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Geochemistry and geochronology of palaeoproterozoic Fe-rich tholeiites and metasediments from the Georgetown Inlier, North Queensland : their petrogenesis, metamorphic history, tectonic setting, and implications for relationship with the Broken Hill and Mt Isa sequences

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Geochemistry and Geochronology of Palaeoproterozoic Fe-rich Tholeiites and Metasediments from the Georgetown Inlier, North Queensland: their Petrogenesis, Metamorphic History, Tectonic Setting, and Implications for Relationship with the Broken Hill and Mt Isa Sequences.

by

Michael J Baker, B.Sc. (Hons.)

Submitted in fulfilment of the Requirements for the Degree of

Doctor of Philosophy



University of Tasmania, September, 2007

Declaration

This thesis contains no material that has been accepted for the award of a degree or diploma at any other institution. To the best of the author's knowledge, this thesis contains no material previously written or published by another person, except where due reference is given in the text. All referenced material is listed in the back of the thesis.

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Geochemistry and geochronology of Palaeoproterozoic Fe-rich tholeiites and metasediments from the Georgetown Inlier, North Queensland: their petrogenesis, metamorphic history, tectonic setting, and implications for relationship with the Broken Hill and Mt Isa sequences.

Abstract

Fe-rich tholeiites intruded the lower Etheridge Group of the Georgetown Inlier of north Queensland during the late-Palaeoproterozoic (ca. 1680-1640 Ma), during a prolonged period of rift-related sedimentation associated with extension in the Georgetown basin. The tholeiites comprise two lithologically distinct units: the extrusive Dead Horse Metabasalt (DHM), a conformable 1000m-thick series of fine- to coarse-grained basalts and pillow lavas; and the intrusive Cobbold Metadolerite (CMD), a multitude of sills and lesser dykes that range from a few metres to over 500 m in thickness. U-Pb LA-ICPMS dating of igneous zircons indicate the tholeiites were intruded in at least two pulses. The first coincided with the extrusion of the DHM and associated CMD sills at ca. 1670 Ma, while a younger pulse occurred during the deposition of the overlying Lane Creek Formation at ca. 1640 Ma.

The DHM and CMD are co-magmatic, and have compositions typical of relatively evolved, low-K, Ferich continental tholeiites and tholeiites generated in plume-triggered volcanic passive margins leading up to continental breakup. Immobile major and trace element patterns suggest that the tholeiites were derived from partial melting of the convecting depleted upper mantle depleted mantle melt, and melts evolved along a trend of increasing Fe-enrichment to strongly Fe-enriched (21.11 wt% total Fe as Fe₂O₃) ferro-tholeiites. Initial ε_{Nd} values range between +2.6 and +5.3 (average +4.0), confirming a depleted mantle source for the tholeiites. T_{DM} ages range between 1.85 Ga and 3.39 Ga (average 2.5 Ga). Several samples show slight LILE- and LREE-enrichment and have low initial ε_{Nd} values (+2.6-2.9) consistent with limited crustal contamination involving local host sedimentary rocks of the lower Etheridge Group.

Between ca. 1600 and 1500 Ma, the lower Etheridge Group was deformed and metamorphosed by at least two regionally significant orogenic events. The first event (D_1) was the most intense, producing a pervasive foliation and metamorphism (M_1) that varied in intensity from upper amphibolite facies in the east, through to lower greenschist facies in the south-west. Metasediments in the south of the inlier record *P*-*T* conditions of up to 650°C and 7.5 kbar for M_1 . D_1 was followed by a period of retrogressive metamorphism that lasted until ca. 1550 Ma, when voluminous granitoids of the Forsayth Batholith were intruded into the lower Etheridge Group. High-temperature, low-pressure metamorphism (M_2) was associated with this event. Initial *P*-*T* conditions for M_2 are inferred from a retrogressive mineral assemblage that returned conditions of 550°C and 3 kbar. EMPA monazite dating returned a possible ca. 1585 Ma age for M_1 and a ca. 1550-1530 Ma age for M_2 . A younger monazite population (ca. 1500

Ma) suggests either M_2 continued for a further 30-50 million years, or an additional weak thermal event affected the Etheridge Group at this time.

The DHM and CMD are broadly correlated temporally and geochemically to basic gneisses of the Broken Hill Group in the Willyama Inlier, and amphibolites of the Soldiers Cap Group in the Mt Isa Eastern Succession. In particular, immobile trace element ratios and REE patterns between these tholeiitic suites are virtually indistinguishable. The crystallisation ages of the DHM and CMD, as determined by U-Pb LA-ICPMS zircon dating also closely resembles established ages for the Broken Hill (ca. 1685 Ma) and Mt Isa Eastern Succession (ca. 1670 Ma) tholeiitic rocks. This relationship is of regional significance, as it suggests the terranes were spatially associated during the Palaeoproterozoic, and places the Georgetown Inlier within the Proterozoic framework of northern Australia, and likely a part of the North Australian Craton in the Palaeoproterozoic.

Geochemistry and geochronology of Palaeoproterozoic Fe-rich tholeiites and metasediments from the Georgetown Inlier, North Queensland: their petrogenesis, metamorphic history, tectonic setting, and implications for relationship with the Broken Hill and Mt Isa sequences.

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Chapter 1.

Introduction

1.1 Outline of the Study

Proterozoic basement complexes in eastern Australia include the Willyama Inlier of western NSW and eastern SA, the Mt Isa – Macarthur Basin Inliers of western Queensland and the Northern Territory, the Georgetown Inlier of north Queensland, and several smaller inlier (Coen, Yambo and Dargalong) north of the Georgetown Inlier. Two of these, the Mt Isa and Willyama Inliers, contain immense base-metal ore deposits. A growing body of evidence is showing strong similarities between these crustal blocks in terms of the nature and timing of Palaeoproterozoic magmatism and subsequent metamorphism. This has led to a number of tectonic models suggesting that these diverse crustal blocks all derive from fragmentation of a Palaeoproterozoic continental block (Giles *et al.*, 2002).

This study focuses on the geological development and magmatic, metamorphic and tectonic evolution of Proterozoic rocks in the central part of the Georgetown Inlier, the largest area of exposed Precambrian rocks in northeastern Queensland, Australia (Fig. 1.1). A major emphasis is on the timing of magmatic and metamorphic events in the western Georgetown Inlier, as this information is critical to a detailed evaluation of tectonic models claiming that the Georgetown, Mt Isa and Willyama Inliers were once contiguous as part of a Palaeozoic continental block. The geochemistry of extensive mafic igneous rocks in the central Georgetown Inlier are of particular interest to my study, and their characteristics are compared with essentially contemporaneous metabasic igneous rocks in the Broken Hill region of the Willyama Inlier (New South Wales) and in the Mt Isa Inlier of far western Queensland.

1.2 Previous Work - a Brief Review

The main features of the Georgetown Inlier were first outlined by reconnaissance mapping in the late 1950s by the Bureau of Mineral Resources (BMR) and the Geological Survey of Queensland (GSQ) (White, 1965). From 1972, BMR and GSQ began more detailed studies of the inlier that included geological mapping at

1:100,000 scale as well as geochronological, geochemical and geophysical studies (Withnall *et al.*, 1988a). Subsequent mapping of the area has been reported by Bain *et al.* (1985), Withnall (1985), and further 1:100,000 scale mapping was released in 2003 by the Queensland Department of Natural Resources and Mines (Withnall & Blight, 2003; Withnall *et al.*, 2003a; Withnall *et al.*, 2003b). Petrological studies focussing on the oldest rocks in the Georgetown Inlier have been carried out at numerous locations within central and western parts of the Inlier. Dating of various igneous and metamorphic rocks of the inlier was completed by Black *et al.* (1979), Black & McCulloch (1984, 1990), Black & Withnall (1993), Black *et al.* (1998, 2005) and Hoskin & Black (2000). Studies involving petrology, structural analysis, geochronology and geochemistry for the purpose of developing a model for the regional tectonic evolution of the Inlier include the work of Withnall (1985; 1996), Withnall *et al.* (1988a), Black *et al.* (1998, 2005), Boger & Hansen (2004), and Cihan & Parsons (2005).



Figure 1.1: Location map of the Georgetown Inlier. Georgetown and Dargalong Inliers are highlighted in darker grey.

1.3 Background – Extensive Proterozoic and Phanerozoic Mafic Magmatism

Earth history is punctuated by numerous episodes in which rapid pulses of mafic magmatism formed extensive provinces of flood basalts, on either continental crust or oceanic crust, known as Large Igneous Provinces or LIP (Coffin & Eldholm, 1994). Well studied Phanerozoic examples include the Karoo, Parana-Entendeka, Deccan, Siberian and Emeishan flood basalts, but Proterozoic examples are also known (eg. Coppermine Range - Cape Smith Fold Belt, Canada; Griselin et al., 1997). All contain vast amounts of igneous material (often in excess of 10^5 km³), and almost always consist of an extrusive component accompanied by an often voluminous suite of doleritic and gabbroic feeder sills and dykes in the underlying country rocks. The extensional environments in which flood basalt provinces occur vary from intracontinental settings in which rifting does not progress to generation of oceanic crust (eg. Siberian Traps), to regions of successful continental fragmentation and the generation of new ocean basins (eg. Parana - Entendeka, leading to opening of the South Atlantic). Consequently, the geochemistry (including isotopic characteristics) of flood basalts varies significantly. This, in turn, assists in the determination of mantle-source components and the types of contamination the magmas may have encountered through their evolution. LIP are generally interpreted to be the prime manifestation of the arrival at the Earth's surface of the head of a mantle plume (White & McKenzie, 1995; Coffin & Eldholm, 1994). Starting mantle plume heads ascend from deep in the mantle, and eventually impinge on the base of the lithosphere. The hot plume material spreads laterally, ponding beneath the lithosphere and leading to epeirogenic uplift (Nyblade & Sleep, 2003). Adiabatic decompression melting within the flattened starting-heads of upwelling mantle plumes is widely believed to be responsible for the genesis of the enormous volumes of magma forming LIP, which may form lava piles several kilometers thick (Self et al., 1997). Eventually the plume material transfers its excess heat to the overlying lithosphere, thinning the lithosphere, and in some instances, leading to creation of an ocean basin.

A second form of LIP occurs on volcanic passive continental margins, and forms at the time of final breakup of continental crust and initiation of steady state ocean crust accretion. Consequent upon final rupture of the continental crust and lithospheric mantle, massive volumes of basalt are erupted in a short time interval (up to 10 m/1000 yr; Larsen & Jakobsdóttir, 1988), leading to the formation of seaward-dipping

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reflector packages (SDRS). SDRS comprise lava flows that dip towards the centre of zones of rifting, probably largely due to post-depositional subsidence (e.g. SDRS of east Greenland; Larsen & Jakobsdóttir, 1988). Their occurrence is believed to result from the breakup and spreading of thinned continental basement adjacent to thick, pre-breakup, sediment-filled basins, with the thickest parts of the newly formed mafic crust associated with seaward-dipping reflectors (Mutter & Zehnder, 1988). Flood basalt provinces were originally considered to be characterised by the eruption of chemically homogeneous, relatively Fe-rich tholeiitic basalts (Kuno, 1969). Since the advent of more detailed geochemical analysis, studies have revealed significant compositional diversity both within and between individual provinces. In a general sense, most continental flood basalts (CFB) have the geochemical characteristics of enriched mid-oceanic ridge tholeiitic basalts, and isotopic compositions vary from MORB to oceanic-island basalt (OIB) affinities (Mahoney, 1987). However, SDRS basalts are generally MORB-like, as are some basalts in oceanic LIP such as the Ontong Java Plateau (Neal *et al.*, 1997).

