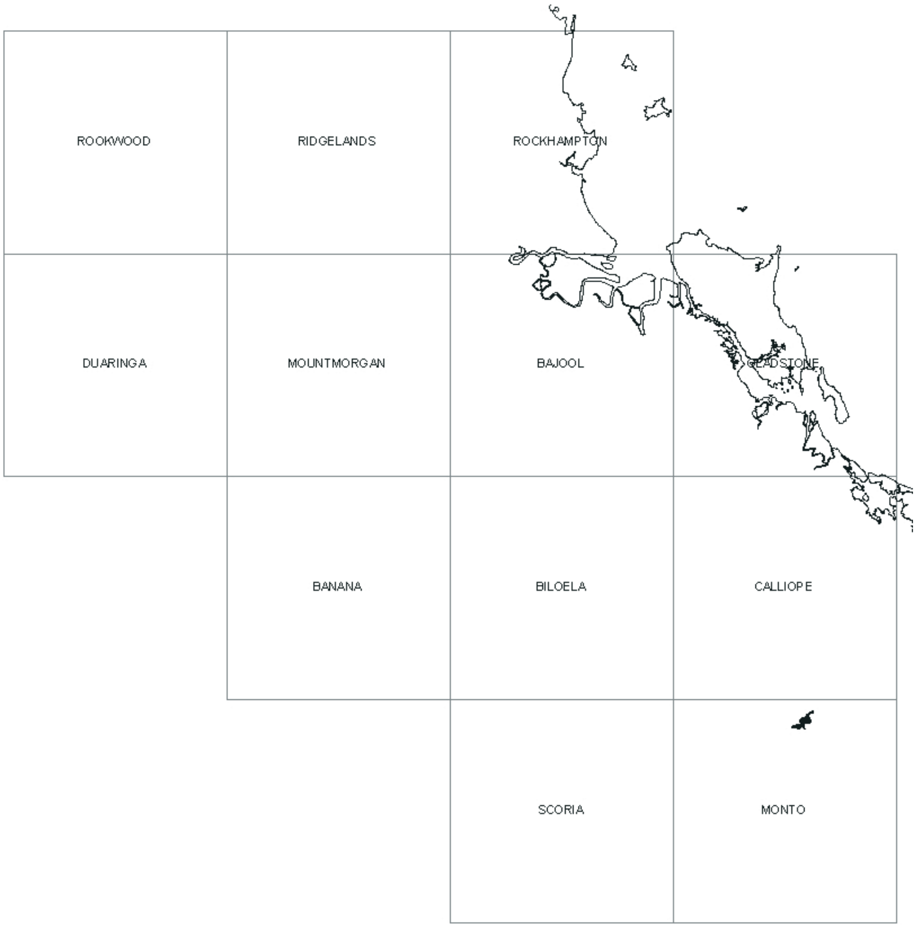


Figure 359. Distribution of the Rule Gabbro

Topographic expression

The Rule Gabbro forms extremely steep, rugged and thickly forested country ranging in altitude from just above 200m in stream valleys to 600m on the higher ridges. Access is difficult except along creeks and old forestry tracks.

Geophysical expression

Although outcrops cover only a relatively small part of the intense doughnut shaped positive magnetic feature, the Rule Gabbro is considered to be responsible for the anomaly. Modelling results indicate that the magnetic source is buried. Along a north-east – south-west profile, the depth to the top of the magnetic body is greater in the north-east part of the doughnut than in the south-west (R. Deakin, *in* Rees 1997). This is consistent with the outcrop pattern of the gabbro. However, modelling of a north-west – south-east profile suggests greater depths to the top of the body on the south-eastern side, even though the gabbro is exposed there (Appendix 5). The results suggest that the rocks exposed at the surface are not the main magnetic source, despite their substantial magnetic susceptibility (average of 4150 x 10^{-5} SI units for the main southern mass, and considerably greater values for the small northern outcrop; N Cook, personal communication, 1998).

The magnetic data clearly show that the Rule Gabbro formed as a single, slightly elliptical pluton about 10km long and 7km wide, with its long axis oriented north-east. It is intruded by the weakly magnetic Robert Granite that forms the centre of the doughnut anomaly.

The Rule Gabbro gives a low radiometric response. Interpretation is hampered by the dual problems of data quality and downslope transport of material, both of which are the result of the rugged topography. Nevertheless, the radiometric data have been used to determine the western boundary of the main southern mass, where access is difficult.

The gabbro has no obvious gravity expression, but it should be noted that there are no stations within 3km of the intrusion.

Figure 360. Photomicrograph showing strong alignment of plagioclase laths with interstitial augite, hornblende, hypersthene and altered olivine, Rule Gabbro. Crossed nicols. Scale bar is 1mm long.

Lithology and petrography

The Rule Gabbro is an altered, grey, medium to coarse grained biotite-hornblende-olivine-augite gabbro. Subhedral laths of plagioclase up to 5mm long, labradorite in composition, are by far the dominant constituent, comprising 60-65% of the rock. Locally they are strongly aligned (Figure 360), giving the rock a prominent foliation. Ophitic augite grains up to at least 8mm across are the main mafic mineral (10–25%). Some augite is rimmed by brown hornblende. Olivine (10–15%) occurs as anhedral to subhedral crystals up to 5mm across, but is mostly altered. It is either enclosed in augite grains, or surrounded by reaction rims of hypersthene or its alteration products. Biotite (up to 5%) forms poikilitic grains up to 3mm across enclosing plagioclase laths. Most specimens have small amounts of interstitial quartz, and all contain minor opaque minerals and accessory apatite. Secondary amphibole, epidote and chlorite are common.

Finer grained rocks of the Rule Gabbro are augite microgabbro in which augite forms stumpy prisms rather than ophitic grains.

Geochemistry

Only one sample of the Rule Gabbro was analysed, and its mafic composition makes it impossible to assign it to a geochemical group. A notable feature is the high Zr content of 270ppm.

Age

Relationships with the Littlemore Granodiorite, Lawyer Granite and Robert Granite, all of which have given Triassic dates, indicate a Triassic age for the Rule Gabbro.

Stratigraphic relationships

Interpretation of airborne magnetic data suggests that the Rule Gabbro intrudes the Littlemore Granodiorite and Lawyer Granite, and is intruded by the Robert Granite. It also intrudes unnamed Permian sedimentary and volcanic rocks along its eastern margin, and is intruded by the Deception Quartz Monzonite. Its relationship with Triassic volcanic rocks, which are interpreted to extend along the eastern side of the Deception Quartz Monzonite as far as the Rule Gabbro, is unknown.

Economic significance

No mineralisation is known to be associated with the Rule Gabbro.

Robert Granite (Rggo)

Introduction

The Robert Granite is a relatively non-magnetic leucocratic granite to quartz monzonite with granophyric texture which occurs in the centre of the doughnut-shaped magnetic high attributed to the Rule Gabbro. It was largely mapped as Triassic Muncon Volcanics by Dear & others (1971), but no volcanic rocks were observed during the recent mapping program.

Distribution

The Robert Granite is an oval intrusion 8km long and 5km wide with its long axis oriented north-east, and is centred about 10km west of the township of Many Peaks (Figure 361).

Derivation of name

The name is derived from Mount Robert, the highest point within the outcrop area.

Type area

The type area is in the headwaters of Prizemans Creek, which provides easily accessible and typical exposures.

Figure 361. Distribution of the Robert Granite

Topographic expression

The Robert Granite forms a dissected plateau ranging in altitude from 500 to over 900m, with deeply incised streams cut down to 200m. The extremely steep topography and thick forest cover make access to most of the intrusion difficult. Locally the granite forms low cliffs (Figure 299).

Geophysical expression

On images derived from airborne data, the Robert Granite appears as a weakly magnetic body because of the intensity of the doughnut shaped anomaly associated with the Rule Gabbro. This is confirmed by modelling (R. Deakin, in Rees, 1997). Magnetic susceptibility averages 1270 x 10^{-5} SI units.

Although it is an alkaline intrusion, it gives a moderate and very patchy radiometric response, with no coincidence between highs for different channels, particularly potassium and thorium. The unexpectedly low response is partly due to the extreme topography. An exception is a small area at the south-western extremity of the granite, where high values for potassium, thorium and uranium coincide. This area was not visited because of difficult access, and the cause of the anomaly is not known.

The Robert Granite has no apparent gravity expression.

Lithology and petrography

The Robert Granite is an altered, pink to greyish pink, medium grained, sparsely porphyritic, leucocratic granite to quartz monzonite, commonly with miarolitic cavities and a granophyric groundmass.

Plagioclase is the most abundant mineral, comprising 30 to 40% of the rock. It occurs both as euhedral phenocrysts up to 10mm long (up to 15% of the rock), and as smaller laths. Cores of sodic andesine are zoned outwards to more sodic compositions, and in many cases to a K-feldspar rim forming up to 20% of the grain. K-feldspar rims are optically continuous with the feldspar in adjacent micrographic intergrowths. Plagioclase is altered and cloudy in all samples, and locally is partly replaced by prehnite, chlorite and calcite.

K-feldspar (25–30% of rock) and quartz (15–35%) occur mainly in micrographic intergrowths which form the groundmass between plagioclase laths. Both minerals are also present as equant grains, of which quartz is the most abundant and larger (up to 1.5mm across). The proportion of quartz is quite variable, and the rocks range in composition from granite to quartz monzonite.

Mafics make up from 5 to 10% of the rock. Hornblende is most abundant, usually forming elongate prisms, but is rarely unaltered. A sample from a prominent rocky knoll just north of Mount Robert contains a strongly pleochroic alkali amphibole, probably arfvedsonite. Biotite is common, but is less abundant than hornblende, and typically is altered to chlorite. Small stumpy prisms of augite occur locally, notably in the arfvedsonite-bearing sample, and possible altered fayalite has also been noted. Opaques are a minor component, and accessory apatite, sphene and zircon are present. Secondary minerals include epidote, chlorite, calcite and prehnite.

Geochemistry

Only one sample of quartz monzonite was analysed from the Robert Granite. The sample tends towards A-type granite chemistry, and falls in the Within Plate Granite fields on the discrimination diagrams of Pearce & others (1984) (Figure 301).

Age

A U-Pb zircon age of 230.2±3.7Ma from the Robert Granite (Appendix 1) indicates a Triassic, probably Late Triassic, age.

Stratigraphic relationships

The Robert Granite intrudes unnamed Permian rocks to the east, the Littlemore Granodiorite to the north, and the Rule Gabbro and Lawyer Granite to the south.

Economic significance

The Robert Granite is not known to be associated with any mineralisation.

Figure 362. Distribution of the Deception Quartz Monzonite

Deception Quartz Monzonite (Rggd)

Introduction

The Deception Quartz Monzonite is a relatively small mass of leucocratic quartz monzonite to granite very similar to the quartz-poor parts of the Robert Granite, and the two intrusions may be related.

Distribution

The Deception Quartz Monzonite covers a roughly triangular area 2.5km from east to west and just over 2km from north to south along Deception Creek, 8km west-south-west of the township of Many Peaks (Figure 362).

Derivation of name

The name is derived from Deception Creek, which flows through the intrusion.

Type area

The type area is along Deception Creek below the junction of its two main tributaries, and upstream along the northern tributary.

Topographic expression

The unit forms steep, dissected country south-east of Mount Robert, ranging in elevation from 200m in Deception Creek to 400m on ridges.

Geophysical expression

The edge of the prominent magnetic high surrounding the Robert Granite cuts right across the intrusion, indicating that the Deception Quartz Monzonite is not the cause of the anomaly. This is supported by magnetic susceptibility readings, which range from very low to moderate, and average 940 x 10^{-5} SI units. The magnetic pattern suggests that the Deception Quartz Monzonite is only a thin, sill-like body intruded into the magnetic unit, which must be present beneath the quartz monzonite at relatively shallow depth. This is consistent with modelling results which show that the top surface of the magnetic body in this area is at a depth of about 500m (Appendix 5).

The intrusion gives a moderate but patchy radiometric response from potassium and uranium, and a low response for thorium. The small separate northern outcrop of the unit is interpreted from the radiometric data.

Lithology and petrography

The Deception Quartz Monzonite is a pink, medium grained, equigranular, leucocratic hornblende-biotite quartz monzonite to granite. Micrographic textures and the presence of miarolitic cavities, some now filled with secondary minerals, are clear evidence of intrusion at shallow levels. Large andesitic dykes are common in the unit and comprise up to 50% of some outcrops.

Subhedral to euhedral plagioclase laths up to 6mm long make up from 50 to 65% of the rock. Relatively homogeneous cores of sodic andesine composition are zoned to rims of more sodic plagioclase and K-feldspar. Cloudy alteration is ubiquitous.

K-feldspar (20–30% of rock) and quartz (10–20%) form micrographic intergrowths interstitial to plagioclase laths. The K-feldspar in these intergrowths is optically continuous with rims on the plagioclase.

Biotite is about twice as abundant as hornblende, and together they comprise from 5 to 10% of the rock. Both are extensively altered to chlorite. Accessory opaques, apatite and zircon are present.

Miarolitic cavities are filled by euhedral epidote and quartz crystals projecting into the centre of the former void, now occupied by calcite and chlorite.

Geochemistry

No samples of the Deception Quartz Monzonite were analysed.

Age

Correlation with the Robert Granite suggests a Triassic age.

Stratigraphic relationships

The Deception Quartz Monzonite intrudes Permian sedimentary and volcanic rocks and the Rule Gabbro.

Economic significance

No mineralisation is known to be associated with this unit.

Munholme Quartz Diorite (PRgmu)

(C.G. Murray)

Introduction

The Munholme Quartz Diorite is a newly defined unit that was not shown on the 1st edition Monto 1:250 000 geological map (Dear & others, 1971).

Distribution

The Munholme Quartz Diorite is a small intrusion 2.5 by 1.5km in extent, located on the western side of Munholme Creek about 40km north of Monto (Figure 363).

Figure 363. Distribution of the Munholme Quartz Diorite

Derivation of name

The name is from Munholme Creek, immediately east of the intrusion.

Type area

The type area is along an unnamed north-east flowing tributary of Munholme Creek in the northern part of the outcrop.

Topographic expression

The Munholme Quartz Diorite forms low, locally steep ridges.

Geophysical expression

The intrusion has a moderate to low radiometric signature that gives a red colour on images derived from airborne data. Despite moderately strong susceptibility readings averaging 2175 x 10^{-5} SI units, it has a relatively low overall magnetic response, and is surrounded by a prominent magnetic halo developed in the contact rocks.

Lithology and petrography

The dominant rock type is a grey, medium grained biotite-hornblende quartz diorite to tonalite, with subsidiary biotite-augite-hornblende diorite.

Plagioclase is the most abundant mineral in all rock types, making up 70% or more. It occurs as subhedral laths with cores of calcic andesine, dusty in some specimens, zoned to more sodic rims. Interstitial quartz grains range from minor to more than 10%. Mafic minerals comprise 20 to 30% of the rocks. Pale green hornblende is dominant, often with augite cores. The amount of relict augite decreases as the proportion of quartz increases. Biotite is always present but is subordinate to hornblende. Very pale secondary amphibole in some specimens may replace hypersthene, and epidote occurs locally. Minor and accessory minerals include opaque grains, apatite, sphene and zircon.

Geochemistry

The Munholme Quartz Diorite is placed in the Monal group (Figure 305).

Age

The intrusion has not been dated. It is assigned a Permian to Triassic age consistent with most plutons in the region.

Stratigraphic relationships

The Munholme Quartz Diorite intrudes the Three Moon Conglomerate and Rockhampton Group, which forms some roof pendants.

Economic significance

The intrusion hosts a variety of mineralisation styles. The Silver Star deposit, also known as the Barite lode, consists of silver and base metal mineralisation hosted by barite-quartz-carbonate veins in altered diorite. Sulphides include pyrite, galena, sphalerite, tetrahedrite, pyrargyrite, proustite and chalcopyrite, with variable gold grades (Mortensen, 1994). To the east of the Silver Star, pyrite, chalcopyrite and molybdenite occur as disseminations and fracture coatings in altered quartz diorite, and replacing the matrix in breccias and conglomerates of the Rockhampton Group (Mortensen, 1994). This mineralisation was classified as a porphyry-style deposit in the compilation of Horton (1982). In the northern part of the intrusion, high gold values in stream sediments led to a drilling program at the Nestor prospect. Gold is associated with sulphides in calcite-quartz and calcite-chlorite-quartz veins in altered diorite to tonalite and in contact rocks of the Rockhampton Group (Pope, 1991).

Bororen Tonalite (Pgbo)

(C.G. Murray)

Introduction

The geophysical pattern of a small intrusion around the small township of Bororen distinguishes it from the rest of the Miriam Vale Granodiorite. Larger outcrops occur in the Miriam Vale 1:100 000 sheet area to the east, where it has been named the Bororen Tonalite.

Distribution

The section of the intrusion on the Calliope 1:100 000 sheet area is 5km long and 2km wide, oriented in a north-north-west direction. A small area along Cedar Creek 15km east-south-east of Many Peaks has also been mapped as Bororen Tonalite based on its geophysical expression (Figure 364).

Derivation of name

The name is taken from the small township of Bororen, near the western margin of the intrusion.

Topographic expression

The Bororen Tonalite is deeply weathered, and forms low topography.

Geophysical expression

The radiometric response of the intrusion is low, and it gives dark, almost black colours on images derived from airborne data. It has a low to moderate magnetic signature, and the pluton at Bororen stands out most clearly on images of the first vertical derivative of the airborne magnetic data.

Figure 364. Distribution of the Bororen Tonalite

Lithology and petrography

The Bororen Tonalite is deeply weathered, and the surface expression is a granitic soil (grus). The original rock type was a biotite tonalite, and the biotite was obviously important in the weathering process (*c.f.* Isherwood & Street, 1976). Remnants of a foliation can still be seen.

Age

The Bororen Tonalite is assigned a Permo-Triassic age.

Stratigraphic relationships

The outcrop near Bororen is surrounded by the Miriam Vale Granodiorite – Jackass Gabbro, suggesting that it intrudes the latter units. Its relationship with the Molangul Granite south-east of Many Peaks is unknown.

Economic significance

No mineralisation is known from the Bororen Tonalite.

Molangul Granite (Rg and Rg?)

(R.J. Bultitude and C.G. Murray)

Introduction

The Molangul Granite is exposed mainly on the Rosedale 1:100 000 Sheet area, but extends west into the Monto 1:100 000 Sheet. It was not visited during the Yarrol Project, but thin sections from the original

Figure 365. Distribution of the Molangul Granite

mapping of the Monto 1:250 000 Sheet (Dear & others, 1971) were examined. This description is based on mapping for the Bundaberg Project.

Distribution

The Molangul Granite forms rugged, mountainous country in the north-east of the Monto 1:100 000 Sheet area, extending for 18km north-south and up to 5km east-west (Figure 365).

Derivation of name

The unit is named after Mount Molangul, which is located on the Rosedale 1:100 000 Sheet area.

Geophysical response

The geophysical response suggests that the Molangul Granite is a composite body. South of Cedar Creek, the granite has a high radiometric and a low magnetic signature, and is displaced by about 5km sinistral movement on the Mount Bania Fault. This area includes the type section on the southern flanks of Mount Molangul. North of the outcrop of Bororen Tonalite mapped in Cedar Creek, the granite has a slightly lower radiometric and greater magnetic response, and is mapped as **Rgm?**. Thin sections from this area include two granite types.

Lithology and petrography

On the Rosedale Sheet area, the dominant rock type is pale to dark pink, fine- to medium-grained, uneven grained to slightly porphyritic, amphibole-bearing syenogranite. The syenogranite is leucocratic, generally containing ~2–4 % Fe-Mg minerals (mainly amphibole). Grainsize ranges from ~0.2 mm to ~5 mm, the average grainsize in most samples being ~1 mm to 1.5 mm. Granophyric intergrowths between quartz and K-feldspar are common. Small (up to ~1 cm across), irregular miarolitic cavities are also common.

The unit is characterised by the presence of an alkali-rich (sodic) amphibole (arfvedsonite?) as the main Fe-Mg mineral (~2-3 %). The arfvedsonite? is pleochroic from brown, pale brown or pale violet-brown to pale bluish green, pale indigo, indigo, or dark blue, and forms tabular and acicular grains. The larger amphibole grains are commonly zoned.

Some of the rocks also contain traces of aegirine (pleochroic from bright green to pale yellow) and/or aenigmatite (red-brown; generally partly rimmed by amphibole).

Opaque oxide, zircon, allanite, apatite, fluorite, titanite, and biotite (mainly partly rimming amphibole grains) are the main accessory minerals.

The rocks are almost invariably slightly to moderately altered.

Thin sections from north of Cedar Creek are of two distinct granite types. One has a granophyric texture, with plagioclase laths rimmed by K-feldspar intergrown with quartz. The rocks are devoid of mafic minerals except for sparse, small opaque grains. The other type consists of approximately equal proportions of plagioclase, quartz and K-feldspar, with green to brown hornblende as the main mafic mineral. Scattered biotite grains are completely altered to chlorite.

Geochemistry

The Molangul Granite is an A-type, alkaline to peralkaline, subsolvus granite. On granite discrimination diagrams of Pearce & others (1984), it falls in the Within Plate Granite field (Figure 301).

The Molangul Granite is characterised by relatively high $Na_2O + K_2O$, Na_2O, Zr, Nb, Y, Sc, F, Zn, Ga, Sn, and REE, and low Mg number, MgO, CaO, and Sr. It is similar to silicic granites of the Miriam Vale Granodiorite, the Castletower Granite, and the Coulston and Winterbourne Volcanics (*c.f.* Figures 174, 294 and 301).

Relationships and age

The Molangul Granite intrudes the Lower Carboniferous Wandilla Formation. Its relationships to the Triassic Bobby Volcanics, to unnamed Permian to Triassic gabbro, and to the Bororen Tonalite are unknown.

The unit has not been isotopically dated. It was thought to be most probably Late Triassic because of similarities to other Late Triassic granites in the region — principally its felsic, highly leucocratic character, its geochemical characteristics (A-type), and its mode of outcrop (forms rough mountainous country). However, recent dating indicates that the Castletower Granite, which exhibits all of these characteristics, is at least in part of mid-Permian age.

Economic significance

No significant mineralisation is known to be associated with the Molangul Granite.

Norton Tonalite (PRgno)

(C.G. Murray)

Introduction

The Norton Tonalite is a newly defined unit that was previously mapped as an isolated intrusion of the Miriam Vale Granodiorite.

Distribution

The intrusion is an irregularly shaped body about 6km across located about 35km south-south-east of Calliope (Figure 366). It is interpreted to be offset by a sinistral strike-slip fault with about 2km of displacement.

Derivation of name

The name is taken from the Norton goldfield that is hosted by the intrusion, and Norton Creek that flows through it.

Figure 366. Distribution of the Norton Tonalite

Topographic expression

The Norton Tonalite forms low hills that contrast with the rugged timber covered area of the Castletower Granite to the east, and the sharp ridges of the Wandilla Formation to the west.

Geophysical expression

The pluton has a moderate radiometric response characterised by dark red hues on images produced from airborne data. It has a low to moderate magnetic signature, significantly higher than that of the surrounding rock units.

Lithology and petrography

The Norton Tonalite is a very uniform, grey, medium grained, leucocratic hornblende-biotite tonalite.

Plagioclase makes up about 60% of the rock, and consists of subhedral to anhedral grains strongly zoned from altered andesine cores to sodic rims. Anhedral quartz comprises about 30%, with very sparse K-feldspar as interstitial to poikilitic crystals. The proportion of mafic minerals is no more than 10% of the rock, with biotite more abundant than hornblende. Accessory minerals include opaque grains, apatite, sphene, zircon, and secondary epidote.

Geochemistry

The Norton Tonalite is part of the Dumgree group. Compared with other members of the group, it has slightly higher TiO_2 and Ce (Figure 302).

Age

The Norton Tonalite has not been dated, and is assigned a Permian to Triassic age.

Stratigraphic relationships

The Norton Tonalite intrudes the Wandilla Formation, and the outcrop pattern also suggests that it cuts the Castletower Granite and the Miriam Vale Granodiorite.

Economic significance

The intrusion hosts numerous sulphide-rich gold-bearing reefs of the Norton goldfield.

Pack Granite (PRgpa)

(C.G. Murray)

Introduction

The Pack Granite was first mapped by Troensegaard (1969) who delineated two outcrops of muscovite-bearing granite to the east of the Mount Seaview Igneous Complex, and referred to these as the Crow Creek granite.

Distribution

The Pack Granite is located 35km south-west of Calliope, just north of Dan Dan Creek. Together the two outcrops cover an area of only about $1km^2$ (Figure 367).

Derivation of name

The intrusions form the southern flanks of Mount Pack, from which the name is taken.

Type area

The type area is along Crow Creek, a tributary of Dan Dan Creek, which flows through the western and larger outcrop.

Topographic expression

The Pack Granite forms fairly steep hills on the southern slopes of Mount Pack, cut by the narrow valley of Crow Creek.

Geophysical expression

The granite has a moderately strong radiometric signature, and appears in shades of pink on images generated from airborne data. It is only weakly magnetic, similar to the surrounding sedimentary rocks, and consistent with magnetic susceptibility readings averaging 20×10^{-5} SI units.

Lithology and petrography

The only rock type is a pink, medium grained, equigranular, leucocratic muscovite-biotite granite to granodiorite with closely spaced joints. Quartz and plagioclase are present in approximately equal amounts, each constituting about 40% of the rock. Quartz occurs as equant anhedral grains up to 2.5mm across, and plagioclase (oligoclase) forms subhedral laths to 4mm long. K-feldspar (15% of rock) is mainly interstitial, but also rims plagioclase. Biotite, muscovite and opaques together make up only about 5% of the rock. Biotite is most abundant, but is completely altered to chlorite. Muscovite forms flakes up to 2mm across. Accessories include apatite and zircon. The cores of plagioclase laths are extensively altered to sericite and a little calcite, and needles of secondary rutile occur in chlorite replacements of biotite.

Geochemistry

A single sample was analysed from the Pack Granite, and falls within the Ridgelands group (Figure 306).

Age

No age determination has been carried out on the Pack Granite, and the intrusion is assigned a Permian to Triassic age.

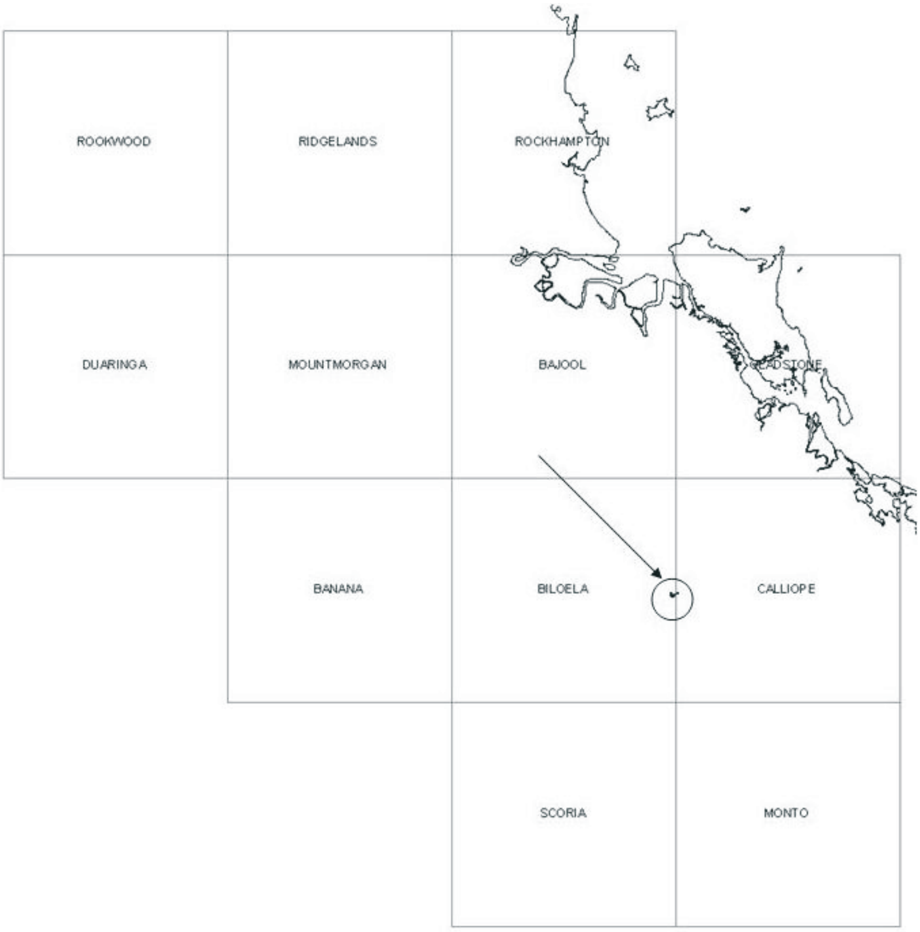


Figure 367. Distribution of the Pack Granite

Stratigraphic relationships

Both of the outcrops intrude the Rockhampton Group, and the western one also intrudes the Mount Alma Formation.

Economic significance

Gold has been mined from a sulphide-bearing quartz vein in the western outcrop of the Pack Granite, and this appears to have been the source for alluvial gold workings in Crow Creek.

Lookerbie Igneous Complex (PRglo)

(M.A. Hayward)

Introduction

The Lookerbie Igneous Complex is an intrusive complex centred 25km south of Biloela. Placer Exploration first reported the occurrence of small granitic and granodioritic intrusions in the area and coined the name Lookerbie Igneous Complex while exploring the Permian Rannes beds (now mapped in this area as the Yaparaba Volcanics) for Carlin-style gold mineralisation (Taylor, 1988).

Distribution

This unit occurs to the north-east of Mount Lookerbie as a north-west trending belt approximately 6km wide by 15km long (Figure 368).

Derivation of name

The unit takes its name from Mount Lookerbie at GR257100 7274700 on the Scoria 1:100 000 Sheet.

Figure 368. Distribution of the Lookerbie Igneous Complex

Type area

Numerous tors crop out on the eastern bank of Grevillea Creek just downstream from the junction with Saddle Gully at GR263100 7275700.

Topographic expression

The unit forms undulating hilly topography and generally crops out poorly. A noticeable exception is the more deeply incised topography and correspondingly better outcrop in the type area as defined above.

Geophysical expression

Unlike the Kariboe Gabbro, the Lookerbie Igneous Complex produces a very low response on a total magnetic intensity image. A weak north-west trending fabric is visible which possibly represents dykes. The Lookerbie Igneous Complex produces only a weak response on the total count radiometric image

Lithology and petrography

Rock types include diorite, gabbro, tonalite, granodiorite, quartz diorite and andesite.

The plutonic rocks are commonly quite strongly altered, manifested by sericitisation of plagioclase and chloritisation of mafic minerals.

Geochemistry

The Lookerbie Igneous Complex is a member of the Ridgelands group (Figure 306).

Age

The unit is of probable Permian–Triassic age.

Figure 369. Distribution of the Mannersley Quartz Microdiorite

Stratigraphic relationships

The generally poor outcrop in the area rendered observation of contacts quite difficult. Rocks of the adjacent Permian Yaparaba Volcanics are hornfelsed near the margins of the Lookerbie Igneous Complex indicating a post-Permian age for the intrusion.

Economic significance

Neither fieldwork or a detailed literature review (Blake & others, 1996) discovered any evidence of known mineralisation associated with this unit.

Mannersley Quartz Microdiorite (PRgma)

(C.G. Murray)

Introduction

The Mannersley Quartz Microdiorite was mapped and briefly described during company exploration programs.

Distribution

This small pluton is located almost 40km west of Calliope, less than a kilometre east of the Zig Zag Tonalite, and separated from it by the ridge of Rockhampton Group forming Mount Grim (Figure 369). It is 2.5km long and almost 1.5km wide.

Derivation of name

The name is taken from the Mannersley porphyry copper prospect (Horton, 1982), which is hosted by the intrusion.

Type area

The type area is along an unnamed tributary of Harper Creek that flows north-east through the intrusion.

Topographic expression

The Mannersley Quartz Microdiorite covers low, undulating country that has been cleared for grazing.

Geophysical expression

The intrusion has a low radiometric signature and a moderate magnetic response, both of which are similar to those of the surrounding Rockhampton Group hornfelsed by the Galloway Plains Igneous Complex to the south.

Lithology and petrography

Detailed company mapping found that the intrusion is dominantly a quartz diorite porphyry that surrounds a smaller stock of medium to coarse grained biotite quartz diorite (Taube, 1976). A sample from the type area is a pink to grey porphyritic quartz microdiorite. It consists of abundant microphenocrysts of strongly zoned plagioclase to 2.5mm long, sparse resorbed quartz grains, and rare biotite and altered ?hornblende (now calcite plus chlorite) in a fine grained groundmass of plagioclase, chlorite (after biotite), quartz, opaques, sphene, apatite, and secondary calcite. The rock contains abundant secondary biotite along joints.

Geochemistry

The Mannersley Quartz Microdiorite falls in the Dumgree group (Figure 302).

Age

No dating has been carried out, and a Permian to Triassic age is assigned.

Stratigraphic relationships

The Mannersley Quartz Microdiorite intrudes the Rockhampton Group.

Economic significance

The intrusion is host to the Mannersley porphyry copper prospect (Horton, 1982). According to Taube (1976), the mineralisation is mainly confined to the late stage coarser-grained quartz diorite.

Zig Zag Tonalite (PRGzz)

(C.G. Murray)

Introduction

The Zig Zag Tonalite was first delineated by exploration company mapping, but not described.

Distribution

The Zig Zag Tonalite is a small oval pluton, just over 2km long and 1km wide, located 40km west of Calliope, near the northern margin of the Galloway Plains Igneous Complex (Figure 370).

Derivation of name

The name is from Zig Zag Creek, which flows southwards through the centre of the intrusion.

Figure 370. Distribution of the Zig Zag Tonalite

Type area

The type area is a low ridge west of Zig Zag Creek, where the tonalite is exposed as sparse boulders.

Topographic expression

The intrusion forms low, cleared country along the valley of Zig Zag Creek, and for the most part is deeply weathered and devoid of outcrop.

Geophysical expression

The Zig Zag Tonalite gives a moderate radiometric response, with orange hues on images generated from airborne data. It is easily distinguished from the dark tones of the Mount Alma Formation to the west, but is similar to the Rockhampton Group at Mount Grim to the east. The intrusion also has a moderate magnetic signature, similar to that of the hornfelsed country rocks around the Galloway Plains Igneous Complex to the south.

Lithology and petrography

Only a single outcrop was located and examined. The rock is a pale grey, medium grained, equigranular hornblende-biotite tonalite. Plagioclase (about 45% of rock) occurs as subhedral laths of zoned andesine up to 3mm long, with cores extensively altered to sericite. Quartz (35% of the rock) and K-feldspar (less than 5%) form interstitial grains. Mafic minerals comprise 15%, biotite (partly altered to chlorite) being more abundant than green hornblende. Accessories include opaques, sphene, apatite, and zircon, and some secondary epidote is also present.

Geochemistry

The Zig Zag Tonalite falls within the Dumgree group (Figure 302).

Age

No dating has been carried out, and the Zig Zag Tonalite is assigned a Permian to Triassic age.

Stratigraphic relationships

The pluton intrudes the Mount Alma Formation to the west, and the Rockhampton Group to the east, and does not appear to be offset by the faulted contact between these two units.

Economic significance

No mineralisation is known to be associated with the Zig Zag Tonalite. However, analysis of fresh material from the outcrop in the type area gave a high copper content of 2120ppm.

Targinie Quartz Monzonite (PRgta)

(C.G. Murray)

Introduction

This intrusion was named the Targinie Adamellite by Kirkegaard & others (1970). It was examined only briefly during the Yarrol Project, but chemical analyses show that it is a quartz monzonite rather than a monzogranite.

Distribution

The Targinie Quartz Monzonite is an elongate intrusion slightly more than 10km long in a north-south direction, and up to 3km wide, centred about 15km west-north-west of Gladstone. As well as the main body, some small outcrops to the west are included (Figure 371).

Figure 371. Distribution of the Targinie Quartz Monzonite

Figure 372. View of Mount Larcom from the east. The rocky outcrops forming the summit are an alteration zone consisting of equal proportions of quartz and alunite. The slopes below the summit are of Targinie Quartz Monzonite

Topographic expression

The Targinie Quartz Monzonite has a varied topographic expression. Most of the quartz monzonite forms low country drained by streams flowing into The Narrows, and is largely concealed by alluvium. To the west, the granitic rocks crop out as the eastern slope of the Mount Larcom Range, and reach considerable elevations (Figure 372).

Geophysical expression

The intrusion has both strong radiometric and magnetic signatures on images derived from airborne data. It shows as white on radiometric images, due to a high content of potassium, thorium and uranium. The magnetic pattern indicates that the intrusion continues further to the north-east than surface exposures would suggest. Readings of the magnetic susceptibility give a result of 2335 x 10^{-5} SI units, consistent with its strong response on images of the airborne magnetic data.

The peak of Mount Larcom, which is interpreted as an altered phase of the Targinie Quartz Monzonite, also has a strong radiometric response, but the magnetic signature is much lower.

Lithology and petrography

The dominant rock type is a medium grained granitoid consisting of microperthite (>40%), sodic andesine (<35%), quartz (<15%), biotite (partly altered to chlorite) and green hornblende (together 10%), minor opaques, and accessory apatite and zircon. In some samples the quartz occurs in micrographic intergrowths with K-feldspar. Biotite is slightly more abundant than hornblende. The equant habit of the opaque mineral suggests that it is probably magnetite.

Alteration

The peak of Mount Larcom is an altered rock consisting of subequal proportions of quartz and alunite (Figure 372). This has previously been referred to as an altered trachytic plug, and assigned a Cretaceous age (Kirkegaard & others, 1970; Donchak & Holmes, 1991), but no other plug of this age shows a similar degree of alteration. Therefore the alteration zone is here interpreted as a phase of the Targinie Quartz Monzonite.

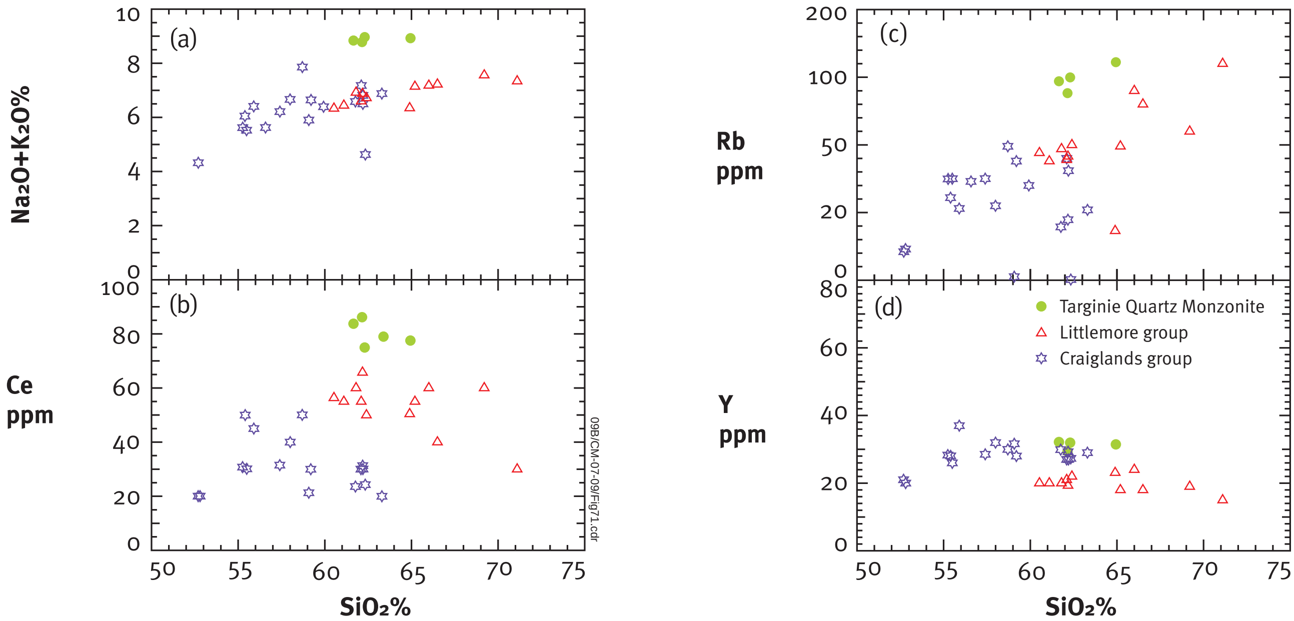


Figure 373. Harker diagrams comparing analyses from the Targinie Quartz Monzonite with the Littlemore and Craiglands groups. The Targinie Quartz Monzonite is notably higher for total alkalis, Ce and Rb and less so for Y. Analyses from GSQ and B. Chappell & R. Bultitude (unpublished data)

Chemistry

Analyses of the Targinie Quartz Monzonite do not fall into any of the other groups of Permian–Triassic intrusions defined on geochemistry. It is most like the Littlemore and Craiglands groups, but is richer in total alkalis, Ce, and Rb, and is richer in Y than the Littlemore group (Figure 373).

Age

The age of the Targinie Quartz Monzonite is unknown, but is probably Permian to Triassic like most intrusives in the project area.

Stratigraphic relationships

The quartz monzonite intrudes the Devonian to Lower Carboniferous Balnagowan Volcanic Member of the Doonside Formation, and elements of the mid-Permian Berserker Group. It is overlain by Tertiary and Quaternary sediments.

Economic significance

The intrusion hosts and appears to be the source for several gold-bearing reefs in the Targinie area. The secondary uranium mineral torbernite was found in one of these deposits, but is of academic interest only (Davies, 1955). The uranium mineralisation is probably related to the enrichment of the Targinie Quartz Monzonite in radioactive elements, including uranium.

Riverston Granodiorite (PRgrv)

(C.G. Murray)

Introduction

The Riverston Granodiorite was previously mapped as an undifferentiated intrusion of Permian to Triassic age (Dear & others, 1971). It was not examined during the Yarrol mapping program, and this description is taken from company exploration reports.

Distribution

The intrusion is a rectangular shaped body 3 by 2km located 8km east-south-east of Calliope, at the northern edge of Lake Awoonga (Figure 374).

Figure 374. Distribution of the Riverston Granodiorite

Derivation of name

The name is derived from Riverston Creek that flows southwards through the intrusion.

Topographic expression

The intrusion forms a low area compared to the surrounding country rocks. Microgranite and porphyritic rhyolite bodies form east-west ridges.

Geophysical expression

The Riverston Granodiorite is clearly outlined on images derived from airborne radiometric and magnetic data. Its radiometric response is dominated by potassium, giving it a deep red colour, and it has a moderate to high magnetic signature,

Lithology and petrography

The main rock type is referred to in reports as a grey, equigranular to porphyritic biotite granodiorite to diorite. Petrographic descriptions do not indicate the presence of K-feldspar, noting that some samples are actually trondhjemite (Amax Australian Ventures Ltd, 1976), but the radiometric signature suggests that the pluton as a whole contains a significant amount of potassium.

Within the area covered by the granodiorite are numerous elongate east-west trending bodies of rhyolite and microgranite. Their shape suggests that they are dykes, but they are more altered and fractured than the granodiorite, and contact relations are ambiguous. If older than the granodiorite, they appear to be relicts of dykes, because they extend beyond the eastern contact of the pluton (Amax Australian Ventures Ltd, 1976).

Age

A Permian to Triassic age has been assigned to the intrusion.

Stratigraphic relationships

The Riverston Granodiorite intrudes the Wandilla Formation.

Economic significance

Old gold workings are centred on an area of intense brecciation, silicification and argillic alteration that forms a small hill and is interpreted as a breccia pipe 300 by 150m. Traces of copper oxides, chalcopyrite, molybdenite and gold occur in the old workings. The host rocks were originally interpreted as microgranite and porphyritic rhyolite, but drilling showed that the main breccia fragments are diorite or granodiorite cemented by coarse quartz, some calcite, and sulphides (including chalcopyrite and molybdenite) (Amax Australian Ventures Ltd, 1976). There is also an area of quartz stockworking in porphyritic rhyolite that is anomalous in copper and molybdenum. Horton (1982) classified the Riverston prospect as a porphyry style deposit.

Diglum Granodiorite (Rgdi)

(C.G. Murray)

Introduction

The Diglum Granodiorite was previously mapped as an unnamed intrusion of Triassic tonalite (Dear & others, 1971). Skarns are developed in large inclusions of country rock in the north-eastern part of the intrusion, and have been prospected by a number of companies. They were also the subject of a thesis by Muggeridge (1973).

Distribution

The intrusion is a oval shaped body 13 by up to 6km trending north-east and located 30km south-south-west of Calliope (Figure 375).

Figure 375. Distribution of the Diglum Granodiorite

Derivation of name

The name is derived from Diglum Creek that flows north-east through the intrusion.

Topographic expression

The intrusion forms a low area adjacent to Diglum Creek and its tributaries.

Geophysical expression

The Diglum Granodiorite is a composite body whose divisions are clearly outlined on images derived from airborne radiometric and magnetic data. The radiometric data indicate a felsic core surrounded by a more mafic rim. The central core of the intrusion gives a moderate radiometric response and appears pink in images, whereas the rim displays dark red colours.

The intrusion is surrounded by a strongly magnetic hornfelsed zone. Within the intrusion itself, the magnetic pattern is the opposite of what would be expected from the radiometrics. The core gives a moderate response, whereas the surrounding more mafic rim has a very low magnetic signature. This is contrary to magnetic susceptibility measurements, and suggests that the rim has a component of reversed remanent magnetisation. A separate intrusion of magnetite-rich gabbro, which does not crop out and is known only from drilling, is associated with a very strong magnetic high at the north-eastern extremity of the pluton.

Lithology and petrography

The central core (**Rgdi$_g$**) is a pinkish-grey, medium grained biotite-hornblende granodiorite. Plagioclase laths (andesine) make up about half the rock, and are enclosed in a fine-grained granophyric groundmass of quartz and K-feldspar, with quartz being more abundant (25 to 35% of rock). The mafic minerals are hornblende and biotite in approximately equal proportions, together comprising about 10% of the rock. Opaque grains are also present. The rim (**Rgdi$_t$**) ranges from biotite-hornblende quartz diorite to augite-hornblende-biotite tonalite and is composed of 55 to 70% plagioclase (andesine). Quartz makes up 10 to 20% of the rocks, K-feldspar less than 5%, and mafic minerals 15 to 20%, with minor opaques. In rocks containing augite, biotite is more abundant than hornblende, whereas the opposite is true if no augite is present.

Exploration drilling has shown that the strong magnetic feature to the north-east is due to a hornblende-augite gabbro with up to 10% magnetite (Neale, 1975; Rigby, 1991).

Age

The intrusion has been dated by K-Ar as Late Triassic (Webb & McDougall, 1968; Muggeridge, 1973).

Stratigraphic relationships

The Diglum Granodiorite intrudes the Mount Alma Formation and the Rockhampton Group.

Economic significance

Numerous remnants of the country rocks form discontinuous belts across the north-eastern and eastern margins of the intrusion. Drilling has shown that these are shallow rafts. Original limestones have been converted to skarns, which contain small and low grade copper mineralisation. More disseminated copper and molybdenum mineralisation occurs at the Booreco Creek prospect associated with altered rocks of the intrusion itself.

Eulogie Park Gabbro (Rgep)

(C.G. Murray)

Introduction

The Eulogie Park Gabbro is a layered gabbro that was named and described by Wilson & Mathison (1968). A summary of the geology was included in Kirkegaard & others (1970), and Bourke (1995) studied the variation in mineral composition from base to top. The intrusion was not examined in detail during the Yarrol mapping project, and this description is based on the sources listed above.

Figure 376. Distribution of the Eulogie Park Gabbro

Distribution

The Eulogie Park Gabbro is located about 25km south of Mount Morgan. It has been broken up into a number of separate outcrops by intrusion of later diorite, and by faulting. The largest segment, located at the north-eastern margin, is semicircular, with a diameter of 5km. The second largest occurs along the western boundary, and is 4km long, elongated north-south. Smaller bodies occur in the west and south. The largest semicircular segment may retain its original orientation, but the other outcrops appear to have been rotated relative to each other. The gabbro and diorite together form an ovoid body 10km long and 5 to 6km across (Figure 376).

Derivation of name

The name is taken from Eulogie Park homestead in the centre of the outcrop area.

Type section

To illustrate the regular igneous layering within the intrusion, Wilson & Mathison (1968) nominated a type section trending north-east across the widest part of the largest remnant, at right angles to the layering.

Topographic expression

The primary igneous layering in the main north eastern mass is reflected in semicircular ridges that rise almost 200m above the level of the diorite to the west. An elongate mass to the west also forms high rounded hills, but smaller remnants to the south are expressed as low bare rises.

Geophysical expression

The Eulogie Park Gabbro gives an extremely low radiometric response in all channels, and appears black in false colour composite images. The radiometric signature of the diorite is marginally higher.

The magnetic response on images generated from airborne data is also very low. Because the magnetic susceptibility of rock samples is high (some are natural magnets), this low response must be due to reverse remanent magnetisation. Only the outermost zone of the main semicircular mass has a strong magnetic signature, suggesting a reversal of the magnetic field during crystallisation. The diorite shows a uniform moderate to strong magnetic response.

The gabbro corresponds to the southern part of the Westwood Gravity High of Darby (1969). This high lies along the Marlborough Gravity Ridge, attributed by Murray & others (1989) to arc volcanics and their possible mafic-ultramafic basement. However, the gabbro intrusion must contribute substantially to the positive anomaly and is regarded as the source of the local gravity high in which it is centred.

Lithology and petrography

According to Wilson & Mathison (1968), the Eulogie Park Gabbro consists of repetitions of six main rock types: olivine gabbro, ferrigabbro, troctolite, leucogabbro, gabbro, and magnetite-rich rocks. The chief minerals are plagioclase, olivine, augite and magnetite, with minor ilmenite and uralite, and accessory hornblende, sulphides, hypersthene and apatite. The main semicircular mass in the north-east displays three broad subdivisions, with a lower section dominated by ferrigabbros and magnetite-rich rocks, a middle section of olivine gabbros and troctolites, and an upper section of olivine gabbros and leucogabbros.

The later diorite is a relatively uniform medium grained rock consisting mainly of plagioclase (andesine An_{44}) and hornblende, accompanied by some augite, hypersthene and quartz, rare biotite and opaques, and accessory apatite and zircon. Towards the western contact of the diorite, the rocks become finer grained, and the proportion of pyroxene increases, with augite slightly more abundant than hypersthene (Kirkegaard & others, 1970).

Structure

Layering is best developed in the main north-eastern segment of the gabbro, and is expressed by variations in the proportions of minerals (rhythmic layering) and a pronounced foliation due to the alignment of tabular plagioclase laths (igneous lamination). Wilson & Mathison (1968) recognised about 65 saucer-shaped layers dipping to the south-west at 30° to 35°, forming a conformable sequence 900m thick. The layers are remarkably continuous, and some as thin as one metre can be traced for up to 5km along strike.

The elongate western portion of the gabbro consists of about 450m of olivine gabbro and ferrigabbro, but the succession cannot be correlated with any part of the main sequence (Wilson & Mathison, 1968). This suggests a total thickness of at least 1350m.

Geochemistry

The gabbro is tholeiitic to transitional in composition (Bourke, 1995). Averaged analyses of the lower, middle and upper sections were presented by Wilson & Mathison (1968).

Both Wilson & Mathison (1968) and Bourke (1995) found that plagioclase becomes more calcic upwards through the layered sequence, with some anomalous oscillations, but that olivine and augite show much less variation in composition. They postulated that these trends are most readily explained by the addition of new magma during crystallisation. Plagioclase compositions from the elongate western section of the gabbro are consistently more sodic than those from the main mass, supporting the suggestion that this represents a separate and possibly later section of the intrusion (Bourke, 1995).

Age

The Eulogie Park Gabbro has not been dated. Plagioclase from diorite 2km north-west of Eulogie Park homestead gave a K-Ar date of 241.6±3Ma (Bourke, 1995; Holcombe & others, 1997b).

The airborne magnetic data suggest that crystallisation of the Eulogie Park Gabbro commenced during a normal interval, and was largely completed during a reversal. The diorite was intruded during a subsequent normally polarised period. The date from the diorite coincides with the Early to Middle Triassic boundary according to Gradstein & others (1994), at the start of a reasonably long period of normal polarity. The Eulogie Park Gabbro must be Early Triassic or older. The Early Triassic was a time of rapid alternations between normal and reversed polarity (Gradstein & others, 1994).

Stratigraphic relationships

The Eulogie Park Gabbro intrudes the Balaclava Formation and is intruded by unnamed diorite. The contact between the main north-eastern segment of the gabbro and the diorite appears to be fault controlled.

Economic significance

The magnetite-rich bands within the Eulogie Park Gabbro were first evaluated as a source of iron ore by Ball (1904), but are too thin to comprise a significant resource. In addition, Brooks (1956) reported a high TiO_2 content ranging from 15.1 to 23.7%, due to intermixed ilmenite (Wilson & Mathison, 1968). These deposits, and alluvial concentrations in streams draining the gabbro, have recently been prospected as a source of ilmenite (Garrad & others, 2000). The gabbro has also been explored unsuccessfully for platinoids.

Kariboe Gabbro (PRgka)

(M.A. Hayward)

Introduction

Gabbroic and dioritic rocks were first reported in the Spring Creek region, south-east of Biloela by Bellamy (1976). Mapping outlined an area of layered gabbro cropping out over approximately 1km^2. The gabbro was interpreted to be bordered by a body of diorite with a mapped extent of approximately 3km^2. Mapping, combined with petrographic and geochemical studies, enabled the recognition of four dominant rock types within the layered gabbro sequence, which Bellamy named the Kariboe Layered Gabbro Intrusion.

Subsequent mapping by White Industries (Ekstrom & Ilsley, 1983) did not distinguish between the diorite phase and Bellamy’s layered gabbro intrusion, assigning the name Spring Creek Diorite to the whole area of basic intrusive rocks.

Distribution

The known limits of this occurrence of basic intrusive rocks have been extended further west and south-west than previously mapped (Figure 377). The unit, which is now mapped over 20km^2, is dominated by hornblende gabbro, but also includes hornblende diorite, tonalite, quartz monzogabbro/diorite and rare hornblendite. Locally the rocks contain xenoliths of metasiltstone/schist. The new name Kariboe Gabbro is synonymous with White Industries usage of Spring Creek Diorite, but is greater in extent. The Kariboe Gabbro therefore includes rocks previously assigned to Bellamy’s Kariboe Layered Gabbro and also his related diorite phase.

Derivation of name

The Kariboe Town Reserve, after which the layered gabbro intrusion is named, was gazetted in 1873. The mining history of Kariboe township was discussed by Robertson (1982).

Type area

A type locality is designated in the vicinity of GR276400 7278800 adjacent to a farm track. Grey, medium to coarse-grained, equigranular diorite crops out just east of the track and grey, fine to medium, strongly layered hornblende gabbro crops out as a low hill approximately 500m west.

Topographic expression

Most of the country occupied by the Kariboe Gabbro is undulating, but locally forms low rounded hills.

Geophysical expression

The Kariboe Gabbro is readily recognised on aerial magnetic images due to its very high response in contrast to adjacent units, although the magnetic anomaly extends beyond the mapped extent suggesting that the contacts may dip shallowly north. Little internal variation within the unit is evident. Magnetic susceptibility values are consistent with the magnetic response. Values have a skewed distribution with a mode at 1200 x 10^{-5} SI Units and a mean of 3500 and a tail extending to about 15000.

The gabbro and diorite can be easily distinguished from the Wingfield Granite on radiometric images due to its very low signature in all three channels in contrast to the granite being high in all channels. However, there

Figure 377. Distribution of the Kariboe Gabbro

is no contrast with the Lochenbar Formation which consists mostly of mafic volcanic rocks and related sedimentary rocks.

Lithology and petrography

Within the layered gabbro sequence, Bellamy (1976) recognised macro layering on two scales: a layering greater than a few centimetres thick, and a second, larger scale layering which allowed subdivision into four zones of dominant rock types. These are ferrigabbro, olivine gabbro, hypersthene gabbro, and porphyritic augite gabbro. The layered gabbro body was interpreted to dip approximately 5° north-west.

As now mapped, the pluton also includes hornblende gabbro to diorite and also tonalite, quartz monzogabbro/diorite and rare hornblendite. Locally the rocks contain xenoliths of metasiltstone and schist.

At GR276100 7278800 in the Scoria 1:100 000 Sheet area, a small hill of hornblende gabbro exhibits some spectacular small-scale layering features. An outcrop on the north-eastern side of this hill exhibits alternating light (plagioclase) and dark (hornblende) bands approximately 1cm thick (Figure 378). This banding is almost identical in outcrop to Plate 8 in Hess (1960) which he describes as “Inch-scale layering”. This term was introduced to describe small scale layering in the Lower Gabbro Zone of the Stillwater Igneous Complex, formed by repetition of plagioclase and pyroxene bands. One model proposed for the formation of such layering is presented by McBirney & Noyes (1979, page 537). This model invokes successive cycles of nucleation and diffusion of a particular component in a magma, acting on a front parallel to the wall/floor of a magma chamber.

On the western side of the hill an outcrop exhibits this same layering which in places is ‘cross-bedded’ (Figure 379). This type of feature has been previously documented from layered intrusions including the Skaergaard intrusion (Wager & Brown, 1968) and the Duke Island ultramafic complex (Irvine, 1974).

As previously mentioned, Bellamy (1976) also recognised and mapped a diorite body which crops out on the southern and western margins of the layered gabbro. This diorite essentially comprises plagioclase and hornblende, with minor biotite and quartz (both up to 2%). Bellamy (1976) proposed that the diorite and the

Figure 378. Inch scale layering in hornblende gabbro, Kariboe Gabbro

Figure 379. Layering showing ‘cross-bedding’, Kariboe Gabbro

layered gabbro were geochemically related and that the diorite probably intruded the gabbro when the latter was still hot, indicating that they are of virtually the same age. The Eulogie Park Gabbro in the south of the Rockhampton 1:250 000 Sheet area displays a similar relationship, having been intruded by a large diorite body (Wilson & Mathison, 1968).

Age

In the absence of definitive dates or intrusive relationships, a tentative Late Permian – Early Triassic age is assigned to the gabbro.

Stratigraphic relationships

The Kariboe Gabbro has metamorphosed rocks of the Late Devonian Lochenbar Formation to hornblende hornfels facies. It is mapped in contact with the Wingfield Granite. White Industries suggested that the mafic rocks predated the Wingfield Granite, but Bellamy (1976) suggested that the Wingfield Granite was older. Fine grained pink aplite/pegmatite dykes intrude the diorite and suggest a genetic link between these and the Wingfield Granite. Accordingly the diorite would be the older of the two phases, although it is possible that the two could be of similar age and have undergone magma mingling as for the contact between the Delubra Gabbro and Cadarga Creek Granodiorite in the Rawbelle Batholith to the south (Withnall & others, 2009).

Economic significance

Copper mineralisation was discovered in the area in 1869 by Robert Buffet Ridler of Yarrol Station. Although the period 1869 to 1874 saw 31 mineral selections granted in the Spring Creek area, only two major deposits were worked. The larger of the two, the **Great Blackall Mine** (GR278600 7279700 on the Scoria 1:100 000 Sheet area) is located in the diorite, near the contact with Bellamy's olivine gabbro. The Great Blackall Mine comprised at least eight shafts, of which only two were extensive. One of these two was **Lizzies shaft** which was approximately 74m deep and had drives at 30 and 50m. The other major shaft was **Vickery's**. The orebody at Great Blackall was marked at the surface by blocks of malachite up to 60kg in weight. An early sample of ore weighing approximately 200kg was reported to be "the red oxide of copper" containing "from 50 to 88.8% of pure metal". The ore from the Great Blackall Mines was carted untreated to Gladstone, a very costly proposition given the remoteness and standard of access in that era. The smelters erected at Great Blackall in 1874 never smelted any ore.

Flanagans workings was the other group of mines in the Spring Creek area. These are located on a ridge about 1km north-east of the Great Blackall Mines (GR279200 7280500 on the Scoria 1:100 000 Sheet area). The ruins of the smelter and slag heap plot near the north-eastern boundary between Bellamy's olivine gabbro and hypersthene gabbro. At least 350t of ore was smelted at Flanagans Workings, with the ore reported to be "chiefly grey oxide averaging 20 to 25% metal".

Pechiney (Australia) Exploration Pty Ltd drilled three cored holes in the Kariboe Gabbro in the early 1970s. Great Blackall DDH1, 2 and 3, named after the major mine in the area, were drilled to 108.51m, 148.13m and 205.96m respectively. Drill logs are presented in French, but generally appear to log microdiorite passing downward into diorite and andesite. Representative core from these holes is stored, and available for viewing, at the Department of Natural Resources and Mines's Exploration Data Centre in Zillmere, Brisbane. Some of the core was sampled in 1983 by Mareeba Mining and Exploration Pty Ltd and submitted for gold and silver analyses. The highest assay results recorded 155ppb gold and 6ppm silver. Inspection of this core confirmed the layered nature of these mafic rocks with at least six distinct rock types noted in DDH 1. The bottom interval (75.82–96.79m) is represented by a coarse breccia with clasts of dominantly dark grey, coarse-grained gabbro/diorite up to 15cm, set in a groundmass of pale cream to medium-grey, mottled diorite. Pyritic veins and calcite + clay veins cut the rock. Pyrite also occurs disseminated through the rock, dominantly in the groundmass. It is uncertain whether the dioritic groundmass of the breccia represents the same body of diorite mapped by Bellamy, but if so, this would support his interpretation of the diorite intruding the layered gabbro. The brecciation may have been caused by hydraulic fracturing and/or stoping of the consolidated gabbro.

It is noteworthy that the Kariboe Gabbro appears to be one of a chain of gabbroic intrusions forming a north-west trending belt extending from north-west of Brisbane to south-west of Rockhampton. From south to north these intrusions are the Somerset Dam Igneous Complex, Wateranga Gabbro, Hawkwood Gabbro, Goondicum Gabbro, Kariboe Gabbro, Eulogie Park Gabbro and Bucknalla Gabbro. Several of these intrusions have been explored for platinum group metals and/or magnetite with varying success.

Some of the samples of banded gabbro collected during mapping were slabbed and polished. The rock takes a good polish and has an attractive black and very light grey banded appearance. Some portions of the Kariboe Gabbro may therefore be worthy of investigation in terms of potential for dark coloured facing stone.

TRIASSIC–JURASSIC

Bucknalla Gabbro (Rjgbu)
(C.G. Murray)

Introduction

The Bucknalla Gabbro was originally referred to as the Westwood intrusion (Hansen, 1971) or the Westwood layered gabbro (Clifford, 1987). However, the name Westwood Complex was used by Wood (1974) and Ford & others (1976) for the western lobe of the Bouldercombe Igneous Complex. To avoid confusion, Carrigg & others (1989) introduced the name Bucknalla Complex. This is changed to Bucknalla Gabbro to reflect the dominant composition.

The Bucknalla Gabbro was not examined in detail during the Yarrol mapping project because of previous detailed studies by Hansen (1971), Clifford (1987), Carrigg & others (1989) and Reeves & Keays (1995).

Distribution

The Bucknalla Gabbro forms a triangular mass with sides about 3km long immediately west of the township of Westwood (Figure 380).

Derivation of name

The name is derived from Bucknalla homestead, located about 1km north-west of Westwood within the outcrop of the intrusion.

Figure 380. Distribution of the Bucknalla Gabbro

Type area

The type area nominated by Carrigg & others (1989) is the prominent ridge 2km south-west of Bucknalla homestead.

Topographic expression

A prominent north-west trending ridge runs through the middle of the intrusion, with scree-covered slopes leading down to a flat area along the eastern boundary.

Geophysical expression

The Bucknalla Gabbro gives an extremely low radiometric response in all channels, appearing black on false colour images generated from airborne data. It has a very strong positive magnetic signature.

It also corresponds to a strong positive gravity anomaly included by Darby (1969) in the Westwood Gravity High, part of the Marlborough Gravity Ridge. The gravity high is centred on the south-western corner of the gabbro, and also includes the unnamed diorite-gabbro intrusion to the west of the Bucknalla Gabbro.

Lithology and petrography

The Bucknalla Gabbro is a layered gabbroic lopolith that consists of 15 mappable but laterally discontinuous units. Rock types include gabbro, olivine gabbro, leucogabbro, anorthosite, hornblende gabbro, feldspathic clinopyroxenite, olivine clinopyroxenite and wehrlite with lesser amounts of troctolite and ferrigabbro (Hansen, 1971; Reeves & Keays, 1995). Plagioclase, clinopyroxene and olivine are cumulus phases; hornblende occurs as late-stage poikilitic grains, and magnetite is mainly interstitial.

Structure

The intrusion is interpreted as a layered gabbroic lopolith that has been tilted to the north. Layering is expressed both as rhythmic layering caused by variation in mineral proportions, and as primary igneous lamination, but locally there appears to be angular discordance between the two (Hansen, 1971). The layering is concave to the north-east (Carigg & others, 1989). Hansen (1971) noted that the mafic-rich units are even textured, whereas the gabbroic units are laminated, and it has been suggested that the pyroxenite may form a discordant pipe-like mass underlying the layered gabbros (Carigg & others, 1989).

Dips are difficult to determine due to poor outcrop and the discontinuous nature of the layering, and estimates range from 35° (Hansen, 1971) to 70° (Clifford, 1987) to the north and north-east. The thickness of the layered sequence may be as much as 2000m.

Geochemistry

Whole rock analyses, as well as numerous determinations of platinum-group metals, were carried out by Reeves & Keays (1995). They described the intrusion as tholeiitic.

Mineral compositions vary erratically with stratigraphic height, and cover a relatively narrow range. A feature is the high Ca content of the plagioclase, most of which is anorthite (Hansen, 1971; Reeves & Keay, 1995).

Age

Plagioclase from the Bucknalla Gabbro was dated at 200.8±6Ma (Holcombe & others, 1997b). This age range falls within a period of dominantly reversed magnetic polarity (Gradstein & others, 1994), suggesting either that crystallisation of the gabbro was a rapid event during a brief period of normal polarity, or that the age is in error.

Stratigraphic relationships

The Bucknalla Gabbro intrudes the Rookwood Volcanics and the Flaggy Quartz Monzodiorite of the Bouldercombe Igneous Complex.

Economic significance

The Bucknalla Gabbro hosts minor Cu-Pd-Au-Pt mineralisation, first reported by Shepherd (1956), and described by Reeves & Keays (1995). Copper-bearing minerals include chalcopyrite, bornite, digenite and covellite. Several minerals containing gold and platinum group elements have been reported, including porpezite (palladian gold) (Ostwald, 1979). Reeves & Keays (1995) considered that the mineralisation was primary magmatic.

MESOZOIC

Warrong Teschenite (Mgwa)

(C.G. Murray)

Introduction

The presence of a mafic intrusion on the Dawes Range south of Dooloo Tops was recognised by exploration company geologists, and it was delineated on the 1st edition 1:250 000 Monto geological map. It is here described and defined as a new unit.

Distribution

The Warrong Teschenite is a small body located on the crest of the Dawes Range about 30km north of Monto, between Pine Mountain Creek to the west and Warrong Creek to the east (Figure 381).

Derivation of name

The intrusion occurs within the drainage area of the headwaters of Warrong Creek, from which the name is taken.

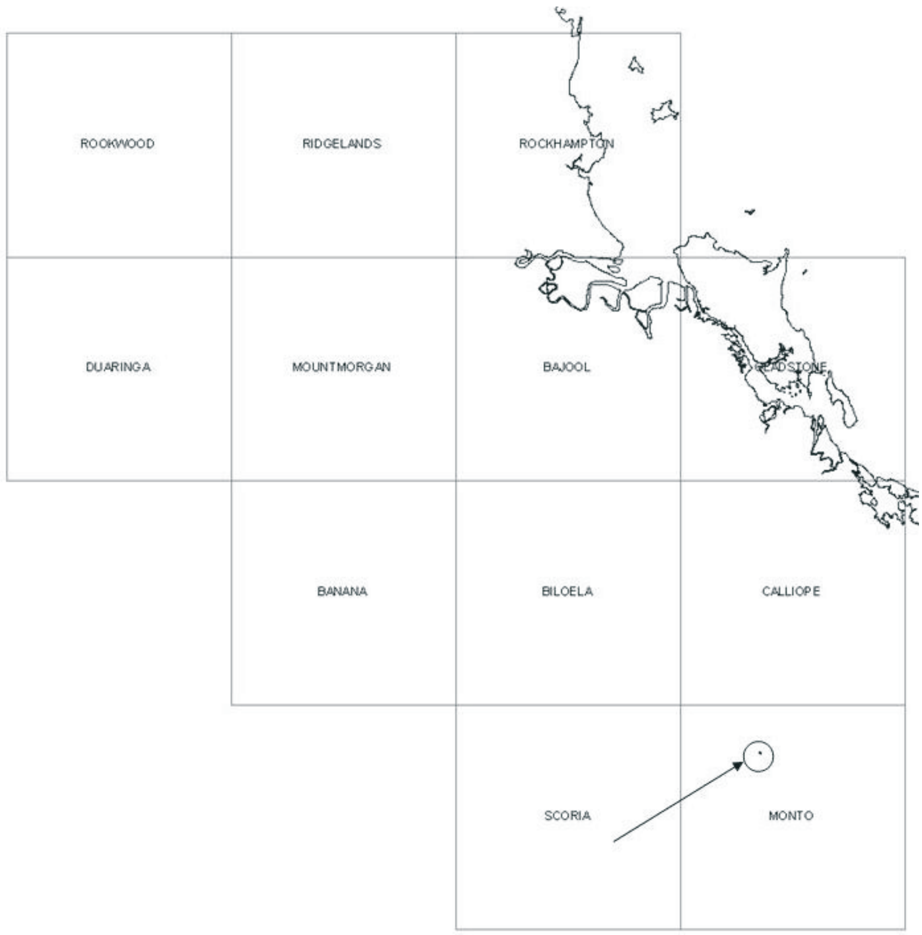


Figure 381. Distribution of the Warrong Teschenite

Type area

The type area is along the western side of the track south from Dooloo Tops homestead, where boulder outcrops can be seen.

Topographic expression

The teschenite forms part of the elevated, well forested Dawes Range.

Geophysical expression

The intrusion coincides with an area of low radiometric response, giving dark green hues on images generated from airborne data. Its radiometric signature contrasts markedly with that of the surrounding rhyolitic Dooloo Tops Volcanics, which give a high response in all channels and white tones on these images.

Although magnetic susceptibility readings are high, the teschenite corresponds to a magnetic low on airborne data, indicating reversely polarised remanence.

Lithology and petrography

Outcrops consist of a medium grained grey gabbroic rock showing poorly defined compositional banding. In thin section, it can be seen to be a typical teschenite.

Plagioclase is the most abundant mineral (40 to 45% of rock), forming euhedral to subhedral laths of calcic labradorite averaging about 1mm long. Analcite makes up from 10 to 15%, and occurs as partly altered interstitial grains. Deeply coloured ophitic to subophitic titanaugite up to 4mm across comprises about 20% of the rock. It typically shows sector zoning, and some grains have a narrow dark rim. Other mafic minerals are anhedral olivine (10% of rock), minor amounts of small prisms of dark brown to dark green hornblende, and rare biotite (Figure 382). Opaques are abundant, and constitute as much as 10% of the rock. Apatite is also relatively common as characteristic euhedral prisms.

Geochemistry

No analyses have been carried out on the Warrong Teschenite.

Age

The unit has been assigned a Mesozoic age.

Stratigraphic relationships

The Warrong Teschenite is assumed to intrude the Dooloo Tops Volcanics, although actual contact relationships were not observed. An unconformable relationship would require that the teschenite formed a prominent peak when the volcanics were erupted, which is considered unlikely. The alkaline nature of the intrusion suggests that it is not part of the Permian to Triassic suite of intrusives, but is more likely to be related to the earliest Cretaceous monzonite and nepheline syenite of the Glassford Igneous Complex.

Economic significance

No mineralisation is known to be associated with the Warrong Teschenite.

CRETACEOUS

Glassford Igneous Complex

(C.G. Murray)

Ridler Monzonite (Kggi)

Introduction

Neale (1968) described a complex syenitic intrusion in the headwaters of Ridler Creek, and informally named it the Ridler Syenite. Further field mapping was carried out by companies exploring for base metal

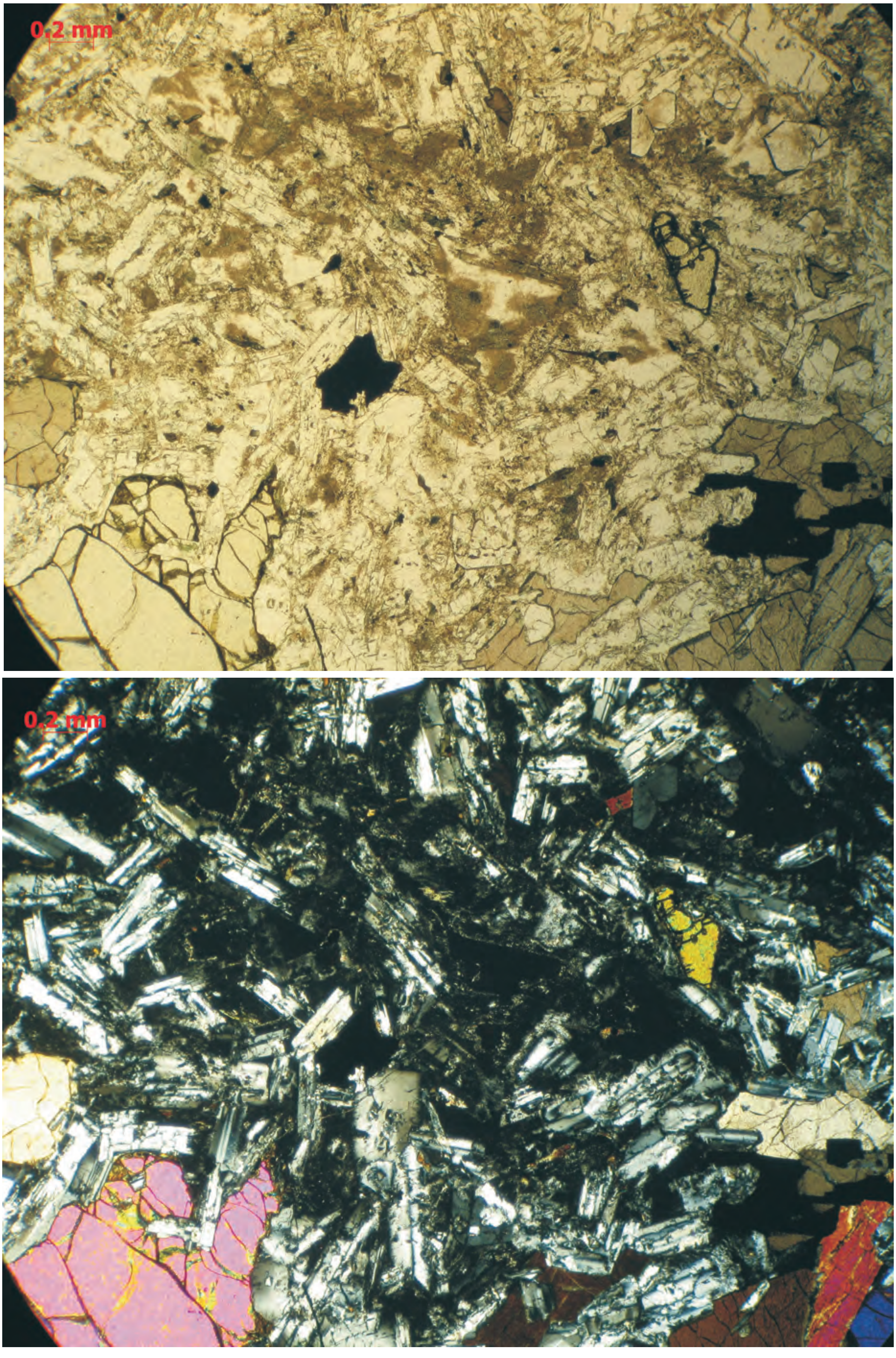


Figure 382. Photomicrograph showing plagioclase laths, olivine, titanaugite, partly altered analcite (for example, in the centre of the slide), opaques, and apatite, Warrong Teschenite. Plane polarised light and crossed nicols. Scale bar is 0.2mm long.

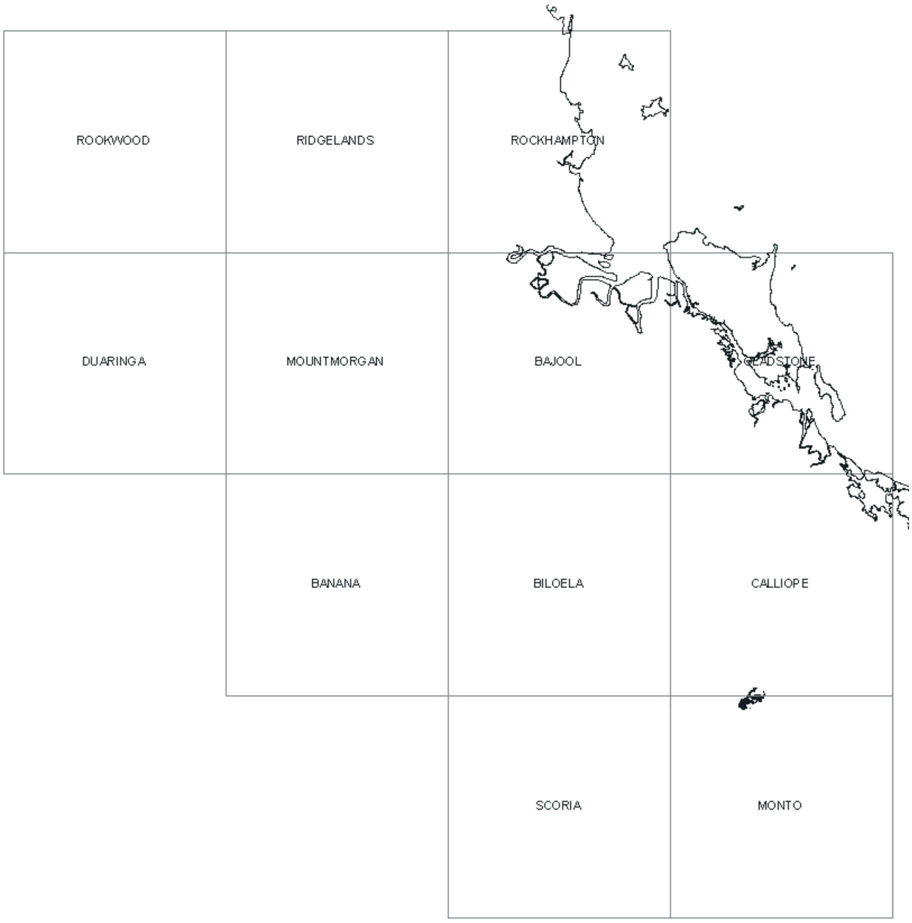


Figure 383. Distribution of the Ridler Monzonite

mineralisation (Climie & others, 1970; Roderick & others, 1971). The recent program of mapping has subdivided these rocks into 7 units, of which the Ridler Monzonite is the oldest and largest. It equates to the Mount Sugarloaf phase of Neale (1968).

Distribution

The Ridler Monzonite is an elongate north-east – south-west trending intrusion 7km long and 2.5km wide which extends along the north side of Ridler Creek from Mount Sugarloaf to the Dawes Range (Figure 383).

Derivation of name

The name is derived from Ridler Creek, which runs along the southern boundary of the intrusion and locally cuts through it.

Type area

The type area is Mount Sugarloaf, on the northern bank of Ridler Creek, and the lower ridges immediately to the west. Good exposures can also be seen in the headwaters of Tollbar Creek.

Topographic expression

The Ridler Monzonite forms hilly, locally steep country dissected by upper Ridler Creek and its tributaries. Elevations range from 450m at Mount Sugarloaf, and almost 600m along the eastern side of the Dawes Range, to just over 200m at the junction of Ridler and Tollbar Creeks.

Geophysical expression

Results from the AIRDATA airborne survey show that the intrusion does not give a uniform magnetic response, suggesting that it may be composite. The type area at Mount Sugarloaf is strongly magnetic, but this may be partly or largely due to the presence of an arcuate outcrop of amphibolite around the eastern side of the mountain. The amphibolite has a magnetic susceptibility of 10650 x 10^{-5} SI units, and some specimens will support a suspended magnet. In contrast, 7 samples of monzonite from the magnetic high average 1690 x 10^{-5} SI units, with two other more mafic samples giving a much higher average reading of 8020 x 10^{-5} SI units. These results are consistent with a model of the airborne data which has a large magnetic mass at depth with three localised magnetic bodies projecting upwards from it, all dipping at moderate angles to the south-east (R. Deakin, *in* Rees, 1997).

A second, less intense magnetic high, centred on the headwaters of Tollbar Creek, appears to be due to the intrusion itself, which in this area has an average magnetic susceptibility of 4920 x 10^{-5} SI units. The remainder of the Ridler Monzonite has a much weaker magnetic signature, and an average susceptibility of 2490 x 10^{-5} SI units. Comparing this value with typical readings from samples from the Mount Sugarloaf magnetic high reinforces the view that this high is due to the amphibolite, not to the intrusion.

In contrast to the varied magnetic signature, the radiometric response is high in all channels, although for thorium it is rather patchy. The response is considerably greater than that of adjacent Triassic intrusive units, and therefore is very useful in determining boundaries.

The Ridler Monzonite coincides with a small but distinct gravity high.

Lithology and petrography

The Ridler Monzonite covers the full compositional range from syenite, through monzonite and quartz monzodiorite, to monzogabbro and gabbro. The average composition is close to monzonite. Generally, the rocks can be distinguished from Triassic granitoids of the Glassford Igneous Complex by their low content of quartz and mafic minerals, finer grain size, lack of alteration, and close jointing.

Syenite is found mainly on Mount Sugarloaf. It is a pale grey to fawn, medium grained equigranular, leucocratic rock. K-feldspar makes up about 55% of the rock as subhedral laths averaging 5mm long, and rarely up to 15mm. Local alignment of K-feldspar laths produces a prominent foliation. Zoned subhedral to euhedral plagioclase laths up to 4mm long (25–30% of rock) occur as inclusions in the larger K-feldspar crystals, and as separate grains. Cores are calcic andesine in composition. Some myrmekite is developed at contacts with K-feldspar. Small interstitial quartz grains make up less than 5% of the rock. The main mafic minerals are colourless augite and green to green-brown hornblende in approximately equal amounts, with minor biotite and opaques. Altogether they comprise no more than 10% of the rock. Sphene and apatite are relatively abundant as euhedral crystals (Figure 384).

Monzonite occurs on the lower ridges west and north of Mount Sugarloaf. It is a grey, medium grained equigranular rock, locally with sparse mafic phenocrysts or xenoliths. Strongly zoned subhedral to euhedral plagioclase laths to 3mm long, with calcic andesine cores, are the main mineral (45–55% of rock). K-feldspar (25–35%) occurs as rims on plagioclase, as interstitial grains, and in some cases as poikilitic crystals more than 5mm across. Myrmekite development is generally restricted to samples containing poikilitic K-feldspar. Quartz makes up less than 5% of the rock and may be completely absent. Mafic minerals (10–25%) include clinopyroxene, hornblende and biotite. In rocks containing quartz, the pyroxene is a colourless augite, and the hornblende (if present) is green. Rocks devoid of quartz also lack biotite, and contain green aegirine or aegirine augite and dark green to brown ferrohastingsite. Both aegirine and ferrohastingsite are zoned, with rims darker than cores. It is possible that these rocks represent a separate phase related to the Radley Nepheline Syenite. Minor opaque grains and accessory sphene, apatite and zircon are present.

Quartz monzonite from the western extremity of the intrusion is similar in mineralogy to quartz-bearing monzonites, with approximately equal proportions of plagioclase and K-feldspar, and less than 10% quartz. The rock is pale grey in colour, leucocratic (mafic minerals about 10%), foliated, and in hand specimen is similar to the syenite forming Mount Sugarloaf.

Quartz monzodiorite is the main rock type along Tollbar Creek upstream from the Tollbar Breccia. It is a relatively uniform grey, medium grained equigranular hornblende quartz monzodiorite with or without biotite. Plagioclase is dominant (about 60%) as subhedral laths up to 4mm long with calcic andesine cores and thin zoned rims. K-feldspar and quartz both make up from 10 to 15% of the rock and occur chiefly as interstitial grains. K-feldspar rims some plagioclase laths. Mafic minerals also comprise 10 to 15% of the rock. Euhedral to subhedral prisms of green-brown hornblende up to 2mm long are much more abundant than biotite, which is completely absent in some specimens. Minor opaque grains and accessory apatite, zircon and euhedral sphene are present.

Monzogabbro and gabbro are found at two separate localities along Tollbar Creek, one north of Mount Sugarloaf, and the other at the head of the creek. Both occur within magnetic highs, suggesting that gabbro might be more abundant at depth. The gabbro is a grey, medium grained equigranular rock with locally

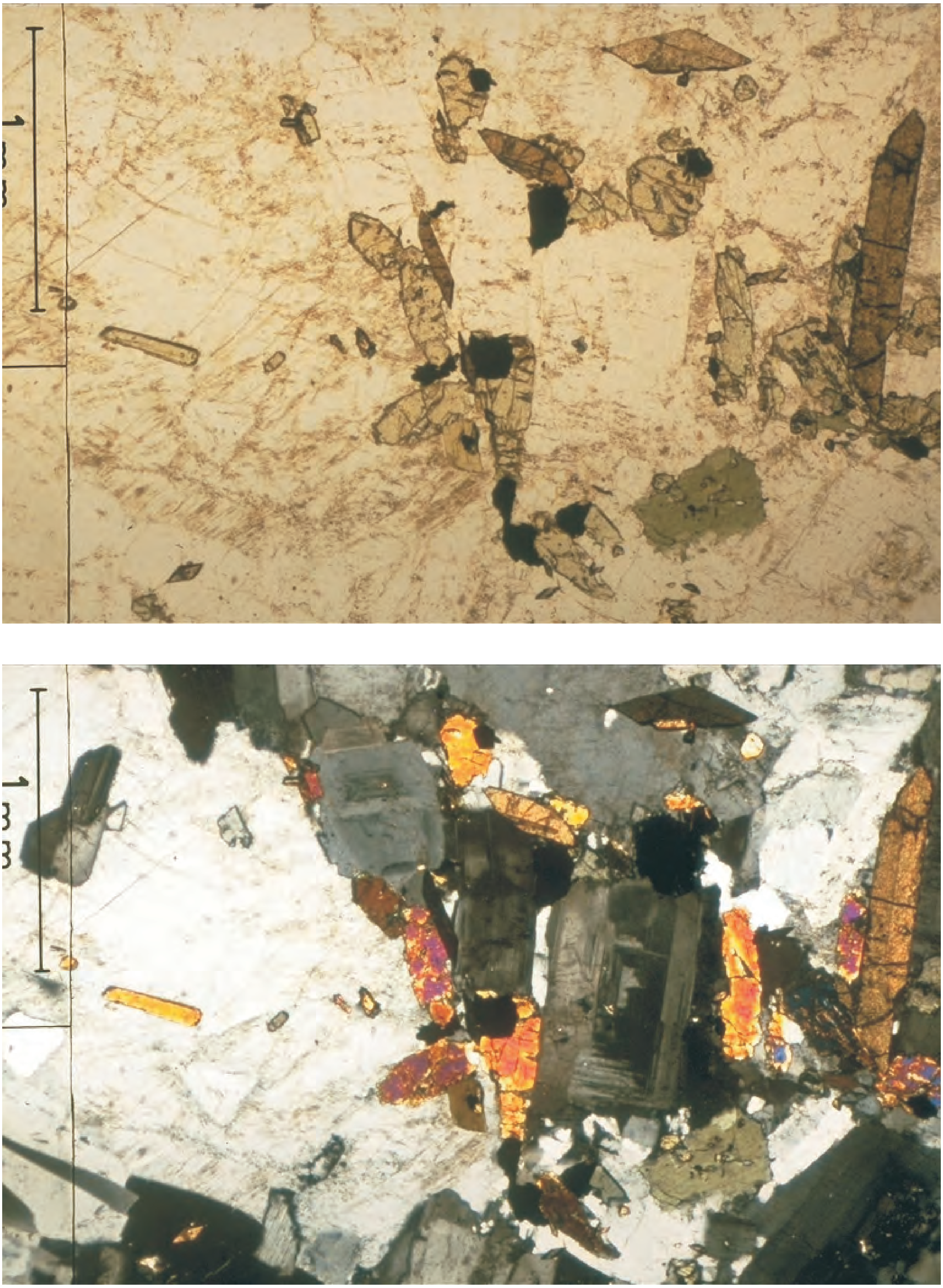


Figure 384. Photomicrograph of monzonite from Mount Sugarloaf, consisting of large K-feldspar laths poikilitically enclosing sodic plagioclase, augite, hornblende, sphene and opaques, with minor quartz. Plane polarised light and crossed nicols. Scale bar is 1mm long.

abundant mafic inclusions. Subhedral to euhedral plagioclase laths to 3mm long, aligned in some cases, are the major constituent (65–70%). Thin zoned rims as sodic as calcic oligoclase surround uniform cores of sodic labradorite, sometimes altered. K-feldspar (about 5%) occurs as small interstitial grains and as rims on plagioclase, and there is minor development of myrmekite. Mafic minerals range from 25 to 35% of the rock. Subhedral prisms of augite are more abundant than biotite. Minor green hornblende may or may not be present as rims on the pyroxene. Opaque grains are relatively abundant, comprising more than 5% of some rocks, and there is accessory apatite and euhedral sphene.