The presence of abundant mafic rocks, including dykes, sills and lavas, all of similar geochemical affinities (Withnall, 1985) interbedded with and intruding sedimentary rocks in the Georgetown Inlier, suggests that similar processes that produced CFB provinces described above may have also have been involved in the development of the Georgetown Inlier mafic rocks. The tectonic setting for the deposition of the major rock package hosting the mafic volcanics in the Georgetown Inlier, the Etheridge Group, is the focus of current debate. It has been linked with the formation of the North Australian Craton (Myers *et al.*, 1996) during the Palaeoproterozoic and Mesoproterozoic, and in recent years a number of models have been put forward to describe this relationship (Etheridge *et al.*, 1987; Withnall *et al.*, 1988; Black *et al.*, 1998; Giles *et al.*, 2002; Boger & Hansen, 2004). The key features of each of these models are described in the following chapter, following a summary of the Proterozoic basement blocks of northeastern Australia, and leading into an outline of the geological evolution of the Georgetown Inlier.

1.4 Aims

This project is primarily aimed at providing a detailed understanding of the petrology, geochemistry, geochronology and tectonic significance of the mafic-igneous rocks

and enclosing sedimentary rock packages that comprise the western, low metamorphic grade part of the Palaeoproterozoic Georgetown Inlier. Such detailed information is critical for a more rigorous comparison of the Georgetown Inlier with other Proterozoic crustal blocks in northern Australia to test whether or not they were once contiguous parts of the same continental block. In particular, the formation and emplacement of the mafic rocks in the Georgetown Inlier and the 'adjacent' Mt Isa Inlier have important implications for the development of continental crust, its rifting and dispersal in forming the North Australian Craton during the Palaeoproterozoic. Other key questions that will be addressed in this project include:

- 1. Do the mafic rocks in the Georgetown Inlier show temporal and spatial variations in composition?
- 2. Are the intrusive dolerites and extrusive basalts geochemically related, as suggested by Withnall (1985) on the basis of reconnaissance data?
- 3. What are the magmatic affinities of the mafic rocks in the Georgetown Inlier, and how do they compare with Phanerozoic and other Proterozoic continental flood basalts?
- 4. How does the mafic magmatism in the Georgetown Inlier compare or contrast with Palaeoproterozoic mafic magmatism in the Broken Hill region in western New South Wales, and the Mt Isa Inlier in western Queensland? In particular, do the high-Fe mafic rocks that characterise the near-ore settings in the Broken Hill region and Mt Isa Inlier occur also in the Georgetown Inlier?
- 5. How similar are detrital zircon populations from the sedimentary packages above and below the major basaltic sequences in the Georgetown Inlier to those from the Mt Isa Inlier and Broken Hill region?
- 6. What do the structure and metamorphism tell us about the post-depositional tectonic evolution of the Etheridge Group? In particular, do the observed structural fabrics correspond to particular monazite age populations or other features of geological significance.

Chapter 2.

Overview of the Regional Geological Setting and Lithologies of the Georgetown Inlier and Etheridge Group, Northeast Queensland

2.1 Regional geology overview

2.1.1 Proterozoic Crustal Blocks of Northern Australia

The North Australian Craton contains several Proterozoic blocks that are inferred to have formed through the accretion of older crustal fragments, along with younger Proterozoic orogenic belts and sedimentary basins. Blocks that accumulated as sedimentary basins of Proterozoic age include those in McArthur and Mt Isa Inliers, Georgetown Inlier, and Victoria River basin (Fig. 2.1; Etheridge *et al.*, 1987; Myers *et al.*, 1996; Scott *et al.*, 2000). Between the older cratonic blocks the structure of the Precambrian basement has mostly been interpreted from combined aeromagnetic and gravity data due to varying amounts of Phanerozoic cover (Myers *et al.*, 1996). Geophysical and isotopic evidence suggests that Archaean and Proterozoic basement rocks were amalgamated to form the North Australian Craton by 1800 Ma (Myers *et al.*, 1996; Tyler *et al.*, 1998; Scott *et al.*, 2000).



Figure 2.1: Cartoon depicting early- to mid-Proterozoic terranes of the North Australian Craton, after Etheridge *et al.* (1987) and Myers *et al.* (1996).

Proterozoic basement complexes in northern Australia show evidence for two main periods of deposition and orogeny (Page, 1988; Etheridge et al., 1987). The first depositional cycle is best represented in the Halls Creek and Pine Creek domains of western and central Australia (Etheridge et al., 1987). It was characterised by shallowmarine clastic sedimentary sequences containing predominately mafic volcanic rocks, overlain by dominantly carbonaceous and turbiditic sedimentary rocks. Limited study of the first cycle mafic volcanic rocks suggested they have 'both MORB and continental affinities' (Etheridge et al., 1987), with some examples showing geochemical similarities to tholeiitic flood basalts of the Karoo province of southern Africa (O'Dea et al., 1997; Giles et al., 2002). The first cycle of basin formation ended with the onset of a period of inter-cratonic orogeny between 1900-1870 Ma (O'Dea et al., 1997) that affected most of the northern Australian Proterozoic terranes. This orogeny was termed the "Barramundian" by Etheridge et al. (1987). The second inter-cratonic cycle of sedimentation and basin formation is generally thought to have begun around ca. 1730-1670 Ma (Etheridge et al., 1987). Characteristics of this orogenic cycle are described in more detail below, as it directly affected the evolution of the Georgetown Inlier.

A largely convergent setting is proposed for the cratonic elements of the North Australian Craton between 1800 and 1575 Ma (Scott *et al.*, 2000). Jackson *et al.* (2000) have divided the period covered by rocks of the Victoria River, Pine Creek, McArthur, Murphy and Mt. Isa regions into the Leichhardt (c. 1800-1750 Ma), the Calvert (c. 1735-1690 Ma), and the Isa (c. 1665-1575 Ma) superbasins (Fig. 2.1). Extension in the north-east of the region began with the opening of the Davenport and Mt Isa rifts between 1835-1780 Ma, and the McArthur rift between 1750-1710 Ma (Myers *et al.*, 1996). Felsic intrusives are also recorded in the western part of the North Australian Craton at this time, in the Pine Creek and Tanami regions (1835-1815 Ma; Myers *et al.*, 1996). The rift stages of northern basins comprise 'continental-type' basalts, and felsic volcanics and intrusives. The igneous rocks are typically overlain by, or interbedded with, diverse terrestrial, peritidal to deep marine sediments that are characterised by extensive fine-grained carbonaceous siltstones and shales (Myers *et al.*, 1996; Scott *et al.*, 2000).

Extension and deposition in the northern basins continued until c. 1710 Ma, with granite intrusion occurring near the Davenport rift at this time. Further extension in

the Mt Isa region between 1680-1650 Ma led to emplacement of granitic rocks, as well as renewed deposition of sedimentary rocks (Mt Isa Group rocks; Myers *et al.*, 1996; Betts & Lister, 2001). In addition to felsic magmatism, at least five discrete major, mafic magmatic events at c. 1780, 1725, 1710, 1590, and 1300 Ma are recognised in the north and central Australian terranes over this period (Scott *et al.*, 2000). Although the presence of bimodal magmatism is often linked to episodes of crustal extension, Scott *et al.* (2000) preferred the interpretation that this magmatism was rather a consequence of a large mantle thermal anomaly beneath the North Australian Craton.

2.1.2 North-eastern Queensland Proterozoic Crustal Blocks

The Georgetown Inlier, at approximately 30,000 km², is the largest of four regions of contemporaneous Precambrian rocks in north-east Queensland. The other areas are the Dargalong, Yambo, and Coen Inliers, which all lie to the north of the Georgetown Inlier (Fig. 1.1).

The Coen Inlier is comprised of a number of Proterozoic metamorphic and granitic rocks, intruded by mid-Palaeozoic granites, and overlain locally by mid- to late-Palaeozoic sedimentary rocks (Blewett *et al.*, 1992). The Proterozoic rocks are comprised of predominately mid- to upper-amphibolite facies schist, gneiss, and some quartzite. The age of deposition and Proterozoic metamorphism in the Coen Inlier is uncertain, however zircon age dating on paragneiss from the group has returned ages of between 1590-1550 Ma (Blewett *et al.*, 1997).

The Yambo Metamorphic Group is comprised of high-grade metasedimentary and meta-igneous rocks including schist, quartzite, ortho- and paragneiss, amphibolite, and mafic granulite. A major thermal event that occurred during the mid-Proterozoic is thought to be associated to metamorphism that locally reached granulite facies conditions (Blewett & Black, 1998). The age of deposition of the sedimentary protoliths to the group is believed to have commenced after about 1643 Ma (Blewett *et al.*, 1997).

The Dargalong Metamorphic Group is comprised of a variety of rock types including migmatitic gneiss, orthogneiss, and amphibolites, along with some medium- to coarse-grained schists and lesser calc-silicate gneiss (Bultitude *et al.*, 1995). During

the mid-Proterozoic the rocks of this group experienced relatively high temperatures associated with a period of regional metamorphism. The metamorphic grade associated with this event reaches upper amphibolite facies. The age of the group is uncertain, but dating of metasedimentary rocks in the adjacent Yambo Inlier indicate deposition probably occurred towards the end of the Palaeoproterozoic (Donchak & Bultitude, 1998). Withnall *et al.* (1997) included the Dargalong Metamorphic Group into the geological framework of the Georgetown Inlier.

2.2 Tectonostratigraphic setting and development of the Georgetown Inlier and Etheridge Group

2.2.1 Geological framework

The location of this study is the regionally extensive Palaeoproterozoic to late Palaeozoic Georgetown Inlier of central northern Queensland (Fig. 1.1). The region is bounded to the west by the Carpentaria Lowlands Region and to the east by the Cairns and Clarke River Regions (Fig. 2.2). The Georgetown Inlier was originally separated into three tectonic subprovinces termed the Croydon, Forsayth, and Greenvale Subprovinces (Withnall *et al.*, 1980). The subdivisions have since been incorporated into the Etheridge Province and Croydon Province of Withnall *et al.* (1997) (Fig. 2.3).

The Mesoproterozoic Croydon Province corresponds to the Croydon Subprovince of Withnall *et al.* (1980). It is comprised of the S-type Croydon Volcanic Group and associated granites of the Esmeralda Supersuite (Fig. 2.3). The age of the province has been dated to about 1550 Ma (Black & McCulloch, 1990; Black & Withnall, 1993). The rocks are disconformably overlain by the Neoproterozoic to early Palaeozoic Inorunie Group (Fig. 2.2; Mackenzie *et al.*, 1985).

The Palaeoproterozoic Etheridge Province is divided into the Forsayth and Yambo Subprovinces (Withnall *et al.*, 1997), and corresponds to the Forsayth Subprovince of Withnall *et al.* (1980). The Yambo Subprovince is represented by the Dargalong Metamorphic Group (Fig. 2.2). The Forsayth Subprovince includes the Etheridge and Langlovale Groups, McDevitt Metamorphics, and various mafic intrusive rocks and Mesoproterozoic granites of the Forest Home and Forsayth Supersuites. Rocks of the Greenvale Subprovince of Withnall *et al.* (1980) including the Halls Reward



Metamorphics are correlated with the Etheridge Group (Withnall, 1989), and have subsequently been included with it (Fig. 2.3; Withnall *et al.*, 1997).

Figure 2.2: Simplified geological map of the Georgetown Inlier (after Withnall *et al.*, 1997). (a) Area of study.

The Etheridge Group represents the main focus of this study. It contains the majority of Palaeoproterozoic rocks in the Georgetown Inlier and crops out over approximately half of the region. The group consists predominately of shallow-marine clastic meta-sedimentary rocks, along with lesser doleritic intrusions and associated basaltic lavas. Biotite-schist, calc-silicate, and leuco-gneiss, and amphibolite occur predominately in the eastern part of the group where the metamorphic grade is higher. The Etheridge Group is thought to have been deposited between 1700-1550 Ma, based on zircon ages obtained from a number of igneous and metasedimentary rock samples within the group (Black *et al.*, 1998, 2005). The Etheridge Group is unconformably overlain to the west by the Langlovale Group and to the north by various Mesozoic and Cainozoic sediments (Fig. 2.2). The eastern and southern parts of the group are overlain by various Palaeozoic volcanics and sediments, and are bounded by the Palmerville and Burdekin River Fault systems (Fig. 2.2).

The Mesoproterozoic Langlovale Group consists of poorly sorted, feldspathic sandstone grading into massive to laminated shales, mudstones, siltstones, and sporadic sandstone beds (Withnall, 1996). The rocks are unconformably overlain by the Croydon Volcanic Group (Fig. 2.2). Due to the Langlovale Groups stratigraphic relationship with the Etheridge and Croydon Volcanic Groups, its age is constrained to about 1550 Ma (Withnall *et al.*, 1997).

The Palaeoproterozoic McDevitt Metamorphics are exposed in the northern part of the Etheridge Group alongside the Dargalong Metamorphic Group (Fig. 2.2). The rocks consist predominately of metapelites (mica schist) and subordinate micaceous metaarenite and quartzite (Withnall *et al.*, 1997). The rocks appear to have followed a similar metamorphic history to the Etheridge Group, with grades locally reaching upper-greenschist to lower-amphibolite facies. The highest grade rocks commonly occur adjacent to contacts with the Dargalong Metamorphic Group (Withnall *et al.*, 1997). The McDevitt Metamorphics have not been directly dated, but are probably of a similar age to the Etheridge Group. The rocks are unconformably overlain by Carboniferous felsic-volcanic rocks, Mesozoic sediments, and Cainozoic basalts, and are intruded by Proterozoic mafic-igneous, and Palaeozoic granitic rocks (Fig. 2.2; Withnall *et al.*, 1997; Donchak & Bultitude, 1998).



Figure 2.3: Proterozoic stratigraphic relationships in the Georgetown Inlier (after Withnall, 1996).

The metasedimentary rocks of the Etheridge Group are intruded by various Mesoproterozoic granites that include the Forest Home and Forsayth Supersuites. The Forsayth Supersuite is mainly S-type, but contains some I-type granites. It intrudes the Etheridge Group in a number of areas, with the main outcrop situated around Forsayth in the western half of the region (Fig. 2.2). The Forest Home Supersuite is dominantly

I-type, and is comprised of a number of small intrusive bodies found in the central part of the region. Numerous unnamed granites have also been identified throughout the higher grade areas of the Etheridge Group (Withnall *et al.*, 1997). Recent zircon age dating and structural observations indicate they may also be of Mesoproterozoic age (Black *et al.*, 1998).

2.2.2 Tectonic setting

The key characteristic which links all models for the formation of the Etheridge Group is that the deposition of sediments comprising the basal units of the group formed during extension of continental crust, which resulted in the formation of a shallow epi-continental or intra-continental sea. This was thought to have occurred during the second of two cycles of episodic extension, sedimentation, and magmatism (ca. 1.73-1.67 Ga; Etheridge et al., 1987) that affected several northern Australian Proterozoic terranes, including the McArthur basin, the Mt Isa Inlier, and the Georgetown Inlier (Etheridge *et al.*, 1987; Withnall 1985; Giles *et al.*, 2002).

In the Georgetown Inlier, deposition commenced with predominantly fine-grained calcareous to dolomitic sandstone, siltstone and mudstone, containing sedimentary structures indicative of a shallow-water or shoreline environment (Withnall *et al.* 1988a). Immediately following this phase of sedimentation, a phase of tholeiitic mafic magmatism occurred (ca. 1.69-1.65 Ga). This magmatic phase is also recorded within the eastern Mt Isa Inlier (Withnall 1985; Giles *et al.*, 2002). Sedimentation in the northeastern Australian basins continued until around 1.59 Ga (Page & Sun, 1998), and included rift-related sediments in the eastern Mt Isa Inlier (O'Dea *et al.*, 1997) and deep-water sediments in the Etheridge Group.

During the Mesoproterozoic a number of tectonothermal events deformed the Etheridge Group, producing definitive structural and metamorphic features. Metamorphic grade ranges from low-greenschist facies in the west and south-west through to granulite facies in the north-east, with peak metamorphic conditions probably reached in the Mesoproterozoic. Although there appears to be general consensus as to the metamorphic history of the region, conjecture still exists as to the tectonic events which led to sedimentation of the Etheridge Group: was it related to similar areas of sedimentation in northern Australia? Or did it form as a separate and

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unrelated sequence, only to be amalgamated into the north Australian craton after deposition ceased?

2.2.3 Models for the Tectonic Evolution of the Etheridge Group

Although there appears to be consensus on the style of deposition recorded in the Georgetown Inlier, the various models (Etheridge *et al.*, 1987; Withnall *et al.*, 1988; Black *et al.*, 1998; Giles *et al.*, 2002; Boger & Hansen, 2004) differ with respect to the tectonic conditions prior to its deposition. Etheridge *et al.* (1987) proposed an ensialic model for northern Australian Proterozoic tectonics. It involved processes largely independent of plate-boundary forces, including lithospheric stretching, mantle-derived magmatism and contemporaneous underplating of the crust prior to sedimentation. The processes were all initiated by the onset of small-scale mantle convection between 2.2-2.0 Ga (Etheridge *et al.*, 1987).

The second cycle, as described above, commenced with a period of crustal extension and rifting followed by post-extensional thermal subsidence with associated finegrained clastic and carbonaceous sedimentary sequences. Felsic, dominantly A-type plutonic rocks were emplaced throughout the second cycle, such as in the Mt Isa Inlier (Etheridge *et al.*, 1987; Scott *et al.*, 2000), whereas mafic rocks were generally emplaced and erupted in periods of crustal extension (Etheridge *et al.*, 1987; Withnall *et al.*, 1988a).

The absence of felsic igneous rocks in the Etheridge Group has also been observed in other northern Australia Proterozoic terranes, such as the Pine Creek Inlier (Withnall *et al.*, 1988a). The absence of coarse clastic rift-related sediments (i.e. associated with rift margin faulting) from the lowermost exposed sections of the Etheridge Group suggests that it may be related to a period of thermal subsidence during the second depositional cycle (Withnall *et al.*, 1988a). The rift-related sequences, if they do exist, must occur below the Etheridge Group and have therefore not been identified. U-Pb zircon ages in more recent work (Black *et al.* 1998; Hoskin & Black 2000) confirm that ages of the oldest recognisable units of the Etheridge Group are around 1700-1650 Ma. As the mafic rocks in the Etheridge Group were emplaced close to the interpreted base of the sequence after 1700 Ma, it has been argued that they may relate to the final stages of the extensional part of the cycle (Withnall *et al.*, 1988a).

The genesis of the mafic rocks in the Etheridge Group has been interpreted by Withnall (1985) and Withnall et al. (1988a) using a similar framework to that of Etheridge et al. (1987). However, in the model of Withnall et al. (1988a) the mafic rocks are considered to be derived directly from the convective mantle upwelling which produced crustal extension. The deposition of the Etheridge Group is interpreted to have taken place on continental crust, in either a stable shelf or epicontinental sea. As no evidence of rifting has been found in the lower parts of the sequence, the cause of the apparent rapid change from a shallow-marine to deepmarine environment that followed the eruption of mafic rocks in the formation remains unclear. Withnall (1985) suggested that the subsidence may be caused by the thinning of ductile crust rather than via rifting, as some authors have previously postulated that continental crust in the early Proterozoic may still have been hot enough to allow sub-solidus ductile flow (Wynne-Edwards 1972; Kröner 1979). Once convection and associated crustal thinning ceased, rapid subsidence would then commence. Withnall (1985) also noted that the mafic rocks show similarities to modern enriched mid-ocean ridge basalts (E-MORB), although they also share characteristics with the low-K flood basalts of Greenland and the Karoo (Cox & Hornung 1967; Clarke 1970; Brooks et al., 1976). In this instance it is the extension and accelerated subsidence of the continental basement that is proposed to have led to the eruption and emplacement of the mafic rocks.

In recent times, as an increasing amount of evidence has become available suggesting that subduction operated in the Proterozoic, plate-tectonic models of the Australian Proterozoic have become more popular (Myers *et al.*, 1996; Scott *et al.*, 2000; Giles *et al.*, 2002). In contrast to the model of Etheridge *et al.* (1987) that involved a combination of lithospheric stretching and mantle-derived magmatism, Giles *et al.* (2002) proposed that the Georgetown Inlier represents a far-field back-arc basin formed in a period of subduction driven extension. They suggested that the northern Australian Proterozoic terranes were originally part of a contiguous crustal block, and that they show contemporaneous tectono-thermal events. The formation of a back-arc basin into which the Etheridge Group was deposited was linked to a continental scale region of intermittently extended crust, which also contained the McArthur Basin and Mt Isa Inlier. A north-dipping subduction system driving the extension is believed to have existed along the Palaeo-southern edge of the central Australian Arunta Inlier



(Zhao, 1994). It would have provided, along with hypothesized slab-rollback, the necessary slab-pull force to initiate the wide region of continental extension (Fig. 2.4).

Figure 2.4: Cartoon depicting relationship between convergent margin near Arunta Inlier and wide zone of intermittent extension (after Giles *et al.*, 2002).

Alternatively, Boger & Hansen (2004) suggested that the Georgetown Inlier is unrelated to the other Australian Palaeoproterozoic terranes. They pointed to a number of key differences in metamorphic evolution and structural history between the Georgetown Inlier and adjacent Proterozoic inliers, including preservation of a two-stage metamorphic event in the Georgetown Inlier, the level of heat flow through both terranes, and the gross structural trend across the region. They propose that the Georgetown Inlier may have been accreted onto the margin of the North Australian Craton prior to 1550 Ma, rather than having always been part of it. This theory has significant implications, as it would imply the Palaeoproterozoic eastern margin of Australia was ocean-facing, and hence would not allow for plate-reconstructions of Rodinia that rely on the continuation of pre-1550 Ma features from Australia to Laurentia (e.g. Burrett & Berry 2000; Karlstrom *et al.* 2001). They also note that the metamorphism experienced by the Georgetown Inlier and the Eastern Fold Belt of the Mt Isa Inlier differ in both style and intensity. They indicated that substantial lowpressure high-temperature metamorphism and anomalously high geothermal gradients (40-60 °C km⁻¹) are absent from the geological record of the Georgetown Inlier. On the other hand, the metamorphic evolution of the Georgetown Inlier involved burial and a 'normal' geotherm (20-30 °C km⁻¹) followed by localized contact metamorphism. However, it is possible that the effects of orogenesis in the Mt Isa Inlier prior to the onset of deposition in the Georgetown Inlier would have substantially affected the "local" geotherm underlying the Mt Isa Inlier. The emplacement of voluminous granites in the Mt Isa Inlier, including the Weberra and Sybella Granites (O'Dea *et al.* 1997), prior to deposition of sediments considered contemporaneous with those of the Etheridge Group, also supports the case that geothermal gradients and later styles of metamorphism may have been affected by tectonic processes prior to 1550 Ma.

Boger & Hansen (2004) also noted that the gross structural grain of the Georgetown Inlier is orthogonal to that of the Mt Isa Inlier and the Coen and Yambo Inliers to the north. Proterozoic structures in the Georgetown Inlier generally trend east-west, whereas Proterozoic structures in surrounding terranes predominantly trend northsouth. Although this is partly true, Proterozoic structural fabrics orthogonal to those described above for the Georgetown Inlier also exist. Withnall (1996) noted that the main episodes of deformation that affected the Georgetown Inlier during the Proterozoic exhibited some variation in the trend of structural fabric across the terrane. He suggested this may be caused by either 'mega-folding' related to one of the later deformation events, or that the terrane may have been part of an arcuate fold belt. He also noted the orthogonal nature of fold trends between the first (D₁) and second (D₂) deformation events in the Georgetown region, and presented U-Pb zircon ages for syn-D₂ granites that questioned the separation of the two events (Black & McCulloch, 1990; Black & Withnall, 1993).