Monzogabbro occurs as dykes cutting gabbro in Tollbar Creek north of Mount Sugarloaf. The rock is finer grained than the gabbro, and pinkish in colour. It contains the same minerals as the gabbro, but in different proportions. Plagioclase (less than 55%) and mafics (about 20%) are less abundant, and K-feldspar (more than 25%) is greater.

Geochemistry

Analyses of the Ridler Monzonite carried out by Neale (1968) and during the present investigation form a coherent group which is distinguished from older Permo-Triassic granitoids, including those of the Glassford Igneous Complex, by the high values of Al_2O_3, light REE and particularly Sr, and low MgO (Figure 385). These geochemical features are shared by the adjacent Radley Nepheline Syenite and Judas Trachybasalt. One feature of the analyses of the Ridler Monzonite is the marked negative Nb anomaly (Figure 386). The $(Th/Nb)_N$ ratio is moderate to large, ranging from 29.2 to 65.6 and averaging 40.3. The Nb anomaly and the $(Th/Nb)_N$ ratios are similar to those of the Permo-Triassic granitoids, and are assumed to reflect common sources and processes, rather than active subduction at the Jurassic-Cretaceous boundary.

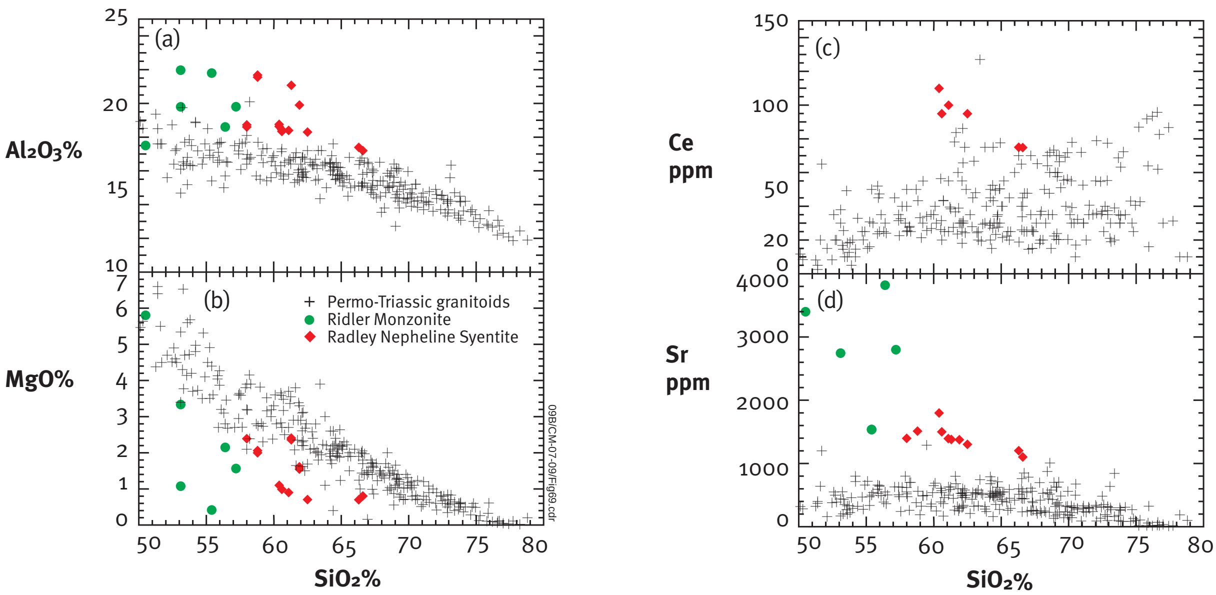


Figure 385. Harker diagrams comparing Al_2O_3, MgO, Ce and Sr values of the Ridler Monzonite and the Radley Nepheline Syenite with those of Permian to Triassic intrusives. Analyses from Neale (1968) and GSQ (unpublished).

Age

Green & Webb (1974) dated biotite from syenite at Mount Sugarloaf at 141±4Ma, which is earliest Cretaceous, but intrusion could have been of latest Jurassic age.

Stratigraphic relationships

The Ridler Monzonite intrudes the Carboniferous Rockhampton Group and the Triassic Monal Granodiorite and Littlemore Granodiorite. An arcuate mass of amphibolite along the eastern side of Mount Sugarloaf (mapped as Rockhampton Group) was apparently dragged up from depth during emplacement of the monzonite. Contact metamorphic rocks in the Rockhampton Group along the north-west side of the intrusion are locally rich in fluorite.

The monzonite is intruded by all the other Cretaceous units of the Glassford Igneous Complex: the Radley Nepheline Syenite, Burns Spur Nepheline Monzosyenite, Judas Trachybasalt, Tollbar Breccia, and unnamed

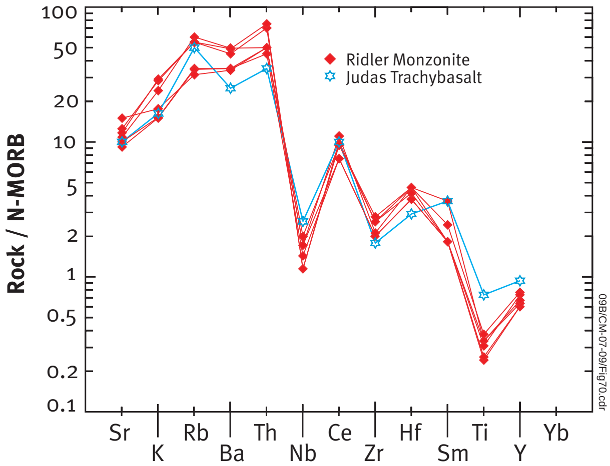


Figure 386. Spidergram plot of analyses from the Ridler Monzonite and Judas Trachybasalt. N-MORB values are from Pearce (1983).

granite and rhyolite. Neale (1968) recorded dykes of the Radley Nepheline Syenite intruding Ridler Monzonite on Mount Sugarloaf.

Economic significance

Anomalous copper levels in rocks and drainage samples in the area of the Ridler Monzonite have been of interest to several exploration companies, but no prospects have been identified within this unit.

Radley Nepheline Syenite (Kggr)

Introduction

Neale (1968) described an alkaline suite of rocks containing the sodalite mineral haüyne as the Tollbar phase of his Ridler Syenite, but did not recognise the widespread occurrence of nepheline. This unit is here defined as the Radley Nepheline Syenite.

Distribution

The Radley Nepheline Syenite forms a triangular-shaped mass about 1.5km^2 in area between Camp Creek in the north and Tollbar Creek in the south (Figure 387).

Derivation of name

The name is derived from the Parish of Radley, in which the unit occurs.

Type area

The type area is in the south-west corner of the unit along a ridge leading up to the Dawes Range to the west.

Topographic expression

The Radley Nepheline Syenite forms low hills with an average elevation of about 300m.

Geophysical expression

The unit occurs in the northern part of the prominent magnetic high associated with the Ridler Monzonite, even though its average magnetic susceptibility is only 700 x 10^{-5} SI units. This is consistent with the results of modelling, which show the presence of a non-magnetic body at the surface in the position of the Radley Nepheline Syenite (R. Deakin, *in* Rees, 1997).

The intrusion gives a strong radiometric response for all channels, but particularly for potassium and thorium.

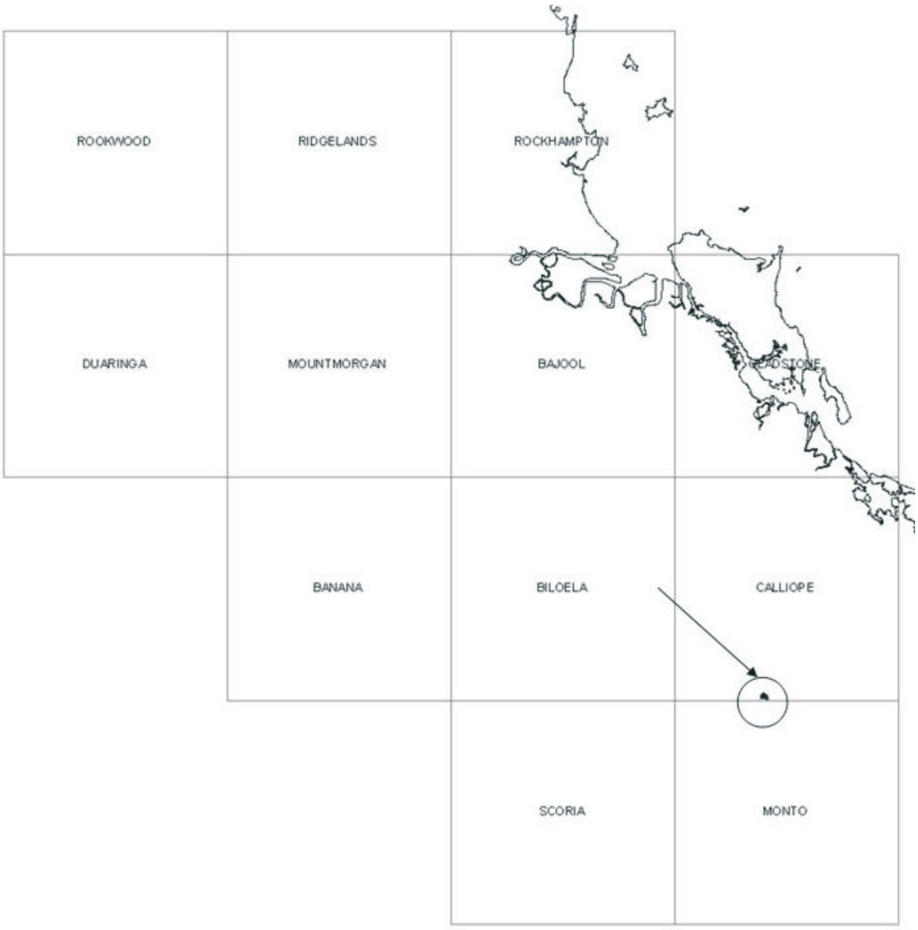


Figure 387. Distribution of the Radley Nepheline Syenite

Lithology and petrography

The Radley Nepheline Syenite is a multi-phase intrusion, with up to three generations of dykes, both coarser and finer grained than the host rock (Neale, 1968). The rocks consist mainly of K-feldspar, with or without subordinate plagioclase, and varying proportions of nepheline and a sodalite group mineral. Mafic minerals include aegirine, ferrohastingsite, melanite garnet and lepidomelane.

K-feldspar is always the dominant mineral in these rocks, ranging from 40 to 75%. It forms simply twinned subhedral laths up to 15mm long, locally aligned. All specimens contain nepheline and/or sodalite or their alteration products. Where fresh, nepheline comprises up to 35% of the rock as stumpy prisms to 2.5mm long (Figure 388). In some cases, nepheline (and possibly sodalite as well) has been completely altered to a fibrous zeolite accompanied by minor calcite and muscovite. Some altered grains enclosed in K-feldspar retain the original hexagonal outline of the nepheline crystals (Figure 389). Although natrolite is the zeolite which usually replaces nepheline (see, for example, Delvigne, 1998), this mineral has been identified by x-ray diffraction (Neale, 1968) and optical properties as thomsonite. Thomsonite has been recorded as replacements and pseudomorphs after nepheline (Deer & others, 1963). The sodalite mineral was determined as haüyne by x-ray diffraction by Neale (1968), who also reported the local occurrence of lazurite. Haüyne occurs as small equant grains comprising up to 25% of the rock.

Plagioclase is always greatly subordinate to K-feldspar. In general, its proportion decreases as the amount of feldspathoids increases, and it never constitutes more than about 20% of the rock. It forms zoned subhedral laths up to 1.5mm long, mainly calcic andesine in composition, although more calcic compositions were recorded by Neale (1968). Locally, twinned sodic plagioclase also forms thin reaction rims around K-feldspar grains at contacts with nepheline.

The content of mafic minerals ranges from less than 5% to about 30%. Any one of aegirine, ferrohastingsite or melanite may be dominant. Aegirine shows the characteristic green pleochroic scheme, and is rimmed and replaced by ferrohastingsite. Ferrohastingsite is pleochroic from deep brown to green to blue-green, with a very small 2V. It usually forms subhedral prisms up to 1.5mm long, many of which are zoned with darker rims

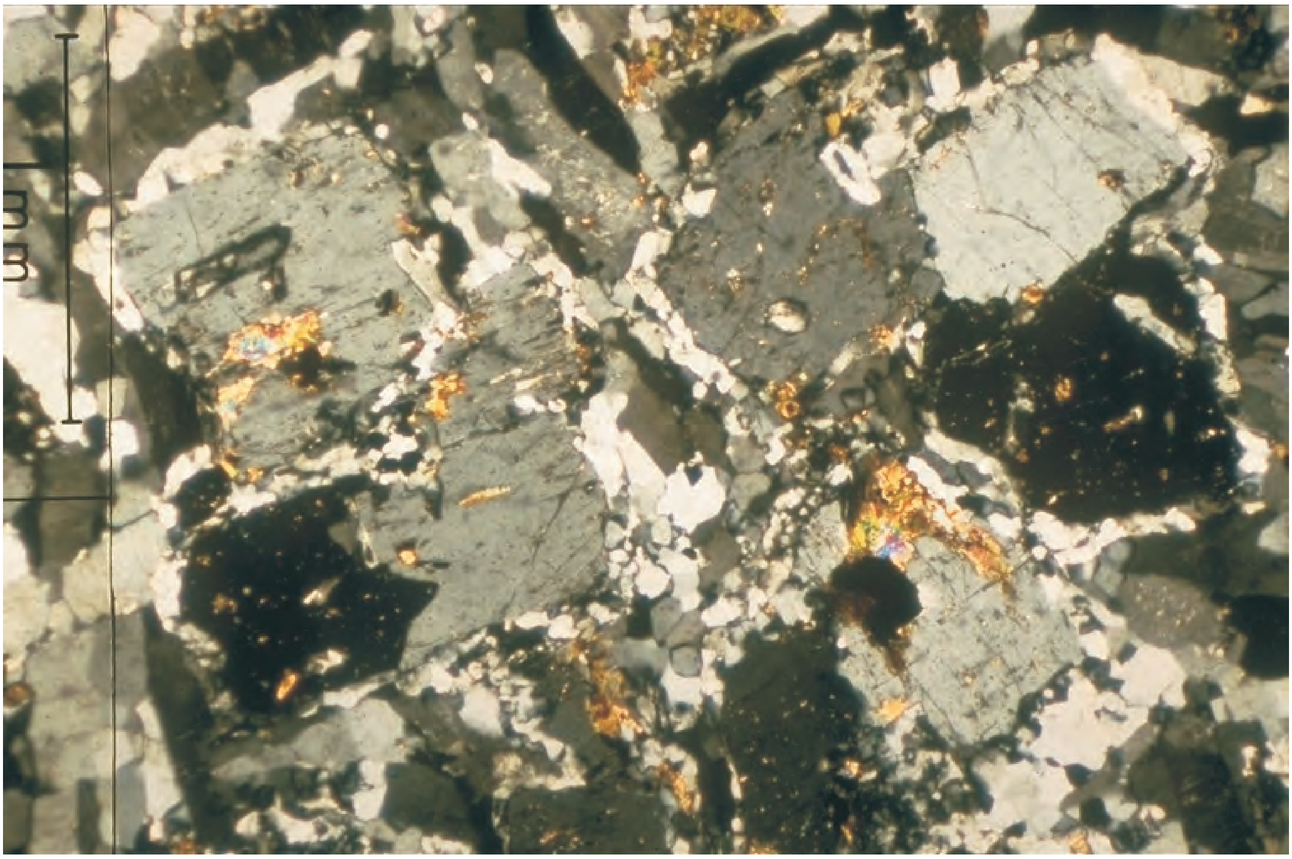


Figure 388. Photomicrograph showing stumpy nepheline prisms, isotropic sodalite, and small plagioclase grains, Radley Nepheline Syenite. Crossed nicols. Scale bar is 1mm long.

than cores, but also occurs as larger grains poikilitically enclosing altered nepheline. Melanite garnet comprises both small equant to subhedral zoned crystals and interstitial grains (Figure 390). Green lepidomelane, muscovite and opaques are minor constituents, and accessory sphene, apatite and zircon are present.

Geochemistry

No samples of the Radley Nepheline Syenite were analysed in the present study, but partial analyses were carried out by Neale (1968). These are similar to analyses from the Ridler Monzonite, but are even higher in Sr (Figure 385).

Age

The Ridler Monzonite, Radley Nepheline Syenite, Burns Spur Nepheline Monzosyenite and Judas Trachybasalt all appear to be related, and are assigned an earliest Cretaceous age, although they could range into the latest Jurassic.

Stratigraphic relationships

The Radley Nepheline Syenite intrudes the earliest Cretaceous Ridler Monzonite and the Triassic Littlemore Granodiorite.

Economic significance

No mineralisation is known to be associated with the Radley Nepheline Syenite.

Burns Spur Nepheline Monzosyenite (Kggb)

Introduction

A second small feldspathoid-bearing intrusion, the Burns Spur Nepheline Monzosyenite, occurs within the Glassford Igneous Complex about 6km south-west of the Radley Nepheline Syenite.

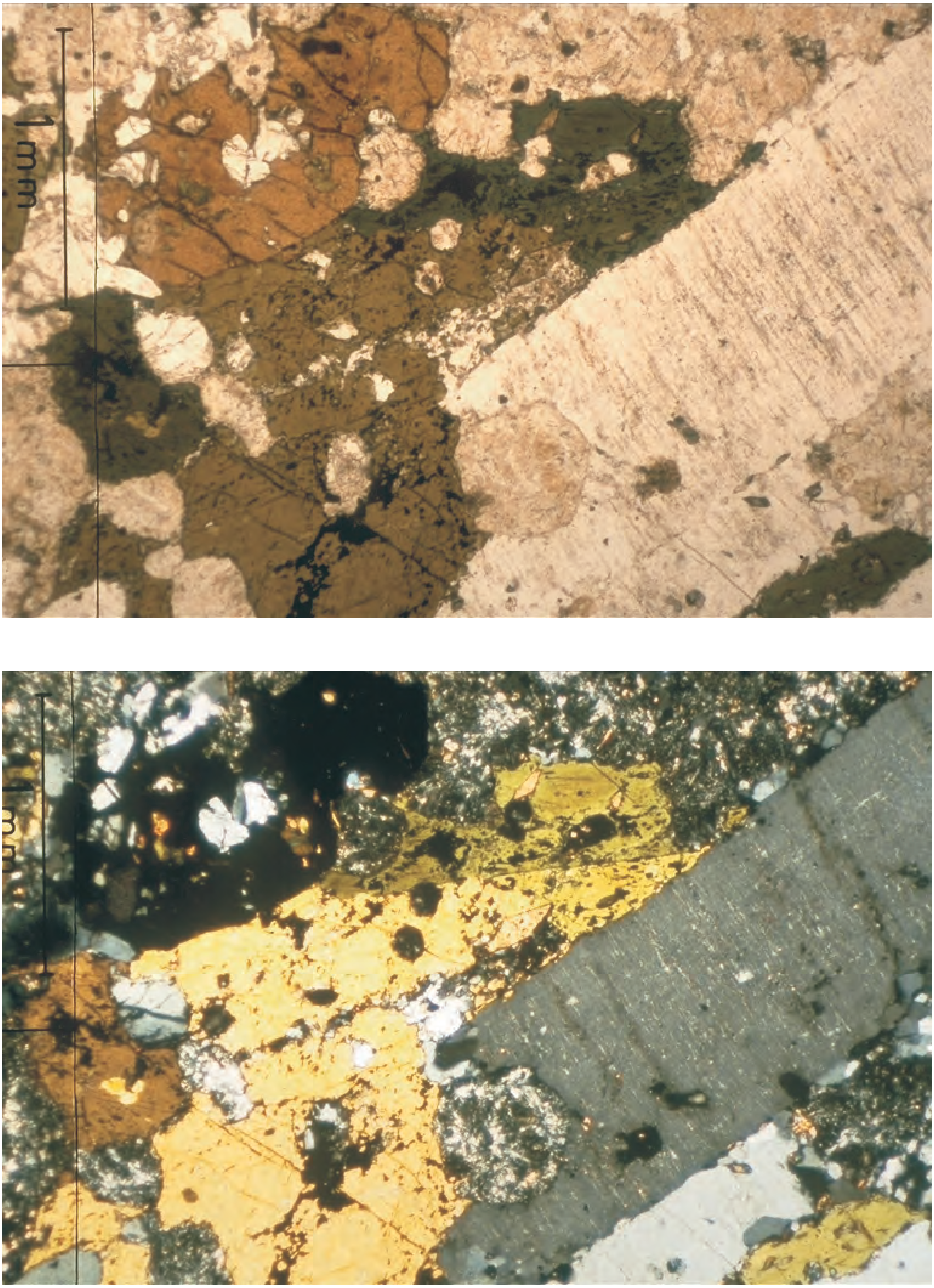


Figure 389. Photomicrograph showing large K-feldspar lath, ferrohastingsite, melanite garnet, altered ?nepheline and/or sodalite, and sphene, Radley Nepheline Syenite. Plane polarised light and crossed nicols. Scale bar is 1mm long.

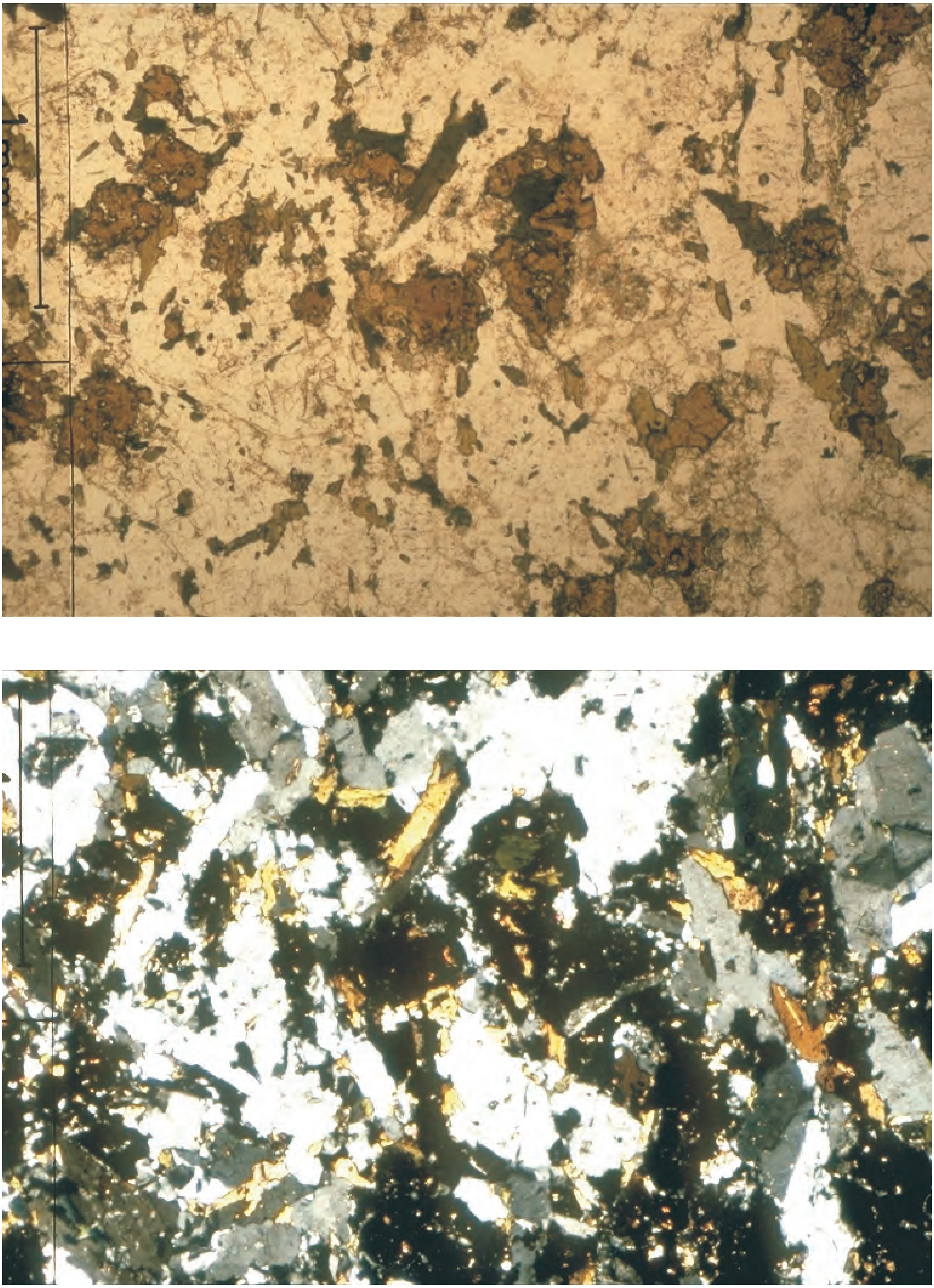


Figure 390. Photomicrograph showing K-feldspar, melanite garnet, ferrohastingsite, sodalite and plagioclase, Radley Nepheline Syenite. Plane polarised light and crossed nicols. Scale bar is 1mm long.

Figure 391. Distribution of the Burns Spur Nepheline Monzosyenite

Distribution

The Burns Spur Nepheline Monzosyenite forms a slightly elongate mass just over 2km long and 1km wide, trending east-north-east. It covers the area of the headwaters of Ridler Creek and its tributaries (Figure 391).

Derivation of name

The name is from the old Burns Spur mine, located in the eastern part of the intrusion (Figure 392).

Type area

The type area is along a stream, either Ridler Creek or a tributary, which flows north-east through the middle of the intrusion and divides it into approximately equal parts.

Topographic expression

The intrusion covers the steep, dissected slope of the Dawes Range, with elevations ranging from 300m in Ridler Creek to almost 600m on the crest of the range.

Geophysical expression

The Burns Spur Nepheline Monzosyenite has a moderate magnetic signature on images of airborne data, consistent with its observed magnetic susceptibility of 1230 x 10^{-5} SI units.

It gives a strong radiometric response in all channels. The thorium channel defines the extent of the unit particularly well.

Figure 392. Adit of the old Burns Spur copper mine

Lithology and petrography

The Burns Spur Nepheline Monzosyenite is a pink to fawn medium to coarse grained equigranular leucocratic nepheline monzosyenite with local concentrations of large finer grained grey inclusions. Sulphides, mainly pyrite, are relatively abundant along irregular joint planes.

Microperthitic K-feldspar, as poikilitic grains up to more than 8mm across, is the most abundant mineral (about 40% of rock). Plagioclase (30% of rock) forms subhedral laths up to 2.5mm long, mainly enclosed in K-feldspar. There is a sharp division between altered cores of calcic andesine and fresh rims of more sodic composition. Nepheline (about 10%) occurs as stumpy euhedral prisms and hexagonal sections pseudomorphed by a secondary mineral identified from its optical properties as thomsonite. Muscovite (5%) is present as small grains usually associated with plagioclase. Mafic minerals constitute little more than 5% of the rock, and consist of melanite garnet (interstitial grains ophitically enclosing plagioclase laths), lepidomelane (partly altered to chlorite), and rare ferrohastingsite (Figure 393). Minor opaques and euhedral sphene are present, and accessory apatite as relatively large grains, zircon, and possibly aenigmatite. Apart from possible thomsonite replacing nepheline, another very pale green, almost isotropic secondary mineral with anomalous brown interference colours is a significant constituent (almost 10% of rock). Its properties most closely resemble chlorite, but it does not appear to have replaced a mafic mineral.

Geochemistry

No samples of the Burns Spur Nepheline Monzosyenite have been analysed.

Age

By correlation with the Radley Nepheline Syenite, the unit is assigned an earliest Cretaceous age.

Stratigraphic relationships

The Burns Spur Nepheline Monzosyenite intrudes the Triassic Monal Granodiorite and the earliest Cretaceous Ridler Monzonite.

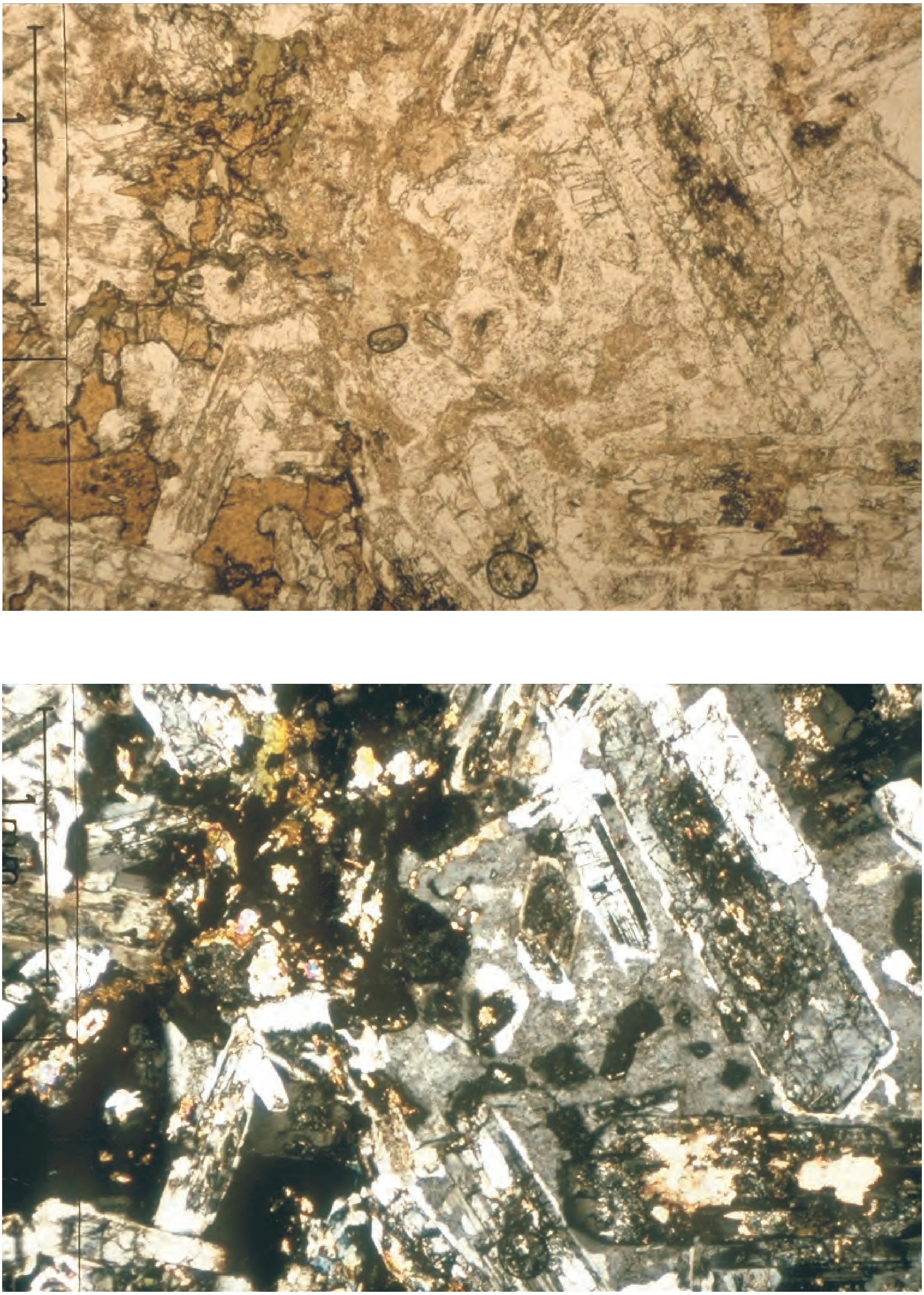


Figure 393. Photomicrograph showing K-feldspar poikilitically enclosing zoned and altered plagioclase laths and altered ?nepheline, with abundant melanite garnet, Burns Spur Nepheline Syenite. Plane polarised light and crossed nicols. Scale bar is 1mm long.

Economic significance

The old Burns Spur mine occurs in the intrusion, which is associated with elevated copper levels.

Tollbar Breccia

Introduction

The presence of breccia in the Glassford Igneous Complex was noted by Neale (1968), who mapped a Breccia phase in the area of Tollbar Creek

Distribution

The Tollbar Breccia occupies an area west of Mount Sugarloaf more than 2km long and averaging about ¾km wide, trending west-north-west between Ridler Creek to the east and Tollbar Creek near the western end (Figure 394).

Derivation of name

The name is derived from Tollbar Creek, which is the type area of the unit.

Type area

The type area is along Tollbar Creek. Outcrops along both sides of the creek are commonly covered with lichen, and the best exposures are in waterworn boulders in the bed of the creek itself, particularly for about 100m upstream from the track crossing.

Topographic expression

The Tollbar Breccia forms low hills with an elevation around 300m.

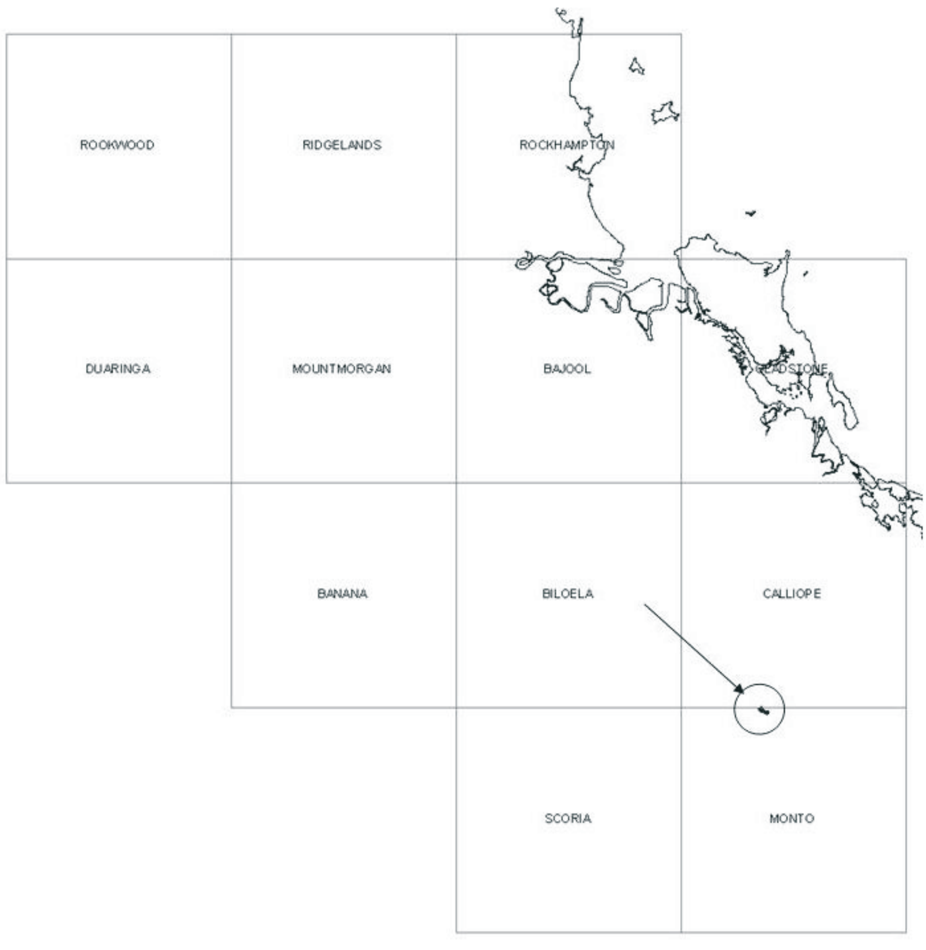


Figure 394. Distribution of the Tollbar Breccia

Geophysical expression

The unit corresponds with a magnetic low on images of airborne data, in keeping with its average magnetic susceptibility of 40 x 10^{-5} SI units.

It has a high radiometric response in the potassium and uranium channels, but a low response for thorium.

Lithology and petrography

The Tollbar Breccia is a coarse breccia with angular to rounded clasts mainly (but not exclusively) of altered granitic rocks in a limonitic matrix. A number of different types of breccia are present (Figures 395–398).

Age

The Tollbar Breccia is thought to be the same age as or slightly younger than the earliest Cretaceous intrusives of the Glassford Igneous Complex.

Stratigraphic relationships

The breccia is entirely surrounded by the Ridler Monzonite, and presumably is derived by brecciation of that unit.

Economic significance

A number of costeans along the banks of Tollbar Creek show that this rock type has been the target of exploration in the past.

Judas Trachybasalt (Kggj)

Introduction

Neale (1968) mapped trachybasalt north of Tollbar Creek, now defined as the Judas Trachybasalt.

Figure 395. Clast-supported breccia consisting of angular clasts of granitic and subordinate volcanic rocks in the Tollbar Breccia

Figure 396. Clast-supported breccia consisting of subangular clasts that are exclusively granitic in the Tollbar Breccia

Figure 397. Matrix-supported breccia consisting of angular to subangular clasts of volcanic rocks in the Tollbar Breccia

Figure 398. Breccia containing a significant proportion of elongate sedimentary clasts and granitic clasts with diffuse outlines in the Tollbar Breccia

Distribution

The Judas Trachybasalt extends over an area of about $1km^2$ north from Tollbar Creek to the summit of Judas Mountain (Figure 399). The unit was examined briefly during the Yarrol Project, but its distribution is taken from Neale (1968).

Derivation of name

The name comes from Judas Mountain, a locally named peak north of Tollbar Creek.

Type area

The type area is along Tollbar Creek, where good outcrops can be seen.

Topographic expression

The Judas Trachybasalt covers the steep country north from Tollbar Creek to Judas Mountain.

Geophysical expression

The unit corresponds with a magnetic high on images of airborne data, in keeping with its average magnetic susceptibility of 4600 x 10^{-5} SI units.

It has a high radiometric response similar to that of the Ridler Monzonite.

Lithology and petrography

Along Tollbar Creek, the Judas Trachybasalt is a grey, porphyritic trachybasalt (Figure 400). Phenocrysts include plagioclase, augite and hornblende to 1cm across.

In thin section, the phenocrysts can be seen to make up about one third of the rock, and are mostly euhedral, although some show resorption by the magma. Two types of plagioclase phenocrysts are present: clear, unaltered crystals with a reaction rim, and cloudy, altered grains lacking any reaction rim. Both types appear to be labradorite in composition. Hornblende is dark brown to dark green, and is zoned with a relatively

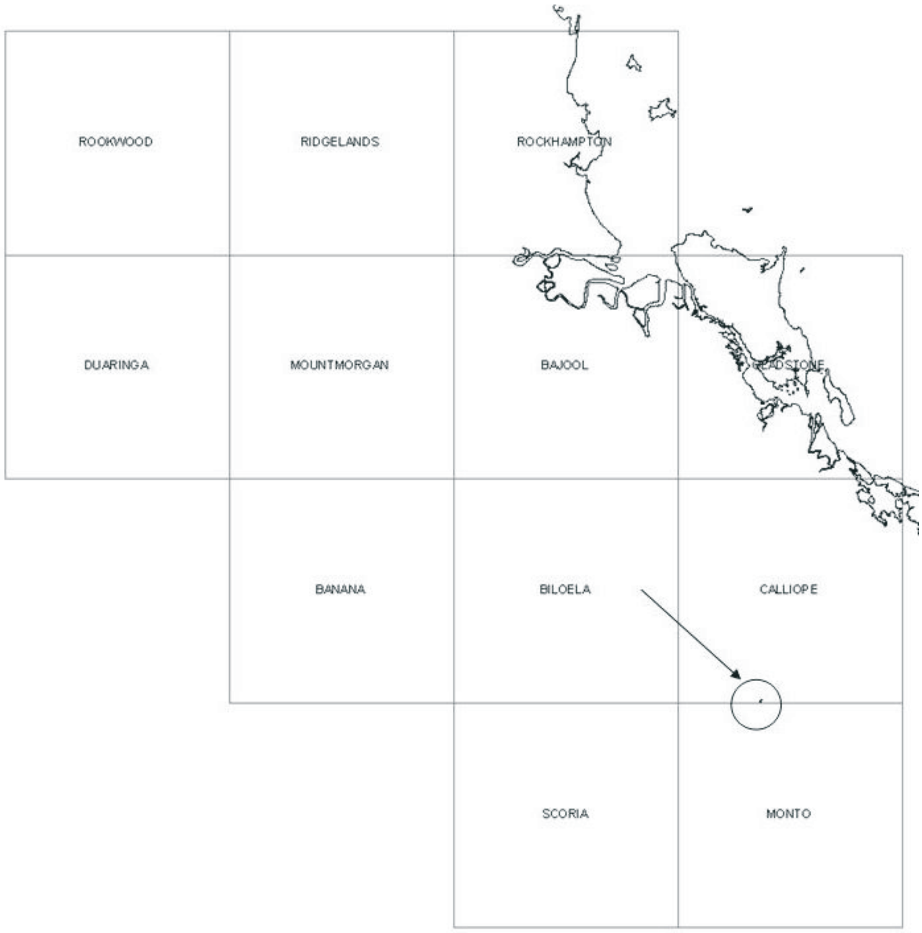


Figure 399. Distribution of the Judas Trachybasalt

narrow rim surrounding a uniform core. Augite is slightly zoned, and pleochroic. Minor minerals include biotite (some grains included in hornblende), apatite (included in all three major phenocryst phases), magnetite, and sphene (with a thin opaque rim). The groundmass is fine-grained, and consists of plagioclase laths, hornblende, and opaque grains (Figure 401). Secondary calcite is present.

Dykes porphyritic in K-feldspar, possibly related to the Radley Nepheline Syenite, cut the trachybasalt.

Geochemistry

A single sample of the Judas Trachybasalt from an outcrop along Tollbar Creek was analysed, and is a potassic trachybasalt. The analysis is similar to results from the Ridler Monzonite, including a high value of Sr and the presence of a significant negative Nb anomaly (Figure 386).

Age

The Judas Trachybasalt is thought to be related to the Ridler Monzonite, and is therefore assigned an earliest Cretaceous age.

Stratigraphic relationships

The nature of the contacts between the trachybasalt and the surrounding Ridler Monzonite is unknown.

Economic significance

The Judas Trachybasalt is not known to be associated with any mineralisation.

Figure 400. Trachybasalt with phenocrysts of augite and hornblende, Judas Trachybasalt

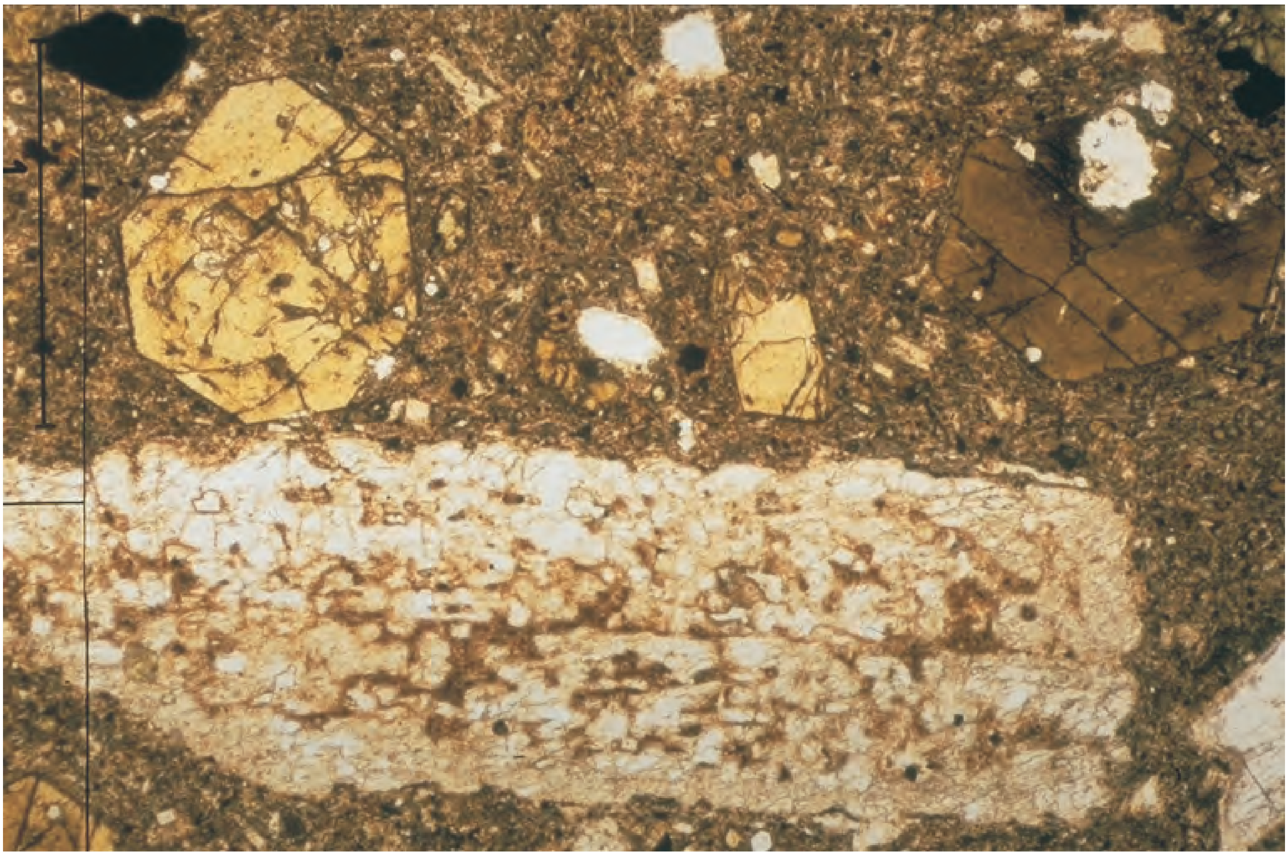


Figure 401. Photomicrograph showing phenocrysts of augite, hornblende and corroded plagioclase in a fine grained groundmass, Judas Trachybasalt. Plane polarised light. Scale bar is 1mm long.

Unnamed granite (Kgg$_g$)

Introduction

A small mass of leucocratic granite about 2.5km south of Mount Sugarloaf differs from all other Cretaceous intrusions within the Glassford Igneous Complex, having a high quartz content.

Distribution

The granite forms an east-west trending area 1.5km long and less than 1km wide on the north side of Ridler Creek (Figure 402).

Topographic expression

The intrusion is centred on a ridge reaching an elevation of about 350m.

Geophysical expression

The granite occurs in an area which gives a low to moderate magnetic signature on airborne data. However, its magnetic susceptibility is negligible (less than 10 x 10^{-5} SI units).

It gives a high radiometric response in all channels. The thorium channel is particularly useful in delineating the extent of the unit.

Lithology and petrography

The granite is leucocratic, ranging in colour from buff to cream with local orange patches. It is medium grained and equigranular, with voids which are either miarolitic cavities or produced by weathering and removal of mafic minerals. The rock consists essentially of K-feldspar (about 60%) and quartz (35%).

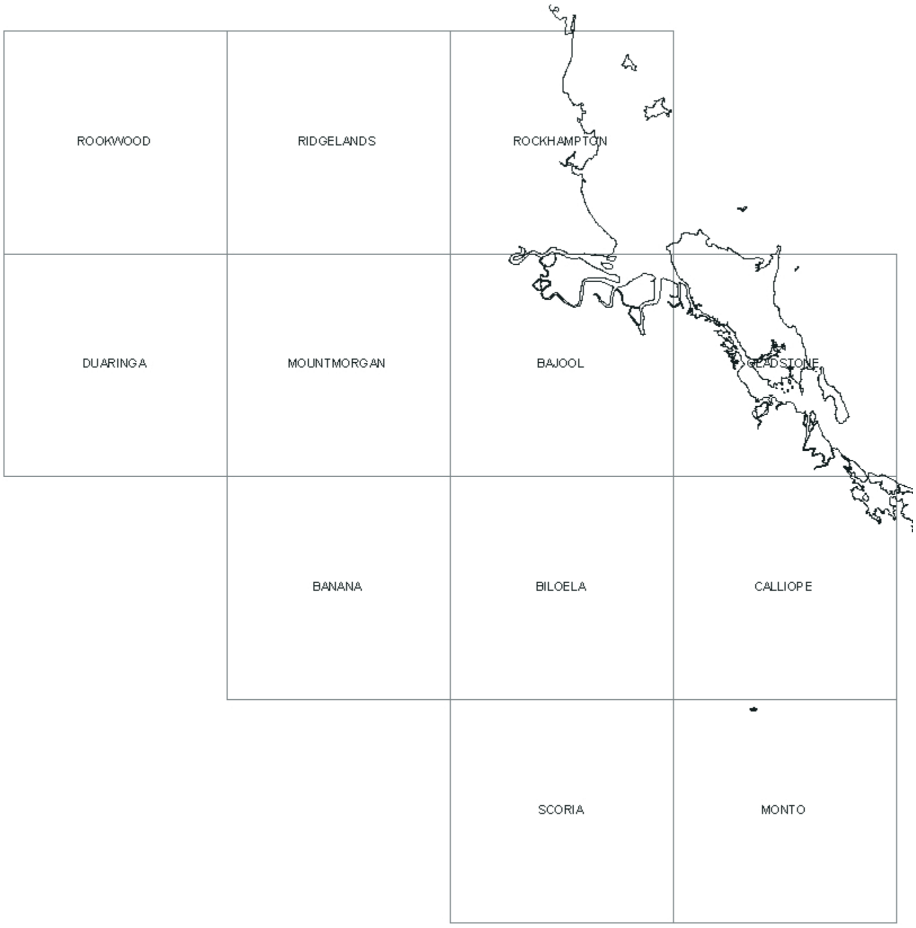


Figure 402. Distribution of the unnamed granite

Perthitic K-feldspar and quartz occur as anhedral grains up to 2mm and 0.5mm across, respectively. Both minerals exhibit an interlocking texture with sutured grain boundaries.

Plagioclase (less than 5% of rock) has two distinct modes of occurrence. Rectangular cores up to 1.5mm long in some of the larger K-feldspar crystals, now pseudomorphed by clay minerals and opaques, are considered to be altered subhedral plagioclase laths. In addition, fresh sodic plagioclase forms thin discontinuous rims on K-feldspar grains.

Minor biotite is present as small, slightly pleochroic crystals. Very small opaque grains along feldspar and quartz boundaries are probably secondary. Apatite is an accessory.

Geochemistry

No samples of the granite were analysed.

Age

A Cretaceous age has been assigned to the intrusion based on its relationships.

Stratigraphic relationships

The granite is almost entirely surrounded by, and therefore assumed to intrude, the Cretaceous Ridler Monzonite. It also intrudes the Triassic Monal Granodiorite.

Economic significance

No mineralisation is known to be associated with this intrusion.

Unnamed rhyolite (Kgg$_r$)

A very small body of rhyolite was mapped by an exploration company (Climie & others, 1970) between the Tollbar Breccia and the unnamed granite (**Kgg$_g$**) (Figure 403). It was not inspected during the mapping for the Yarrol project.

Goondicum Gabbro (Kgoo)

(C.G. Murray)

Introduction

The Goondicum Gabbro was named by Dear & others (1971). It was subsequently described as a layered gabbro by Walsh (1972), who recognised its main features. It was not inspected during the Yarrol mapping project, and this description is taken from Walsh (1972) and Groen (1993).

Distribution

The Goondicum Gabbro is an almost circular intrusion 6km across, centred about 30km east of Monto (Figure 404).

Derivation of name

The name is from Goondicum homestead located in the eastern part of the gabbro.

Type area

No type area has been nominated.

Topographic expression

The Goondicum Gabbro forms relatively low undulating country surrounded by a rim of hills representing the contact metamorphic aureole. A central hill is composed of Tertiary basalt. The feature is locally called the Goondicum crater, and is very prominent on aerial photographs.

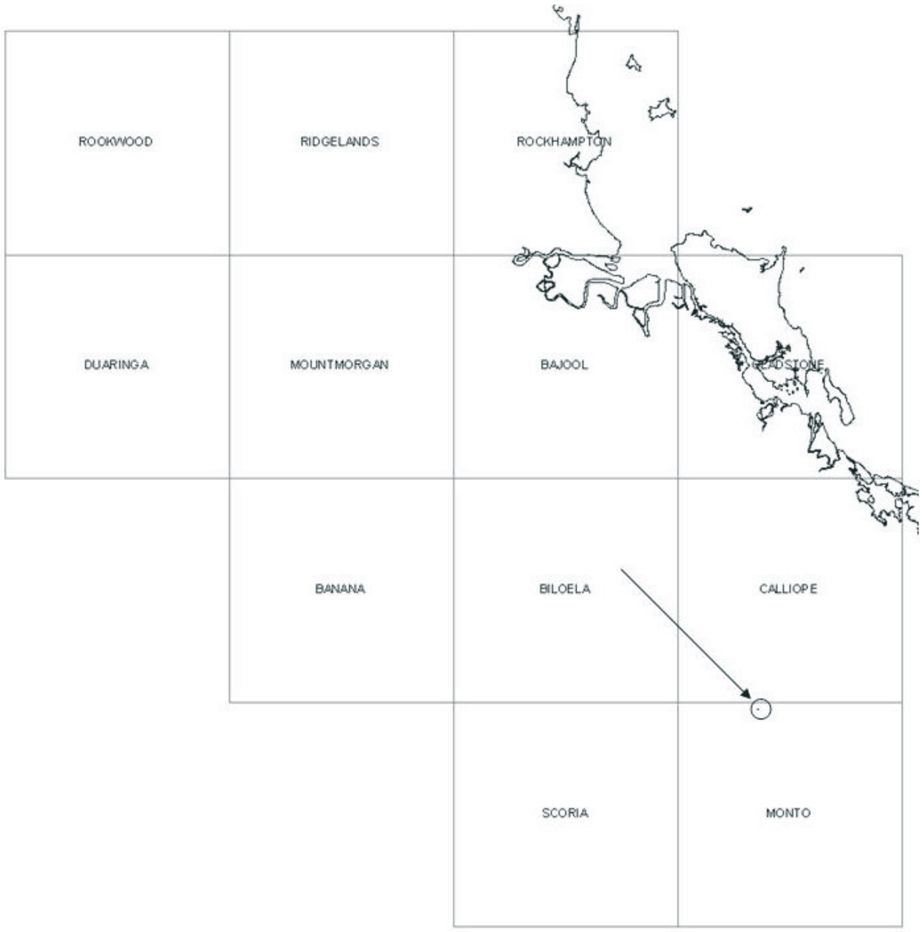


Figure 403. Distribution of the unnamed rhyolite (Kgg_r)

Geophysical expression

The gabbro has a very low radiometric expression in all channels, and an extremely strong magnetic signature. It also has a distinct gravity expression, coinciding precisely with a sharp positive anomaly along a belt of gravity highs that continues to the south (Lonsdale, 1965).

Lithology and petrography

The Goondicum Gabbro is a lopolithic intrusion with well developed layers dipping towards the centre. Both compositional layering and igneous lamination are present. Rock types include olivine gabbro, troctolite, leucogabbro, anorthosite, ferrigabbro, layers rich in magnetite and ilmenite, and marginal hornblende gabbro. The main cumulus minerals are plagioclase (labradorite), olivine, and titaniferous augite. Biotite is common, and locally abundant.

Syenite forms concentric dykes within the gabbro. It consists of K-feldspar, arfvedsonite or riebeckite, aegirine-augite, opaques, aenigmatite, and minor interstitial quartz. Some dykes are more monzonitic in composition, and contain plagioclase, Fe-rich clinopyroxene and olivine, and minor biotite (Groen, 1993).

Structure

Most of the intrusion consists of laminated gabbros that can be divided into 19 evenly spaced alternating leucogabbro and olivine gabbro layers (Groen, 1993). Walsh (1972) estimated that the dip towards the centre averaged about 35°, and ranged from 25° to 50°, whereas Groen (1993) suggested a dip of 20° to 30°. The total thickness represented in surface exposures is therefore about 1500m.

Walsh (1972) noted that magnetite-rich ferrigabbros form two elongate bodies conformable with the layering. In contrast, Groen (1993) considered that these two arcuate bodies dipped more steeply than the layering, and interpreted them as later intrusions into a partially solidified crystal mush.

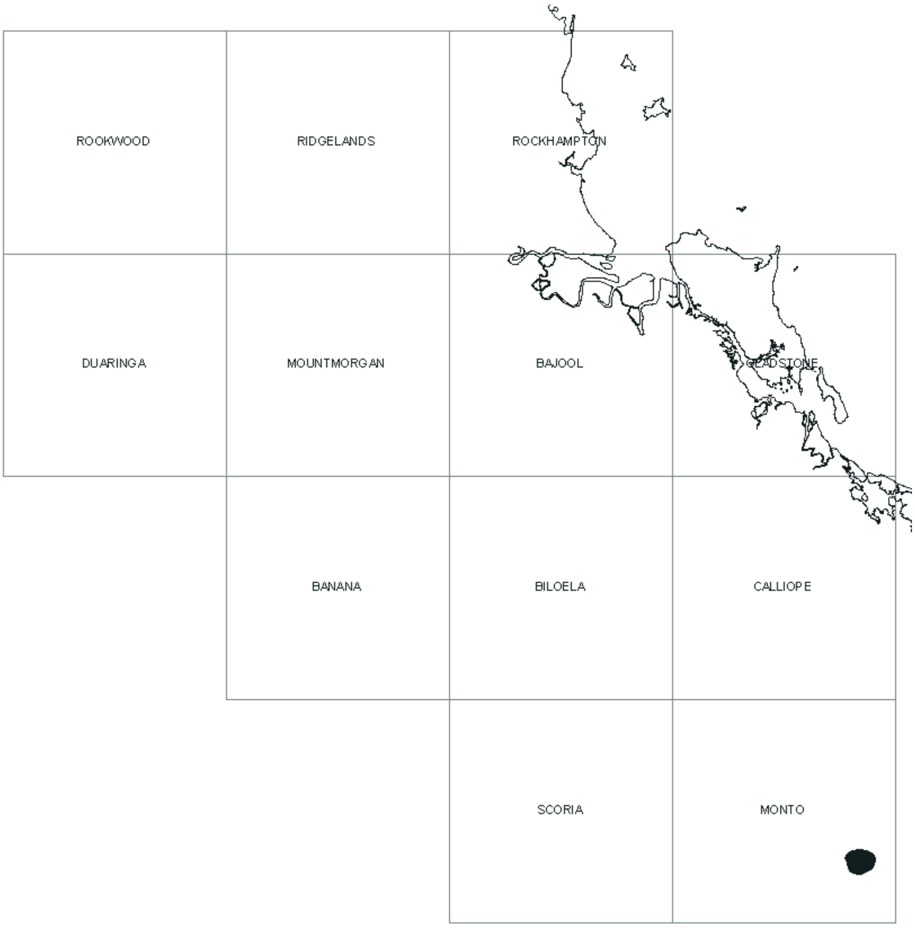


Figure 404. Distribution of the Goondicum Gabbro

Some of the syenite dykes are conformable with the layering, and some are steeper (Walsh, 1972). A pegmatitic syenite in the centre of the pluton dips outwards towards the margin at 40° (Groen, 1993).

Geochemistry

The gabbroic rocks are mildly alkaline, as suggested by the ubiquitous presence of titanaugite and the relative abundance of biotite, and the syenite is peralkaline (Groen, 1993). The restriction of the syenite dykes to the area covered by the Goondicum Gabbro suggests a genetic relationship.

Age

Hornblende from gabbro near the southern margin of the intrusion gave an age of 96Ma, or mid-Cretaceous (Groen, 1993). This date corresponds to an extended period of normal polarity (Gradstein & others, 1994), and is consistent with the strong magnetic signature suggesting normal remanence.

Relationships

The Goondicum Gabbro intrudes the Wandilla Formation and is intruded or overlain by Tertiary basalt.

Associated mineralisation

The intrusion has been explored unsuccessfully for platinum group elements. Some ilmenite and plagioclase feldspar was produced from alluvial and colluvial deposits derived from the gabbro (Johnson & Lee, 1996), but this operation has now ceased.

Mount Hedlow Trachyte (Klh)
(A.D.C. Robertson)

Introduction

Kirkegaard & others (1966, 1970) proposed the name Mount Hedlow Trachyte for a series of felsic plugs described by Jardine (1923). Willmott & others (1986) briefly described the unit, and Harding (1969), Sutherland & others (1988, 1996), and Robertson & Sutherland (1994) have discussed the age of the trachyte and its relationship to the Cretaceous–Cainozoic volcanism of eastern Queensland.

Distribution

Nineteen felsic plugs dominate this volcanic unit, which extends over a total area 30 by 15km between Rockhampton and Yeppoon (Figure 405). Individual plugs are small, and the largest, Mount Wheeler, is 1.5 by 1km. Along the coast south of Yeppoon they form headlands and islands at Double Head, Bluff Rock, Bluff Point, and Pinnacle Point (Figure 406).

Topographic expression

The plugs form prominent domes (Figure 406). They range from small rounded spires (Iron Pot) to large elongate masses (Mount Wheeler) and the triple-peaked Pine Mountain. The plugs generally are not jointed, and exfoliation has resulted in rounded, smooth masses. Within the Jim Crow Basin (elevation of about 40m), the plugs rise up to 200m in height. Plugs within the Cawarral serpentinite belt reach heights of 393m, while along the coast they do not exceed 100m.

Geophysical expression

The plugs have a strong radiometric signature. Most are white on images derived from airborne data, but some are red. Whether this difference relates to variation of composition is not known. Scree from the plugs has

Figure 405. Distribution of the Mount Hedlow Trachyte

Figure 406. Bluff Rock south of Yeppoon, with Great Keppel Island in the background, Mount Hedlow Trachyte

enlarged the geophysical response in some cases. The magnetic response is low, comparable with the surrounding sedimentary and metasedimentary rocks.

Basalt plugs which have been included in the Mount Hedlow Trachyte have a low radiometric and a high magnetic signature.

Lithology

The Mount Hedlow Trachyte is predominantly composed of rhyolite and trachyte, with rhyolite comprising 60% of the unit. Much of the rhyolite is fine-grained and grey in hand specimen. In thin section, the rhyolite contains phenocrysts of sanidine, anorthoclase, and quartz accompanied by minor amounts of glomeroporphyritic sodic pyroxene in a groundmass of fine feldspar laths and granular quartz and iron oxide. Groundmass feldspar invariable exhibits trachytic texture and anorthoclase phenocrysts show sodic overgrowths. Riebeckite may occur in the groundmass in the more sodic varieties and biotite with corroded margins can appear as a minor phenocryst phase in some rhyolites.

Trachyte is fine-grained, grey to off white in colour. Sanidine is the dominant phenocryst phase and may be accompanied by anorthoclase. These phenocrysts are usually set in a groundmass of fine feldspar laths, granular iron oxide, sodic pyroxene and minor quartz. In the more sodic varieties, small glomeroporphyritic riebeckite gives the rock a spotted appearance. The feldspar in the groundmass usually exhibits trachytic texture to varying degrees.

Two basalt plugs, denoted on the Rockhampton 1:100 000 Sheet area as $\mathbf{Kih_b}$, have been included in the Mount Hedlow Trachyte.

Geochemistry

The Mount Hedlow Trachyte has typical within plate compositions with high contents of rare earths, Nb, Zr, Hf and Y (Sutherland & others, 1996) (Figure 407).

Age

K-Ar data for the Mount Hedlow Trachyte are sparse. The plug at Jim Crow Mountain has given a K-Ar date of 71±1.4Ma (Harding, 1969). Zircon collected from trachyte at Black Mountain and from rhyolitic plugs at Cawarral, Jim Crow Mountain and Camp Hill East, were subjected to fission track analyses. High track densities in all zircons except one zircon from Camp Hill East prevented dating, as the zircon usually did not survive the track etching process (Sutherland & others, 1996). The lone fission track determination on the

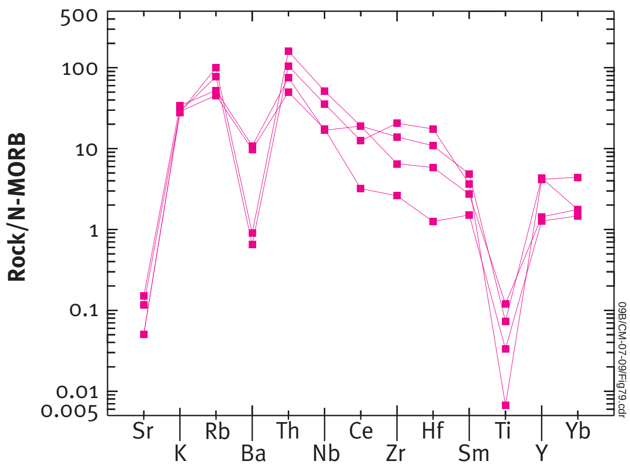


Figure 407. Spidergram plot of analyses from the Mount Hedlow Trachyte, showing its within plate character. Analyses from Sutherland & others (1996). N-MORB values from Pearce (1983).

zircon from Camp Hill East gave a result of 50.5±4.4Ma. U-Pb ages (SHRIMP method) derived from zircon from Black Mountain, Cawarral and Jim Crow Mountain ranged from 73 to 80±3Ma (see Table 1B in Sutherland & others, 1996). One zircon grain from Jim Crow Mountain which gave a U-Pb age of 242±5Ma is presumably a xenocryst from the underlying basement. The above results confirm the previously held view that the Mount Hedlow Trachyte is of Late Cretaceous age. The 50.5Ma age derived from Camp Hill East may indicate that eruption of the Mount Hedlow Trachyte could have extended into the Early Tertiary, or the age could be erroneous.

Structure

Sutherland & others (1996) suggested that the emplacement of plugs within the Jim Crow Basin was confined by suspected basin-bounding faults, but this hypothesis is unlikely, as many of the plugs fall outside the basin limits, and the basin is at least 40 million years older. The nature of any structure which controlled emplacement is not obvious.

Stratigraphic relationships

The Mount Hedlow Trachyte intrudes Upper Devonian to Lower Carboniferous accretionary wedge rocks of the Wandilla Formation and the Balnagowan Volcanic Member of the Doonside Formation. It also intrudes Lower Cretaceous sediments in the Jim Crow Basin (Esso Exploration and Production Australia Inc., 1981). The plugs are surrounded by flat-lying olivine basalt which has been assigned to the Upper Cretaceous Alton Downs Basalt, and the basalt and felsic rocks are assumed to form a bimodal assemblage. Both basalt and Mount Hedlow Trachyte are overlain by Quaternary alluvium.

Economic significance

No economic mineralisation has been found associated with the Mount Hedlow Trachyte.

UNDIFFERENTIATED INTRUSIONS

(C.G. MURRAY)

Small unnamed intrusions occur throughout the Yarrol Project area (Figure 408). They are described briefly in Table 6.

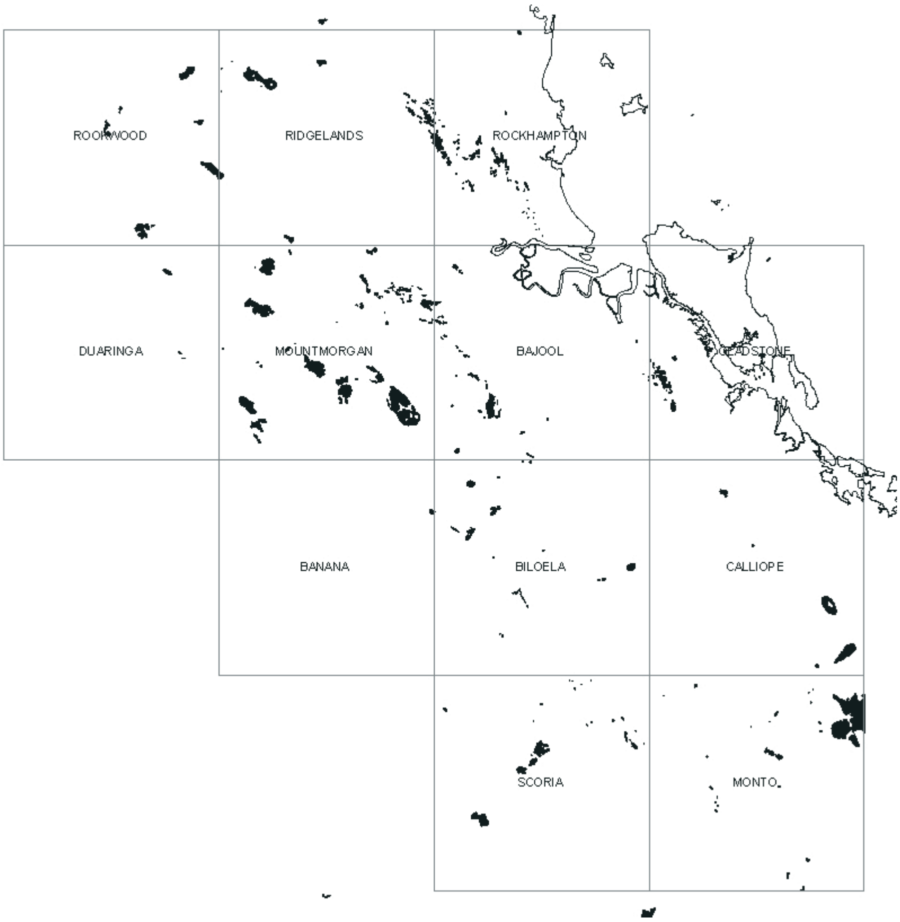


Figure 408. Distribution of unnamed intrusions

Table 6. Undifferentiated intrusions in the Yarrol Project area.

Symbols such as Pg, PRg, Rg and PRia refer to map unit symbols on the relevant 1:100 000 geological map sheets; those shown in magenta are based on geophysical interpretation.

Location and distribution	Geophysical expression	Lithology	Stratigraphic relationships	Age
ROOKWOOD 1:100 000 SHEET AREA				
PRg. North of Fitzroy River. Ovoid area 10 by 6km (continues onto Ridgelands sheet)	Low radiometric signature; moderate magnetic response similar to Wattlebank Granodiorite	?Granodiorite. The Broken Hill Proprietary Company Limited (1967) mapped decomposed granite west of the Fitzroy River and monzonite to the east	Intrudes Proterozoic to Palaeozoic Princhester Serpentinite	?Permian to Triassic
PRg. South of Fitzroy River. Elongate north-east trending mass 4 by 1.5km	Low radiometric signature; low to moderate magnetic response (not coincident with mapped outcrop)	?Granodiorite	Intrudes uppermost Silurian to Middle Devonian Craigilee beds	?Permian to Triassic
PRg. 8km east of Fitzroy River. Small north-west oriented mass, 0.5 by 0.2km	Too small to affect geophysical signature	?Granodiorite	Intrudes uppermost Silurian to Middle Devonian Craigilee beds	?Permian to Triassic
PRg. Bank of Fitzroy River. East-west oriented mass 2 by up to 1km	In area of moderate radiometric and magnetic response	?Granodiorite	Intrudes uppermost Silurian to Middle Devonian Craigilee beds	?Permian to Triassic

Table 6 (continued)

Location and distribution	Geophysical expression	Lithology	Stratigraphic relationships	Age
PRg. Rosewood Creek east of Round Mount. North-west oriented mass 6.5 by 1.5km (continues onto Ridgelands sheet)	At boundary between low and moderate radiometric signature; moderate magnetic response	Granodiorite, locally with abundant andesitic xenoliths	Intrudes uppermost Silurian to Middle Devonian Craigilee beds, Lower Carboniferous Rockhampton Group, Upper Carboniferous to lowermost Permian Lorray Formation, and uppermost Carboniferous to Lower Permian Youlambie Conglomerate	?Permian to Triassic
PRg_d. At Rookwood homestead. Almost circular mass 3.5km across	Moderate radiometric signature; strong magnetic response	?Diorite to quartz diorite	Intrudes mid-Permian Rookwood Volcanics and Back Creek Group; overlain by Tertiary sediments Ta	?Permian to Triassic
RIDGELANDS 1:100 000 SHEET AREA				
Pg_b. 2km west of Ducks Nest Hill. Roughly east-west intrusion 2km east-west and up to 1.5km north-south	In area of moderate radiometric signature; ?moderate magnetic response	Gabbro, diorite	Intrudes undifferentiated Lower Permian	?Permian
PRg. North of Fitzroy River. Ovoid area 10 by 6km (continues onto Rookwood sheet)	Low radiometric signature; moderate magnetic response similar to Wattlebank Granodiorite	?Granodiorite. The Broken Hill Proprietary Company Limited (1967) mapped decomposed granite west of the Fitzroy River and monzonite to the east	Intrudes Proterozoic to Palaeozoic Princhester Serpentinite	?Permian to Triassic
Pg_b. Between Marble Ridges homestead and Fitzroy River. West-north-west oriented intrusion 8 by up to 2.5km	Low radiometric signature; low magnetic response	Gabbro, diorite	Intrudes Proterozoic to Palaeozoic Princhester Serpentinite and uppermost Silurian to Middle Devonian Craigilee beds	?Permian
PRg_b. Magnetic anomaly 8km north of Ridgelands	Very strong magnetic response	East (1945) described spoil from old wells as diorite	Intrudes Upper Devonian to lowermost Carboniferous Mount Alma Formation	?Permian to Triassic
PRg_b. Belt extending north-west from Rockhampton. Belt is 40km long and up to 4km wide; largest single intrusion is 5 by 2km (continues onto Rockhampton sheet) (Crouch & Parfrey, 1998)	Individual intrusions have low radiometric signature; moderate magnetic response. Many intrusions appear to be sills (Willmott & others, 1986).	Gabbro, dolerite, diorite	Intrude Lower Carboniferous Rockhampton Group and Lower Permian Berserker Group	?Permian to Triassic
PRg. 25km north-west of Kalapa North. North-west oriented mass 6.5 by 1.5km (continues onto Rookwood sheet)	At boundary between low and moderate radiometric signature; moderate magnetic response	?Granodiorite	Intrudes uppermost Silurian to Middle Devonian Craigilee beds, Lower Carboniferous Rockhampton Group, Upper Carboniferous to lowermost Permian Lorray Formation, and uppermost Carboniferous to Lower Permian Youlambie Conglomerate	?Permian to Triassic

Table 6 (continued)

Location and distribution	Geophysical expression	Lithology	Stratigraphic relationships	Age
PRg. North-west trending group of intrusions 20km south-west of Ridgelands; largest intrusion is arcuate and 3.5 by up to 2km	Strong magnetic response	Geophysical response is typical of gabbro or diorite	Within area of Mount Salmon Volcanics	?Permian to Triassic
Rg_g. 2km west of Kalapa North. West-north-west oriented intrusion 2.5 by up to 1km	In area of moderate radiometric and magnetic response	Granite, granodiorite	Intrudes Triassic Native Cat Andesite	?Triassic
ROCKHAMPTON 1:100 000 SHEET AREA				
Rg. Along Cabby Creek, north of Yeppoon. Intrusion 1 by 0.5km.	Not in detailed coverage; company aeromagnetic survey indicates that granodiorite is larger at depth	Two phases of granodiorite, the earlier porphyritic and the later (which is the source of mineralisation) equigranular, with abundant secondary biotite. Associated mineralisation is mainly Cu-Mo, concentrated in breccias around the margin of the intrusion (Wasp prospect)	Intrudes Lower Carboniferous Wandilla Formation and unnamed Permian–Triassic volcanics	Triassic. Reported isotopic dates on biotite are 218±5 and 229±5Ma (Late Triassic)
PRg. Just east of Rockhampton–Yeppoon road, 4.5km north-east of Jim Crow Mountain. Small intrusion 0.6 by 0.4km	At edge of detailed coverage	?Granodiorite	Intrudes unnamed Proterozoic to Palaeozoic serpentinite	?Permian–Triassic. Zircon from the trachyte of Jim Crow Mountain dated at 242±5Ma (Sutherland & others, 1996) may be from this intrusion
PRg_g. Along Rockhampton–Yeppoon road, 5km south-west of Jim Crow Mountain. North-north-east oriented intrusion 2 by up to 0.8km	Low radiometric response, indicative of deep weathering; low magnetic response	?Granodiorite	Intrudes Chalmers Formation of Lower Permian Berserker Group	?Permian to Triassic
PRg_g. Just west of Yarrol Fault 2km south of Tungamull–Keppel Sands road. North-north-west oriented intrusion 2 by up to 0.4km	Low radiometric response; rather diffuse low to moderate magnetic signature	?Granodiorite	Intrudes Chalmers Formation of Permian Berserker Group; overlain by Tertiary to Quaternary slopewash (**TQcc**)	?Permian to Triassic
PRg_b and **PRg_d**. Belt extending north-west from Rockhampton. Belt is 40km long and up to 4km wide; largest single intrusion is 5 by 2km (continues onto Ridgelands sheet) (Crouch & Parfrey, 1998)	Individual intrusions have low radiometric signature; moderate magnetic response. Many intrusions appear to be sills (Willmott & others, 1986).	Gabbro, dolerite, diorite	Intrude Permian Berserker Group	?Permian to Triassic
PRia. Small intrusions near Mount Chalmers (Crouch & Parfrey, 1998)	Generally low radiometric and magnetic response	Andesite to basalt	Intrude Permian Berserker Group	?Permian to Triassic
PRir. Small intrusions near Mount Chalmers (Crouch & Parfrey, 1998)	Intrusions too small to give distinctive geophysical response	Porphyritic rhyolite	Intrude Permian Berserker Group and **PRia**	?Permian to Triassic

Table 6 (continued)

Location and distribution	Geophysical expression	Lithology	Stratigraphic relationships	Age
DUARINGA 1:100 000 SHEET AREA				
Pg_b. North of Fitzroy River 3km west-south-west of Fitzroy Pocket.	Low radiometric signature; high magnetic response (similar to surrounding Rookwood Volcanics)	Gabbro, diorite	Intrudes Lower Permian Rookwood Volcanics	?Permian
PRg_d. South of Capricorn Highway 2km south-west of Herbert Creek homestead. Larger intrusion is only 1 by 0.3km	Very low radiometric response; moderate magnetic signature	Diorite, quartz diorite, commonly altered	Intrude Permian Back Creek Group of Gogango Overfolded Zone	?Permian to Triassic
MOUNT MORGAN 1:100 000 SHEET AREA				
Rg_d. Small intrusion 10km north-west of Westwood	Intrusion too small to give distinctive geophysical signature	Diorite	Intrudes Permian Rookwood Volcanics	?Triassic
PRg_b. Headwaters of Fred Creek, 8km north-west of Westwood. Intrusion 3.5 by up to 1.5km	Low radiometric signature; strong magnetic response	Gabbro	Intrudes Upper Permian Moah Creek beds and ?**PRg_d**; intruded by **PRg_g**; overlain by Lower Cretaceous Dalma Basalt	?Permian to Triassic
PRg_d. Along Fred Creek, 9km north-north-west of Westwood. Two intrusions 2 by 1km and 3 by up to 1km	Low to moderate radiometric signature; moderate magnetic response	Diorite	Intrudes mid-Permian Rookwood Volcanics and Upper Permian Moah Creek beds; intruded by **PRg_g** and ?**PRg_b**; overlain by Lower Cretaceous Dalma Basalt	?Permian to Triassic
PRg_g. 10km north-north-west of Westwood. Arcuate intrusion 1.5 by 1.5km	Moderate radiometric response; low magnetic signature	?Granodiorite	Intrudes **PRg_b** and **PRg_d**; overlain by Lower Cretaceous Dalma Basalt	?Permian to Triassic
Rg_g. Near Neerkol Creek 10km west of Kalapa. Elongate intrusion 1 by 0.2km	Moderate radiometric response; low magnetic signature (including on the first vertical derivative)	?Granodiorite	Intrudes Triassic Native Cat Andesite; overlain by Lower Cretaceous Dalma Basalt	?Triassic
PRg_g. 4km west of Table Mountain. Two small masses about 1km across	Moderate radiometric response; in an area giving a moderate magnetic signature	?Granodiorite	Intrude Middle Devonian Raspberry Creek Formation and Upper Carboniferous-lowermost Permian Lorray Formation	?Permian to Triassic
PRg_b. Along Capricorn Highway 6km west-south-west of Westwood. East-west oriented intrusion 7 by up to 3.5km	Low radiometric response; magnetic signature indicates that this is a composite intrusion, ranging from strong to weak (?reversed)	Gabbro and granite. Granite in the southern part of this intrusion has the same geochemistry as the felsic phases of the Rookwood Volcanics and the Kyle Mohr Igneous Complex, and must be co-magmatic	Intrudes mid-Permian Rookwood Volcanics	?Permian to Triassic
Kib. 12km west-south-west of Mount Morgan. Small basalt plugs in the Kyle Mohr Igneous Complex	Too small to give distinctive geophysical signatures	Basalt with numerous mantle or lower crustal inclusions	Intrude mafic phase of the mid-Permian Kyle Mohr Igneous Complex and the uppermost Carboniferous–Lower Permian Youlambie Conglomerate	Lower Cretaceous

Table 6 (continued)

Location and distribution	Geophysical expression	Lithology	Stratigraphic relationships	Age
PRg_d. Headwaters of Turner Creek 6km west of Mount Morgan. Intrusion with maximum dimension about 1km	Low radiometric response; moderate magnetic signature, particularly on the first vertical derivative	Diorite	Intrudes Upper Devonian to Lower Carboniferous Balaclava Formation; faulted against uppermost Carboniferous to Lower Permian Youlambie Conglomerate; overlain by Lower Jurassic Precipice Sandstone	?Permian to Triassic
PRg_d. Centre Creek 10km north-west of Mount Morgan. North-north-west trending elongate intrusion 5 by less than 1km	Very low radiometric response; in an area giving a moderate magnetic signature	Dark grey microdiorite	Intrudes Middle Devonian Raspberry Creek Formation, Upper Devonian to Lower Carboniferous Balaclava Formation, and Lower Carboniferous Rockhampton Group; overlain by Lower Jurassic Precipice Sandstone	?Permian to Triassic
Mg_b. 5km north of Mount Morgan.Elongate intrusion trending east-west, more than 1km long and 0.3km wide	Geophysical response cannot be distinguished from that of the surrounding Mount Morgan Trondhjemite	Gabbro	Intrudes Upper Devonian Mount Morgan Trondhjemite	?Mesozoic. Geochemistry indicates that it is not part of the Mount Morgan Trondhjemite (Messenger, 1996)
Dir. Small intrusions in the Mount Morgan area.	Geophysical response cannot be distinguished from surrounding units	Dacite to rhyolite	Intrudes Middle Devonian Capella Creek Group; relations with Upper Devonian Mount Morgan Trondhjemite and Hoopbound Formation uncertain; overlain by Lower Jurassic Precipice Sandstone	Upper Devonian. Comagmatic with the Mount Morgan Trondhjemite
PRg_d. At Struck Oil prospect 5km east-north-east of Mount Morgan. Very small intrusion	Too small to give a distinctive geophysical signature	Diorite	Intrudes Middle Devonian Raspberry Creek Formation	?Permian to Triassic
PRg_g. At Struck Oil prospect 5km east-north-east of Mount Morgan. Intrusion less than 0.5km across	Prominent moderate radiometric signature; low magnetic response surrounded by a moderately magnetic hornfelsed rim	Biotite granodiorite with abundant secondary muscovite. Associated with porphyry-type mineralisation (Djaswadi, 1979)	Intrudes Middle Devonian Raspberry Creek Formation	?Permian to Triassic
PRg. Station Creek 10km east of Mount Morgan. Intrusion about 4.5 by 2km (continues onto Bajool sheet)	Low radiometric signature; moderate to low magnetic response, zoned from centre (low) to rim (moderate)	?Granodiorite	Intrudes Middle Devonian Capella Creek Group	?Permian to Triassic
Kib. Near Bottle Tree Creek, about 19km south-south-west of Westwood. Basalt plugs up to 0.5km in diameter	Low radiometric response; very low (reversed) magnetic signature	Nephelinitic basalt with inclusions including garnet	Overlain by Tertiary Biloela Formation and Tertiary to Quaternary alluvium (**Qpa** and **Qa**)	?Lower Cretaceous
PRg_d. Small intrusion 4km north-west of The Black Range, 15km north-west of Dululu	Low radiometric and magnetic response	?Diorite	Intrudes uppermost Carboniferous to Lower Permian Youlambie Conglomerate and Permian Rookwood Volcanics; overlain by Quaternary alluvium (**Qa**)	?Permian to Triassic

Table 6 (continued)

Location and distribution	Geophysical expression	Lithology	Stratigraphic relationships	Age
PRg$_d$. Intrusion 8 km north-west of Dululu. North-west oriented intrusion 7 by up to 3km	Low to moderate radiometric signature; moderate to strong magnetic response	Diorite	Intrudes Upper Devonian to Lower Carboniferous Balaclava Formation; probably intruded by **PRg$_g$**; unconformably overlain by Tertiary Biloela Formation	?Permian to Triassic
PRg$_g$. Intrusion 7 km north-west of Dululu. Equant intrusion about 1.5km across	Low to moderate radiometric signature; low magnetic response	?Granodiorite	Probably intrudes **PRg$_d$**	?Permian to Triassic
PRg$_d$. Intrusion centred 3km east-north-east of Dululu. North-south oriented intrusion 6 by up to 3km	Low to moderate radiometric signature; moderate to strong magnetic response	Diorite	Intrudes Upper Devonian Hoopbound Formation and Upper Devonian to Lower Carboniferous Balaclava Formation; overlain by Quaternary alluvium (**Qa**)	?Permian to Triassic
PRg$_g$. Very small intrusion on Burnett Highway at Wura	Too small to give a distinctive geophysical signature	?Granodiorite	Intrudes Upper Devonian Hoopbound Formation; overlain by Quaternary alluvium (**Qa**)	?Permian to Triassic
PRg$_d$. Intrusion east of Burnett Highway 11km north-east of Dululu. Intrusion 4 by 3km	Low to moderate radiometric signature; low to moderate magnetic response with strongly magnetic metamorphic rim	Diorite	Intrudes Upper Devonian Hoopbound Formation	?Permian to Triassic
PRg$_g$. Intrusion straddling the Burnett Highway 11km north-east of Dululu. Intrusion about 0.8 by 0.4 km	Moderate radiometric signature; low magnetic response	?Granodiorite	Surrounded by Quaternary alluvium (**Qa**); probably intrudes Upper Devonian Hoopbound Formation	?Permian to Triassic
Pg$_b$. Series of relatively large intrusions in north-north-west trending belt about 10km west of Wowan. Largest intrusion is 5.5 by 2km	Low radiometric and magnetic signature	?Gabbro to diorite	Intrude uppermost Carboniferous to Lower Permian Camboon Volcanics	?Permian
PRg$_d$. Several magnetic highs covering a relatively large area east of Wowan	High magnetic response	?Diorite	Beneath Tertiary Biloela Basin	?Permian to Triassic
Kib. Small basalt plug in unnamed diorite at Eulogie Park	Geophysical signature cannot be distinguished from that of the unnamed diorite	Basalt	Intrudes unnamed diorite **Rg$_d$**	Upper Cretaceous. K-Ar date on whole rock is 68.6±0.5Ma (Sutherland & others, 1996)
Rg$_d$. Diorite surrounding the Eulogie Park Gabbro. North-west trending intrusion 11 by up to 4.5km	Low radiometric signature; moderate to strong magnetic response	Diorite	Intrudes Upper Devonian Hoopbound Formation and Upper Devonian to Lower Carboniferous Balaclava Formation; faulted against uppermost Carboniferous to Lower Permian Youlambie Conglomerate; faulted against and intrudes Eulogie Park Gabbro	Triassic. K-Ar date on plagioclase is 241.6±3Ma (Holcombe & others, 1997b)