Despite these observations, the inferred structural disparity along with the differing styles of metamorphism between the Georgetown Inlier and other Proterozoic terranes in north-eastern Australia led Boger & Hansen (2004) to conclude that the Georgetown Inlier may have been part of a separate and unrelated lithospheric block prior to 1550 Ma. Their work has important ramifications not only for models of the

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eastern Proterozoic margin of Australia, but also for Australia's place in the Proterozoic make-up of Rodinia.

The work of Black *et al.* (1998) focussed on constraining the age of deposition of the Etheridge Group. Their study showed that the lower portion of the Etheridge Group, including the mafic igneous rocks was deposited over a period of more than 40 m.y. A U-Pb zircon age of 1695.8 \pm 1.5 Ma was determined for a granite that intrudes the sequence, indicating that deposition must have begun prior to 1700 Ma. They also noted that the age of deposition correlates with younger units in the Eastern Fold Belt of the Mt Isa Inlier, where dating by Page & Sun (1998) limited the deposition of the sedimentary precursor to felsic gneiss within the Soldiers Cap Group to before 1676 \pm 5 Ma. O'Dea *et al.* (1997) showed that the upper units of the Eastern Fold Belt correlate in age to that of the Surprise Creek Formation in the western Mt Isa Inlier. This formation is composed of conglomerate, sandstone and thinly-bedded sandstone and siltstone and is overlain by the predominantly carbonaceous siltstones of the Mt Isa Group (1652 \pm 7 Ma; Page & Sun, 1998), which hosts several world-class Pb-Zn-Ag deposits. The Mt Isa Group is thought to have been deposited in a shallow marine environment (Neudert & Russell 1981).

Subsequent work done by Black *et al.* (2005) on the age of the Etheridge Group has provided further constraints on the maximum age of deposition. They collected zircons from leucogneiss of the Einasleigh Metamorphics which returned ages in two populations of 1707 ± 8 and 1703 ± 14 Ma. Should the leucogneiss represents the altered remains of sediments derived from volcanics or epiclastic deposits, then the ages would appear to constrain the maximum age of deposition of that part of the sequence to around 1700 Ma.

The minimum ages for sedimentation noted above also match those determined for another major region of base-metal mineralisation, the Broken Hill region of the Willyama Inlier of western New South Wales. SHRIMP U-Pb zircon dating of amphibolite facies felsic metavolcanics in the Broken Hill Group by Page & Laing (1992) defined a major magmatic population at 1690 \pm 5 Ma. The upper part of this group contains the Hores Gneiss, a formation containing metamorphosed tuffaceous sediments and rhyodacitic tuff that hosts the massive Pb-Zn-Ag orebodies at Broken Hill. Noting these similarities, Black *et al.* (1998) suggested that the Georgetown region may have also provided a favourable environment for base-metal mineralisation.

The absence of exposed features that relate the Georgetown Inlier to an intracontinental rift basin raises the possibility that the terrane may in fact represent a continental margin rift setting, with no continental material existing outboard (east) of the terrane. The absence of felsic volcanics of a similar age to the Etheridge Group also suggest that the terrane was deposited in a tectonically quiet environment, possibly a passive-margin setting. Sedimentary features in the lowermost exposed members of the Etheridge Group indicate the depositional environment varied from wave-dominated shoreline to deep sub-tidal settings (Withnall, 1985). The upward fining of sediments in the sequence is also indicative of a relatively calm, uniformly subsiding shallow-water environment.

2.3 Review of Etheridge Group lithostratigraphy

The Etheridge Group, that contains over half of all exposed Palaeoproterozoic rocks in the Georgetown Inlier, is estimated to be between 6 and 11 km thick (Withnall, 1996), and its base is not exposed. Exposed rocks of the Etheridge Group have been subdivided into a number of lithologically distinct units (Fig. 2.2; Withnall & MacKenzie, 1980; Withnall, 1996). These include the Einasleigh Metamorphics, Juntala Metamorphics, Bernecker Creek Formation, Robertson River Subgroup, the Townley, Heliman, and Candlow Formations, and the Langdon River Mudstone. The Cobbold Metadolerite is also included with the Etheridge Group due to its relationship with members of the Robertson River Subgroup in the study area; however it is not stratigraphically considered a part of the Etheridge Group (Withnall, 1996). A detailed summary of the lithologies listed above, based primarily on the work of previous authors, is presented here, with detailed information on lithologies collected over the course of this study presented in the following chapter.

2.3.1 Einasleigh Metamorphics

The Einasleigh Metamorphics is exposed mostly in the south and east of the Georgetown Inlier (Fig. 2.2). It consists of biotite gneiss, calc-silicate gneiss, migmatite, and lesser leucogneiss, amphibolite, schist, and quartzite. The dominant

units are biotite and calc-silicate gneiss that generally form well-layered, laterally continuous units which appear to reflect original sedimentary layering (Fig. 2.5). Hornblende-gneiss occurs in places and is thought to be intermediate between the biotite and calc-silicate gneiss (Withnall, 1996). The calc-silicate gneiss is thought to be derived at least in part from calcareous to dolomitic psammitic rocks similar to those of the Bernecker Creek Formation (Withnall *et al.*, 1988a); however it is uncertain whether all calc-silicate gneiss in the Einasleigh Metamorphics is equivalent to the Bernecker Creek Formation. Most of the larger amphibolite bodies are believed to be of intrusive origin, and are thought to be equivalent to the Cobbold Metadolerite (Withnall, 1996). The leucogneiss has no known low grade equivalent, but is thought to have been derived from sediments of igneous origin (Black *et al.*, 2005).



Figure 2.5: Einasleigh Metamorphics biotite gneiss. Grain size variation and mineralogy possibly reflect original sedimentary layering. Percy River near Percyvale.

The regional increase in metamorphic grade eastwards in the Etheridge Group has enabled the recognition of stratigraphy traced from lower grade rocks into the higher grade and more intensely deformed Einasleigh Metamorphics. On this basis Withnall (1984, 1996) and Withnall *et al.* (1988a) suggested that the Einasleigh Metamorphics are probably equivalent to the Bernecker Creek and Daniel Creek Formations, and that parts of the Einasleigh Metamorphics may also represent units of the Etheridge Group below the basal exposed unit, the Bernecker Creek Formation. However, this correlation is less certain in the easternmost parts of the region where metamorphic grade has all but obliterated pre-metamorphic detail.
2.3.2 Juntala Metamorphics

The Juntala Metamorphics are located in the southern part of the Georgetown Inlier (Fig. 2.2). The unit consists predominately of mica schist that is believed to have originally formed from either mudstone or shale (Withnall, 1996). It has a gradational contact to the west and north with calc-silicate and biotite gneiss of the Einasleigh Metamorphics. To the east it is bounded by the Werrington Fault which juxtaposes the rocks against calc-silicate gneiss of the Einasleigh Metamorphics (Withnall, 1996). Withnall (1996) tentatively correlated the Juntala Metamorphics to the schist facies of the Corbett Formation.

2.3.3 Bernecker Creek Formation

The Bernecker Creek Formation is the lowermost exposed unit of the Etheridge Group (Fig. 2.3), cropping out mainly in the cores of regional-scale anticlines in the south-western part of the Georgetown Inlier (Fig. 2.6). However, due to the higher grade of metamorphism further east in the Etheridge Group obliterating original sedimentary features and stratigraphic relationships, Withnall (1996) has suggested that parts of the Einasleigh Metamorphics could be stratigraphically lower. The Bernecker Creek Formation consists predominantly of calcareous to dolomitic, fine-grained, subfeldspathic sandstone, siltstone, and mudstone. It is at least 2000 m thick, and its base is not exposed. Bedding in the sandstones is rarely more than a metre thick, and sedimentary layers are commonly between a few cm and a metre thick. Calc-silicate gneiss and granofels belonging to the Bernecker Creek Formation are observed further east near 'Welfern', and are considered equivalent to calc-silicate gneiss of the Einasleigh Metamorphics (Withnall, 1996). The formation is conformably overlain by the Daniel Creek Formation.

The lowermost section in the formation consists mainly of up to 100 m of laminated mudstone, siltstone, and minor fine-grained sandstone (Withnall & Mackenzie, 1980). Higher in the sequence, the frequency and thickness of sandstone layers increase, and they are interbedded with thin laminated siltstone and mudstone beds. In the uppermost part of the unit, carbonaceous mudstone or shale, and siltstone layers are overlain by relatively thick, fine-grained sandstone layers. In the larger sandstone

beds, sedimentary features include (but are not limited to) trough-cross-bedding (Fig. 2.7), ripple marks, and soft-sediment deformation. Observed soft-sediment deformation structures include load casts, boudinaged layers, and contorted or disturbed bedding (Fig. 2.8). The observed sedimentary features have led previous authors (Withnall & Mackenzie, 1980; Withnall *et al.*, 1988a; Withnall, 1996) to suggest a wave-dominated shore-line to tidal flat as the possible environment of deposition for the formation.



Figure 2.6: Generalised geology of the central Georgetown Inlier (after Withnall, 1996). Outcrop too small for this scale, including individual units of the Cobbold Metadolerite, have been omitted.



Figure 2.7: Cross-bedding in Bernecker Creek Formation sandstone; Percy River near Ortona.



Figure 2.8: Soft-sediment deformation in Bernecker Creek Formation sandstone; Percy River near Ortona.

2.3.4 Robertson River Subgroup

The Robertson River Subgroup is subdivided into the Daniel Creek Formation, Dead Horse Metabasalt, and the Corbett and Lane Creek Formations (Fig. 2.3; Withnall, 1983). The subgroup consists predominately of low-grade metasedimentary rocks and metabasalt in the south (Ortona area; Fig. 2.2), and higher-grade 'schist-facies' equivalent rocks and amphibolite north of the Robertson River, and to the east near

Gum Flats (Fig. 2.2; Withnall, 1996). Metamorphosed mafic rocks of the Cobbold Metadolerite are stratigraphically bounded by the metasedimentary rocks of the Robertson River Subgroup. The subgroup is conformably overlain by the Townley Formation to the north and north-west, and unconformably overlain to the south-west by Mesozoic sediments (Fig. 2.2).

2.3.4.1 Daniel Creek Formation

The Daniel Creek Formation conformably overlies the Bernecker Creek Formation. It differs from the Bernecker Creek Formation in being less calcareous or dolomitic, and generally more pelitic (Fig. 2.9; Withnall *et al.*, 1988a). The Daniel Creek Formation is composed of mudstone, shale or siltstone, and fine-grained sandstone, and is 1000-2000 m thick (Withnall *et al.*, 1988a). The formation is suggested to have been deposited in a sandy-deltaic environment (Withnall *et al.*, 1988a). A sequence of alternating quartzite and mica schist mapped as the Mount Helpman Member in the 'Robinhood' area (Withnall & Blight, 2003) is compositionally similar to the Daniel Creek Formation, and was therefore included within it; however the mica schist phase only becomes dominant in the Daniel Creek Formation proper (Withnall, 1996).



Figure 2.9: Lensoid bedding (centre-left) and cross-bedding (centre-right) in the Daniel Creek Formation; Percy River near Ortona.

The lower part of the formation consists of mudstone, siltstone, and lesser finegrained sandstone. The siltstone and sandstone are commonly laminated and preserve some sedimentary features including cross-bedding and lensoid bedding (Fig. 2.9). Further up in the formation, fine-grained sandstone including calcareous sandstone is more abundant (Withnall, 1996), and form beds up to 1 m thick. Mudstone and siltstone with lesser sandstone form the uppermost part of the formation, underlying the Dead Horse Metabasalt; however in the Gilbert River near 'Rungulla' the metabasalt overlies a sequence containing more abundant sandstone. This observation also prompted Withnall (1996) to suggest that the sedimentary facies in the Daniel Creek Formation are partly diachronous.

2.3.4.2 Dead Horse Metabasalt

The Dead Horse Metabasalt lies between the Daniel Creek and Corbett Formations, and consists of several hundred metres of metamorphosed mafic pillow lava, hyaloclastic breccia and structureless, massive aphanitic flows, along with some minor interbedded mudstone and siltstone. Lower grade metabasalts are metamorphosed to lower greenschist facies, and igneous textures are generally well preserved. The main outcrop area is between 'North Head', 'Gilberton' and 'Robinhood' (Fig. 2.6; Withnall *et al.*, 2003a; Withnall *et al.*, 2003b). Previous studies of mafic rocks in the Etheridge Group have focussed on their classification as either doleritic or basaltic, especially where the two layers do not overlap. However, it has become apparent over the course of this study that significant variation exists within the metabasalt layers, particularly in respect to grain size and texture. Consequently, ground truthing of areas mapped via airborne surveys as metabasalt has in fact revealed numerous metadolerite layers.

In the lower grade area of the Robertson River Subgroup, the base of the mapped metabasalt units is mostly composed of massive, fine- to medium-grained structureless flows, which commonly resemble metadolerite in texture and grainsize. The rocks are green to dark-green, and consist of randomly oriented albite laths, acicular actinolite or hornblende (at higher metamorphic grades), epidote/clinozoisite, chlorite, calcite, and minor titanite and magnetite. The middle part of the formation generally consists of fine- to very fine-grained, locally aphyric basalt flows, and the upper part consists of layers of pillow lavas and hyaloclastic breccias (metabasalt clasts in siliceous matrix), with thin inter-layered, fine-grained sheet flows or sills and minor sedimentary rocks (Fig. 2.10). In many outcrops the metabasalt is weakly to

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strongly foliated, which along with metamorphism has obliterated most igneous textures, and locally flattened pillowed basalts parallel to the foliation.

Figure 2.10: Stratigraphic relationships of the 'South Head' area. Compositional changes within Dead Horse Metabasalt outcrop also shown.

2.3.4.3 Corbett Formation

The Corbett Formation overlies the Dead Horse Metabasalt, and differs from the Daniel Creek Formation in consisting almost entirely of mudstone. The formation crops out between Gilberton in the south, and as far north as Forsayth and 'Ironhurst' (Fig. 2.6; Withnall & Blight, 2003; Withnall *et al.*, 2003a; Withnall *et al.*, 2003b). The formation is approximately 1000 m thick, and is suggested to have been deposited in a subtidal to deep-subtidal environment (Withnall *et al.*, 1988a).

The lower part of the formation consists of grey, slightly carbonaceous laminated mudstone. The upper part consists of generally green, sparsely laminated mudstone or phyllite. A quartz-rich phyllite layer marks the top of the formation (Withnall, 1996). In higher grade areas the mudstone grades into mica schist similar to that of the Daniel Creek Formation. The mica schist of the Corbett Formation differs from that of the Daniel Creek Formation in that it is generally poorly layered and contains abundant biotite porphyroblasts (Withnall, 1996). Staurolite is locally present in the mica schist as large poikiloblasts containing small quartz grains, and breaks down to form garnet plus kyanite with increasing grade. Where the quartz-rich phyllite marker

bed is absent in the higher-grade rocks, the presence of biotite porphyroblasts distinguish the Corbett Formation from the overlying Lane Creek Formation (Withnall, 1996).

A series of outcrops containing almost exclusively thickly bedded quartzite has been mapped as the Tin Hill Quartzite Member of the Corbett Formation in the area between 'Robinhood' and the Forsayth road (Withnall & Blight, 2003). It consists of a single layer of massive quartzite 5-40 m thick, and has been traced east towards the Newcastle Range. No low grade equivalent has been recognised. Withnall (1996) suggested that the quartzite was originally a chemical sediment or chert.

2.3.4.4 Lane Creek Formation

The Lane Creek Formation conformably overlies the Corbett Formation, and is the uppermost formation in the Robertson River Subgroup. It is composed of commonly carbonaceous and locally calcareous mudstone and siltstone, and lesser sandstone and limestone (Withnall *et al.*, 1988a). Its main outcrop extends in a broad belt from Forsayth in the south to 'Ironhurst' in the north, where it is overlain unconformably by Palaeozoic-Cainozoic sediments (Withnall, 1996). Various smaller outcrops occur to the west and south around 'North Head' and the Robertson River, where it is overlain by the Townley Formation (Withnall & Blight, 2003; Withnall *et al.*, 2003b). The formation is up to 2000 m thick, and suggests a continuation of the deep-water facies conditions that are inferred for the Corbett Formation (Withnall *et al.*, 1988a). Withnall *et al.* (1988a) also noted the presence towards the top of the formation of minor mudclast-sandstone containing sedimentary features that may indicate a return to a shallow-water depositional environment.

Highly carbonaceous, laminated mudstone and siltstone is characteristic of the low grade part of the Lane Creek Formation, and distinguish it from the underlying Corbett Formation (Withnall, 1996). Minor components of the formation include finegrained sandstone and thinly bedded limestone, which are interbedded with the more dominant lithologies. The mudstone and siltstone grade into mica schist in the central part of the region (Fig. 2.6), along with minor quartzite and calc-silicate rocks. The schists commonly contain andalusite and cordierite porphyroblasts, and at higher grades muscovite and sillimanite. Sillimanite porphyroblasts are commonly retrogressed to sericite. In the 'Ironhurst' area migmatite has developed (Withnall, 1996).

2.3.5 Upper Etheridge Group lithologies

The upper units of the Etheridge Group include, in order of decreasing age: the Townley, Heliman, and Candlow Formations, and the Langdon River Mudstone (Fig. 2.3). They are made up of varying proportions of siltstone, carbonaceous mudstone, silicious siltstone, fine-grained sandstone and rare limestone. The formations have been metamorphosed from greenschist through to lower amphibolite facies (Withnall, 1996). They are unconformably overlain the Mesoproterozoic Langlovale Group to the west, and by Palaeozoic and Mesozoic felsic-volcanic and sedimentary rocks to the east and north (Fig. 2.2).

2.3.6 Cobbold Metadolerite

The Cobbold Metadolerite occurs as numerous largely conformable intrusions throughout the Robertson River Subgroup (Withnall & Blight, 2003; Withnall *et al.*, 2003a; Withnall *et al.*, 2003b). The intrusions occur mostly as sills, and are anywhere between a few metres to over 500 m in thickness. Many of the outcrops can be followed along strike for several km. Dykes are uncommon, although some examples are present around 'Ortona' (Withnall, 1981; Withnall *et al.*, 2003a). The Cobbold Metadolerite is believed to have been intruded in at least two pulses (Withnall, 1985). One is believed to have coincided with the extrusion of the Dead Horse Metabasalt, with a second, younger pulse thought to have occurred immediately after the deposition of the Lane Creek Formation (Withnall, 1996). No mafic intrusive rocks of Proterozoic age occur above the Lane Creek Formation in the Etheridge Group.

2.4 Mesoproterozoic Granites

Mesoproterozoic granites in the central part of the Georgetown Inlier form the Forsayth Batholith. The batholith outcrops over \sim 700 km² and is subdivided into nine named, and several other unnamed plutons (Black & Withnall, 1993). Units relevant to this study are mentioned here; however a comprehensive list of Proterozoic granites of the Georgetown Inlier is given by Withnall *et al.* (1997).

2.4.1 Forsayth Granite

The Forsayth Granite's main exposure occurs just to the north of Forsayth (Fig. 2.6; Withnall & Blight, 2003). It consists of commonly foliated, grey to dark-grey, medium- to coarse-grained, strongly porphyritic biotite granite. Conventional U-Pb zircon dating has returned an age of 1550 ± 7 Ma for the Forsayth Granite (Black & McCulloch, 1990).

2.4.2 Digger Creek Granite

The Digger Creek Granite consists of numerous small exposures that outcrop between Forsayth and as far south as 'Percyvale' (Fig. 2.6; Withnall & Blight, 2003). The main exposures occur near 'Robinhood' station (Fig 2.6), where the granite has caused the recrystallization of 'schist-facies' rocks of the Robertson River Subgroup. The granite consists of medium- to coarse-grained muscovite leuco-granite and muscovite pegmatite. A U-Pb zircon SHRIMP ion-microprobe age of 1554 \pm 10 Ma has been obtained for the Digger Creek Granite (Black & Withnall, 1993).

2.4.3 Ropewalk Granite

The Ropewalk Granite is also composed of numerous small exposures, the largest of which are between Forsayth and 'Robinhood' (Fig. 2.6; Withnall & Blight, 2003). The granite consists of grey to dark-grey, foliated, medium-grained muscovite-biotite and quartz granite, and commonly contains melanocratic pockets of biotite (Sample 208; Appendix 1).

2.4.4 Mount Hogan Granite

The Mount Hogan Granite is exposed in the south of the region near 'Mount Hogan' station (Fig. 2.6; Withnall *et al.*, 2003a). The granite consists of pink to reddishbrown, medium- to coarse-grained biotite-muscovite granite. A U-Pb zircon SHRIMP ion-microprobe age of 1549 ± 25 Ma has been obtained for the Mount Hogan Granite (Black & Withnall, 1993).

2.5 Review of the Structural evolution of the Etheridge Group

Previous authors have identified six major deformation events that have affected the Etheridge Group (D_1 - D_6 ; Withnall *et al.*, 1997). The first two folding events ($D_1 \& D_2$) were inferred to have occurred close to each other at about 1550 Ma, and are both associated with metamorphism. The metamorphism ranges from lower-greenschist to upper-amphibolite and locally granulite facies, with peak metamorphic conditions reached during D_2 (Withnall, 1996). D_1 is best preserved in the western, lower grade part of the region, where later overprinting effects are at their weakest. Withnall (1996) indicated that D_1 appears to be of a relatively uniform intensity over the entire region. D_1 generally produced tight to isoclinal, upright to overturned folds, with a strong axial-plane foliation. F_1 wavelengths are generally of the order of 1-10 km (Withnall *et al.*, 1997).

The age of D_1 remains uncertain. It was directly dated by Black *et al.* (1979) using Rb-Sr whole-rock and mineral techniques to 1570 ± 20 Ma. However, the Rb-Sr whole-rock technique for dating metamorphic rocks has long been considered obsolete (Black & Withnall, 1993). Subsequent dating of post-D₂ granitoids (Black & McCulloch, 1990; Black & Withnall, 1993) has shown that the age of D₁ obtained using the Rb-Sr technique was anomalously young. Recent dating of leucogneiss from the Einasleigh Metamorphics by Black et al. (2005) showed that a significant metamorphic event occurred at 1562 ± 4 Ma. However, it remains unclear whether this age is representative of D_1 , or alternatively represents the onset of the second major episode of regional deformation and metamorphism. If the latter is correct, it would indicate that D_2 lasted for at least 12 million years, and that D_1 is likely to have occurred ~20 million years earlier at ca. 1580 Ma, when peak tectonothermal events were drawing to a close in the Broken Hill Block and the Mt Isa Eastern Succession (Black et al., 2005). EMPA dating of monazites from amphibolite facies metasedimentary rocks of the Robertson River Subgroup by Cihan et al. (2006) suggest the latter is true and they give a weighted average age for the peak metamorphic conditions associated with D_1 of 1586 ± 6 Ma.