Table 6 (continued)

Location and distribution	Geophysical expression	Lithology	Stratigraphic relationships	Age
BAJOOL 1:100 000 SHEET AREA				
PRg_b. West of Bruce Highway 4km north-west of Archer siding. North-west trending intrusion 3 by up to 1km	Low radiometric signature; strong magnetic response	?Gabbro (area of no outcrop); skarns drilled around contact	Intrudes Lower Carboniferous Rockhampton Group.	?Permian to Triassic
PRg. Station Creek 15km west of Bajool. Intrusion about 4.5 by 2km (continues onto Mount Morgan sheet)	Low radiometric signature; moderate to low magnetic response, zoned from centre (low) to rim (moderate)	?Granodiorite	Intrudes Middle Devonian Capella Creek Group	?Permian to Triassic
$PRg_b{}^2$?. Two magnetic anomalies 8km south of Bajool and 12km south-east of Bajool, about 1km across	Low radiometric signature; very low (reversed) magnetic response	?Gabbro, diorite	The anomaly south of Bajool is within the Bajool Quartz Diorite; that to the east appears to intrude hornfelsed country rocks	?Permian to Triassic.
PRgb. Small intrusion 15km east of Raglan	In area of moderate radiometric signature; prominent magnetic high in area of low magnetic response	?Gabbro	Intrudes Lower Carboniferous Rockhampton Group	?Permian-Triassic
$PRg_b{}^1$. North-west trending belt of magnetic anomalies east of Bruce Highway, extending onto Gladstone sheet. Largest anomaly 3 by up to 0.5km	In area of moderate to strong radiometric signature; moderate magnetic response	?Gabbro, diorite	Intrude mixed Lower Carboniferous Rockhampton Group and Early Permian Berserker Group	?Permian to Triassic
Dir. Belt of small intrusions along the Dee Range. Largest at Mount Helen is oriented north-west and 3 by 0.3km	In area of low radiometric signature and low to moderate magnetic response	Dacite to rhyolite	Intrude Middle Devonian Mount Warner Volcanics and Raspberry Creek Formation	Middle to Upper Devonian. Similar geochemistry to the Mount Morgan Trondhjemite
Dg_g. 3km south-south-east of Mount Gelobera. Small circular intrusion about 0.5km across	In area of low radiometric signature; at margin between low and high magnetic response	Granodiorite, tonalite	Intrudes Middle Devonian Raspberry Creek Formation; faulted against Upper Devonian to Lower Carboniferous Balaclava Formation	?Middle to Upper Devonian
Dg_b. 7km south-south-west of Prior Park homestead. Two small intrusions up to about 0.5km across	In area of low radiometric signature; strong magnetic response	Gabbro	Intrude Middle Devonian Ginger Creek Member of Raspberry Creek Formation	?Middle to Upper Devonian
Dg_b?. South of Prior Park homestead. North-south oriented intrusion about 10 by up to 2.5km	Low radiometric signature; strong magnetic response	Gabbro	Intrudes Middle Devonian Raspberry Creek Formation including Ginger Creek Member	?Middle to Upper Devonian
PRg. 13 km west of Bracewell. Small circular intrusion about 0.4km across	In area of low radiometric signature; moderate magnetic response	Altered biotite tonalite. Associated with porphyry style mineralisation (Mount Bennett)	Intrudes Upper Devonian to Lower Carboniferous Mount Alma Formation	?Permian to Triassic
PRg. On Manton Creek. Ovoid intrusion a little more than 1km in its greatest dimension	Moderate radiometric signature in area of high values; moderate to strong magnetic response	?Granodiorite	Intrudes uppermost Carboniferous to Lower Permian Youlambie Conglomerate	?Permian to Triassic

Table 6 (continued)

Location and distribution	Geophysical expression	Lithology	Stratigraphic relationships	Age
PRg_g. 13km south of Cecilwood homestead. Three small intrusions, the largest 0.7 by 0.5km	In area of low radiometric and magnetic signature	Biotite granodiorite with secondary muscovite, epidote and chlorite. Chemistry typical of the Dumgree group (Figure 302). Associated with porphyry style mineralisation (Briggs or Riverhead prospect)	Intrude Upper Devonian to Lower Carboniferous Balaclava Formation	?Permian to Triassic
GLADSTONE 1:100 000 SHEET AREA				
$PRg_b{}^1$. North-west trending belt of magnetic anomalies east of Bruce Highway, extending onto Bajool sheet. Largest anomaly 3 by up to 0.5km	In area of moderate to strong radiometric signature; moderate magnetic response	?Gabbro, diorite	Intrude mixed Lower Carboniferous Rockhampton Group and Early Permian Berserker Group	?Permian to Triassic
PRg_d. 6km north-west of Targinie. Intrusion 1 by 0.8km	Moderate radiometric signature stands out in an area of high response; magnetic response low	Diorite, monzonite	Intrudes mixed Lower Carboniferous Rockhampton Group and Early Permian Berserker Group and Lower Permian Chalmers Formation	?Permian to Triassic
PRg_b. North-north-west trending belt 5km east of Mount Larcom township. Largest intrusion 3 by 1.5km	Very low radiometric signature; moderate to strong magnetic response	Gabbro, diorite	Intrude mixed Lower Carboniferous Rockhampton Group and Early Permian Berserker Group; overlain by Tertiary to Quaternary red soil (**TQr_r**)	?Permian to Triassic
BANANA 1:100 000 SHEET AREA				
PRg. Centred 13km north-east of Jambin. North-south oriented intrusion 9 by up to 4km	Very low radiometric response; northern part of intrusion is strongly magnetic, and extends to the west under the Biloela Basin, but the southern part has a low to moderate magnetic signature	Granite, diorite. 2 analyses are closest to the Miriam Vale group	Intrudes uppermost Carboniferous to Lower Permian Youlambie Conglomerate and Permian Smoky beds; overlain by Tertiary Biloela Formation and Quaternary alluvial and residual sediments (**Qr**, **Qpa** and **Qa**)	?Permian to Triassic
PRg_b. 15km north-east of Jambin. Circular intrusion about 1km across (continues onto Biloela sheet)	Very low radiometric response; strong magnetic signature	Orthopyroxene-hornblende-clinopyroxene gabbro to diorite	Intrudes uppermost Carboniferous to Lower Permian Youlambie Conglomerate and Permian Craiglands Quartz Monzodiorite	?Permian to Triassic
PRg. Small intrusions near Moura railway about 11km east of Jambin	Low radiometric and magnetic response	Granite, granodiorite	Intrude Permian Smoky beds; overlain by Tertiary Biloela Formation	?Permian to Triassic
BILOELA 1:100 000 SHEET AREA				
Mir. Mount Eugenie about 40km north of Biloela. Equant intrusion about 1.5km across	Moderate to high radiometric signature; low magnetic response	Intrusive rhyolite. Within plate, A-type chemistry	Intrudes uppermost Carboniferous to Lower Permian Youlambie Conglomerate	Mesozoic (probably Late Triassic)
PRg_d. Small intrusions in headwaters of Don River about 45km north of Biloela. Two circular intrusions less than 1km across	In area of moderate to high radiometric signature; low magnetic response	Clinopyroxene-hornblende diorite	One intrudes Upper Devonian to Lower Carboniferous Balaclava Formation; the other intrudes uppermost Carboniferous to Lower Permian Youlambie Conglomerate	?Permian to Triassic

Table 6 (continued)

Location and distribution	Geophysical expression	Lithology	Stratigraphic relationships	Age
PRg_b. Beside the upper Calliope River, less than 1km across	In area of moderate radiometric signature; low to moderate magnetic response	Clinopyroxene gabbro, diorite	Intrudes Upper Devonian to Lower Carboniferous Balaclava Formation	?Permian to Triassic
PRg_b. 30km north of Biloela. Circular intrusion about 1km across (continues onto Banana sheet)	Very low radiometric response; strong magnetic signature	Orthopyroxene-hornblende-clinopyroxene gabbro to diorite	Intrudes uppermost Carboniferous to Lower Permian Youlambie Conglomerate and Permian Craiglands Quartz Monzodiorite	?Permian to Triassic
PRg_g. Four intrusions along Bell and Back Creeks about 30km north of Biloela. Largest intrusion 3.5 by 1km	Moderate to low radiometric signature; moderate to strong magnetic response, especially the largest intrusion	Granodiorite, hornblende-biotite granite. One granite body along Back Creek is part of the Ridgelands group	Intrude uppermost Carboniferous to Lower Permian Youlambie Conglomerate, mid-Permian Yarrol Formation and Smoky beds, and Upper Permian Inverness Volcanics	?Permian to Triassic
PRg_d. Small intrusion along the Dawson Highway	In area of low radiometric and magnetic response	Hornblende diorite	Intrudes Upper Devonian to Lower Carboniferous Three Moon Conglomerate	?Permian to Triassic
PRg_g. Small intrusion south of the Rocky Point Granodiorite	In area of low radiometric signature; moderate magnetic response	?Granodiorite	Intrudes Upper Devonian to Lower Carboniferous Three Moon Conglomerate	?Permian to Triassic
PRg_g. Three intrusions in the Mount Seaview area	In areas of low to moderate radiometric signature; the largest intrusion north-east of Mount Seaview is about 2km across and is strongly magnetic, but the other two give negligible response	?Granodiorite	Intrude Upper Devonian to Lower Carboniferous Mount Alma Formation and Lower Carboniferous Rockhampton Group	?Permian to Triassic
Rg. Elongate intrusions about 20km north-east of Biloela	No radiometric or magnetic expression	Granite	Intrude uppermost Carboniferous to Lower Permian Youlambie Conglomerate	?Triassic
PRg_d. East of Biloela	In area of low radiometric response and low to moderate magnetic signature	Diorite	Intrudes Upper Devonian Lochenbar Formation	?Permian to Triassic
CALLIOPE 1:100 000 SHEET AREA				
PRg. Intrusion about 1.5km across, near Badjin Creek 8km south-south-west of Calliope	In an area of moderate radiometric and low magnetic response	?Granodiorite	Intrudes Lower Carboniferous Rockhampton Group; overlain by Quaternary alluvium (**Qa**)	?Permian to Triassic
PRg. Small intrusion less than 1km across at Glengarry homestead, 20km south of Calliope	In an area of moderate radiometric and low magnetic response	?Granodiorite	Intrudes Lower Carboniferous Rockhampton Group; overlain by Quaternary alluvium (**Qa**)	?Permian to Triassic
PRg_b. Intrusion 5km long east of Norton Creek 15km west of Miriam Vale	Very low radiometric response; reversely magnetised	Gabbro (float has been observed in an adjacent stream)	Sill intruding Castletower Granite	?Upper Triassic
PRg_g. Small intrusion less than 1km across at Mount Rand Creek 25km south-west of Miriam Vale	In area of low to moderate radiometric and magnetic response	?Granodiorite	Intrudes Lower Carboniferous Wandilla Formation	?Permian to Triassic

Table 6 (continued)

Location and distribution	Geophysical expression	Lithology	Stratigraphic relationships	Age
PRg$_g$. North-east oriented intrusion 6.5km long and up to 2.5km wide at Oaky Creek 15km south-south-west of Miriam Vale	Radiometric and magnetic response is the same as that of the Norton Tonalite	?Tonalite	Intrudes Lower Carboniferous Wandilla Formation, ?Upper Triassic Miriam Vale Granodiorite, and ?Upper Triassic Castletower Granite. May be related to the mineralisation of the Boyne River goldfield to the west	?Upper Triassic
MONTO 1:100 000 SHEET AREA				
PRg$_b$. Very small intrusion north of Munholme Creek, 35km north of Monto	In area of moderate radiometric and low magnetic response	Altered biotite-olivine-augite gabbro	Intrudes Upper Devonian to Lower Carboniferous Three Moon Conglomerate	?Permian to Triassic
PRg. Intrusion less than 1km across on Monal Creek north of the Monal Granodiorite 38km north of Monto	In an area of low radiometric and magnetic response	?Granodiorite	Intrudes the Upper Devonian to Lower Carboniferous Three Moon Conglomerate	?Permian to Triassic
PRg$_d$. Very small intrusion south of Glassford Creek 16km west-south-west of Many Peaks	In an area of moderate radiometric response and low magnetic signature	Diorite (description by exploration company)	Intrudes the Lower Carboniferous Rockhampton Group	?Permian to Triassic
PRg. Intrusion less than 1km across on Boggy Creek 10km west-south-west of Many Peaks	In an area of moderate radiometric response and low magnetic signature	Company	Intrudes the Lower Carboniferous Rockhampton Group	?Permian to Triassic
PRg. Very small intrusions west and south of Many Peaks	In an area of low radiometric response and moderate magnetic signature	?Granodiorite	Intrude the Lower Devonian Calliope beds	?Permian to Triassic
PRg$_d$. Intrusion less than 1km across on Double Creek 8km south of Many Peaks	In an area of moderate radiometric response; magnetic response is moderate, but stands out as an anomalous high point	?Diorite	Intrudes the Lower Carboniferous Wandilla Formation	?Permian to Triassic
PRg. Intrusion less than 1km across on a tributary of Minerva Creek 10km south of Many Peaks	In an area of moderate radiometric response and low magnetic signature	?Granodiorite	Intrudes the Lower Carboniferous Wandilla Formation	?Permian to Triassic
PRg. Intrusion 4 by 3km centred on Mount Beatrice 12km south-east of Many Peaks	Strong radiometric and low magnetic response	Granite; geochemistry is similar to the Miriam group, but TiO_2, Al_2O_3, and particularly Rb are high	Intrudes the Lower Carboniferous Wandilla Formation; relationship with the Triassic Bobby Volcanics and unnamed ?Permian to Triassic gabbro unknown	?Permian to Triassic
PRg$_b$. Intrusion 5km long and up to 3km wide along the Kolan River 13km south-east of Many peaks	Very low radiometric and strong magnetic response	Gabbro	Intrudes the Lower Carboniferous Wandilla Formation; relationship with the Triassic Bobby Volcanics, and ?Permian to Triassic Molangul Granite, Bororen Tonalite and unnamed granite unknown	?Permian to Triassic
PRg$_b$. Very small circular intrusion near Monal Creek 20km north of Monto	In area of low radiometric and low to moderate magnetic response	Gabbro	Intrudes the Upper Devonian to Lower Carboniferous Three Moon Conglomerate	?Permian to Triassic

Table 6 (continued)

Location and distribution	Geophysical expression	Lithology	Stratigraphic relationships	Age
PRg. 4 intrusions up to 1.5km across extending for 4km north-west from Kalpowar to Mount Cannindah	In area of low to moderate radiometric and low magnetic response	Biotite granite, hornblende diorite (extensively altered); geochemistry is closest to the Monal group, but Y and Zr and particularly CaO are lower; related to porphyry-style mineralisation at Mount Cannindah (Bedford, 1975; Creenaune & Harvey, 1996)	Intrude the Lower Carboniferous Rockhampton Group; faulted against the Upper Carboniferous to lowermost Permian Lorray Formation; relationship to unnamed Triassic volcanics unknown	?Permian to Triassic
Rib. Very small plug 3km north-west of Kalpowar	In area of moderate radiometric and low magnetic response	Basalt	Intrudes the Lower Carboniferous Rockhampton Group; relationship to unnamed ?Permian to Triassic granodiorite unknown	Triassic
PRg. 6 small intrusions extending for 7km between Bukali and Mungungo	In area of low radiometric and magnetic response	Granodiorite	Intrude the Upper Devonian to Lower Carboniferous Three Moon Conglomerate and the Lower Carboniferous Rockhampton Group	?Permian to Triassic. One intrusion was dated by Ar-Ar at 277±12Ma (Webb, 1973, 1977)
PRg. Intrusion less than 1km across on Barney Creek in the Yarrol Syncline 22km north-east of Monto	In area of strong radiometric and low magnetic response	Granodiorite	Intrudes the Upper Carboniferous to lowermost Permian Lorray Formation	?Permian to Triassic
PRg$_d$. Three intrusions up to 1.5km long in the Yarrol area 25km east-south-east of Monto	Their very low radiometric and moderate magnetic signature forms anomalies in an area of strong radiometric and low magnetic response	Diorite. Associated with mineralisation, particularly in the Yarrol goldfield (Whittaker, 1997)	Intrude the Upper Carboniferous to lowermost Permian Lorray Formation	?Permian to Triassic

STRUCTURE

The Yarrol Project area has been subjected to episodic deformation. This deformation has been studied within individual structural blocks, but few age data are available. The lack of ages makes it difficult to correlate between blocks and compile a definitive treatment of the structural evolution of the entire area.

The main episodes of deformation were:

(a) Early Late Devonian arc — continent collision

(b) Latest Devonian — Carboniferous accretion

(c) Latest Carboniferous — mid-Permian extension

(d) Mid Permian — Late Triassic compression, folding and thrusting (Hunter-Bowen Orogeny)

(e) Cretaceous and Tertiary faulting

This discussion treats deformation in chronological order.

(a) Early Late Devonian

An angular unconformity has been described near Mount Morgan between the Middle Devonian Raspberry Creek Formation and the Upper Devonian Mount Hoopbound Formation (Kirkegaard & others, 1970; Leitch

& others, 1992). The age of the unconformity can confidently be assigned to the early Late Devonian because the Mount Hoopbound Formation also overlies the 380Ma-old Mount Morgan Trondhjemite (Brown & Wilson, 1965, 1966; Anonymous, 1968; Kirkegaard & others, 1970; Messenger & Golding, 1996).

This unconformity occurs throughout the project area. Malone & others (1969) described an angular discordance between the uppermost Silurian to Middle Devonian Craigilee beds and Lower Carboniferous strata (probably Rockhampton Group) to the east of Armagh homestead which they interpreted as an unconformity, but the contact is now mapped as a fault. Evidence for an unconformable relationship in this area is provided by the occurrence of fossiliferous Middle Devonian or older limestone clasts in the Mount Alma Formation (Kirkegaard & others, 1970, page 45). The Mount Holly beds of Kirkegaard & others (1970) are now mapped as two unconformable units, the uppermost Silurian to Lower or Middle Devonian Erebus beds and the Upper Devonian to Lower Carboniferous Mount Alma Formation, which contains large allochthonous limestone blocks from the older unit and also from the uppermost Silurian to Lower Devonian Calliope beds. It was originally thought that continuous deposition from Middle to Late Devonian was represented by the Kroombit beds (Dear & others, 1971), but the sequence is clearly divisible into the Middle Devonian Marble Waterhole beds and Upper Devonian Lochenbar Formation in fault contact. Elsewhere in the project area the unconformity is indicated by the occurrence of fossiliferous limestone clasts or basement highs in younger sequences (Kirkegaard & others, 1970, pages 10, 34, 48–49, 103, 138; Dear & others, 1971, page 10). Some of these occurrences suggest that the time break was much longer than at Mount Morgan.

The Late Devonian unconformity was proposed as evidence for an orogeny by Malone & others (1969), Kirkegaard & others (1970) and Dear & others (1971), associated with the local generation of folding and cleavage in Middle Devonian and older rocks, and this has been accepted by some subsequent workers (eg. Day & others, 1978, 1983). Malone & others (1969) stated that folds in the Craigilee beds pre-dated the formation of the Craigilee Anticline. At Kawana near Rockhampton, Morand (1993b) described tight mesoscopic folding of cleavage in Lower Devonian limestone. The distribution of limestone lenses at this locality (Krosch & Kay, 1977, figure 19) proves that they are allochthonous blocks. Although the enclosing Lower Carboniferous Rockhampton Group is steeply dipping, it lacks mesoscopic folds like those in the limestone (Murray & Blake, 2005). Devonian folding is indicated by field relations at Mount Morgan, where the undeformed Mount Morgan Trondhjemite cuts across folded rocks of the Capella Creek Group (Taube, 1986). A section along the Moura-Gladstone railway line (Morand, 1993a, figure 7) displays much tighter folding in limestone-bearing strata now mapped from radiometric data as Erebus beds than the gently folded Mount Alma Formation to the east.

Kirkegaard & others (1970) suggested that cleavage in their Mount Holly beds formed during the Devonian orogeny. However, Morand (1989, 1993b) noted that folds and cleavage in the uppermost Silurian to Middle Devonian rocks (mainly the Mount Holly beds and Barmundoo beds) have similar orientations to those in later sequences, and proposed a single deformation in mid- to Late Permian time which increased in intensity to the east towards the Yarrol Fault. Although subsequent mapping has changed the interpretation of the geology in this area, dividing the Mount Holly beds into two units separated by an unconformity, and placing the Barmundoo beds in the Mount Alma Formation above the unconformity, the structural observations of Morand (1993b) are still valid. Plots of cleavage in the Erebus beds and Mount Alma Formation are identical (dip azimuths 67° towards 057° and 73° towards 059° respectively, Figures 409 and 410), and similar to plots

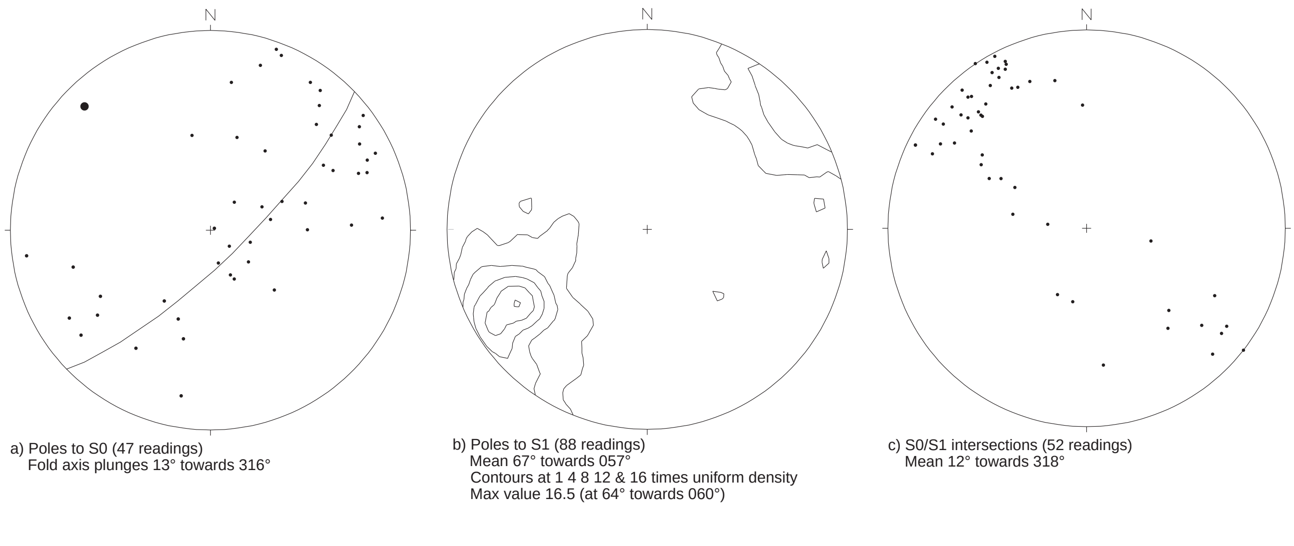
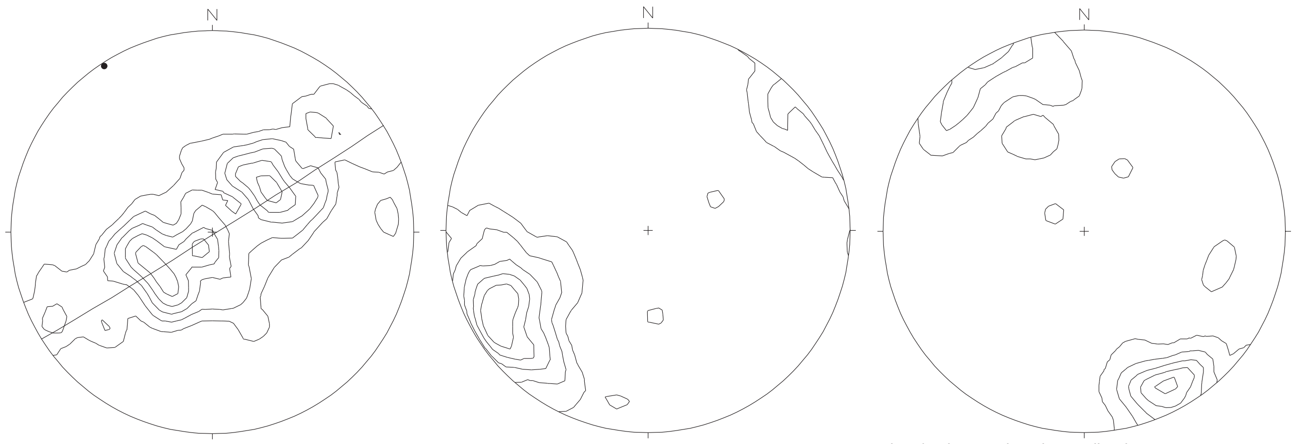


Figure 409. Plots of structural data from the Erebus beds (a) shows poles to bedding, (b) shows poles to cleavage, and (c) shows bedding/cleavage intersections.

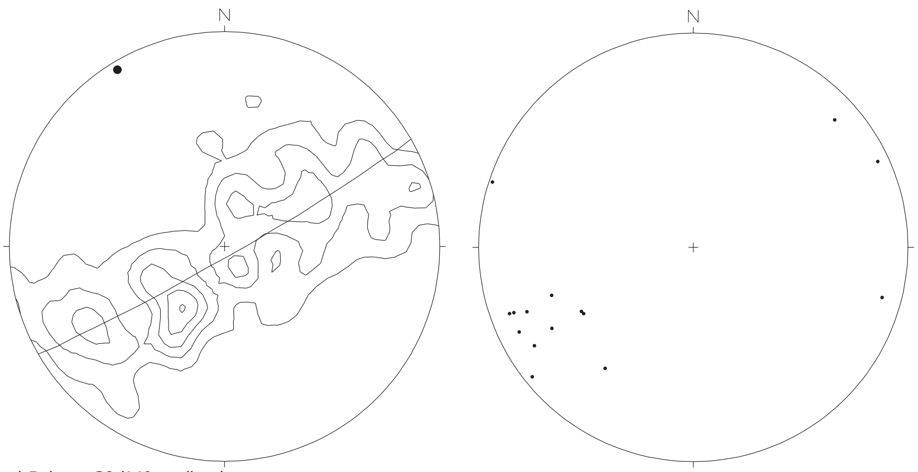


a) Poles to S0 (169 readings)
Contours at 1 2 3 4 & 5 times uniform density
Max value 5.6 (at 29° towards 231°)
Fold axis plunges 1° towards 328°)

b) Poles to S1 (80 readings)
Mean 73° towards 059°
Contours at 1 3 6 9 & 12 times uniform density
Max value 16 (at 74° towards 060°)

c) S0/S1 intersections (75 readings)
Contours at 1 5 10 15 & 20 times uniform density
Max value 21 (at 16° towards 152°)

Figure 410. Plots of structural data from the Mount Alma Formation (a) shows poles to bedding, (b) shows poles to cleavage, and (c) shows bedding/cleavage intersections.



a) Poles to S0 (146 readings)
Contours at 1 2 3 4 & 5 times uniform density
Max value 5.2 (at 29° towards 034°)
Fold axis plunges 4° towards 330°)

b) Poles to S1 (15 readings)
Mean 15° towards 065°

Figure 411. Plots of structural data from the Rockhampton Group in the region of the Erebus beds (a) shows poles to bedding, and (b) shows poles to cleavage.

of cleavage in the Mount Holly beds and Barmundoo beds of Morand (1993b, figure 5). Cleavage within the Erebus beds and the Mount Alma Formations is usually more strongly developed proximal to the Queenslander Fault and its splays, not near the Yarrol Fault as claimed by Morand (1993b). The relation between faulting and cleavage in the Mount Alma Formation is shown in Figure 81. The Queenslander Fault and related faults are thrusts that cut through both formations and are interpreted to be related to the phase of deformation that produced the folding and cleavage of the Mount Alma Formation and Rockhampton Group. This phase of deformation is thought to have occurred in the mid-Permian to Late Triassic Hunter-Bowen Orogeny. Poles to bedding in the Erebus beds define fold axes plunging shallowly towards the north-west (about 13° towards 316°), compared with the Mount Alma Formation fold axes which plunge more shallowly (1° towards 328°) (Figures 409 and 410). It is concluded that there is no evidence for a Devonian cleavage-forming event in the Erebus beds. The cleavage was generated at the same time as that in the Mount Alma Formation, which is consistent with the close structural intercalation of belts of both units at such a scale that individual units are not mapped out on all parts of the Bajool 1:100 000 sheet area. The orientations of bedding and cleavage in the Rockhampton Group in this area (Figure 411) are also similar to those of the Erebus beds and Mount Alma Formation, although cleavage is not well developed in the Rockhampton Group (see also figure 5 of Morand, 1993b).

Cleavage is poorly developed in the Capella Creek Group (see figure 5 of Morand, 1993b), and readings come mainly from the Raspberry Creek Formation. There is evidence for cleavage formation at the end of Middle Devonian time. Cleavage is best developed in the southern part, where it is so strong that it can easily be seen

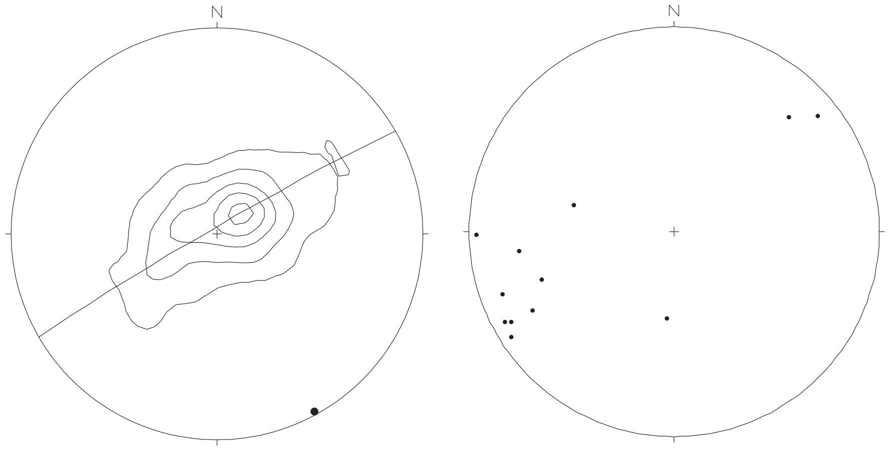


Figure 412. Plots of structural data from the Capella Creek Group (a) shows poles to bedding, and (b) shows poles to cleavage.

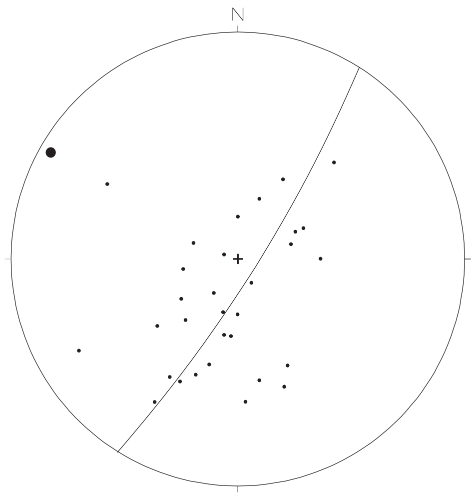


Figure 413. Poles to bedding in the Marble Waterhole beds

in coarse-grained sedimentary units such as conglomerates. However, in this area no cleavage was found in the overlying Mount Alma Formation, and the cleavage readings shown in Figure 410 come from outcrops near the core of the Alma Syncline and towards the Queenslander Fault. Similarly, no cleavages have been found in the Upper Devonian to Lower Carboniferous Balaclava Formation that also overlies the southern part of the Raspberry Creek Formation. The orientations of the cleavages in the Capella Creek Group and Mount Alma Formation differ somewhat. That in the Capella Creek Group trends more northerly than that in the Mount Alma Formation (dip azimuths 75° towards 070° versus 73° towards 059° respectively), but the significance of this difference is doubtful in view of the small number of readings from the Capella Creek Group. The poles to bedding of these two units show that fold axes of the Capella Creek Group and Mount Alma Formation both plunge shallowly to either south-east or north-west (about 2.5° towards 149° and about 1° towards 328° respectively).

The Marble Waterhole beds are not in stratigraphic contact with the adjacent Upper Devonian Lochenbar Formation, and are preserved as a fault bounded unit. However, the plunge of fold axes defined by the poles to bedding in the Marble Waterhole beds is also shallow (6° to 302°) (Figure 413).

Observations from other areas support a localised mid-Devonian cleavage-forming event. In the Craigilee beds north of Armagh homestead, Malone & others (1969) noted that some of the limestones were closely cleaved, and this is a feature of the Devonian limestone blocks at Kawana near Rockhampton (Morand, 1993b). Cleavage has not been observed in any limestone younger than Middle Devonian in the project area.

It is concluded that deformation definitely occurred at the beginning of the Late Devonian, but that the effects of this event were variable. The only area where it produced a definite angular unconformity was south of Mount Morgan, and this unconformity can be assigned to a short time range due to dating of the Mount Morgan Trondhjemite. The widespread occurrence of allochthonous fossiliferous limestones in younger sequences either as pebbles or as lenses suggests that the unconformity was regional in extent, and may have been longer lasting. Evidence of Devonian folding is only sporadically present. It appears that cleavage was not generally formed, and is preserved at only a few localities.

The significance of the mid-Devonian deformation has been a major issue in discussions on the nature of the Middle Devonian and older rocks. Interpretation of these rocks as volcanic island arc assemblages (Marsden, 1972; Day & others, 1978, 1983; Offler & Gamble, 2002; Murray & Blake, 2005) necessarily implies that they were exotic terranes subsequently accreted to the Australian continent. Arguments for a continental margin setting have attributed the deformation to local changes of plate motion (Morand, 1993b) or extension (Messenger, 1996; Bryan & others, 2003) rather than an island arc – continental collision. It has generally been assumed that an island arc – continent collision would cause more deformation than has been observed in the earliest rocks of the Yarrol project area. However, the model proposed by Murray (2007) involving an east-dipping subduction zone beneath the island arc guarantees accretion of the arc (or arcs) to the continent, and avoids a genuine arc – continent collision by preserving an intervening slab of oceanic crust. Thus the observed deformation is consistent with either scenario.

(b) Latest Devonian — Carboniferous

The easternmost strata of the Yarrol project area are interpreted as a classic accretionary wedge formed in latest Devonian and Carboniferous time above a west-dipping subduction zone. Some of the deformation-related structures in these rocks are attributed to this event, and have been described by Fergusson & others (1988, 1990a,b, 1993) who divided the accretionary wedge into two components, the Wandilla and Shoalwater terranes.

The Wandilla terrane (Doonside and Wandilla Formations) consists of mudstone, greywacke, chert, greenstone and tuff. It is largely lenticular mélange in which fragments of greywacke and tuff are enclosed by a mudstone matrix with a scaly fabric defined by anastomosing shear zones. The mélange was formed by layer-parallel extension during coaxial flattening (Fergusson & others, 1988, 1990a,b). Mud-seam mélange is less extensive. Mélange formation is interpreted to have occurred during the subduction process before the rocks were lithified, and to be produced by offscraping at the toe of the accretionary wedge (Fergusson & others, 1990b). Tight folds in coherent strata (usually of higher metamorphic grade), locally accompanied by slaty cleavage, are attributed to this early deformational phase, D_1 (Fergusson & others, 1990a, 1993). Faults sub-parallel to bedding are thought to be of this age and to indicate the probable existence of an imbricate thrust stack typical of accretionary wedges.

This stage of deformation is only represented in the Shoalwater terrane in the Shoalwater Bay region in the Port Clinton 1:250 000 area (Fergusson & others, 1990a).

This deformational episode has not been recognised west of the Yarrol Fault in the Yarrol Project area.

(c) Latest Carboniferous — mid-Permian extension

At the end of the Carboniferous, subduction ceased and the area was dominated by a period of modest extension and basin formation (Fielding & others, 1997a; Holcombe & others, 1997b). This resulted in a significant change from the palaeogeographic pattern associated with the latest Devonian and Carboniferous forearc basin, although the preserved extent of Early Permian sequences suggests that the extensional basin or basins had a general north-north-west orientation. As pointed out by Holcombe & others (1997a, page 53), no regional fault patterns have been identified which could be associated with this event. Extension did not lead to the development of deep basins, and the subsidence rate remained the same (Yarrol Project Team, 1997). The Youlambie Conglomerate, Rookwood Volcanics and Berserker Group represent this extensional phase.

Much of the Youlambie Conglomerate was laid down on basement rocks ranging in age from Middle Devonian to Early Carboniferous, indicating a substantial interval of non-deposition and/or erosion (Yarrol Project Team, 1997). Some contacts show a moderate to strong angular unconformity, and at others the units appear to be disconformable (eg. Dear, 1968). The variation in relationships at this basal contact can be attributed to different degrees of tilting of basement blocks during extension. Deposition was more continuous in some areas through the Late Carboniferous (Pennsylvanian), especially in the east of the Yarrol Province. In the area east of Monto which is now the site of the Yarrol Syncline, marine conditions prevailed through

Figure 414. Detail of the contact between the folded granitic dyke and gneiss of the Shoalwater Formation, showing cross-cutting relationship.

the Carboniferous into the Permian. Here, the upper part of the Lorray Formation (formerly placed in the Burnett Formation) is of equivalent age to the Youlambie Conglomerate (Briggs, 1998, figure 17), and the two units are separated by as little as 11km. The present distribution of the Lorray Formation in the Yarrol Syncline is the result of folding during the Hunter-Bowen Orogeny.

Relationships at the base of the Berserker Group are similar. In this case, no unconformity has been observed, but its existence is suggested by the consistent observation that the older basement rocks dip more steeply than the Permian strata.

The ages assigned to the Youlambie Conglomerate, Rookwood Volcanics and Berserker Group indicate that extension started at the end of the Carboniferous and persisted until mid-Permian time (about 270Ma). No evidence of folding of the Permian basin-fill or basement rocks during this extensional event has been observed.

Rocks of the accretionary wedge may have undergone extension and folding at this time. Fergusson & others (1993) assigned most if not all deformation subsequent to subduction and accretion to the Hunter-Bowen Orogeny, consistent with the Rb-Sr dates reported by Leitch & others (1993). However, recent U-Pb zircon dating suggests that metamorphism and folding of this sequence could have been considerably older. Carson & others (2006) reported ages from zircons in a folded granitic dyke on the rock platform at East Point at the south-eastern end of Facing Island. The dominant structure in the amphibolite facies metamorphics of the Shoalwater Formation at this locality is interpreted as a differentiated layering (S_2; Fergusson & others, 1990a). The dyke appears to have been formed during D_2 (although it locally crosscuts S_2; Figure 414) and to have been folded by D_3, which produced recumbent mesoscopic folds. The zircons from this dyke have dark, homogenous rims with low Th/U values typical of metamorphic zircon. Two spot analyses of 269.8±2.0 and 270.8±3.2Ma are interpreted to give the best age of metamorphism and D_2. These determinations are comparable with more definitive results from amphibolite facies rocks in the Bundaberg 1:250 000 sheet area, also derived from the Shoalwater Formation, that give metamorphic ages of 276.8±4.6 and 274.9±3.2Ma from zircon rims (Carson & others, 2006). They are also similar to K-Ar dates of 260Ma (whole rock) and 268Ma (biotite) (constants not specified) reported by Kirkegaard & others (1970) from gneiss at Gatcombe Head on southern Facing Island. The U-Pb zircon results suggest that folding and metamorphism of the Shoalwater Formation may have been older than the Rb-Sr ages of 230 to 255Ma reported by Leitch & others (1993) and to have preceded the Hunter-Bowen Orogeny.

Fergusson & others (1993, page 410) noted similarities between D_2 structures in the Wandilla terrane and the North D'Aguilar Block. In the North D'Aguilar Block, D_2 has been related to a major extensional event close to the Carboniferous–Permian boundary (Little & others, 1992, 1993a,b). Studies in other parts of the New England Orogen suggest that peak high-T – low-P metamorphic conditions produced amphibolite facies rocks earlier than the Hunter-Bowen Orogeny and were associated with formation of rift basins at the beginning of the Permian (Dirks & others, 1992). The results from the Bundaberg 1:250 000 sheet area and Facing Island suggest that the peak of metamorphism was much later in central Queensland, although extension commenced at the end of the Carboniferous.

(d) Mid-Permian – Late Triassic compression, folding and thrusting (Hunter-Bowen Orogeny)

Major deformation in the New England Orogen and the Bowen Basin involving west-directed thrusting and associated folding took place in mid-Permian to late Triassic time. The folding event was first recognised in the southern New England Orogen, where Carey & Browne (1938) introduced the term Hunter-Bowen Movement to explain an angular unconformity at the top of the Upper Marine Series. They attributed the folding to compression from the east. Subsequently, younger deformational events have been recognised from seismic profiles, particularly in the Bowen Basin, where the entire sequence is cut by west-directed thrusts (Korsch & others, 2009b). All this episodic thrusting appears to be related to the same thrust system, and the name Hunter-Bowen Orogeny is now applied to the whole mid-Permian to Late Triassic interval (Fergusson, 1991; Holcombe & others, 1997a; Korsch & others, 2009b). The thrust system has been imaged by deep seismic reflection profiles across the Bowen Basin (Korsch, Wake-Dyster & Johnstone, 1990, 1992a; Korsch & others, 1997). The first thrust event (the Aldebaran Event of Korsch & others, 2009b) is correlated with the Hunter-Bowen Movement of Carey & Browne (1938), and marked the start of the foreland basin phase of Bowen-Gunnedah-Sydney Basin development (Jones & others, 1984; Veevers & others, 1993; Fielding & others, 1997b; Korsch & others, 2009b). Deformation attributed to the Hunter-Bowen Orogeny has also been recognised in north Queensland (Davis & others, 1998, 2002).

The Hunter-Bowen Orogeny was the main folding event in the northern New England Orogen, but its effects differed markedly in exposed structural zones. These zones are: (1) the accretionary wedge in the east; (2) the Yarrol Province or Block in the centre; and (3) the Gogango Overfolded Zone in the west. In addition, the Marlborough Block has been thrust over the Yarrol Province.

All post-subduction structures in the **accretionary wedge** were attributed by Fergusson & others (1993) and Leitch & others (1993) to the Hunter-Bowen Orogeny. However, recent U-Pb zircon dating suggests that the metamorphism coeval with D_2 may have been related to earlier extension. If this is correct, the main deformation of the accretionary wedge, particularly the Wandilla terrane, was not related to the Hunter-Bowen Orogeny. Structures which are likely to have been produced during the Hunter-Bowen Orogeny are D_3 and D_4 of Fergusson & others (1990a). These structures reflect renewed east-west compression following D_2, and occurred at lower metamorphic grade (Fergusson & others, 1990a). D_3 produced conjugate strike-slip faults related to north-east–south-west compression which have brecciated cherts in the Wandilla Formation, and tight to broad folds with an axial plane crenulation cleavage (Fergusson & others, 1988, 1990b, 1993). Folding is much more common in the Shoalwater terrane than in the Wandilla terrane (Fergusson & others, 1990a, table 1), and recumbent folding attributed to D_3 occurs in higher grade rocks of the Shoalwater Formation at the southern end of Facing Island. D_4 has only been recognised from the Shoalwater terrane, and formed mesoscopic to macroscopic open upright folds with either no or a poorly developed axial plane cleavage (Fergusson & others, 1990a). A problem with this interpretation is the presence in the accretionary wedge of significant areas of overturned strata. These were previously explained as the result of rotation of steeply-dipping accretionary packages during thrust transport to the west (Henderson & others, 1992, 1993).

Morand (1993b) made a structural study of the most deformed belt in the **Yarrol Province** between Bajool and Ubobo. Although subsequent mapping based on airborne radiometrics has changed the interpretation of the geology in this area, his structural observations are still valid. He concluded that this belt was deformed as part of a west-vergent fold and thrust belt in the Hunter-Bowen Orogeny to produce a set of folds which trend north-west and are generally very gently plunging (Figures 409–413). The axial plane cleavage is variably developed, but steep. In the more deformed part of the belt, the attitude of bedding ranges from vertical to dipping moderately to the east-north-east (Morand, 1993b). His observation that folds become tighter, dips are steeper, and cleavage intensifies towards the Yarrol Fault is not consistent with the relative lack of deformation in the Berserker Group, which crops out immediately west of the Yarrol Fault. The Berserker Group is correlated with the Yarrol Formation, which is tightly folded in the Yarrol Syncline east of Monto. Other factors, such as the thickly bedded to massive nature of the Berserker Group or proximity to thrust faults, must have influenced the intensity of folding. The present mapping has found that within this more deformed belt, the intensity of cleavage development is greater in the vicinity of subsidiary, but major, faults such as the Queenslander Fault. This is consistent with the view that faulting, folding and cleavage formation were related to a single thrust event. The exposed faults are steep, and are assumed to flatten to the east. South-east plunging folds in the area north of Mount Larcom (Morand, 1993b, figure 5, Darts Creek Anticline) and the Mount Salmon area presumably indicate a later episode of folding, but could reflect proximity to major faults.

The fold-thrust belt model is supported by structural studies by Holcombe & others (1997a) at the eastern edge of the Gogango Overfolded Zone. Here, moderately to steeply west-dipping Carboniferous rocks of the Yarrol Province (Rockhampton Group and Lorray Formation) are interpreted as the western limb of a breached fault-propagation fold (Holcombe & others, 1997a, figure 4, section C–D). However, it seems clear that the Yarrol Province does not exhibit the classic structures of a thin-skinned fold-thrust belt like the Gogango Overfolded Zone (Fergusson, 1991). Little of the sequence appears to be missing or repeated due to the relatively small displacements on faults, and dips are generally shallow to moderate (see cross section in figure 2 of Bryan & others, 2001). Moderate to steep dips are only regionally developed in the belt studied by Morand (1993b) and its southern continuation to the Yarrol Syncline. The Yarrol Province appears to have been a relatively rigid block that moved westward as a whole, deforming the rocks of the Gogango Overfolded Zone, and the underlying detachment could be at a depth of several kilometres (Fergusson, 1991). The amount of shortening was estimated at 15% by Henderson & others (1993).

The Abercorn Trough, which underlies the Mulgildie Basin, may provide evidence for extension during the Hunter-Bowen Orogeny. This north-south, fault-bounded basin was described by Whitaker & others (1974). It is less than 10km wide and about 90km long, and contains up to 1500m of Middle Triassic sedimentary and volcanic rocks. The trough is considered to have formed during a Middle to Late Triassic tensional phase, although the bounding faults could have been initiated as reverse faults. An alternative interpretation is that the Abercorn Trough is actually a foreland basin and that its present form relates to later events.

The age of folding in the Yarrol Province is poorly constrained. Constraints on the age of folding are angular relationships between volcanic and sedimentary units, and isotopic dates from intrusives that post-date folding. The youngest unit folded in the Yarrol Syncline east of Monto is the Owl Gully Volcanics, shown as belonging to the *Echinalosia preovalis* zone of late Artinskian age (about 280–275Ma) by Briggs (1993, 1998). This provides only an older mid-Permian limit to the age of folding. The relationship between the Berserker Group and Warminster Formation indicates at least a disconformity at this time. The angular relationship between these two units is uncertain due to the poor outcrop of the Warminster Formation and the gently folded nature of the Berserker Group (Crouch & Parfrey, 1998; Crouch, 1999a, b). The oldest rocks in the Yarrol Province that post-date the folding event are Triassic volcanics which are probably Middle or Late Triassic.

The interpreted angular unconformity between the Upper Permian Dinner Creek Conglomerate and the Triassic Native Cat Andesite near Stanwell west of Rockhampton indicates that the main folding event in this area must have been Late Permian or younger. Only plant macrofossils have been recovered from the Dinner Creek Conglomerate, and it has been assigned a latest Permian age by Briggs (1993, 1998) because it overlies the poorly dated Upper Permian Moah Creek beds. However, it could be substantially older than the range of 255–250Ma given by Briggs (1993, 1998). In the eastern limb of the Candlelight Syncline, the Dinner Creek Conglomerate overlies the Upper Carboniferous (Pennsylvanian) Lorray Formation and the Lower Carboniferous (Mississippian) Rockhampton Group with little or no angular break, and all units dip to the west at about 50° (Kirkegaard & others, 1970, page 108). The Native Cat Andesite has a strong subduction signature, and is probably older than Late Triassic. Based on this meagre evidence, the age of folding cannot be more tightly constrained than latest Permian to Early Triassic.

The Permo-Triassic granites of the Yarrol Province, together with other south-eastern Queensland intrusives, have been interpreted as post-orogenic in the sense that they postdate the main folding event (Day & others, 1978, 1983). Kirkegaard & others (1970, page 108) noted that several plutons, particularly those oriented east-west to north-easterly, cut north-north-westerly trending folds. The ages of these plutons should therefore provide an upper limit for the age of folding, but two problems exist. As well as the obvious problem of the accuracy of the isotopic age determinations, the relative age of granite emplacement and folding is uncertain. This does not apply to the eastern, more deformed part of the Yarrol Province, but is a problem in the less deformed rocks along the crest and western limb of the Gracemere Anticline. Examples are the Devonian Mount Morgan Trondhjemite and Pomegranate Tonalite and the mid-Permian Kyle Mohr Igneous Complex. All three are massive and apparently undeformed, yet must have been older than and relatively unaffected by the Hunter-Bowen Orogeny apart from possible tilting, behaving as part of a rigid block. Structural data from the intruded rocks reinforces this conclusion. An arcuate, east-west trending syncline parallel to and about 1km from the northern margin of the Kyle Mohr Igneous Complex probably represents a rim syncline. The orientation and position of this syncline indicates that it was not produced by folding associated with the Hunter-Bowen Orogeny. The status of other intrusions into relatively undeformed rocks with no pronounced north-north-westerly trending structural fabric, such as the late Permian Craiglands Quartz Monzodiorite and Flaggy Quartz Monzodiorite (both about 257Ma), is uncertain. A prominent foliation in the Flaggy Quartz Monzodiorite was related to emplacement, not later deformation. Intrusions which are more definitely later than the main folding event fall in the age range from latest Permian to Late Triassic. The oldest group of these includes the Ridgelands Granodiorite, Bajool Quartz Diorite, Redshirt Granite and Mount Seaview Igneous Complex, which have given K-Ar, Rb-Sr and U-Pb ages between 253 and 249Ma, indicating that folding must have been latest Permian or earlier.

The similarity of structural style and the presence of moderate dips in the Rockhampton Group and Lorray Formation in both the Yarrol and Candlelight Synclines are consistent with a single major folding event in the Yarrol Province. The evidence summarised above indicates a latest Permian age.

The **Marlborough Block** is a flat-lying thrust sheet that has been interpreted to have been emplaced by out-of-sequence thrusting after the main fold-thrust event of the Yarrol Province (Holcombe & others, 1997a). Within the Marlborough Block, the Princhester Serpentinite forms alternating slices with metamorphic rocks of the Marlborough Metamorphics. Both faulted contacts, metamorphic foliation and bedding dip at moderate angles to the east (Henderson & others, 1993; Holcombe & others, 1997a; Harbort, 2001). The main fabric in the metamorphic rocks is parallel to the sedimentary layering where that can be recognised, and is locally overprinted by a second deformation (Kirkegaard & others, 1970; Henderson & others, 1993).

Regional gravity data (Darby, 1969, plate 11) indicate that the Marlborough Block is a flat sheet less than 3km thick, and Murray (1974) postulated that it was emplaced to its present position by at least 60km of westward thrusting. The mechanical problem of emplacing a thin sheet over such a distance led Harbort (2001) and Harbort & others (2001) to propose that the fault slices of the Marlborough Block are the lower part of an eroded duplex in an out-of-sequence nappe structure that was thrust over the rocks of the northern part of the Yarrol Province and adjacent Gogango Overfolded Zone. Holcombe & others (1997a) named the basal thrust the Marlborough Thrust, and noted that it “is a brittle structure with little deformation, even within metres of the contact”. Internal faults within the block are associated with kilometre-scale ductile deformation and although now steep, are thought to be rotated thrusts or remnants of extensional faults. The serpentinite is largely massive with localised cleaved zones, mainly in regions disrupted by late (Cretaceous and Tertiary) normal faulting, and strongly schistose zones near the contacts with surrounding units. The bases of the serpentinite sheets are metamorphosed to upper greenschist and amphibolite facies and generally display high angle, ductile thrusting with east-over-west shear sense indicators.

The timing of final emplacement of the Marlborough duplex is provided by the age of Early to mid-Permian rocks which underlie the basal thrust, and by the age of intrusive rocks that either are cut by the basal detachment or pierce both the detachment and the internal thrusts. Harbort & others (2001) referred to 20 previously published K-Ar ages and 3 Rb-Sr ages supplemented by 22 ^{40}Ar/^{39}Ar step-heating results that constrained emplacement of the Marlborough Block to 253Ma (latest Permian). If correct, this means either that the K-Ar dates of 264 and 269Ma obtained by Webb (1969) from hornblende of the Ridgelands Granodiorite (now the Wattlebank Granodiorite) intruding serpentinite on the south bank of the Fitzroy River near Eden Bann homestead are incorrect, or that the granodiorite was intruded before the thrusting event and is part of the nappe sheet. Near Milman railway siding and Eden Bann, the Princhester Serpentinite has been thrust over the Berserker Group and correlatives. Therefore, the overlap of these ages with that obtained from U-Pb zircon dating of the Ellrott Rhyolite of the Berserker Group (268.2±3.9Ma; Appendix 1) suggests that the dates obtained by Webb (1969) must be older than the emplacement event for the Princhester Serpentinite. The metamorphic rocks interleaved with the ultramafics of the Marlborough Block have been dated by Rb-Sr at 245±2Ma and 242±16Ma (Henderson & others, 1993) and by Ar-Ar at 248.8±0.5Ma and 242.8±0.4Ma (Holcombe & others, 1997a,b), but these results are appreciably younger than the emplacement age deduced by Harbort & others (2001).

Other possible late, out-of-sequence thrusts in the Yarrol Province displace previously folded strata, and are recognised because they place younger strata over older. They include the Bracewell Fault west of the township of Mount Larcom, and the Tiverton Fault which cuts the Yarrol Syncline sequence east of Monto. The age of these faults is unknown, but could be the same as that of the Marlborough Thrust. It is possible that more of these late faults exist.

The structure of the **Gogango Overfolded Zone** was described by Withnall & others (2008). The zone is characterised throughout by the development of strong east-dipping slaty cleavage in mudstone and siltstone (Figure 128), although sandstone and volcanic rocks also commonly have a well-developed foliation, generally expressed as an anastomosing fracture cleavage or dissolution cleavage. Stretching lineations generally plunge east, and shear-sense indicators such as small-scale duplexes are consistent with east-over-west transport. The intensity of deformation increases eastwards towards the major thrusts that form the western margin of the Rookwood Volcanics and the Yarrol Province. Alternating belts of volcanic and sedimentary rocks generally show a consistent eastward younging and are interpreted as an imbricate stack of thrust sheets.

Bedding dominantly dips towards the east to east-north-east at about 30°. The general trends change from east-north-east in the central and southern sectors to north in the northern sector. The slaty cleavage generally has the same strike with mainly easterly dips. The cleavage dips are steeper than those for bedding, consistent with the dominant younging and a westward vergence. Mesoscopic fold closures are present locally. Holcombe & others (1997a) state that broad areas of steep to overturned dips are an indication that thrust-propagation folds are common. However, most folds are upright, and the description “overfolded” zone is somewhat of a misnomer. Thus the Gogango Thrust Zone may be a more appropriate term for the belt (Korsch & others, 1997).

Contractional shortening was considerable, consistent with the intensity of cleavage development. It was estimated at 70 to 80% by Fergusson (1991) and at greater than 50% by Henderson & others (1993). The zone terminates relatively abruptly to the west of Marlborough and farther north the Bowen Basin succession is only gently folded. The termination is probably a complex zone of linked tear faults that are the south-westerly continuation of the Stanage Fault Zone (Morand, 1993c; Henderson & others, 1993; Leitch & others, 1994b; Holcombe & others, 1997a). The latter was a major tear fault system to the nappe emplacement of the Marlborough Block with dextral displacement of about 50km.

Strata as young as Late Permian were deformed in the Gogango Overfolded Zone. Thus the age of thrusting and folding was probably latest Permian, the same as the marked angular unconformity between the Dinner Creek Conglomerate and the Native Cat Andesite.

(e) Cretaceous and Tertiary faulting

Faults of Cretaceous and Tertiary age are widespread in the study area, and in many cases overprint previous thrusts and associated tear faults (Kirkegaard & others, 1970; Fielding & others, 1994; Holcombe & others, 1997a). Some reverse movement took place on a few of these faults, and the associated compression appears to have caused gentle folding.

Most faults interpreted as Cretaceous in age are oriented parallel to the earlier north-north-westerly oriented structural grain, but one prominent example trends east-west. Exposure is poor, and the faults are mainly apparent from tilting of adjacent rocks or from aeromagnetic patterns. Some border Mesozoic basins or known Cretaceous rocks. They include:

- The north-north-west trending Mulgildie Fault cuts rock units of and forms the eastern margin of the Mulgildie Basin. The Jurassic sediments adjacent to this fault are vertical to steeply dipping where Palaeozoic rocks have been thrust over the basin sequence along the high-angle reverse Mulgildie Fault (Figure 204). This movement may have been associated with gentle folding. Away from the eastern boundary fault, the Mulgildie Basin is a broad syncline with dips of up to 5°. Reid (1927) recognised a north-north-west trending normal fault, subsequently named the Anyarro Fault, along the western boundary of the Mulgildie Coal Measures, associated with steep dips. The coal measures appear to be downthrown to the east along this fault, which might be partly responsible for their preservation in the Mulgildie-Monto area. However, further south on the Eidsvold 1:100 000 sheet, the displacement is apparently reversed, and the Mulgildie Coal Measures have been mapped only west of the fault zone. The Mulgildie and Anyarro Faults form the boundaries of the Middle Triassic Abercorn Trough which underlies the Mulgildie Basin, and could be of this age. Alternatively, they could be later structures that define the present form of the Abercorn Trough. One problem with a Cretaceous age for the faulting is that it does not explain the cessation of deposition in the Mulgildie Basin in mid-Jurassic time.
- The east-west trending Stanwell Fault forming the northern boundary of the Stanwell Basin, was described as “one of the notable faults in Queensland, combining as it does striking topographical evidence supported by adequate geological evidence” by Reid & Morton (1928). As along the Mulgildie Fault, the adjacent basin rocks have been tilted to steep dips, possibly accompanied by local folding.
- North-north-west and north-west trending faults form a half-graben downfaulted more on the eastern side, and occupied by basalts which trend along the Fitzroy River valley (Mather, 1980). These faults are very prominent on the first vertical derivative of the aeromagnetic data.
- A strong north-west magnetic lineament runs through and partly forms the boundary of the Mount Salmon Volcanics;

- A north-west trending fault along the western side of the Jim Crow Basin also forms the boundary of the outcrop of the Berserker Group and displaces the accretionary wedge east of the Yarrol Fault;

- A prominent north-west trending fault that cuts the Eulogie Park Gabbro is most obvious on aeromagnetic images, and may be of Cretaceous age.

- A north-west trending fault displacing the Norton Tonalite has had about 2km of interpreted sinistral strike-slip movement.

The age of these faults is defined only by the ages of the affected rocks, and is not known more precisely than post-Early Cretaceous. The assumption that they are all of the same mid- to Late Cretaceous age may be incorrect, particularly in the case of the Mulgildie and Anyarro Faults.

The formation of early Tertiary (Eocene) grabens and half grabens has been interpreted to mark an extensional event associated with spreading in the Coral Sea and northern Tasman Sea. A number of basins are half-grabens with the western boundary formed by a normal fault with up to 1km of displacement. These faults have a north-north-west trend and reflect the basement structure. The best examples are The Narrows Graben and the Duaringa Basin, but the Biloela Basin may also have this form, and the beds in the Nagoorin Graben dip to the west at about 15°. Two small unnamed Tertiary basins west of Gladstone are interpreted to be similar half-grabens, but the larger unnamed basin underlying the Calliope River south-west of Gladstone appears to be faulted along its eastern margin. The bounding faults remained active during and probably after deposition, as dips steepen towards them, and even become vertical or locally overturned (Lindner, 1980). The conversion of some basins to a graben structure (for example, the Nagoorin Graben) was probably due to later faulting along a line parallel to and east of the western boundary faults. Numerous faults cut the graben sequences. The age of this later faulting is unknown, but it is interpreted to have been a continuation of the process which created the basins. It was probably not related to the widespread eruption of Tertiary basalt in the mid-Oligocene, because the basalt does not appear to be faulted, and overlies the deeply weathered Biloela Basin sequence.

Movement on the bounding faults is dominantly normal, implying a significant amount of approximately east-west extension for the formation of these basins in Eocene time. In most cases, strike-slip motion can only have been minor. An exception may be the Rossmoya Basin, which appears to be bordered on both the north-eastern and south-western margins by major faults evident on the magnetic images. The shape of this basin is consistent with formation by sinistral strike-slip movement.

Controlling faults have not been recognised in the Casuarina and Yaamba Basins. The Yaamba Basin is faulted along its northern and north-eastern boundaries, and the sequence is cut by later north-west trending fault zones (Olive & others, 1984).

Tertiary faults cut the Princhester Serpentinite and the lateritic profile developed on it, but have produced only gentle tilting.

A unique north-north-west trending fault from the Kolan River to Bajool is the continuation of a composite fault zone which extends north from Brisbane. Various names have been applied to different sections, including North Pine Fault, Perry Lineament, and Mount Bania Fault. It was first described as a single structure by Veevers & others (1993, 1994), and later discussed by Holcombe & others (1997a). The southern half, as far north as Mount Perry, has a sinistral displacement of 7 to 9km. As it is traced north, the fault zone steps westward and the sinistral displacement appears to decrease, finally ending in a series of splay faults on the Bajool 1:100 000 Sheet area. Its northernmost expression is a north-north-west trending fault visible on aeromagnetic images which displaces magnetic features at the northern margin of the Bajool Quartz Diorite. The fault trends into the Nagoorin Graben, and also is interpreted to form the boundary of unnamed Tertiary basins south-west and west of Gladstone. The oldest units displaced are Late Triassic (Veevers & others, 1994; Holcombe & others, 1997a), so the fault could have had a long history. However, it was probably also active in the Tertiary, and a very young age is supported by the fact that it has strong topographic expression in some areas, notably as paired faults visible on air photos of the Esk Trough. Attempts to model the formation of the Nagoorin Graben by strike-slip movement on this fault were unsuccessful, but it may be related to a probable strike-slip fault identified on a seismic section through the graben sequence.

TECTONIC EVOLUTION

NEOPROTEROZOIC

The oldest rocks known from the area of the Yarrol Province are metagabbroic to metabasaltic dykes and included blocks within the Princhester Serpentinite that were dated by a Nd-Sm isochron as 562±22Ma (Bruce & others, 2000b), just older than the Proterozoic–Cambrian boundary (Gradstein & others, 2004). These rocks have compositions close to normal mid-ocean ridge basalt (N-MORB), and record partial melting of the Princhester Serpentinite at an oceanic spreading centre in Neoproterozoic time (Bruce & others, 2000b). The serpentinite subsequently interacted with tholeiitic and calc-alkaline basalt magmas at a spreading centre (or centres) in a supra-subduction zone (SSZ) setting, and represents the mantle component of a Late Devonian SSZ ophiolite (Murray, 2006, 2007).

The Neoproterozoic history of the Princhester Serpentinite can be correlated with the final phase of lithospheric extension to form the palaeo-Pacific Ocean. Glen (2005, pages 36–39) regarded the Princhester Serpentinite as the product of rifting in a proto-Pacific Ocean after supercontinent breakup. Cambrian SSZ ophiolites and volcaniclastic sediments along the Peel Fault in the southern NEO (Cawood & Leitch, 1985; Aitchison & others, 1992, 1994; Aitchison & Ireland, 1995) represent the subsequent convergent (subduction) phase. However, there is no record of early Palaeozoic SSZ ophiolites or magmatic arcs in the northern NEO, representing a major difference from the southern section. The Princhester Serpentinite must have been outboard of any early Palaeozoic subduction zone along the margin of the Australian continent (Gondwana), and the 562Ma partial melting event at an oceanic spreading centre cannot be regarded as representing the early history of the NEO (Murray, 2007).

ORDOVICIAN

A very small outcrop of Ordovician limestone within the Calliope beds just west of the Yarrol Fault is either a seamount capping scraped off during subsequent subduction, or a remnant from a precursor of the island arc represented by the Calliope beds.

LATEST SILURIAN – MIDDLE DEVONIAN

The tectonic setting of latest Silurian to Middle Devonian sequences has been the subject of debate. Marsden (1972) was the first to place these rocks in an island arc setting. His interpretation was accepted by Murray (1975) and Day & others (1978), who grouped several discrete inliers and fault blocks in a single tectonic assemblage, the Calliope Volcanic Arc. This was identified as a probable exotic terrane by Veevers & others (1982). An angular unconformity at Mount Morgan between Middle and Upper Devonian sequences was thought to be related to collision of this island arc assemblage with the Australian continent.

Morand (1993a,b) applied the name Calliope Volcanic Assemblage to the uppermost Silurian to Middle Devonian rocks, and presented evidence on their structure, provenance and palaeogeography that supported a continental margin rather than island arc environment, in line with the model proposed by Henderson (1980). Messenger (1996) carried out detailed geochemical studies of both the Mine Corridor Volcanics (correlated with the Mount Warner Volcanics of the Middle Devonian Capella Creek Group) and Mount Morgan Trondhjemite at Mount Morgan. He concluded that they formed a co-genetic magmatic assemblage similar to the modern Kermadec arc. Despite these geochemical similarities, Messenger (1996) and Messenger & others (1997, 1998) proposed that the Mount Morgan rocks formed not in a primitive intra-oceanic arc, but in an arc built on juvenile or immature continental crust at the margin of Gondwanaland.

The geochemistry of basaltic and trondhjemitic rocks in the latest Silurian to Middle Devonian sequences (Morand, 1993a; Hackman, 1995; Messenger, 1996; Offler & Gamble, 2002; P. Blevin, unpublished data; Murray & Blake, 2005) is most consistent with an intra-oceanic arc origin, as reviewed by Murray & Blake (2005). They pointed out that modern island arcs with primitive compositions like those of the Mount Warner Volcanics are typically associated with an active back-arc basin (Woodhead & others, 1993), and that the increasing age of latest Silurian to Middle Devonian sequences east from the Capella Creek Group supports an origin as remnant arcs split off by back-arc basin spreading above an east-dipping subduction zone. Murray & Blake (2005) and Murray (2007) also proposed that collision of the latest Silurian to Middle Devonian assemblage with the continent near the beginning of Late Devonian time led to a reversal of subduction direction, a model that was first proposed, but rejected, by Morand (1993a), and subsequently put forward for the southern New England Orogen by Aitchison & Flood (1995).

A tectonic model for the generation of the latest Silurian to Middle Devonian intra-oceanic arc assemblages, and their collision with the Australian continent (Gondwana) in the Late Devonian, is depicted in Figure 415 (after Murray, 2007). The Middle Devonian Capella Creek Group at Mount Morgan formed in a primitive intra-oceanic arc above an east-dipping subduction zone, with older sequences to the east (Erebus and Calliope beds) representing inactive remnant arcs (Murray & Blake 2005). At about 380Ma, the Australian continent encountered and blocked the subduction zone. The effects of the collision were recorded in the Capella Creek Group as well as in sequences as far west as the Adavale Basin in central Queensland (McKillop & others, 2007). The Mount Morgan Trondhjemite, dated by U-Pb zircon at 380Ma (Golding & others, 1994b; Yarrol Project Team, 2003), cuts folded rocks of the Mount Warner Volcanics, the middle unit of the Capella Creek Group (Taube, 1986), and there is an angular unconformity at the top of the Group (Kirkegaard & others, 1970; Leitch & others, 1992).

Slab pull, although the main driver of plate movements (Forsyth & Uyeda, 1975), could not drag down the buoyant continental crust, and the slab broke off and sank, a process that could have been as rapid as a few million years after the initiation of subduction of continental lithosphere (Figure 415c).

LATE DEVONIAN

Upper Devonian rocks that unconformably overlie the Middle Devonian Capella Creek Group were traditionally regarded as the earliest part of a transitional arc – forearc basin assemblage that formed along the continental margin above a west-dipping subduction zone, a tectonic setting that persisted until the end of Carboniferous time (Murray, 1975d; Day & others, 1978; Murray & others, 1987).

This view was challenged by Bryan & others (2001), who proposed that the entire Upper Devonian and Carboniferous succession previously interpreted as a forearc basin sequence (Murray & others, 1987) was in fact deposited in a back-arc basin formed at a rifted continental margin. They used the compositions of basalts and basaltic andesites in Upper Devonian units between Mount Morgan and Monto to support a back-arc basin setting. Murray & Blake (2005) disputed this interpretation, and compared basalts and basaltic andesites in the Mount Hoopbound Formation, Lochenbar Formation and Three Moon Conglomerate with mature or evolved island arcs like the modern Lesser Antilles, Marianas, Vanuatu and Aleutian arcs. The back-arc basin basalts first recognised by Bryan & others (2001) are only a minor component, and are restricted to the eastern extremity of these Upper Devonian units. These back-arc basin basalts have compositions similar to rocks from the modern Scotia Sea and North Fiji Basin (Murray & Blake, 2005), indicating that deposition at least commenced in an oceanic environment. The change in basalt geochemistry from back-arc basin basalts in the east to island arc compositions in the west suggests that the volcanics were related to or inherited from the same east-dipping subduction zone that produced the Middle Devonian and older rocks.

Island arc tholeiites from the 380Ma V2 group of mafic intrusions and included blocks within the Princhester Serpentinite (Bruce & Niu, 2000a) are linked to Late Devonian basalts between Mount Morgan and Monto by their similar geochemistry (Murray & Blake, 2005), suggesting a common source and process in magma generation. The REE chemistry of the Princhester Serpentinite can be explained by a single process of reaction between depleted peridotite and magmas of the V2 group by equilibrium porous flow, creating a chromatographic column effect that preferentially enriched the peridotite in LREE. The Princhester Serpentinite (and the Northumberland Serpentinite on South Island of the Percy Isles 135km to the north-north-east that reacted with boninitic liquids) is the mantle component of a Late Devonian supra-subduction zone (SSZ) ophiolite formed at forearc, intra-arc and back-arc spreading centres above the east-dipping subduction zone (Figure 415b,c) (Murray, 2006, 2007).

The most widespread Upper Devonian unit, the Mount Alma Formation, was deposited in an intra-oceanic back-arc basin to the east of an active arc which sourced most of the volcaniclastic sediments (Figure 415c). However, remnant arcs within and along the eastern margin of the back-arc basin were local sources of sedimentary detritus, including large allochthonous limestone blocks (Mawson & others, 1995; Yarrol Project Team, 1997; Blake & others, 1998). This pattern of deposition continued into latest Devonian (Famennian) time.

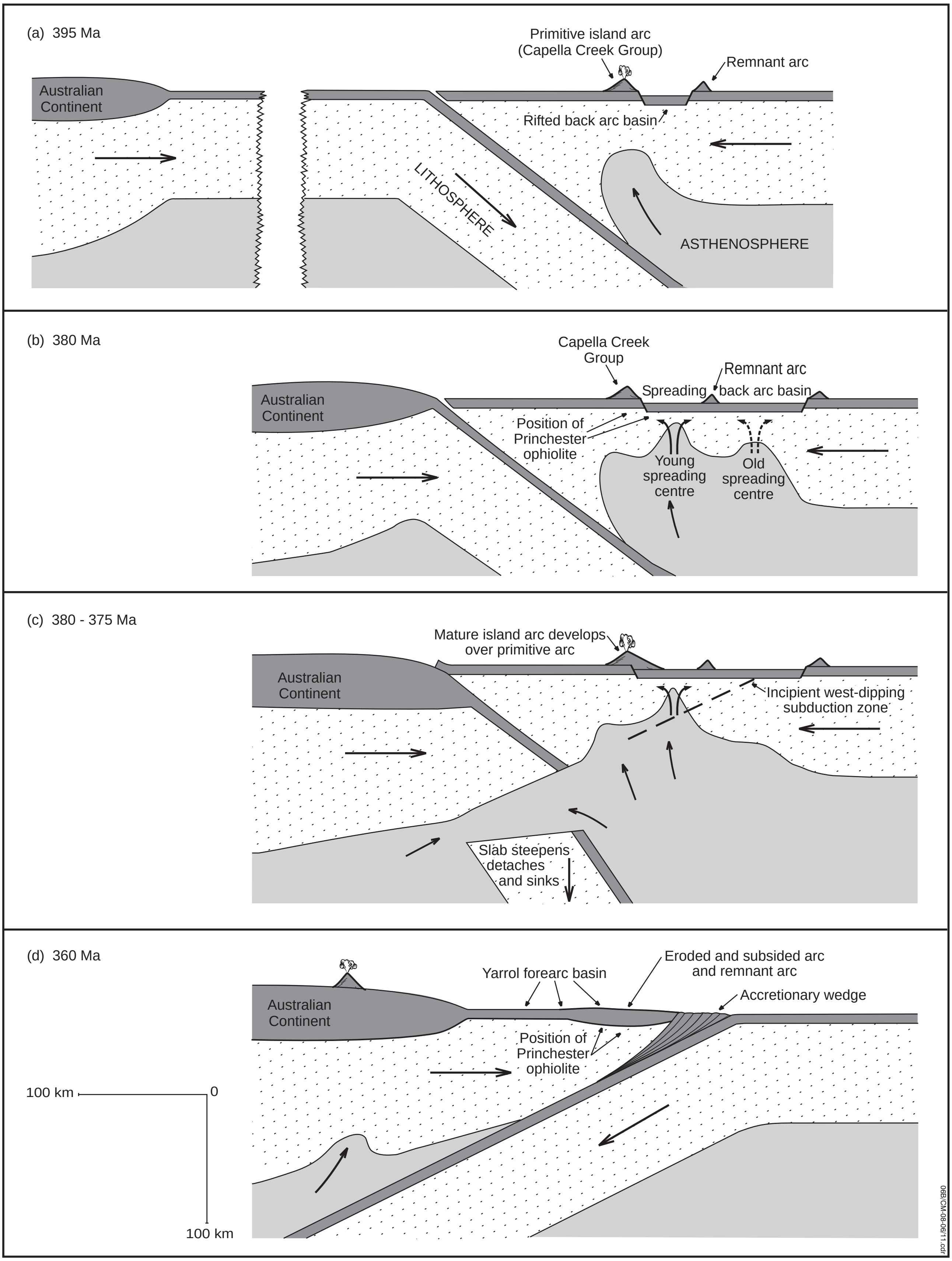


Figure 415. Tectonic evolution of the Yarrol Province in Middle Devonian to earliest Carboniferous time (after Murray, 2007).

LATEST DEVONIAN – CARBONIFEROUS

Continuing plate convergence after blockage of a subduction zone is the most probable mechanism capable of generating the forces required to create a new subduction zone (Mueller & Phillips, 1991). An inevitable result of the arc-continent collision was the reversal of subduction direction and formation of a new west-dipping subduction zone beneath the continental margin, as proposed for the southern New England Orogen in Late Devonian time (Aitchison & Flood, 1995). The new subduction zone was most likely to be initiated at a pre-existing zone of weakness, such as a spreading centre within the back-arc basin (Wakabayashi & Dilek, 2003) or along the margin of the basin (Casey & Dewey, 1984) (Figure 415c). This pattern of west-dipping subduction continued through much of Carboniferous time, and is recorded in parallel belts comprising an eastern accretionary wedge, a central forearc basin, and remnants of a western continental margin arc that form the basic tectonic elements of the New England Orogen (Murray, 1975d; Day & others, 1978; Murray & others, 1987).

The reversal of subduction direction trapped the Devonian arcs and underlying oceanic crust, including the future Princhester ophiolite, in a forearc position above the growing accretionary wedge to form a classic Cordilleran-type ophiolite (Figure 415d). The transition from back-arc to forearc basin setting occurred during the latest Devonian, near the end of deposition of the Mount Alma Formation. As the forearc basin became established, the Devonian arcs were eroded, and they subsided to the point where they no longer influenced depositional patterns. The forearc basin sequence extended across the outcrop belt of the Capella Creek Group, which had previously separated different depositional facies in the Late Devonian (Blake & others, 1998). Deposition of the forearc basin sequence therefore began on oceanic crust, as postulated by Henderson & others (1993).

By 360Ma, the subduction zone may have been operating for 10 million years, and continental margin arc volcanism had started. Widespread felsic volcanism commenced in central coastal Queensland at this time (Bryan & others, 2004), consistent with a change from mafic to felsic volcanism throughout the New England Orogen (Leitch 1974), and a marked increase in the radiometric signature of volcaniclastic sediments. Similar changes in magma chemistry have been recorded for other examples of collisions between intra-oceanic arcs and continents (*e.g.* Draut & Clift, 2001).

The Lower Carboniferous (Mississippian) Rockhampton Group reflects this change to more felsic compositions, and is readily distinguished from earlier units on radiometric images. It consists mainly of volcaniclastic sediments derived from the western arc, but is characterised by the repeated development of oolitic carbonate banks along the western margin which supplied ooids to deeper water sandstones to the east. By the time the overlying Upper Carboniferous (Pennsylvanian) to lowermost Permian Lorray Formation was deposited, the volcanic arc was less active and more deeply eroded, so that the volcaniclastic component declined and the plutonic component increased. This is marked by another sharp increase in the radiometric signature.

To the east, an accretionary wedge assemblage developed throughout the Carboniferous (Korsch & others, 2009a), with strong provenance linkages to the Early Carboniferous forearc basin sequence (McKellar 1967; Fleming & others, 1975; Murray & others, 1987) (Figure 415d). The western part of the accretionary wedge consists of ribbon cherts with interbedded shales and some associated mafic volcanics which formed the seafloor of the palaeo-Pacific Ocean. The middle section is composed of volcaniclastic sandstones with a small but persistent component of calcareous ooids which were derived from the forearc basin to the west. Dating of detrital zircons shows that the volcaniclastic component of these rocks was sourced almost exclusively from the Carboniferous continental margin arc (Korsch & others, 2009a). The easternmost accretionary wedge rocks are interbedded quartose sandstones and thin dark grey shale layers. The change from volcaniclastic to quartose compositions marks a distinct provenance switch (Leitch & others, 2003). Dating of detrital zircons indicates that, although the Carboniferous volcanic arc continued to be a source, the majority of material came from cratonic areas of inland and northern Queensland, either by transport from the north along the trench, or via rivers that cut through the arc (Carson & others, 2006; Korsch & others, 2009a).

LATEST CARBONIFEROUS TO MID-PERMIAN

The convergent cycle ended about the end of the Carboniferous. The pattern of deposition changed in response to a change from compressional to extensional tectonics (Holcombe & others, 1997a), and rocks formed during this episode were included in the Early Permian extensional package of Fielding & others (1997a). Over much of the area, older basement rocks were eroded and tilted to varying degrees, so that the contact with the Early to mid-Permian extensional package ranges from a disconformity to an angular unconformity. Continuous deposition from the Carboniferous to the mid-Permian is only recorded in the

Yarrol Syncline east of Monto, where the present distribution of rock units is due to later folding during the Hunter-Bowen Orogeny. Extension was mild, and it appears that subsidence was neither extreme nor rapid, and was little changed from the preceding convergent phase associated with west-dipping subduction (Yarrol Project Team, 1997).

The Youlambie Conglomerate, the oldest unit of the extensional package, is characterised by conglomerate with clasts of granite and purple rhyolite. Rhyolitic clasts and dykes within conglomerate at the base of the Youlambie Conglomerate along the Dawson Highway north-east of Biloela which have been dated as latest Carboniferous have a strong subduction input, suggesting that their chemistry was still related to the previous tectonic environment.

The Berserker Group and Rookwood Volcanics are also assigned to this tectonic setting, indicating that extension continued to mid-Permian time. Two pieces of evidence suggest that the Berserker Group formed in a back-arc environment: the presence of classic Kuroko-style mineralisation at Mount Chalmers and in smaller deposits is indicative of a back-arc setting (Marani & others, 1997); and the chemistry of the volcanic rocks shows some subduction influence. The Rookwood Volcanics have typical oceanic back-arc chemistry, which indicates that “sub-continental lithosphere at that time was very thin or absent, such that an evolved lithosphere did not contribute significantly to the sources of the magmas” (Stephens & others, 1996a). Their chemistry is consistent with the hypothesis that the Carboniferous forearc sequence was deposited on oceanic crust. The Kyle Mohr Igneous Complex, dated at 270.6±3.7Ma, is a bimodal intrusion that was probably co-magmatic with the Rookwood Volcanics. Recent dating suggests that the Miriam Vale Granodiorite and Castletower Granite are of similar age.

The position of the volcanic arc to the east of the Berserker Group and Rookwood Volcanics may be marked by the coeval Owl Gully Volcanics east and north of Monto. The andesitic Owl Gully Volcanics have a strong subduction signature, and their chemistry is consistent with eruption in a continental margin arc or a mature island arc with a maximum crustal thickness of about 30km.

The accretionary wedge east of the Yarrol Fault may also show evidence of this extensional event. Dating of metamorphic overgrowths on zircon from the higher-grade parts of the assemblage has given mid-Permian ages (Carson & others, 2006). Thus the peak of metamorphism may have been mid-Permian, and not Triassic as proposed by Leitch & others (1993). This raises the possibility that flat to gently-dipping structures in the accretionary wedge, related to the Hunter-Bowen Orogeny by Fergusson & others (1988, 1990, 1993) were in fact related to the mid-Permian extensional event.

The extensional phase was followed by brief sag phase (the mid-Permian thermal subsidence package of Fielding & others, 1997a). In the Yarrol Project area, this is represented only by the lower part of the Moah Creek beds (Fielding & others, 1997a).

HUNTER-BOWEN OROGENY

The Hunter-Bowen Orogeny was a protracted mid-Late Permian to Middle Triassic event (265–230Ma) during which west-directed thrusts deformed the New England Orogen and the Bowen Basin to the west (Korsch & others, 2009b). Deformation was strongly partitioned from east to west, with the result that the orogeny had very different effects in different areas.

Initial deformation of the accretionary wedge was related to its formation in the Carboniferous. All subsequent folding was attributed to the Hunter-Bowen Orogeny by Fergusson & others (1988, 1990, 1993). Recent zircon dating suggests that the main episode of metamorphism and folding was mid-Permian. If that is the case, the only structures definitely produced by the Hunter-Bowen Orogeny are overprinting broad folds with variably developed steeply-dipping axial plane cleavage (D_3 and D_4 of Fergusson & others, 1990, 1993).

The intensity of folding and cleavage development in the Yarrol Province west of the Yarrol Fault is very variable. The most deformed belt from Bajool to Ubobo was studied by Morand (1993b), and is characterised by large scale upright folds with steeply-dipping axial plane cleavage. It is interpreted as a thick-skinned fold-thrust belt produced by west-directed thrusting, with the detachment zone at a depth of several kilometres or more (Fergusson, 1991). Morand (1993b) considered that the folds become tighter and cleavage intensifies towards the Yarrol Fault to the east, but this is contradicted by the gently folded nature of the mid-Permian Berserker Group. In this case, lithology may have been one of the factors controlling deformation. However, proximity to major faults within the deformed belt appears to have been of paramount importance. The Yarrol Province behaved as a relatively thick and rigid crustal block that moved westward to cause the deformation in the Gogango Overfolded Zone and the Folded Zone of the Bowen Basin (Fergusson, 1991). Thrusting is most obvious along the western boundary with the Gogango Overfolded Zone (Holcombe & others, 1997a). The age of the main folding event was Late Permian.

Three later, out-of-sequence thrusts have been recognised. The main is the Marlborough Thrust at the base of the Marlborough Block. The Marlborough Block has been interpreted as an eroded duplex in a nappe structure that was thrust over the folded rocks of the northern part of the Yarrol Province and adjacent Gogango Overfolded Zone (Harbort, 2001; Harbort & others, 2001). The Bracewell and Tiverton Faults, in the Bajool and Yarrol Syncline areas respectively, place younger rocks on older, indicating that they post-dated the main folding event in the Yarrol Province, and may have the same age as the Marlborough Thrust.

Deformation was concentrated in the Gogango Overfolded Zone, which is a stack of imbricate thrust sheets developed in rocks of the Bowen Basin immediately west of the Yarrol Province. Bedding dips to the east at about 30°, and the cleavage, best seen in siltstones interbedded with sandstones, is steeper, giving a consistent vergence which indicates east over west transport. Contractional shortening was more than 50% (Fergusson, 1991; Henderson & others, 1993). Sediments as young as Late Permian were involved, and the age of thrusting and folding was probably the same as that in the Yarrol Province.

The Hunter-Bowen Orogeny was active throughout and was the main cause of development of the Bowen Basin as a major foreland basin. The rising thrust sheets in and adjacent to the New England Orogen became the dominant source of sediments for the basin, commencing in mid-Permian time and continuing until deposition ceased in the Middle Triassic (Jones & others, 1984; Fielding & others, 1997a,b; Brakel & others, 2009). The elongate Middle Triassic Abercorn Trough, situated beneath the Mulgildie Basin west of Monto, may also have been a foreland basin, although it has been interpreted as a rift basin related to a tensional phase (Whitaker & others, 1974).

PERMO-TRIASSIC MAGMATISM

Extensive magmatism in Late Permian to Late Triassic time was largely coincident with the Hunter-Bowen Orogeny, and can therefore be regarded as syntectonic, although some plutons cross-cut folded rocks. Four small batholiths and several composite plutons are present. Granodiorite is the dominant composition, but a full range from gabbros (including layered gabbros) to syenogranites is present. A few plutons are bimodal, with gabbros around the margin and granite or granodiorite cores. Mafic magmas related to subduction pooled in and partly melted the lower crust. Some of the plutons are zoned or nested, and their form and spacing is similar to that of intrusions in the Andes (Cobbing & others, 1981). Geochemically the intrusives belong to the Clarence River Supersuite of Bryant & others (1997), but it is not possible to define individual suites because every pluton is unique in some way. The peak of intrusive activity was close to the Permian-Triassic boundary. The early magmatism was related to active subduction, with a change to extensional character and A-type chemistry in the Late Triassic (Gust & others, 1993, 1996). Mineralisation associated with the intrusions includes gold-bearing veins and porphyry-style copper-gold-molybdenum, but deposits are small and currently uneconomic.

Continental volcanics are exclusively of Triassic age, and post-date the main folding event associated with the Hunter-Bowen Orogeny, being flat-lying and lacking deformation except adjacent to faults. Most show distinct subduction-related chemistry and can be matched with volcanics from the Southern Volcanic Zone of the Andes. Some are comagmatic with the plutonic rocks. Late Triassic volcanics are dominantly felsic, and show a transition from subduction-related to extensional character, similar to that displayed by the intrusions.

LATE TRIASSIC, JURASSIC AND EARLY CRETACEOUS BASINS AND MAGMATISM

The Upper Triassic Callide Coal Measures represent the preserved fill of an asymmetric basin considered by Jorgensen (1997) to be a half-graben. The main controlling structures lay along its western margin. The basal conglomeratic facies accumulated within high energy, alluvial channel and fan environments, and is overlain by alluvial plain deposits including coal. Palaeocurrent data indicate a consistent south to south-eastward direction of sediment transport throughout the history of the basin (Biggs & others, 1995; Jorgensen, 1997).

In earliest Jurassic time, the extensive but thin sand sheet of the Precipice Sandstone was deposited by a series of braided streams that flowed from south to north. Much of the quartzose sand was derived from the Permo-Triassic granites (Playford & Cornelius, 1967). The only area where younger Jurassic sediments are preserved is the Mulgildie Basin west of Monto. The eastern part of this basin overlies the Middle Triassic Cynthia beds of the Abercorn Trough. The younger strata of the Mulgildie Basin were deposited in fluviolacustrine environments, and the succession includes oolitic ironstones and the mid-Jurassic Mulgildie Coal Measures. The Mulgildie Basin was undoubtedly more extensive to the east, because its present eastern boundary is a major fault, the Mulgildie Fault, and the sequence thickens to the east (Whitaker & others, 1974). Only the northern end of this basin extends into the Yarrol project area, which is difficult to reconcile with the south to north streams that laid down the Precipice Sandstone. Either the Mulgildie Basin developed subsequently and the stream direction reversed, or the basin sequence has been completely eroded from a large area. The basin does not contain any rocks younger than mid-Jurassic, but it is not known whether this reflects non-deposition or subsequent erosion.

Magmatic activity resumed close to the Jurassic-Cretaceous boundary, when a relatively small alkaline complex, including nepheline syenite, was intruded into Permo-Triassic granites of the eastern part of the Glassford Igneous Complex.

Soon after, a marine incursion reached the Stanwell area in earliest Cretaceous (Neocomian) time. Only a thin, steeply-dipping sequence is preserved south of the Stanwell Fault, and there is no evidence for the existence of a substantial basin. The marine sequence is slightly older than the Early Cretaceous marine transgression in the Maryborough and Surat Basins (Whitehouse, 1946; Day & others, 1983), suggesting that the sea may have entered south-eastern Queensland through the Rockhampton region.

This marine incursion was followed by extensive Early Cretaceous (Aptian, about 120Ma) volcanism in the same area. Initial trachytic to rhyolitic volcanism, comprising both extrusives and intrusive plugs (Wycarbah Volcanics), was succeeded by widespread basalt flows (Dalma Basalt), and finally by a resumption of large scale eruption of trachytic and rhyolitic volcanics (Mount Salmon Volcanics). Minor sediments including thin coal seams are preserved beneath the Dalma Basalt. The volcanism occurred at the beginning of a period of intracratonic sag and extension in north-eastern Australia from 120 to 95Ma (Symonds & others, 1996), and has typical within-plate chemistry.

Still later in the Early Cretaceous (Albian), another marine incursion is recorded in the Stanwell Basin and in the Jim Crow Basin north-east of Rockhampton. Only thin deposits accumulated over a short time period. The Stanwell Basin was deltaic, and may have marked the western limit of the marine incursion.

Another episode of within-plate trachytic to rhyolitic magmatism is marked by the occurrence of the Late Cretaceous Mount Hedlow Trachyte, which forms a series of plugs in the area between Rockhampton and Yeppoon.

No subsequent deposition is recorded in the Yarrol project area until the Eocene. The environment was still extensional, but the development of rift basins was confined to the offshore Capricorn Basin and the Townsville and Queensland Troughs (Symonds & others, 1996). Widespread normal faults in the Yarrol project area are believed to be at mainly of Late Cretaceous age.

A number of lacustrine oil shale basins formed during Eocene time. They are relatively deep considering their small size. Most are half-grabens with a significant normal fault on the western side, although subsequent faulting has led to a graben structure in some cases. The faults are parallel to structural trends in the basement, and are oriented in a north-north-west direction. The basins were probably generated mainly by east-west extension, although a transtensional origin involving a limited amount of lateral movement on the bounding faults cannot be excluded. Their formation has been related to spreading in the Tasman and Coral Seas, but the basins post-date sea-floor spreading, which ended early in the Eocene. Undoubtedly the formation of these basins reflects the overall extensional environment at this time. A younger limit for the age of the basin fill is provided by upper Oligocene basalt which overlies the Biloela Basin. The Tertiary basins contain very large resources of oil shale.

Within plate continental basalts were erupted in the southern part of the Yarrol Project area in late Oligocene time (about 25Ma). All of the basalt flows show normal magnetism; in contrast, related plugs have both normal and reversed magnetism, consistent with the rapid reversals during this time period.

Recent Cainozoic deposits accumulated in an erosional environment (Jones, 2006), and reflect the present topography. They are mainly restricted to narrow zones along streams, at the base of slopes, and along the coastline.

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APPENDIX 1
GEOCHRONOLOGY

U-Pb ZIRCON DATING (SHRIMP)

(C.M. Fanning)

Rockhampton Group

MHROYCB: rhyolitic ignimbrite

Below base of overlying Youlambie Conglomerate along Youlambie Creek, Cania area, MONTO
MGA 306854 7260751
(Table 1, Figure 1)

Twenty grains were analysed from this sample. All twenty grains give a weighted mean age of 356.5±3.3Ma with a MSWD of 1.5. Removing two analyses whose error ranges fall outside this mean reduces the scatter, but does not significantly change the age. The preferred age therefore is 356.8±3.8Ma (external error, MSWD = 1.05)

Youlambie Conglomerate

MHRO970C: *rhyolitic ignimbrite*
Road cutting on Dawson Highway, 25km north-east of Biloela, BILOELA MGA 262661 7319645
(Table 2, Figure 2)

Twelve grains were analysed from this sample. One analysis has slightly higher common Pb but its age is not significantly different from the rest of the population. One grain, 4.1, is somewhat younger and has probably had Pb loss. Excluding this analysis produces a weighted mean of 300.8±3.1Ma (MSWD = 1.7). Removing two further analyses whose error ranges fall outside this mean reduces the scatter and results in an age of 300.6±2.7 Ma (internal error, MSWD = 0.8).

MHRO971(1): *rhyolitic ignimbrite*
Road cutting on Dawson Highway, 22km north-east of Biloela, BILOELA MGA 258242 7318366
(Table 3, Figure 3)

Sixteen grains were analysed from this sample. One analysis (12.1) is significantly older at ~348Ma and is presumed to be inherited. Excluding this analysis produces a weighted mean of 302.7±3.0Ma (MSWD = 1.9). Removing two further analyses whose error ranges fall outside this mean reduces the scatter and results in an age of 303.6±3.7Ma (external error, MSWD = 1.3).

MHRO971(2): *rhyolitic ignimbrite*
(Table 4, Figure 4)

Fifteen grains were analysed from this sample. One inherited grain (11.1) is very much older at ~1252Ma. Excluding this analysis produces a weighted mean of 299.5±3.3Ma (MSWD = 2.3). Removing three further outliers reduces the scatter and results in an age of 300.0±3.8Ma (internal error, MSWD = 1.3).

MHRO971(3): *rhyolitic ignimbrite*
(Table 5, Figure 5)

Fourteen grains were analysed from this sample. One analysis (14.1) is slightly older than the rest of the population at ~321Ma and is presumed to be inherited. Excluding this analysis produces a weighted mean of 303.3±3.0Ma (MSWD = 1.8). Removing two further analyses whose error ranges fall outside this mean reduces the scatter and results in an age of 303.1±3.4Ma (external error, MSWD = 0.91).

Berserker Group

Chalmers Formation

SCYL0116: ***volcanic breccia***
2km south-south-east of Mount Chalmers, ROCKHAMPTON MGA 259425 7419479
(Table 6, Figure 6)

Sample yielded abundant equant to elongate subhedral zircons mostly with bipyramidal terminations. CL images show that these are dominantly simple zoned magmatic grains. Fourteen areas were analysed. Thirteen of the 14 give a weighted mean age of 277.1±3.5Ma (MSWD = 1.3, probability 0.20). The excluded analysis has a slightly younger 206/238 age of 260Ma and the area analysed may be interpreted as having lost radiogenic Pb.

Ellrott Rhyolite

SCYL0030: ***dark grey volcanic breccia***
1km south-west of Tungamull, ROCKHAMPTON MGA 263094 7419384
(Table 7, Figure 7)

Sample contained few zircons, of which 12 were analysed. Most are euhedral zoned crystals or are fragments of such grains. Grain 1 is clearly older at ~305Ma. Ten of the remaining 11 grains show slight dispersion about a weighted mean age of 268.2±3.9Ma (MSWD = 1.7, probability 0.072).

SCYL0081: ***dark grey massive porphyritic dacite or rhyolite***
400m north of Cabbage Tree Hill, ROCKHAMPTON MGA 255617 7421202
(Table 8, Figure 8)

Elongate to lozenge shaped zircons with pyramidal terminations or fragments of such grains were separated. A significant number of grains have tube-like internal cavities consistent with magmatic crystallisation in a surface or near-surface environment. The CL images show dominantly simple zoned magmatic internal structures. Fifteen grains were analysed, and a weighted mean of all analyses has excess scatter (MSWD = 3.1) about an age of 278Ma. If the oldest analysis is excluded (~290Ma), the weighted mean age is 276.0±3.9Ma (MSWD = 2.1, probability 0.012), which is the best estimate for the magmatic crystallisation age.

Mount Morgan Trondhjemite

CMM239: ***hornblende trondhjemite***
Gully near Nine Mile Creek south-east of Mount Morgan, MOUNT MORGAN MGA 236922 7376184
(Table 9, Figure 9)

Fifteen grains were analysed. They form a simple population of magmatic zircon which has a weighted mean $^{206}Pb/^{238}U$ age of 379.9±4.27Ma (MSWD = 0.407, probability 0.973).

Pomegranate Tonalite

PBRO528: ***grey tonalite***
2km south-south-west of Pomegranate homestead, BAJOOL MGA 259425 7419479
(Table 10, Figure 10)

Fifteen grains were analysed. One analysis is markedly discordant, and can be discarded. The remaining 14 have a weighted mean of 369.0±4.2Ma (MSWD = 0.732, probability 0.733), and this is interpreted as the crystallisation age.

Kyle Mohr Igneous Complex

CMM462B: *leucocratic hornblende-biotite granite*
Track up ridge 15km south-east of Old Westwood, MOUNT MORGAN MGA 219514 7373395
(Table 11, Figure 11)

Beautiful simple population of euhedral doubly terminated zircon crystals. CL images show simple magmatic zoning. Fifteen grains were analysed. One area is more enriched in common Pb and yields a younger 206/238 age around 260Ma. The remaining 14 analyses have no excess scatter, giving an age of 270.6±3.7Ma (MSWD = 0.92, probability 0.53).

Craiglands Quartz Monzodiorite

CMM106B: *biotite-hornblende diorite*
Side of ridge to west of track 6km north-west of Craiglands homestead, BANANA MGA 245123 7331852
(Table 12, Figure 12)

Fifteen grains were analysed, and gave a mean age of 257.0±5.3Ma with an unacceptably large scatter (MSWD = 4.6). Considering only 10 of the results gives an age that is essentially the same (256.8±2.6Ma) with no excess scatter (MSWD = 2.6, probability 0.56).

Mount Seaview Igneous Complex

CMM245: *leucocratic biotite granite*
Ridge of Griffiths Hill north of The Tableland, BILOELA MGA 290564 7312679
(Table 13, Figure 13)

The sample yielded a small number of mostly elongate euhedral zircons with some equant to squat grains also present. CL images show a relatively simple magmatic internal structure. Sixteen grains were analysed, and a weighted mean of all 16 has slight scatter about a mean age of 252.6±5.3Ma (MSWD = 1.8, probability 0.029). The analyses can be divided into an older group of 10 (258.0±4.3Ma, MSWD = 0.37, probability 0.95) and a younger group of 6 (238.3±6.9Ma, MSWD = 0.17, probability 0.97), suggesting either that they come from intrusives of different age, or that some of the areas have lost radiogenic Pb. There is no correspondence between the squat versus elongate grain shapes and the older and younger sub-groups. The overall weighted mean at 252.6±5.3Ma is the current best estimate of the age of the sample.

Galloway Plains Igneous Complex

Redshirt Granite

CMM145A: *hornblende-biotite granite*
Railway cutting 1.7km south-east of Mount Redshirt, BILOELA MGA 271131 7333522
(Table 14, Figure 14)

The sample gave a relatively simple population of elongate zircons with pyramidal terminations, or fragments of such grains. A few blocky grains may represent inherited components. Sixteen analyses were carried out on 14 grains, but 3 analyses are not presented as there were machine problems during data collection. The 13 acceptable analyses have no excess scatter about a weighted mean 206/238 age of 251±4Ma (MSWD = 0.71, probability 0.74), and this is interpreted as the magmatic age of the sample.

Voewood Granite

CMM242: *hornblende-biotite granite*
South side of hill east of Voewood homestead, BILOELA MGA 280440 7330806
(Table 15, Figure 15)

Slender, elongate zircons were separated from this rock. Most have pyramidal terminations and some have internal cavities consistent with magmatic crystallisation in a near-surface environment. Fifteen grains were analysed, but gave no unique age. Rather, there is a continuum of 206/238 ages ranging from around 250Ma to 215Ma. The cumulative probability plot shows two general groupings, a dominant one at about 240Ma and a subordinate group at about 215Ma. This result is not consistent with the zircon morphology and CL images show a dominantly simple magmatic paragenesis. It is possible that the rock has inherited zircons from a slightly older magmatic source. The weighted mean of the 15 analyses (233.6±6.8Ma; MSWD = 6.3, probability 0.000) is taken as the magmatic age.

Glassford Igneous Complex

Robert Granite

CMM418: *porphyritic hornblende monzonite*
Track to Mount Robert 10km west of Many Peaks, MONTO MGA 323113 7285923
(Table 16, Figure 16)

Thirteen zircon grains were analysed. These gave a weighted mean age of 227.46±3.24Ma (MSWD = 1.06, probability 0.389). The scatter can be significantly decreased by discarding 3 younger ages, with a resultant weighted mean age of 230.2±3.7Ma (MSWD = 0.258, probability 0.985), and this is interpreted as the magmatic age.

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K-AR DATING

(A.W. Webb)

Pomegranate Tonalite

PBRO528: hornblende tonalite

Sample	%K#	^{40}Ar*(x10^{-10}moles/g)	^{40}Ar*/^{40}Ar$_{Total}$	Age†
PBRO528 hornblende	0.221	1.7887	0.905	415±3
GR259425 7419479	0.221			
PBRO528 hornblende	0.221	1.8010	0.920	417±3

* Denotes radiogenic ^{40}Ar
† Age in Ma with error limits given for the analytical uncertainty at one standard deviation
# The mean K value is used in the age calculation
Constants:
^{40}K = 0.01167 atom%
$ë_{â}$ = 4.962 x $10^{-10}y^{-1}$
$ë_{å}$ = 0.581 x $10^{-10}y^{-1}$

RB-SR AND TIMS ZIRCON DATING

(C.G. Murray)

Rb-Sr data

Sample	Rb, ppm	Sr, ppm	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr \pm 2σ$
Native Cat Andesite				
ADR7A (andesite) GR216288 7400698	25.05	620.2	0.1168	0.70391±2
ADR9 (andesite) GR216359 7401185	14.64	717.2	0.0590	0.70347±6
ADR8 (rhyolite) GR216350 7400894	148.0	101.6	4.214	0.71477±6
Muncon Volcanics				
ADR264 (basaltic andesite) GR291264 7271835	26.36	772.1	0.0987	0.70388±8
Winterbourne Volcanics				
ADR199 (rhyolite) GR291028 7304757	81.37	21.73	10.834	0.73618±14
Wycarbah Volcanics				
ADR11 (rhyolite) GR260471 7307125	228.28	7.622	86.642	0.84431±13
Dalma Basalt				
ADR501 (basalt) GR207050 7395222	14.87	743.9	0.0578	0.70306±2
Mount Salmon Volcanics				
ADR387 (rhyolite) GR213720 7402684	133.3	34.80	11.086	0.71688±5
Tertiary basalt				
ADR4 GR285022 7283232	30.88	597.1	0.1496	0.70453±3
Ridgelands Granodiorite				
MHRO1024 (whole rock) GR211099 7432373	44.15	384.3	0.3323	0.70472±5
MHRO1024 (hornblende)	6.237	53.49	0.3118	0.70463±8
MHRO1024 (biotite)	337.0	9.048	107.8	1.09116±14
CMM238 (whole rock) GR204859 7428436	56.47	253.6	0.6454	0.70598±5
CMM238 (plagioclase)	56.17	194.7	0.835	0.70664±9
CMM238 (biotite)	330.8	12.13	78.90	0.98400±20
Unnamed granodiorite				
Kalapa North 1 (whole rock) GR210623 7399910	67.65	376.4	0.5199	0.70497±5
Kalapa North 2 (whole rock) GR210255 7399674	25.17	665.1	0.1095	0.70372±4

Only the Ridgelands Granodiorite gives an acceptable Rb-Sr date, which is 251±2Ma with an initial ratio of 0.70359±5 (Figure 17).

Approximate ages can be calculated for the Winterbourne Volcanics and Wycarbah Volcanics, by using basalts of similar age. The combination of rhyolite from the Winterbourne Volcanics and basaltic andesite from the Muncon Volcanics gives 213Ma, which is consistent with the assumed Late Triassic age of the former unit (Figure 18). A single analysis from the Mount Sugarloaf plug in the Wycarbah Volcanics gives an age of 113Ma (Early Cretaceous) when combined with the result from a nearby sample of Dalma Basalt (Figure 19).

U-Pb data

Sample	Weight (mg)	^{238}U (ppm)	$^{206}Pb^{1}$ (ppm)	$^{206}Pb/^{204}Pb$ (meas)	$Pb_c{}^{2}$ (ppm)	$^{206}Pb/^{238}U$ ratio[3] (±%error)[4]	Age (Ma)	$^{207}Pb/^{235}U$ ratio[3] (±%error)[4]	Age (Ma)
Winterbourne Volcanics									
ADR170 (white) GR293989 7280609	1.5	218.5	6.419	3393	0.052	0.03394 (±0.324)	215.1±0.7	0.23664 (±0.355)	215.7±0.7
ADR170 (amber)	1.5	207.5	6.325	1262	0.051	0.03522 (±0.491)	223.2±1.1	0.24691 (±655)	224.1±1.3
Wycarbah Volcanics									
ADRMM53 (white) GR208140 7391942	5.52	233.7	3.806	6357	0.017	0.01882 (±0.277)	120.2±0.3	0.12667 (±0.327)	121.1±0.4
ADRMM53 (amber)	9.56	294.1	4.321	8039	0.017	0.01697 (±0.279)	108.5±0.3	0.11362 (±0.310)	109.3±0.3
ADR3 GR209483 7391267	3.21	103.0	1.788	1197	0.047	0.02006 (±0.337)	128.0±0.4	0.13555 (±0.479)	129.1±0.6
Sebastopol (plug south-south-west of Westwood)	4.97	194.0	6.309	5944	0.034	0.03757 (±0.302)	237.8±0.7	0.26696 (±0.432)	240.3±0.9
Mount Hedlow Trachyte									
Hedlow 48	2.40	220.3	5.030	2331	0.069	0.02638 (±0.279)	167.8±0.5	0.18190 (±0.350)	169.7±0.6

Notes: [1] radiogenic ^{206}Pb
[2] common Pb corrected for fractionation, spike and blank Pb
[3] measured ratio corrected for fractionation and spike contribution
[4] the total per cent error found by summing the error contribution to the computed ratio by all measured errors (one standard error of the mean).

Common lead correction made using Stacey-Kramers lead of the appropriate age. The total Pb blank varied from 14 to 19pg.

The sample from the Winterbourne Volcanics was taken from a dyke in a tributary of South Kariboe Creek 1.5km from the outcrop of the volcanics. Both ages are Late Triassic, as is the approximate Rb-Sr determination detailed above.

Zircon from the Mount Sebastopol plug of the Wycarbah Volcanics is presumably a xenocryst from a Permo-Triassic granitoid. Dates from two other samples, including both a volcanic breccia (ADRMM53) and the Sugarloaf Mountain plug (ADR3) are Early Cretaceous, although the age from amber zircon in sample ADRMM53 is too young.

The date obtained from zircon extracted from a sample of Mount Hedlow Trachyte cannot be readily explained, in view of the evidence for a Late Cretaceous age and the lack of Jurassic magmatism in the region.

Table 1. Summary of SHRIMP U-Pb zircon analytical data for sample MHROYCB

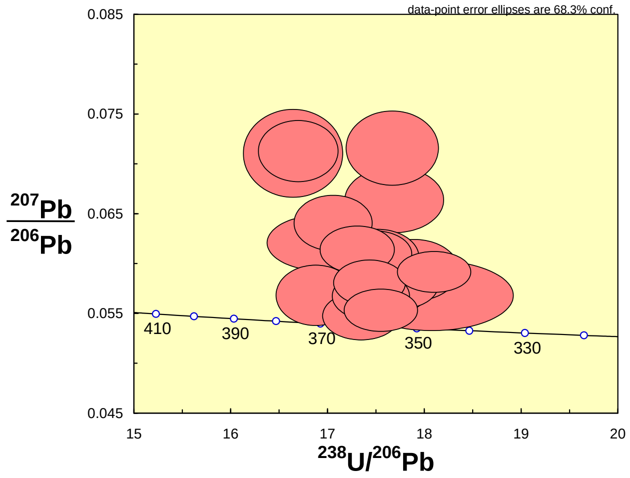
Rockhampton Group (below basal Youlambie Conglomerate): ignimbrite, Cania area

							Total				Radiogenic		Age (Ma)	
Grain. spot	**U (ppm)**	**Th (ppm)**	**Th/U**	**$^{206}Pb^*$ (ppm)**	**$^{204}Pb/^{206}Pb$**	**f_{206} %**	**$^{238}U/^{206}Pb$**	±	**$^{207}Pb/^{206}Pb$**	±	**$^{206}Pb/^{238}U$**	±	**$^{206}Pb/^{238}U$**	±
1.1	87	41	0.47	4.1	0.000588	0.42	18.073	0.555	0.0568	0.0023	0.0551	0.0017	345.8	10.5
2.1	157	82	0.53	7.7	0.000499	0.38	17.443	0.262	0.0567	0.0016	0.0571	0.0009	358.1	5.3
3.1	109	48	0.44	5.3	0.000890	0.52	17.618	0.294	0.0578	0.0020	0.0565	0.0010	354.1	5.9
4.1	125	58	0.46	6.4	0.000020	0.35	16.874	0.269	0.0568	0.0020	0.0591	0.0010	369.9	5.9
5.1	178	95	0.53	8.9	0.000123	0.95	17.303	0.252	0.0614	0.0016	0.0572	0.0009	358.8	5.2
6.1	166	82	0.50	8.2	0.000866	0.91	17.481	0.256	0.0609	0.0015	0.0567	0.0008	355.4	5.2
7.1	149	73	0.49	7.4	0.000993	0.12	17.343	0.263	0.0547	0.0016	0.0576	0.0009	360.9	5.4
8.1	162	70	0.43	8.0	-	0.72	17.382	0.272	0.0595	0.0041	0.0571	0.0010	358.1	5.8
9.1	256	136	0.53	12.2	0.000579	0.72	18.097	0.249	0.0591	0.0014	0.0549	0.0008	344.3	4.7
10.1	59	22	0.38	3.0	0.001523	2.12	16.640	0.341	0.0710	0.0029	0.0588	0.0012	368.5	7.6
11.1	135	68	0.51	6.8	-	1.27	17.052	0.268	0.0640	0.0018	0.0579	0.0009	362.8	5.7
12.1	122	53	0.43	6.0	0.001367	0.86	17.511	0.284	0.0606	0.0019	0.0566	0.0009	355.0	5.7
13.1	97	45	0.46	4.7	0.000221	2.25	17.662	0.315	0.0716	0.0025	0.0553	0.0010	347.2	6.3
14.1	133	84	0.63	6.8	0.000478	2.15	16.693	0.272	0.0713	0.0020	0.0586	0.0010	367.2	6.0
15.1	124	52	0.42	6.0	0.000086	0.73	17.892	0.303	0.0593	0.0020	0.0555	0.0010	348.1	5.9
16.1	124	60	0.49	6.3	-	1.01	16.917	0.360	0.0621	0.0018	0.0585	0.0013	366.6	7.7
17.1	118	52	0.44	5.7	0.001065	1.60	17.687	0.337	0.0664	0.0022	0.0556	0.0011	349.0	6.6
18.1	140	69	0.49	6.8	0.000101	0.54	17.738	0.266	0.0579	0.0016	0.0561	0.0009	351.7	5.2
19.1	174	86	0.49	8.6	0.001075	0.55	17.429	0.245	0.0581	0.0015	0.0571	0.0008	357.7	5.0
20.1	158	69	0.44	7.8	0.000582	0.20	17.547	0.249	0.0553	0.0014	0.0569	0.0008	356.6	5.0

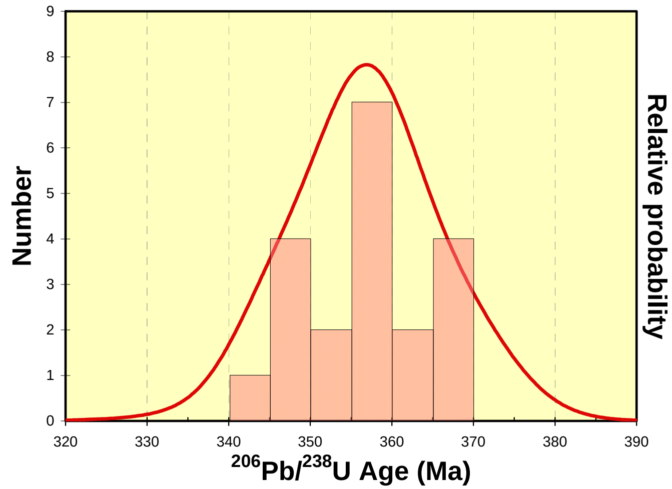
Notes:
1. Uncertainties given at the one ó level.
2. Error in AS3 reference zircon calibration was 0.75% for the analytical session (not included in above errors but required when comparing data from different mounts).
3. f_{206} % denotes the percentage of ^{206}Pb that is common Pb.
4. Correction for common Pb made using the measured $^{238}U/^{206}Pb$ and $^{207}Pb/^{206}Pb$ ratios following Tera & Wasserburg (1972) as outlined in Williams (1998).

Age	**± no std**	**± include std**
356.8	2.7	**3.8**

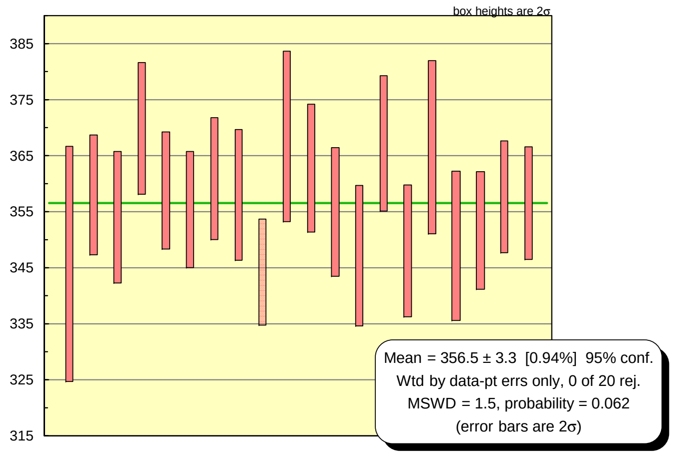
Figure 1. Plots of SHRIMP U-Pb zircon data for sample MHROYCB



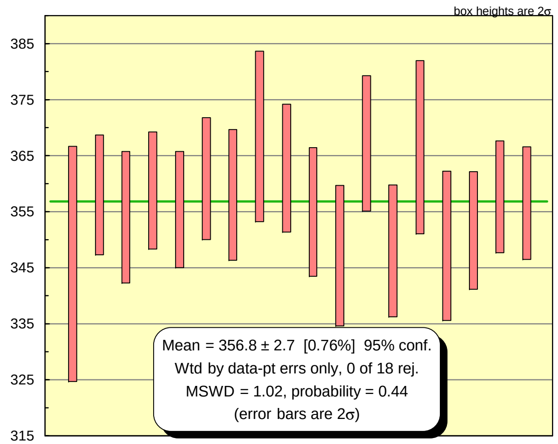
(a) Tera-Wasserburg plot, all analyses



(b) Probability density plot – all grains



(c) Weighted average – all grains



(d) Weighted average – best estimate

Table 2. Summary of SHRIMP U-Pb zircon analytical data for sample MHRO970C

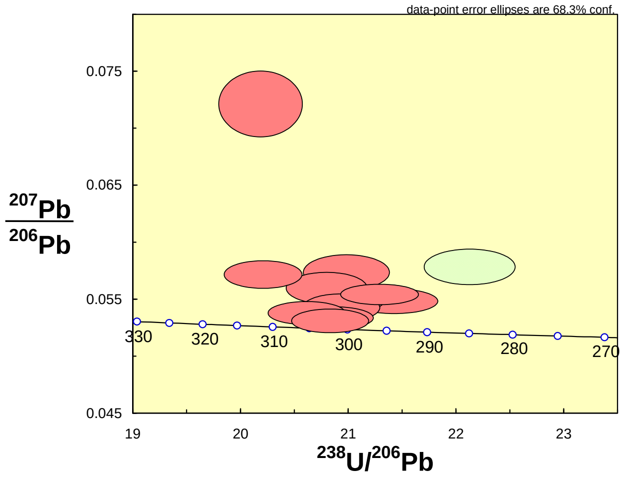
Youlambie Conglomerate: rhyolitic ignimbrite. Road cutting on Dawson Highway, 25km north-east of Biloela, BILOELA, MGA 262661 7319645

							Total				Radiogenic		Age (Ma)	
Grain. spot	**U (ppm)**	**Th (ppm)**	**Th/U**	**$^{206}Pb^*$ (ppm)**	**$^{204}Pb/^{206}Pb$**	**f_{206} %**	**$^{238}U/^{206}Pb$**	±	**$^{207}Pb/^{206}Pb$**	±	**$^{206}Pb/^{238}U$**	±	**$^{206}Pb/^{238}U$**	±
1.1	372	239	0.64	15.2	0.000289	0.33	20.956	0.252	0.0550	0.0008	0.0476	0.0006	299.5	3.6
2.1	585	644	1.10	24.0	0.000165	0.24	20.930	0.237	0.0542	0.0008	0.0477	0.0005	300.2	3.4
3.1	754	353	0.47	30.4	0.000306	0.40	21.287	0.239	0.0554	0.0006	0.0468	0.0005	294.8	3.3
4.1	329	212	0.64	12.8	0.000551	0.73	22.123	0.279	0.0578	0.0010	0.0449	0.0006	283.0	3.5
5.1	279	203	0.73	11.4	0.000472	0.63	20.978	0.264	0.0574	0.0010	0.0474	0.0006	298.3	3.7
6.1	615	292	0.47	25.6	0.000206	0.17	20.609	0.237	0.0538	0.0007	0.0484	0.0006	304.9	3.5
7.1	600	324	0.54	23.8	0.000134	0.33	21.419	0.268	0.0548	0.0007	0.0465	0.0006	293.2	3.6
8.1	459	257	0.56	19.5	0.000588	0.57	20.201	0.238	0.0572	0.0008	0.0492	0.0006	309.7	3.6
9.1	613	304	0.50	25.3	-	0.12	20.867	0.238	0.0533	0.0006	0.0479	0.0006	301.4	3.4
10.1	324	179	0.55	13.4	0.000288	0.45	20.797	0.248	0.0560	0.0009	0.0479	0.0006	301.4	3.6
11.1	477	240	0.50	20.3	0.001687	2.45	20.181	0.256	0.0721	0.0019	0.0483	0.0006	304.3	4.0
12.1	528	240	0.45	21.8	0.000081	0.09	20.826	0.236	0.0531	0.0007	0.0480	0.0005	302.1	3.4

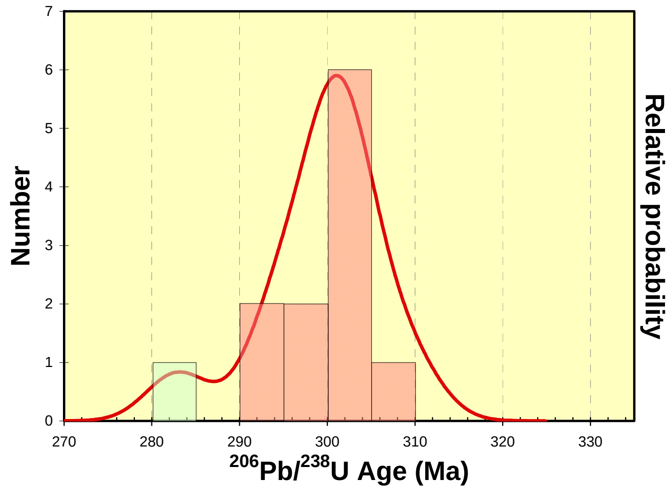
Notes:
1. Uncertainties given at the one ó level.
2. Error in AS3 reference zircon calibration was 0.48% for the analytical session (not included in above errors but required when comparing data from different mounts).
3. f_{206} % denotes the percentage of ^{206}Pb that is common Pb.
4. Correction for common Pb made using the measured $^{238}U/^{206}Pb$ and $^{207}Pb/^{206}Pb$ ratios following Tera & Wasserburg (1972) as outlined in Williams (1998).

	Age	**± no std**	**± include std**
most	**300.8**	3.1	**3.4**
with rejects	**300.6**	2.3	**2.7**

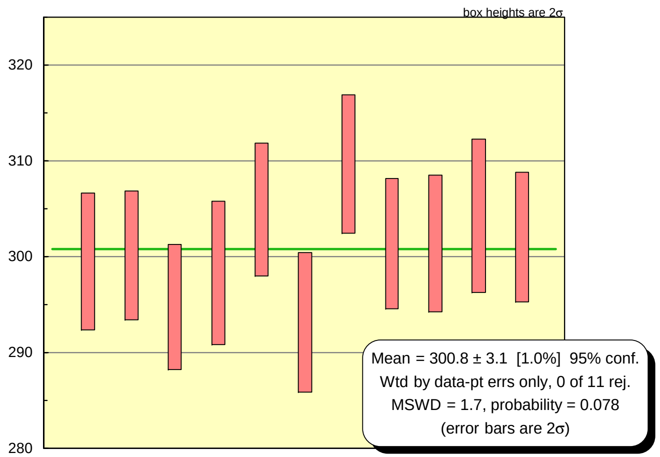
Figure 2. Plots of SHRIMP U-Pb zircon data for sample MHRO970C



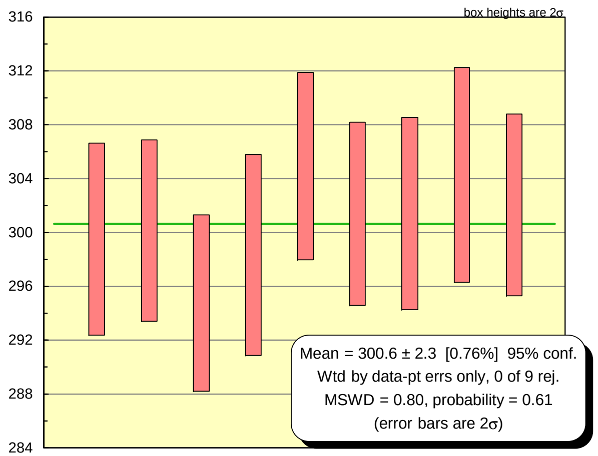
(a) Tera-Wasserburg plot, all analyses



(b) Probability density plot – all grains



(c) Weighted average – all grains



(d) Weighted average – best estimate

Table 3. Summary of SHRIMP U-Pb zircon analytical data for sample MHRO971(1)

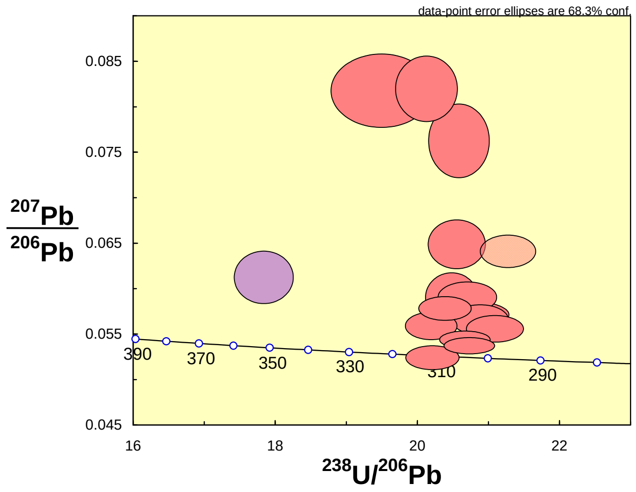
Youlambie Conglomerate: rhyolitic ignimbrite. Road cutting on Dawson Highway, 22km north-east of Biloela, BILOELA, MGA 258242 7318366

							Total				Radiogenic		Age (Ma)	
Grain. spot	**U (ppm)**	**Th (ppm)**	**Th/U**	**^{206}Pb* (ppm)**	**^{204}Pb/ ^{206}Pb**	**f_{206} %**	**^{238}U/ ^{206}Pb**	±	**^{207}Pb/ ^{206}Pb**	±	**^{206}Pb/ ^{238}U**	±	**^{206}Pb/ ^{238}U**	±
1.1	341	182	0.53	14.1	0.000681	0.83	20.700	0.272	0.0590	0.0011	0.0479	0.0006	301.7	3.9
2.1	450	273	0.61	18.6	0.000101	0.59	20.774	0.338	0.0571	0.0009	0.0479	0.0008	301.3	4.9
3.1	667	487	0.73	26.9	0.001004	1.49	21.270	0.259	0.0641	0.0012	0.0463	0.0006	291.9	3.6
4.1	469	228	0.49	19.3	0.000243	0.53	20.882	0.262	0.0566	0.0011	0.0476	0.0006	300.0	3.7
5.1	327	170	0.52	13.7	0.001744	2.98	20.579	0.280	0.0762	0.0027	0.0471	0.0007	297.0	4.3
6.1	868	339	0.39	36.4	0.000681	0.82	20.480	0.241	0.0590	0.0018	0.0484	0.0006	304.9	3.6
7.1	91	62	0.69	4.0	0.001809	3.63	19.490	0.471	0.0818	0.0027	0.0494	0.0012	311.1	7.6
8.1	374	194	0.52	16.0	0.001946	3.68	20.123	0.287	0.0819	0.0024	0.0479	0.0007	301.4	4.6
9.1	581	285	0.49	24.5	0.000625	0.66	20.382	0.244	0.0578	0.0009	0.0487	0.0006	306.8	3.6
10.1	735	693	0.94	31.3	0.000071	0.41	20.189	0.238	0.0559	0.0010	0.0493	0.0006	310.4	3.6
11.1	336	161	0.48	14.1	0.000188	1.55	20.548	0.266	0.0649	0.0018	0.0479	0.0006	301.7	3.9
12.1	1057	453	0.43	50.9	-	0.97	17.830	0.275	0.0613	0.0019	0.0555	0.0009	348.5	5.4
13.1	735	330	0.45	30.5	0.000081	0.16	20.727	0.234	0.0537	0.0006	0.0482	0.0005	303.3	3.4
14.1	269	141	0.52	10.9	0.000698	0.41	21.085	0.266	0.0556	0.0010	0.0472	0.0006	297.5	3.7
15.1	624	283	0.45	25.9	0.000002	0.25	20.664	0.234	0.0544	0.0006	0.0483	0.0006	303.9	3.4
16.1	303	182	0.60	12.9	0.000360	<0.01	20.205	0.247	0.0524	0.0009	0.0495	0.0006	311.5	3.8

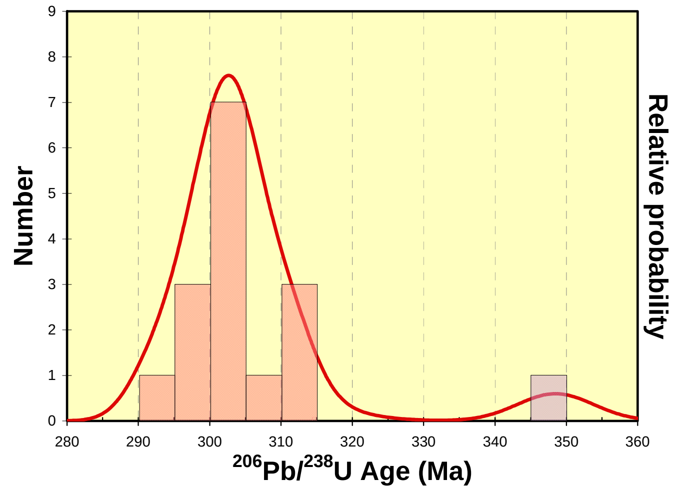
Notes:
1. Uncertainties given at the one ó level.
2. Error in AS3 reference zircon calibration was 0.71% & 0.48% for the two analytical sessions (not included in above errors but required when comparing data from different mounts).
3. f_{206} % denotes the percentage of ^{206}Pb that is common Pb.
4. Correction for common Pb made using the measured ^{238}U/^{206}Pb and ^{207}Pb/^{206}Pb ratios following Tera & Wasserburg (1972) as outlined in Williams (1998).

Age	**± no std**	**± include std**
303.6	2.6	**3.7**

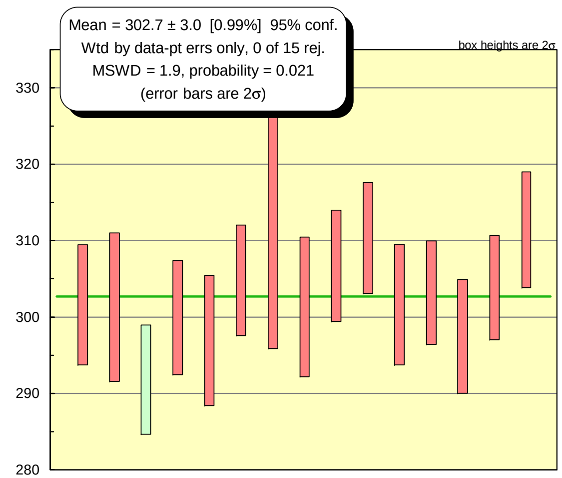
Figure 3. Plots of SHRIMP U-Pb zircon data for sample MHRO971(1)



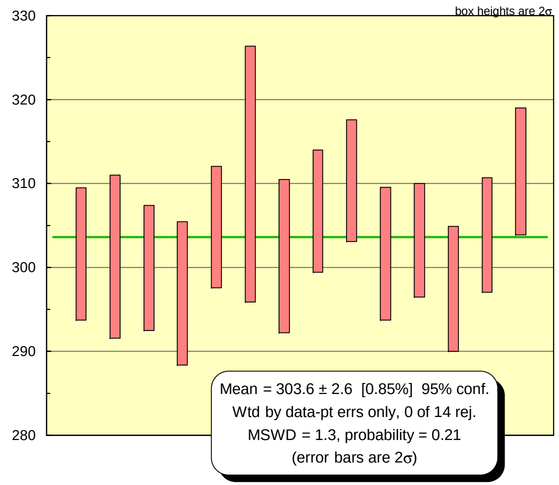
(a) Tera-Wasserburg plot, all analyses



(b) Probability density plot – all grains



(c) Weighted average – most grains



(d) Weighted average – best estimate

Table 4. Summary of SHRIMP U-Pb zircon analytical data for sample MHRO971(2)

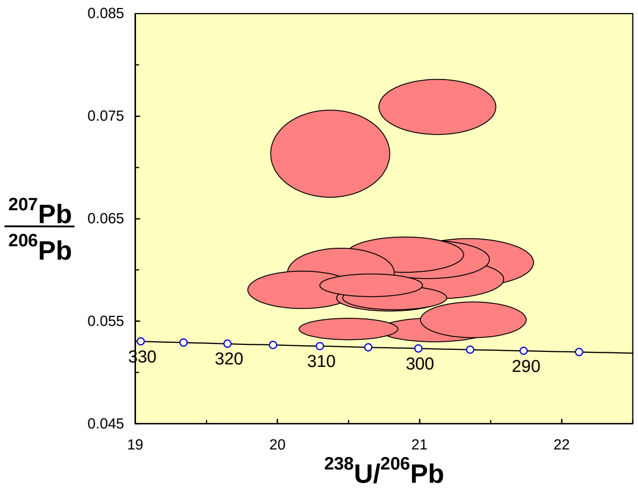
Youlambie Conglomerate: rhyolitic ignimbrite. Road cutting on Dawson Highway, 22km northeast of Biloela, BILOELA, MGA 258242 7318366

							Total				**Radiogenic**		**Age (Ma)**	
Grain. spot	**U (ppm)**	**Th (ppm)**	**Th/U**	**^{206}Pb* (ppm)**	**^{204}Pb/ ^{206}Pb**	**f_{206} %**	**^{238}U/ ^{206}Pb**	±	**^{207}Pb/ ^{206}Pb**	±	**^{206}Pb/ ^{238}U**	±	**^{206}Pb/ ^{238}U**	±
1.1	675	247	0.37	28.0	0.000289	0.61	20.821	0.243	0.0572	0.0008	0.0477	0.0006	300.6	3.5
2.1	382	267	0.70	15.5	0.002189	2.96	21.123	0.273	0.0759	0.0018	0.0459	0.0006	289.5	3.9
3.1	350	118	0.34	14.2	0.000985	0.85	21.130	0.304	0.0591	0.0013	0.0469	0.0007	295.6	4.2
4.1	358	168	0.47	14.6	0.001462	1.09	21.056	0.286	0.0610	0.0012	0.0470	0.0007	295.9	4.0
5.1	420	264	0.63	17.3	0.001083	1.15	20.894	0.274	0.0615	0.0011	0.0473	0.0006	298.0	3.9
6.1	654	300	0.46	27.5	0.000728	0.90	20.443	0.248	0.0597	0.0016	0.0485	0.0006	305.2	3.7
7.1	521	335	0.64	22.2	0.000386	0.68	20.170	0.254	0.0580	0.0012	0.0492	0.0006	309.9	3.9
8.1	249	169	0.68	10.0	0.000380	1.07	21.341	0.304	0.0607	0.0015	0.0464	0.0007	292.1	4.2
9.1	279	207	0.74	11.8	0.000492	2.36	20.367	0.277	0.0713	0.0028	0.0479	0.0007	301.9	4.3
10.1	783	336	0.43	32.5	-	0.76	20.657	0.238	0.0585	0.0007	0.0480	0.0006	302.5	3.5
11.1	371	43	0.12	71.7	0.000005	4.77	4.440	0.053	0.1227	0.0018	0.2145	0.0031	1252.5	16.3
12.1	609	433	0.71	25.2	-	0.61	20.787	0.246	0.0572	0.0008	0.0478	0.0006	301.1	3.5
13.1	403	230	0.57	16.4	0.000031	0.23	21.101	0.254	0.0541	0.0008	0.0473	0.0006	297.8	3.6
14.1	513	302	0.59	20.6	0.000511	0.37	21.376	0.246	0.0551	0.0012	0.0466	0.0005	293.7	3.4
15.1	660	249	0.38	27.7	0.000400	0.22	20.494	0.230	0.0542	0.0007	0.0487	0.0006	306.5	3.4

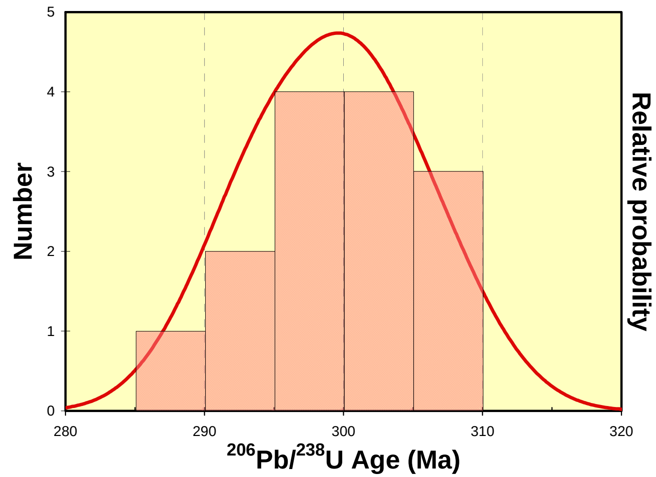
Notes:
1. Uncertainties given at the one ó level.
2. Error in AS3 reference zircon calibration was 0.71% & 0.48% for the two analytical sessions (not included in above errors but required when comparing data from different mounts).
3. f_{206} % denotes the percentage of ^{206}Pb that is common Pb.
4. Correction for common Pb made using the measured ^{238}U/^{206}Pb and ^{207}Pb/^{206}Pb ratios following Tera & Wasserburg (1972) as outlined in Williams (1998).

Age	**± no std**	**± include std**
300.0	2.8	**3.8**

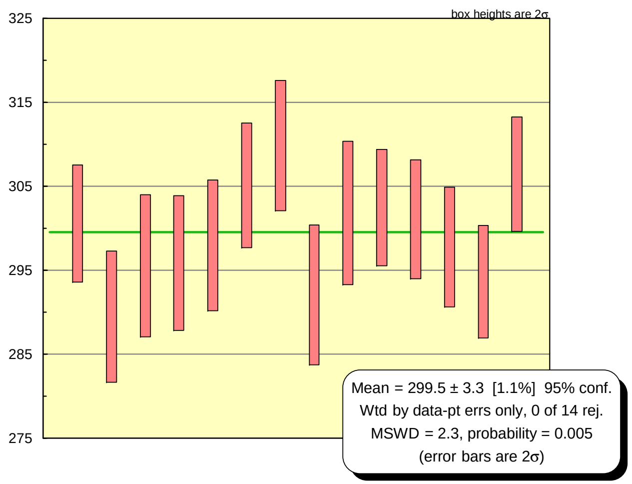
Figure 4. Plots of SHRIMP U-Pb zircon data for sample MHRO971(2)



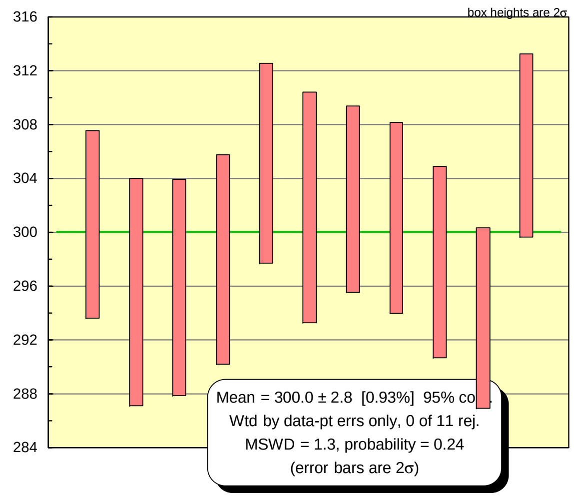
(a) Tera-Wasserburg plot, magmatic grains



(b) Probability density plot – all grains



(c) Weighted average – magmatic grains



(d) Weighted average – best estimate

Table 5. Summary of SHRIMP U-Pb zircon analytical data for sample MHRO971(3)

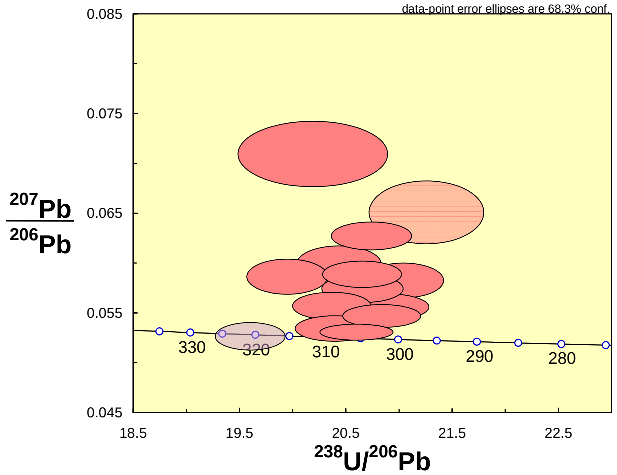
Youlambie Conglomerate: rhyolitic ignimbrite. Road cutting on Dawson Highway, 22km north-east of Biloela, BILOELA, MGA 258242 7318366

							Total				Radiogenic		Age (Ma)	
Grain. spot	U (ppm)	Th (ppm)	Th/U	^{206}Pb* (ppm)	^{204}Pb/ ^{206}Pb	f_{206} %	^{238}U/ ^{206}Pb	±	^{207}Pb/ ^{206}Pb	±	^{206}Pb/ ^{238}U	±	^{206}Pb/ ^{238}U	±
1.1	404	167	0.41	16.3	0.000354	1.61	21.253	0.356	0.0651	0.0021	0.0463	0.0008	291.7	4.9
2.1	651	363	0.56	26.8	0.000406	0.40	20.885	0.259	0.0556	0.0009	0.0477	0.0006	300.3	3.7
3.1	497	301	0.60	21.4	0.000433	0.74	19.941	0.251	0.0586	0.0012	0.0498	0.0006	313.1	3.9
4.1	678	357	0.53	28.6	0.000453	0.39	20.362	0.244	0.0557	0.0009	0.0489	0.0006	307.9	3.7
5.1	146	110	0.75	6.2	0.002596	2.30	20.185	0.464	0.0709	0.0022	0.0484	0.0011	304.7	7.0
6.1	491	278	0.57	20.4	0.000354	0.62	20.650	0.251	0.0574	0.0009	0.0481	0.0006	303.0	3.7
7.1	743	295	0.40	30.6	-	0.29	20.832	0.241	0.0547	0.0007	0.0479	0.0006	301.4	3.4
8.1	571	212	0.37	23.3	0.000148	0.74	21.037	0.250	0.0582	0.0012	0.0472	0.0006	297.2	3.5
9.1	313	230	0.73	13.2	0.000073	0.93	20.431	0.262	0.0599	0.0012	0.0485	0.0006	305.2	3.9
10.1	546	275	0.50	22.7	0.000045	0.81	20.648	0.245	0.0589	0.0009	0.0480	0.0006	302.5	3.6
11.1	321	199	0.62	13.3	0.000411	1.29	20.738	0.249	0.0627	0.0009	0.0476	0.0006	299.8	3.6
12.1	814	342	0.42	33.9	0.000040	0.07	20.594	0.227	0.0531	0.0005	0.0485	0.0005	305.4	3.3
13.1	308	123	0.40	13.0	0.000347	0.11	20.389	0.246	0.0534	0.0008	0.0490	0.0006	308.3	3.7
14.1	835	346	0.41	36.6	0.000121	<0.01	19.590	0.216	0.0526	0.0009	0.0511	0.0006	321.0	3.5

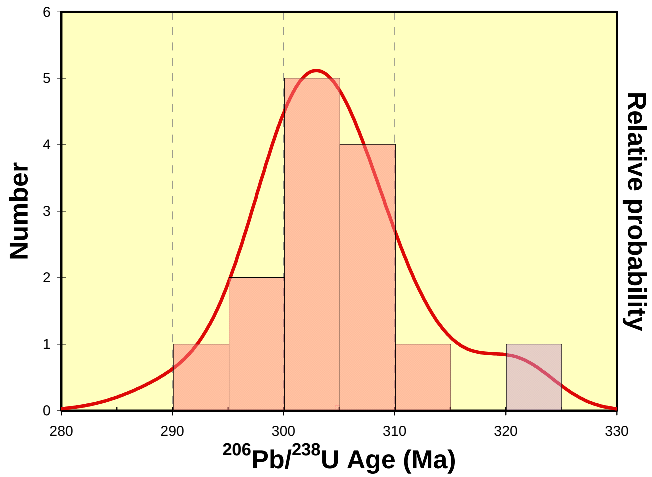
Notes:
1. Uncertainties given at the one ó level.
2. Error in AS3 reference zircon calibration was 0.71% & 0.48% for the two analytical sessions (not included in above errors but required when comparing data from different mounts).
3. f_{206} % denotes the percentage of ^{206}Pb that is common Pb.
4. Correction for common Pb made using the measured ^{238}U/^{206}Pb and ^{207}Pb/^{206}Pb ratios following Tera & Wasserburg (1972) as outlined in Williams (1998).

Age	± no std	± include std
303.1	2.2	**3.4**

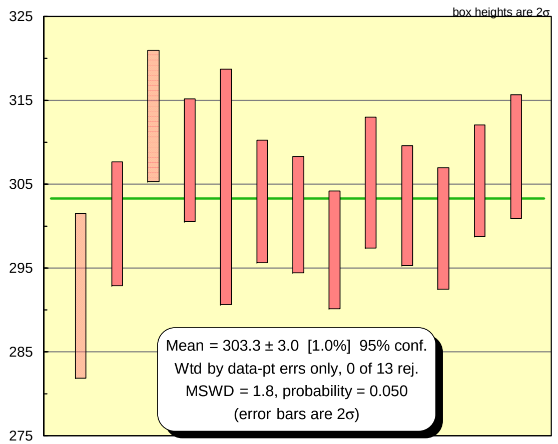
Figure 5. Plots of SHRIMP U-Pb zircon data for sample MHRO971(3)



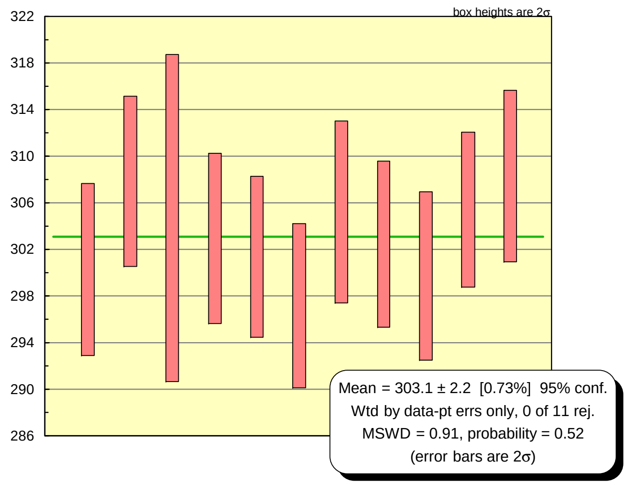
(a) Tera-Wasserburg plot, all analyses



(b) Probability density plot – all grains



(c) Weighted average – all grains



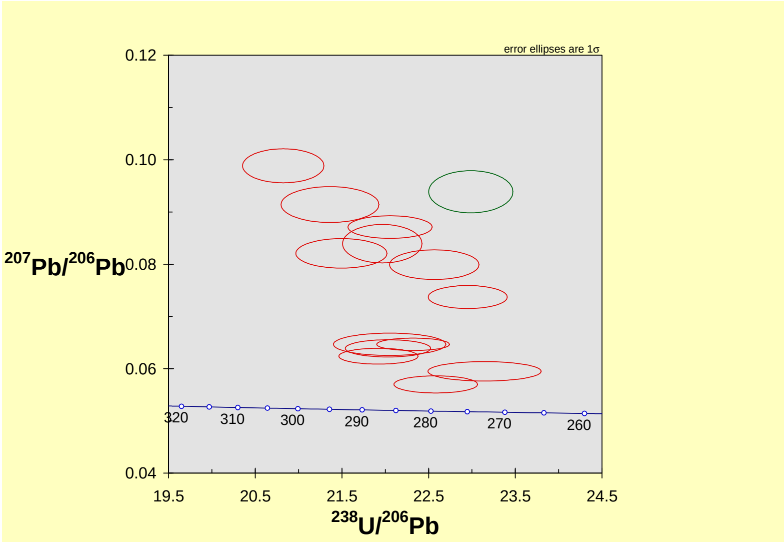
(d) Weighted average – best estimate

Table 6. Summary of SHRIMP U-Pb zircon results for sample SCYL0116.

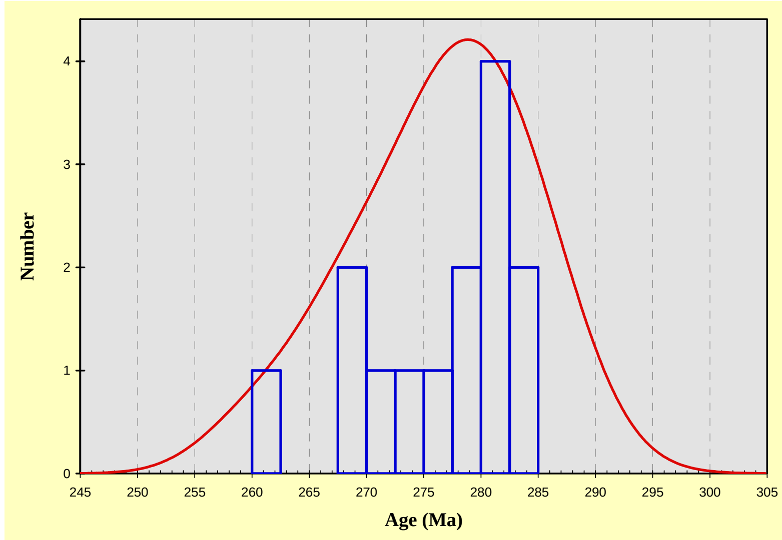
							Radiogenic		Age (Ma)	
Grain. spot	U (ppm)	Th (ppm)	Th/U	Pb* (ppm)	$^{204}Pb/^{206}Pb$	f_{206} %	$^{206}Pb/^{238}U$	±	$^{206}Pb/^{238}U$	±
1.1	102	60	0.59	5	0.001506	0.03	0.0424	0.0007	267.6	4.3
2.1	109	76	0.70	6	0.000075	0.00	0.0428	0.0010	270.1	6.1
3.1	90	47	0.52	5	0.002146	0.06	0.0433	0.0008	273.5	4.9
4.1	70	38	0.54	4	0.000750	0.02	0.0446	0.0011	281.4	6.6
5.1	90	47	0.52	5	0.002064	0.04	0.0437	0.0008	275.7	4.7
6.1	89	53	0.60	5	0.003912	0.06	0.0445	0.0010	280.6	6.0
7.1	102	67	0.66	6	0.000988	0.02	0.0447	0.0008	281.9	5.1
8.1	240	246	1.02	14	0.000866	0.02	0.0441	0.0007	278.1	4.2
9.1	106	82	0.77	6	0.001312	0.02	0.0448	0.0009	282.3	5.6
10.1	115	54	0.48	6	0.003131	0.06	0.0452	0.0009	285.0	5.2
11.1	100	51	0.51	5	0.001725	0.03	0.0428	0.0008	269.9	5.0
12.1	160	106	0.66	9	0.000330	0.01	0.0440	0.0008	277.5	4.8
13.1	75	38	0.51	4	0.003530	0.05	0.0412	0.0007	260.4	4.5
14.1	113	71	0.63	6	0.000339	0.01	0.0450	0.0008	283.8	4.8

Notes :
1. Uncertainties given at the one σ level.
2. f_{206} % denotes the percentage of ^{206}Pb that is common Pb.
3. Correction for common Pb made using the measured $^{238}U/^{206}Pb$ and $^{207}Pb/^{206}Pb$ following Tera & Wasserburg (1972) as outlined in Compston & others (1992).

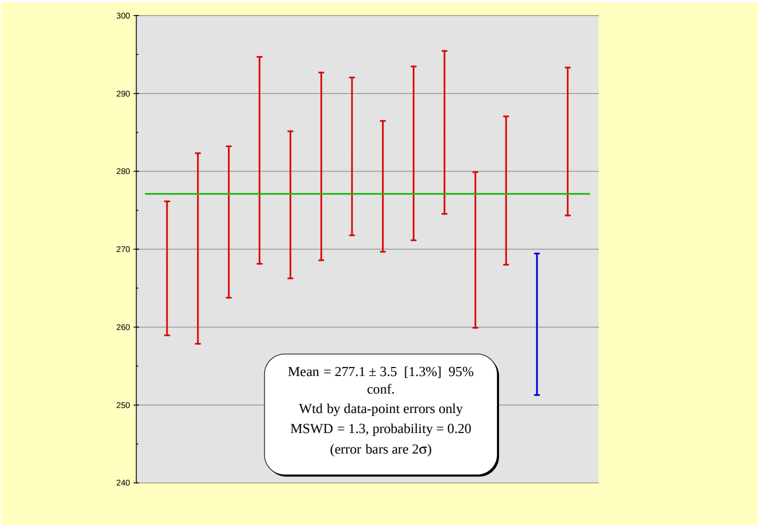
Figure 6. Plots of SHRIMP U-Pb zircon data for sample SCYL0116



(a) Tera-Wasserburg plot, all analyses



(b) Probability density plot – all grains



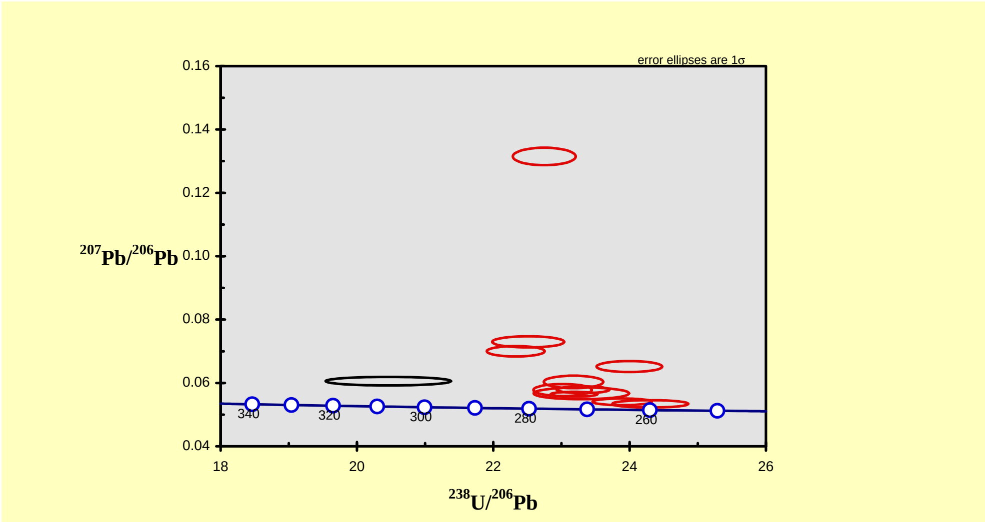
(b) Weighted average – all grains

Table 7. Summary of SHRIMP U-Pb zircon results for sample SCYL0030

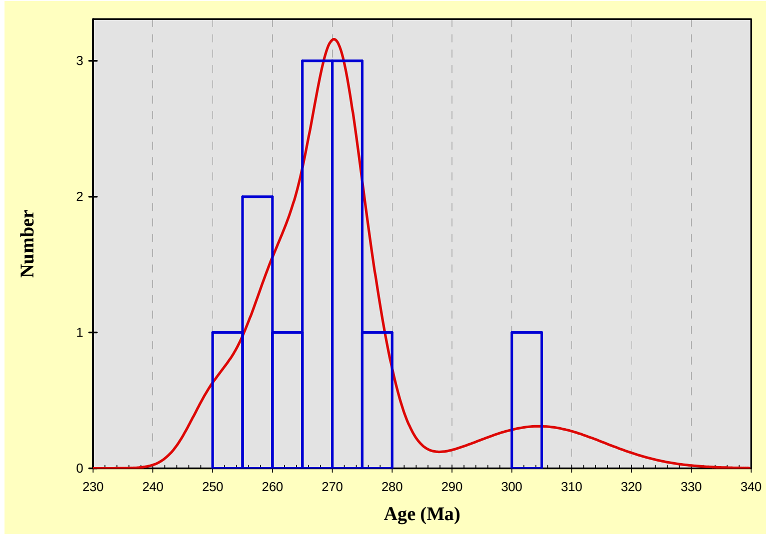
							Radiogenic		Age (Ma)	
Grain. spot	U (ppm)	Th (ppm)	Th/U	Pb* (ppm)	$^{204}Pb/^{206}Pb$	f_{206} %	$^{206}Pb/^{238}U$	±	$^{206}Pb/^{238}U$	±
1.1	600	455	0.76	36	0.000921	0.02	0.0484	0.0018	304.5	10.9
2.1	164	108	0.66	9	0.000200	<0.01	0.0438	0.0007	276.1	4.2
3.1	266	126	0.47	13	0.000619	0.01	0.0426	0.0006	268.7	3.6
4.1	176	129	0.73	9	0.000319	0.01	0.0427	0.0007	269.4	4.1
5.1	190	131	0.69	9	0.005766	0.06	0.0396	0.0007	250.2	4.1
6.1	562	457	0.81	31	0.000447	0.01	0.0429	0.0005	270.6	3.2
7.1	475	281	0.59	29	0.000084	<0.01	0.0411	0.0008	259.4	4.8
8.1	207	114	0.55	13	0.000324	0.01	0.0427	0.0011	269.3	6.5
9.1	372	235	0.63	23	0.001082	0.02	0.0410	0.0007	258.8	4.2
10.1	456	456	1.00	31	0.000154	<0.01	0.0417	0.0006	263.6	3.9
11.1	266	274	1.03	19	0.000502	0.03	0.0431	0.0007	272.1	4.1
12.1	343	277	0.81	23	0.002143	0.03	0.0432	0.0008	272.9	5.2

Notes :
1. Uncertainties given at the one σ level.
2. f_{206} % denotes the percentage of ^{206}Pb that is common Pb.
3. Correction for common Pb made using the measured $^{238}U/^{206}Pb$ and $^{207}Pb/^{206}Pb$ following Tera & Wasserburg (1972) as outlined in Compston & others (1992).

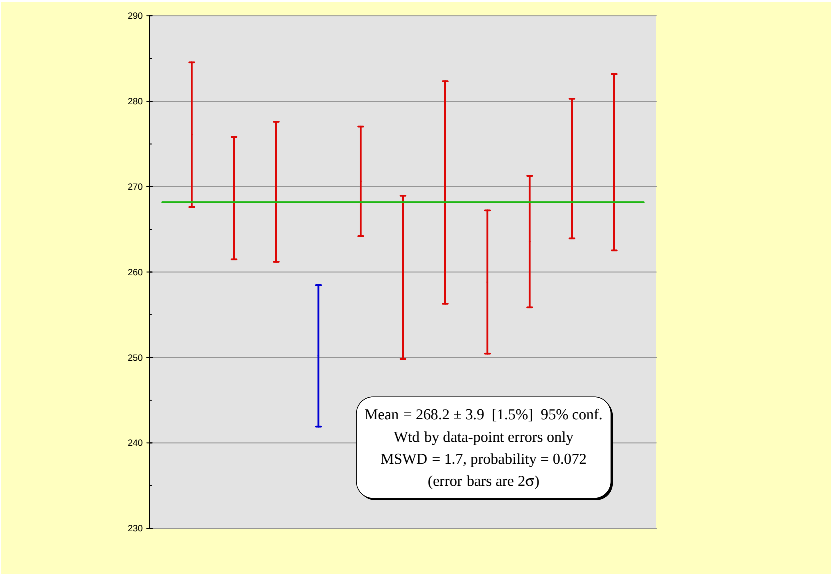
Figure 7. Plots of SHRIMP U-Pb zircon data for sample SCYL0030



(a) Tera-Wasserburg plot, all analyses



(b) Probability density plot – all grains



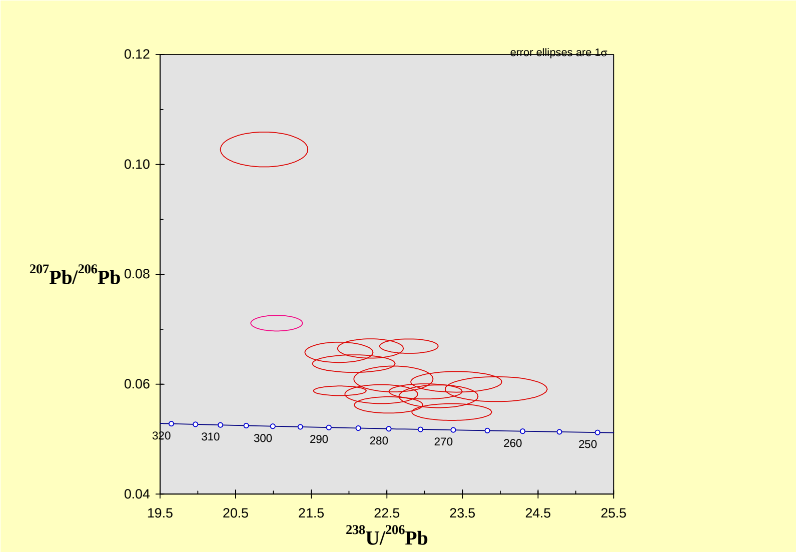
(b) Weighted average – best estimate

Table 8. Summary of SHRIMP U-Pb zircon results for sample SCYL0081

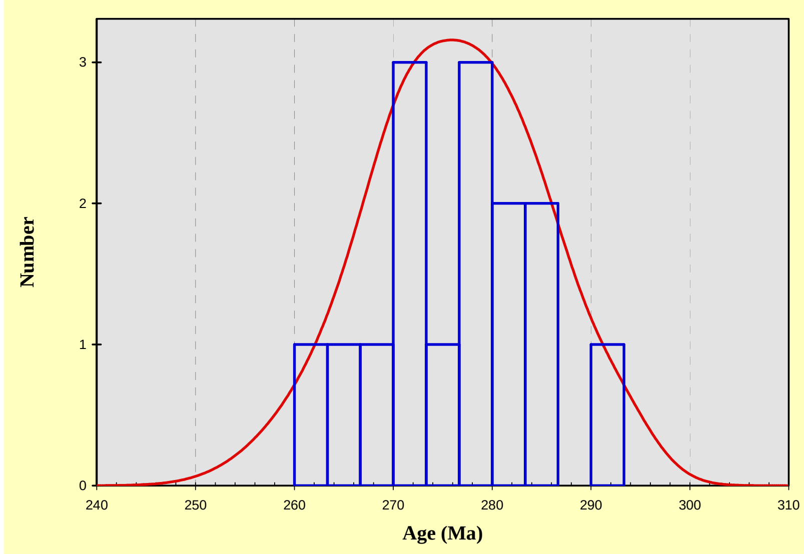
							Radiogenic		Age (Ma)	
Grain. spot	U (ppm)	Th (ppm)	Th/U	Pb* (ppm)	$^{204}Pb/^{206}Pb$	f_{206} %	$^{206}Pb/^{238}U$	±	$^{206}Pb/^{238}U$	±
1.1	192	111	0.58	11	0.001070	0.02	0.0464	0.0006	292.4	3.8
2.1	126	68	0.54	6	0.001349	0.02	0.0431	0.0007	271.9	4.6
3.1	87	43	0.49	5	0.002284	0.06	0.0441	0.0007	277.9	4.4
4.1	150	93	0.62	8	0.001084	0.02	0.0430	0.0006	271.6	3.7
5.1	42	20	0.47	2	0.004003	0.08	0.0449	0.0010	282.9	6.3
6.1	172	85	0.50	9	0.001016	0.02	0.0449	0.0008	283.4	4.7
7.1	301	230	0.76	17	0.000538	0.01	0.0453	0.0006	285.7	3.7
8.1	230	124	0.54	14	0.000951	0.02	0.0428	0.0008	270.2	4.9
9.1	252	144	0.57	17	0.000781	0.02	0.0446	0.0009	281.5	5.6
10.1	158	103	0.65	11	0.000248	0.01	0.0438	0.0008	276.1	5.2
11.1	183	73	0.40	11	0.000918	0.03	0.0414	0.0010	261.3	5.9
12.1	313	253	0.81	22	0.000157	0.01	0.0442	0.0007	278.5	4.5
13.1	204	101	0.49	13	0.000287	0.03	0.0442	0.0008	279.0	4.8
14.1	204	129	0.63	13	-	0.01	0.0422	0.0009	266.7	5.5
15.1	271	146	0.54	17	-	0.01	0.0426	0.0008	269.1	4.9

Notes :
1. Uncertainties given at the one σ level.
2. f_{206} % denotes the percentage of ^{206}Pb that is common Pb.
3. Correction for common Pb made using the measured $^{238}U/^{206}Pb$ and $^{207}Pb/^{206}Pb$ following Tera & Wasserburg (1972) as outlined in Compston & others (1992).

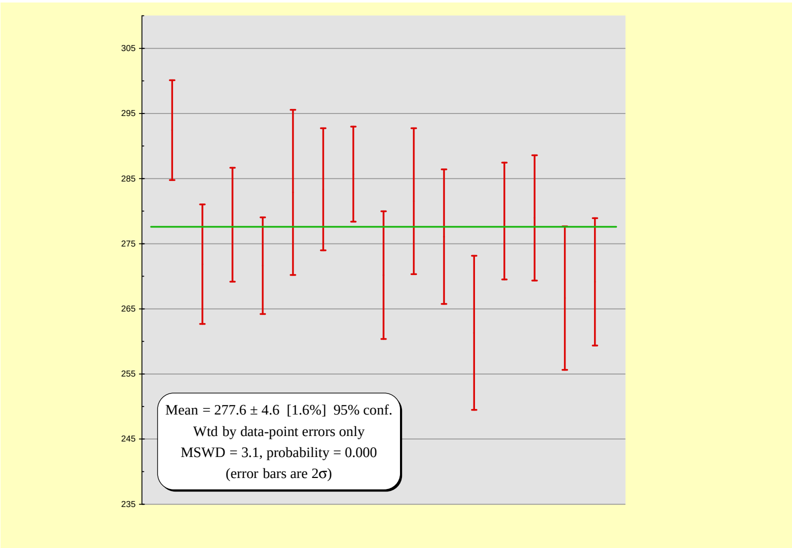
Figure 8. Plots of SHRIMP U-Pb zircon data for sample SCYL0081



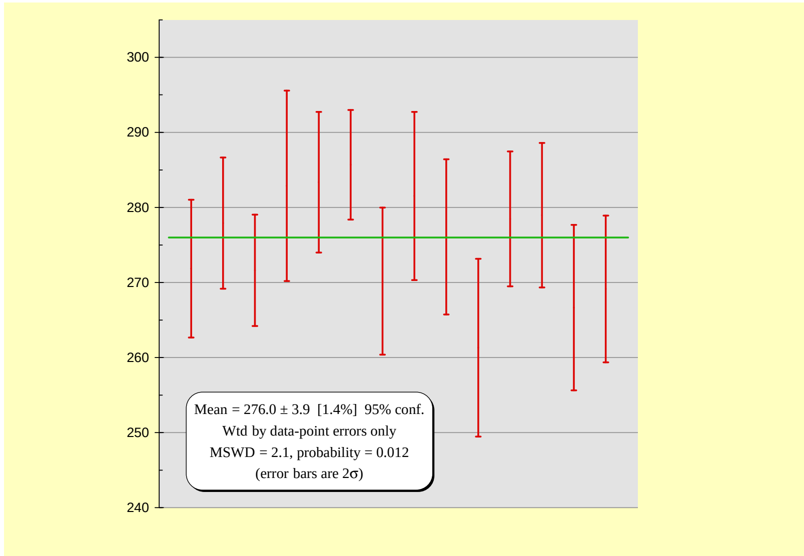
(a) Tera-Wasserburg plot, all analyses



(b) Probability density plot – all grains



(c) Weighted average – all grains



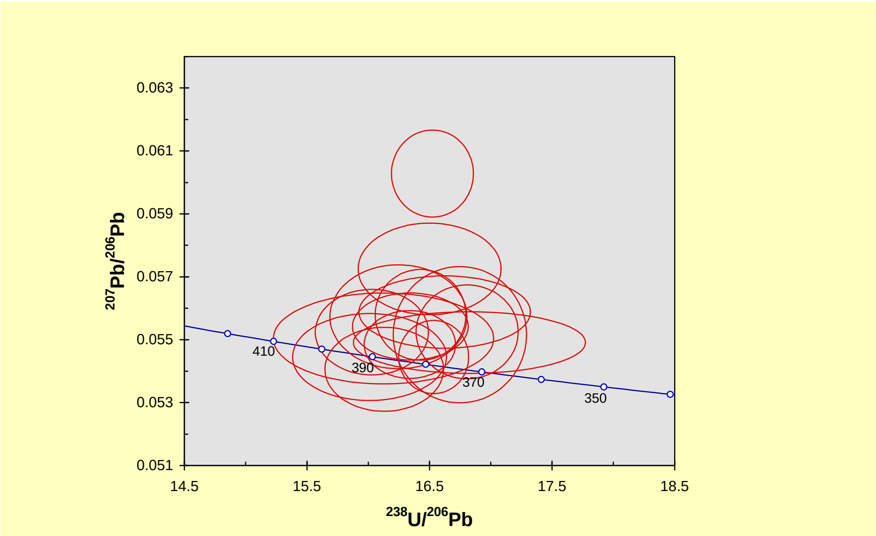
(d) Weighted average – best estimate

Table 9. Summary of SHRIMP U-Pb zircon results for sample CMM239

							Radiogenic		Age (Ma)	
Grain. spot	U (ppm)	Th (ppm)	Th/U	Pb* (ppm)	$^{204}Pb/^{206}Pb$	f_{206} %	$^{206}Pb/^{238}U$	±	$^{206}Pb/^{238}U$	±
1.1	179	98	0.55	12	0.000200	0.02	0.0601	0.0010	376.0	6.1
2.1	225	145	0.64	16	0.000010	0.01	0.0620	0.0015	387.8	9.3
3.1	168	64	0.38	11	-	0.01	0.0623	0.0015	389.6	8.9
4.1	246	196	0.80	18	0.000206	0.01	0.0600	0.0021	375.8	12.6
5.1	215	159	0.74	15	-	0.01	0.0594	0.0012	372.1	7.3
6.1	200	135	0.68	14	0.000434	0.03	0.0604	0.0018	377.9	10.6
7.1	229	133	0.58	16	0.000109	0.00	0.0612	0.0011	382.6	6.9
8.1	226	165	0.73	16	0.000146	0.01	0.0608	0.0011	380.2	6.8
9.1	145	76	0.52	10	0.000583	0.03	0.0596	0.0016	373.4	9.7
10.1	555	399	0.72	39	0.000094	0.01	0.0594	0.0027	371.9	16.7
11.1	231	183	0.79	17	0.000095	0.01	0.0611	0.0014	382.3	8.8
12.1	111	51	0.46	7	0.000139	0.01	0.0614	0.0017	384.4	10.5
13.1	155	75	0.48	11	0.000010	0.02	0.0620	0.0028	387.5	17.2
14.1	225	109	0.48	15	0.000196	0.01	0.0624	0.0020	390.4	12.2
15.1	455	257	0.57	31	0.000170	0.01	0.0605	0.0009	378.5	5.2

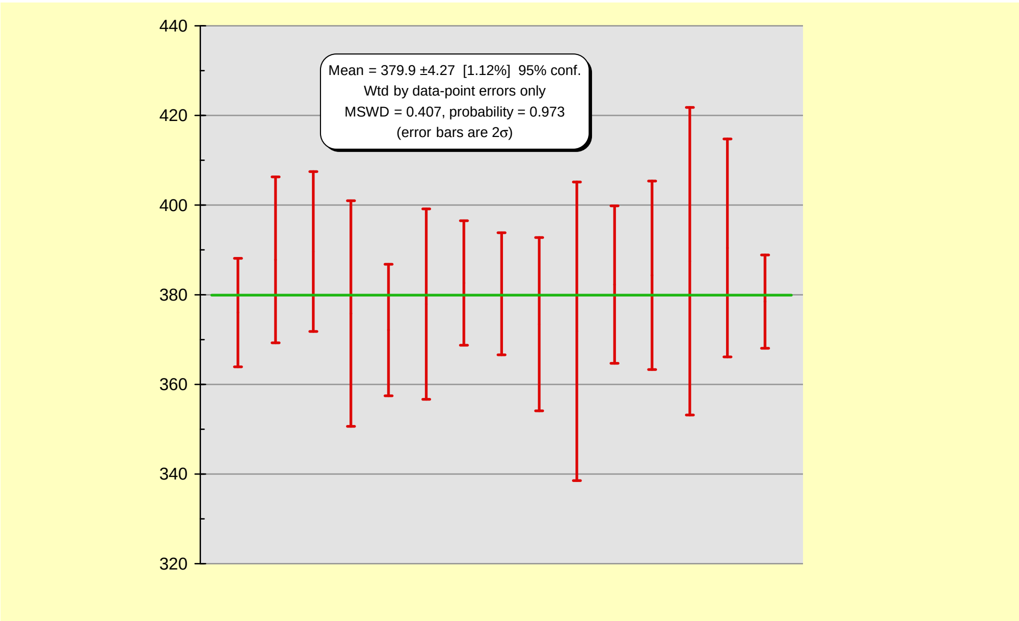
Notes : 1. Uncertainties given at the one σ level.
2. f_{206} % denotes the percentage of ^{206}Pb that is common Pb.
3. Correction for common Pb made using the measured $^{238}U/^{206}Pb$ and $^{207}Pb/^{206}Pb$ following Tera & Wasserburg (1972) as outlined in Compston & others (1992).

Figure 9. Plots of SHRIMP U-Pb zircon data for sample CMM239



(a) Tera-Wasserburg plot, all analyses

(b) Probability density plot – all grains



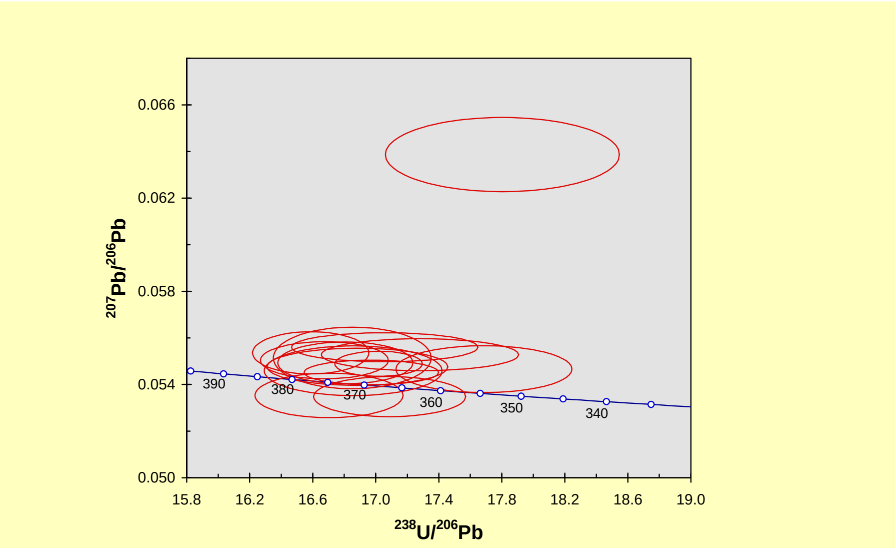
(c) Weighted average – all grains

Table 10. Summary of SHRIMP U-Pb zircon results for sample PBRO528

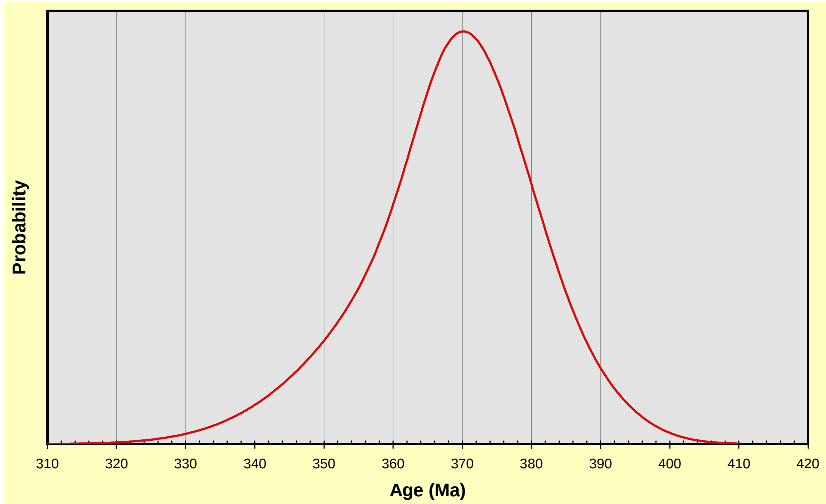
							Radiogenic		Age (Ma)	
Grain. spot	U (ppm)	Th (ppm)	Th/U	Pb* (ppm)	$^{204}Pb/^{206}Pb$	f_{206} %	$^{206}Pb/^{238}U$	±	$^{206}Pb/^{238}U$	±
1.1	417	408	0.98	30	0.003841	0.05	0.0619	0.0012	387.3	7.5
2.1	574	467	0.81	41	0.000089	<0.01	0.0602	0.0011	376.9	6.7
3.1	754	695	0.92	54	0.000082	0.01	0.0585	0.0017	366.7	10.1
4.1	864	893	1.03	65	-	<0.01	0.0599	0.0014	375.2	8.4
5.1	635	393	0.62	43	0.000171	0.01	0.0594	0.0012	372.3	7.6
6.1	595	875	1.47	49	0.000187	0.01	0.0599	0.0012	375.1	7.3
7.1	818	940	1.15	60	0.000024	<0.01	0.0565	0.0015	354.4	8.9
8.1	730	954	1.31	57	0.000060	<0.01	0.0578	0.0017	362.2	10.4
9.1	411	441	1.07	31	-	<0.01	0.0593	0.0014	371.2	8.8
10.1	869	1161	1.34	68	0.000010	<0.01	0.0587	0.0008	367.8	4.8
11.1	451	305	0.68	31	-	<0.01	0.0593	0.0016	371.4	9.9
12.1	872	1289	1.48	72	0.000076	<0.01	0.0592	0.0016	370.7	10.0
13.1	511	632	1.24	39	0.000121	<0.01	0.0586	0.0014	366.9	8.2
14.1	421	332	0.79	28	0.000894	0.02	0.0555	0.0019	348.2	11.6
15.1	1047	1111	1.06	78	0.000072	<0.01	0.0589	0.0012	368.9	7.4

Notes : 1. Uncertainties given at the one σ level.
2. f_{206} % denotes the percentage of ^{206}Pb that is common Pb.
3. Correction for common Pb made using the measured $^{238}U/^{206}Pb$ and $^{207}Pb/^{206}Pb$ following Tera & Wasserburg (1972) as outlined in Compston & others (1992).