The effects of D_2 are best seen in the eastern half of the Georgetown Inlier, as both the intensity of folding and deformation increase eastwards. In westernmost outcrops of the Etheridge Group D_2 forms local, weak, northerly trending crenulations that lie sub-parallel to the axial planes of large open folds which deform F_1 folds (Withnall,

1996). In easternmost outcrops D_2 forms generally north trending tight to isoclinal folds with wavelengths up to 2 km, which are associated with either a strong crenulation cleavage, or schistosity (Withnall, 1996). Where this deformation is strongest, D_2 overprints the earlier metamorphic fabric as well as primary features including sedimentary layering.

Using Rb-Sr techniques Black *et al.* (1979) obtained an age of 1469 ± 20 Ma for D₂. However, subsequent dating of Proterozoic felsic granites and volcanic rocks that postdate D₂ using SHRIMP and conventional multi-grain zircon techniques (Black & McCulloch, 1990; Black & Withnall, 1993) has shown that this age is obsolete, and that D₂ is about 80 Ma older, probably occurring at around 1550 Ma (Black & Withnall, 1993). Recent dating of leucogneiss from the Einasleigh Metamorphics by Black *et al.* (2005) returned an age of 1555 ± 3 Ma, which they correlated to the peak of metamorphism associated with the D₂ event. Monazites analysed by Cihan *et al.* (2006) yielded a weighted average age of 1542 ± 8 Ma, which they correlate to marking the beginning of D₂ in the Robertson River Subgroup.

 D_3 produced folding of varying intensity across the Georgetown Inlier. In the central part of the inlier near 'Robinhood', D_3 produced generally open to tight, upright to overturned folds with axial planes striking approximately south-easterly to easterly (Withnall, 1996). Further east, and to the south of the Gilberton Fault, D_3 produced generally weak open folds with wavelengths of less than 1 km to over 10 km (Withnall, 1996). In these areas, D_3 is also associated with small-scale folds, weak crenulations, and low-grade retrogressive metamorphism, although due to the weak development of D_3 foliations, it is difficult to distinguish these features from later events.

The age of D₃ was dated at 967±28 Ma using Rb-Sr techniques from samples of outcrops supposedly possessing well-developed S₃ schistosities (Black *et al.*, 1979). Subsequent studies of the same outcrops have suggested that the foliation S₃ may in fact be S₂, and that the age obtained by Black *et al.* (1979) may represent a partially reset D₂ age (Withnall, 1996). Bell (pers. comm., 1984 in Withnall, 1996), on the basis of shallowly dipping axial planes in some F₃ folds, suggests that D₃ may be early Palaeozoic in age, and related to deformation in the Greenvale Subprovince of Withnall (1989). Cihan *et al.* (2006) noted a large grouping of monazite ages in their study which gave a weighted mean age of 1512 ± 5 Ma. They postulated that this age

may represent a thermal pulse or hydrothermal activity that may or may not have been associated with the later stages of D_2 , and could therefore represent the age of D_3 .

 D_4 has had little effect on the western and southern parts of the Etheridge Group, and the event is best recorded in the eastern, higher-grade parts of the inlier. F_4 folds are rare; however, Withnall (1996) suggests open, mesoscopic F_4 folds are common near Einasleigh and further north. Black *et al.* (1979) suggested an age of about 400 Ma for D_4 based on K-Ar and Rb-Sr mica ages, but did not directly attempt to date the event. McNaughton (1980) obtained K-Ar muscovite ages of 410 ± 12 Ma and 414 ± 12 Ma from the Einasleigh Metamorphics east of the study area, which he interpreted to have crystallised syn-D₄. Other geochronological studies from areas surrounding the Georgetown Inlier have also returned ages around 400 Ma (e.g. Withnall *et al.*, 1991) indicating that this age is of regional significance; however, it is yet to be determined conclusively that this age corresponds to D₄.

Later deformation events (D_5 - D_6) produced open folding and weak crenulations in the Etheridge Group. D_5 is represented by open folds with east-striking axial planes (Withnall, 1996). Minor crenulations have also been recognised in the Robertson River area. D_6 is mainly represented by crenulations. The ages of D5 and D6 have not been determined, but are believed to be of late-Palaeozoic age (Withnall, 1996).

2.6 Faulting

Three major fault systems can be found in the central part of the Georgetown Inlier (Fig. 2.6). The Robertson, Gilberton and Delaney fault systems all or partly occur within the study area, and are described in detail below.

2.6.1 Robertson Fault System

The Robertson Fault System has been mapped for approximately 45 km northwest from the northern end of the Agate Creek Volcanics to 2 km south of the Gongora Granodiorite (Withnall & Blight, 2003; Withnall *et al.*, 2003b). The fault system splits into two branches north of the Robertson River (Fig. 2.6; Withnall *et al.*, 2003b).

The eastern branch and main stem of the Robertson Fault both show a sinistral movement of 5 km, while at least 2 km of dextral movement has occurred on the western branch (Withnall, 1996). Vertical displacement involving west-block-down

movement of up to 150 m also occurred along the western and main stems of the fault during the Pliocene (Withnall, 1996). This displacement is observable along the section of the fault which runs parallel to the Robertson River, where the Jurassic Hampstead Sandstone is now in fault-contact with metasediments of the Etheridge Group (Fig. 2.6; Withnall *et al.*, 2003b).

The age of the Robertson Fault is uncertain. It is thought to be at least Pliocene in age, but could possibly be as old as late Palaeozoic, based on late Palaeozoic volcanics and intrusives lying along the south-easterly projection of the fault system (Withnall, 1996).

2.6.2 Gilberton Fault System

The Gilberton Fault System has been mapped for approximately 55 km northeast from its southern contact with Jurassic sediments 20 km southwest of Gilberton, to its intersection with the Ballynure Fault, east of 'Gum Flats' station (Fig. 2.6; Withnall *et al.*, 2003a).

Considerable dextral movement of at least 50 km along the Gilberton Fault has been hypothesised (Withnall *et al.*, 1980), as along the southern half of the fault high grade rocks of the Einasleigh Metamorphics are in contact with low grade rocks of the Robertson River Subgroup. The juxtaposition of rocks of different metamorphic grades can also be explained by vertical movement; however, the distance over which faulting is thought to have occurred makes this unlikely (Withnall *et al.*, 1980).

In outcrop, the Gilberton Fault ranges from narrow zones of fault gouge (Withnall, 1996), through to zones of intense shearing and fracturing up to 100 m wide. Mylonitic leucogranite is recognised close to the southern side of the fault south of 'Gum Flats', indicating that some ductile deformation was involved during movement along parts of the fault.

The age of the Gilberton Fault is thought to be late Ordovician or early Silurian (Withnall, 1989), based on the presence of mylonite pebbles in early Silurian conglomerates in the Broken River Province to the southeast of the Georgetown Inlier (Withnall *et al.*, 1988b). Mylonitisation along the Gilberton Fault is believed to have occurred at this time.

2.6.3 Delaney Fault System

The Delaney Fault is discontinuous, and forms a series of north-striking, narrowly offset segments that continue for at least 100 km (Fig. 2.2; Withnall, 1996). The northern limit of the fault has been mapped to 3 km south of the Yataga Granodiorite, and the southern limit is mapped to the Gilberton Fault, approximately 5 km southwest of Gilberton (Fig. 2.6; Withnall *et al.*, 2003a). A number of small, mafic dykes occur along the length of the fault, and are interpreted to be Carboniferous to early Permian in age (Withnall & Blight, 2003).

In most areas, the fault has caused no observable displacement of stratigraphy. One notable exception is in the central part of the region, where the fault has displaced the contact of the Robin Hood Granodiorite by approximately 1 km (Withnall, 1996; Withnall & Blight, 2003). Small vertical displacements have otherwise been suggested to account for the lack of lateral movement (Withnall, 1996).

The age of the Delaney Fault is uncertain. Withnall (1996) suggests that it may have formed between the early Devonian and Permian, based on the displacement of the Robin Hood Granodiorite and the intrusion of mafic dykes.

Chapter 3.

Geology and Structure of key lithologies within the lower Etheridge Group

3.1 Introduction

Based on geological maps by Withnall & Blight (2003), Withnall *et al.* (2003a), and Withnall *et al.* (2003b), I spent four weeks carrying out detailed geological sampling of key units across the western Georgetown Inlier, and completed several transects aimed at elucidating the geochemical, structural and metamorphic evolution of the relatively low-grade section of the inlier. Over 300 samples were collected for this study, and a representative selection is listed in Appendix 1.

This chapter describes in detail the composition, structure and field relationships of the lower Etheridge Group lithologies which form the focus of this study, in particular the mafic igneous rocks of the Dead Horse Metabasalt and Cobbold Metadolerite. Geological relationships that clarify the timing, relative ages, and geological sequence of events for the various lithologies are also described. The result is a geological framework which allows the tectonic significance of the mafic igneous units to be confidently examined.

The location of the area of study is shown in Figure 2.2. A geological map encompassing the area of study is shown in Figure 2.6. Lithologies not shown include Palaeozoic and younger rocks, felsic pegmatites, and various small mafic igneous veins, sills and dykes belonging to the Cobbold Metadolerite and Einasleigh Metamorphics.

3.2 Etheridge Group

Apart from several granitoid samples of the Forsayth Batholith and Mt Hogan Granite, the remaining samples from this study were collected from lithologies belonging to the lower Etheridge Group. Sampling was concentrated on the Dead Horse Metabasalt and Cobbold Metadolerite members; however numerous samples of surrounding metasediments including the Einasleigh Metamorphics, Bernecker Creek Formation, and Robertson River Subgroup were also collected. Outcrop of metaigneous and metasedimentary rock is generally poor across the region, with best exposures found in road cuttings and in river beds where seasonal water flow and sand abrasion have exposed fresh outcrops.

3.2.1 Einasleigh Metamorphics

Biotite gneiss belonging to the Einasleigh Metamorphics in the Gum Flats area (Fig. 2.6) is generally grey and consists mainly of coarse-grained biotite, quartz, andesine to albite, calcite, locally garnet, and accessory haematite (Fig. 3.1; Sample 317, Appendix 1). Thin discontinuous layers composed of aggregates of retrogressive sericite replacing sillimanite are also visible in the gneiss. With decreasing feldspar and increasing muscovite, biotite gneiss grades into mica schist. With increasing grade, the biotite gneiss grades into migmatite near 'Percyvale' (Fig. 2.6); and consist mainly of concordant leuco-granitic pods (leucosomes) and biotite-rich selvedges (melanosomes) ranging in thickness from <1 cm to up to 50 cm (Fig. 3.2).



Figure 3.1: Thin section photo of Einasleigh Metamorphics biotite gneiss. Mineral labels: bt = biotite; qz = quartz; ab = albite; gt = garnet. Sample 317, Appendix 1.

Mafic layers up to 1 m across are found in areas of migmatisation, and are probably equivalent to amphibolites from lower grade areas. Samples of the amphibolite contain foliated, euhedral hornblende, albite, and minor quartz, clinopyroxene, and magnetite (Sample 318, Appendix 1). Samples of calc-silicate gneiss vary from pink through to grey and green, and contain varying proportions of quartz, microcline, plagioclase, hornblende, clinopyroxene, sphene, and locally garnet, epidote/clinozoisite, and calcite. Magnetite and haematite/ilmenite are also present as accessory minerals.



Figure 3.2: Migmatitic Einasleigh Metamorphics; Percy River near Percyvale.

3.2.2 Bernecker Creek Formation

Representative samples of the Bernecker Creek Formation were collected from the Percy River north of 'Ortona' (Fig. 2.6), and are mostly fine- to very fine-grained sandstones (Appendix 1). Representative samples 7, 13 and 124 contain a high percentage of grain-supported quartz (60-70% in sample 7, 50-60% in sample 13, 50% in sample 124). Minor mineral phases include calcite (Sample 7) and epidote (Sample 13). In sample 124, calcite comprises approximately 30% of the sample, and muscovite comprises a further 5-10% (Fig. 3.3; Sample 124, Appendix 1). Accessory opaque minerals are predominately magnetite and chalcopyrite. Detrital zircon is rare, but can be found in some samples of fine-grained sandstone. Grainsize variation in sedimentary minerals, such as quartz, is minimal, indicating that sandy layers in the formation are generally well sorted and mature. However, metamorphic minerals,

such as epidote in sample 13, show greater variation in grainsize, with larger grains forming microscopic layering in the sediments (Fig. 3.4; Sample 13, Appendix 1).