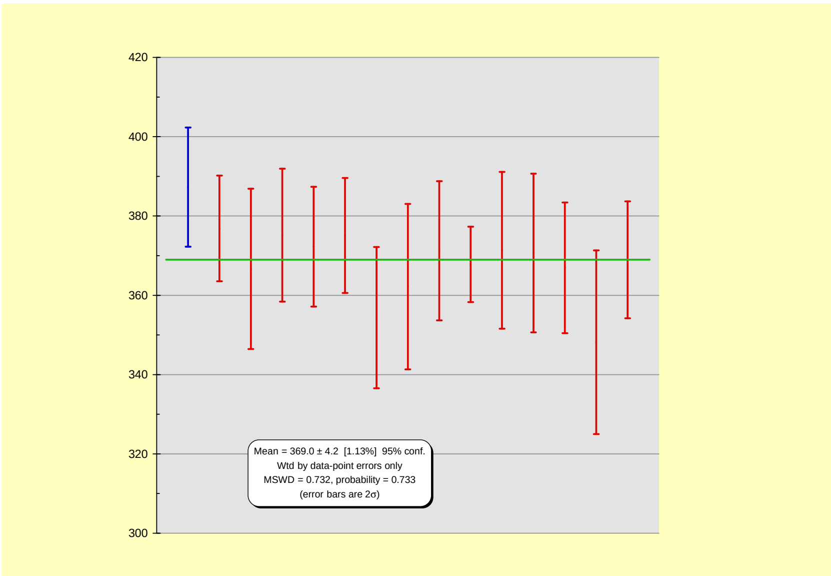
Figure 10. Plots of SHRIMP U-Pb zircon data for sample PBRO528



(a) Tera-Wasserburg plot, enlarged



(b) Probability density plot – all grains



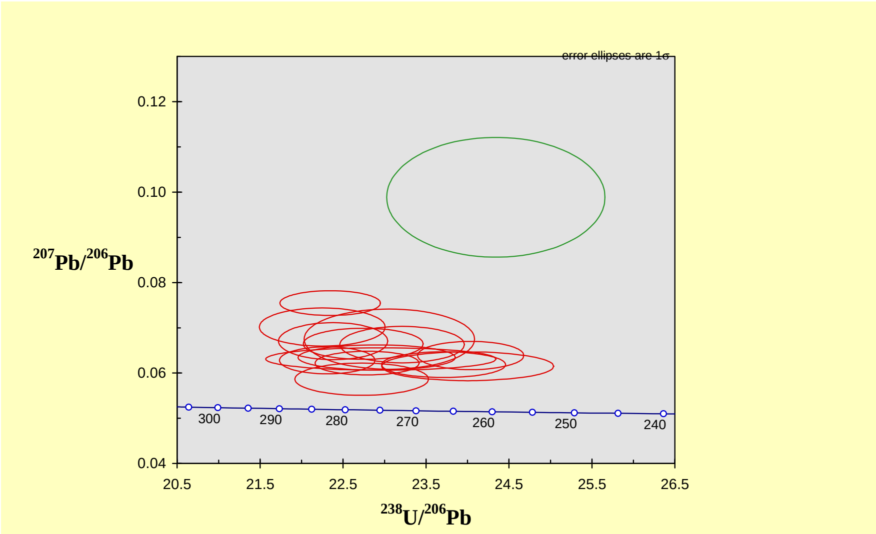
(c) Weighted average – all grains

Table 11. Summary of SHRIMP U-Pb zircon results for sample CMM462B

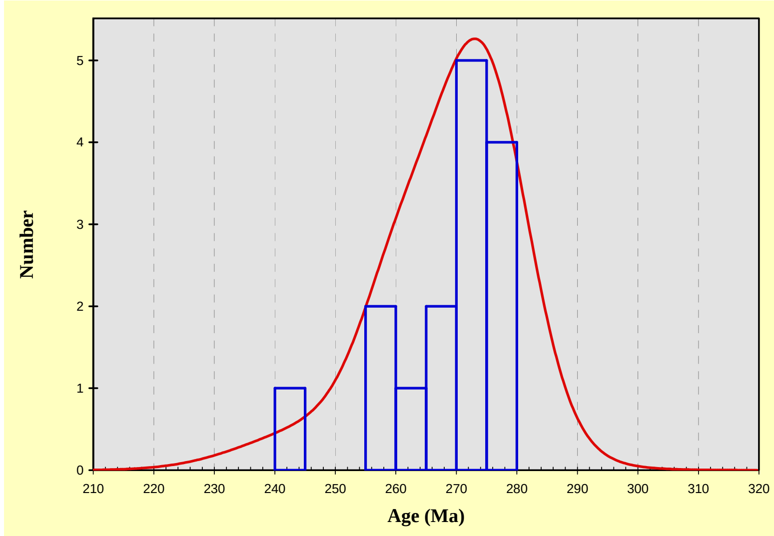
								Radiogenic		Age (Ma)	
Grain. spot	U (ppm)	Th (ppm)	Th/U	Pb* (ppm)	$^{204}Pb/^{206}Pb$	f206	f_{206}%	$^{206}Pb/^{238}U$	±	$^{206}Pb/^{238}U$	±
1.1	136	98	0.72	6	-	0.00873	0.04	0.0436	0.0013	275.2	7.9
2.1	95	69	0.72	5	0.001682	1.43E-02	0.07	0.0429	0.0021	271.0	13.1
3.1	115	92	0.80	5	0.001337	1.23E-02	0.07	0.0412	0.0015	260.0	9.0
4.1	130	116	0.89	6	-	1.32E-02	0.04	0.0433	0.0010	273.3	6.1
5.1	110	83	0.76	5	0.000551	1.85E-02	0.07	0.0432	0.0011	272.4	7.0
6.1	105	80	0.76	5	-	2.98E-02	0.04	0.0434	0.0010	274.0	6.0
7.1	107	85	0.79	5	0.000010	1.93E-02	0.01	0.0438	0.0011	276.5	6.6
8.1	93	54	0.58	4	0.002185	1.98E-02	0.05	0.0425	0.0016	268.4	9.8
9.1	98	74	0.76	4	0.002178	1.53E-02	0.06	0.0410	0.0009	258.8	5.6
10.1	103	86	0.84	5	0.000842	0.01276	0.02	0.0416	0.0011	263.0	6.7
11.1	86	50	0.58	4	-	0.01473	0.08	0.0430	0.0015	271.5	9.0
12.1	51	34	0.65	2	0.007477	5.90E-02	0.30	0.0387	0.0018	244.5	11.2
13.1	70	46	0.66	3	0.004293	0.02315	0.14	0.0439	0.0012	277.0	7.7
14.1	120	86	0.71	6	0.001595	0.0183	0.05	0.0423	0.0011	267.0	7.0
15.1	116	82	0.71	6	0.001258	0.01394	0.05	0.0442	0.0009	278.8	5.8

Notes :
1. Uncertainties given at the one σ level.
2. f_{206} % denotes the percentage of ^{206}Pb that is common Pb.
3. Correction for common Pb made using the measured $^{238}U/^{206}Pb$ and $^{207}Pb/^{206}Pb$ following Tera & Wasserburg (1972) as outlined in Compston & others (1992).

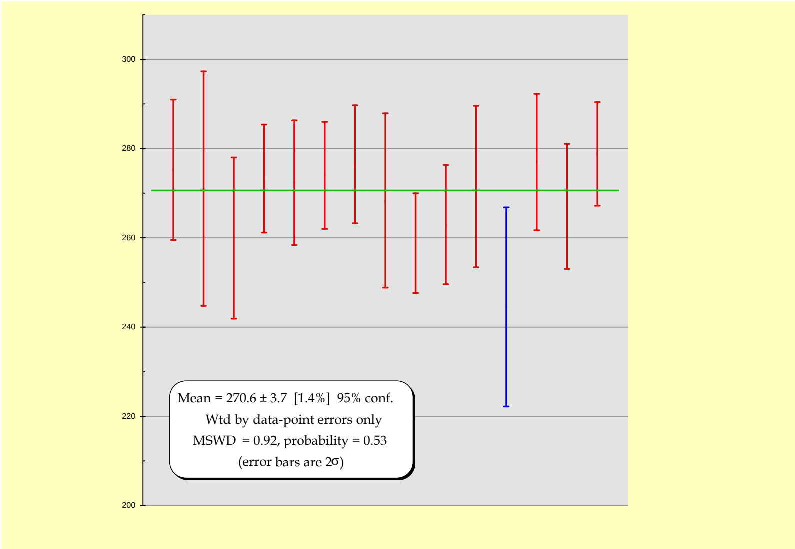
Figure 11. Plots of SHRIMP U-Pb zircon data for sample CMM462B



(a) Tera-Wasserburg plot, all analyses



(b) Probability density plot – all grains



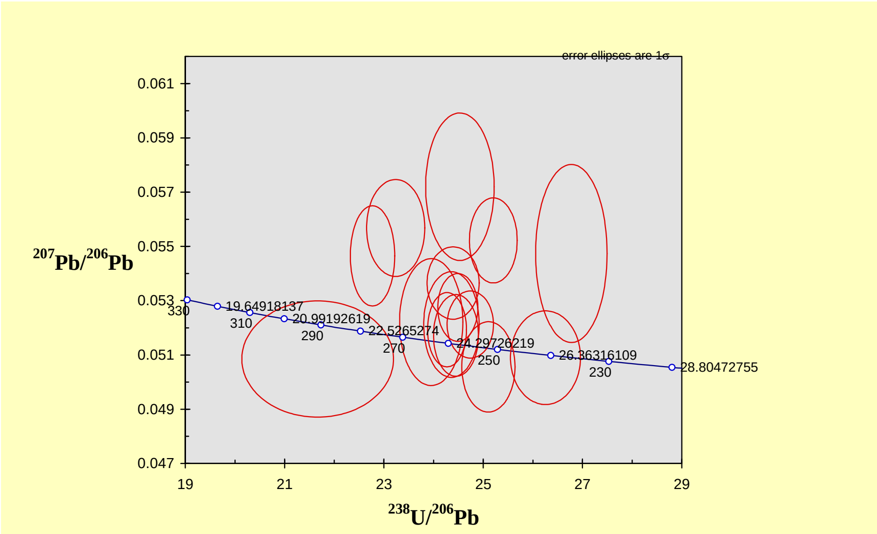
(c) Weighted average

Table 12. Summary of SHRIMP U-Pb zircon results for sample CMM 106B

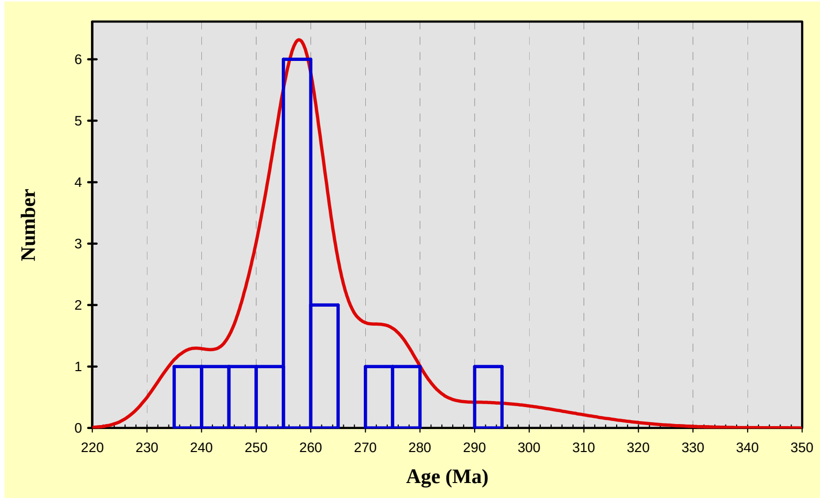
							Radiogenic		Age (Ma)	
Grain. spot	U (ppm)	Th (ppm)	Th/U	Pb* (ppm)	$^{204}Pb/^{206}Pb$	f_{206} %	$^{206}Pb/^{238}U$	±	$^{206}Pb/^{238}U$	±
1.1	183	128	0.70	9	0.000839	0.05	0.0405	0.0009	255.8	5.8
2.1	493	515	1.04	25	0.000010	<0.01	0.0395	0.0006	249.7	3.8
3.1	318	202	0.64	16	0.000081	<0.01	0.0410	0.0008	259.1	4.8
4.1	283	217	0.77	13	0.000561	0.02	0.0372	0.0008	235.4	5.1
5.1	596	746	1.25	34	0.000181	0.01	0.0408	0.0006	257.6	3.4
6.1	295	231	0.78	14	0.000262	0.03	0.0381	0.0008	241.1	5.2
7.1	399	341	0.86	21	-	0.01	0.0417	0.0009	263.4	5.6
8.1	265	225	0.85	13	0.000390	0.01	0.0399	0.0007	252.1	4.3
9.1	513	472	0.92	26	0.000360	0.03	0.0404	0.0006	255.2	3.9
10.1	567	698	1.23	32	0.000073	0.00	0.0412	0.0006	260.1	3.4
11.1	384	251	0.65	20	0.000041	0.01	0.0437	0.0007	276.0	4.4
12.1	346	302	0.87	21	0.000343	0.01	0.0462	0.0027	291.1	16.4
13.1	468	389	0.83	24	0.000010	<0.01	0.0409	0.0007	258.3	4.5
14.1	258	154	0.59	13	0.000753	0.05	0.0428	0.0009	270.2	5.5
15.1	537	541	1.01	29	0.000145	0.01	0.0409	0.0006	258.3	3.9

Notes :
1. Uncertainties given at the one σ level.
2. f_{206} % denotes the percentage of ^{206}Pb that is common Pb.
3. Correction for common Pb made using the measured $^{238}U/^{206}Pb$ and $^{207}Pb/^{206}Pb$ following Tera & Wasserburg (1972) as outlined in Compston & others (1992).

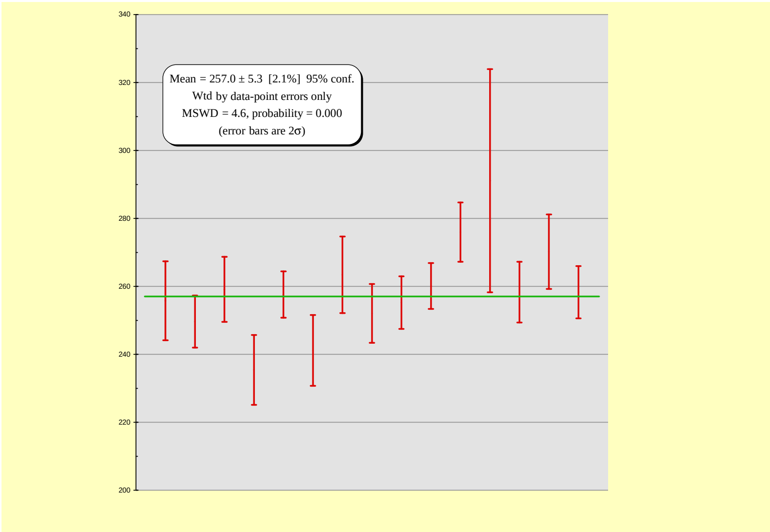
Figure 12. Plots of SHRIMP U-Pb zircon data for sample CMM106B



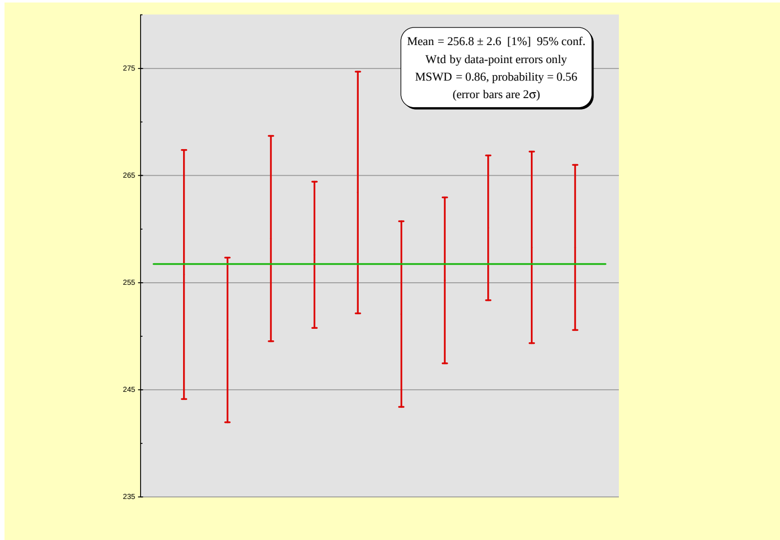
(a) Tera-Wasserburg plot, all analyses



(b) Probability density plot – all grains



(c) Weighted average – all grains



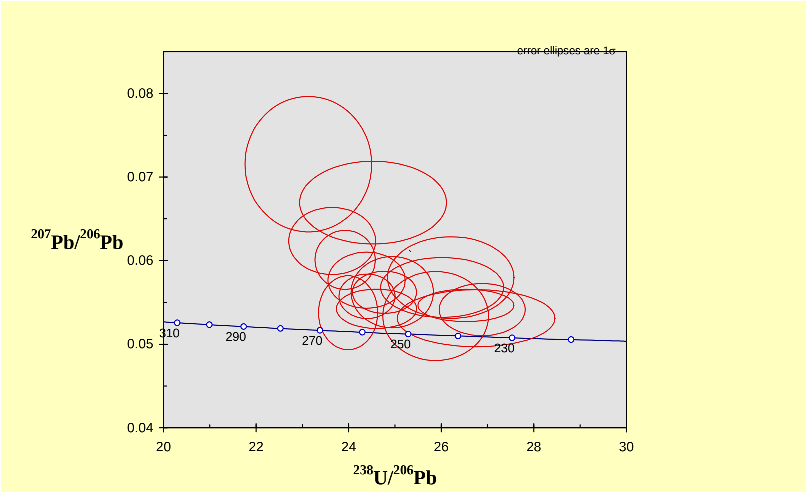
(d) Weighted average – best estimate

Table 13. Summary of SHRIMP U-Pb zircon results for sample CMM245

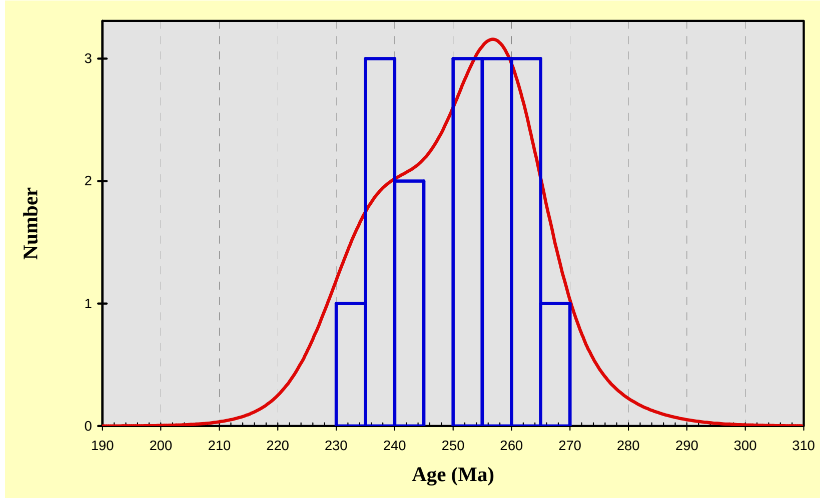
							Radiogenic		Age (Ma)	
Grain. spot	U (ppm)	Th (ppm)	Th/U	Pb* (ppm)	$^{204}Pb/^{206}Pb$	f_{206} %	$^{206}Pb/^{238}U$	±	$^{206}Pb/^{238}U$	±
1.1	83	61	0.73	5	0.000511	0.05	0.0408	0.0008	257.6	5.2
2.1	37	24	0.66	2	0.000818	0.03	0.0417	0.0014	263.5	8.4
3.1	72	68	0.95	5	0.001176	0.04	0.0413	0.0009	261.2	5.8
4.1	57	36	0.63	3	0.001626	0.04	0.0401	0.0009	253.6	5.7
5.1	47	37	0.80	3	-	0.10	0.0379	0.0016	239.5	10.1
6.1	41	31	0.74	2	0.002119	0.09	0.0400	0.0021	252.7	13.2
7.1	104	111	1.07	6	0.000896	0.05	0.0375	0.0012	237.5	7.5
8.1	56	40	0.72	3	0.001380	0.05	0.0385	0.0014	243.8	8.7
9.1	48	37	0.77	3	0.002089	0.08	0.0421	0.0021	266.1	12.8
10.1	54	37	0.69	3	-	0.04	0.0416	0.0009	262.5	5.7
11.1	37	26	0.71	2	-	0.13	0.0371	0.0011	234.6	6.6
12.1	43	28	0.67	2	-	0.09	0.0382	0.0016	241.5	9.9
13.1	51	36	0.71	3	0.000623	0.04	0.0407	0.0011	257.1	7.1
14.1	62	29	0.47	3	0.000938	0.07	0.0373	0.0020	236.1	12.1
15.1	70	55	0.78	4	0.000708	0.02	0.0405	0.0012	255.9	7.3
16.1	45	32	0.71	3	0.001639	0.07	0.0399	0.0012	251.9	7.3

Notes : 1. Uncertainties given at the one σ level.
2. f_{206} % denotes the percentage of ^{206}Pb that is common Pb.
3. Correction for common Pb made using the measured $^{238}U/^{206}Pb$ and $^{207}Pb/^{206}Pb$ following Tera & Wasserburg (1972) as outlined in Compston & others (1992).

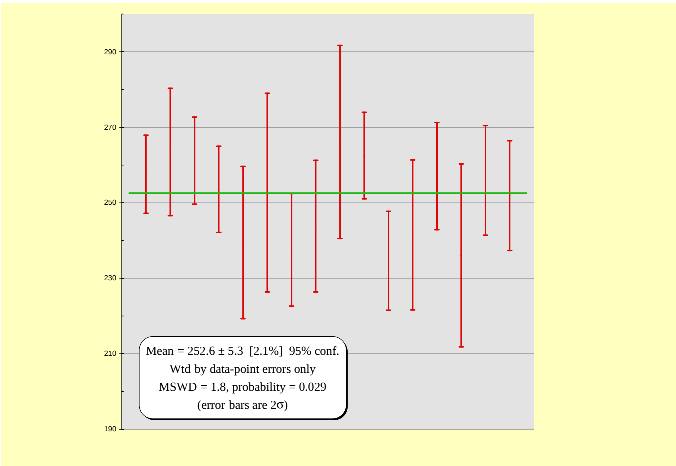
Figure 13. Plots of SHRIMP U-Pb zircon data for sample CMM245



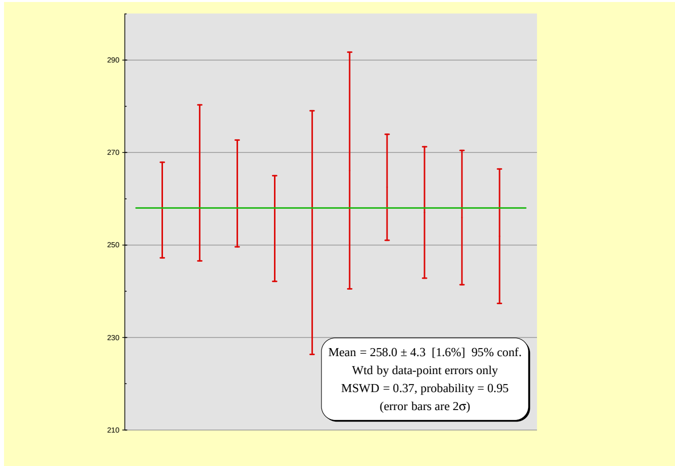
(a) Tera-Wasserburg plot, all analyses



(b) Probability density plot – all grains

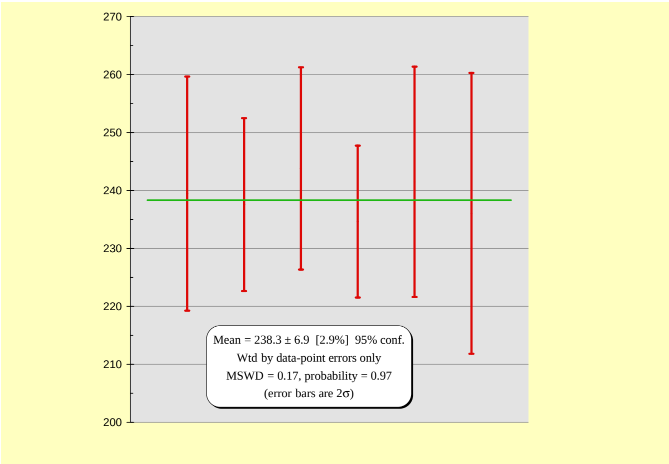


(c) Weighted average – all grains



(d) Weighted average – older group

Figure 13. Plots of SHRIMP U-Pb zircon data for sample CMM245



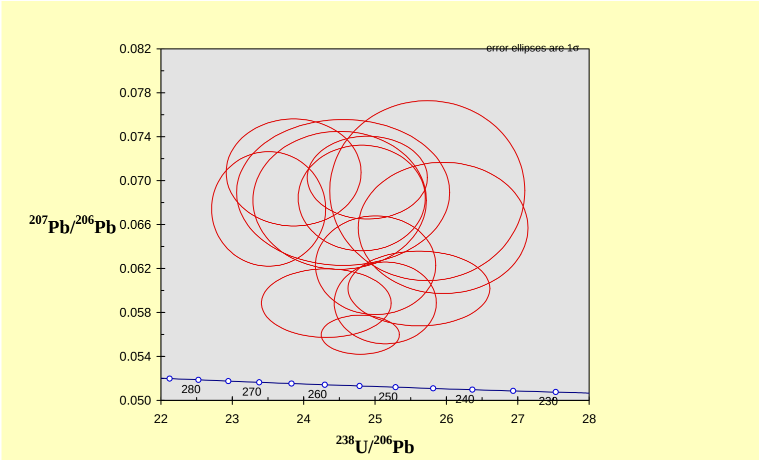
(e) Weighted average – younger group

Table 14. Summary of SHRIMP U-Pb zircon results for sample CMM145A

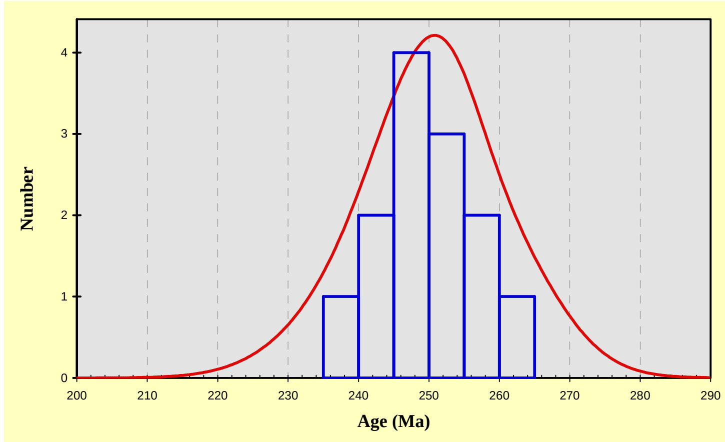
									Radiogenic		Age (Ma)	
Grain. spot	U (ppm)	Th (ppm)	Th/U	Pb* (ppm)	$^{204}Pb/^{206}Pb$	f_{206} %	238/206	±38/6	$^{206}Pb/^{238}U$	±	$^{206}Pb/^{238}U$	±
1.1	65	61	0.94	6	0.001818	0.06	23.9930	0.679	0.0417	0.0012	263.2	7.3
2.1	60	44	0.74	5	-	0.10	25.9020	0.826	0.0386	0.0012	244.2	7.7
3.1	48	39	0.81	4	0.002882	0.14	25.1060	1.259	0.0398	0.0020	251.8	12.4
4.1	49	28	0.58	4	0.002971	0.11	26.4300	1.002	0.0378	0.0014	239.4	8.9
5.1	93	92	0.99	8	0.000723	0.03	24.5500	0.754	0.0407	0.0013	257.4	7.8
6.1	55	55	1.00	5	0.000371	0.07	25.0340	1.028	0.0400	0.0016	252.5	10.2
7.1	59	56	0.95	5	0.000010	0.03	25.4960	0.711	0.0392	0.0011	248.0	6.8
7.1b	58	55	0.96	5	0.002627	0.12	25.3570	0.708	0.0394	0.0011	249.4	6.8
8.1	55	35	0.63	4	0.002144	0.14	24.4540	0.801	0.0409	0.0013	258.4	8.3
9.1	81	80	0.99	7	0.002463	0.10	25.3560	0.754	0.0394	0.0012	249.4	7.3
10.1	208	243	1.17	18	0.000343	0.01	24.9390	0.453	0.0401	0.0007	253.4	4.5
11.1	89	93	1.04	8	0.003947	0.14	25.3840	0.598	0.0394	0.0009	249.1	5.8
12.1	44	21	0.48	3	0.002944	0.10	26.3170	1.163	0.0380	0.0017	240.4	10.4

Notes :
1. Uncertainties given at the one σ level.
2. f_{206} % denotes the percentage of ^{206}Pb that is common Pb.
3. Correction for common Pb made using the measured $^{238}U/^{206}Pb$ and $^{207}Pb/^{206}Pb$ following Tera & Wasserburg (1972) as outlined in Compston & others (1992).

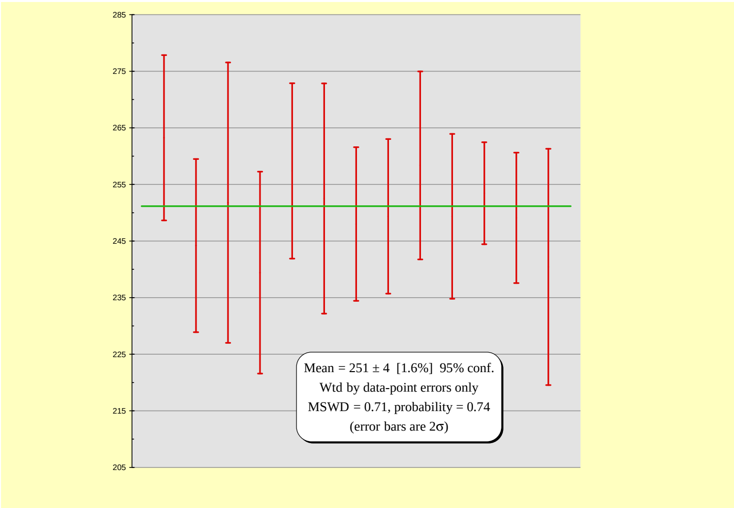
Figure 14. Plots of SHRIMP U-Pb zircon data for sample CMM145a



(a) Tera-Wasserburg plot, all analyses



(b) Probability density plot – all grains



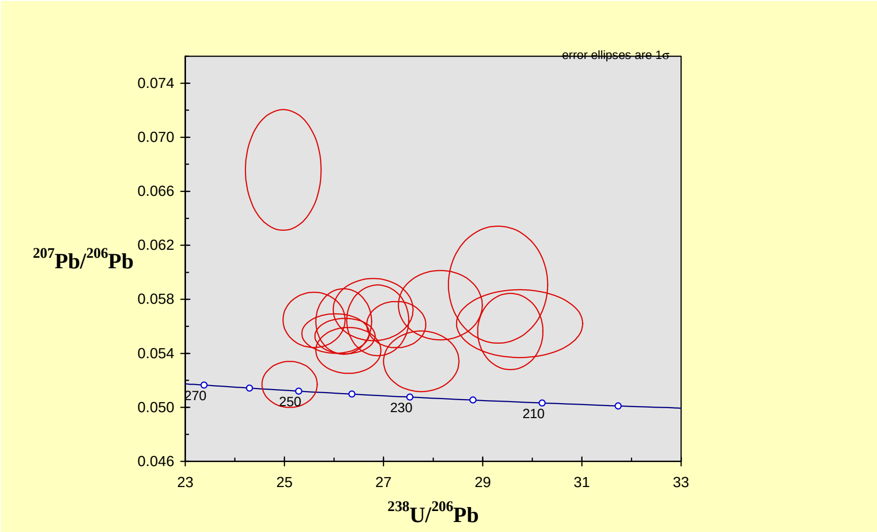
(c) Weighted average – all grains

Table 15. Summary of SHRIMP U-Pb zircon results for sample CMM242

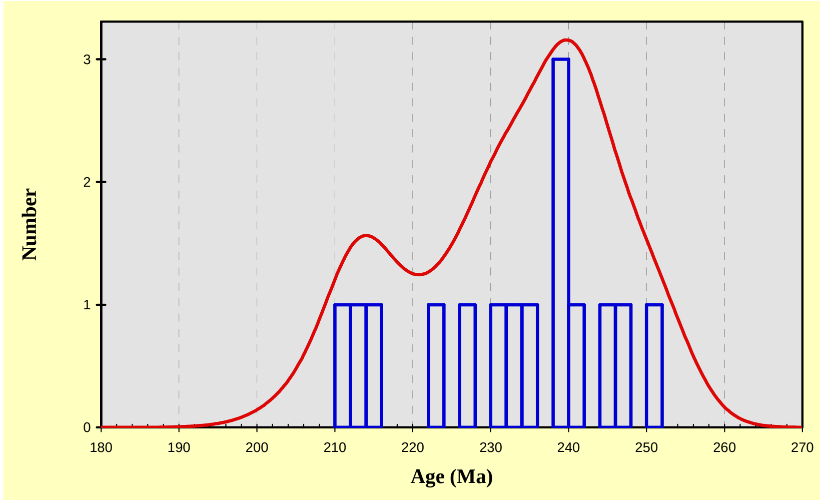
							Radiogenic		Age (Ma)	
Grain. spot	U (ppm)	Th (ppm)	Th/U	Pb* (ppm)	$^{204}Pb/^{206}Pb$	f_{206} %	$^{206}Pb/^{238}U$	±	$^{206}Pb/^{238}U$	±
1.1	188	117	0.62	10	0.000010	0.01	0.0359	0.0008	227.5	5.0
2.1	294	113	0.38	14	0.000147	0.02	0.0334	0.0012	211.8	7.3
3.1	449	175	0.39	17	0.000631	0.03	0.0382	0.0008	241.7	5.1
4.1	93	44	0.47	4	0.004222	0.15	0.0392	0.0010	247.9	6.2
5.1	342	192	0.56	14	0.000023	0.01	0.0398	0.0007	251.6	4.5
6.1	269	109	0.40	10	0.000758	0.02	0.0365	0.0007	230.8	4.1
7.1	288	221	0.76	12	-	0.01	0.0379	0.0008	239.7	4.8
8.1	308	123	0.40	11	0.000884	0.03	0.0338	0.0010	214.1	6.0
9.1	313	148	0.47	12	0.000216	0.02	0.0379	0.0007	240.0	4.5
10.1	257	143	0.56	10	0.000394	0.02	0.0352	0.0009	223.2	5.4
11.1	215	107	0.50	8	0.000571	0.04	0.0370	0.0009	234.4	5.7
12.1	357	191	0.54	14	0.000630	0.02	0.0379	0.0007	239.9	4.2
13.1	302	134	0.44	10	0.000376	0.04	0.0336	0.0006	213.3	3.9
14.1	263	133	0.51	10	0.000389	0.02	0.0370	0.0007	233.9	4.4
15.1	224	92	0.41	9	-	0.01	0.0388	0.0008	245.4	4.8

Notes :
1. Uncertainties given at the one σ level.
2. f_{206} % denotes the percentage of ^{206}Pb that is common Pb.
3. Correction for common Pb made using the measured $^{238}U/^{206}Pb$ and $^{207}Pb/^{206}Pb$ following Tera & Wasserburg (1972) as outlined in Compston & others (1992).

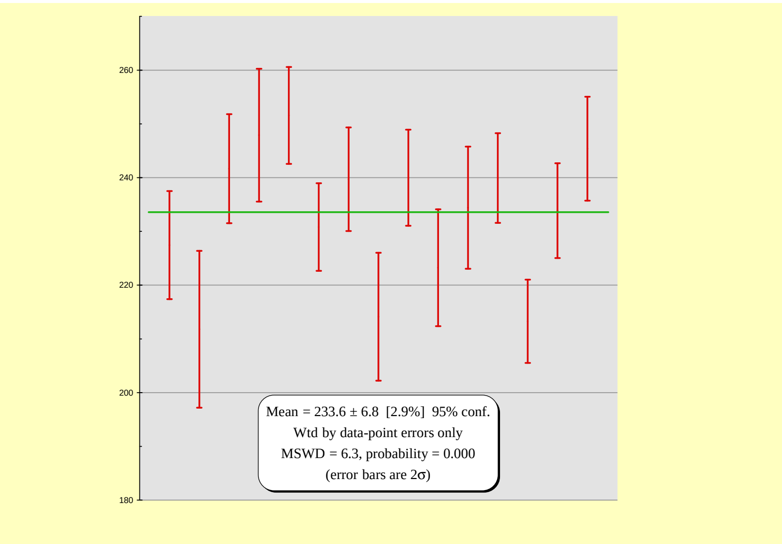
Figure 15. Plots of SHRIMP U-Pb zircon data for sample CMM242



(a) Tera-Wasserburg plot, all analyses



(b) Probability density plot – all grains



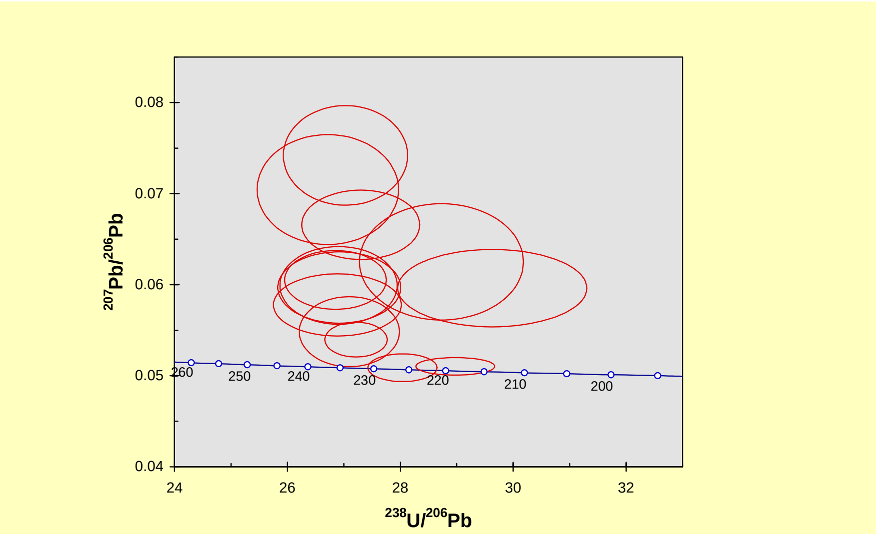
(c) Weighted average – all grains

Table 16. Summary of SHRIMP U-Pb zircon results for sample CMM418

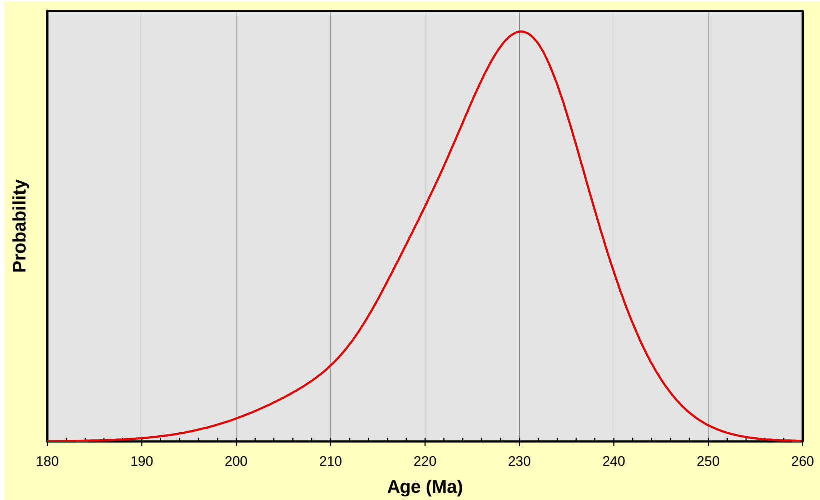
							Radiogenic		Age (Ma)	
Grain. spot	U (ppm)	Th (ppm)	Th/U	Pb* (ppm)	$^{204}Pb/^{206}Pb$	f_{206} %	$^{206}Pb/^{238}U$	±	$^{206}Pb/^{238}U$	±
1.1	684	953	1.39	32	-	<0.01	0.0345	0.0007	218.7	4.2
2.1	58	29	0.50	2	-	0.14	0.0367	0.0010	232.5	6.2
3.1	48	23	0.48	2	-	0.10	0.0368	0.0010	232.9	6.3
4.1	62	31	0.50	2	0.002184	0.14	0.0343	0.0014	217.4	9.0
5.1	56	29	0.51	2	0.002611	0.11	0.0334	0.0016	211.6	9.7
6.1	210	188	0.90	8	0.000113	0.01	0.0366	0.0006	231.7	3.8
7.1	41	16	0.40	1	0.002314	0.07	0.0365	0.0014	231.2	8.8
8.1	79	47	0.60	3	-	0.02	0.0367	0.0012	232.6	7.3
9.1	49	27	0.55	2	-	0.11	0.0367	0.0012	232.6	7.6
10.1	51	19	0.36	2	-	0.19	0.0369	0.0013	233.4	7.9
11.1	680	942	1.39	30	0.000051	<0.01	0.0357	0.0006	225.9	4.0
12.1	44	28	0.63	2	0.000337	0.07	0.0359	0.0012	227.4	7.6
13.1	46	22	0.47	2	0.001156	0.11	0.0359	0.0011	227.4	7.1

Notes :
1. Uncertainties given at the one σ level.
2. f_{206} % denotes the percentage of ^{206}Pb that is common Pb.
3. Correction for common Pb made using the measured $^{238}U/^{206}Pb$ and $^{207}Pb/^{206}Pb$ following Tera & Wasserburg (1972) as outlined in Compston & others (1992).

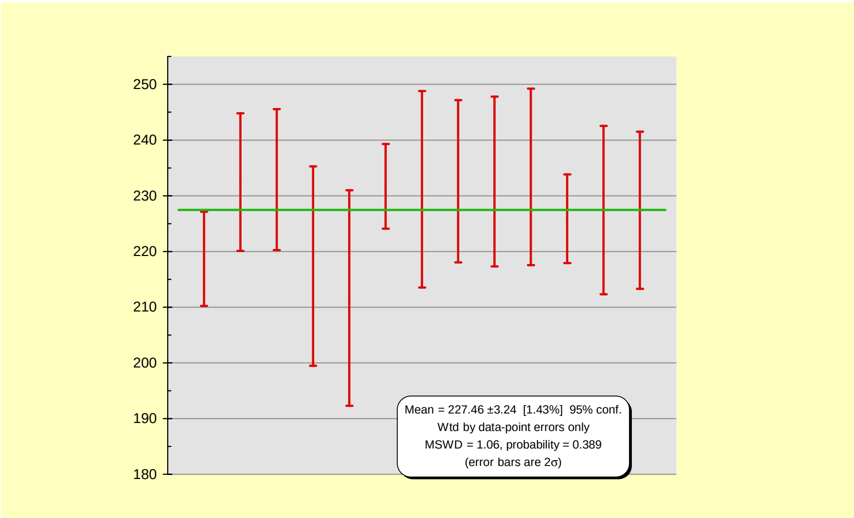
Figure 16. Plots of SHRIMP U-Pb zircon data for sample CMM418



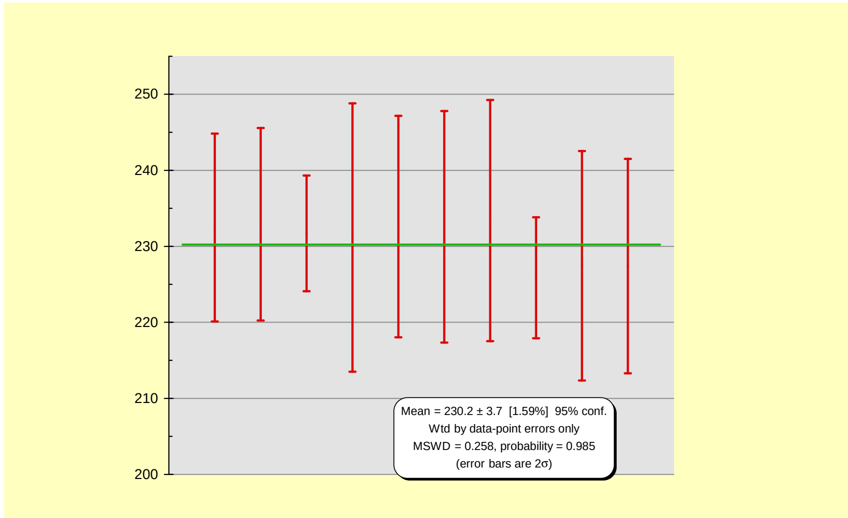
(a) Tera-Wasserburg plot, all analyses



(b) Probability density plot – all grains



(c) Weighted average – all grains



(d) Weighted average – best estimate

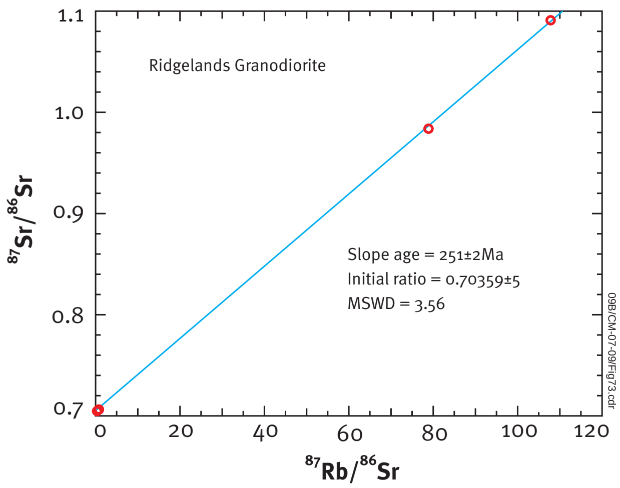


Figure 17. Rb-Sr isochron for the Ridgelands Granodiorite

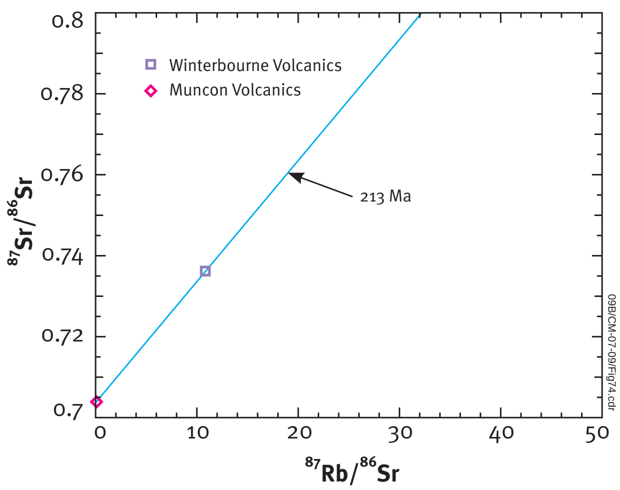


Figure 18. Approximate Rb-Sr isochron for the Winterbourne Volcanics

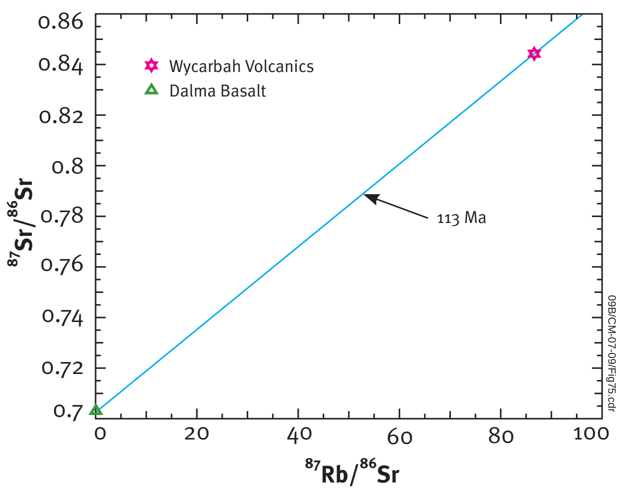


Figure 19. Approximate Rb-Sr isochron for the Wycarbah Volcanics

APPENDIX 2
ORDOVICIAN – LOWER CARBONIFEROUS CONODONT DATING

(B.G. Fordham)

PALAEOZOIC STRATIGRAPHIC ELEMENTS OF THE YARROL PROVINCE

Subdivision of the Yarrol Province and the adjacent Gogango Overfolded Zone for geological synthesis has usually been treated in terms of tectonic elements. One framework, established by Day & others (1978; see Day & others, 1983; Murray, 1986), recognises a Silurian–Middle Devonian Calliope Island Arc, an Upper Devonian–Cisuralian Yarrol Forearc Basin, including a Berserker Graben, and a Cisuralian Grantleigh Trough. This scheme, however, has been questioned: recent mapping and geochemistry (Murray & others, 1997; Murray & Blake, 2005) queried the integrity of the Calliope Arc as a single entity; Murray & others (1987) considered the continuity of the Yarrol Forearc Basin was broken by a Late Carboniferous nonconvergent regime; Crouch & Parfrey (1998) and Crouch (1999a,b) doubted the existence of the Berserker Graben; and Fielding & others (1994, 1997a) considered the Grantleigh Trough merely an eastern extension of the Bowen Basin.

An alternative descriptive framework, based on stratigraphic elements (Fordham, *in* Yarrol Project Team, 1997), is likely to be more reliable. This uses “stratigraphic assemblages” to recognise groups of rock units, and their lithostratigraphic correlatives, which share conformable contacts. Stratigraphic assemblages are thus those rock packages lacking mutual stratigraphic contacts — analogous to the structural “terrane” — within intervals separated by regional unconformities.

Palaeozoic stratigraphic assemblages of the Yarrol Block (Figure 1) fall into four main intervals separated by unconformities near the Middle Devonian–Upper Devonian, Visean–Serpukhovian, and Cisuralian–Guadalupian boundaries. This does not include poorly understood pre-Silurian rocks: a very thin Lower/Middle Ordovician sliver(?) crops out just west of the Yarrol Fault near Santa Glen and, nearby, a conglomerate, presumed to be Devonian, contains Late Ordovician limestone cobbles.

Below an apparently regional hiatus with an extent of at least uppermost Middle Devonian, there are four Upper Silurian–Devonian stratigraphic assemblages: the Craigilee beds, the Calliope beds, the Erebus beds, and the Mount Morgan Stratigraphic Assemblage. The Craigilee beds occur in the north-west of the Yarrol Province; the Calliope beds occur near the eastern margin of the Province near the Yarrol Fault; the Erebus beds are confined to east of the Dee Range; and the Mount Morgan Stratigraphic Assemblage, scattered through the western part of the Yarrol Block, is represented by the Capella Creek Group (and Mount Morgan Trondhjemite) near Mount Morgan, and the Marble Waterhole beds in the Kroombit Creek area. Each of the four Silurian–Devonian stratigraphic assemblages lacks an exposed base but is known down to the uppermost Silurian except for the Mount Morgan Stratigraphic Assemblage.

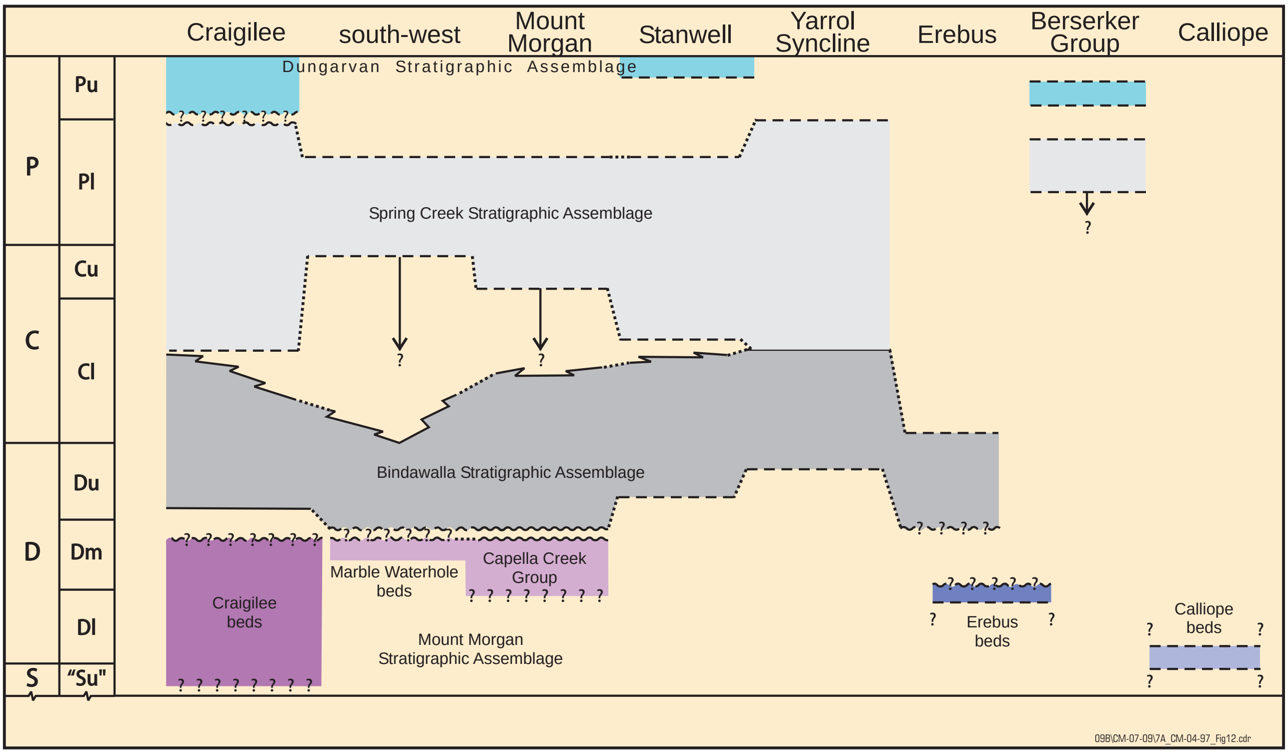


Figure 1. Schematic Palaeozoic stratigraphy of the Yarrol Block (modified after Fordham in Yarrol Project Team, 1997, figure 2).

Contacts have been placed between both easterly stratigraphic assemblages and Upper Devonian units of the Bindawalla Stratigraphic Assemblage which are considered to represent major hiatuses, and basal intervals of these Upper Devonian units contain limestone bodies interpreted to be allochthonous blocks. The Upper Devonian Mount Alma Formation contains limestone-clast conglomerates probably sourced from the Calliope beds. Thus the uppermost interpreted parts of the Calliope beds (upper Pragian) and Erebus beds (upper Emsian) may be stratigraphically reduced erosional surfaces. However, with one exception, the Upper Devonian conglomerates and interpreted blocks can be demonstrated to have sourced only the uppermost levels of the older stratigraphic assemblages. Indeed, more closely spaced sampling of limestones either side of the top of the Erebus beds has not convincingly demonstrated that the interpreted blocks are as old as the uppermost limestones of the Erebus beds. The exception comes from the Mount Etna area where enormous limestone bodies interpreted to be allochthonous include a Lochkovian interval.

The westerly Mount Morgan Stratigraphic Assemblage extends as least as high as middle Givetian but the regional extent of the upper-Givetian, and possibly early-Frasnian, gap is unclear. Although near Mount Morgan it is represented by the well-known uncomformity, constrained by middle-Givetian dates below and mid-Frasnian above, the hiatus may have been very short depending on the accumulation rates assumed for the intervals either side. Elsewhere contacts are faulted and, in the north-west, dates from the Craigilee beds reach even higher in the Givetian, leaving very little time for the gap.

The Upper Devonian–Visean unconformity interval contains only the Bindawalla Stratigraphic Assemblage. This assemblage has apparently unconformable contacts with all four Silurian–Devonian stratigraphic assemblages. It consists of several subregional Upper Devonian units overlain by the extensive Tournaisian–Visean Rockhampton Group. There is weak evidence of local hiatuses and erosion of Upper Devonian units prior to or within the earlier history of the Rockhampton Group.

The Serpukhovian–Cisuralian unconformity interval includes the extensive Spring Creek Stratigraphic Assemblage and, in the north-east, the Berserker Group. The uppermost units of the Spring Creek Stratigraphic Assemblage are the Artinskian Yarrol Formation and overlying Owl Gully Volcanics, both apparently marine. These overlie an unreliably dated, possibly Serpukhovian–Artinskian, suite of marine and nonmarine facies, from east to west: the marine Lorray Formation, the almost entirely terrestrial Youlambie Conglomerate, and the marine Rookwood Volcanics. The latter contacts the easternmost units of the Bowen Basin (as most commonly defined).

In the western Yarrol Block, the hiatus separating the Spring Creek Stratigraphic Assemblage from the underlying Bindawalla Stratigraphic Assemblage is clear-cut. Distinctive granitoid-rich sediments, including boulder conglomerates, of the Youlambie Conglomerate sharply contact Upper Devonian units and, more commonly, lower parts of the Rockhampton Group. However, in the east, the Lorray Formation apparently overlies the Rockhampton Group without any obvious break in the Yarrol Syncline (Roberts & others, 1976, 1993), though the coincident rapid increase in radiometric response probably corresponds to the influx of granitoid detritus.

Poor biostratigraphic control precludes confident support for stratigraphic integrity of the Spring Creek Stratigraphic Assemblage. A conglomerate within the upper Lorray Formation in the Yarrol Syncline has been considered to represent a break spanning all or almost all of the Pennsylvanian (Jones, 1995, 1996: within the “Burnett Formation”). The change at the base of the Yarrol Formation is apparently abrupt both in rock type (Dear & others, 1971, page 40) and fossil content (Maxwell, 1964, page 7) and may betray an early-Cisuralian gap.

The Berserker Group includes Artinskian intervals, demonstrated by fossils correlative with the Buffel Formation of the Bowen Basin and by a radiometric date from volcaniclastics. Apparently intraformational lavas from the upper half of the Group, radiometrically dated late Cisuralian (or possibly early Guadalupian, depending on the time scale), are consistent with a top below the mid-Aldebaran (apparently early Guadalupian) hiatus of the Bowen Basin. Lower limits on the Group are poorly constrained. Contacts with the Spring Creek Stratigraphic Assemblage are not known but unconformities with the underlying Bindawalla Stratigraphic Assemblage have been interpreted with units as low as the Mount Alma Formation. The missing section is most easily attributed to the interval of uplift implicated by accumulation of the Youlambie Conglomerate (Spring Creek Stratigraphic Assemblage). However, the lack of obviously erosional facies in the lower parts of the Berserker Group suggests that at least some parts of the site of deposition were emergent or in some way above base level during this period. There is little basis on which to speculate on the duration of overlap, if any, of accumulation of the Berserker Group with the Spring Creek Stratigraphic Assemblage.

The uppermost unconformity interval of the Palaeozoic of the Yarrol Province is implied by tentative recognition, as the base of the interval, of a regional hiatus correlative with the mid-Aldebaran hiatus of the

Bowen Basin. The top of the interval is poorly defined by Triassic units. The interval includes the Dungarvan Stratigraphic Assemblage and, in the north-east, the Warminster Formation.

NEW PALAEOZOIC MICROPALAEONTOLOGICAL RESULTS

Almost all the results reported herein are conodont dates from large limestone bodies (See Tables 1 and 2 for summary age determinations, and lists of key taxa). The facies and structural complexity of the Yarrol Province preclude in general a fully confident interpretation of autochthoneity for such bodies. However, in any particular instances, unless there are specific reasons to further suspect the depositional context of these bodies, the stratigraphy implied by the dates is taken at face value.

The micropaleontological sampling conducted during the geological mapping followed a reconnaissance strategy more or less solely directed to address regional mapping issues. Many of these samples were needed to develop concepts of stratigraphic units. This was particularly crucial given the difficulty in subdividing the volcaniclastic-dominated Yarrol Province succession and the major revisions required to definitions of pre-Carboniferous stratigraphic units in particular, in order to equip the exercise with properly mappable units. Other reasons for sampling included testing for structural disruption or broad younging directions, establishing the full stratigraphic extent of units, and the delineation of the extent of possible hiatuses. In the collation of age determinations for a stratigraphic unit, only the minimal stratigraphic extent implied by the suite of age ranges was taken as the extent demonstrated for the unit. Of course, in lieu of detailed sampling along sections or precise results from boundary beds, some of the stratigraphic limits quoted herein may be too narrow.

Published conodont results from other sources have been incorporated into the results presented herein (Druce, 1969a,b,c; Webb, 1973, 1977; Mory & Crane, 1982; Jenkins & others, 1993; Mawson & others, 1995). This has been done by entering the taxa recorded by the source and interpreting the age significance of each sample individually — in conformity with the procedure used for the Geological Survey of Queensland collections. Thus age determinations for horizons from published sampled sections do not take into account the position of the horizon nor any interpreted zonal boundaries. In the case of the collections studied by Druce (1969a,b,c), new taxic determinations have been made by direct examination of the original collections. Age determinations made in the thesis of Little (1972) have been taken without revision and taxa have not been entered. Conodont collections from other relevant theses (Bradley, 1994; Simpson, 1995; Randall, 1996) have been incorporated by direct examination.

Time scales should, of course, always be considered problematic but the present state of the mid-Palaeozoic is in a particular state of flux. More at issue for the present purpose is the need for a time scale explicitly built from biozones in order that the evidence from biostratigraphic ages can be incorporated systematically, rather than merely assigned more or less arbitrarily. The only published mid-Paleozoic scale that satisfies this demand, Fordham (1992), is now well surpassed by subsequent results. However, the author has maintained the currency of this biostratigraphically based scale, and this has been applied herein.

Packages of the Palaeozoic stratigraphy of the Yarrol Province generally exhibit average accumulation rates of the order of 100m/my., barring major undetected hiatuses. This value is therefore applied to poorly controlled sequences to arrive at broad and tentative indications of stratigraphic position.

Summary details for units dealt with in this section are given in Table 3.

Click here for Table 1

Click here for Table 2

Table 3. Summary details for units with new micropaleontological results. Limits are minimum estimates (maxima for upper limits) based on known extents

Unit	**Lower limit**	**Upper limit**		
	Biozone	Chronostratigraphy	Biozone	Chronostratigraphy
Rockhampton Group	early *sulcata* Zone	very lower Tournaisian	early Late *Gnathodus bilineatus* Zone	latest Visean
Conglomerate at Santa Glen	–	Frasnian?	–	Famennian?
Mount Alma Formation	–	Frasnian?	early *sulcata* Zone	very lower Tournaisian
Three Moon Conglomerate	upper Lower *hassi* Zone–very lower Lower *rhenana* Zone +	lower Frasnian +	*Chonetes* Zone of Maxwell (1953, 1954) –	?upper Famennian –
Balaclava Formation	*Atrypa* cf. *desquamata kimberlyensis* occurrences	lower Frasnian?	middle Lower *duplicata* Zone – middle Lower *crenulata* Zone –	very lower Tournaisian
Lochenbar Formation	upper Upper *hassi* Zone–*jamieae* Zone +	lower Frasnian +, ?near Givetian–Frasnian boundary or upper Givetian	*Cyrtospirifer sinensis* var. *australis* Fauna of Dear (1968)	?middle or upper Famennian
Mount Hoopbound Formation	Upper *rhenana* Zone–*linguiformis* Zone +	very upper Frasnian +, ?near Givetian–Frasnian boundary	*Sulcatospirifer* Bed of Maxwell (1953) (top ?approx. = Upper *postera* Zone)	?upper Famennian
Marble Waterhole beds	Early *varcus* Zone +	early Givetian +	Late *varcus* Zone –	upper Givetian –
Ginger Creek Member	Middle *varcus* Zone– Upper *hermanni–cristatus* Zone	middle/upper Givetian	Middle *varcus* Zone– Upper *hermanni–cristatus* Zone –	middle/upper Givetian –
Raspberry Creek Formation	*kockelianus* Zone	upper Eifelian	Middle *varcus* Zone –	middle Givetian –, ?near Givetian–Frasnian boundary
Mount Warner Volcanics	*australis* Zone +	middle Eifelian +	*kockelianus* Zone	upper Eifelian
Craigilee beds	*Oulodus elegans detorta* Zone(?) +	upper Pridoli +	Lower *hermanni–cristata* Zone -	upper Givetian -
Erebus beds	*woschmidti* Z. +	upper Pridoli +	*costatus* Z. -	lower Eifelian -
Calliope beds	*hesperius* Z. +	Lockhovian, Ludlow? +	*kindlei* Z. -	middle Pragian -
Unnamed Lower/Middle Ordovician beds near Santa Glen	–	lower Floian +	–	lower Floian -

Unnamed Lower/Middle Ordovician beds near Santa Glen

This thin interval is interpreted as a faulted sliver between fault blocks containing the Calliope beds. Conodonts in the single thin lime-mudstone interval are apparently rare and quite meagre. They appear to be close to taxa from the lower Floian[1] of western Newfoundland (beds 9–11, Factory Cove Member, Shallow Bay Formation, Cow Head Group; Stouge & Bagnoli, 1988). This will need confirmation from follow-up taxonomic work.

Calliope beds

The lowest demonstrated part of the Calliope beds is an east-dipping sequence in a fault block near the western bank of Lake Awoonga south-east of Taragoola. A volcaniclastic interval, possibly 240m thick, is overlain by an approximately 200m-thick limestone interval. Limestone pods and marly beds are scattered through the uppermost parts of the volcaniclastic interval, demonstrating stratigraphic continuity (with regard to the final depositional site of the limestone) through to the limestone interval. The limestone pods are very close to the Silurian–Devonian boundary (BF801, 803); the top of the limestone unit, upper Lockhovian (BF863). A typical accumulation rate (see introductory section) would suggest that the volcaniclastics may extend down at least to the Ludlow. The easterly dipping fault block apparent immediately to the west similarly contains volcaniclastics followed to the east by mid- or late-Lochkovian limestones (GSMO0104, PBMO0001A, BF862), thus supporting this reconstruction of the lower Calliope beds and establishing thrust repetition with sheet thicknesses of the order of 0.5km .

Elsewhere — near Netherleigh (PBMO0098A), Bompa (PBMO0121A), and Yarwun (PBRO0611B), and in the western part of the Taragoola area (BF814) — only late-Pragian ages, with some slightly older or younger possibilities, are known. This is also the age of limestones from the eastern belt of the Mount Alma Formation: both the large outcrops in the Santa Glen area (BF850, 851, PBMO0090A) and limestone-clast conglomerates in the Yarwun area (PBRO0602). This supports both an upper limit of upper Pragian for the intact Calliope beds as well as the interpretation in the Yarwun area of an unconformable stratigraphic contact between the Calliope beds and the Mount Alma Formation that resulted from a destructively erosional event which tore up the upper parts of the Calliope beds. However, as no older limestones have yet been found among this interpreted allochthonous material of the Mount Alma Formation, these results are also consistent with an intact stratigraphic succession from the Calliope beds to the Mount Alma Formation across a boundary within the *kindlei* Zone (middle Pragian). In fact, the age ranges so far obtained, possibly coincidentally but also consistently, all support this: those from the Calliope beds could all be no younger than *kindlei* Zone, whereas those from the Mount Alma Formation may not be older than *kindlei* Zone. A similar conundrum for the interpretation of allochthonous sourcing of limestones applies to the interpretation of redeposition of Erebus beds into the western part of the Mount Alma Formation (see below).

There appear to be two discrete intervals of limestone in the Calliope beds: lower–middle Lochkovian and upper Pragian, though the relatively small number of zonal determinations and their ranges preclude confidence in this. In the absence of a reliable section through the known extent of the Calliope beds, the implied total known duration for the unit of approximately 10my would tentatively suggest a minimum thickness of the order of 1km, assuming typical accumulation rates for the Yarrol Block. This, together with the thicknesses of thrust sheets apparent in the south-east Taragoola area, would suggest that the present map underestimates the degree of faulting in domains of the Calliope beds.

The two westerly dipping thrust blocks demonstrated in the south-east Taragoola area may imply an east-over-west-directed imbricate thrust system, and this would be expected to be associated with the Yarrol Fault. However, the structure is complicated by a steep west-dipping contact between westerly volcaniclastics and easterly limestone in the western part of the Taragoola area (a quarry exposure at the previously mentioned site, BF814). This contact is considered a thrust (thus presumed a back thrust) as the limestone — upper Pragian — is assumed to be the upper part of the Calliope beds. Also, in the Santa Glen area the fault blocks are westerly dipping, and westward younging in the Mount Alma Formation is supported by the eastward position of allochthonous blocks(?) in each fault slice. This apparent west-over-east imbricate system may be due to anomalous geometries imposed by the westward translation (looking south) of the Yarrol Fault starting south-east of Santa Glen.

1 = The type lower Middle Ordovician, depending on future decisions of the Ordovician Subcommission of the International Stratigraphic Commission, IUGS

Erebus beds

The anticlinorial structure of the area occupied by the Erebus beds precludes identification of even a broadly representative section. As expected by this structure, the oldest date, middle to late Pridoli, comes from an axial area — near Deversoir (PBRO0634), consistent with McKellar's (*in* McKellar & others, 1970) macrofossils from Collection 615/7. Sites in line with 2–3km either side of Deversoir (in particular: ED/YAR103/1, 107/8, 105/2 to the west; BF763–8, 771, 774, 775 to the east) are mostly lower to middle Lochkovian. The occurrence of Pragian (and/or possibly Emsian) dates (BF763, 774) within the eastern belt may reflect the anticlinorial complexity.

A younger set of dates (BF783, 871–3, PBMO0048, PBRO0374, 376, 377, 412) all overlap in the late-Emsian *serotinus* Zone and, taking the anticlinorial structure into account, can all be plausibly placed in the uppermost parts of the unit, not far below the interpreted erosional contact with the Mount Alma Formation. Above this contact in the general region of the type area the limestone bodies interpreted to be allochthonous blocks yield ages similar to this younger set (BF776–782, 790, ED/YAR101/4, 151/4, MT&FG/Duncan02, 03/50m north, 07–13, 15, PBRO0379B, 430A, 545A, 623A–5A, 628A, 629A). However, although some of these confirm the *serotinus*-Zone age (BF782), not only do the ranges tend to be shifted subtly towards a younger average but also there are likely to definite indications of individual ages younger by one to three zones, that is, the *patulus* Zone and, crossing over to the early Eifelian, the *partitus* and *costatus* Zones (in particular: BF779, 780, MT&FG/Duncan02, 07, 08, 10, PBRO0379B, 430A, 545A). This apparent demonstration in the type region of the Erebus beds that the very uppermost parts of the unit were favoured sources for the megaconglomerate facies interpreted in the lower Mount Alma Formation is entirely consistent with such an interpretation. However, in this region not one of the ages from the interpreted allochthonous blocks is demonstrably older than any of the autochthonous(?) lenses from below the contact. This, then, is also consistent with a simple stratigraphic sequence across the interpreted erosional contact and, unless the rocks presently assigned to the lower Mount Alma Formation which host the interpreted allochthonous blocks are misleading, would imply more or less continuous Devonian accumulation in this region. This is a definite problem for the allochthonous interpretation, particularly as long as supporting megaconglomerate features associated with the interpreted limestone blocks or definite angular discordances at the erosional contact are lacking. In addition, blocks not clearly near the erosional contact (PBRO0379B, PBRO0629, and including the Marmor body, PBRO0545A) contain some of the youngest age indications — more easily accommodated by a simple stratigraphy. A conodont result supporting the allochthonous interpretation is the dating of the bodies at Mount Etna as Lochkovian (ED/YAR150/1; and possibly as young as middle Emsian, SBC/SCYL1070). However, this poorly outcropping area cannot be considered well understood.

Discrete limestone intervals are apparent in the Erebus beds: middle or upper Pridoli, lower Lochkovian (*eurekaensis* and *delta* Zones), upper Pragian or lower Emsian (within the range, *pireneae* to early *perbonus* Zones), and upper Emsian–lower Eifelian (*serotinus–costatus* Zones). Although the range uncertainties of the conodont dates may conceal further intervals, a small number of such intervals is apparent from the regional outcrop pattern of limestones in the type region. The duration for the known Erebus beds of approximately 28my tentatively suggests a thickness of the order of 2.8km (with typical accumulation rates for the Yarrol Block), though significant hiatuses separating thinner sequences, cannot be discounted.

Craigilee beds

Our formalisation of the undifferentiated Silurian–Devonian of the north-eastern Duaringa (Malone & others, 1969) and north-western Rockhampton (Kirkegaard & others, 1970) 1:250 000 Sheet areas as the Craigilee beds belies major difficulties which still remain in correlating the formation's several outcrop domains. The informal units used herein, SDc_1, SDc_2, and SDc_3, are largely based on reinterpretation of detailed mapping in the Morinish area by Bradley (1994) as well as regional mapping in the Mount Cassidy–Round Mountain belt by J. Domagala. The informal units can be broadly recognised in each of easterly dipping Lower Devonian–Middle Devonian sequences apparent in the Mount Cassidy, Bannockburn, and Morinish areas. However, in detail the correlations are unclear and point to major changes in relative thicknesses of individual units over now quite closely located domains, if not significant differences in lithostratigraphic sequence which make formational identity somewhat suspect. This is compounded by more structurally complex but poorly outcropping domains elsewhere, including the Armagh area where Silurian dates may or may not indicate lower intervals.

Unit SDc_1 at Armagh. The several limestone bodies distributed across strike east of Armagh all yield dates which are consistent with an upper-Pridoli level (especially JDF022, 23, 297), though conodont dating in the very upper Silurian is somewhat imprecise. Although ranges in these dates could disguise younging between these bodies, the stratigraphic thickness which this would imply seems excessive if recent estimates of the duration of the Pridoli (Tucker & others, 1998,, Fordham, 1998) are accepted[2]. These bodies are therefore presumed to be repetitions. If this is due to faulting and this assumed minimal, the widest belt (JDF023–JDF298) would be consistent with a thickness of approximately 700m, above or below the limestone interval.

Craigilee beds in the Morinish area. The stratigraphy in this area has been reconstructed by combining Bradley's (1994) Fault Block 3 and 1 and adding a fault in the eastern part of Fault Block 1; his correlation with Fault Block 2 is left unchanged. This places his Facies in the following Lower Devonian–Middle Devonian sequence: Unit SDc_1 (3A/2A/1F–3B/2B/1G–3C/2C/1H–3D/2D–3E/2E), Unit SDc_2 (1A), and Unit SDc_3 (1B–1C–1D–1E). The Morinish area is disrupted throughout by faults and bedding is moderate to steep and variable. Therefore the sections measured by Bradley (1994, Appendix C) must be considered provisional (below, the thicknesses quoted are from his original sections through Facies 1A–[lowermost] 1H, but the reconstructed stratigraphy is Facies 1F–1H, 1943–2992m, followed, after an unmeasured interval corresponding to Facies 3C/2C–3E/2E, by Facies 1A–1E, 0–1943m).

Unit SDc_1. The known sequence of SDc_1 appears to be of the order of at least 1.5km thick (Bradley's section for Facies 1F–1H amounts to 1049m but lacks Facies 2C–2E which appear to be very thick in the northern part of the area). Its lower part contains a limestone-dominated interval (Facies 3B/ 2B/1G) 300–400m thick (Bradley, 1994, p.11) starting in the middle or upper Lockhovian (*eurekaensis* Zone–*pesavis* Zone, lower Facies IB: GSMO0034, 35, 37, 39). This is underlain by poorly outcropping volcaniclastics that have a minimum thickness of 784m in the eastern part of the area (Bradley, 1994, appendix C, Facies 1F, 1943–2727m). The upper part of the limestone-dominated interval reaches the lower-Emsian *excavatus* Zone (either side of the contact of Facies 1B–1C [BBY/Cono_A, GSMO1098] and of Facies 3B–3C [GSMO1097]). This interval is then overlain by quartzphyric volcanics (Facies 3C) suggestive of but presumably too low for the Eifelian (only?) Mount Warner Volcanics of the Capella Creek Group. The remainder of Unit SDc_1 is presumed to be at least 0.5km thick (see above). A limestone in its upper part yielded late-Pragian or early-Emsian conodonts (*kindlei* Zone–*dehiscens* Zone, Facies 2E, GSMO0043). Within the present mapping this must be allochthonous and indicative of uplift, presumably in the mid-Emsian.

Unit SDc_2. This unit lacks limestone. It is 810m thick (Bradley, 1994, Appendix C, Facies 1A, 0–810m) and contains abundant volcanics. Bradley (1994) considered the SDc_1–SDc_2 contact (Facies 3E–1A) a thrust. Also, he interpreted the SDc_2–SDc_3 contact (Facies 1B–1C) as an unconformity based on discordant strike trends, though a structural explanation (or, even, volcanic doming?) seems equally possible.

Unit SDc_3. The known sequence of SDc_3 is 1133m thick (Bradley, 1994, Appendix C, Facies 1B–1E, 810–1943m). Limestones associated with conglomerates, from possibly a single interval close to the known top, yield a range of dates in the middle to late Givetian (over at least the interval Early *varcus* Zone – Early *hermanni–cristata* Zone, GSMO0045, 46B, BBY/Cono_D). The age range is surprising, but could be ascribed to different intervals displaced by faulting, to redeposition, or, to a minor extent, to problems with taxic ranges.

Craigilee beds in the Bannockburn area. **Unit SDc_1.** Near the Melrose–Glenroy road intersection, a number of limestone bodies occur in close proximity within volcaniclastics and in places exhibit intergrading with the latter. The several dates are consistent with a *delta*-Zone level but a longer interval through the Lochkovian is also possible. This is therefore considered an intact Lochkovian sequence equivalent to the lowermost limestones in the Morinish area. In the area east to Ten Mile Creek, this interval may be overlain by a considerable thickness (> 1km?) of mostly thin-bedded volcaniclastics, though the latter are likely to be relatively incompetent rocks subject to considerable structural disruption. In the transition interval there are distinctive ribbon interbeds of limestone and siltstone. Similar bedding occurs in calcsilicates in the south-eastern Morinish area (GSMO0516: Facies 1G, Fault Block 1 of Bradley, 1994) suggesting a correlation despite the difficulty of dating at this level of metamorphism.

Two limestone bodies in the southern part of the Bannockburn area form a younging series up section (GSMO1135, early *delta* Zone?; GSMO0014, *kindlei* Zone–*dehiscens* Zone) and so may be in place.

Unit SDc_3. As presently mapped, this area lacks Unit SDc_2 and so the SDc_1–SDc_3 sequence may, in fact, be two fault blocks. The eastern, uppermost, part of the Unit SDc_3 includes apparently autochthonous middle-Givetian limestone horizons (for example, GSMO1083, Middle *varcus* Zone?) commonly associated with calcareous sandstones and limestone-clast conglomerates, one of which has yielded a similar date (GSMO0073).

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2 a short duration for the Pridoli (2.7my) has since been adopted by the IUGS time scale (Gradstein & others, 2004)

A limestone-clast conglomerate with an upper-Pragian or lower-Emsian clast (JDF282, *kindlei* Zone–*gronbergi* Zone) in the south-west of the SDc_3 domain is not easy to assimilate. Given the relative proximity of the site to the middle-Givetian horizons, it is presumed to be sourced from SDc_1 limestones which were uplifted and remained exposed in SDc_3 time. Alternatively, it may be an SDc_1 intraformational conglomerate and point to problems in the subdivision used for the Craigilee beds.

An intriguing fault block in the north-eastern Bannockburn area contains small poorly outcropping upper-Givetian limestone bodies (lenses?) (Lower or Upper *hermanni–cristata* Zone, GSMO0076). Here, contact with siltstones typical of the Upper Devonian Mount Alma Formation is covered, but a stratigraphic contact—telling for the contentious Middle Devonian–Upper Devonian regional hiatus—is possible.

Craigilee beds in the Mount Cassidy area. North from Craigilee there are at least a few structural blocks containing the Craigilee beds. A domain near Muldoon Creek which cuts across the meridional regional strike contains lower-Lochkovian limestones with volcaniclastics [*hesperius* Zone–very low *eurekaensis* Zone(?), JDF303, 304, GSMO0053] and so may provide overlap between the Armagh and Bannockburn/Morinish sequences.

The large fault block to the north at Hillview lacks biostratigraphic control. However, its lithostratigraphy is a plausible base for the adjacent block to the east containing an apparent SDc_2–SDc_3 sequence. The SDc_2 domain here includes an upper-Emsian limestone body (*serotinus* Zone, JDF276) of uncertain stratigraphic context. The overlying SDc_3 volcaniclastics include laterally extensive thin lime-mudstone horizons, apparently devoid of macro- and micro-fossils (JDF279, GSMO0056), which may be correlative with the uppermost (middle Givetian) part of Unit SDc_3 at Bannockburn (for example, GSMO1083). However, not far below these beds in the Mount Cassidy area are limestone bodies and conglomerates with dates (JDF251, 272) similar to the SDc_2 limestone body. The associated volcaniclastics, especially in the lower parts of Unit SDc_3 include distinctive volcanic breccias with apparently freshly deposited clasts, and so could represent direct continuation of the more-proximal volcanism of Unit SDc_2. At present the SDc_2–SDc_3 succession in the Mount Cassidy area is taken as a very rapidly accumulated largely upper-Emsian sequence. However, until the clearly authochthonous limestone beds are dated, there is an equal possibility that this SDc_3 sequence is Givetian and that an hiatus caused by uplift and erosion separates it from SDc_2, thus supporting the unconformity proposed by Bradley, (1994) in the Morinish area.

Immediately south of the Hillview fault block is quite a thick limestone body (GSMO0015), cutting across strike, which may belong to Units SDc_1 or SDc_2.

East of the main Silurian–Middle Devonian belt there are probably one or more thin fault slices (GSMO0021) within the Bindawalla Stratigraphic Assemblage succession.

Summary, Craigilee beds. The Craigilee beds are a Pridoli–upper-Givetian volcaniclastic and volcanic sequence, over 4km thick, with limestone intervals variously developed in at least the very upper Pridoli–upper Lochkovian, lower Emsian, very upper Emsian, and middle–upper Givetian. However, there may have been one or more periods of uplift and erosion, possibly in the mid-Emsian or within part of the very latest Emsian–early Givetian. In addition, there appear to be major variations in thicknesses of units even in adjacent fault blocks. The inclusion, therefore, of this entire Silurian–Middle Devonian succession within the concept of Talent & others (1982) for the Craigilee beds, centred originally on the Armagh limestones, may prove too encompassing.

Mount Morgan Stratigraphic Assemblage

Capella Creek Group

Mount Warner Volcanics. Limestone apparently occurs as discontinuous lenses in a few horizons in the middle to upper parts of the Mount Warner Volcanics (see Messenger & Taube, 1994, figure 3). At the Upper Nine Mile Creek prospect in the Dee Range there are limestones in the “crenulated cherty tuff” and “jasperoid” units of the Banded Sequence in the middle part of the formation (Taube & McLeod, 1987; = Banded Mineralised Sequence of Taube & Messenger, 1994). A limestone probably equivalent to the latter horizon occurs at the Ajax prospect (Taube & Messenger, 1994). There is also a limestone horizon at the base of the Banded Mine sequence in the Mount Morgan mine (Taube, 1986; = Jasper Sequence of Messenger & Taube, 1994). The Eifelian age of the latter reported by Palmieri (in Taube, 1986, page 1325) was subsequently refined as *costatus* Zone–*kockelianus* Zone (middle Eifelian) and matched with that of the Nine Mile Creek limestones (Fordham in Messenfger & Taube, 1994; Fordham & Taube, 1994). Alexei Taube (letter, 24 February 1983) also found limestones in the upper part of the Mount Warner Volcanics: in the Hanging Wall Sequence of Taube & McLeod (1987) at Upper Nine Mile Creek. These may be equivalent to limestones of the Upper Mine Pyroclastics of the Mount Morgan mine (Taube, 1986; = Upper Mine Sequence of Taube & Messenger, 1994).

Subsequent sampling has now further firmed and refined the correlation of the Banded Mine Sequence at Mount Morgan with the Banded Mineralised Sequence in the Dee Range. Fordham's original correlation was partly based on finding "aff. *Polygnathus trigonicus* Bischoff & Ziegler, 1957 of Klapper, 1971 (late form of Sparling, 1983)" — the taxon now attributed to Palmieri's original conodont from Mount Morgan (Geopeko/D. D. H. 38/8, 262' 08"–263' 06") — also in the "jasperoid" of the Banded Mineralised Sequence in the Dee Range (Geopeko/1982/03). This confirmed the level of the Mount Morgan ore and the equivalent subeconomic mineralisation in the Dee Range as mid-Eifelian (*costatus* Zone–*kockelianus* Zone), most probably *australis* Zone. *Tortodus australis* has now been recovered from subsurface near the mine (Perilya/PD37A, 227–228m) confining the maximum age range for the ore level to *australis* Zone–*kockelianus* Zone. A little higher in the same drill-hole (215–6, 214–5m), difficult-to-assign taxa are best accommodated by a date of *kockelianus* Zone. The jasper unit of the Banded Mine/Mineralised Sequence thus most likely lies within the *kockelianus* Zone and may start in the immediately underlying *australis* Zone.

From the slightly higher level of the Hanging Wall Sequence in the Dee Range (Geopeko/1982/13), Fordham had originally recovered "Polygnathus sp. B of Sparling, 1983", indicating *australis* Zone–*kockelianus* Zone. Further results from samples probably at this level now make a level of *kockelianus* Zone very likely: *T. kockelianus* and a possible intermediate, *T. australis*–*T. kockelianus*, from Limestone Creek (ATLC1), and *T. australis* from nearby (PBRO0127).

In summary, the level of prime mineralisation (Banded Mine/Mineralised Sequence) in the jaspers of the middle Mount Warner Volcanics, including the Mount Morgan ore, as well as the immediately overlying volcaniclastics of the upper part of the formation (Upper Mine / Hanging Wall Sequence) very probably lies in the *kockelianus* Zone (upper Eifelian). The mineralised interval may start at least as low as the immediately underlying *australis* Zone (middle Eifelian) but controls on the lower limits of the interval are absent as are controls from the lower part of the formation or the underlying Mount Dick beds.

Raspberry Creek Formation. An interval with minor limestones extends from the middle part of the underlying Mount Warner Volcanics to the lower part of the Raspberry Creek Formation. This lower part of the Raspberry Creek Formation lies within the same conodont zone, *kockelianus* Zone, as the upper Mount Warner Volcanics. This is demonstrated by several dates taken in combination: those no younger than this level [Geopeko/1982/10 with *Polygnathus* aff. *benderi*; Geopeko/1982/11, *Tortodus australis*; ?PBRO0248, *Icriodus* (?)*augustus*]; one no older (PBRO0235A, *T. kockelianus*).

Effective controls for the remainder of the formation are confined to near and west of the Dee Range where an upper contact is absent and a substantial portion of the upper part of the formation may be lost from exposed synclinal cores. Even where the upper contact remains, as it is west of the Mount Morgan Trondhjemite, the incorporation of debris from the latter in conglomerates near the base of the overlying Mount Hoopbound Formation point to the loss at least in parts, near the Middle Devonian–Late Devonian boundary, of the thickness of Raspberry Creek Formation which sat above the intruded Trondhjemite. This would be expected to be of the order of 0.5–1km as the intrusion is likely to have been relatively high-level.

In this Dee Range area there are not as yet any precise results from the minor limestones in the middle of the sequence. A prominent limestone horizon well exposed in the headwaters of Capella Creek north of Beschs Hill from near the upper part of this sequence is, from a combination of results, in the *ensensis* Zone (very upper Eifelian): BF830 is no higher; PBRO0272C and 237A are no lower. Higher limestones from the core of the syncline in the area of the headwaters of Capella Creek, from immediately east and south of the Ginger Creek fault block, and from near Marble Mountain may also come from a single interval. If so, it would be no lower than and probably confined to the Middle *varcus* Zone (for example, PBRO164A, 236, 239B, 245B, 1001C, ATSO101, 179, 293).

These two limestone intervals, most probably in the *ensensis* and Middle *varcus* Zones, may correspond with those in the Station Creek synclinal core (Hayward & others, 1999), though dates from the latter (respectively: MHRO248; MHRO035D and 036) are not precise enough to confirm this.

Ginger Creek Member. This felsic facies is best exposed in a fault block bordering Ginger Creek. The Member may be a lateral equivalent of the upper part of the main, intermediate, sequence of the Raspberry Creek Formation preserved nearby, given that that the two dates (PBRO217B, 240A) from the Member include the Middle *varcus* Zone in their ranges. It is, however, also possible that the facies is higher in the Givetian (no higher than Upper *hermanni*–*cristatus* Zone, based on PBRO217), especially as the dated limestones are from the lower part of the admittedly thin (100m?) sequence demonstrable in the fault block. If so, the block could be an up-sequence interval displaced down-dip in the core of the syncline well evidenced to the north in the area of the headwaters of Capella Creek.

Summary of Capella Creek Group. The lower parts of the Group are undated: 540m of Mount Dick beds; and 200m of the Footwall Tuff (Taube & McLeod, 1987; = Footwall Sequence of Taube & Messenger, 1994) in the lower part of the Mount Warner Volcanics (see Messenger & Taube, 1994, figure 3 for sections).

Much of the remaining 600m (at Upper Nine Mile Creek)–1900m (Mount Morgan mine) of the Mount Warner Volcanics, as well as the very lower part of the overlying Raspberry Creek Formation, may lie within the *australis* and *kockelianus* Zones (middle–upper Eifelian). There are probably four main, though thin and discontinuous, limestone horizons through this interval. The second lies in the jasper-dominated part of the Banded Mine/Mineralised Sequence containing the Mount Morgan ore.

The best-controlled sequence of the Raspberry Creek Formation, from near the headwaters of Capella Creek, is difficult to measure but is estimated to be 1350m thick. Almost all of this (1200m?) appears to span the *kockelianus* Zone–*ensensis* Zone (upper Eifelian). It includes minor limestones in the lower part and a significant horizon in the *ensensis* Zone.

Thus, the >2km of upper Mount Warner Volcanics to middle? Raspberry Creek Formation appears to have accumulated over only approximately 3my (*australis* Zone–*ensensis* Zone). This very rapid volcaniclastic accumulation followed the deposition of the Mount Morgan ore.

The upper Raspberry Creek Formation (>300m preserved?) is difficult to reconstruct. It includes a limestone horizon probably in the Middle *varcus* Zone (middle Givetian). The Ginger Creek Member may be at this level or higher (its base could be as high as the Upper *hermanni–cristatus* Zone). Either way, it seems that the rapid accumulation regime of the middle–upper Eifelian was slowed in the early–middle Givetian, barring a significant lower-Givetian gap in the sequence.

An extra approximately 0.5–1km would apparently need to be added to the preserved sequence to account for the dissection of the Mount Morgan Trondhjemite prior to deposition of the Mount Hoopbound Formation. This uppermost part of the unit, which may be preserved west of the Mount Morgan Trondhjemite, may be very upper Givetian and/or very lower Frasnian.

Marble Waterhole beds

Although a number of dates, with a minimal span of Early *varcus* Zone–Late *varcus* Zone (early–middle Givetian) are now available for this limestone-rich unit, complex faulting has prevented reconstruction of a sequence. Even a measured section at Marble Waterhole appears to contain contradictory dates: Middle *varcus* Zone in the “lower” part (MHRO762, 791) and *ensensis* Zone or Early *varcus* Zone in the “upper” part (MHRO789, 788). However, the conodont taxa recovered from the measured section are not considered reliable enough indices to clearly indicate repetition within the section. The Late *varcus* Zone date comes from limestones near Allen Creek.

The available dates may give a misleadingly narrow impression of the stratigraphic range occupied by the formation as there is little basis for integrating sequences lacking limestones.

Bindawalla Stratigraphic Assemblage

Mount Hoopbound Formation

Conodonts from Trotters Creek described by Druce (1969b; YAR126) remain the only firm biostratigraphic control on this unit. The date of Late *rhenana* Zone–*linguiformis* Zone (very late Frasnian) is based on Klapper’s (1990, p.1005) retention of Druce’s (1969b) identification of *Ancyrognathus asymmetricus* (this would supersede Druce’s, 1974, pages 6–7, revision of the date to early Frasnian). The few small limestone bodies at YAR126 are poorly outcropping and so of dubious stratigraphic context. They are, however, along strike from a persistent horizon (Hawkins & Whitcher, 1962) of fossiliferous calcareous sandstone and limestone-clast conglomerate which includes limestone bodies most easily interpreted as lenses (for example, at MHRO0139). The limestone material is thus tentatively considered intraformational and generally indicative of the age of final deposition. The structure in the area of Druce’s locality is complicated by faulting (Nimmo, 1992) and so the most reliable indication of the stratigraphic position of the site is probably by correlation with the fossiliferous horizon. This horizon is considered most likely to correspond with the conglomerate 610m above the base of the Formation in the composite section of Randall (1996, figure13, level 620m; R. E. Randall, personal communication, 2000).

The only other conodont evidence comes from a limestone-clast conglomerate lower in the formation (RRRR035A/04, 420m above base of Formation, 430m level, above mentioned section). The *Polygnathus* specimen is transitional *P. dubius*–*P. webbi* — if the respective taxic diagnoses are interpreted literally — and would point to an age in the general time of the overlap in the ranges of these taxa (Early *asymmetricus* Zone [ie, falsiovalis Z.—transitans Z.]: very early Frasnian). This is, however, a dubious basis for accurate age assignment.

Further controls on the formation could come from Maxwell's (1953, 1954) brachiopod assemblages from the "Thomson Clastics" and "Boulder Creek Grits" but these assemblages are yet to be reliably tied to modern zonations. Thus the stratigraphic limits of the formation must still be considered poorly known. Simplistic interpolation of unit thicknesses from the area south-west of Mount Morgan would, however, provide broad estimates — based on a very approximately estimated total thickness approximately of 1870m and controls furnished by only the Trotter Creek date in the lower third of the formation and the dated oolite from low in the overlying Rockhampton Group ("Neil's Creek Clastics" , see below), and barring significant changes in accumulation rates and major hiatuses. These interpolations suggest the base of the unit to be close to the Givetian–Frasnian boundary and the top, at the contact with the Balaclava Formation, to be ? up the Famennian. This would also produce estimates — of dubious reliability — for Maxwell's units: "Thomson Clastics" / *Cyrtospirifer* bed, middle Famennian (in the general vicinity of Upper *marginifera* Zone–Uppermost *marginifera* Zone); and "Boulder Creek Grits" / *Sulcatospirifer* Bed, middle–upper Famennian (in the general vicinity of Lower *trachytera* Zone–Upper *postera* Zone).

Lochenbar Formation

Conodont controls for the Lochenbar Formation broadly parallel those available for the Mount Hoopbound Formation. The measured sequence south of Kroombit Tourist Park, which starts at a faulted contact, contains a few limey volcaniclastic intervals over the interval 750–950m. In the lower parts of this interval these limey beds form limestone lenses with a characteristic nodular texture which may be a boundstone formed by a tabulate coral or stromatoporoid. This level is very probably lower Frasnian (late Late *hassi* Zone–*jamieae* Zone: MHRO0435A/2), though fully reliable index taxa are absent. Given that the older date from conglomerates in the lower Mount Hoopbound Formation (RRRR035A/04, 420m above the base of the Formation, see above) is highly speculative, the MHRO0435A/2 date is the oldest useful date from the Bindawalla Stratigraphic Assemblage. This date also suggests differences in timing or average rates of accumulation in the early (dated) deposition of the Lochenbar Formation and the Mount Hoopbound Formation. If accumulation for both units started broadly coevally — near the Givetian–Frasnian boundary — then the Lochenbar Formation underwent much higher average rates (approximately: >770m in <2my, versus 610m in 6my for the Mount Hoopbound Formation). Alternatively, the Lochenbar Formation may have begun with only a very minor or without any break, following the last (?Late *varcus* Zone) Marble Waterhole beds (allowing some 4my for this lower part of the section), and still sustained higher average accumulation rates. Given the uncertainty in these estimates, the top of the exposed Lochenbar Formation at Kroombit Creek (approximately 2km in total) can only be suggested as reaching somewhere in the upper Frasnian–Famennian.

Limestones from south-west of the Callide Dam (MHRO0682, 684A) appear to be at a similar level to those just discussed (MHRO0435A/2) from the lower part of the section south of Kroombit Tourist Park.

The only other key conodont dates from the Lochenbar beds are from the ?100m-thick limestone in a domain trending across regional strike in the western part of the Kroombit Creek belt. Only the late-Frasnian result from near the top of this limestone (late Early *rhenana* Zone–middle Late *rhenana* Zone(?): MHRO0382C) is considered reliable[3]. This could be at the same general level as the Trotter Creek limestone of the Mount Hoopbound Formation (YAR126; see above).

Balaclava Formation

Base. The base of this unit appears to be highly diachronous. In the Glengowan area south-south-east of Mount Morgan it apparently is laterally equivalent to the entire sequence of the Mount Hoopbound Formation and unconformably overlies typical Raspberry Creek Formation and, along strike to the south-east, its Ginger Creek Member. It was in this area that McKellar & others (1970, page 146, Collections 43/2, 4) reported *Atrypa* cf. *desquamata kimberlyensis* from the lower, presumed herein Frasnian, part of the Balaclava Formation (their "undifferentiated Upper Devonian"). This fossil is common at this level and so its abundance may be tentatively taken as a Frasnian indicator. It is not clear if the unqualified *A. d. kimberlyensis* reported by McKellar & others (1970, page 146, Collection 294) from the Raspberry Creek Formation at the Station Creek headwaters (also placed by them in their "undifferentiated Upper Devonian") represents a different biostratigraphic marker.

3 Sample A from near the base contains *Icriodus? chojnicensis,* a middle-Famennian index taxon from Sandberg & Dreesen's (1984) icriodid zonation. This taxon, however, has unclear origins and so is of dubious reliability. Given that the top of this limestone from the western part of the Kroombit Creek area is probably upper Frasnian, this occurrence may be a rare test (a failure) of the icriodid scheme.

The mid-Famennian conodont dates from the Alma Creek area south-south-west of Mount Morgan (Middle *crepida* Zone–Late *crepida* Zone, MHRO0094D; late Late *rhomboidea* Zone–early Late *marginifera* Zone, MHRO0117) are from limestones which grade into volcaniclastics, apparently from near the base of the unit overlying, in that area, the Mount Hoopbound Formation.

West of Mount Morgan in the classic area mapped by Maxwell (1953, 1954), the Balaclava Formation also overlies the Mount Hoopbound Formation but relative thicknesses of the units there, combined with only a few conodont controls from above and below the Balaclava Formation, tentatively point to a base for the latter in the upper Famennian, approximately near the base of the Lower *expansa* Zone. In this area this basal portion of the unit, in which Maxwell placed his *Tenticospirifer* Bed, is now confirmed to contain the *Tulcumbella tenuistriata* Zone (Locality 273-4; see Appendix 3).

The base of the Balaclava Formation thus appears to be highly diachronous — basal Frasnian to upper parts of the Fammenian. Similarly, apart from the apparently highly abrupt lateral passage of the unit into the Mount Hoopbound Formation in the Grasstree Creek area, the top of the underlying Mount Hoopbound Formation appears to vary from lower Famennian in the Alma Creek area to upper Famennian west of Mount Morgan.

Middle. The middle of the Balaclava Formation contained, in Maxwell’s (1953, 1954) account of the area west of Mount Morgan, in ascending stratigraphic order, his *Avonia* and *Chonetid* Beds, both included in his *Chonetes* Zone. Roberts (see Appendix 3) has confirmed the occurrence of the former Bed from the Mount Battery Block (273-3) and the latter Bed from the Golden Gully Block (GSMO0543A, B), but has not found the two Beds in sequence. Roberts, in support of McKellar (1967, page 29). has been able to confirm the *Chonetid* Bed as belonging to the *Tulcumbella tenuistriata* Zone and, contrary to Maxwell’s stratigraphic ordering of these two Beds, considers that the *Avonia* Bed may be a higher level than that of the *Chonetid* Bed. In relation to the collections from west of Mount Morgan, Roberts considered the collection from south of Mount Morgan in the Grasstree Creek area (PBRO0438; Collection 46 of McKellar & others, 1970, page 146) to be most comparable with Maxwell’s *Avonia* Bed, thus supporting Dear’s (1968, page 10) record of a collection characteristic of this Bed from “blue cherty siltstones” from apparently the same site.

Top. The base of the Rockhampton Group appears to be relatively synchronous throughout most of the Yarrol Block and to be close to or a little above the Devonian–Carboniferous boundary (see below). This seems to be the case west of Mount Morgan (see Rockhampton Group) where the Balaclava Formation is overlain by the Rockhampton Group and the top of the Balaclava Formation is probably close to the Devonian–Carboniferous boundary.

In the area of the headwaters of the Don River south of Mount Morgan, the Balaclava Formation is unconformably overlain by the Upper Carboniferous–Lower Permian Youlambie Conglomerate. Here the Balaclava Formation includes oolitic limestones, apparently indistinguishable from those of the Rockhampton Group, one of which is lower Tournaisian (middle Lower *duplicata* Zone–middle Lower *crenulata* Zone: PBRO0494). Thus at its southern extent the Balaclava Formation is correlative with at least the lower part of the Rockhampton Group.

Three Moon Conglomerate

The only conodont date from this unit — from a limestone-clast boulder conglomerate near Dooloo Creek in the eastern part of the unit’s extent (late Early *hassi* Zone–very early Early *rhenana* Zone: IDR/QP0726) — closely matches the early-Frasnian date from the limestone in the lower part of the Lochenbar Formation. The Dooloo Creek conglomerate is presumed to have an intraformational source and to similarly be in the lower part of the unit.

Dear (1968, page 10) considered fossils from approximately 100m below the contact with the Rockhampton Group (his “Cania Formation”) at Cania (GSQL2250) to be correlative with sandstones of the Balaclava Formation at Grasstree Creek interbedded with siltstones correlative with Maxwell’s *Avonia* Bed (see above). Thus the very upper part of the Three Moon Conglomerate is most likely very upper Famennian.

Mount Alma Formation

The base of the unit is lower Frasnian, based on newly discovered corals and brachiopods. Apart from the large limestone bodies interpreted as allochthonous (see below), the only limestones known from the unit are lime-mudstones, mostly occurring as thin lenses in flysch. East of Mount Alma such limestones (PBRO0341) appear to lie in the middle of that part of the unit exposed in an anticlinal fault block. These contain fragmentary Upper Devonian conodonts that are most likely Frasnian.

Upper-Emsian – lower-Eifelian limestone bodies in the lower part of the western belt of the unit near contact with the Erebus beds are considered allochthonous, as are the Pragian limestone bodies in the eastern belt, although there are problems with these interpretations (see under Erebus beds and Calliope beds, respectively). Limestone bodies near Mount Etna of unclear stratigraphic position within the Mount Alma Formation include Lochkovian intervals.

The top of the unit is well controlled in the Cannindah Creek area where there are a few lime-mudstone horizons in the uppermost part. The uppermost lime-mudstone (GSMO1102A) directly contacts oolite (GSMO1102B) at the base of the overlying Rockhampton Group and most likely is no higher than lower *sulcata* Zone. A lime-mudstone horizon 50m stratigraphically below this is most likely Upper *praesulcata* Zone. There are several other lime-mudstone sites in the Mount Alma Formation which are probably very upper Famennian (Craigilee area, JDF287; near Stanmore, GSMO0519, 1069; Takilberan Creek, GSMO0457; near Weitalaba, PBMO0129). However, in the upper-Morinish area, horizons within metres of an apparent stratigraphic contact with the Rockhampton Group are middle Famennian [Early *marginifera* Zone–Late *marginifera* Zone (?), GSYB240; late Late *crepida* Zone, GSYB287]. This may be evidence of local hiatuses at the Mount Alma Formation–Rockhampton Group contact. There is, in fact, evidence for erosion of the Mount Alma Formation during the very early history of the Rockhampton Group — from middle Famennian lime-mudstone clasts in conglomerates not far above the base of the Group 30 km south-west of Calliope (Late *marginifera* Z.–Late *velifer* Z., =?Late *trachytera* Z., GSMO0393A//02).

Conglomerate at Santa Glen

Conodonts from adjacent but separate boulders in a limestone-boulder conglomerate at Santa Glen are Upper Ordovician and upper Emsian respectively. Possible Ordovician trilobites have been found nearby. As there is not any evidence that the Calliope beds are higher than Pragian, inclusion of this conglomerate with the nearby volcaniclastics is suspect. The upper-Emsian age suggests provenance in the Erebus beds and the redeposition may warrant comparison with allochthonous blocks interpreted for the basal Mount Alma Formation in its western belt.

The Upper Ordovician boulder is a unique remnant of that interval in the Yarrol Block. The conodonts are moderately abundant and diverse and appear broadly typical of Ashgill collections from eastern Australia (Palmieri, 1978, 1984; Zhen & others, 1999). This will need confirmation from follow-up taxonomic work.

Rockhampton Group

Conodonts from the limestone intervals of the Rockhampton Group have been well documented across the mapping area by Mory & Crane (1982) and Jenkins & others (1993), who respectively covered: from near the base of the unit (near the base of the Tournaisian) to the upper part of the Tournaisian; and the uppermost Tournaisian to the uppermost-sampled limestones near the top of the unit (upper Visean). Apart from spot samples for various regional purposes, a relatively large number of Rockhampton Group sites were sampled for conodonts during this mapping to test sequence-stratigraphic correlations away from the more detailed sampled sections of Mory & Crane (1982) and Jenkins & others (1993). These conodont results are incorporated into the section on the Rockhampton Group by Simpson & others in the main report.

The most reliable evidence for the age of the base of the Rockhampton Group is considered to come from Cannindah Creek where lime-mudstones of the Mount Alma Formation directly contact oolite of the Rockhampton Group without indicators of significant stratigraphic gaps (see “Mount Alma Formation”, above, for other details). These lime-mudstones are most likely early *sulcata* Zone (earliest Tournaisian). Conodonts were not recovered from the contacting oolite, though the next higher oolite horizon nearby is early Early *crenulata* Zone (GSMO1103). The remaining evidence across the mapping area is not inconsistent with a regionally more or less synchronous base for the Rockhampton Group within the *sulcata* Zone. However, a few limestone sites near the base of the Rockhampton Group suggestively include taxa which are also consistent with the very latest Famennian *praesulcata* Zone (PBMO0035A and 038A from 30km south-west of Calliope; GW97/7/136 from 1.5m above the base of the type Gudman Oolite west of Rockhampton, Webb, 2005). Also, there is evidence of a hiatus with and erosion of the underlying Mount Alma Formation (see “Mount Alma Formation”, above), which indicates a disrupted, possibly significantly diachronous, start to accumulation of the Rockhampton Group.

The highest dated Rockhampton Group in the mapping area appears to be best demonstrated by limestones at the top of the unit near the contact with the Lorray Formation in the Baywulla area (Jenkins & others, 1993, table 4): probably the early (Visean) part of the Late *Gnathodus bilineatus* Zone, which straddles the Visean–Serpukovian boundary. Conodont sampling of calcareous horizons and probably reworked limestone bodies around the boundary between the Rockhampton Group and Lorray Formation at the east limb of the Yarrol Syncline was largely unsuccessful, possibly reflective of the temperate paleolatitude by this time. However, in this area a probably reworked oolite of latest Tournaisian or Visean age (BF842), grades into calcareous sandstones and granule conglomerates just within the radiometrically hot Lorray Formation (a thrust fault has been mapped at the boundary, but the stratigraphic succession appears to be intact). This points to a number of possible regimes around the Rockhampton Group–Lorray Formation boundary, including significant erosion, possibly starting before the cessation of oolite limestone deposition of the Rockhampton Group.

In the Bukali area north of Monto, limestones near the contact with the Youlambie Conglomerate are early Tournaisian (MHRO0803, 804, JW08495668, JW09025583), suggesting either that the Early Carboniferous Rockhampton Group paleobasin quickly ceased accumulation, or that the Permian erosion was deeply incisive in this region.

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APPENDIX 3
DEVONIAN–CARBONIFEROUS SHELLY FOSSILS

(J. Roberts)

Lochenbar Formation

MHRO1098 (35km east-south-east of Mount Scoria, GR292199 7277794)

Cyrtospirifer reidi Maxwell
?Atrypids, possibly two forms
Athyrid

Age: Frasnian.

MHRO1108L (near Lochenbar homestead 30km east of Biloela, GR276662 7293061)

Spinatrypina (Spinatrypina) prideri prideri (Coleman)
Desquamatia (Synatrypa) kimberleyensis (Coleman)
Tenticospirifer sp. - smaller than *T. grandis* Maxwell
Fenestella sp.
Solitary rugose coral

Age: Frasnian.

MHRO1108N

Spinatrypina (Spinatrypina) prideri prideri (Coleman)

Age: Frasnian.

Balaclava Formation

PBRO620 (Centre Creek 2.5km north-east of Glengowan homestead, GR253765 7355191)

Desquamatia (Synatrypa) kimberleyensis (Coleman)
Atrypid - ornament similar to large specimen from MHRO1108L
Spiriferoid - small fine ribbed , narrow, pointed posteriorly, rounded anteriorly.
Nervostrophia bunapica Veevers

Age: Frasnian.

PBRO265 (east of Grasstree Creek 4km east of Glengowan homestead, GR255854 7352804)

Tenticospirifer sp. Same as at MHRO1108L
Gastropod fragments

Age: Late Devonian, probably Frasnian.

PBRO479B (13km south-south-east of Glengowan homestead, GR262855 7345483).
Lowest fossil locality in the Balaclava Formation in the Gunpowder Creek area.

Cyrtospirifer sp.
Rhynchonelloid indet
Spiriferoid?

Age: Late Devonian, but the material is so poorly preserved because of the coarse grainsize that it is difficult to say more.

PBRO 491 (15km south-east of Glengowan homestead, GR262959 7343812). At a similar level to conodont sample PBRO494B , three-quarters of the way up the Balaclava Formation. The conodont zone is early *duplicata* Zone - middle to early c*renulata* Zone.

Schizophoria sp.
Rhipidomella australis (McCoy)
Schuchertella sp.
Leptagonia analoga (Phillips)

Rugosochonetes magnus (Maxwell)
Spinocarinifera kennedyensis (Maxwell)
Productina globosa Roberts
Brachythyris davidi (Dun)
Crassumbo kennedyensis (Maxwell)
Prospira prima Maxwell
Prospira typa = *Unispirifer cf. striatoconvolutus* of Qian
Spirifer sp. cf *S. sol* but with much narrower sinus
Kitakamithyris sp. (not *rouchelensis* - spines more numerous)
Cleiothyridina segmentata Roberts
?Schumardella sp.
Hamburgia? hillae Qian
Mitchellinia sp.

Age: Earliest Tournaisian, *Tulcumbella tenuistriata* Zone.

MHRO063 (14km east-north-east of Dululu, GR233570 7366820)

Tulcumbella tenuistriata (Maxwell)
Spinocarinifera kennedyensis (Maxwell)
Rugosochonetes kennedyensis (Maxwell)
Prospira typa/morganensis

Age: Earliest Tournaisian, *Tulcumbella tenuistriata* Zone.

GSMO 538 (10km south-west of Mount Morgan, GR227320 7375722)
See Figure 1 for relative stratigraphic position.

Tulcumbella tenuistriata (Maxwell)

Age: Earliest Tournaisian, *Tulcumbella tenuistriata* Zone.

GSMO543A (8km south-south-west of Mount Morgan, GR226656 7378961). *Chonetes* horizon of Maxwell, same as GSQL983. See Figure 1 for relative stratigraphic position.

Tulcumbella tenuistriata (Maxwell)
Austrochoristites solidus (Campbell & Engel) according to Qian
Hamburgia? hillae Qian
Angustispatulata campbelli Qian

Age: Early Tournaisian *Tulcumbella tenuistriata* Zone.

GSMO543B
Just beneath GSMO543A. See Figure 1 for relative stratigraphic position.

Tulcumbella tenuistriata (Maxwell)
Retichonetes kennedyensis (Maxwell)
Rugosochonetes magnus Maxwell
Spinocarinifera kennedyensis (Maxwell)
Brachythyris cf. *davidi* (Dun) but more narrow
Cleiothyridina sp.
?Shumardella sp.
Hamburgia? hillae Qian

Age: Early Tournaisian *Tulcumbella tenuistriata* Zone.

GSMO 543F
Same locality as GSMO543A and GSMO543B

Tulcumbella tenuistriata (Maxwell)
Austrochoristites solidus (Campbell & Engel)
Spinocarinifera kennedyensis (Maxwell)
Hamburgia? hillae Qian

Age: Early Tournaisian *Tulcumbella tenuistriata* Zone.

PBRO438 (Grasstree Creek 2km south-south-east of Glengowan homestead, GR253589 7352456)

Schizophoria sp.
Retichonetes kennedyensis (Maxwell)
Rugosochonetes magnus Maxwell
Acanthocosta cf. teicherti Roberts (*A. teicherti* is an early Tournaisian species from the Bonaparte Basin)
Spiriferoid with reduced dental plates, flat sinus bordered by bifurcating costae (adults may have obsolete costae in sinus), simple lateral costae, short mucronate hinge, compact and subquadrate in outline. Unlike any I have seen before.
Brachythyris davidi (Dun)
Austrochoristites solidus (Campbell & Engel)
Crassumbo kennedyensis (Maxwell)
Crurithyris sp.
Cleiothyridina round type
Shumardella sp.
Angustispatulata campbelli Qian
Terebratuloid

Age: Early to middle Tournaisian. Younger than *Tulcumbella tenuistriata* Zone and older than *Schellwienella burlingtonensis* Zone.

MHRO1091 (9km north-north-east of Dululu, GR223200 7368400)

Retichonetes kennedyensis (Maxwell)
Rugosochonetes magnus Maxwell
Pharcidodoscus boulderensis Roberts
Productoid - brachial valve interior more like *Spinocarinifera* than *Pharcidodiscus*.
Shumardella sp.
?Cleiothyridina sp.
Unispirifer typa/morganensis - sinal costae like one from Grasstree Ck *Angustispatulata campbelli* Qian

Age: Probably Early Tournaisian. Younger than *Tulcumbella tenuistriata* Zone and older than *Schellwienella burlingtonensis* Zone.

PBRO489 (14km south-east of Glengowan homestead, GR261989 7343812). Top of Balaclava Formation, overlying conodont sample PBRO494B, and 50m below Youlambie Conglomerate. The conodont zone was early *duplicata* Zone - middle to early c*renulata* Zone.

Schizophoria sp.
Syringothyris australis Maxwell
Ribbed spiriferoid
Cleiothyridina sp.
Orthotetoid ornament
Solitary coral
Mitchellinia sp.

Age: Tournaisian. Younger than *Tulcumbella tenuistriata* Zone - either *Schizophoria* or *Schellwienella burlingtonensis* Zones.

GSMO 539A (10km south-west of Mount Morgan, GR227082 7375163)
See Figures 1 and 2 for relative stratigraphic position.

Retichonetes kennedyensis (Maxwell)
Rugosochonetes magnus Maxwell
Spiriferoid - Pro/Unispirifer
Brachythyris davidi (Dun)
Cleiothyridina segmentata type
Hamburgia? hillae Qian
Angustispatulata campbelli Qian

Age: Tournaisian

GSMO 540 (10km south-west of Mount Morgan, GR227070 7375212)
See Figures 1 and 2 for relative stratigraphic position.

Schizophoria sp.
Rugosochonetes magnus Maxwell
Unispirifer morganensis (Maxwell)
Brachythyrid - elongate
Brachythyrid - larger, ovate
Crassumbo kennedyensis (Maxwell)
Cleiothyridina segmentata type
Angustispatulata campbelli Qian
Hamburgia? hillae Qian

Age: Tournaisian

GSMO 541 (10km south-west of Mount Morgan, GR227141 7375340)
See Figures 1 and 2 for relative stratigraphic position.

Retichonetes kennedyensis (Maxwell)
Rugosochonetes magnus Maxwell
?Brachythyris davidi (Dun)
Cleiothyridina sp.
Hamburgia? hillae Qian
Angustispatulata campbelli Qian

Age: Tournaisian

GSMO 542 (10km south-west of Mount Morgan, GR 227497 7375167)

Nautiloids, plant fragments, possible productid.

PBRO596A (8km north-west of Mount Morgan, GR228007 7387990)

Schizophoria sp.
Rugosochonetes magnus Maxwell
?Rugauris sp. Could be what Dear (1963) called *Pustula dubia*. Dear's name cannot be used.
Brachythyrid. May be the same as *Brachythyris* sp. ? cf. *pinguis* of Maxwell and *B. caniensis* in Dear (1963). Dear's name cannot be used.
Prospira typa/morganensis (single pedicle valve)
?Prospira prima Maxwell (2 half valves of simple-ribbed spiriferoid, not lamellose)
Cleiothyridina australis Maxwell
Trilobite

Age: Probably Tournaisian. A precise age requires more material for all species. If the productid (*?Rugauris*) and the brachythyris are the same as those from Dear (1963), the age would be mid-late Tournaisian.

GSQL946 (53-3) (West bank of Boulder Creek, GR226500 7378900)

Schizophoria sp.
?Schuchertella sp.
Rugosochonetes magnus Maxwell
Spinocarinifera kennedyensis (Maxwell) Qian considers this to be *Pharcidiscus boulderensis*, but I am doubtful
Prospira typa Maxwell
Crasumbo kennedyensis (Maxwell)
Cleiothyridina sp.
Crurithyris sp.

Age: Tournaisian.

GSQL956 (5) (West of McBride Creek, Gelobera Range, GR241700 7361300)

Schizophoria sp.
?Schuchertella sp.
Spinocarinifera kennedyensis (Maxwell)

Rugosochonetes magnus Maxwell
Unisprifer morganensis (Maxwell) finer ornament than type locality
Crasumbo kennedyensis (Maxwell)
Angustispatulata campbelli Qian

Age: Tournaisian

Northern traverse of Mount Battery Block, west of Mount Morgan

Locality **273-1**
See Figure 2 for relative stratigraphic position.

Schizophoria sp.
Spinocarinifera kennedyensis (Maxwell)
Retichonetes kennedyensis (Maxwell)
Rugosochonetes magnus Maxwell
Cleiothyridina

Age: Early Tournaisian, *Tulcumbella tenuistriata* Zone.

(NOTE Localities 273-1 and 273-4 (supposed site of the *Tenticospirifer* Zone of Maxwell) may be in float from an adjacent creek.)

Locality **273-2**
See Figures 1 and 2 for relative stratigraphic position.

Rugosochonetes magnus Maxwell
C*leiothyridina*
Hamburgia? hillae Qian
Angustispatulata campbelli Qian

Age: Early Tournaisian, no diagnostic species of a zone.

Locality **273-3** = L1288 of Maxwell (1954)
See Figures 1 and 2 for relative stratigraphic position.

Schizophoria sp.
Schellwienella sp.
Tulcumbella tenuistriata (Maxwell)
Retichonetes kennedyensis (Maxwell)
Rugosochonetes magnus Maxwell
Spinocarinifera kennedyensis (Maxwell)
Plectospira simplex Qian
Crassumbo kennedyensis (Maxwell)
Unispirifer morganensis (Maxwell)
Spiriferoid, less thickend around muscle field than *Prospira typa*
Cleiothyridina segmentata Roberts
?Crurithyris sp. small. narrow sinus, 2 low plicae on each lateral slope, fine striate micro-ornament
Hamburgia? hillae Qian
Angustispatulata campbelli Qian

Age: Early Tournaisian, *Tulcumbella tenuistriata* Zone.

Locality **273-4**
See Figures 1 and 2 for relative stratigraphic position.

Tulcumbella tenuistriata (Maxwell)
Spinocarinifera kennedyensis (Maxwell)
Ribbed spiriferid

Age: Early Tournaisian, *Tulcumbella tenuistriata* Zone.

Three Moon Conglomerate

CMM726 (dam near head of Callide Creek 7.5km south-east of Mount Seaview, GR295675 7307344)

Tenticospirofer grandis Maxwell
Cyrtospirifer reidi Maxwell

Age: Famennian

Mount Alma Formation

PBRO584 (3km east of The Caves, GR242461 7434935)

Spinatrypa sp.
Stromatoporoid

Age: Pre-Famennian, probably Frasnian.

GSMO531 (4km west-south-west of Bindawalla homestead, GR292752 7318219)

Retichonetes kennedyensis (Maxwell)
Pharcidodiscus boulderensis Roberts
Productid cf *Acanthocosta*
Austrochoristities solidus (Campbell)
Crurithyris sp.

Age: Early Tournaisian. Younger than *Tulcumbella tenuistriata* Zone, possibly equivalent to Grasstree Creek locality or *Schizophoria* Zone of Maxwell.

Rockhampton Group

GSMO544A (9km south-south-west of Mount Morgan, GR225582 7379438). Equivalent to Maxwell's *Schizophoria* Zone and approximately equivalent to GSQL985 and 273-6. See Figures 1 and 2 for relative stratigraphic position.

Schizophoria sp.
Leptagonia analoga (Phillips)
Rhipidomella australis (M'Coy)
Schuchertella oversbyi Qian
Retichonetes kennedyensis (Maxwell)
Rugosochonetes magnus Maxwell
Pharcidodiscus boulderensis Roberts
?Fluctuaria sp.
Productid indet, spinose, not as ribbed as at PBRO438; ?*Acanthocosta*
Brachythyris davidi (Dun)
Brachythyris sp.
Crassumbo kennedyensis (Maxwell)
?Austrochoristites solidus (Campbell)
Prospira prima Maxwell
Prospira/Unispirifer ?typa
Plectospira simplex Qian
Crurithyris sp.
Cleiothyridina round species
Shumardella sp.
Angustispatulata campbelli Qian
Hamburgia? hillae Qian
Pseuarietities? ammonitiformis (Etheridge)
Protocanites careyi Campbell, Brown & Coleman

Age: Early-mid Tournaisian, younger than *Tulcumbella tenuistriata* and older than *Schellwienella burlingtonensis* Zones.

GSQL955 (previously Neils Creek Clastics) (Divide between Oaky and Boulder Creek, GR226500 7374900)
See Figure 1 for relative stratigraphic position.

Schizophoria sp.
Leptagonia analoga (Phillips)
Rugosochonetes magnus Maxwell
Pharcidodiscus boulderensis Roberts
?Fluctuaria sp.
Prospira prima Maxwell
Prospira typa Maxwell
Brachythyris sp. of Maxwell (1954, pl.4, fig.13a, b from UQ L1289)

Age: Tournaisian.

GSMO545 (west of Boulder Creek 9km south-south-west of Mount Morgan[=GSQ L987], GR225191 7379028).
Equivalent to Maxwell's *Cleiothyridina* Zone. See Figure 1 for relative stratigraphic position.

Angustispatulata campbelli Qian
Tylothyris planimedia Cvancara (Spiriferid - wide, lamellose, 11-12 rounded 'plicae' on lateral slope; striate radial micro-ornament in simple sinus; reasonably high ventral interarea.)
Crassumbo kennedyensis (Maxwell)
Rugosochonetes magnus (Maxwell)
Angustispatulata campbelli Qian
Spicules - probably a sponge - an almost complete individual is preserved
Gastropods, bivalved molluscs

Age: Tournaisian

GSMO574C (700 m west of New Cannindah homestead 13km north-north-east of Monto, GR321771 7254042)

Rugosochonetes sp.
Productina globosa Roberts
Crassumbo kennedyensis (Maxwell)
Tylothyris sp. more coarsely ribbed than *planimedia* Cvancara
?Prospira typa Maxwell
Cleiothyridina australis Maxwell

Age: Middle Tournaisian. Younger than *Tulcumbella tenuistriata* Zone and older than *Schellwienella burlingtonensis* Zone.

GSMO 600B (Spring Creek west of Malakoff homestead 21.5km north of Monto, GR305868 7275852)

Bivalved molluscs

GSMO637 ('Fossil Hill' 8km south of Ridgelands in equivalents of the Malchi Formation overlying a Gudman Oolite equivalent, GR222896 7418139)

Schizophoria sp.

Age: Probably Carboniferous.

GSMO 652 (Callide Dam 12km west-south-west of Bindawalla homestead, GR283900 731600)

Rhipidomella australis (McCoy)
Schizophoria sp.
Leptagonia analoga (Phillips)
Schuchertella oversbyi Qian
Rugosochonetes magnus (Maxwell)
Productina globosa Roberts
Pharcidodiscus boulderensis Roberts
Acanthocosta cf. teicherti Roberts
?Prospira typa Maxwell
Spirifer sol (Campbell & Engel)

Spiriferoid similar to one at 273-3 (and at PBRO438 Grasstree Creek, though latter is more convex on slopes of pedicle valve). Hinge wider than long, costae simple, sinus deep with flattish floor and steep sides, possibly weakly costate (not well preserved) though internal moulds lack costae in sinus, ventral muscle field pointed. dental plates short, restricted to margins of delthyrium, ventral interarea flat and triangular.
Brachythyris cf. elliptica Roberts
??Acuminothyris cf. triangularis Roberts (similar to *Tylothyris* but has greater number of ribs)
Tylothyris sp. same species as at 574C
?Syringothyris sp.
Crurithyris sp.
Cleiothyridina segmentata Roberts
Athyris sp.
Angustispatulata campbelli Qian
?Shumardella sp.

Age: Tournaisian, probably *Schizophoria* Zone, though the *Brachythyris* cf *elliptica* and ornament that could belong to *Schellwienella burlingtonensis* would suggest a late Tournaisian age. The productids suggest a slightly older age.

GSMO1003A and B (7km west-north-west of Mount Morgan, GR227357 7385849)

Productoid indet.
Smooth spiriferoid indet.

Age: Indeterminate.

MHRO783 (on ridge above alluvial plant on Cania Road, GR289858 7276978). = L2254 of Dear (1968).

Retichonetes kennedyensis Maxwell
Rugosochonetes magnus Maxwell
Schellwienella burlingtonensis Weller
Linoproductid, long trail on pedicle valve, few rugae
Productina globosa Roberts
?Rugauris sp. = *Pustula dubia* (*nomen nudum* in Dear, 1963)
Spirifer sol Campbell & Engel
Prospira prima Maxwell
Spiriferoid, smooth sulcate
Brachythyris sp. = *boolgalensis* (*nomen nudum* in Dear, 1963)
?Crassumbo sp.
Cleiothyridina australis Maxwell
Angustispatulata campbelli Qian
Pectenid
Solitary coral
Crinoid plates
Trilobite

Age: Mid-late Tournaisian, *Schellwienella burlingtonensis* Zone, probably the lower part of the zone.

MHRO1067 (30km north-east of Biloela, GR273300 7315300)

Spirifer sol Campbell & Engel

Age: Tournaisian.

Northern traverse of Mount Battery Block, west of Mount Morgan

273-5 (GR224823 E, 7378668N.)
Equivalent to Maxwell's *Spirifer* Zone. See Figure 1 for relative stratigraphic position.

Schellwienella burlingtonensis (Weller)
Retichonetes kennedyensis (Maxwell)
Rugosochonetes magnus Maxwell (larger, wider than lower in sequence)
Prospira prima Maxwell
Cleiothyridina round species
Angustispatulata campbelli Qian

Age: Mid-late Tournaisian. *Schellwienella burlingtonensis* Zone.

273-6 (below limestone at small house, GR225548E 7378488N).
Same horizon as GSMO544A. See Figure 1 for relative stratigraphic position.

Schizophoria sp.
Leptagonia analoga (Phillips)
Rhipidomella australis (M'Coy)
Schuchertella oversbyi Qian
Retichonetes kennedyensis (Maxwell)
Rugosochonetes magnus Maxwell
Pharcidodiscus boulderensis Roberts
?Fluctuaria sp.
? Acanthocosta sp.
Productid indet, spinose, not as ribbed as at PBRO438
(*Acanthocosta* cf. *teicherti*)
Brachythyris davidi (Dun)
?Austrochoristites sp. narrow, well ribbed
Crassumbo kennedyensis (Maxwell)
Prospira prima Maxwell
Prospira typa Maxwell
Spiriferid indet. large internal
Crurithyris sp.
Cleiothyridina cf. *australis* Maxwell
Angustispatulata campbelli Qian
Hamburgia? hillae Qian
Ammonoid - *?Ammonellipsites*

Age: Early-mid Tournaisian, younger than *Tulcumbella tenuistriata* and older than *Schellwienella burlingtonensis* Zones.

Lorray Formation

GSMO1106 (north of Cannindah Creek 21.5km east-north-east of Monto, GR330433 7256742). Shell rich calcareous sandstone on eastern limb of Yarrol Syncline. Previously mapped as Dakiel Formation (McKellar, 1967).

Alispirifer sp.
Composita magnicarina Campbell
Kitakamithyris cf. imensa (Campbell)
Neospirifer ornament
Myonia pollocki Maxwell

Age: Namurian or possibly younger. Probably *Levipustula levis* Zone. *Compositia magnicarina* is restricted to the *L. levis* Zone in NSW, as with *K. imensa*, and *Myonia pollocki* is known elsewhere from the lower Rands Formation (now included in the Lorray Formation).

GSMO1109 (approximately 50m higher in sequence than GSMO1106)

Levipustula levis Maxwell
Alispirifer contractus Maxwell
Spiriferellina neerkolensis Maxwell
Spinuliplica spinolosa Campbell
Limipecten multistriatus Maxwell

Age: Namurian. *Levipustula levis* Zone.

MHRO1001B (4km north-west of Mount Salmon, GR196127 7429420)

Levipustula levis Maxwell

Age: Namurian.

REFERENCE

MAXWELL, W.G.H., 1953: Upper Palaeozoic formations in the Mount Morgan district – stratigraphy and structure. *Papers Department of Geology University of Queensland*, **4**(4).

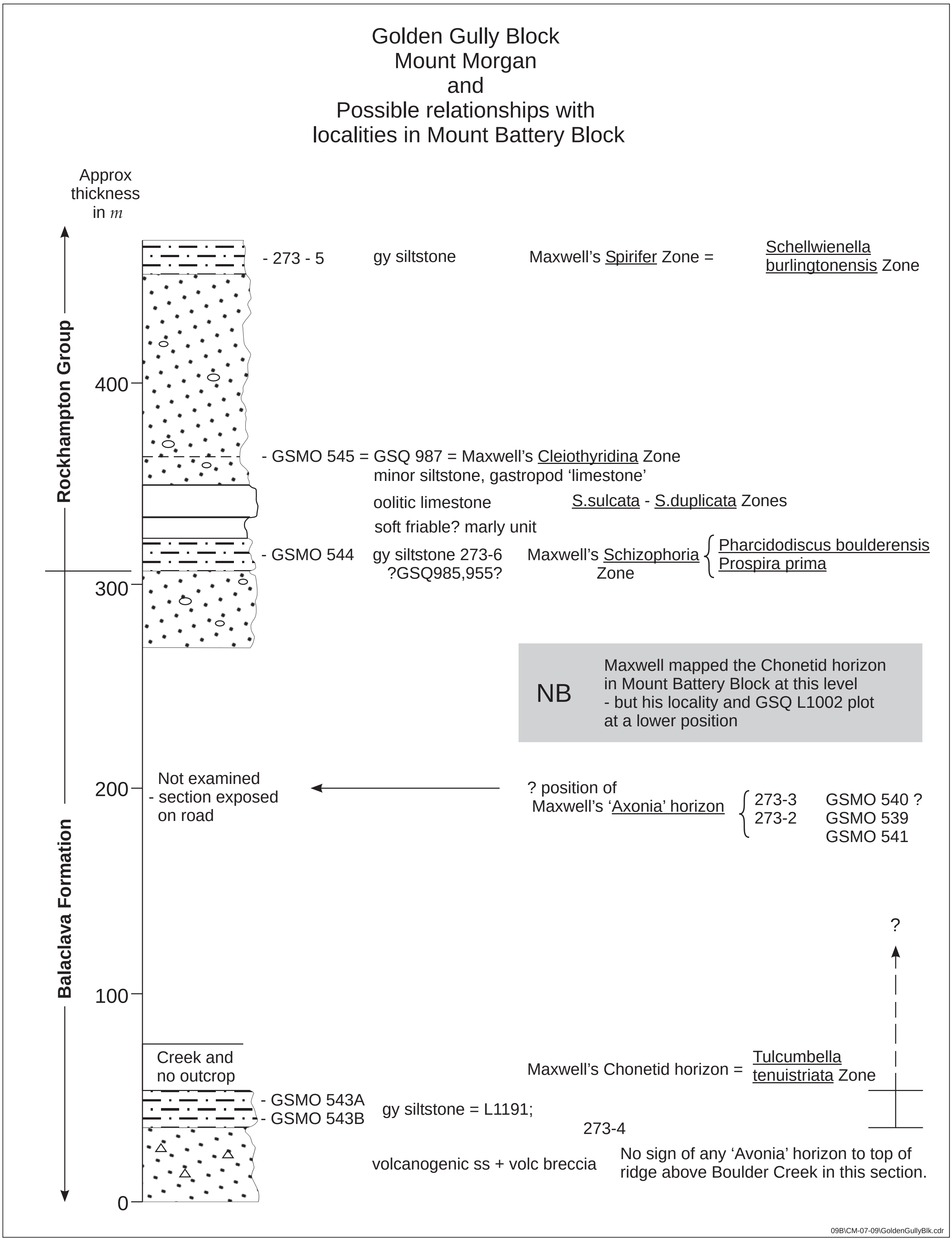

Figure 1. Lithological log of traverse line across the Golden Gully Block, as mapped by Maxwell (1953) west of Mount Morgan, showing the relative positions of fossil localities and the fossil zones defined by Maxwell.

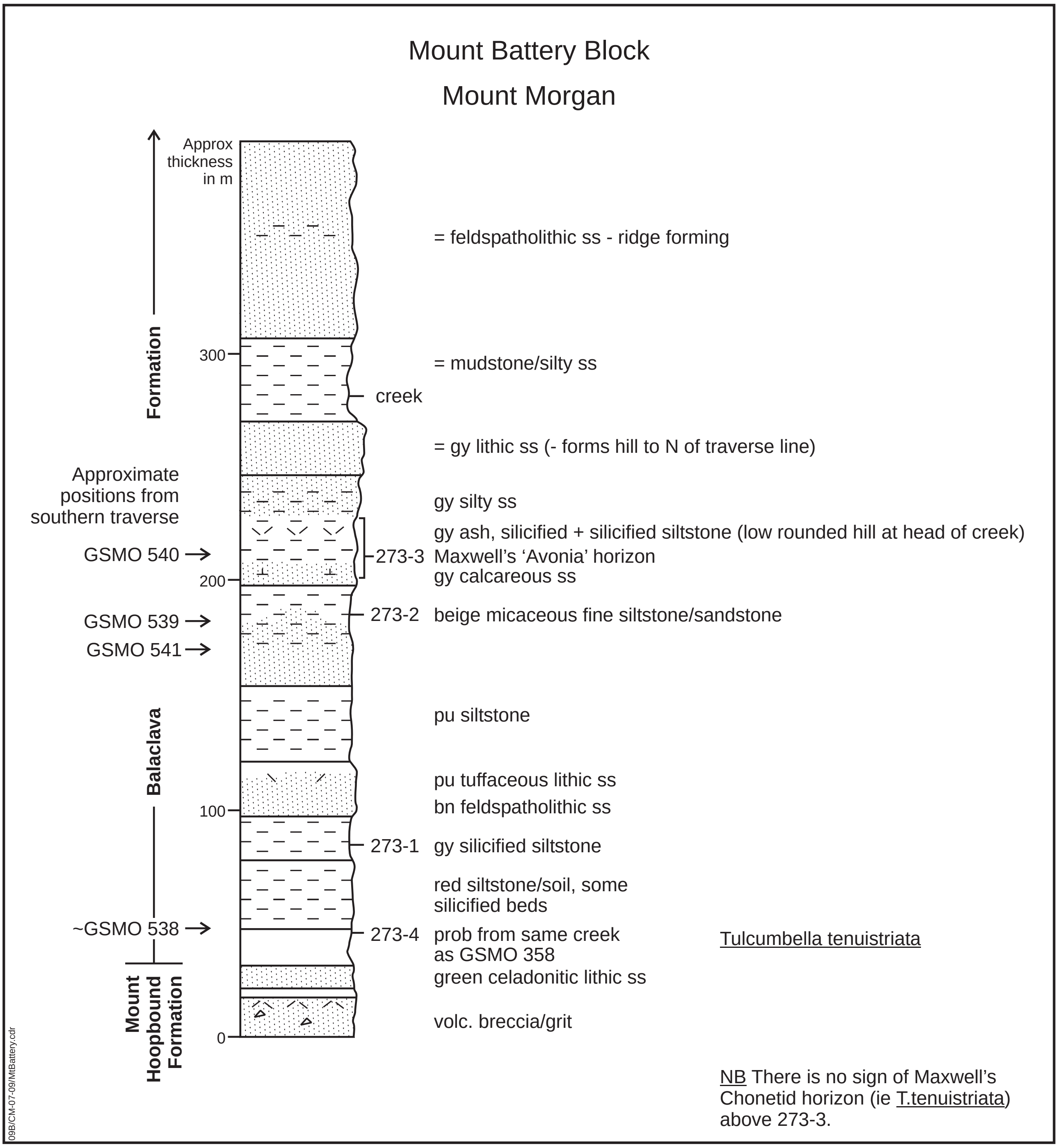

Figure 2. Lithological log of traverse line across the Mount Battery Block, as mapped by Maxwell (1953) west of Mount Morgan, showing the relative positions of fossil localities and the fossil zones defined by Maxwell.

APPENDIX 4
DEVONIAN CORALS

(P. Blake)

This appendix lists the coral species found in the Devonian units in the Yarrol Province. It is based on the PhD study of Blake (2005) which was published as an AAP Memoir (Blake, 2010).

Following the species names are the Location Numbers where that species is found. In brackets after the official Location Number is the original Field Point number for collections made during the latest phase of mapping.

Calliope beds

Aphyllum simplexum Blake, 2009- GSQL4156 (PBMO001).
Rhizophyllum parvum Strusz, 1970- GSQL4156 (PBMO001).
Favosites forbesi Milne-Edwards & Haime, 1851 - GSQL4156 (PBMO001).
Favosites. grandipora Etheridge, 1890 – GSQL4156 (PBMO001).

Erebus beds

Aphyllum simplexum Blake, 2009 - UQL3523.
Pseudamplexus princeps (Etheridge, 1907) – UQL3523.
Blysmatophyllum sp. A – UQL2472.
Mictocystis? sp. cf. *M. endophylloides* Etheridge, 1908 – GSQL4173 (PBRO500) and UQL3523.
Dohmophyllum sp. A – GSQL4158 (PBMO048).
Embolophyllum sp. A – UQF41852.
Papiliophyllum jelli Blake, 2009 – GSQL4173 (GSQ Raglan), GSQL4173 (PBRO500), UQL2427, UQL2475, UQL3523 and UQL3524.
Papiliophyllum sp. A – UQL3524.
Thamnophyllum sp. A – UQL2473.
Favosites careyi Jell & Hill, 1970a – GSQL4173 (PBRO500).
Favosites duni Etheridge, 1920 – GSQL4173 (PBRO500).
Favosites goldfussi d'Orbigny, 1850 – GSQL4158 (PBRO048) and GSQL4173 (PBRO500).
Favosites gothlandicus Lamarck, 1816 – GSQL4173 (PBRO500).
Favosites grandipora Etheridge, 1890 - UQL2475.
Squameofavosites bryani (Jones, 1937) – GSQL4158 (PBRO048), GSQL4173 (PBRO500), UQL3524 and UQL3523.
Heliolites daintreei Nicholson & Etheridge, 1879 – GSQL4173 (PBRO500) and UQL3524.
Heliolites amplusa Blake, 2009 – GSQL4173 (PBRO500) and UQL3523.
Heliolites comminus Blake, 2009 – GSQL4158 (PBRO048) and GSQL4173 (PBRO500).
Romingeria sp. cf. *R. emergens* (Quenstedt, 1881) – UQL3523 and UQL3524.
Syringopora jonesi Jell & Hill, 1970b – UQL3524.

Craigilee beds

Tryplasma wellingtonense Etheridge, 1895 – GSQK72 and GSQL4134 (JDF25).
Aphyllum simplexum Blake, 2009- GSQL4052 (GSMO016), GSQL4134 (JDF25), GSQK72
Microplasma ronense (Mansuy, 1913) – GSQL4052 (GSMO016).
Pseudamplexus princeps (Etheridge, 1907) - GSQL4129 (GSMO012).
Xystriphyllum insigne Hill, 1940 – GSQL4131 (GSMO489).
Chostophyllum gregorii (Etheridge, 1892b) – GSQL4176
Favosites goldfussi d'Orbigny, 1850 – GSQL4052 (GSMO016), GSQL4129 (GSMO012), GSQL4130 (GSMO036), and GSQK72.
Favosites grandipora Etheridge, 1890 - GSQK72, GSQL4052 (GSMO016), and GSQL4130 (GSMO036).
Favosites sp. A – GSQL4052(GSMO016).
Squameofavosites bryani (Jones, 1937) – GSQL4052 (GSMO016), GSQL4129 (GSMO012), GSQL4130 (GSMO036).
Squameofavosites nitidus (Chapman, 1914) – GSQL4129 (GSMO012).
Squameofavosites craigileei Blake, 2009 – GSQK72, GSQL4052 (GSMO016) and GSQL4129 (GSMO012).
Thamnopora boloniensis (Gosselet, 1877) – GSQL4129 (GSMO012) and GSQL4052 (GSMO016)
Heliolites daintreei Nicholson & Etheridge, 1879 – GSQK72, GSQL4052 (GSMO016), and GSQL4129 (GSMO012).
Heliolites comminu Blake, 2009 – GSQL4052 (GSMO016), GSQL4129 (GSQL012), GSQL4130 (GSMO036),

and GSQL4176 (Mount Cassidy Diggings).
Syringopora jonesi Jell & Hill, 1970b –GSQL4129 (GSMO012).

Mount Warner Volcanics

Tabulophyllum sp. A – UQL1143.
Temnophyllum stainesi Blake, 2009 – UQL1143.
Phillipsastrea? carinata Hill, 1942b – UQL1149.
Favosites goldfussi d'Orbigny, 1850 – GSQL4163 (PBRO128).
Alveolites suborbicularis Lamarck, 1801 – UQL1143.

Raspberry Creek Formation

Tryplasma? abyssum Blake, 2009 – GSQL4167 (PBRO182).
Calceola sp. -
Cystiphylloides immanum Blake, 2009 – GSQL3832 (PBRO273), GSQL4136 (MHRO035), and GSQL4160 (PBRO024).
Pseudomicroplasma australe (Etheridge, 1892b) – GSQL4167 (PBRO182)
Breviphyllum mirourum Blake, 2009 – GSQL4170 (PBRO237)
Metriophyllum? sp. A – UQL1153.
Barrandeophyllum sp. A – UQL1144.
Sanidophyllum davidis Etheridge, 1899 – GSQL4169 (PBRO218), GSQL4170 (PBRO237), and UQL1142
Tabulophyllum sp. A – GSQL1
Acanthophyllum (A.) clermontense (Etheridge, 1911) – GSQL4167 (PBRO182).
Dohmophyllum clarkei Hill, 1942a – GSQL1, GSQL4167 (PBRO182) and GSQL4170 (PBRO237).
Grypophyllum jenkinsi Strusz, 1966 – GSQL4165 (PBRO170), GSQL4167 (PBRO182) and GSQL4169 (PBRO218).
Xystriphyllum sp. cf *X. magnum* Hill, 1942c – GSQL1
Stringophyllum bipartitum Hill, 1942a – GSQL4170 (PBRO237).
Stringophyllum quasinormale Hill, 1942a – GSQL4137 (MHRO036), GSQL4164 (PBRO164), GSQL4166 (PBRO181), GSQL4169 (PBRO218), GSQL4170 (PBRO237), UQL1142
Sunophyllum sp. cf. *S. proteum* Zhen & Jell, 1996 – GSQL2.
Disphyllum stupendum Blake, 2009 – GSQL4170 (PBRO237)
Amaraphyllum amoenum Pedder, 1970 – GSQL4167 (PBRO182) and GSQL4171 (PBRO272).
Temnophyllum stainesi Blake, 2009 – GSQL1.
Thamnopora boloniensis(Gosselet, 1877) – GSQL1, GSQL4137 (MHRO036), GSQL4164 (PBRO164), GSQL4166 (PBRO181), GSQL4167 (PBRO182),GSQL4169 (PBRO218), GSQ4170 (PBRO237), GSQL4174 (RRRO035), GSQL4179 and UQL1142.
Alveolites caudatus Hill, 1954 – GSQL1, GSQL2, GSQL4137 (MHRO036), GSQL4160 (PBRO024), GSQL4170 (PBRO237) and UQL1150
Alveolites suborbicularis Lamarck, 1801 – GSQL1, GSQL3799 (PBRO015), GSQL3800 (PBRO023), GQL3832 (PBRO273), GSQL4160 (PBRO024), GSQL4161 (PBRO050), GSQL4164 (PBRO164), GSQL4165 (PBRO170), GSQL4167 (PBRO182), GSQL4169 (PBRO218), GSQL4170 (PBRO237), GSQL4171 (PBRO272), UQL1149.
Heliolites porosus (Goldfuss, 1826) – GSQL3832 (PBRO273), GSQL4160 (PBRO024), GSQL4165 (PBRO170), GSQL4167 (PBRO182), GSQL4169 (PBRO218), GSQL4170 (PBRO237), GSQL4171 (PBRO272), UQL1142
Multithecopora tubus Blake, 2009 – GSQL1 and GSQL4170 (PBRO237).

Ginger Creek Member

Stringophyllum quasinormale Hill, 1942a – GSQL4168 (PBRO217)
Thamnopora boloniensis(Gosselet, 1877) – GSQL4168 (PBRO217)
Alveolites suborbicularis Lamarck, 1801 – GSQL4168 (PBRO217)
Heliolites porosus (Goldfuss, 1826) – GSQL4168 (PBRO217).

Marble Waterhole beds

Tryplasma? abyssum Blake, 2009 – GSQL4139 (MHRO378) and GSQL4153 (MHRO790).
Tryplasma? careoseptum Blake, 2009 – GSQL4153 (MHRO790).
Calceola sp.
Cystiphylloides immanum Blake, 2009 – GSQL4153 (MHRO790).
Microplasma caespitosum (Schlüter, 1882) – GSQL4140 (MHRO379).
Pseudomicroplasma australe (Etheridge, 1892b) – GSQL4140 (MHRO379), GSQL4148 (MHRO488) and GSQL4153 (MHRO790).
Breviphyllum sp. A – GSQL4153 (MHRO790).
Endophyllum columna columna (Hill, 1942a) – GSQL4108.
Sanidophyllum davidis Etheridge, 1899 – GSQL4139 (MHRO378), GSQL4148 (MHRO488) and GSQL4153 (MHRO790).
Tabulophyllum sp. A – GSQL4108.
Acanthophyllum (A.) clermontense (Etheridge, 1911) – GSQL4153 (MHRO790).
Dohmophyllum clarkei Hill, 1942b – GSQL4108, GSQL4145 (MHRO442), GSQL4148 (MHRO488) and GSQL4153 (MHRO790).
Grypophyllum jenkinsi Strusz, 1966 – GSQL4108 and GSQL4153 (MHRO790).
Stringophyllum bipartitum Hill, 1942a – GSQL4146 (MHRO485) and GSQL4148 (MHRO488).
Stringophyllum quasinormale Hill, 1942a – GSQL4139 (MHRO378), GSQL4147 (MHRO486), GSQL4148 (MHRO488), GSQL4150 (MHRO692).
Sunophyllum sp. cf. *S. proteum* Zhen & Jell, 1996 – GSQL4139 (MHRO378) and UQL1330.
Disphyllum stupendum Blake, 2009 – GSQL4153 (MHRO790).
Amaraphyllum amoenum Pedder, 1970 – GSQL4139 (MHRO378), GSQL4148 (MHRO488), GSQL4153 (MHRO790), GSQL4154 (MHRO792).
Thamnopora boloniensis(Gosselet, 1877) – GSQL4108, GSQL4139 (MHRO378), GSQL4143 (MHRO401), GSQL4147 (MHRO486), GSQL4148 (MHRO488), GSQL4153 (MHRO790), GSQL4154 (MHRO792), UQL1330, UQL6702.
Alveolites caudatus Hill, 1954 – GSQL4108, GSQL4139 (MHRO378), GSQL4143 (MHRO401), GSQL4146 (MHRO485), GSQL4147 (MHRO486), GSQL4148 (MHRO488), and GSQL4153 (MHRO790).
Alveolites suborbicularis Lamarck, 1801 – GSQL4139 (MHRO378), GSQL4146 (MHRO485), GSQL4147 (MHRO486), GSQL4148 (MHRO488), GSQL4154 (MHRO792), UQL1330.
Heliolites porosus (Goldfuss, 1826) – GSQL4108, GSQL4139 (MHRO378), GSQL4140 (MHRO379), GSQL4146 (MHRO485), GSQL4148 (MHRO488), GSQL4150 (MHRO692), GSQL4153 (MHRO790), and GSQL4155 (MHRO792).
Multithecopora tubus Blake, 2009 – GSQL4139 (MHRO378) and GSQL4154 (MHRO792).
Multithecopora sp. A. – GSQL4148 (MHRO488).

Hoopbound Formation

Smithiphyllum petercollsi Blake, 2009 – GSQL4138 (MHRO121) and GSQL4174 (RRRO035).
Temnophyllum kroombitense Blake, 2009 – GSQL959.
Thamnopora boloniensis(Gosselet, 1877) – UQL6702.
Alveolites caudatus Hill, 1954 – UQL6702.
Alveolites suborbicularis Lamarck, 1801 – GSQL4174 (RRRO035).

Lochenbar Formation

Smithiphyllum petercollsi Blake, 2009 – GSQL4142 (MHRO832), GSQL4144 (MHRO435), GSQL4149 (MHRO680), GSQL4152 (MHRO776c), GSQL4177 (PBMO517), UQL6701.
Smithiphyllum finseni Blake, 2009 – GSQL4177 (PBRO517).
Tabulophyllum sp. A – UQL6700.
Tabulophyllum sp. B – UQL6700.
Disphyllum caespitosum (Goldfuss, 1826) – GSQL4142 (MHRO832).
Charactophyllum sp. A – GSQL4144 (MHRO435).
Charactophyllum sp. B – GSQL4151 (MHRO725).
Piceaphyllum menyouense (Hill & Jell, 1971) – GSQL4141 (MHRO380), GSQL4144 (MHRO435), GSQL4178 (PBMO518) and UQL6701.
Temnophyllum kroombitense Blake, 2009 – GSQL4142 (MHRO382) and UQL6700.
Thamnopora boloniensis(Gosselet, 1877) – GSQL4142 (MHRO382) and GSQL4151 (MHRO725).

Alveolites caudatus Hill, 1954 – GSQL4149 (MHRO680), GSQL4151 (MHRO725) and UQL6700.
Alveolites murrayi Blake, 2009 – GSQL4142 (MHRO382), GSQL4144 (MHRO435), GSQL4149 (MHRO680), GSQL4151 (MHRO725) and GSQL4152 (MHRO776c).
Alveolites suborbicularis Lamarck, 1801 – GSQL4073 (MHRO1108c), GSQL4141 (MHRO380), GSQL4144 (MHRO435), GSQL4151 (MHRO725), GSQL4152 (MHRO776c) and UQL6700.
Multithecopora sp. A. – GSQL4151 (MHRO725).

Three Moon Conglomerate

Smithiphyllum petercollsi Blake, 2009 – GSQL4133 (GSMO1217).
Smithiphyllum finseni Blake, 2009 – GSQL4133 (GSMO1217).
Columnaria jenkinsi Wright & others, 1990 – GSQL4133 (GSMO1217).
Disphyllum castellum Blake, 2009 – GSQL4132 (GSMO1130) and GSQL4133 (GSMO1217).
Argutastrea hullensis (Hill, 1954) – GSQL4133 (GSMO1217).
Temnophyllum kroombitense Blake, 2009 – GSQL4133 (GSMO1217).
Hexagonaria? sp. A – GSQL4133 (GSMO1217).
Thamnopora boloniensis(Gosselet, 1877) – GSQL4133 (GSMO1217).
Alveolites caudatus Hill, 1954 – GSQL4133 (GSMO1217).
Alveolites murrayi Blake, 2009 – GSQL4132 (GSMO1130) and GSQL4133 (GSMO1217).
Alveolites suborbicularis Lamarck, 1801 – GSQL4133 (GSMO1217)

Mount Alma Formation

Temnophyllum sp. A – GSQL4162 (PBRO089).
Thamnopora boloniensis (Gosselet, 1877) – GSQL4157 (PBMO012d) and UQL2632.
Alveolites caudatus Hill, 1954 – GSQL4157 (PBMO012d) and UQL2632.
Alveolites suborbicularis Lamarck, 1801 – GSQL4172 (PBRO430) and UQL2632.

Channer Creek beds

Charactophyllum sp. A – UQL2145.
Piceaphyllum menyouense (Hill & Jell, 1971) – UQL2145.
Temnophyllum turbinatum Hill, 1954 – UQL2145.
Temnophyllum kroombitense Blake, 2009 – UQL2145.
Alveolites caudatus Hill, 1954 – UQL2145.
Heliolites porosus (Goldfuss, 1826) – UQL2145.

Locality number	**1:100 000 Sheet**	**Grid Reference**	**Descriptive location**
GSQK72	Rookwood	802885 7422067	Near Armagh homestead.
GSQL1	Mount Morgan	239100 7380100	5.5 km ESE of Mount Morgan.
GSQL2	Mount Morgan	238500 7382100	5.5 km ESE of Mount Morgan.
GSQL959	Mount Morgan	235964 7372461	9 km SSE of Mount Morgan.
GSQL3799	Mount Morgan	240680 7381896	7 km ESE of Mount Morgan.
GSQL3800	Mount Morgan	239187 7381425	6 km ESE of Mount Morgan.
GSQL3832	Bajool	249006 7367513	22 km SE of Mount Morgan.
GSQL3838	Bajool	275153 7350863	2 km NW of Mount Alma.
GSQL4052	Rookwood	803051 7421883	Near Armagh homestead.
GSQL4073	Biloela	276768 7293247	1.5 km S of Kroombit homestead.
GSQL4108	Biloela	282400 7290950	6.5 km SE of Kroombit homestead.
GSQL4129	Rookwood	806143 7436497	2 km SE of Melrose homestead.
GSQL4130	Ridgelands	201146 7435495	10.5 km ESE of Melrose homestead.
GSQL4131	Ridgelands	201420 7435055	10.5 km ESE of Melrose homestead.
GSQL4132	Monto	313519 7273347	25 km N of Monto.
GSQL4133	Monto	313173 7272670	25 km N of Monto.
GSQL4134	Rookwood	802836 7422059	Near Armagh homestead.
GSQL4135	Mount Morgan	237369 7371189	12 km SSE of Mount Morgan.
GSQL4136	Mount Morgan	241196 7382158	7.5 km E of Mount Morgan.

GSQL4137	Mount Morgan	241014 7381816	7.5 km E of Mount Morgan.
GSQL4138	Mount Morgan	234649 7376052	7 km S of Mount Morgan.
GSQL4139	Biloela	281124 7293179	4.5 km ESE of Kroombit homestead.
GSQL4140	Biloela	278731 7293991	2 km ESE of Kroombit homestead.
GSQL4141	Biloela	277370 7292666	2 km SSE of Kroombit homestead.
GSQL4142	Biloela	263063 7299050	14.5 km WNW of Kroombit homestead.
GSQL4143	Biloela	282664 7290964	7 km SE of Kroombit homestead.
GSQL4144	Biloela	276774 7292965	1.5 k m S of Kroombit homestead.
GSQL4145	Biloela	278295 7294712	1.5 km E of Kroombit homestead.
GSQL4146	Biloela	281068 7293383	4.5 km ESE of Kroombit homestead.
GSQL4147	Biloela	280691 7292873	4 km SE of Kroombit homestead.
GSQL4148	Biloela	281143 7292757	4.5 km ESE of Kroombit homestead.
GSQL4149	Biloela	271482 7297835	6.5 km WNW of Kroombit homestead.
GSQL4150	Biloela	279971 7297358	4 km NW of Kroombit homestead.
GSQL4151	Biloela	262211 7299435	15 km WNW of Kroombit homestead.
GSQL4152	Biloela	277359 7291589	3 km S of Kroombit homestead.
GSQL4153	Biloela	281224 7293004	4.5 km ESE of Kroombit homestead.
GSQL4154	Biloela	281875 7292799	5.3 km ESE of Kroombit homestead.
GSQL4155	Biloela	281151 7293140	4.5 km ESE of Kroombit homestead.
GSQL4156	Calliope	321397 7332651	12 km SSE of Calliope.
GSQL4157	Calliope	299669 7343317	18 km W of Calliope.
GSQL4158	Calliope	299579 7343685	18 km west of Calliope.
GSQL4159	Calliope	327952 7318425	27.5 km SSE of Calliope.
GSQL4160	Mount Morgan	239074 7381545	5.5 km ESE of Mount Morgan.
GSQL4161	Mount Morgan	242253 7381236	9km ESE of Mount Morgan.
GSQL4162	Bajool	246451 7389856	14 km NE of Mount Morgan.
GSQL4163	Mount Morgan	240644 7380480	7.5 km SE of Mount Morgan.
GSQL4164	Bajool	246013 7371801	14.5 km SE of Mount Morgan.
GSQL4165	Bajool	250021 7367808	22.5 km SE of Mount Morgan.
GSQL4166	Bajool	250718 7363511	26 km SE of Mount Morgan.
GSQL4167	Bajool	250726 7364550	25 km SE of Mount Morgan.
GSQL4168	Bajool	253110 7359427	30.5 km SE of Mount Morgan.
GSQL4169	Bajool	254068 7359989	31 km SE of Mount Morgan.
GSQL4170	Bajool	249507 7366383	23 km SE of Mount Morgan.
GSQL4171	Bajool	248944 7367789	21.5 km SE of Mount Morgan.
GSQL4172	Bajool	272045 7375086	6 km W of Raglan.
GSQL4173	Bajool	279514 7370904	5 km SSE of Raglan.
GSQL4174	Bajool	240215 7369927	15 km SSE of Mount Morgan.
GSQL4176	Rookwood	803700 7432850	10 km N of Armagh homestead.
GSQL4177	Biloela	276754 7292992	25 km ESE of Biloela.
GSQL4178	Biloela	276770 7293067	25 km ESE of Biloela.
GSQL4179	Mount Morgan	241300 7382300	8 km E of Mount Morgan.
UQL1330	Biloela	278350 7294500	1.6 km E of Kroombit Homestead.
UQL1142	Mount Morgan	240450 7384750	7 km ENE of Mount Morgan.
UQL1143	Mount Morgan	238100 7382800	7.5 km E of Mount Morgan.
UQL1144	Mount Morgan	238450 7381600	5 km ESE of Mount Morgan.
UQL1149	Mount Morgan	235900 7379350	4 km SSE of Mount Morgan.
UQL1150	Mount Morgan	238400 7381900	5.5 km ESE of Mount Morgan.
UQL1153	Mount Morgan	241500 7386300	8.5 km ENE of Mount Morgan.
UQL2145	Monto	338150 7238600	30 km ESE of Monto.
UQL2472	Bajool	289050 7364850	16 km SE of Raglan.
UQL2473	Bajool	279600 7370950	4.5 km SSE of Raglan.
UQL2474	Bajool	279800 7369850	5.5 km SSE of Raglan.
UQL2475	Bajool	281600 7367200	9 km SSE of Raglan.
UQL2632	Ridgelands	242700 7419250	8 km N of the Rockhampton airport.
UQL3523	Bajool	279800 7369400	6 km SSE of Raglan.
UQL3524	Bajool	281400 7366900	9 km SSE of Raglan.
UQL6700	Biloela	276700 7394100	300 m S of Kroombit homestead.
UQL6701	Biloela	277300 7392500	2 km SSE of Kroombit homestead.
UQL6702	Mount Morgan	231050 7380450	3.5 km SW of Mount Morgan.
UQL6704	Calliope	298800 7342800	19 km W of Calliope.

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APPENDIX 5
GEOPHYSICAL INTERPRETATION

(W. Stasinowsky)

(*with comments in italics on the relationship to geological mapping by C.G. Murray*)

This interpretation is also part of the GIS compiled by Scott (2006), which includes the algorithms and other MapInfo and ArcMap files.

REGIONAL STRUCTURE

Before looking in detail at the Yarrol datasets, an overview interpretation of continental features was made using Australian and Queensland datasets from AGSO and Queensland DME. The positions of features from the regional datasets may not correspond with features interpreted in the more detailed datasets because of the very broad resolution in the regional datasets.

Gravity

An interpretation conducted as an AMIRA project (Horton & others, 1993) had continental scale linears (see overlay "AMIRA_G.erv" – locations digitised and approximate only). Locations of small highs, lows and steep gradients on the AMIRA interpretation were updated by the current interpretation. Additional continental scale features intersecting the Yarrol area were interpreted on regional gravity from AGSO and the Queensland DME (see overlay ("Reg_Grav.erv" and algorithm "Reg_Grav.alg").

On the western margin of the Yarrol area, a north-west striking linear marked the western boundary of an intense gravity ridge. It extended south-east past the north-eastern margin of a low. Two similar striking linears were also interpreted further west outside the Yarrol area.

Several north-east striking linears were also interpreted intersecting the area and coincided with gradients for significant parts of their strike.

The north-west and north-east striking linears were interpreted as major crustal faults and may form a conjugate set.

On the eastern margin of the area, a northerly striking linear was interpreted west of several gravity ridges interpreted to be thrust related (see "Gravity Interpretation" below).

Most of the linears and gradients were interpreted to be fault related and were thought to originate from major crustal boundaries with significant depth extent.

The only feature to coincide with a regional magnetic feature (see below) was an east-south-east striking linear across the southern part of the area.

Magnetics

An interpretation conducted as an AMIRA project (Horton & others, 1993) had continental scale linears (see overlay "AMIRA_M.erv" — locations digitised and approximate only). Locations of small highs, lows and steep gradients on the AMIRA interpretation were updated by the current interpretation. Additional continental scale features intersecting the Yarrol area were interpreted on regional magnetic datasets from AGSO and the Queensland DME (see overlay "Reg_Mag.erv" and algorithm "Reg_Mag.alg").

Numbers shown on the AMIRA_M.erv overlay are the numbers on the AMIRA magnetic interpretation.

The AMIRA "Magnetic Break/Linears" and the additional regional linears were all interpreted to be fault related. They were interpreted to be crustal scale faults but their depth extent could not be interpreted. In general, they were thought to be from features closer to the surface than the gravity linears. This was probably why there was so little correlation between the regional gravity and magnetic features. The only regional magnetic feature to coincide with a regional gravity feature (see above) was an east-south-east striking linear across the southern part of the area.

Skippy seismic project

The Skippy project conducted by the Research School of Earth Sciences at ANU (Kennett, 1997) used regional seismic events to form a three dimensional map of seismic velocities in the crust to about 400km depth.

They determined that a zone of higher velocity rocks exists beneath the Yarrol area. They equated high velocity with cooler rocks but higher velocity may also equate to denser mantle material thrust up under the area. The higher velocity material appeared to dip to the east, which was consistent with thrusting interpreted on the eastern side of the Yarrol area (see below).

GRAVITY INTERPRETATION

The gravity dataset was interpreted with linears and closed anomalies. The closed anomalies were interpreted with three levels to provide more definition with each grading from weak to moderate to strong. These were all subjective interpretations and were meant to convey local information only rather than information between different anomalies. Relativities between anomalies could be gained from the gravity interpretation image directly rather than the line interpretations (see algorithm “Gravity.alg”).

The colour scale for both the interpretation and the image, (see algorithm “Gravity.alg”), was selected as hotter colours reflecting higher values and cooler colours representing lower values (Figure 1).

Several linears and closed anomalies were interpreted. The gradient and undifferentiated linears were interpreted to be faults. The undifferentiated linears were often a series of gradient features interspersed with either high or low features. The gradients often were in opposite sense on different parts of the linear. This was thought to be caused by faults which either had cross cutting faults causing different blocks to have different sense of throw or the fault being the boundary for intrusions. Intrusions were interpreted to have both positive and negative relative densities (see below).

Certain prominent features have been investigated and are discussed from the south. Names of the features have mainly been taken from the magnetics interpretation and an overlay of the magnetic intrusion interpretation helped with the gravity interpretation (see algorithm “Gravity.alg”).

A large negative anomaly was evident crossing the southern boundary and was identified as the Rawbelle Batholith. Its north-east boundary was interpreted to be fault controlled.

The Glassford Complex appeared relatively neutral apart from phase 3, (see “Glassford Complex” below), which was a weak low and phase 12 which appeared to be a weak high. The complex overlapped a strong low to the east which also covered the Nagoorin Basin. Neither the complex nor the basin was thought to be the origin of the strong low as it was offset and larger than both. It was thought to be another batholith at depth. The Nagoorin Basin could have enhanced the negative but the whole anomaly extended east under an interpreted thrust front (see below).

Two linear highs extending north-north-west from the Nagoorin Basin were interpreted as arising from thrust fronts. The easterly one was evident for the whole length on the magnetics while the western one was only evident on the magnetics in the north.

Kroombit Tops had a moderate gravity low associated with its eastern edge but the whole intrusion only had a weak gravity low. This could be interpreted as the root of the intrusion being on the eastern side or a separate phase intruded there. Diglum and Diglum West did not have any significant gravity anomaly.

Galloway Plains had a moderate low centred under phase 18 (see “Galloway Plains” below). The remainder of the complex was a weak low and suggested it was acidic in composition. The Langmorne intrusion did not appear connected to the Galloway Plains anomaly on the magnetics but the gravity not only suggested a similar composition but also a link via an enclosing weak low around both.

The Craiglands Complex appeared linked to Galloway Plains on the magnetics but had a moderate gravity high suggesting it was a different composition. A weak gravity high enclosed both Craiglands and Eulogie Park. Eulogie Park was a very intense magnetic anomaly with strong remanence. Together with the strong gravity high, it was consistent with a basic intrusion.

The southern part of Bajool had a weak gravity low. It did not extend north-west into the region of strong magnetic banding. This added to the evidence suggesting the bands were not rims within the intrusion but probably magnetic skarn or alteration development.

The Bouldercombe Complex was as complicated on the gravity as it was on the magnetics. The main outcropping east and west lobes appeared to be gravity lows relative to the intense highs surrounding them. There was a gravity high along the northern edge of the outcropping area which was interpreted to correspond

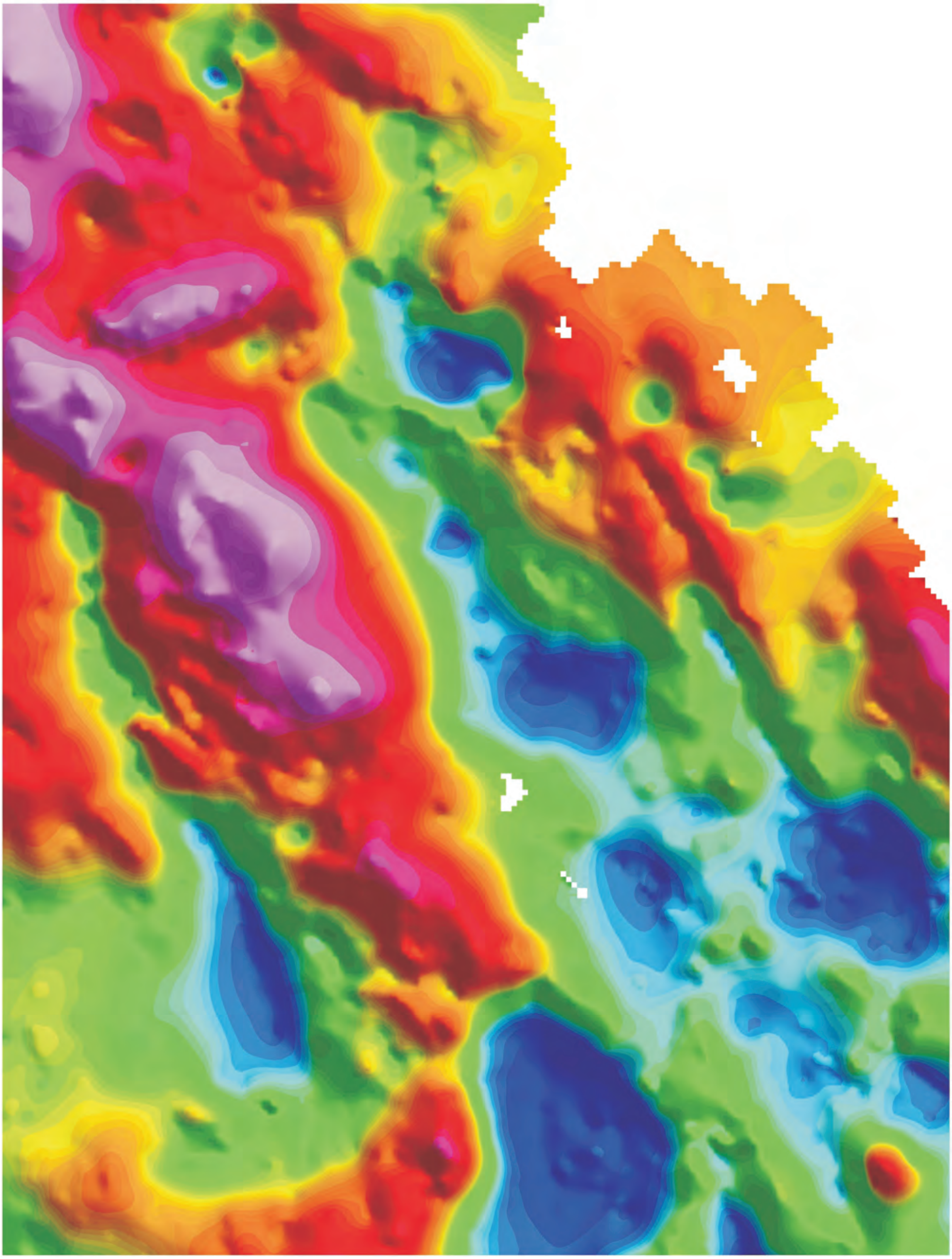


Figure 1. Gravity image covering the Yarrol Project area, with features referred to in the text

to phase 1 (see “Bouldercombe Complex” below). The very magnetic phase 19 had a strong gravity high immediately west of it which was interpreted to be the same unit. This was thought to be basic.

Moderate to strong gravity highs through and north of Rookwood were also thought to be basic intrusives. Shallow or outcropping areas were interpreted on the very northern boundary.

A relative low was coincident with an intrusion north of Ridgelands. This was evident as a separate phase on the radiometrics and confirmed as such on the gravity. The low did not extend to the south-west where different radiometric phases were interpreted.

Other than the Nagoorin Basin, the only two basins to exhibit obvious gravity lows were the Casuarina and Yaamba Basins. The Jim Crow, Rossmoya, Parkhurst and Stanwell Basins did not exhibit any significant low although the Jim Crow Basin did have a weak low in part of it. The Yaamba Basin had a very pronounced low for its size while the Casuarina Basin had a more intense low within a weaker low. The more intense low corresponded to the deeper sections from the magnetics modelling (see “Casuarina Basin” below).

RADIOMETRICS INTERPRETATION

The radiometric channels were converted to ppm for uranium and thorium and to per cent for potassium. A three channel red-green-blue mix was found useful for 90% of the interpretation with good definition on many boundaries (see algorithm “Rad.alg”). In areas where the three channel mix was too dark or too light, the image window was zoomed in to the specific area and an automatic 90% clip was applied to all channels to create greater contrast and definition.

Ratios were generated but found to offer little additional detail.

Radiometric boundaries were interpreted using the above methods to generate a surface map of the area (see overlay “Rad.erv”).

Boundaries of radiometric zones were interpreted with solid lines. These included areas of water such as dams, rivers and the coast. Some features were interpreted as structure zones and were marked with dashed lines. The structure zones were thin linear highs or lows in one or more radiometric channels and were usually marked where there was also a sharp linear radiometric boundary indicating a fault. The structure zones were interpreted to be where ground water either enhanced or depleted radiometric elements in the near surface.

MAGNETICS INTERPRETATION

The colour scale for both the image, (see algorithm “Mag.alg”), was selected as hotter colours reflecting higher values and cooler colours representing lower values.

The magnetic dataset was interpreted for structure, intrusions and dykes. Each of these is discussed in separate sections below.

To aid the interpretation of some anomalies and features, modelling and Fourier analysis were carried out. Modelling included both forward and automatic modelling. A brief outline of each of these methods is given below.

Forward modelling

Conventional forward modelling was performed using Encom’s ModelVision software. Initial models started with vertical prisms and were modelled in 2.5D (finite strike). Some models used prisms with polygonal cross sections and two models were conducted in 3D using multiple lines of data. No basin depth models were done in 3D.

While most of the models had susceptibilities consistent with those measured in the field, some were higher by a factor of 2. This suggested that the models had greater depths than in reality. If the real bodies did not have sharp contacts, the model would need to be deeper to fit the measured curve. This is most often noticed where magnetic alteration is involved and the amount of magnetite dissipates away from the source. The model in these cases will always appear deeper and the susceptibility would need to be higher to give the same magnitude response. This was supported by the fact that no model outcropped but field mapping detected highly susceptible rocks.

Modelling was usually done using profiles extracted from the magnetic TMI grid. This enabled profiles to be extracted with any strike and proved satisfactory because of the large size of the bodies being modelled. However, where depths were suspected of being shallow, there may have been some degradation of the solution because of the subsampling.