The Bernecker Creek Formation grades eastwards into calcareous mica-schist and quartzite that contains calc-silicate minerals including hornblende, plagioclase, epidote, and clinopyroxene. With increasing metamorphic grade and intensity of deformation, most sedimentary features observable in the lower grade rocks are obliterated. Bedding is usually preserved, but becomes increasingly difficult to distinguish from tectonic structures.

Higher grade samples of the Bernecker Creek Formation were collected near 'Gum Flats' station to the east of 'Ortona' (Fig. 2.6). Samples from this area are mostly of mica schist, and sedimentary features are mostly obliterated by overprinting of metamorphic textures. Mineral texture varies from coarse- to fine-grained depending on the sedimentary precursor, and the dominant mineral phases include calcite, muscovite, biotite, quartz, and lesser plagioclase feldspar and scapolite (Fig. 3.5, Samples 292; Appendix 1).



Figure 3.3: Crossed-polarised thin section photo of Bernecker Creek Formation fine-grained sandstone; Gilbert River, Ortona. Sample 124, Appendix 1.



Figure 3.4: Thin section photo of Bernecker Creek Formation sandstone containing metamorphic epidote; Percy River, Ortona. Sample 13, Appendix 1.



Figure 3.5: Thin section photo of coarse-grained, calcareous Bernecker Creek Formation mica schist. Mineral labels: cal = calcite; bt = biotite; qtz = quartz. Sample 292, Appendix 1.

3.2.3 Robertson River Subgroup

Numerous samples of the constituent lithologies within the Robertson River Subgroup were collected over the course of this study. Samples of the Dead Horse Metabasalt were primarily collected for geochemical analysis, with several samples also employed for geochronological and lithological analysis. Samples of the Daniel Creek Formation, Corbett Formation, and Lane Creek Formation were also collected for lithological, structural, geochronological, and thermobarometric analysis as described in this and the following chapters.

3.2.3.1 Daniel Creek Formation

Representative low-grade samples of the Daniel Creek Formation were collected from the Percy River north of 'Ortona' (Fig. 2.6), and are mostly siltstone (Appendix 1). Sample 97 from this locality contains very fine-grained, aligned quartz and muscovite, porphyroblasts of a prograde metamorphic mineral (possibly staurolite) pseudomorphed by chlorite, and biotite (Sample 97, Appendix 1). In thin section, muscovite-rich/quartz-poor and muscovite-poor/quartz-rich layering is visible, and probably indicates small-scale changes in the clastic composition of original sedimentary layers. Metamorphic biotite has grown in the matrix along with quartz.

Low-grade rocks of the Daniel Creek Formation are mainly located between Gilberton and the Agate Creek gemfields (Fig. 2.6; Withnall *et al.*, 2003a), and near 'South Head' (Fig. 2.6; Withnall *et al.*, 2003b). Elsewhere the low-grade rocks grade into mica schist and quartzite. As with the Bernecker Creek Formation, the higher grade of metamorphism and more intense deformation have obliterated most small-scale sedimentary structures; however bedding is generally preserved. Mica schist belonging to the Daniel Creek Formation was sampled from the 'Gum Flats' area (Fig. 2.6). Samples of the mica schist mostly contain fine-grained quartz, biotite, muscovite, lesser plagioclase, and in some samples small garnet porphyroblasts and aluminosilicate (Fig. 3.6; Sample 311, Appendix 1). Biotite is generally the dominant mica phase. Aluminosilicate in the mica-schist was probably andalusite, and where originally present, has been mostly retrogressed to sericite. The quartzite contains mostly quartz, plagioclase, and lesser white mica.



Figure 3.6: Thin section photo of Daniel Creek Formation biotite schist containing fine-grained quartz, biotite, muscovite, and lesser plagioclase. Sample 311, Appendix 1.

3.2.3.2 Dead Horse Metabasalt

Samples of the Dead Horse Metabasalt collected for this study can be split into two groups based on their sampling location. The South Head rocks were collected from the Gilbert River southeast of 'South Head' station (Fig. 2.6; Withnall *et al.*, 2003b), and the Ortona rocks were sampled from the area surrounding 'Ortona' station (Fig. 2.6; Withnall *et al.*, 2003a). In both localities the metamorphic grade of the rocks is predominantly greenschist facies.

The South Head rocks are best exposed to the southeast of 'South Head' station, where they occur as a discontinuous series of outcrops along the base of the Gilbert River for over 4 km (Fig. 2.10). Over the length of the section, the dominant texture of the metabasalt varies considerably, from an inferred margin of fine grained doleritic material, through coarser grained dolerite, then fine grained basalt flows, to predominantly pillowed basalt layers (Fig. 2.10). The predominantly coarse-grained dolerite layers are up to 700 m thick, and are composed of actinolite, albite, and chlorite, along with minor sericite, epidote, relic igneous plagioclase and clinopyroxene, and accessory haematite and chalcopyrite (Fig. 3.7; Sample 241, Appendix 1). They are similar in appearance and mineralogy to samples of the



Cobbold Metadolerite from 'Ortona' that occur stratigraphically below the Dead Horse Metabasalt, and may correlate with it.

Figure 3.7: Thin section photo of doleritic Dead Horse Metabasalt showing low-grade metamorphic assemblage. Mineral labels: chl = chlorite; ab = albite; ep = epidote; ser = sericite. Sample 241, Appendix 1.

Stratigraphically overlying the doleritic layers are predominantly fine-grained, locally massive basalt flows up to 500 m thick. Individual basalt flows are generally no thicker than 20m, and are usually separated by thin doleritic layers and minor pillowed layers. The basalt flows are generally finer-grained, yet mineralogically similar to the doleritic layers. In both the basaltic and doleritic samples accessory haematite and chalcopyrite are metamorphic, and have grown in place with quartz. Haematite primarily occurs as concentric growth bands around chalcopyrite, and the growth bands may also include limonite (Fig. 3.8; Sample 229, Appendix 1). The growth bands may have formed by supergene alteration of the chalcopyrite grains.

The inferred top of the sequence is composed predominantly of a thick sequence of pillowed basalt layers up to 300 m thick. Individual pillows are generally no larger than 1 m in diameter, and pillowed basalt layers are generally less than 10 m thick (Fig. 3.9). Pillow cores and rims are mineralogically similar, containing very fine-grained chlorite, igneous plagioclase, and minor actinolite, epidote, haematite, and magnetite. Pillow rims generally contained a higher proportion of matrix glass,

primarily altered to chlorite (Fig. 3.10; Sample 234, Appendix 1). Spaces between pillows were most likely filled by quenched, basaltic glass that has subsequently weathered (Fig. 3.9).



Figure 3.8: Thin section photo of concentric haematite bands surrounding chalcopyrite grain in aphyric metabasalt; Dead Horse Metabasalt, South Head. Sample 229, Appendix 1.



Figure 3.9: Pillowed basalt outcrop; Gilbert River, South Head.



Figure 3.10: Thin section photo of Dead Horse Metabasalt pillowed basalt rim containing finegrained chlorite, albite, and lesser calcite; Gilbert River, South Head. Sample 234, Appendix 1.

The Ortona rocks are best exposed to the north and south of 'Ortona' station, where they occur as numerous, laterally extensive layers up to 1 km thick (Fig. 2.6; Withnall *et al.*, 2003a). Samples of the metabasalt were collected mostly from outcrops in the Percy and Gilbert River beds, with the largest sequence of outcrops occurring to the northwest of 'Ortona' in the Gilbert River. Here the metabasalt has been folded by a west-recumbent regional scale F_1 anticline (Fig. 2.6; Withnall, pers. comm., 2003). The Ortona metabasalts generally display the same variation in texture as the South Head samples, but on a smaller scale. Alternating layers of fine-grained metabasalt, and coarser-grained metadolerite can be found in all sections; however, pillowed basalt layers are restricted to the largest exposed sequence of outcrops in the Gilbert River. The exact sequence of layering is difficult to identify in the metabasalt due to the randomly occurring nature of outcrops; however, metadolerite sills have been mapped stratigraphically below some metabasalt layers in the region (Withnall *et al.*, 2003a), and therefore may lie at the base of the sequence.

The fine-grained metabasalt and coarser-grained metadolerite layers are each less than 30 m in thickness and are mineralogically similar, containing actinolite, chlorite, albite, and minor epidote, magnetite, and relic igneous clinopyroxene and plagioclase (Fig. 3.11; Sample 111, Appendix 1). Samples of pillowed basalt from outcrops in the area are composed primarily of albite and chlorite. Accessory minerals include

actinolite, epidote, titanite, and relic igneous clinopyroxene and plagioclase grains (Fig. 3.12; Sample 80, Appendix 1).



Figure 3.11: Thin section photo of fine-grained Dead Horse Metabasalt; Percy River, Ortona. Mineral labels: act = actinolite; chl = chlorite; ab = albite; ep = epidote. Sample 111, Appendix 1.



Figure 3.12: Thin section photo of Dead Horse Metabasalt pillowed basalt containing finegrained chlorite, albite, and accessory epidote and titanite; Gilbert River, Ortona. Sample 80, Appendix 1.

In the higher grade area of the Robertson River Subgroup around 'Robinhood', the Dead Horse Metabasalt consists of fine-grained, foliated amphibolite. In samples of the higher grade rocks, the dominant amphibole is actinolite grading into hornblende, chlorite is absent, and metamorphic albite is present along with quartz and minor Ti-magnetite (Fig. 3.13; Sample 261, Appendix 1). Some outcrops of the amphibolite in the Robertson River are coarser-grained with a gneissic appearance, and have been intruded by thin dioritic veins containing a higher percentage of quartz and plagioclase, and less hornblende (Fig. 3.14).



Figure 3.13: Thin section photo of Dead Horse Metabasalt amphibolite containing fine-grained hornblende, quartz, albite, and minor Fe-Ti oxide; Robertson River, Robinhood. Sample 261, Appendix 1.

3.2.3.3 Corbett Formation

Representative low-grade samples of the Corbett Formation were collected from the Percy River north of 'Ortona' (Fig. 2.6), and vary from siltstone to mudstone (Appendix 1). Sample 278B from Ortona contains very fine-grained, aligned quartz, muscovite, and lesser biotite (Fig. 3.15; Sample 278B, Appendix 1). Higher grade samples were collected from between Forsayth and 'Robinhood' (Fig. 2.6), and are predominately mica schist. Representative samples commonly contain fine-grained foliated biotite, muscovite, and blocky quartz (Fig. 3.16; Sample 306; Appendix 1). Garnet and staurolite are also identified in some samples. Garnet is commonly fine-

grained and surrounded by very fine-grained biotite. Staurolite is poikiloblastic, containing fine-grained quartz and retrogressive muscovite.



Figure 3.14: Dead Horse Metabasalt amphibolite containing dioritic vein; Robertson River near Robinhood.



Figure 3.15: Thin section photo of Corbett Formation mica schist; Gum Flats. Sample 278B, Appendix 1.



Figure 3.16: Thin section photo of Corbett Formation mica-schist containing metamorphic garnet porphyroblasts wrapped by later biotite and muscovite. Sample 306, Appendix 1.