FFT analysis

Any waveform can be analysed in terms of its spectrum. In the same way that light waves can be split into their spectrum by applying an appropriate filter (prism or diffraction grating), any data series can be mathematically split into its spectrum by applying the appropriate filter (a Fourier Transform).

The resultant spectrum indicates the components of the original waveform in terms of the frequency and corresponding amplitude at each frequency. Fourier theory states that any waveform can be decomposed into sine and cosine waveform components with appropriate phase and amplitude.

The depth of a magnetic body can be determined from its measured profile. The broader the anomaly, the deeper the originating body.

When there are multiple bodies with similar depth characteristics, the depth can be inferred from analysis of the spectrum of the magnetic profile over the bodies.

By assuming that a magnetic volcanic layer underlies sediments, the thickness of sediments can be determined by analysing the depth to the top of the volcanics. If the volcanics were assumed to be an ensemble of magnetic bodies, Fourier analysis of the magnetic profile should yield the appropriate depth.

For Fourier analysis, data were analysed along flight lines over the selected basins. It was not suitable to select profiles from the grid because the spectrum would be compromised by the gridding. Extracting profiles from the grid would also lead to fewer data points per profile which would not give good definition on the spectrum.

Because the Fourier analysis algorithm used required a power of two for the number of data points, either 1024 or 2048 points were selected depending on the width of the basin. With a sample interval of approximately 7.5m, profile segments were approximately 7.7km and 15.4km. The result was an average for sediment thickness across the whole basin. Where the basin was faulted, the analysis was likely to show several depths. If the analysis could not resolve the different levels, the spectrum would be inconclusive. This limitation was factored into the interpretation and other methods were used as appropriate.

Despite its limitations, Fourier analysis can give results where other methods fail, particularly for the case where the basin has a single level with steep sides. Where anomalies existed within the basin basement, the agreement between Fourier, forward modelling and Automag solutions was very good.

The magnetic profile data were fed into an Excel spreadsheet. The Fourier Transform was computed and the natural log of the power was plotted against the frequency in cycles per km. Slopes were drawn on the graph using the formula below and by supplying intercept and slope values. These were manually changed until a good fit was achieved for any desired spectrum segment. An automatic method for this step was not desirable as considerable experience was required to determine valid segments in the spectrum.

The depth formula for a magnetic pole was calculated for each slope. A magnetic pole calculation would be appropriate where a long body was oriented vertically such that only one magnetic pole was presented at the surface of the body being investigated. This formula was used in general and was confirmed as appropriate if the spectrum decayed uniformly rather than having a peak. This shape was confirmed for all the spectra used. The formula for a magnetic dipole was not used as none of the spectra presented the dipole characteristic shape.

The formula used for a magnetic pole was:

$$\text{Depth} = \frac{(\mathrm{Ln}(a) - \mathrm{Ln}(b))}{4\pi(b - a)}$$

Where Ln(a) = Natural Log of Amplitude of First Point
Ln(b) = Natural Log of Amplitude of Second Point
a = Wavelength of First Point
b = Wavelength of Second Point

The more spectral points falling on the spectral segment and the more values agreeing in spectra on adjacent lines, the higher is the confidence of interpretation.

Results were used in conjunction with the other depth estimation methods.

Automag modelling

Automag is a program included with Encom's ModelVision package. It automatically interprets body location and depths based on Naudy's method. Automag was run on several sections to give a third independent assessment of depth. It was used in conjunction with forward modelling and Fourier analysis.

The Naudy method employs fitting a dyke or edge to an anomaly. In Automag, the dyke can have one or more of its components free to move to give the best fit. These include position, depth, dip and susceptibility. Because Automag was primarily used to determine the basement depth, the dip was often fixed to vertical to minimise the variables. Allowing the dip to vary often produced spurious shallow bodies with dips close to horizontal.

Automag only works on a finite number of samples in a window. With samples every 7.5m and depths being up to 2000m, the data had to be subsampled quite severely in Automag so the window could be wide enough to cover anomalies from the desired depths. In some instances, this introduced variability in the depth estimates from noise in the profiles.

YARROL AREA STRUCTURE

A number of images including first vertical derivative (Figure 2) and artificial shade were used to interpret faulting from the magnetics. Because of the nature of the magnetic response and the enhancements used, the interpretation focussed on near surface faulting. Deeper faulting was interpreted using the gravity dataset (see "Gravity Interpretation" above).

All the faulting was interpreted in detail (see overlay "AllFault.erv"). Several of these features were then investigated to determine whether they were thrusts. The results of investigation of each area are discussed in turn below.

After these investigations, a number of faults were interpreted to be thrusts (see overlay "Thrusts.erv"). The thrust interpretation overlay only indicated the thrust front in general. It should be used in conjunction with the "AllFault.erv" overlay to see the full complexity of the thrust fronts with their high angle transverse faults.

Limitations

A number of models were used to determine whether a fault was normal, reverse or thrust. The lower face of a magnetic body is difficult to model because variations in the overlying magnetic body could give rise to the responses attributed to the lower face. Models attempting to fit thrusts to the lower face were therefore more ambiguous than models where the interest was in the top magnetic face such as in normal faults and basin depth determinations.

Despite the limitations, modelling was conducted and found to give evidence for thrust faulting. Thrust models were always modelled with other possibilities to limit the number of geologically valid interpretations.

Berserker Range (BR)

Coordinate at approximate centre: GR257000 E, 7425000 N

Modelling was conducted to determine the dip of the main north-north-west striking fault (Yarrol Fault) (Figure 3). Four models containing three bodies were attempted with vertical dip, dip to the east, dip to the west and a thrust model (see models BR-A to BR-D). The best fit was the dip to the east followed closely by the thrust model. The high angle reverse dip of 75° to the east gave a more gentle rise to the west which was more consistent with the data. The thrust model had a much sharper negative and was thought a less likely scenario.

The evidence suggested that the fault dipped east at a high angle.

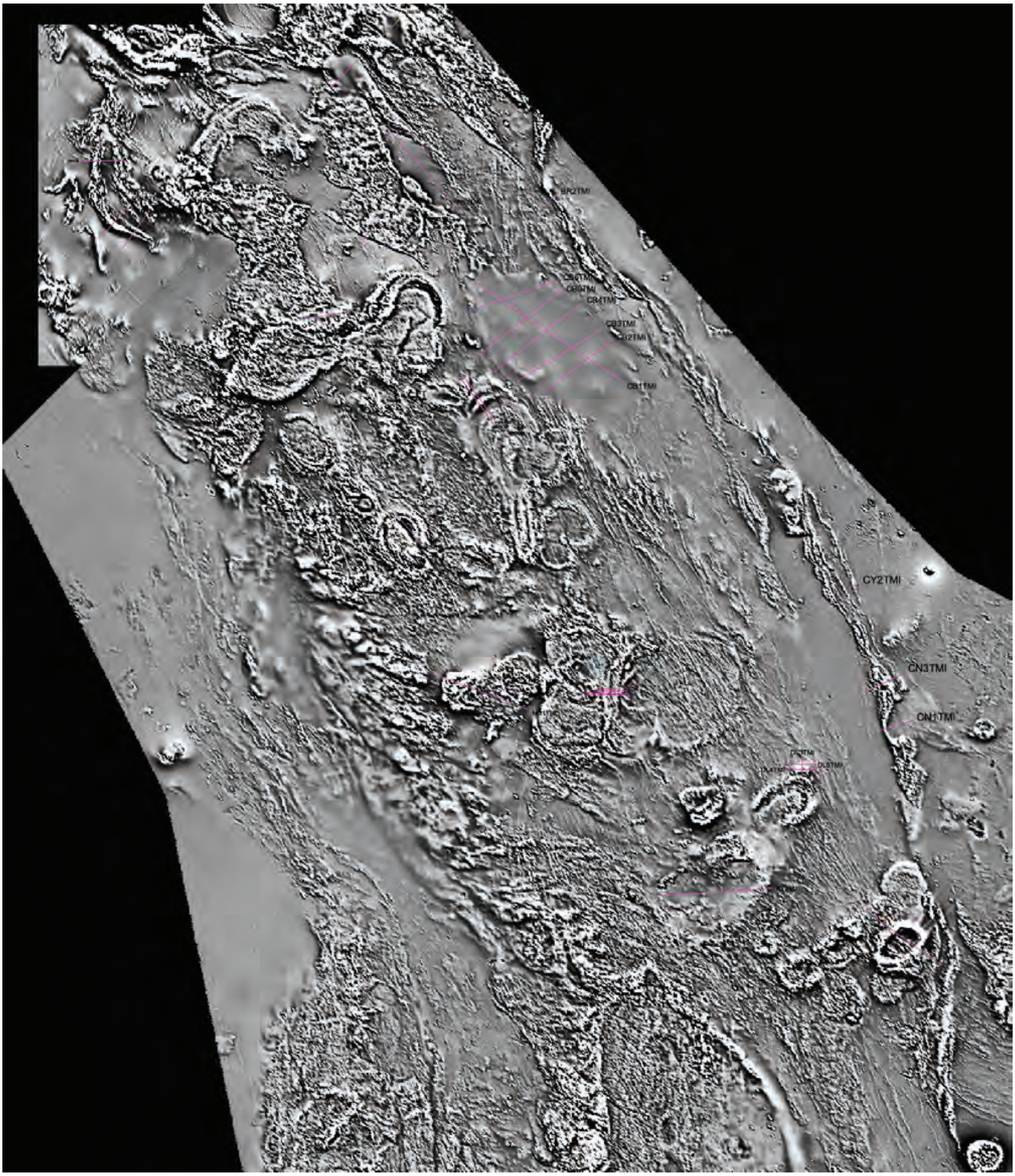


Figure 2. Image of airborne magnetic data over the Yarrol Project (first vertical derivative)

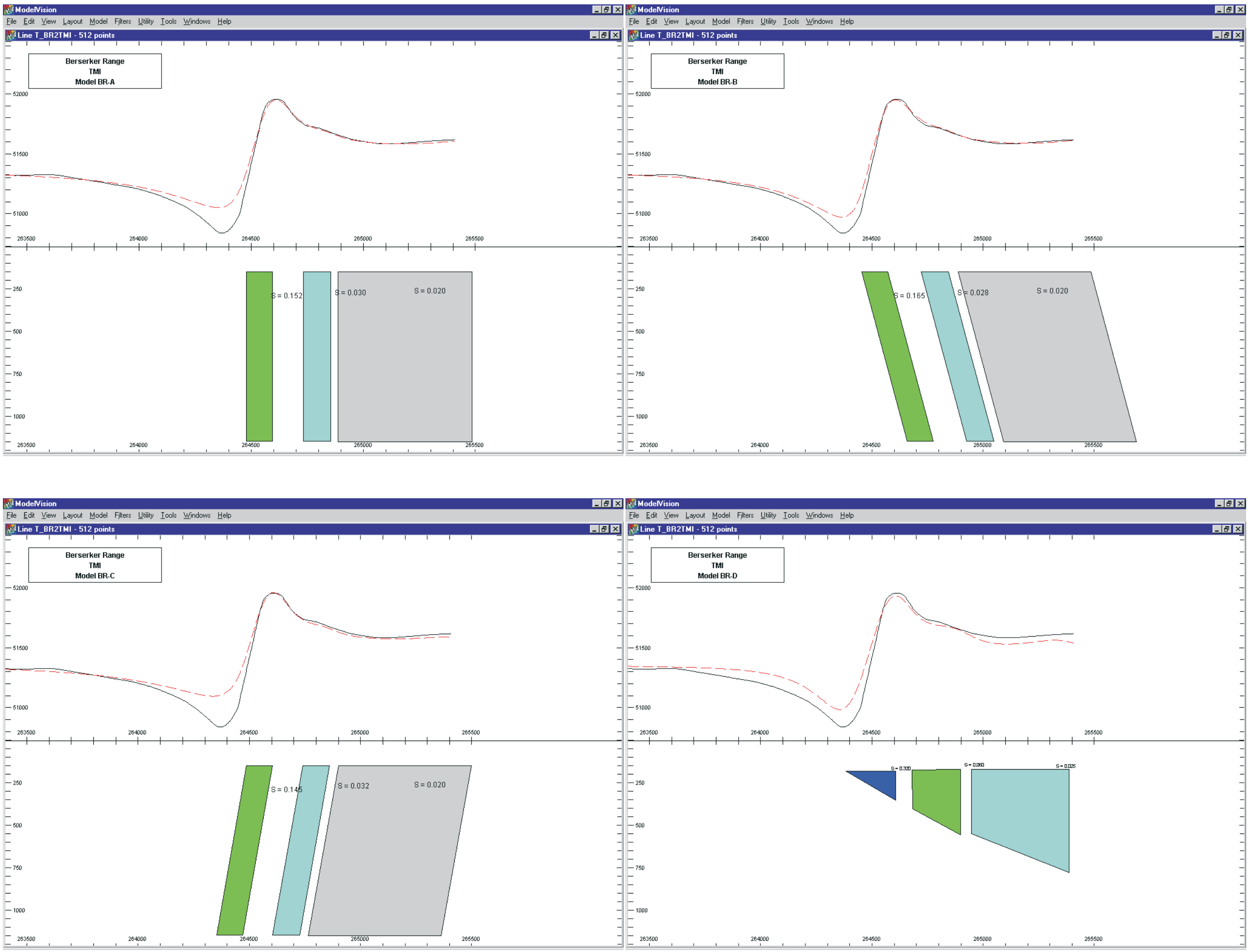


Figure 3. Models across the Yarrol Fault in the Tungamull area

Calliope – North (CN)

Coordinate at approximate centre: GR315000 E, 7349000 N

The area at Calliope North was interpreted to be an area of thrust faulting. Forward modelling with shallow bodies all dipping at 30º to the east was consistent with this interpretation. Most of the bodies were modelled at the surface except the most western body which was modelled at less than 40m depth (Figure 4).

The amplitude of the central portion of the model was higher than the amplitude of the corresponding data. However, the time required to get this area to match was thought unproductive and would probably involve inserting bodies with little relevance to reality. The primary question had been answered by having the other bodies match, particularly the western edge of the first two bodies and the low at the western edge of the bodies.

Calliope South (CS)

Coordinate at approximate centre: GR323000 E, 7332000 N

Modelling at Calliope South was concerned with confirming the interpretation of a thrust fault.

Line CN1TMI to the south intersected the interpreted thrust and an adjacent intrusion. The model demonstrated that the data were consistent with bodies dipping to the east at 40º (Figure 5a). This suggested that the thrust carried part of the intrusion in the overthrust block. This was also consistent with the weak response of the intrusion which also did not appear to have much depth extent.

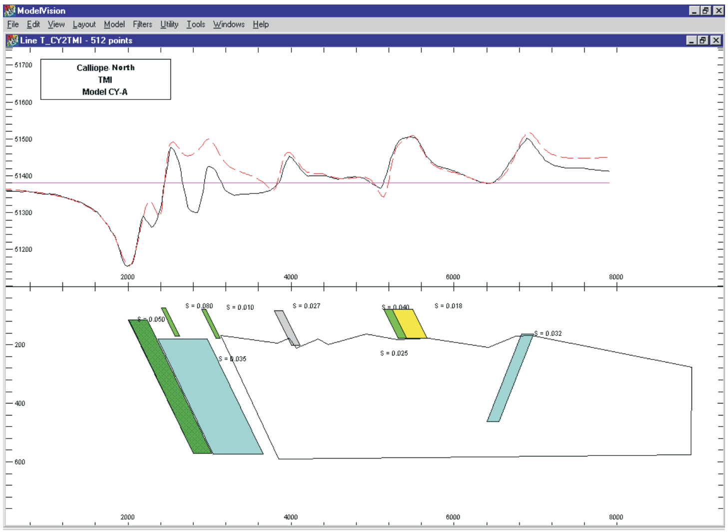


Figure 4. Model along a north-east trending line from the Rockhampton Group (west) across the Calliope beds and Yarrol Fault to the Doonside Formation in the Calliope area.

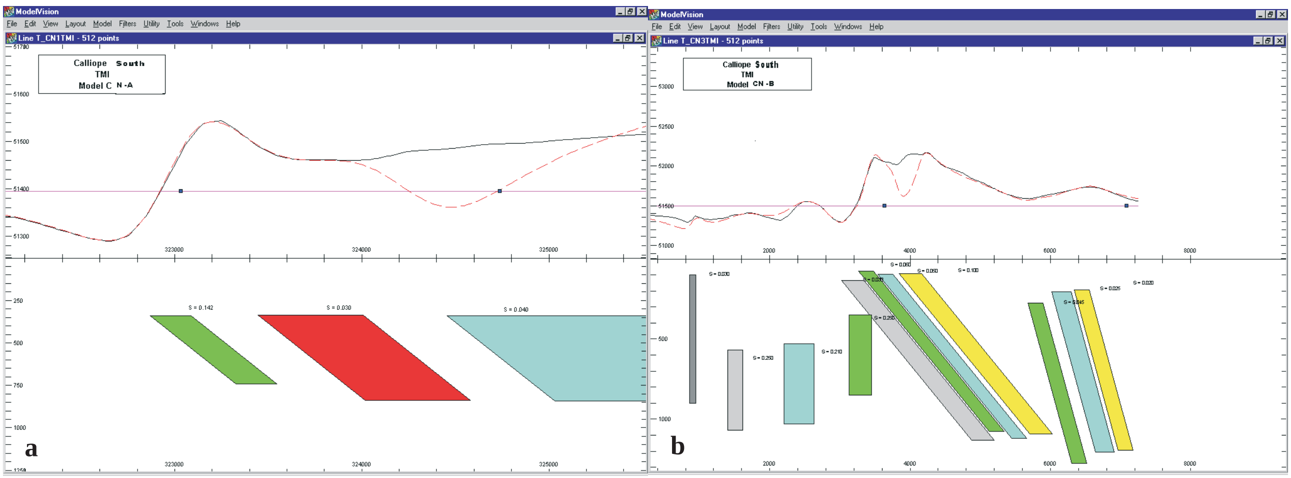


Figure 5. Models along east-north-east trending lines across the Castletower Granite and Lake Awoonga.

Line CN3TMI traversed over an interpreted multiple thrust package. The model showed deep vertical bodies to the west in the footwall block (Figure 5b). A series of shallow bodies dipping at 30° to the east represented the thrust package and fitted well on both the leading and trailing parts of the data. Getting the central part to also match was difficult and not as relevant to the question of fault dip. The mismatch was probably the result of the package having shorter depth extent bodies than those modelled. The time necessary to get multiple short depth extent bodies to match was thought unnecessary and would not be as convincing as the more uniform bodies used.

The area to the east had bodies with 60° dip. Models with 30° dip did not fit as well as the ones with 60° dip. However, they did fit better than bodies dipping west. This area may have been a steeper part of the thrust package or may represent an entirely different unit. (*It is close to the projected trace of the Yarrol Fault, and this result is consistent with the interpretation of the Berserker Range profile*).

Gracemere (GM)

Coordinate at approximate centre: GR230000 E, 7412000 N

Evidence for both normal and thrust faulting has been suggested for this area so six different models were attempted. Simple bodies with an easterly dip of 40º dip appeared to fit better than either vertical or westerly dip (see models GM-A to GM-C, Figure 6a-c).

A thrust model with vertical bodies truncated by a thrust was attempted and achieved a good match except for the western low in the modelled data (see model GM-D, Figure 6d).

Three other versions of the thrust model were then produced (see models GM-E to GM-G, Figure 6e-g). They all involved having the western apex shallower than the rest of the bodies and all achieved better results than the previous models.

Model GM-E was a modification of GM-D with the apex of the western body brought closer to the surface. This achieved a good match with the low but blew out on the high. Model GM-F attempted a modification of the simple shallow dip bodies and model GM-G combined the attributes of all the thrust models and would be consistent with magnetic material down the thrust plane. It achieved good correlation in the high areas but lost correlation in the low.

The evidence from all the models suggested that the fault was a thrust with shallow dip to the east but steeper dip near outcrop. There may or may not be magnetic material down the thrust plane. (*The magnetic unit to the north-east is a thin layer of Cretaceous basalt which is preserved above a basement of non-magnetic Carboniferous Rockhampton Group by normal faulting*).

Rookwood (RW)

Coordinate at approximate centre: GR188000 E, 7413000 N (Zone 56 grid, but point is in Zone 55)

The primary aspect addressed in the Rookwood area was whether the fault on the western edge of a shallow magnetic unit was a thrust. In plan, it had the character of a thrust with a curved front and several cross-cutting faults which could be interpreted as transverse faults in a thrust system. The potential transverse faults were striking north-east which was consistent with other thrust systems in the Yarrol area.

In the north of the area, modelling on line RW1 gave good fits for tabular bodies with steep dips (see model RW1-A, Figure 7a). Contrary to initial thoughts, the dips on the east and west were not parallel but opposite. The western bodies had dips of 82º to the east while the eastern body had a dip of 75º to the west. This suggested the area might be a syncline. The interpretation of the western bodies was ambiguous because the low to the west did not fit well and the three adjoining bodies had varying susceptibilities, notorious for forcing a fit. A more gentle dip to the east would have fitted the low better but would not have fitted the high. The eastern body was however unambiguously dipping west and was inconsistent with a thrust at that location.

The second line, RW2 transected the highest magnetic response in the area which also coincided with a major bend in the fault. The profile exhibited a broad low to the south-west which did not fit any of the thrust models tested (not shown). The thrust models had sharper lows. The model which did fit very well had several bodies contained within an overall body exhibiting a normal fault (Figure 7b). The susceptibilities of the contained bodies would be the sum of the surrounding body with that of the contained body. The evidence was quite strong at this location for a normal fault.

With the large high at the fault bend, it was possible that alteration was superimposed on a thrust to mask its response. A third line further south was therefore modelled to test this hypothesis.

Line RW3 was modelled (see model RW3-A, Figure 7c) to test whether a thrust was consistent with data in the lower intensity area. This was not confirmed because the data provided a good fit to simple models with dips of around 45º to the south-west.

The whole Rookwood fault system had little evidence for a thrust and it was thought more likely to be a normal fault dipping to the west and south-west. (*This region is in the Gogango Overfolded Zone for which structural interpretations indicate stacked thrust sheets dipping to the east. The existence of shallow thrusts has been demonstrated by exploration drilling*).

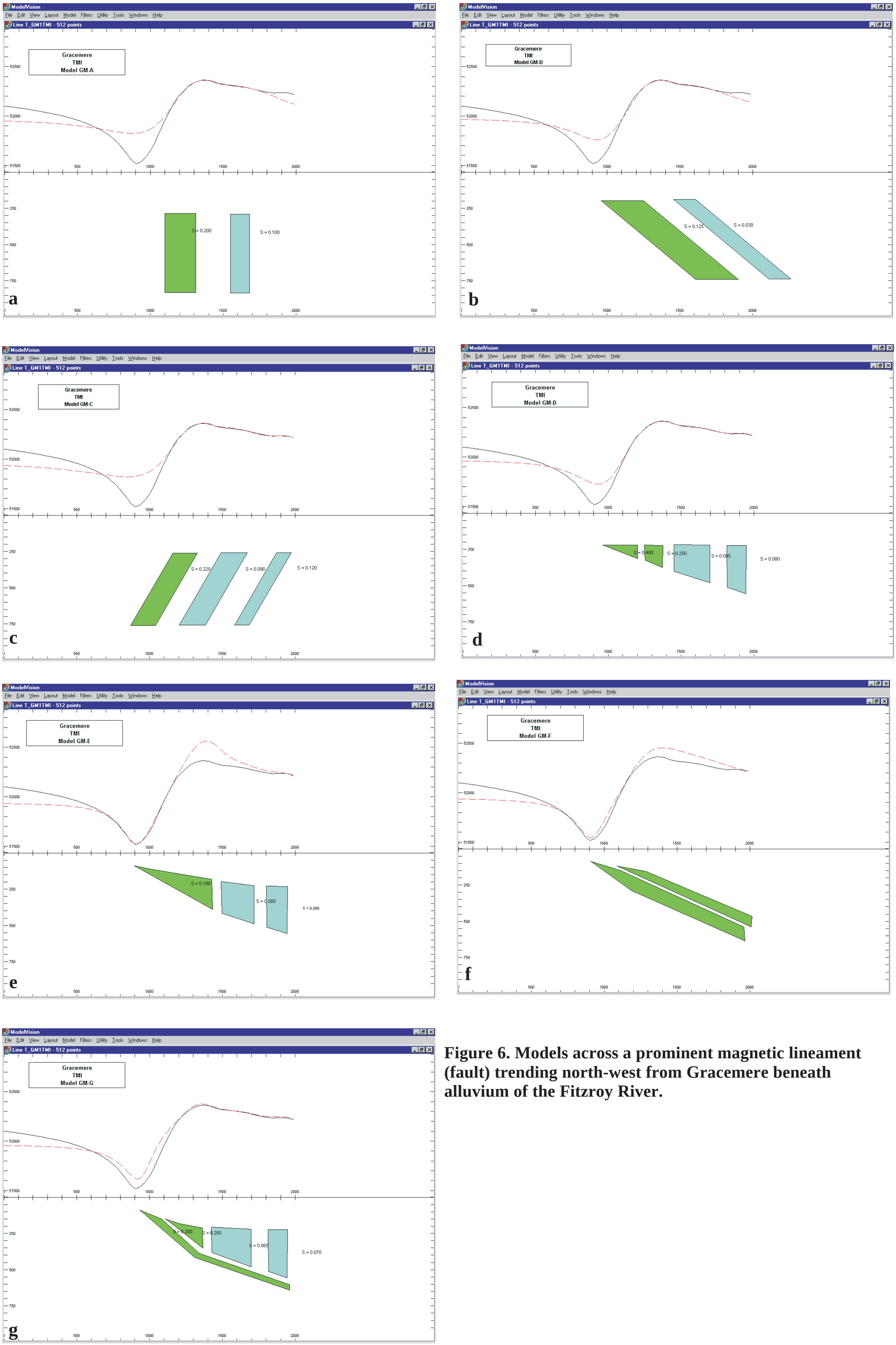


Figure 6. Models across a prominent magnetic lineament (fault) trending north-west from Gracemere beneath alluvium of the Fitzroy River.

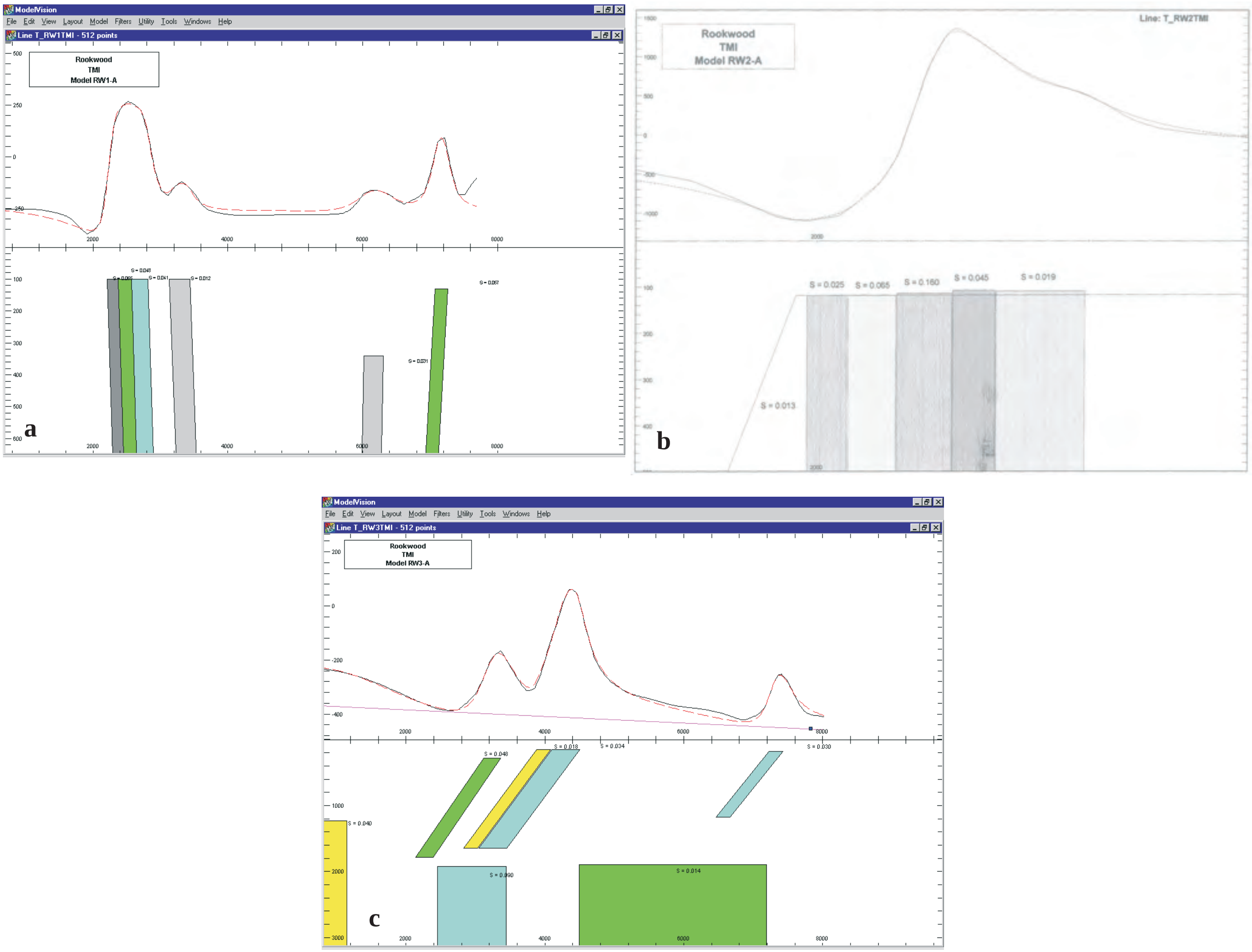


Figure 7. Models across the Gogango Overfolded Zone north-west of Westwood

INTRUSIONS

Several intrusions were interpreted in the area. Some of the major ones had detailed interpretation conducted as to the timing of their various phases and some had modelling conducted to determine their depth (see below). Intrusions that did not have detailed analysis or modelling have not been discussed further but are shown on overlay “Intrusions.erv”.

To avoid duplication, relationships between some of the intrusions and gravity features are discussed under “Gravity Interpretation” above.

Bouldercombe Complex (BC)

The image of airborne. magnetic data over the Bouldercombe Igneous Complex is shown in Figure 8. There are two near surface lobes to this complex:
Coordinate at approximate centre of western lobe: GR216000 E, 7390000 N
Coordinate at approximate centre of eastern lobe: GR239000 E, 7398000 N

Initially, the complex was viewed as two lobes forming a dumbbell shape striking approximately east-west (initially evident on algorithm BC-A). On further inspection, major intrusions were interpreted extending to the north at depth (see algorithms BC-B and BC-C).

Timing relationships were interpreted between the various bodies (see algorithm BC-D).

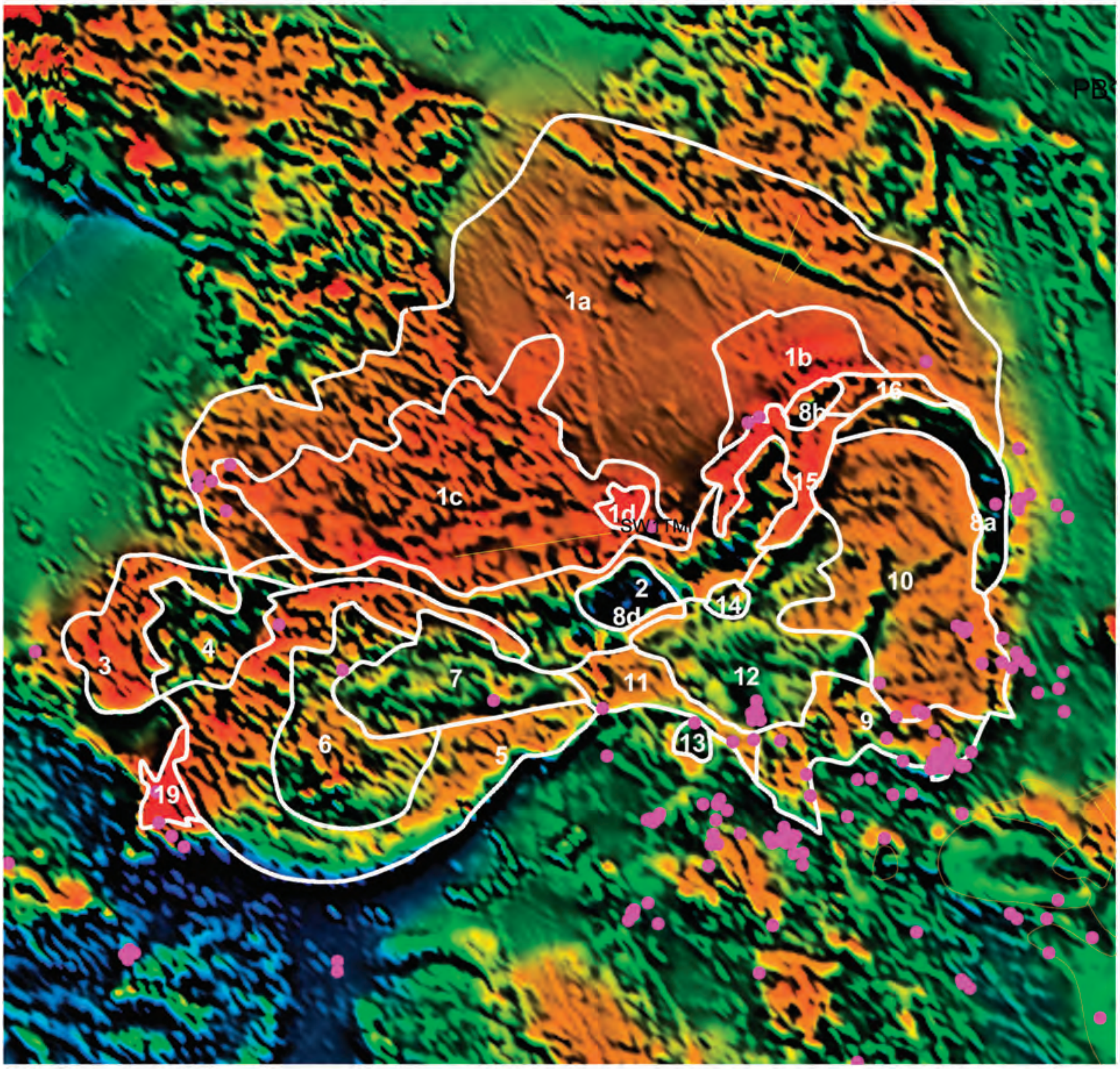


Figure 8. Image of airborne magnetic data over the Bouldercombe Igneous Complex

Body 1 was the largest body being 18km north-south and 34km east-west. It dipped to the north-east and extended under a thrust front at Gracemere. It was interpreted to be composed of several sections (1a to 1c) which were not necessarily different phases but more likely different depths to the same unit. It was unclear whether 1d was also a shallow part of the intrusion. It may have been a different intrusion or have some magnetic alteration enhancement. (*Body 1 does not crop out*).

Body 1 was interpreted to be the oldest with bodies 2, 8 and 15 all directly intersecting it. The north and east boundaries were not interpreted with any confidence as they were very deep. 1 was intersected by 2 on the western side where 1 was shallowest. Body 3 was intruded by 4, which also intruded 2. No clear relationship could be established between 1 and 3. (*Body 3 is unnamed diorite and gabbro. Bodies 2 and 4 do not crop out, their locations being covered by younger volcanic and sedimentary units*).

Body 5 had a curved margin and clearly intruded 4 and 2. Body 6 was interpreted to be different to 5 and intruded it. If the more magnetic anomalies in the north were the result of alteration, it could be possible for 5 and 6 to be the same body. 7 intruded both 6 and 5 and was interpreted with greater confidence to be a separate body to 5 based on the differences in response on the southern margin of 7. (*Bodies 5, 6 and 7 correspond to the Flaggy Quartz Monzodiorite, the Umbrella Granodiorite and the Moonkan Granite respectively*).

Body 8 was interpreted to be in 4 parts. They were correlated based on their strong negative response. The body was interpreted as emplaced with a negative remanence and later intersected with other bodies (see below). 8d was a particularly strong negative and may have been a separate intrusion with negative remanence. (*Body 8d is the Quarry Gabbro. 8b and 8c, together with 15, comprise the Kabra Quartz Monzodiorite, and 8a together with 16 and part of 10 is the Gracemere Gabbro*).

Bodies 10 and 11 looked very similar in character and were possibly the same body. 10 intruded 8 and possibly 9 if 9 was a separate body. There was a character change between 9 and 10 so they were interpreted as separate bodies. 11 intruded body 2 but probably not body 7. 11 may even have been the same body as 5, given the similar character of 5 along its south-east margin. 12 was interpreted to intrude bodies 8, 9, 10 and 11. It had quite a different character to the others. (*Bodies 10, 11 and 12 are all mapped as the Bundaleer Tonalite, in places covered by younger sediments of the Precipice Sandstone. Body 9 is the Gavial Gabbro and the Devonian Capella Creek Group*).

Small intrusions 13 and 14 were interpreted to be later than 11 and 12 respectively. 13 appeared adjacent to 11 and possibly intruded it while 14 clearly intruded 8 and 12. (*Body 13 is Devonian Capella Creek Group, and 14 is an unnamed intrusive unit in the Bouldercombe Igneous Complex*).

Body 15 was highly magnetic and was likely to be the result of several factors. It had at least part of its response attributed to magnetic alteration aureole. The western limb was probably attributed to the emplacement of 8 and was consistent with the aureole extending around the eastern boundary of 8. The eastern limb of 15 could have been attributed to the emplacement of 10 and 12 and was consistent with the eastern boundary of 10 having a magnetic rim. A more difficult problem was the explanation for the magnetic high intersecting the magnetic lows of 8. This was attributed to alteration effects along a fault resetting the magnetic remanence and/or emplacing additional magnetite. (*Body 15, together with 8b and 8c, has been mapped as the Kabra Quartz Monzodiorite*).

Body 16 could have been a separate intrusive event between 8a and 8b but could also have been an alteration effect.

More detailed work should reveal which scenario is better suited to the facts.

Bajool (B)

Coordinate at approximate centre: GR255000 E, 7378000 N

Langmorn (L)

Coordinate at approximate centre: GR266000 E, 7360000 N

The image of airborne. magnetic data over this area is shown in Figure 9.

Bajool was an intersecting series of intrusions. It is interpreted here with the Langmorn series of intrusions, as there was a link between the two on the magnetic interpretation (*Langmorn equals the Cecilwood Quartz Diorite*). On the gravity interpretation, there was a major fault and gravity ridge evident between the two and Langmorn appeared to be linked to and similar to Galloway Plains.

Automag solutions on a line in the north (see model MG-A, Figure 10a), indicated steep dips and a gentle deepening to the north. Most dips were to the north except for a strong anomaly in the centre of the line which dipped steeply to the south.

The overall dip of the main bodies as indicated by the separate anomalies was to the north except at the northern end where it crossed onto some shallow anomalies. The other exception was the central strong anomaly which was also shallower.

Forward modelling confirmed the Automag solutions (see model MG-B, Figure 10b), but depths were slightly greater. Line MG1 was noticed to be slightly offset from the maximum of the central anomaly so a third line was modelled over the peak maximum (see model MG-C, Figure 10c). This confirmed a northerly dip in the main body from around 350m at the southern end of the line to over 500m towards the northern end. At the very northern end, the anomalies shallowed as the line crossed onto another phase.

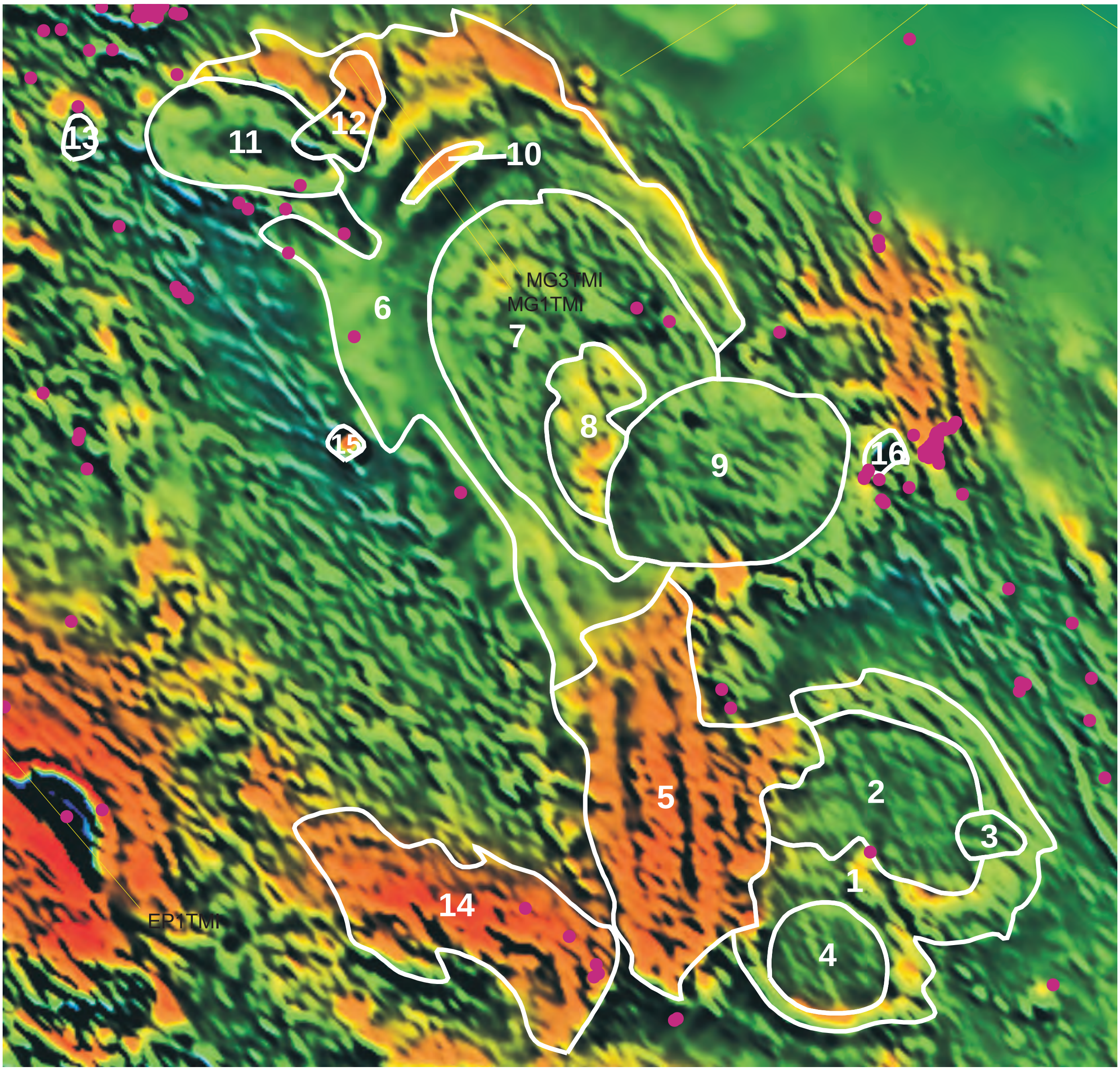


Figure 9. Image of airborne magnetic data over the Bajool Quartz Diorite and Cecilwood Quartz Diorite

The central anomaly was both shallower and dipped south rather than north for all the other anomalies. It was interpreted to be either an alteration effect or skarn developed above the intrusion. Its depth was modelled at about 400m but it was likely that it was shallower than this, given that both alteration and skarn magnetite would not have sharp boundaries.

The various phases were interpreted to establish timing relationships. Body 1 was interpreted to be the oldest as all others could be shown to intersect it. Bodies 2, 3, 4 and 5 all directly intersected it. Body 2 intersected 1 and was intersected by both 3 and 5. (*Bodies 2 and 4 are mapped as a single intrusive phase, the Cecilwood Quartz Diorite, with Body 1 primarily comprising the surrounding contact aureole*).

It was not clear whether body 5 was an intrusion or simply had alteration enhanced susceptibilities. It did not have a classic intrusion character and the internal fabric had a north-south strike. It may have had an intrusion below it. In any event, it was intersected by 6. (*Body 5 consists of the Devonian Capella Creek Group and an unnamed Devonian gabbro. To the north it is mainly covered by Quaternary alluvium*).

It was not clear whether 7 was another intrusion inside 6 or whether 7 was simply the shallow part of 6. As established in the modelling, the dip was away from 7 so the interpretation was ambiguous. Body 8 intruded 7 and 9 intruded both 7 and 8. (*The Bajool Quartz Diorite is largely covered by Quaternary alluvium with very few fresh outcrops, and Bodies 7, 8, 9 and 16 are all included in it*).

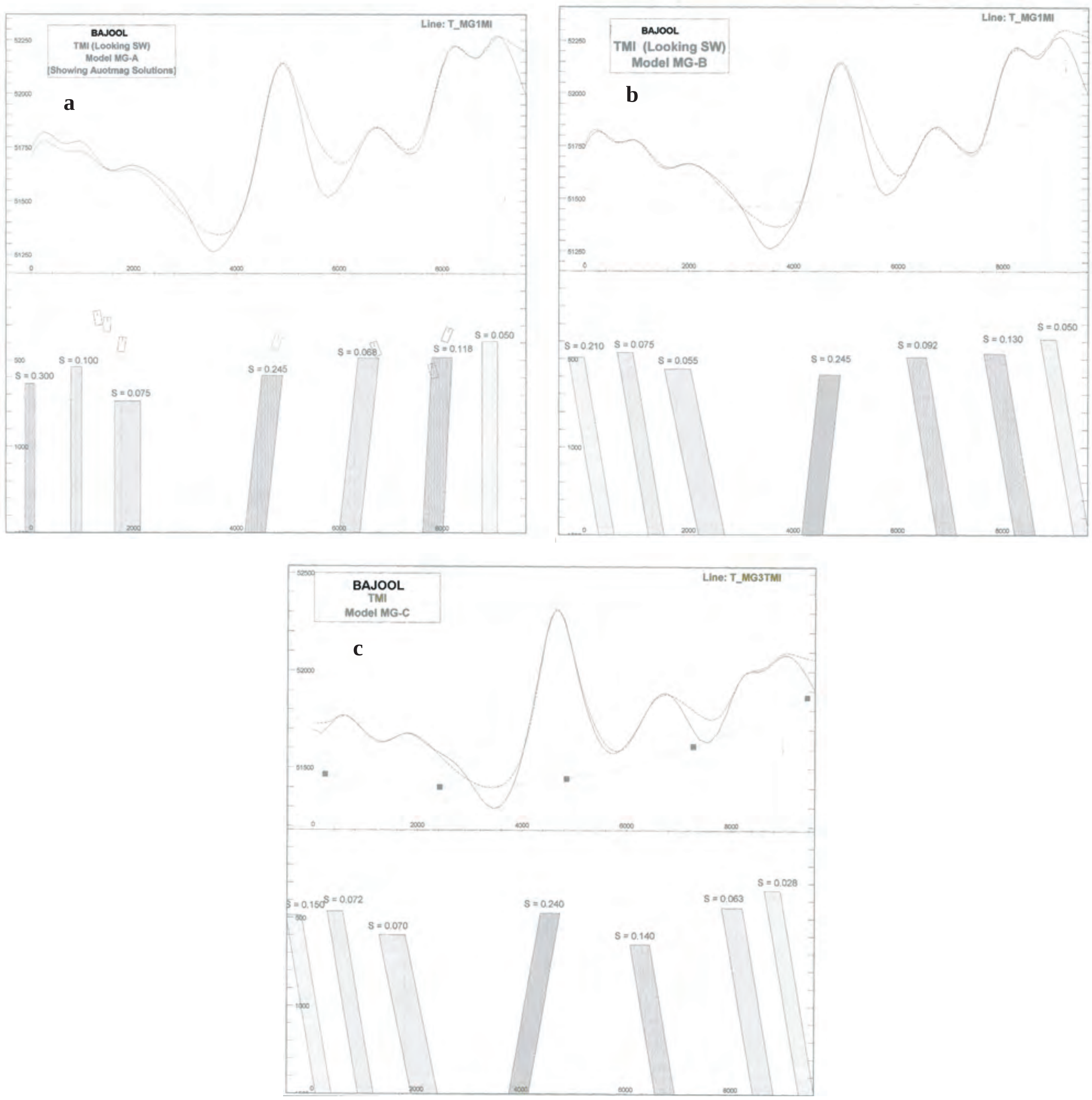


Figure 10. Models across the northern boundary of the Bajool Quartz Diorite

Body 10 was the possible skarn or alteration discussed above. (*Bodies 6 and 10, which are covered by Quaternary alluvium, are interpreted as country rocks surrounding the Bajool Quartz Diorite*).

11 intruded 6 and was intruded by 12. 12 was interpreted as a separate intrusion, however it may have been alteration above intrusion 6. The latter explanation was attributed to magnetic highs at the northern edge of body 6. (*Body 11 is an unnamed granodiorite, and 12 is mapped as country rocks of the Upper Devonian-lowest Carboniferous Mount Alma Formation*).

Body 13 was a separate small intrusion which had a weak aureole. (*This unnamed granodiorite intrusion hosts the Struck Oil porphyry copper-molybdenum prospect*).

It was not clear whether body 14 was an intrusion at the surface or had alteration enhanced susceptibilities due to an underlying intrusion. It was thought more likely to be alteration at the surface but may have had an intrusion below. It did not intersect any other bodies. (*Body 14 includes Devonian Capella Creek Group and a very small intrusion of Devonian granitoid*).

Small intrusions were interpreted at 15 and 16. 15 was a sharp high and was either an outcropping unit or some form of culture. It had a weak linear anomaly striking north-north-west through it which might have been a dyke or a power line. (*Body 15 is the TV tower on Mount Hopeful, with a power line extending to the north-north-west*). 16 was thought unlikely to be culture as it was negative. It was more likely to be a small gabbro with negative remanence. (*It is included in the Bajool Quartz Diorite*).

(*A feature of this magnetic image is the presence of a north-north-westerly trending fault with a small sinistral displacement which cuts both the Cecilwood Quartz Diorite and the Bajool Quartz Diorite. This fault can be traced as far south as Brisbane, and various names have been applied to it*).

Eulogie Park (EP)

The image of airborne magnetic data over the Eulogie Park Gabbro is shown in Figure 11.

Coordinate at approximate centre: GR237000 E, 7358000 N

The Eulogie Park intrusive complex was dominated by a strong negative magnetic response in the centre (see algorithms EP-A and EP-B). It also had a major fault cutting it.

Timing relationships were established (see algorithm EP-C). Body 1 was the first body emplaced. Body 2 could have been the same as body 1 across a major fault but it had a stronger susceptibility, more similar to body 3. If it were the same as body 1 then there would have to be significant vertical displacement to bring different phases of the same body adjacent. It was a simpler interpretation to have it the same rock type as body 3, which was the favoured interpretation.

Body 4 had significantly stronger susceptibility and had a different radiometric response to both 3 and 5. It was therefore interpreted to be at least a different phase of either of these two bodies but thought to be a separate intrusion. With either interpretation, there would need to be significant vertical displacement to emplace the narrow silver in its current location.

Body 5 had a strong negative magnetic response and a different radiometric response and was interpreted to be a separate intrusion. It intruded body 2 and was therefore probably later than 3 but no relationship could be established between 4 and 5. The western margin of 5 was faulted with significant vertical displacement so it was not possible to establish whether 4 or 5 was first. (*Mapping indicates that Body 5 is the Eulogie Park Gabbro, a layered gabbro with reverse remanent magnetism. The outer rim of this layered gabbro has positive remanent magnetism and includes part of Body 2. The remainder of Body 2 is interpreted as a contact aureole. Bodies 1, 3 and 4 are an unnamed diorite, which in Body 1 includes large inclusions of layered gabbro like that of Body 5. Body 6 is an unnamed diorite with a strong magnetic aureole. Body 7 does not crop out*).

Two models across body 5 were conducted to determine whether the margins had any significant dip (Figure 12). Both were 2D models and the first model used strikes perpendicular to the traverse. The second used strikes more consistent with the real strike interpreted from the image. Both needed to have a negative body in the south to make the model fit but used a low regional to simplify the modelling process.

The shallowest depth on the northern margin was 220m while the southern negative anomalies were more consistent with depths from 320 to 450m. The deeper central anomalies had depths around 500m.

Craiglands (CIC)

The image of airborne. magnetic data over the Craiglands Quartz Monzodiorite and the Galloway Plains Igneous Complex is shown in Figure 13.

Coordinate at approximate centre: GR248000 E, 7334000 N

A series of intrusions north of Biloela are aligned roughly east-west with Galloway Plains to the east. They include the Craiglands Quartz Monzodiorite, previously mapped as the Mount Gerard complex. The majority of these intrusions were interpreted to be older than the majority of Galloway Plains since body 34 was intruded by body 4 of Galloway Plains.

The oldest intrusions were interpreted to be in the west with body 23 being the deepest and possibly the oldest. There was no evidence to support it being the oldest other than the general trend of older in the west to

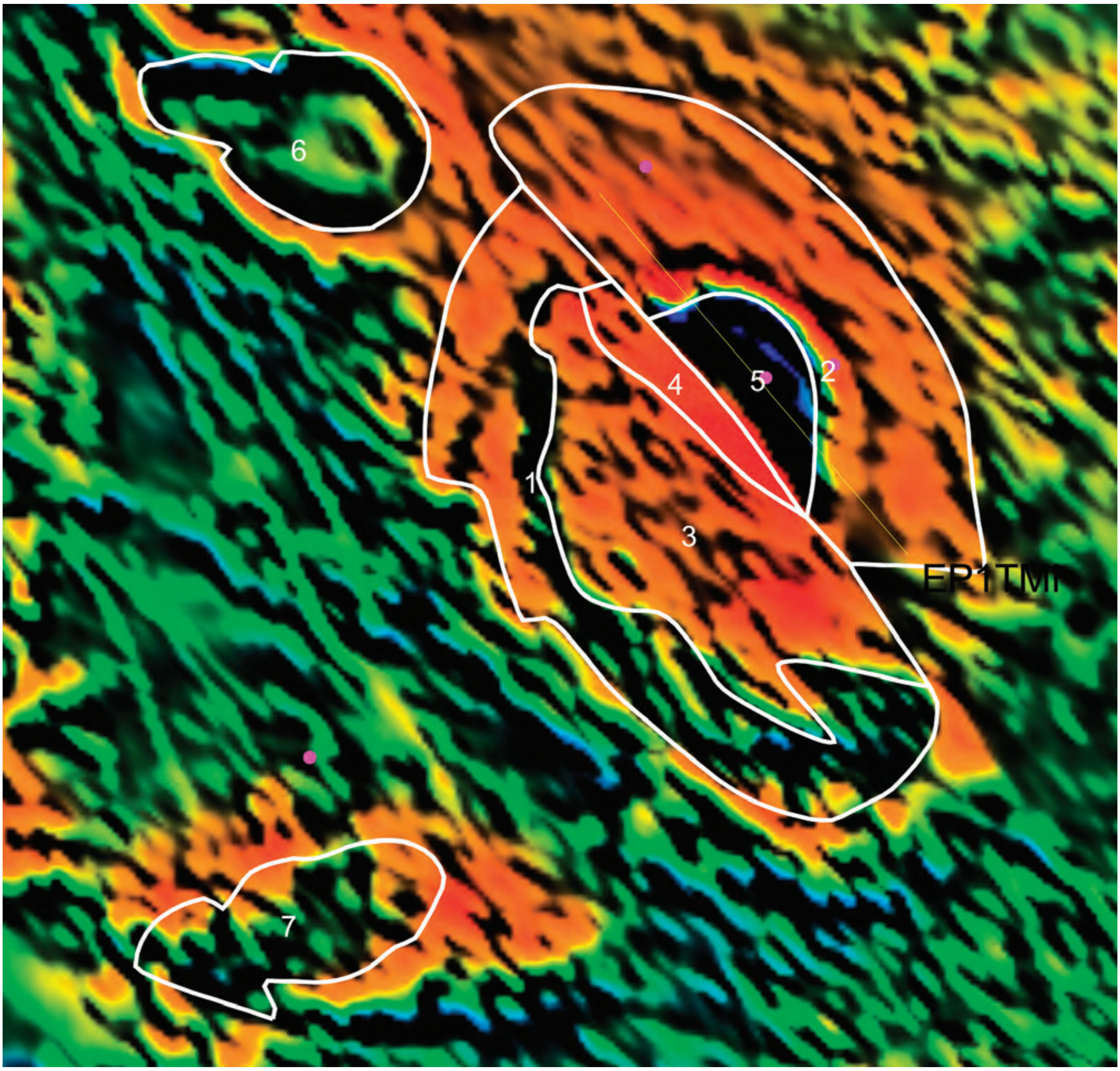


Figure 11. Image of airborne magnetic data over the Eulogie Park Gabbro.

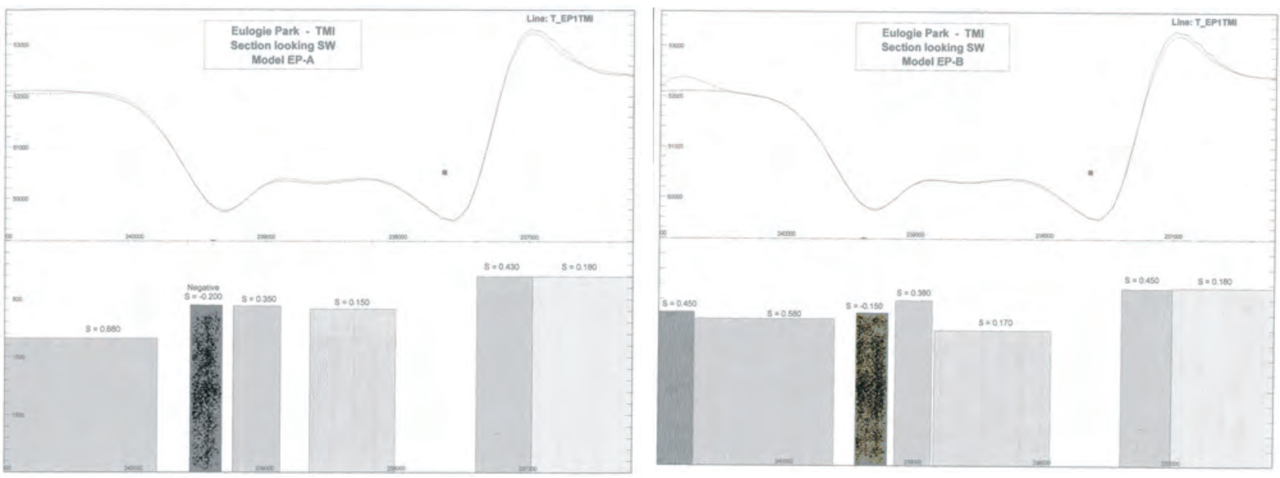


Figure 12. Models across the Eulogie Park Gabbro

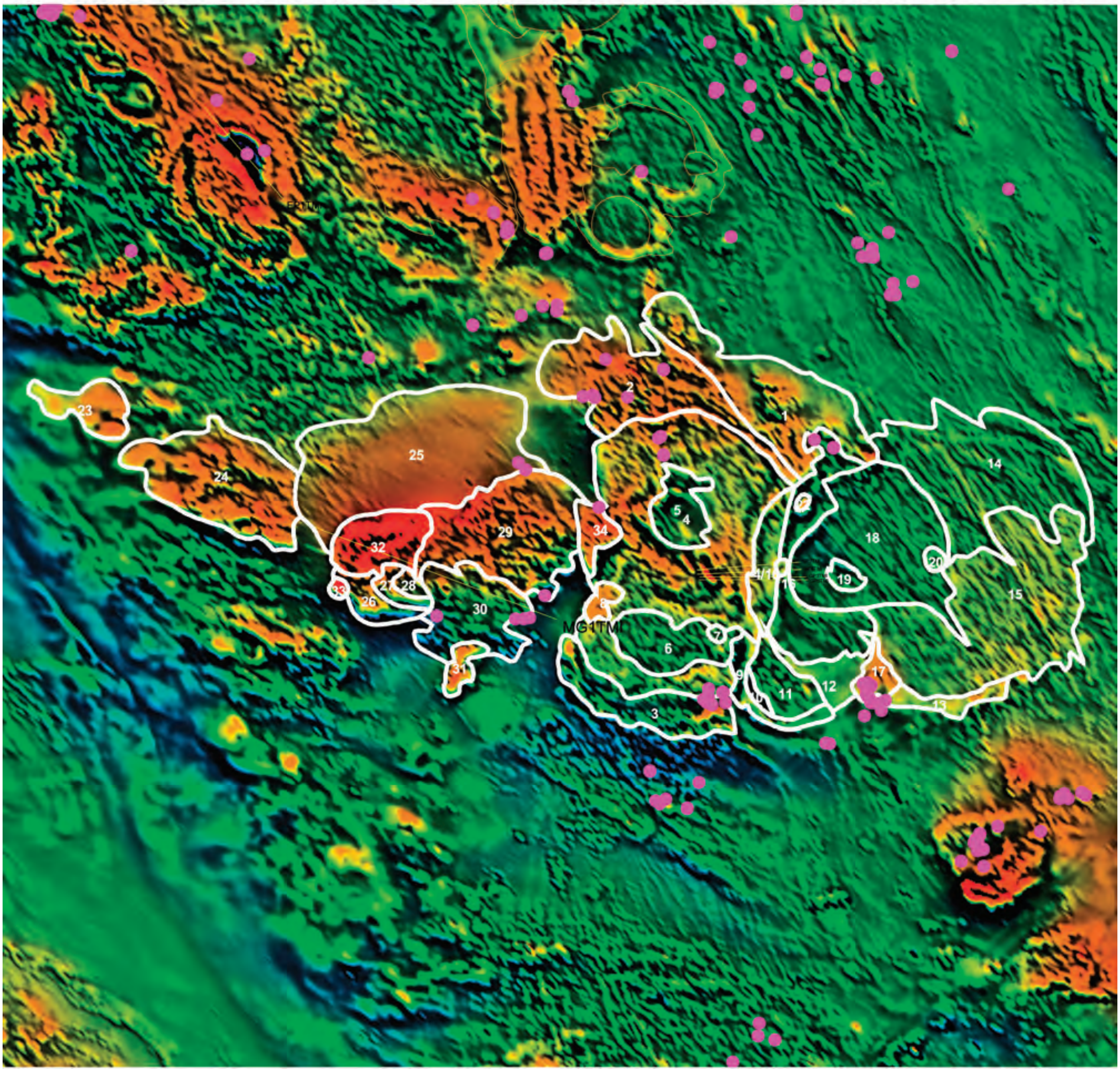


Figure 13. Image of airborne magnetic data over the Craiglands Quartz Monzodiorite and Galloway Plains Igneous Complex

younger in the east which pervaded both the Craiglands intrusives and Galloway Plains. (*Body 23 is concealed beneath the Tertiary Biloela Basin*).

Body 24 was older than all other bodies in the Craiglands and Galloway Plains with the possible exceptions of 1, 2, 3 and 23. No relationship could be established between these four and 24 or directly between 24 and 26, 27 and 28. (*The south-eastern corner of Body 24 is mapped as the northern part of an unnamed Permo-Triassic granitoid. The abrupt change in magnetic character of the mapped granitoid suggests that it consists of more than one phase. The north-eastern part of Body 24 is the Permian Smoky beds, and the western two-thirds is buried beneath the Biloela Basin*).

Body 25 was deep and uniform but appeared to intersect 24. Alternatively, 25 could have been an extension of body 29 but the transition at the boundary of 29 was too sudden and no fault was interpreted. (*Body 25 is not exposed. The rocks are Lower Permian Youlambie Conglomerate*).

No relationship could be established between 26 and 25, but 26 was intruded by 27 which was subsequently intruded by 28. They were interpreted as separate bodies rather than different phases because the susceptibilities varied from high in 26 to low in 27 to moderate in 28. All bodies had alteration aureoles. The big low to the south of 26 was interpreted to be due to the combined effect of 25, 26 and 32. (*Bodies 26, 27*

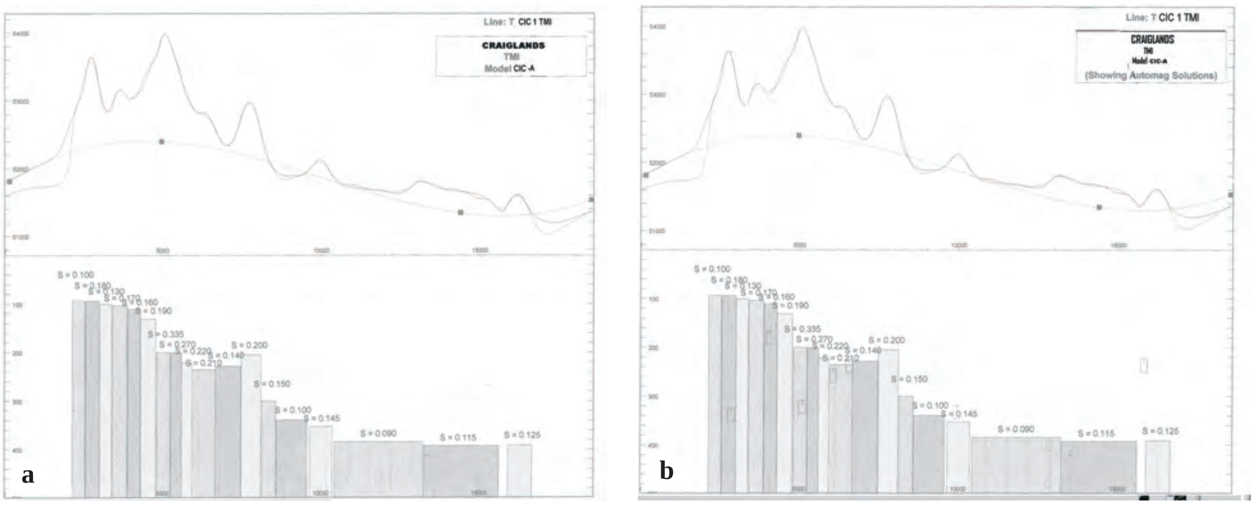


Figure 14. Model across the Craiglands Quartz Monzodiorite

and 32 are all part of the Craiglands Quartz Monzodiorite, which is quite uniform in outcrop and magnetic susceptibility. Body 28 is a small outcrop of Smoky beds).

Bodies 28 and 25 were intruded by 29 which was intruded by 30. 29 and 30 were interpreted to be different bodies because of the small but sharp change in susceptibility across the boundary. (*Body 29 is mainly the north-eastern portion of the Craiglands Quartz Monzodiorite and the co-magmatic Inverness Volcanics, together with a small part of the Smoky beds. This portion of the Craiglands Quartz Monzodiorite has a lower magnetic susceptibility than the western outcrop. Body 30 is the major outcrop of the Smoky beds*).

Body 30 was intruded by 31 and 26 by 33 while 32 intruded all bodies from 25 through to 29. Bodies 31, 32 and 33 were all interpreted to have significant magnetic alteration enhancement of their susceptibilities. (*Body 31 is an unnamed granodiorite pluton, and Body 33 is a small gabbro intrusion that cuts the Craiglands Quartz Monzodiorite*).

Body 34 intruded 29 and was intruded by body 4 of Galloway Plains. (*Body 34 is an unnamed gabbro which is included in the Galloway Plains Igneous Complex*).

A model section was generated across bodies 32 and 30 (see model CIC-A, Figure 14a). This showed the anomalies from the magnetic intrusive units close to outcropping in the west and becoming deeper to the east to a depth of about 300m. The deepening occurred in several stages corresponding to the different units traversed. Automag solutions confirmed this trend and correlated well with the forward model (Figure 14b).

Galloway Plains (GP)

Coordinate at approximate centre: GR276000 E, 7332000 N

The Galloway Plains Intrusive Complex exhibits several intrusive phases that extend through to the Craiglands intrusions. The relationship between some of these phases could easily be determined, but in the centre, magnetic responses from arcuate rims of adjacent intrusions overlie each other, making it difficult to determine the sequence of events.

Initial modelling was essentially 2.5D (ie cross-section with limited strike, see Figure 15) and focussed on several widely spaced east-west profiles across the area of intersecting rims (see model GP-A, Figure 16a). These only served to confuse the picture as shallow bodies to the east were adjacent to deeper bodies in the centre and intermediate depth bodies further west. Automag was attempted to try to gain more information (see model GP-B, Figure 16b). This showed several deeper bodies getting deeper to the west with several shallower near surface bodies persisting across the whole area.

Forward models with deep bodies only could not match the data so models with both shallow and deep bodies were attempted. Interpretation of the image suggested that the shallower bodies had NNW strikes while the deeper rim anomalies had a north strike at this location. To combine these two, a 3D model using several close spaced lines was attempted (see model GP-C, Figure 16c-f).

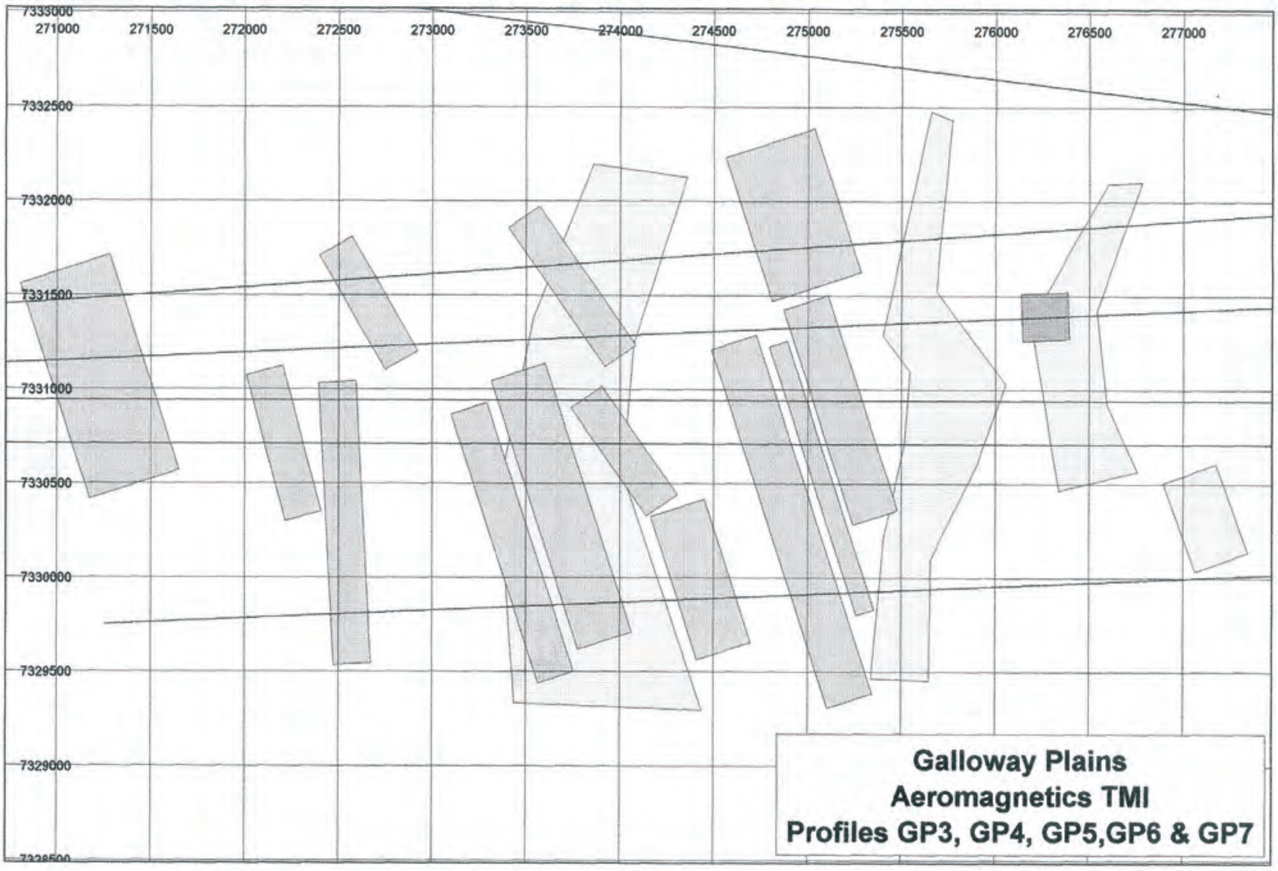


Figure 15. Locations of profiles for models across part of the Galloway Plains Igneous Complex

The modelling using 2D did not differentiate different strikes at different depths. Once a 3D model was used, the shallower north-north-west striking elements of the western intrusive phase could be distinguished from the deeper north striking elements of the eastern phase. The rims of the eastern phase were modelled using bodies that became successively deeper to the west indicating a westerly dip in the top of the eastern phase. The tops of bodies representing the western phase remained shallow at about 50 to 100m depth.

There were two possible explanations for the model. One had the eastern phase emplaced and then faulted with a normal fault close to but internal to its western margin. The western phase then intruded against this fault and overlay the deeper rim of the eastern phase. A more likely explanation had the western phase emplaced and the eastern phase then intruded below it. A high angle fault then lifted the eastern side by at least 200m to bring the current configuration into place.

Additional evidence for this model was found in the change in depth of the rim of the eastern phase across a fault (GR274800 E, 73237000 N). The fault between the two phases was primarily evident on the radiometrics and had a northerly strike at about 276000 E at the northing of the models.

Other phases were interpreted with timing relationships (see algorithm GP-D).

The earliest bodies were 1 and 2. These were probably the same but had a fault boundary between them. (*These bodies have been identified during mapping as contact metamorphosed rocks of the Rockhampton Group, Mount Alma Formation, Balaclava Formation and Raspberry Creek Formation*). The relationship to body 3 could not be established but all of 1, 2 and 3 were intruded by 4. (*The majority of Body 4 equates to the Rocky Point Granodiorite, and Body 3 is a contact aureole on the southern side of the Galloway Plains Igneous Complex*). The gradual increase in magnetic response to the north across body 4 was consistent with a normal magnetic response across a wide body. It was possible that the dumbbell shape of 4 was the result of two separate intrusions but no clear boundary existed and the composition of the north and south parts appeared very similar. (*Based on geochemical analyses, part or all of Bodies 4, 6, 7, 8, and 9 are differentiated as the Rocky Point Granodiorite*).

The majority of the Craiglands intrusions were interpreted to be older than the majority of Galloway Plains since body 4 intruded body 34 of the Craiglands intrusions.

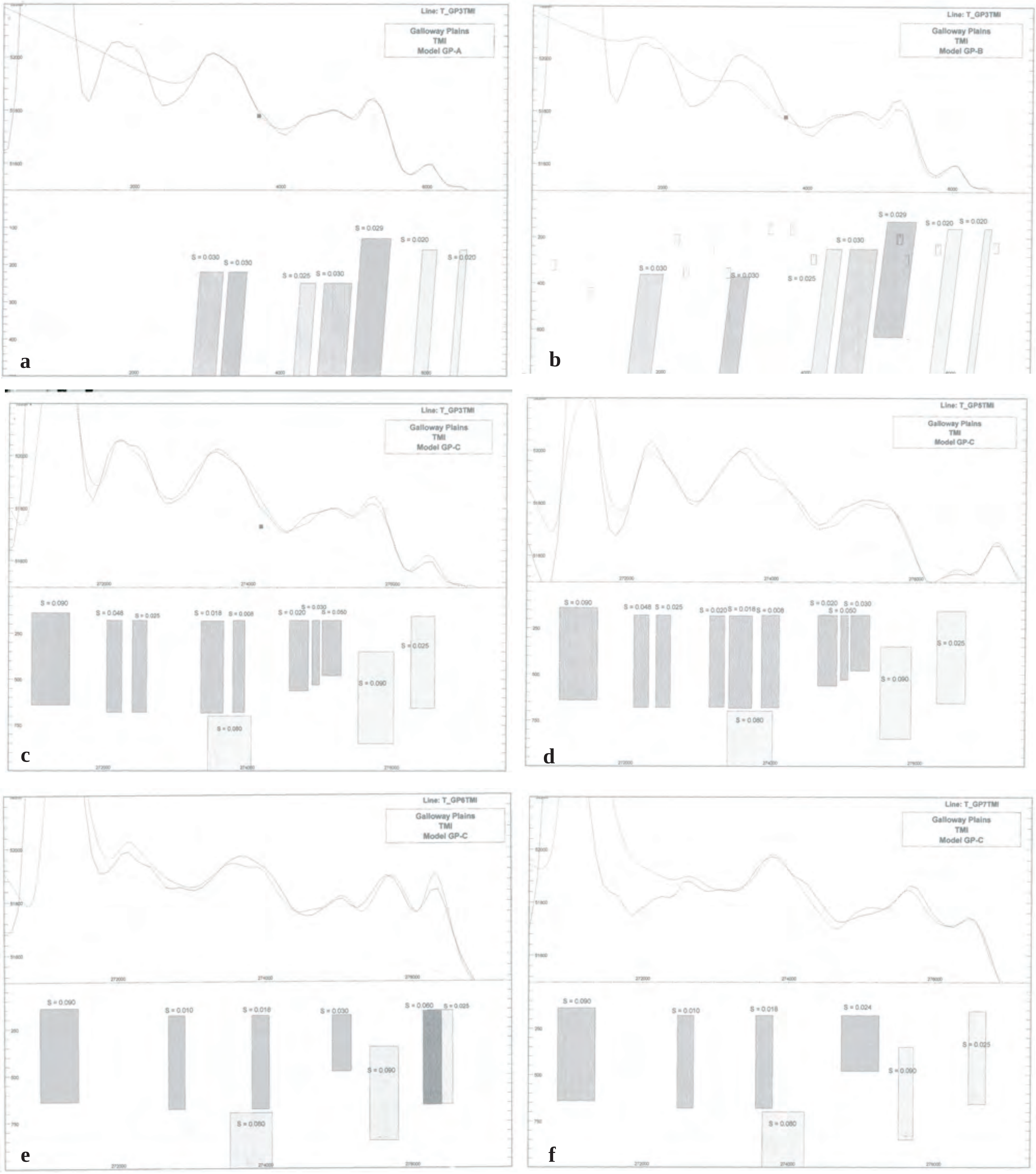


Figure 16. Models across part of the Galloway Plains Igneous Complex

Body 5 intruded the middle of 4 and generated some magnetic alteration around it. More intense magnetic bodies in the centre of 4 were thought to also be a product of intrusion 5 underneath generating alteration effects. (*Within Body 4, the Dumgree Tonalite is differentiated from the Rocky Point Granodiorite by chemical composition. Body 5 corresponds to the Redshirt Granite*).

Body 6 intruded 4 and was intruded by 7. Body 8 also intruded and had a higher susceptibility.

Body 9 looked like it was intruded by the southern part of 4 but looked like it intruded 6. This circular paradox was explained by using a similar mechanism as proposed from the model. It was interpreted that 9 was later than both 6 and 4 but the margin of 4 overlaid and obscured the margin of 9.

Body 10 may have been an alteration aureole around 11 but had a negative response. If it was a separate intrusion it was likely to be earlier than 11 which almost entirely replaced it.

Body 11 intruded 4, 9 and 10 and was intruded by 12 and 16. The boundary between 11 and 12 was primarily across a fault but the small part of it not affected by the contact aureole of 16 suggested it intruded 11. If the whole of the aureole was attributed to 16 then it may be possible for 11 and 12 to be the same intrusion. (*Bodies 10, 11, 12, 14, 15, 16, 18 and 20, and the southern part of Body 9, are included in the Bocoolima Granodiorite. This unit is deeply weathered and very poorly exposed, and probably contains a number of different phases. The difference in magnetic response between Bodies 11 and 12 is largely attributed to topographic and weathering effects*).

Body 13 could have been a contact aureole for 15 but was broader and less regular than the others. It was therefore interpreted to be a separate intrusion that was almost replaced by 15. 13 was thought unlikely to be the result of 15 dipping south because there was no evidence on the south-eastern boundary for this. (*Body 13 is mapped as the contact aureole of the Bocoolima Granodiorite in Three Moon Conglomerate*).

Body 15 had a slightly more intense response than 14 and was therefore interpreted to be a separate body. Most of the boundaries between the two were faults and it may be possible for 15 to be a different phase of 14. However, in one location on the northern boundary of 15 (GR291500 E, 7335600 N), the convex shape of 15 indicated it intruded 14.

Body 16 intruded 11, 12 and 15 and was thought likely by the modelling to also intrude 4.

Body 17 was possibly alteration and intersected 12, 13, 15 and possibly 16. 18 also intersected 16 and had an alteration rim. It was interpreted to be different to 16 because 18 intersected the margin of 16, just north of 17. (*Body 17 is the contact aureole of the Bocoolima Granodiorite in Three Moon Conglomerate, and is the location of the Maxwelton goldfield*).

Bodies 19 and 20 both intruded 18, with 20 being particularly visible on the radiometrics. (*Body 19 corresponds to the western outcrop of the Voewood Granite*).

Bodies 21 and 22 both intruded 16 but they may have been very late. (*Body 22 has been drilled and found to be granodiorite or tonalite no different from the Bocoolima Granodiorite. The magnetic source must lie at greater depth*).

Diglum (DL)

Coordinate at approximate centre: GR306000 E, 7312000 N

Diglum West (DW)

Coordinate at approximate centre: GR291000 E, 7311000 N

The image of airborne. magnetic data over the Diglum–Kroombit Tops area is shown in Figure 17.

Diglum, Diglum West and Kroombit Tops form an interconnecting series of intrusions. Kroombit Tops appears to intersect Diglum and is discussed separately below. The whole area was interpreted together to enable the relationships to be interpreted.

The traditional image featuring TMI or 1VD showed much detail (see algorithm DK-A) but did not resolve either the subtle margin or some of the intrusions. The weak magnetic responses of Kroombit Tops were best interpreted using greyscale 1VD (see algorithm DK-B) while the deeper intrusions were best interpreted using TMI on TMI vertical shade (see algorithm DK-C).

Many of the boundaries were faulted and the fault interpretation was overlaid before attempting to interpret the intrusive boundaries. Several intrusives had alteration margins that were mapped as separate phases in the intrusion interpretation but were omitted in the summary of the timing (see algorithm DK-D).

Diglum North and Diglum north-west correspond to shallower areas of the whole complex. Deeper intrusions were evident as broad areas of high magnetic signature but their margins were difficult to map. This was because of their dipping flanks which made it difficult to detect a sharp change in magnetic level and because

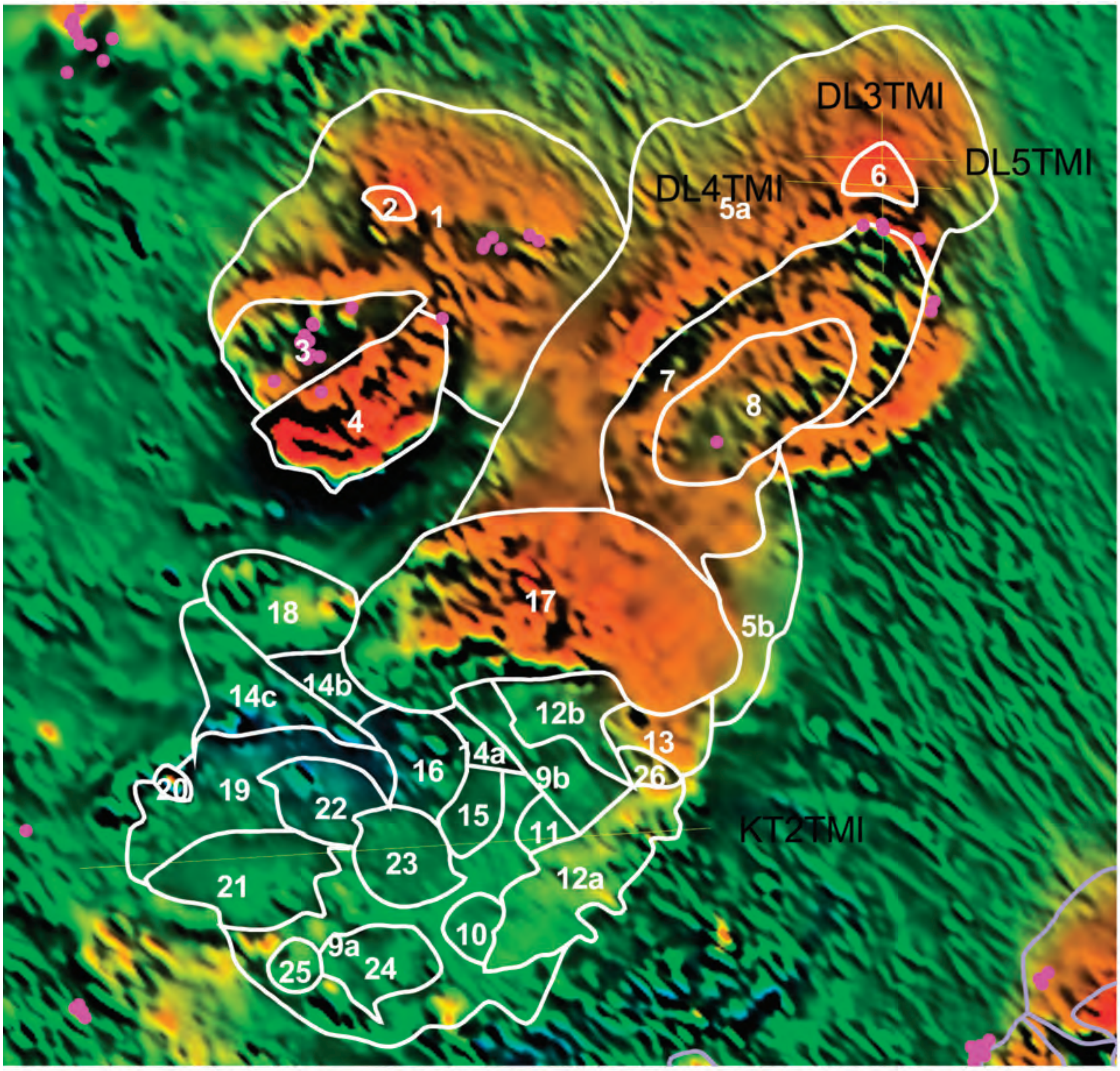


Figure 17. Image of airborne magnetic data over the Diglum Granodiorite, Mount Seaview Igneous Complex, and the Kroombit Tops region.

of the ambiguity of mapping the margin of a 3D body in 2D. The margin was interpreted to be the limit of the intrusion at all the depths which gave a contribution to the signal, probably less than 2km.

Two shallow intense magnetic anomalies were named as Diglum north-west (GR292700 E, 7317000 N) and Diglum North (GR308600 E, 7317800 N). Detailed inspection revealed both magnetic highs had their margins structurally controlled and were composed of several bodies. Their intense magnetic highs were therefore interpreted to be one of three possible scenarios:

1. They may be a very magnetic intrusion such as a gabbro or may be a near surface extension of a deeper intrusion below each. They may have preferentially intruded along pre-existing faults that gave their linear form and correlation with structures. This was thought the least likely because many of the faults postdate the intrusion as evident by the margins being displaced by faults.

2. The magnetic units may be skarns developed above the deep intrusion. These could have preferentially developed along structures.

3. The structures may alternatively host magnetic alteration emplaced by solutions coming from the intrusions or from solutions originating via dewatering of surrounding units. Convection currents set up by the underlying intrusions could have provided a mechanism for focusing the solutions into the faults.

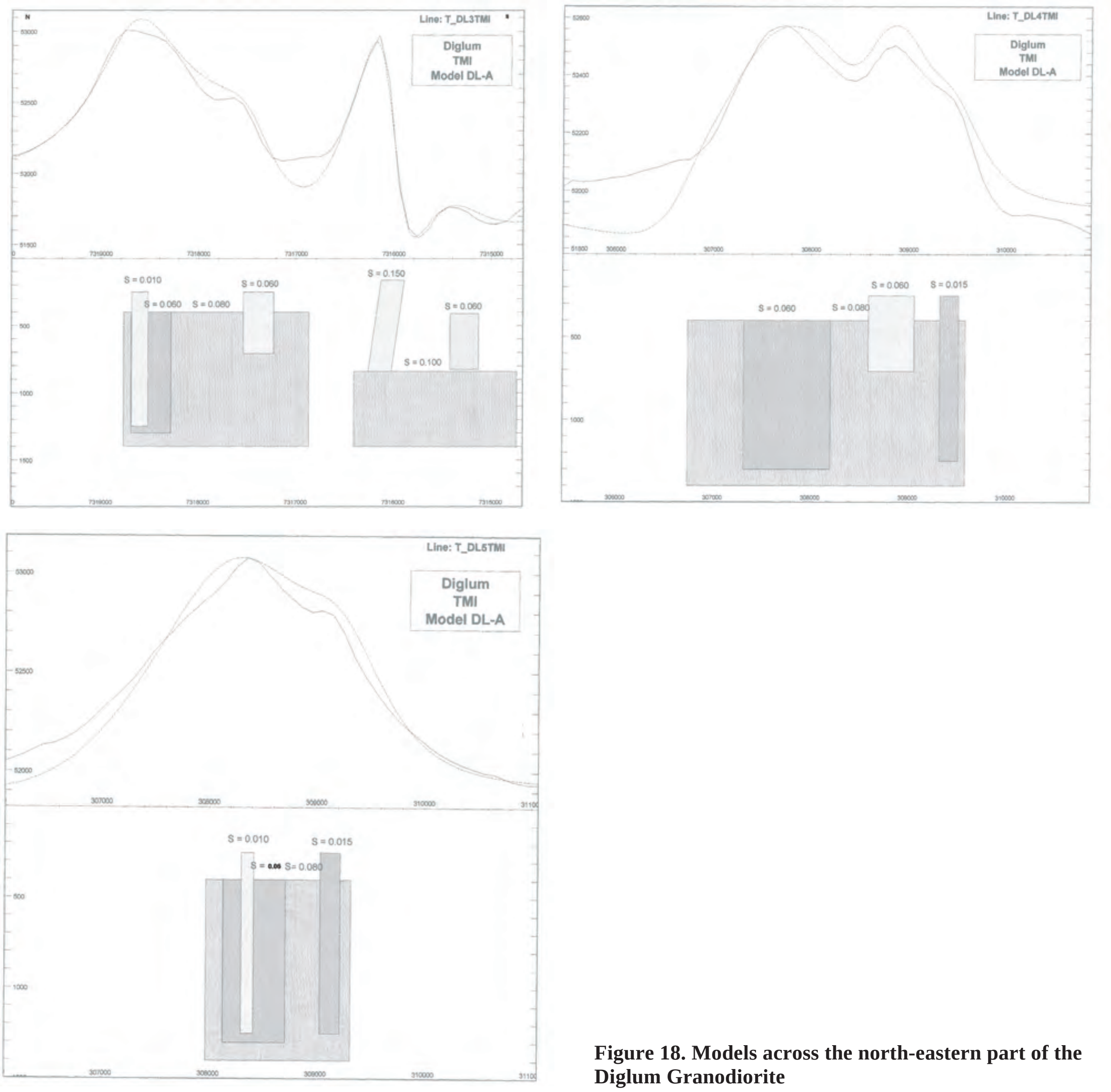


Figure 18. Models across the north-eastern part of the Diglum Granodiorite

Either of the last two scenarios indicated a potential site for mineralisation. (*Diglum north-west or Body 2 is an unnamed granodiorite, and Diglum North or Body 6 is a magnetite-rich gabbro that is shallowly buried beneath alluvium and soil and has been drilled during mineral exploration*).

Modelling was conducted in 3D on Diglum North (see model DL-A, Figure 18) because many of the bodies were short strike length. The starting model was devised from the image and the final model was consistent with the image. The fit was not perfect as the model was limited to bodies with consistent susceptibility. Some variability was simulated by having bodies within bodies. The total susceptibility of the inner bodies was then the sum of its own susceptibility and the susceptibilities of all enclosing bodies.

Depths in the model ranged from 100m to 320m. It was thought that the deeper depths were probably an over estimation as the variability of susceptibility would make the bodies appear deeper than in reality. Depths of between 100 and 200m were thought likely.

The timing of the various intrusive bodies was difficult to establish uniquely (see algorithm DK-D). It is likely that body 1 was the earliest because it can be shown to be intersected by all other bodies. Diglum north-west (body 2) postdates body 1 unless it is an extension of the same body. If not, it could be at any time after intrusion 1. (*The interpretation of Body 1 is based purely on geophysics, with no surface expression*).

Intrusions 3 and 4 intersect 1, have different character, and are separated by a fault. They were interpreted to be the same body since intrusion 3 has a magnetic aureole which could have covered the body if it was emplaced at depth. The southern half could then have been faulted down with the present surface preserving the covering aureole in the south but intersecting and removing it in the north. (*Bodies 3 and 4 are mapped as granitic and dioritic phases of the Mount Seaview Igneous Complex. The magnetic aureole to the north is also diorite. The magnetics indicates a subsurface extension of the diorite to the south*).

Intrusion 5a intersects 1 and body 5b was interpreted to be the same body based on a similar depth and form. Diglum North (body 6) postdates 5a unless it is a shallow extension of the same. Body 7 intrudes 5a and has a significant aureole developed. Body 8 is inside body 7 and has been interpreted as a separate event because its southern boundary does not parallel that of body 7. (*Bodies 7 and 8 correspond to the two phases of the Diglum Granodiorite, and Body 6 is a magnetite-rich gabbro mapped in the Diglum Granodiorite but which may represent a separate intrusion. These are surrounded by a strongly magnetic contact aureole in the Mount Alma Formation and Rockhampton Group that represents the inner part of Body 5*).

Body 5 is intersected by 13 and body 17 intersects 7 and 5, indicating that most of the Kroombit Tops intrusive complex is later than Diglum. These are discussed further under Kroombit Tops (see below).

Kroombit Tops (KT)

Coordinate at approximate centre: GR293000 E, 7296000 N

Kroombit Tops consists of a series of intrusive phases in a complex relationship. These were subsequent to the series of intrusions which made up the Diglum Complex (see above).

The many intrusive phases were best interpreted using a greyscale 1VD (see algorithm DK-B). This showed many small intrusive phases with subtle alteration rims. (*This interpretation is based purely on geophysics, with no surface indications of intrusions. The area is covered by Triassic volcanics*).

An east-west profile, (line KT2), across the area was used to provide depth information about the complex. The broad intrusive response was modelled with two complex polygons at depth. These gave a very good match and could indicate actual intrusions at depth. Alternatively, they may simply express the regional from the combined intrusions.

Individual anomalies were modelled to determine the depths to the top of the magnetic aureoles surrounding each phase. The anomalies chosen were stronger and provided a means for modelling with some confidence (see model KT-A, **Figure 19a**). In order to gain some information about the more subtle anomalies, Automag was run with sensitive settings (see model KT-B, **Figure 19b**). Together, they indicated that the eastern side was subcropping. It deepened to about 200m in the centre and continued at about 200m depth to the western margin.

Timing relationships were established together with Diglum (see text on Diglum above and algorithm DK-D).

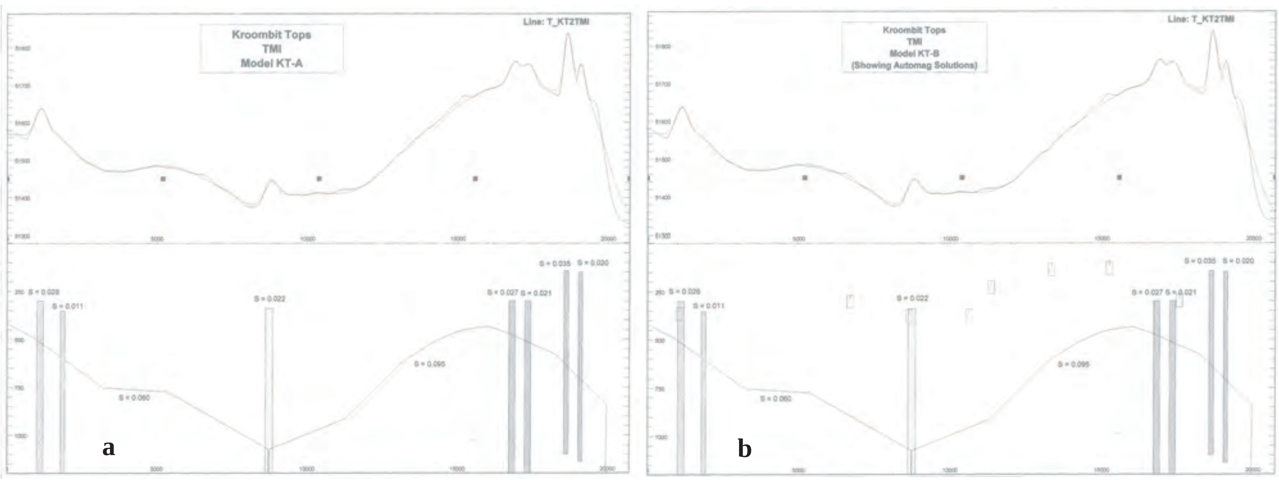


Figure 19. Model across the Kroombit Tops region

Intrusion 9 appears to be the oldest phase in the Kroombit Tops complex as it is intersected by most other intrusions. Body 5b is intersected by 13 and body 17 intersects both 5 and 7, indicating that most of the Kroombit Tops intruded after at least part of Diglum. No timing relationship could be devised between bodies 13 and 17 and Diglum bodies 2, 3, 4, 6, 7 or 8 and no relationship could be established between any Diglum phase and bodies 9, 10, 11, 12 or 14.

Bodies 9a and 9b were interpreted to be the same intrusion as their character was similar. They were separated by a fault boundary.

Bodies 10 and 11 clearly intrude 9a and body 12a clearly intrudes 9a, 10 and 11. Body 12a was thought to be the same as 12b based on their similar character but 13 was interpreted to be different as it was more magnetic. 13 was interpreted to intrude 12b.

Body 26 clearly intruded both 12a and 13. It had a strong magnetic aureole. Its western boundary was faulted, probably after emplacement as it did not have an magnetic aureole.

Bodies 14a, 14b and 14c were interpreted to be the same based on their character. They may even be the same as Body 9 although they had slightly stronger magnetic relief.

Body 15 intruded 9a but had a fault boundary with 14a which was inconclusive. It was intruded by 16 which also intruded 14a and 14b.

Body 17 intruded bodies 12, 13, 14 and 16. It was shallow for much of its area. Part of the area looked deeper as it was flown higher due to topography. The body could be traced through this transition and was interpreted to be all one body despite some gradual decline in susceptibility from east to west.

Body 17 was intruded by 18 which had fault boundaries with 14.

Body 19 was interpreted to intrude 14 and 16 with the evidence being stronger on 16. It was a larger intrusion and had some magnetic aureole developed.

Body 20 was a shallow intense magnetic body which intruded 19. It may prove to be similar to Diglum north-west and Diglum North (see discussion above).

Body 21 was interpreted to intrude 20 and 22 intruded both 20 and 21. 23 intruded 22 and 16 as well as 9a.

Body 24 intruded 9a and was intruded by 25 but no relationship could be established with other bodies.

Glassford (GF)

The image of airborne. magnetic data over the Glassford Igneous Complex is shown in Figure 20.

Coordinate at approximate centre: GR323000 E, 7290000 N

The Glassford Complex was a series of intrusions with one having a very prominent alteration aureole.

Modelling was first conducted on two sides of the prominent alteration aureole and a good match was achieved with vertical bodies at about 250m depth (Figure 21a,b). A more complex model was then conducted across a larger section of the complex. This was modelled with vertical dips to give indications of depth (Figure 21c). Depths varied from 250m on the north-west rim to 500m on the south-eastern rim. Depths increased to around 700m to the north-west which was consistent with the interpreted deeper intrusion in that area. (*The strongly magnetic rim is attributed to the Rule Gabbro, which crops out to the south-east and in one small exposure to the north. The difference between the magnetic interpretation and the mapped geology suggests that magnetic susceptibilities may be greater in the north*).

While most of the models had susceptibilities consistent with those measured in the field, some were higher by a factor of 2. This suggested that the models had greater depths than in reality. If the real rims did not have sharp contacts, the model would need to be deeper to fit the measured curve. They would therefore need to have greater susceptibilities to give the same response. This was supported by the fact that no model outcropped but field mapping detected highly susceptible rocks.

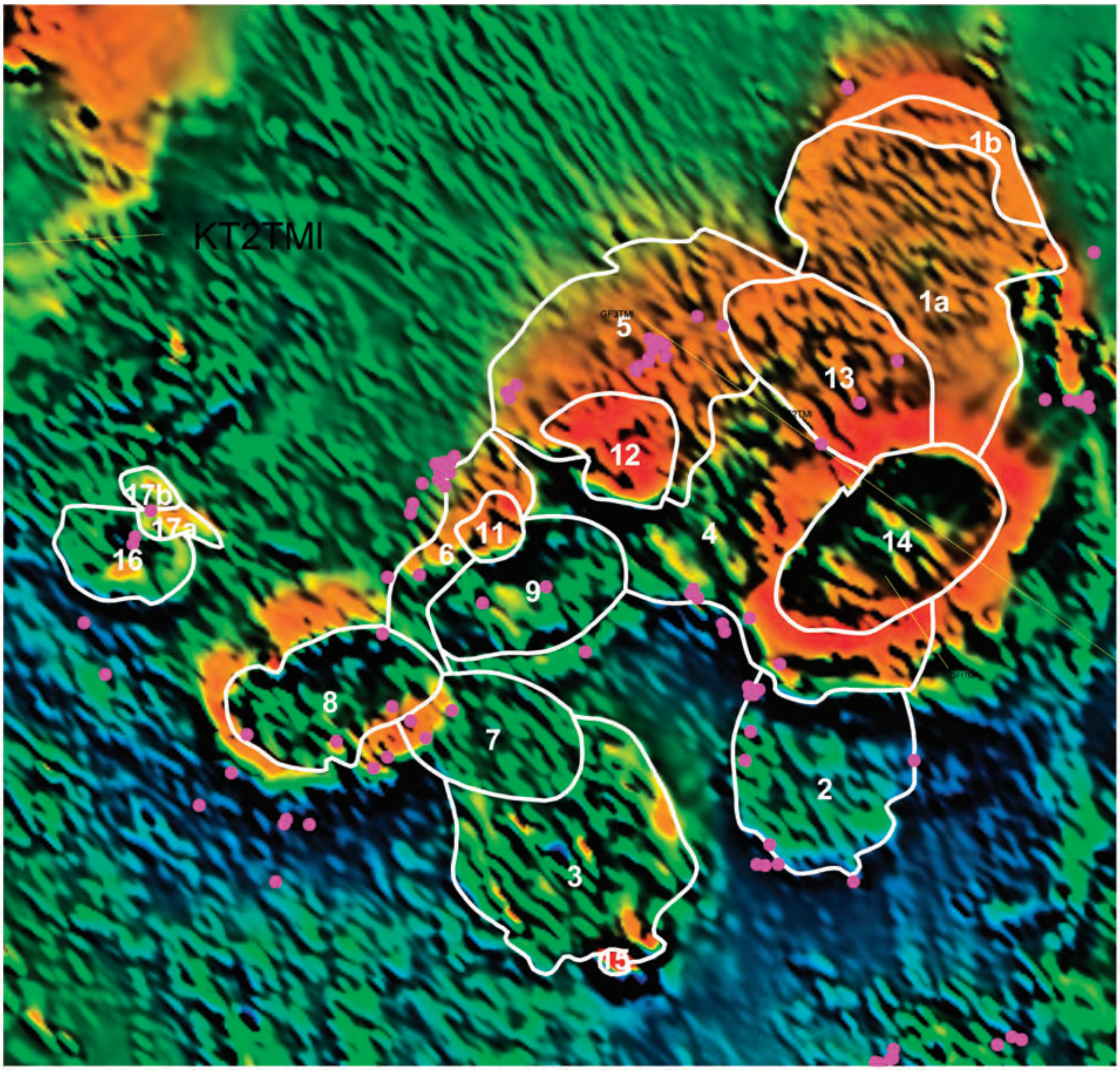


Figure 20. Image of airborne magnetic data over the Glassford Igneous Complex

Automag solutions matched the model fairly well when constrained with vertical dips (Figure 21d). When the correlation between the model and data is more constrained but with dips allowed to be free, two Automag solutions were slightly deeper than the forward model and had steep dips to the south-east (Figure 21e).

Timing relationships were inferred between the different units. The oldest units were either 1, 2 or 3. No relation could be established between these three. 1b was interpreted to be a different phase of 1a with a fault boundary bringing them adjacent. Body 4 was younger than 2 but no relationship could be established with 1 and 3. (*Bodies 1 and 2 crop out and equate broadly with the Littlemore Granodiorite and the Lawyer Granite respectively. The Littlemore Granodiorite also comprises Body 13 and part of Body 4. Body 4 also includes the Rule Gabbro, part of the Lawyer Granite, the Tollbar Breccia, and country rocks of the Rockhampton Group*).

Body 5 was interpreted to be different to 4 based on their different susceptibilities but the boundary was difficult to interpret in terms of establishing which intruded the other. 5 was interpreted more likely to intrude 4 based on the western margin. Body 6 intruded 4 but did not appear to intersect 5. It was not clear whether 7 intruded 6 but both were intruded by 8. (*Bodies 5, 6 and 7 largely comprise country rocks, some of which may be aureoles. Body 8 is the Monal Granodiorite which is surrounded by a strongly magnetic contact aureole in the Three Moon Conglomerate*). Body 3 was intruded by both 7 and 15. (*Body 3 does not crop out, but may be concealed beneath a cover of Triassic Dooloo Tops Volcanics. Body 15 is a Tertiary basalt plug at Pine Mountain*).

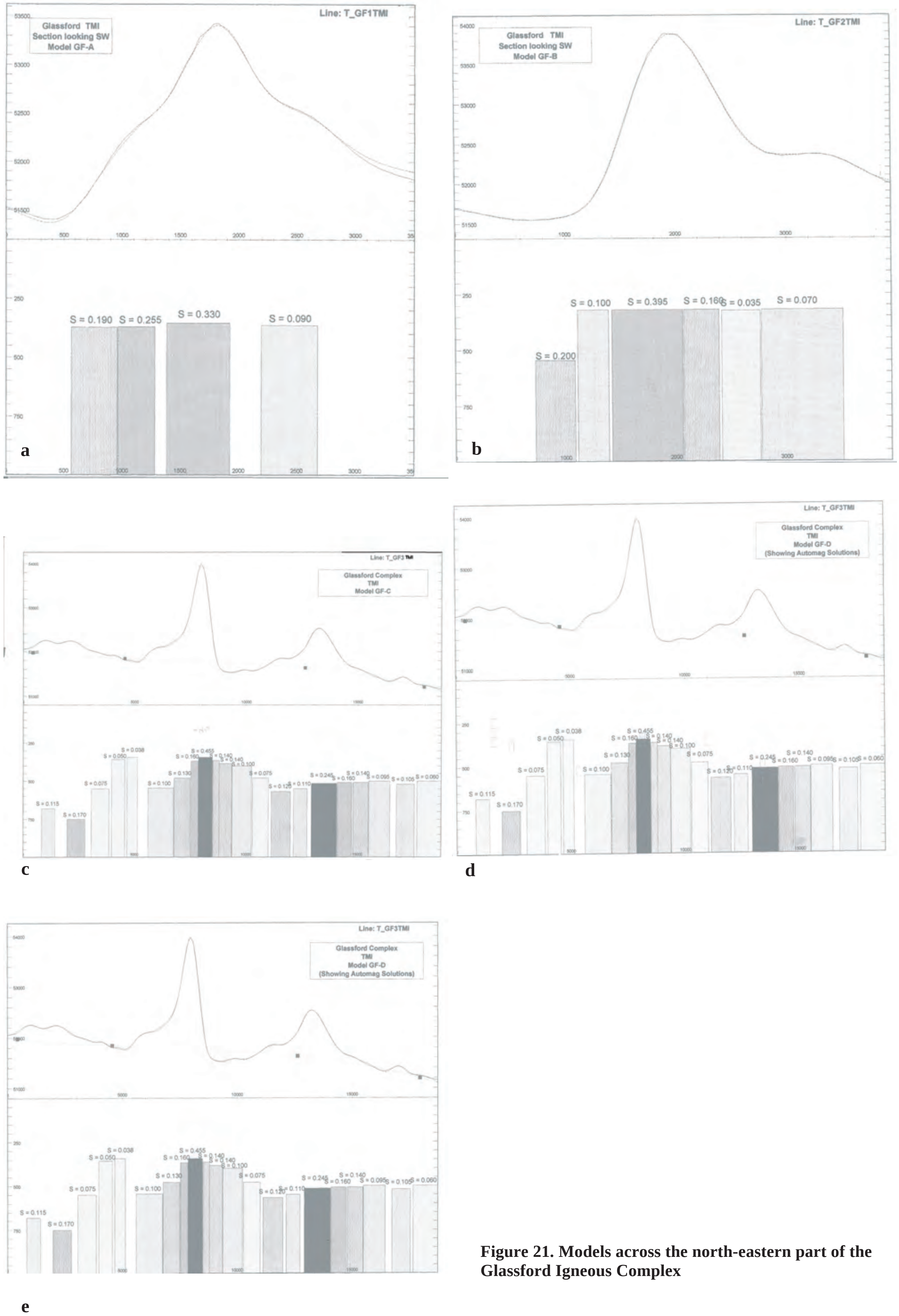


Figure 21. Models across the north-eastern part of the Glassford Igneous Complex

Body 12 intruded 5 and was very magnetic. It may have been a gabbro, skarn or alteration effect (see discussion on Diglum North above). (*Body 12 is the Radley Nepheline Syenite and part of the Ridler Monzonite*).

Body 13 intruded 4, 5 and 1a and was intruded by 14. Body 14 had a very intense alteration aureole with amplitude similar to body 12. (*Body 13 is mapped as the Littlemore Granodiorite. Body 14 is the Robert Granite, and is surrounded by a pronounced magnetic high due to the Rule Gabbro, which is exposed mainly to the south of the granite*).

Intrusions 16 and 17 were separate to the others and no timing relationship could be established. 17a and 17b were separated across a fault and had slightly different magnitudes but were thought to be part of the same event. They may have been similar to Diglum North in origin (see above). (*Body 16 equates to the Munholme Quartz Diorite, which is surrounded by a relatively wide magnetic contact aureole*).

Rawbelle Batholith (RB)

Coordinate at approximate centre: GR285000 E, 7255000 N

The Rawbelle Batholith extended north into the southern part of the area. It had a significant gravity low and an obvious magnetic rim. It did not have much internal magnetic variation.

DYKES

Numerous dykes cut the area, particularly in the south-east (see algorithm Dykes). Where the dykes intersected sedimentary rocks or other units with little magnetic relief, they were easy to interpret. Few dykes had a magnetic relief of greater than 50nT and often they were less than 20nT. Where they intersected units with complex and high magnetic relief, it was often difficult to detect their subtle magnetic signatures. Over magnetic igneous bodies, the magnetic relief was random and aliased between flight lines. When this was gridded, an artificial fabric was generated perpendicular to the flight line direction where anomalies were stretched to fit between flight lines. This fabric was sub-parallel to the dykes and made them difficult to be reliably detected through these complexes.

To the south-east, there were numerous dykes striking east-south-east. This strike was only occasionally present in the Yarrol area. By far the prevalent strike in the Yarrol area was north-west to north-north-west. (*The east-south-east trending dykes are only present east of the Yarrol Fault*).

Immediately north-north-west of the Glassford Complex, the dykes were the most numerous and fanned out slightly from the centre of the Glassford Complex. Not all the dykes did this suggesting there were at least two episodes of dykes with one possibly being related to a phase of the Glassford Complex.

Some dykes were interpreted intersecting most intrusives. This indicated that some of the dykes were a late phase. In other areas, the dykes did not appear to carry into the intrusives, e.g. Diglum and Kroombit Tops had few dykes despite the large number to their south-east. This could be interpreted that many of the dykes postdate the intrusive but may also indicate that the intrusives were not conducive to development of dykes or that the intrusives had susceptibilities the same or greater than those of the dykes.

One constraint on their age was the observation that the dykes were particularly absent from the near surface in the larger sedimentary basins such as the Casuarina Basin.

BASINS

Casuarina Basin (CB)

Coordinate at approximate centre: GR267000 E, 7391000 N

The largest basin in the area appeared to have significant faulting both at the margins and within the basin. Its basement had several steps to a maximum depth of almost 2000m.

Forward modelling successfully fitted bodies down to 1600m (Figure 22a). Naudy modelling suggested a slightly shallower basement depth of 1520m across the centre of the basin (Figure 22b) and shallower depths elsewhere(Figure 22c-g). Fourier analysis had high confidence layers at 1580m and 2000m (**Figure 23**).

At the very edges of the basin, the western side appears in the magnetic data to have a sharp contact with basement rocks via a small fault while the eastern margin appeared to gently dip under cover for several kilometres. This was consistent with previous interpretations of Tertiary half grabens in the region that incorporated a faulted western margin.

Interpretation of the magnetics as a whole did not support this as the basin appeared to be deeper to the east, stepping down in a series of fault blocks from a shallower western margin. The basement on the western margin had a sharp but small fault giving the sharp contact with sediments. The basement on the eastern margin dipped gently under cover for a few kilometres before being faulted down by a much larger fault, giving a deeper basement along the western side. There did not appear to be as much variation along the north-north-west line.

Jim Crow Basin (JC)

Coordinate at approximate centre: GR258000 E, 7430000 N

Visually, the Jim Crow Basin had numerous shallow bodies making it difficult to determine an accurate depth to basement. Forward modelling would be difficult because the shallow bodies would dominate. For this reason, only FFT analysis was conducted. This showed magnetic layers at about 1700m and 350m as well as the usual near surface noise (**Figure 24**).

Nagoorin Basin (NB)

Coordinate at approximate centre: GR330000 E, 7303000 N

The Nagoorin Basin had a long north-south strike but was only narrow. It also had numerous intrusions and near surface features making it difficult to determine depths. One area near the north margin was relatively clear of near surface noise so it was submitted to FFT analysis (**Figure 25**). All the layers were low confidence but a short segment did indicate a depth of between 600m and 650m.

Parkhurst Basin (PB)

Coordinate at approximate centre: GR241000 E, 7423000 N

FFT analysis over the Parkhurst Basin gave a very high confidence layer at 35m below surface and a moderate confidence layer at 1186m (**Figure 26**). The moderate confidence layer correlated well with the forward modelling which had a good fit from a body at 1000m below surface (**Figure 27a**). The forward modelling indicated basement at about 300m depth on the western side. The basement was faulted down by about 300m on the western side and then deepened to 1000m in several steps before rising gradually to the east. The line did not continue onto basement in the east because the anomalies at the surface were so strong that they would swamp the weak signals from the basin. These strong signals were approximated by a body with negative susceptibility coinciding with the strong low south-west of the basin’s north-east margin.

Automag solutions suggested bodies approximately 200-300m shallower (**Figure27b**).

(*The Parkhurst “Basin” is actually a belt of non-magnetic Rockhampton Group flanked to west and east by younger, more strongly magnetic units. The nature of the basement to the Rockhampton Group is unknown*

Rossmoya Basin (RB)

Coordinate at approximate centre: GR237000 E, 7448000 N

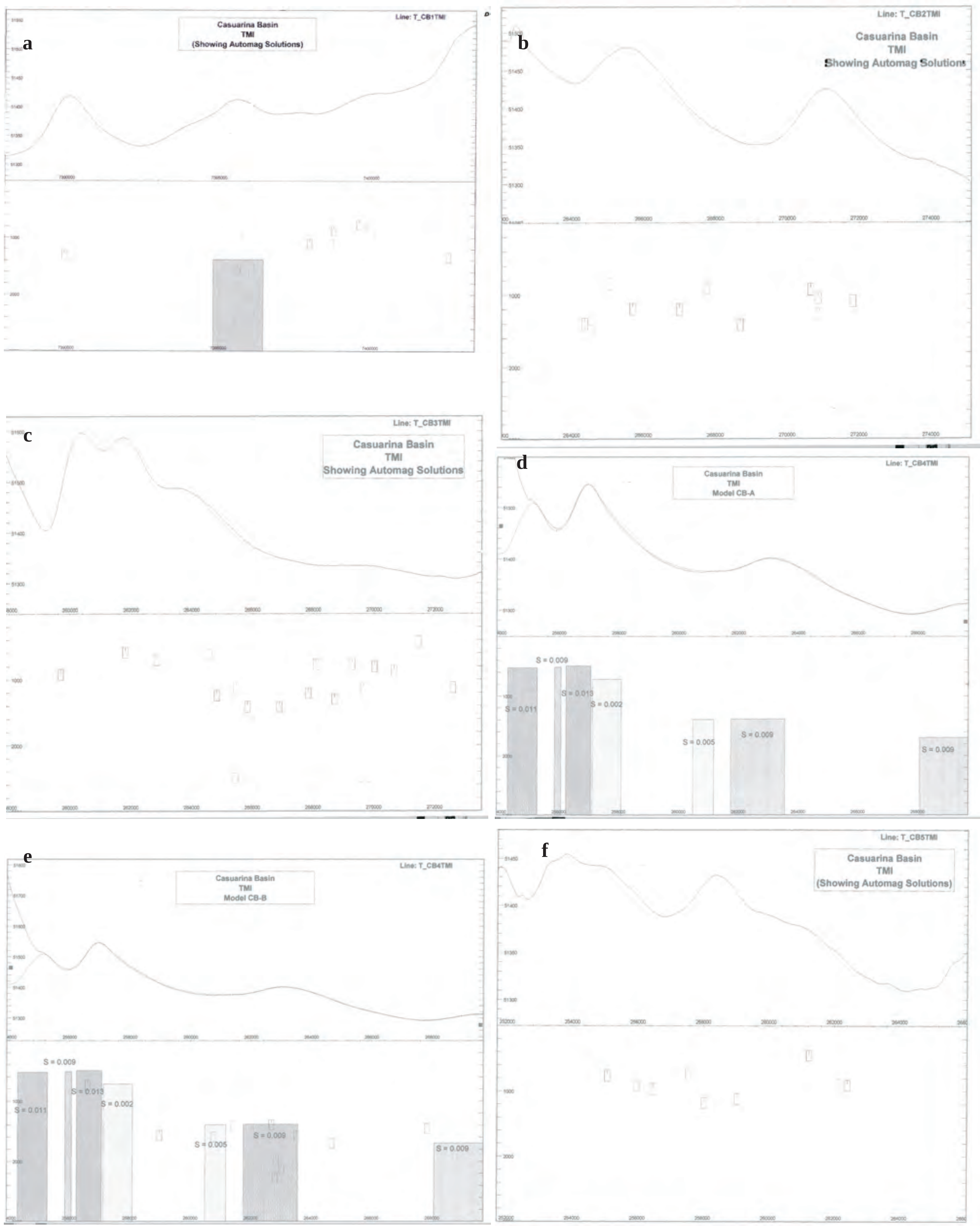


Figure 22a-e. Models across the Casuarina Basin

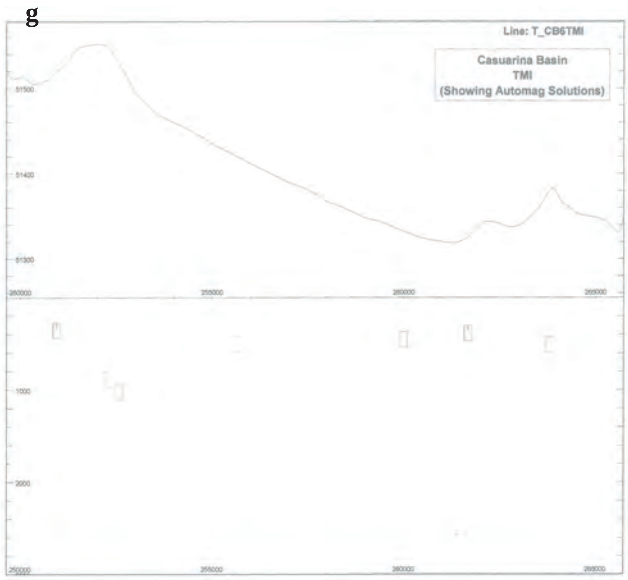


Figure 22g. Models across the Casuarina Basin

Forward modelling of the Rossmoya Basin suggested sediment thicknesses of 300–400m (**Figure 28a**). The basin appeared uniform in a north-south direction but was slightly deeper to the east. Automag solutions correlated well in the western half but suggested bodies up to 250m deeper in the eastern half (**Figures 28b,c**).

FFT analysis indicated moderate confidence layers at 91m and 2183m (**Figure 29**). The deeper signature was probably from the broad basin response rather than individual anomalies within the basin.

Stanwell Basin (SB)

Coordinate at approximate centre: GR225000 E, 7398000 N

Forward modelling of the Stanwell Basin suggested basement depths of not more than 300m (**Figure 30**). The model had a broad second order regional applied. Basement dips were consistent with an anticline in the centre of the basin with vertical dips on each end of the line.

Yaamba Basin (YB)

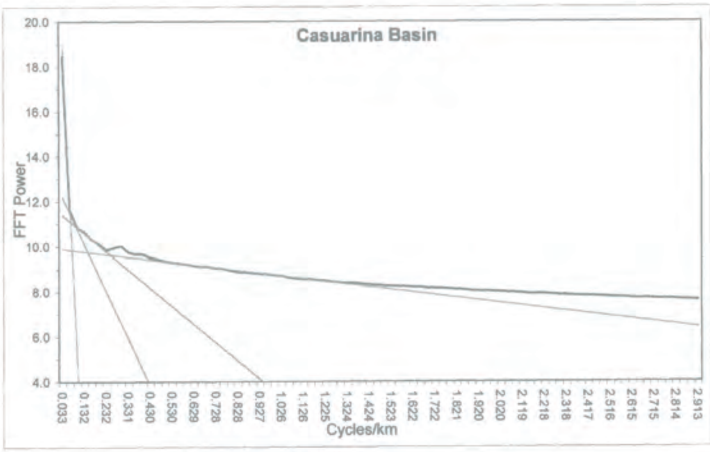
Coordinate at approximate centre: GR228 000 E, 7 440 000 N

Forward modelling of the Yaamba Basin suggested depths of up to 1300m (**Figure 31a**). When Automag was run, it also generated a 1300m body but had shallower sources at between 400 and 700m (**Figure 31b**). The shallow sources could have been the basement, noise or possibly weak magnetic layers within the sediments eg thin basalts. If they were the base of sediments, then the basement would prove to be relatively non-magnetic.

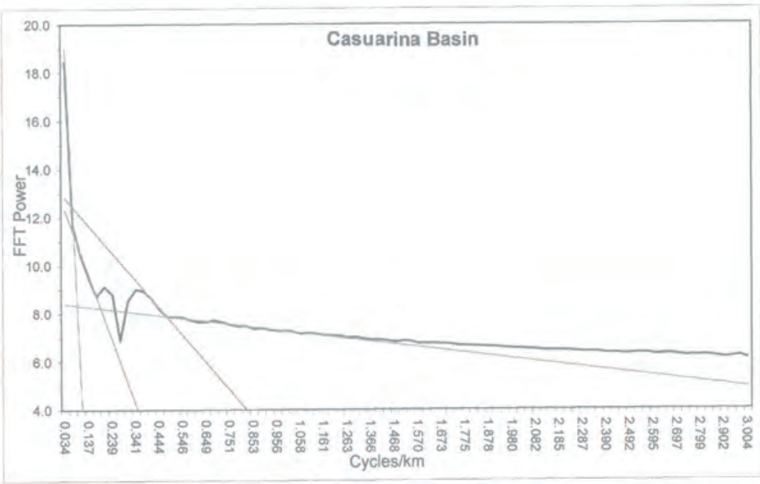
FFT analysis indicated layers at 145m and 765m (**Figure 32**). The latter would correspond to the shallower layers in Automag but the deep layer fitted so well in the forward model it would be difficult to discount it. The more likely scenario was for sediments to 1300m.

PROSPECTIVE AREAS

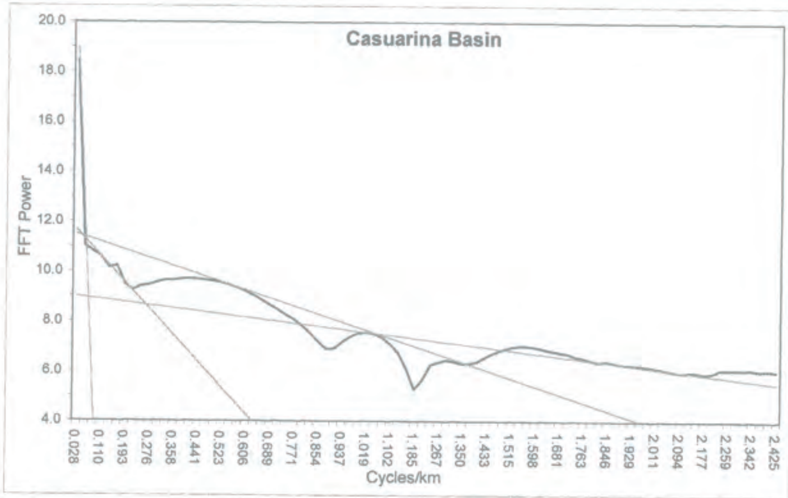
Two large deposits in the area, Mount Morgan and Mount Chalmers, were not immediately obvious as geophysical anomalies. However, they both lay on significant gravity gradient linears interpreted as large faults. Mount Morgan lay close to the intersection of two of these while Mount Chalmers lay close to the intersection of a gravity gradient and a significant fault interpreted from the magnetics.



Aeromagnetic Line No:	**112140**				
Line Coordinates:	**251386E, 7396976N to 262035E, 7406684**				
Layer:	1	2	3	4	
Depth below Sensor:	16825	1683	649	96	metres
Average Layer Depth below Ground:	**16745**	**1602**	**569**	**16**	**metres**



Aeromagnetic Line No:	**112481**				
Line Coordinates:	**260005E, 7385385N to 270357E, 7395746**				
Layer:	1	2	3	4	
Depth below Sensor:	16317	2098	886	93	metres
Average Layer Depth below Ground:	**16245**	**2026**	**813**	**21**	**metres**



Aeromagnetic Line No:	**112580**				
Line Coordinates:	**275823E, 7395516N to 263003E, 7382671**				
Layer:	1	2	3	4	
Depth below Sensor:	20218	1011	309	116	metres
Average Layer Depth below Ground:	**20141**	**934**	**232**	**38**	**metres**

Figure 23. FTT analysis of magnetic profiles across the Casuarina Basin

The area is prospective for a number of different types and styles of deposits. Discussion below focusses on broad geophysical categories and does not try to predict specific types.

Only a few representative sites of each style have been nominated (see overlay “PROSPECT.erv”), and the list is by no means exhaustive. Focus was also on areas that do not currently have many mineral occurrences in order to promote exploration in non-traditional areas. There was no follow-up to check whether the sites nominated were actually correct for their nominated style. They were simply nominated as interesting sites for further work.

Sedimentary basins

Sedimentary basins were evident mainly from the magnetics but several had significant gravity lows.

The basins have a significant development of sediments and would be prospective for coal. Two different deposit styles could be searched for. The first involves a conventional seam that had a relatively uniform thickness over a large area. This style would probably have developed thicker in a thicker sequence of sediments such as on the eastern side of the Casuarina Basin. This assumed that both sides of the basin developed simultaneously. This may not have happened. The deeper part of the basin may have developed first with nothing on the shallow side. Then the wider area could have subsided allowing a second phase of sedimentation to develop across the whole area.

The former scenario would have a gradual thickening of each unit towards the deeper part of the basin while the second scenario would have relatively uniform units developed across the whole basin. Shallow drilling could give evidence for favouring one scenario over the other.

If sedimentation was too quick, the first style would have become diluted with sediments and not developed to an economic deposit. In this case a second style could develop as a local deposit where a relatively stable backwater fostered accumulation of coal. In the latter case, the more prospective areas would be the shelf areas and edges of the basins.

Only the more prospective zones of the Casuarina and Yaamba were marked on the overlay (see overlay “PROSPECT.erv”).

Conducting detailed magnetics and gravity would be the best ways to narrow the exploration areas for coal, particularly the local deposit style.

Alteration zones

Alteration zones are well known sites for mineralisation. There are two types that can be recognised easily from the geophysics. Magnetic alteration is evident as intense magnetic highs and is probably also a local gravity high. The other style involves oxidising solutions and is characteristic of magnetite destruction. This should not be confused with magnetic lows that would be a type of the former with remanence superimposed. Rather it is an absence of magnetic response tending towards a background value. The latter may also have a local gravity high but not necessarily so.

In the case of the Yarrol area, the magnetic dataset is adequate to detect magnetic alteration of both sorts but the gravity dataset is not sampled dense enough to be useful on this detailed scale.

Several magnetic alteration anomalies were identified (see algorithm PROSPECT). In addition to those marked, the magnetic alteration rims of many of the intrusions could also be prospective.

Skarns

Skarns develop adjacent to intrusions in calcareous units. Although it was difficult to determine whether the units were calcareous or not, the irregular and intense magnetic character of skarns allowed some sites to be identified as possible skarn sites (see algorithm PROSPECT).

Structural zones

Structural zones allow expansion and pressure release for fluids and are therefore good sites for mineralisation. Often these are recognised in conjunction with one of the above styles. Special mention of some structural sites was made to highlight some of the more prospective sites (see algorithm PROSPECT).

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APPENDIX 5A: MAGNETIC MODELS

Magnetic models were conducted using Encom's Model Vision program. Most models used data extracted from the gridded images.

Thirty-two 2D models were completed at 21 sites. At some sites, several models were generated to explore different scenarios. An additional 14 models had Automag solutions applied and overlain on the models. Although the model with the Automag solution was often the same as the forward model, it was given a different name so it could be stored in a different file. The models were saved as session files that contained both the data and the model as well as format information. These session files were prefixed with "m_" and had ".ses" extensions.

Two 3D models were completed where it was difficult to achieve a 2D result. This happened where the magnetic bodies had short strike lengths and different strikes.

Models were conducted at the following sites:

3D Models

Site	Model Name	Line Names	Start Coords	End Coords
Galloway Plains	GP-C	GP3TMI GP5TMI GP6TMI GP7TMI	270481E, 7330939N 270418E, 7330710N 270251E, 7331113N 270235E, 7331413N	277906E, 7330939N 278003E, 7330710N 277594E, 7331446N 277494E, 7331930N
Diglum	DL-A	DL-3TMI DL-4TMI DL-5TMI	308600E, 7319919N 305455E, 7317657N 306090E, 7318464N	308651E, 7314675N 310869E, 7317400N 311061E, 7318341N

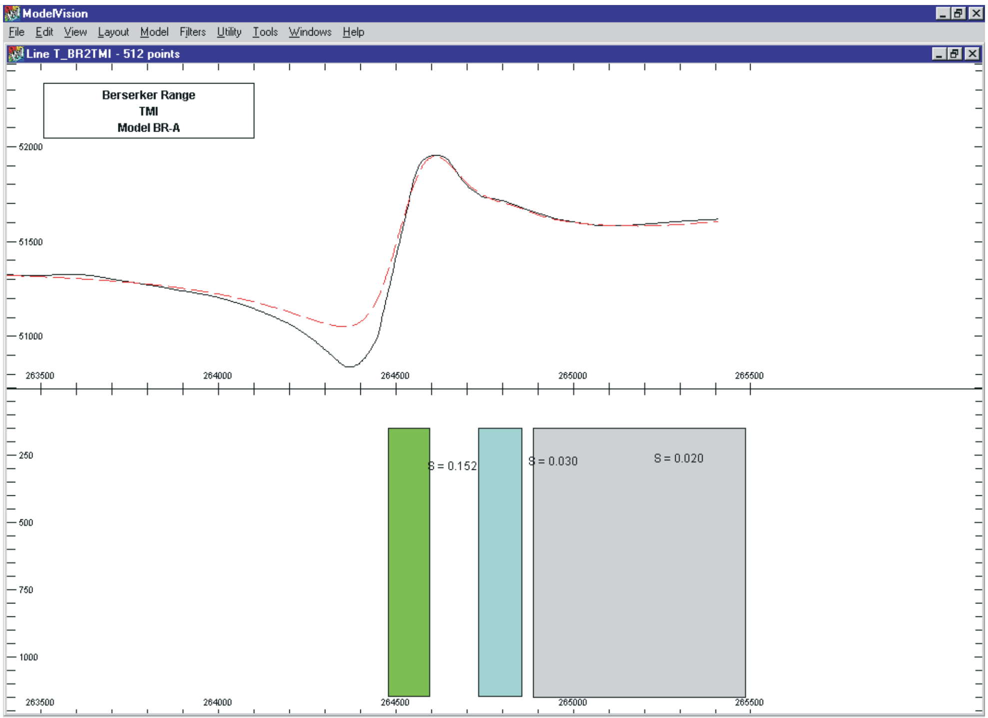
2D Models

Site	Model Name	Line Name	Start Coords	End Coords
Berserker Range	BR-A	BR2TMI	263403E, 7417471N	265408E, 7419158N
	BR-B	BR2TMI	"	"
	BR-C	BR2TMI	"	"
	BR-D	BR2TMI	"	"
Calliope-North	CY-A	CY2TMI	311804E, 7346104N	318858E, 7349693N
Calliope South Site 1	CN-A	CN1TMI	321328E, 7324072N	328132E, 7326071N
Calliope South Site 2	CN-B	CN3TMI	319803E, 7331252N	326353E, 7334377N
Casuarina Basin	CB-A	CB4TMI	253908E, 7387471N	269686E, 7399879N
	CB-B*	CB1TMI* CB2TMI* CB3TMI* CB4TMI* CB5TMI* CB6TMI*	250868E, 7403413N 262044E, 7383280N 257853E, 7385745N 253908E, 7387471N 251690E, 7390676N 249635E, 7398318N	277328E, 7385745N 275110E, 7393470N 273548E, 7395442N 269686E, 7399879N 266070E, 7402016N 265742E, 7403906N
Eulogie Park	EP-A	EP1TMI	235956E, 7363663N	241444E, 7357242N
	EP-B	EP1TMI	"	"
Galloway Plains	GP-A	GP3TMI	270481E, 7330939N	277906E, 7330939N
	GP-B*	GP3TMI*	"	"
Craiglands	CIC-A	CIC1TMI	243367E, 7334488N	260581E, 7327884N
	CIC-A*	CIC1TMI*	"	"
Glassford Site 1	GF-A	GF1TMI	325602E, 7285768N	327585E, 7282876N
Glassford Site 2	GF-B	GF2TMI	322916E, 7290603N	325870E, 7287979N
Glassford Site 3	GF-C	GF3TMI	317908E, 7293876N	333113E, 7282933N

2D Models (continued)

Site	Model Name	Line Name	Start Coords	End Coords
	GF-D*	GF3TMI*	"	"
Gracemere	GM-A	GM1TMI	234436E, 7409121N	235630E, 7412107N
	GM-B	GM1TMI	"	"
	GM-C	GM1TMI	"	"
	GM-D	GM1TMI	"	"
	GM-E	GM1TMI	"	"
	GM-F	GM1TMI	"	"
	GM-G	GM1TMI	"	"
Kroombit Tops	KT-A	KT2TMI	282355E, 7295136N	303031E, 7296473N
	KT-B*	KT2TMI*	"	"
Bajool	MG-A*	MG1TMI*	254111E, 7378468N	248502E, 7386307N
	MG-B	MG1TMI	"	"
	MG-C	MG3TMI	248817E, 7386814N	254440E, 7378941N
Parkhurst Basin	PB-A	PB1TMI	234660E, 7431437N	247049E, 7417406N
	PB-B*	PB1TMI*	"	"
Rookwood Site 1#	RW1-A	RW1TMI	181110E, 7424915N	188809E, 7424996N
Rookwood Site 2#	RW2-A	RW2TMI	185071E, 7411668N	190007E, 7417582N
Rookwood Site 3#	RW3-A	RW3TMI	188585E, 7409327N	193212E, 7415877N
Rossmoya Basin	RB-A	RB9TMI	234437E, 7446919N	238453E, 7450838N
	RB-B*	RB9TMI*	"	"
Stanwell Basin	SW-A	SW1TMI	220504E, 7397065N	227366E, 7398092N
Yaamba Basin	YB-A	YB9TMI	225539E, 7437988N	230261E, 7442839N
	YB-B*	YB9TMI*	"	"

* These models have Automag solutions superimposed.
\# Grid for Zone 56 although the points are in Zone 55.



APPENDIX 5B: FFT ANALYSES

Fourier (FFT) analysis was conducted on flight lines over several basins to determine basement depth. For a description of the method, see “FFT Analysis” in the main body of text.

FFT analyses

Site	Flight Line	Start Coords	End Coords
Casuarina Basin	112140 112481 112580	251385E, 395976N 260005E, 7385385N 275823E, 7395516N	262035E, 7406684N 270357E, 7395746N 263003E, 7382671N
Jim Crow Basin	111590	254000E, 7429711N	260113E, 7435783N
Nagoorin Basin	111590	328803E, 7300866N	334028E, 7306110N
Parkhurst Basin	111430	237611E, 7422381N	242758E, 7427455N
Rossmoya Basin	110940	234501E, 7446962N	239800E, 7452226N
Yaamba Basin	110940	225594E, 7438077N	230903E, 7443356N

APPENDIX 6
MEASURED SECTIONS

Measured sections had been made of many of the Devonian and Carboniferous formations of the Yarrol Province, however, they were thought to have been lost. Recently some were relocated by chance, but they were discovered too late to be integrated into the main report without major re-editing of a virtually completed document, therefore it was decided to include them at the end of the document as an Appendix.

The Legend for the figures is Figure 1. The measured sections are placed in the text from oldest to youngest. The first section is from the Type Section of the Raspberry Creek Formation (Figures 2 and 3). The second measured section is from the Marble Waterhole beds (Figure 4). Two sections were measured through parts of the Balaclava Formation (Figures 5 and 6). The Type Section of the Mount Alma Formation is displayed in Figures 7, 8 and 9. The last measured section is from the Rockhampton Group (Figure 10).

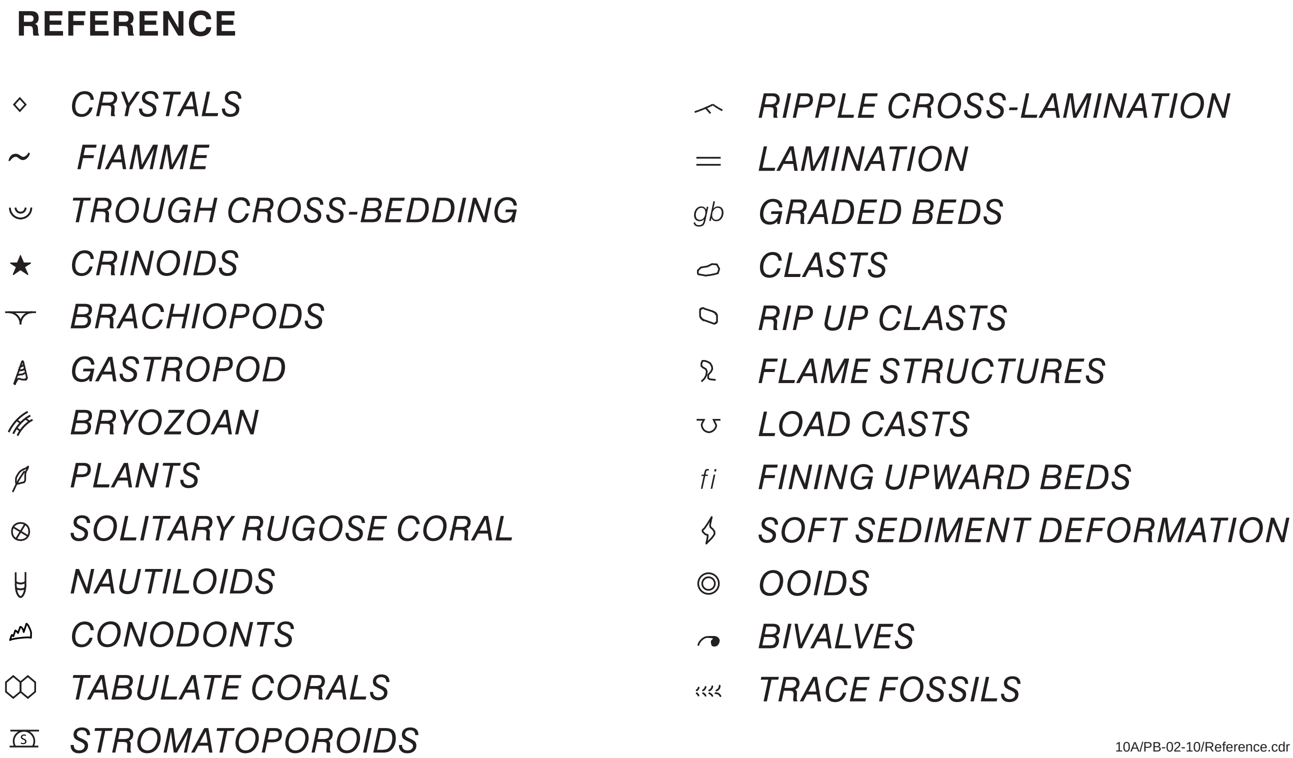


Figure 1. Legend for the symbols on the measured sections

RASPBERRY CREEK FORMATION - TYPE SECTION - MOUNT MORGAN 1:100,000 SHEET AREA - SHEET 1

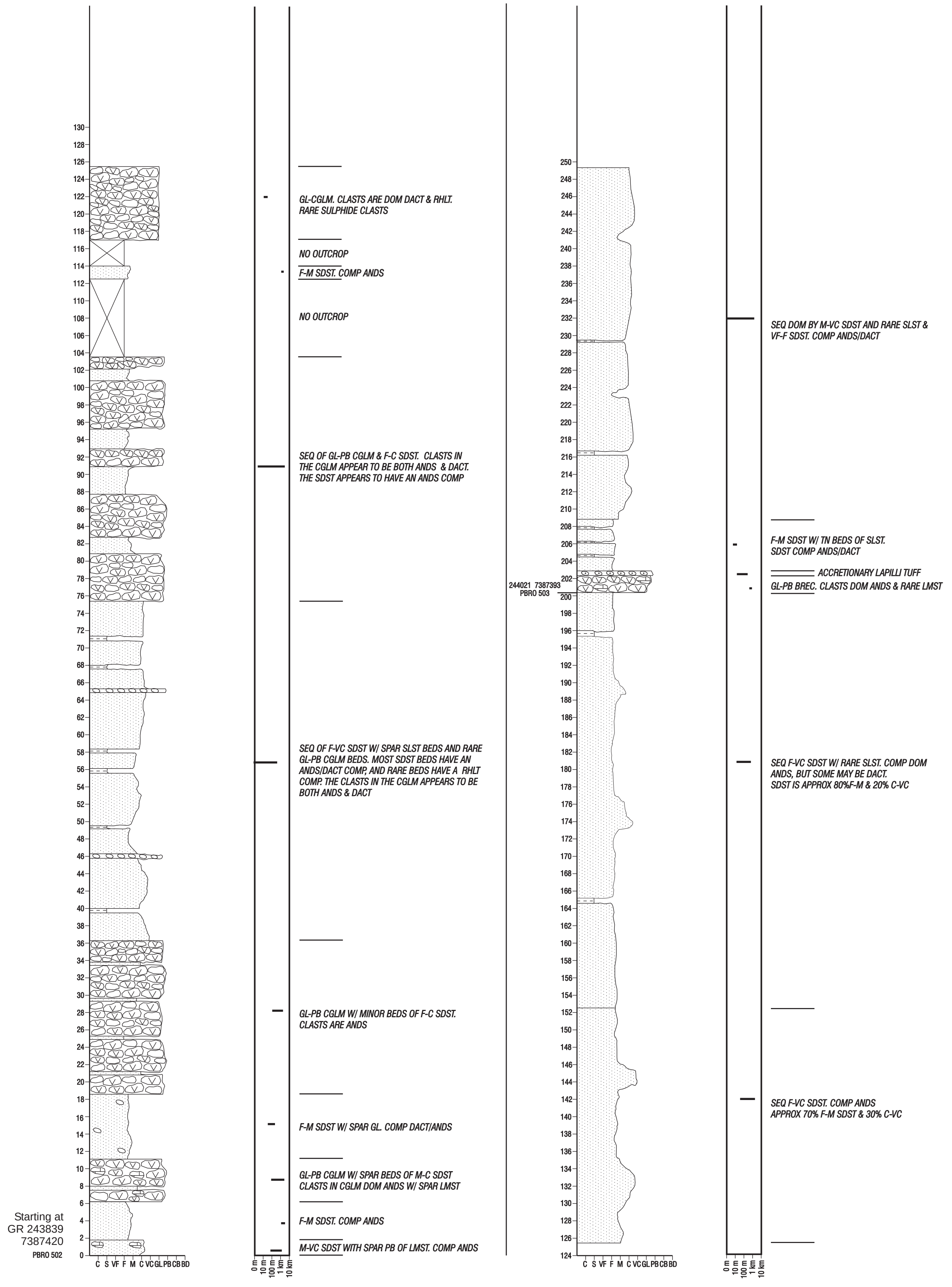


Figure 2. Type section of the Raspberry Creek Formation

RASPBERRY CREEK FORMATION - TYPE SECTION - MOUNT MORGAN 1:100,000 - SHEET 2

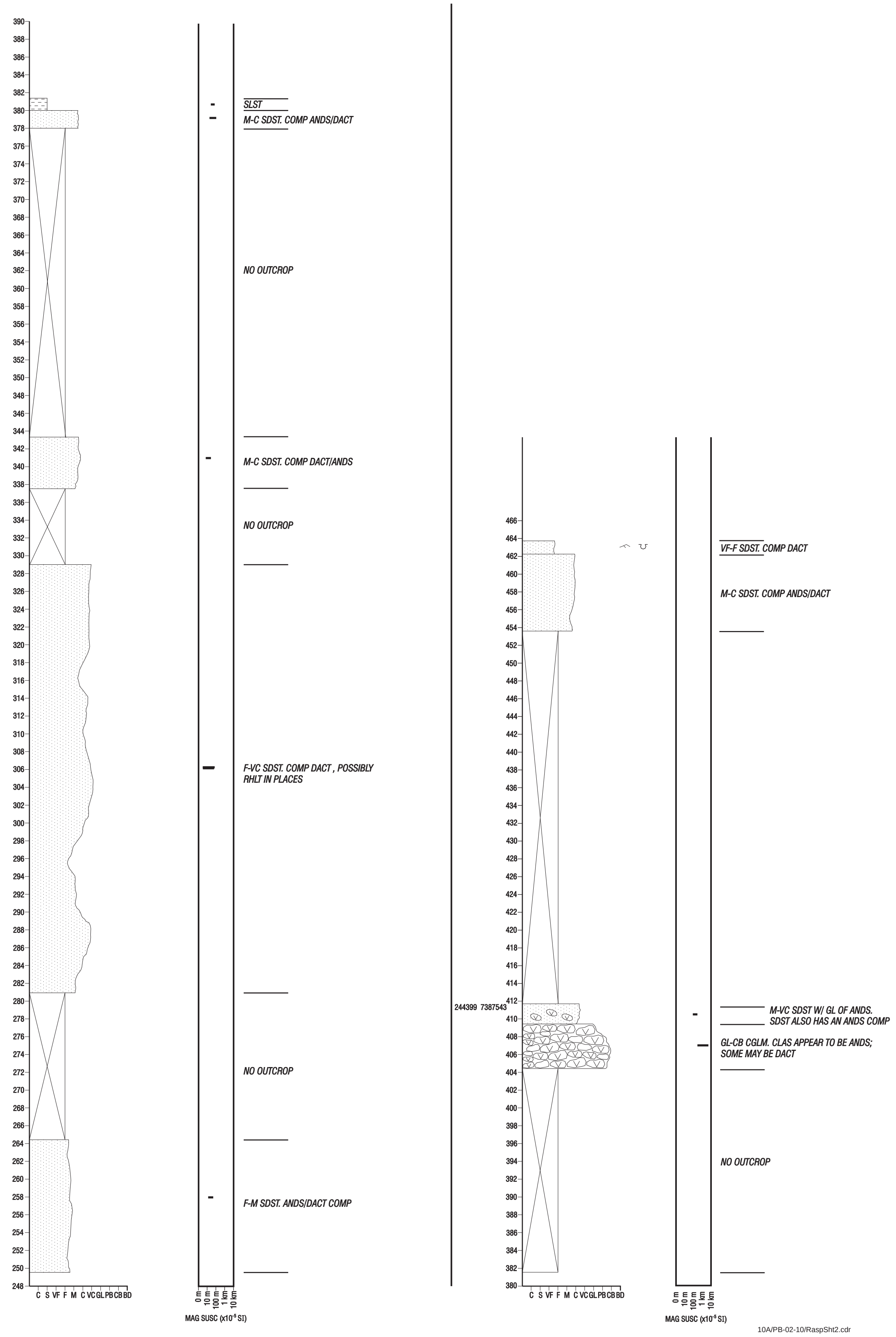


Figure 3. Type section fo the Raspberry Creek Formation (continued)

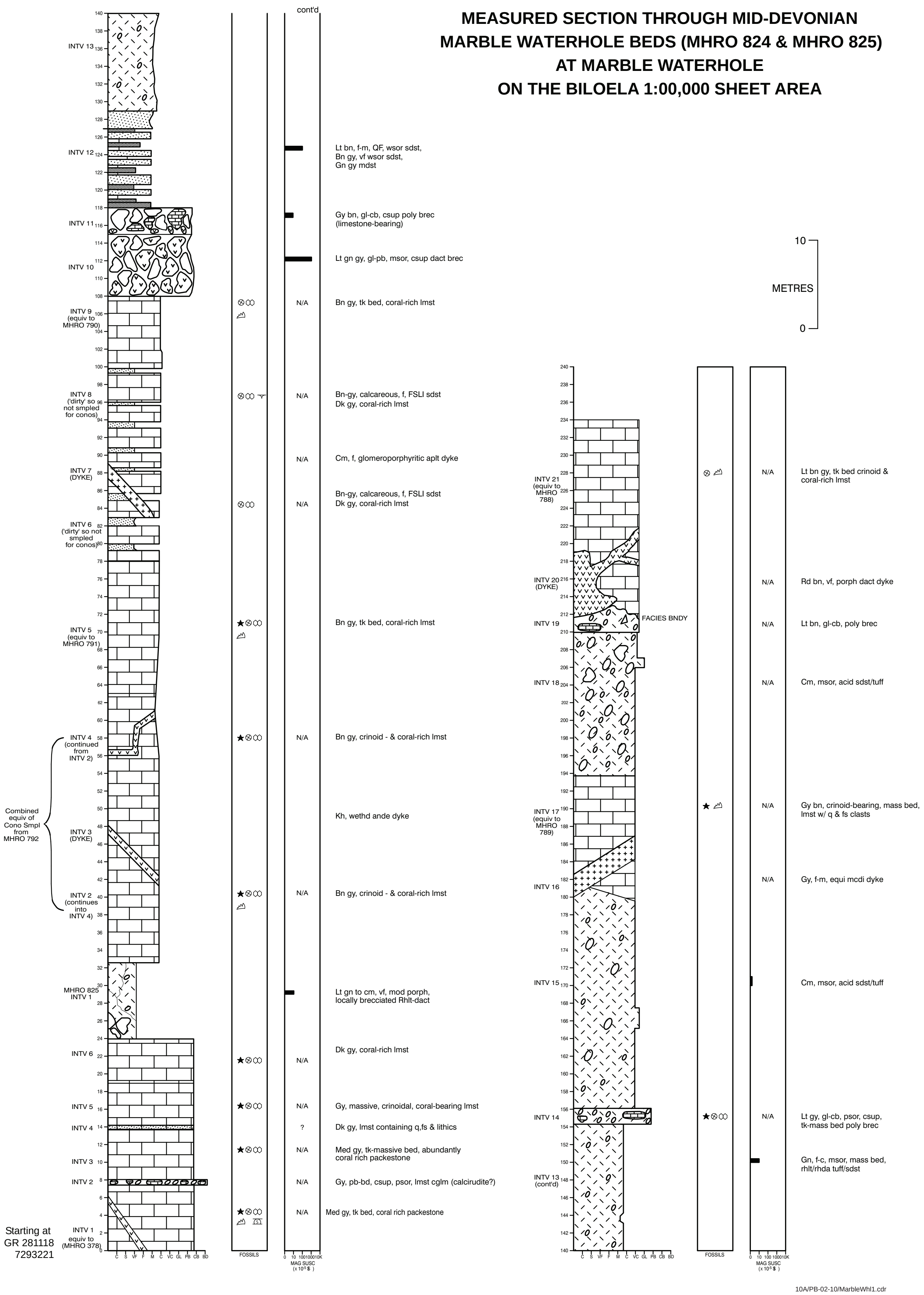


Figure 4. Measured section through the Marble Waterhole beds

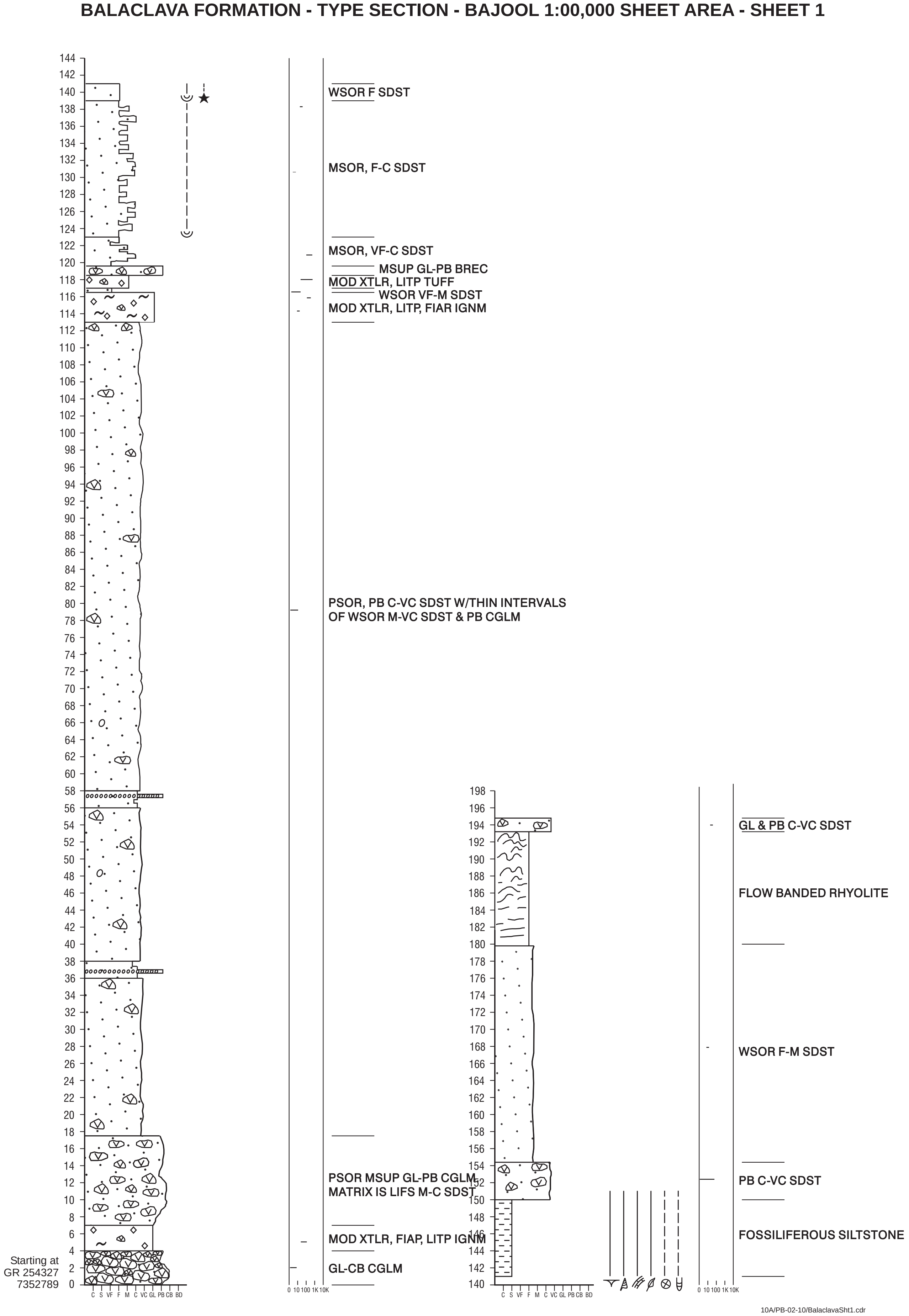


Figure 5. Type section for the Balaclava Formation

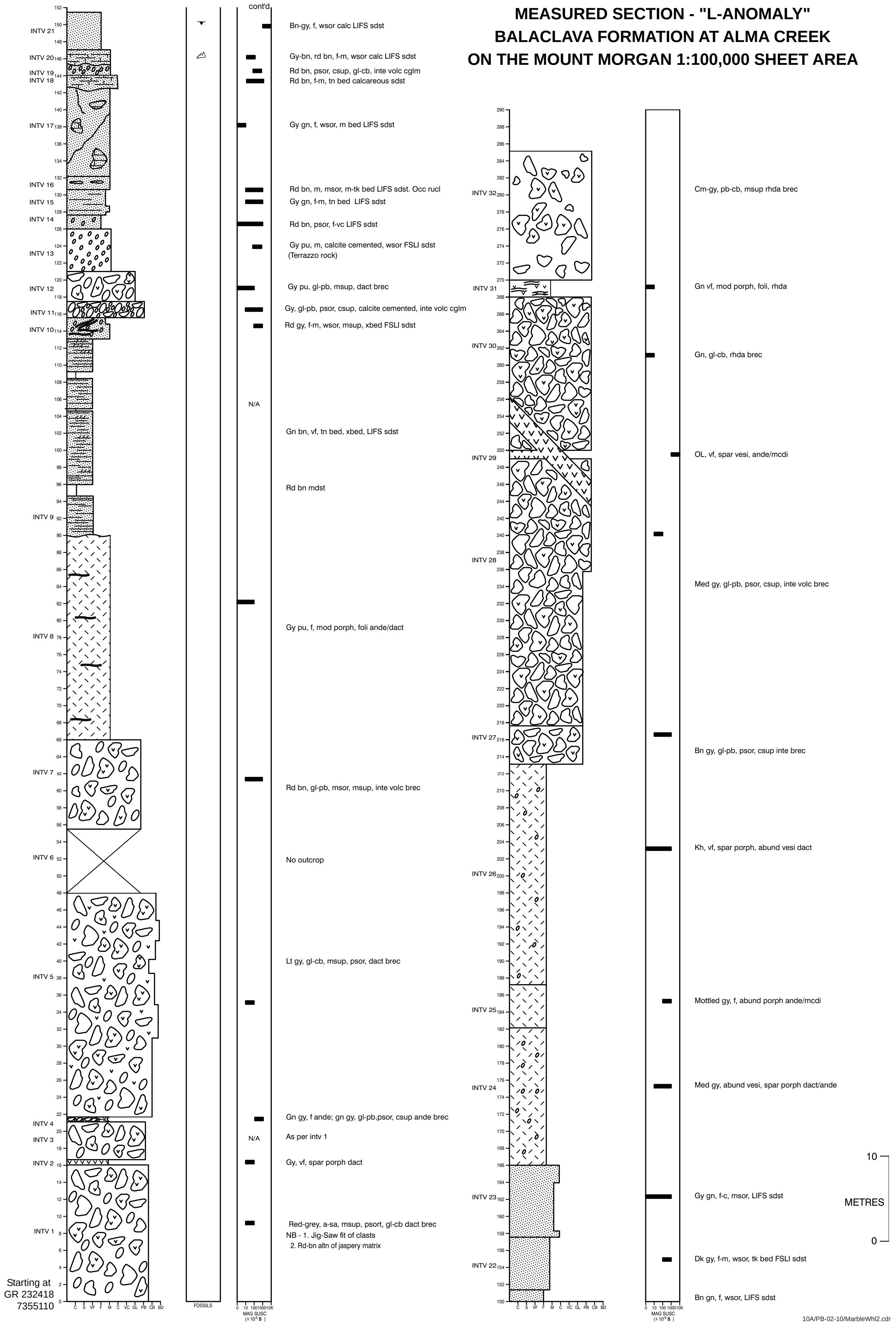


Figure 6. Measured section through the Balaclava Formation along Alma Creek

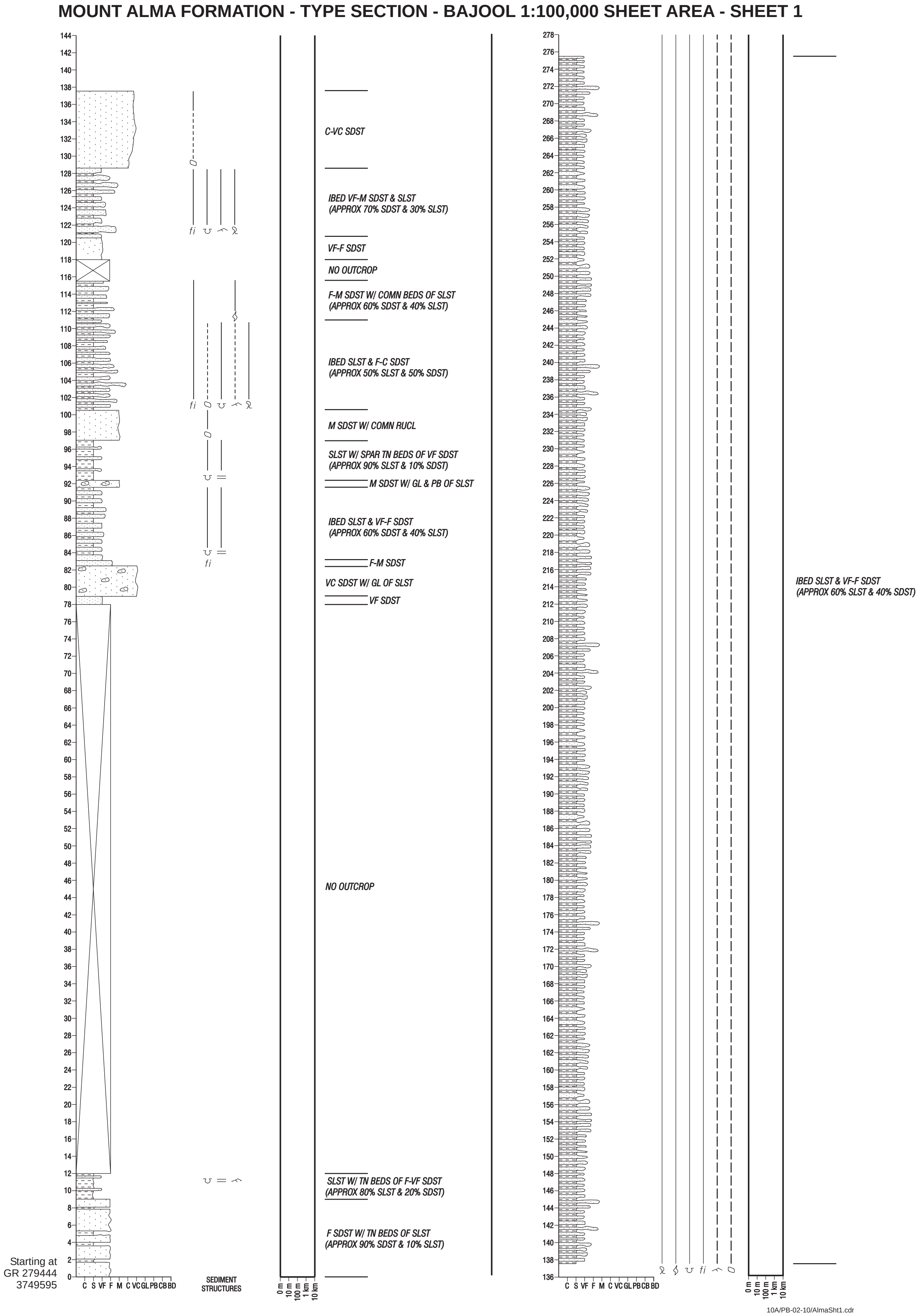


Figure 7. Type section for the Mount Alma Formation on Mount Alma

MOUNT ALMA FORMATION - TYPE SECTION - BAJOOL 1:100,000 SHEET AREA - SHEET 2

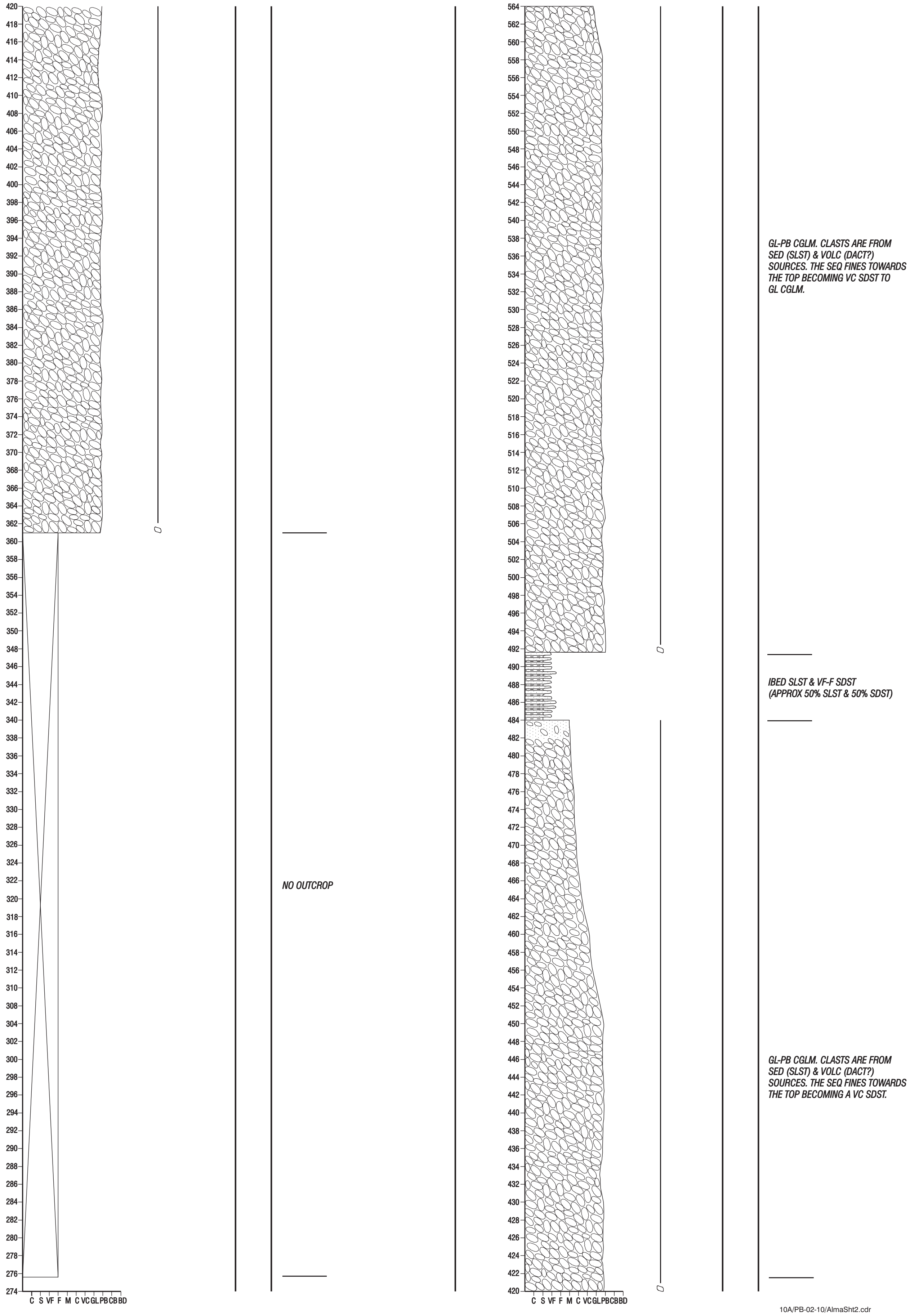


Figure 8. Type section for the Mount Alma Formation on Mount Alma (continued)

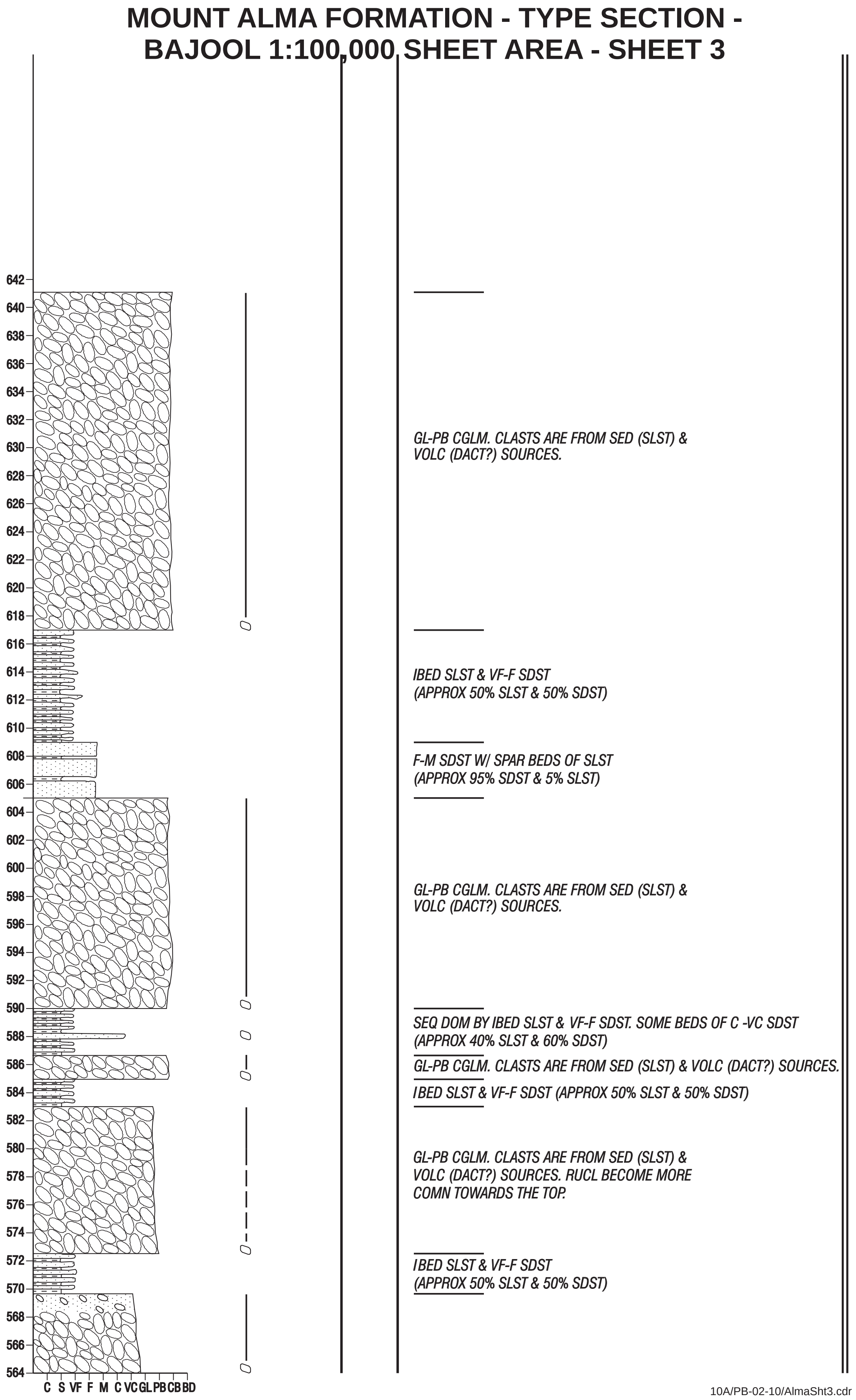


Figure 9. Type section for the Mount Alma Formation on Mount Alma (continued)

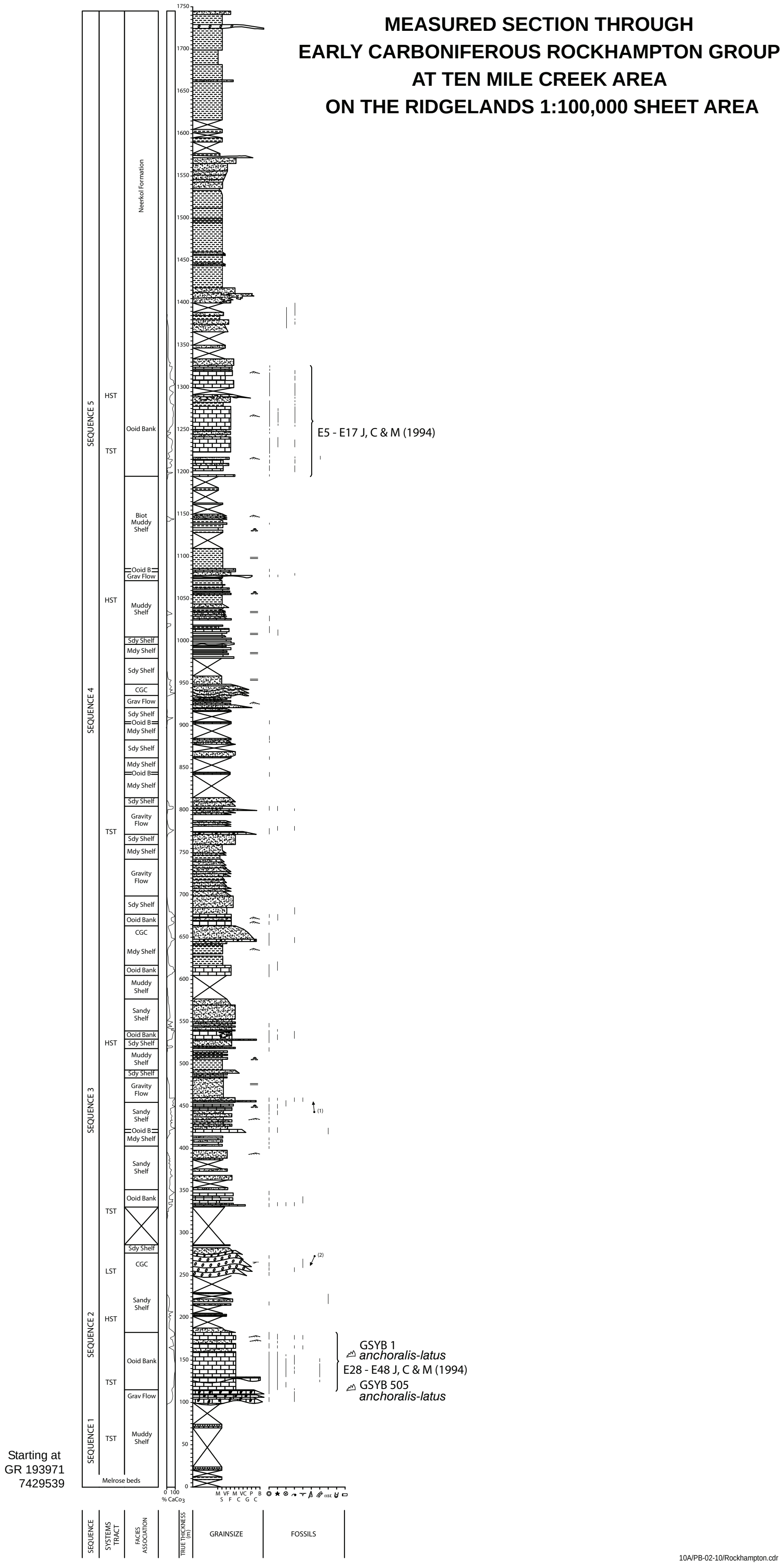


Figure 10. Measured section through the Rockhampton Group