3.2.3.4 Lane Creek Formation

Representative low-grade samples of the Lane Creek Formation were collected from west of Georgetown near Stockyard Creek and South Head (Fig. 2.6), and are predominately comprised of siltstone (Appendix 1). Samples of the siltstone from South Head contain very fine-grained, foliated, quartz and muscovite (Fig. 3.17; Sample 245, Appendix 1). Higher grade samples were collected from between Forsayth and 'Robinhood' (Fig. 2.6), and are predominately mica schist (Fig. 3.18; Sample 304, Appendix 1).



← Figure 3.17: Thin section photo of very fine-grained Lane Creek Formation quartz-muscovite schist; South Head. Sample 245, Appendix 1.

The higher grade mica schist commonly contains foliated, elongate muscovite that is microscopically layered between blocky quartz grains. Minor mineral phases include biotite, and locally garnet. Patches of unfoliated quartz, chlorite and muscovite occur throughout the sample, and may represent replacement of metamorphic andalusite.



Figure 3.18: Thin section photo of fine-grained Lane Creek Formation mica-schist; Robinhood. Sample 304, Appendix 1.

3.2.4 Cobbold Metadolerite

In the low grade area of the Robertson River Subgroup, including near 'Ortona' and Stockyard Creek, the metamorphic grade of the metadolerite is lower- to midgreenschist facies. Samples interlayered with or below the Dead Horse Metabasalt are fine- to coarse-grained, and commonly contain anhedral metamorphic actinolite, albite, chlorite, quartz, epidote, and sericite replacing igneous plagioclase feldspar (primarily labradorite), rare igneous clinopyroxene, and accessory magnetite and chalcopyrite (Fig. 3.19; Sample 121; Appendix 1).



Figure 3.19: Thin section photo of Cobbold Metadolerite from layer adjacent to Dead Horse Metabasalt; Ortona. Mineral labels: ab = albite; chl = chlorite; ep = epidote; qz = quartz. Sample 121, Appendix 1.

Samples of the metadolerite from stratigraphically above the Dead Horse Metabasalt are petrographically similar to those that occur with the metabasalt. Variations between the two metadolerite groupings, if they exist, would be observed in the concentrations of some trace elements, and are therefore described in Chapter 6. Metadolerite outcrops from lower grade areas contain in places minor felsic segregations that have possibly formed as late-stage felsic differentiates, or as peprites along the margins of the intrusion and sediments (Fig. 3.20). Where the felsic segregations occur within the metadolerite, a decrease in the modal percentage of felsic minerals is observable around their margin (Fig. 3.21). Felsic components of the metadolerite are commonly composed of coarse-grained calcite, albite, quartz, and muscovite, and minor chlorite, sphene, magnetite, and accessory zircon.



Figure 3.20: Peperite at contact of Bernecker Creek Formation (left) and Cobbold Metadolerite (to right of photo, not shown); Percy River, Ortona.

One interesting section of exposed metadolerite in the Percy River near 'Ortona' is composed of spheroidal 'felsic' xenoliths surrounded by dolerite (Fig. 3.22). The section is approximately 1 m wide, and is in contact with the Bernecker Creek Formation. However, the xenoliths mineralogically different to other felsic segregations found within the metadolerite as these contain coarse-grained, anhedral clinozoisite and lesser epidote (Fig. 3.23; Sample 178, Appendix 1). The mineralogy of metadolerite surrounding the xenoliths appears equivalent to metadolerite from other locations (Fig. 3.23). The origin of the xenoliths is unclear, but they may have been locally derived from adjacent Bernecker Creek Formation calcareous quartz sandstones, and were possibly rounded by chemical reaction rather than mechanical abrasion.



Figure 3.21: Felsic segregations in low-grade Cobbold Metadolerite; Percy River, Ortona.



Figure 3.22: 'Felsic' xenoliths in low-grade Cobbold Metadolerite, Percy River, Ortona. Red colouration in blebs is a weathering effect unrelated to mineralogy.



Figure 3.23: Thin section photo of clinozoisite-epidote-rich 'felsic' xenolith (left) and low-grade Cobbold Metadolerite (lower right); Percy River, Ortona. Sample 178, Appendix 1.

In the high grade areas of the Robertson River Subgroup, including near 'Robinhood' and 'Gum Flats', actinolite in the metadolerite is replaced by blue-green hornblende, and albite is the dominant feldspar. Epidote and chlorite become minor phases, and are absent in amphibolite facies samples. As with the surrounding metasediments and metabasalt, the metadolerite is commonly deformed, and igneous textures and minerals are rarely preserved. Samples from the 'Robinhood' area mostly reached amphibolite facies, and are generally foliated. They commonly contain euhedral hornblende, quartz, albite, minor anhedral biotite, and magnetite (Fig. 3.24; Sample 215, Appendix 1). Samples from the 'Gum Flats' area reflect a change in metamorphic grade from upper greenschist facies to amphibolite facies. Samples of the metadolerite from close to the Gilberton Fault are at greenschist facies, containing actinolite (e.g.: Sample 291; Appendix 1). Further north, hornblende replaces actinolite and albite, while quartz and magnetite are common (e.g.: Sample 300, Appendix 1). Hornblende and quartz show evidence of recrystallising igneous clinopyroxene, and sericite replacing igneous plagioclase. Higher grade metadolerite samples were collected from within areas mapped as the Einasleigh Metamorphics by Withnall et al. (2003a), and are described in that section.



Figure 3.24: Thin section photo of amphibolite facies Cobbold Metadolerite containing finegrained hornblende, quartz, albite, and minor biotite; Robinhood. Sample 215, Appendix 1.

3.3 Mesoproterozoic Granites

Four samples of Mesoproterozoic granite that intruded the Etheridge Group were collected during this study, including three samples of the Ropewalk Granite, and one sample of the Mount Hogan Granite. Samples of the Ropewalk Granite consist of medium- to coarse-grained biotite and lesser muscovite, along with quartz, titanite, orthoclase, and lesser plagioclase (Samples 208, 209, 210; Appendix 1). The sample of the Mount Hogan Granite consists of medium- to coarse-grained quartz, microcline, and fine-grained muscovite (Sample 135; Appendix 1).

3.4 Folding in the Etheridge Group

Of the six episodes of deformation recognised in the Etheridge Group, only the first four events are described in detail below. D_5 and D_6 had limited effect on the Etheridge Group and were not recognised in outcrops over the course of this study; as such they are not described here.

3.4.1 First Deformation Event

Rocks of the study area generally show well-developed axial planar foliations (S_1) that have intensely folded bedding (S_0), and vary from a slaty cleavage through to schistosity depending on the grade of metamorphism. In the study area F_1 folds are generally between 4-8 km in wavelength and isoclinal (Fig. 3.25). Near 'Ortona' the axial planes of F_1 folds are generally steep to vertical and trend west-southwest (Fig. 3.26a). Eastwards, in the area south of 'Gum Flats', axial planes retain their steep to sub-vertical nature but trend northeast (Fig. 3.27). North of 'Gum Flats' the steeply dipping nature of the event is maintained; however, the trend is more easterly (Fig. 3.27).



Figure 3.25: Core of isoclinal, upright F_1 fold, Bernecker Creek Formation; Percy River, Ortona. In the north of the study area, approximately 20 km west of 'Robinhood' (Fig. 2.6), the majority of D_1 structures have been overprinted by the more intense D_2 event, hence the location and nature of F_1 folds in this area is uncertain. Structural data obtained from this area indicate the overprinting effects of D_2 (Fig. 3.28). In Figure 3.28, the general trend recognised for S_1 at 'Ortona' and 'Gum Flats' appears to continue, with steeply dipping west-southwesterly planes denoting S_1 . The repetition

of stratigraphy in the area may relate to a later stage of folding, most likely F_3 ; however, the overall northward younging of the stratigraphy from 'Ortona' through to 'Robinhood' has previously led Withnall (1996) to suggest the rocks may lie on the northern limb of a major F_1 anticline.

3.4.2 Second Deformation Event

Within the study area the effects of D_2 are variable. Near 'Ortona' there is little or no folding associated with it; however, local, weak crenulation cleavages are attributed to D_2 . They generally form a steeply dipping, east-northeasterly trending cleavage (Fig. 3.26b). Towards the east and north of 'Ortona' D_2 forms generally open folds associated with a weak crenulation cleavage. D_2 is more prominent in the north of the study area near 'Robinhood' and east near 'Gum Flats' and 'Percyvale'. At these localities D_2 has resulted in tight to isoclinal, short wavelength folding associated with a strong, differentiated cleavage or schistosity. Structural data collected from the area surrounding 'Robinhood' indicates a clustering of poles to S_2 in the southwest quadrant (Fig. 3.28). Withnall (1996) has attributed this effect to later folding associated with D_3 .

 S_2 crenulation cleavages in the area south of 'Gum Flats' strike northeast to north and are steeply to sub-vertically dipping (Fig. 3.27). North of 'Gum Flats' S_2 remains steeply dipping; however, the strike is generally towards the east (Fig. 3.27). Large northeast-trending folds in the area that fold both S_0 and S_1 are interpreted to be F_2 . Poles to S_1 in the north of the area illustrate the northeast trending nature of these folds (Fig. 3.27). The change in the general trends of both S_1 and S_2 in the area surrounding 'Gum Flats': from northeast in the south, to east in the north, is consistent with the presence of a major, open F_3 fold in the area. Withnall (1996) suggested that this fold has an axial plane which passes through the middle of the area, and strikes approximately east. North of 'Gum Flats', further rotation of axial planes to F_2 folds towards the southeast support this argument (Fig. 3.27).


Figure 3.26: Structural data for the area near 'Ortona' station. The area corresponds to the 'Ortona domain' of Withnall (1996). (a) S₀ and S₁ data for north and south Ortona. Boundary between areas is northing 7876000 ('Ortona' station). (b) Combined L₁ and S₂ data for 'Ortona' area. Equal area stereonet projection used.



Figure 3.27: Structural data for the area near 'Gum Flats' and 'Percyvale' stations. The area corresponds to the eastern 'Mount Hogan domain' and western 'Welfern domain' of Withnall (1996). Boundary between areas is northing 7885100 ('Gum Flats' station). Equal area stereonet projection used.



Figure 3.28: Structural data for the area near 'Robinhood' station. The area corresponds to the 'Robinhood domain' of Withnall (1996). Equal area stereonet projection used.

3.4.3 Third Deformation Event

In the study area, the effects of D_3 are variable. Foliation associated with D_3 has not been recognised in the study area except as the axial planes of crenulations of S_2 . North-west of 'Robinhood' D_3 is dominant, and has produced tight overturned folds with northerly dipping axial planes and fold axes plunging towards the northwest (Withnall, 1996). As discussed above, this has effectively confined poles to S_2 to the southern hemisphere (Fig. 3.28). Near 'Ortona' F_3 folds have not been recognised. However, Withnall (1996) suggested that in this area the effect of D_3 may have been to further tighten F_1 folds, with no separate F_3 folds being generated.

3.4.4 Fourth Deformation Event

 D_4 has had little effect on the study area, apart from possibly producing a broad warping in the trends of F_3 axial planes near 'Robinhood'. As such, no definitive qualitative structural data relating to this event was obtained. Elsewhere the effects of

 D_4 are minor, including forming small kinks and crenulations in pre-existing fabrics. In the south of the study area near 'Ortona' and 'Gum Flats', no D_4 -related structures were recognised.

Chapter 4.

U-Pb igneous and detrital zircon, and EMPA monazite age constraints on the deposition, igneous emplacement, and post-depositional evolution of the lower Etheridge Group

4.1 Introduction

A combination of conventional multigrain LA-ICPMS U-Pb zircon dating and electron microprobe U-Th-Pb analysis of monazite (EMPA) is used on a number of igneous and metasedimentary samples in order to better constrain the evolution, emplacement and depositional ages the lower Etheridge Group. Zircons were primarily sourced from samples of the Cobbold Metadolerite and Dead Horse Metabasalt exhibiting granophyric or peperitic textures and high Zr contents (Zr > 150 ppm), and were collected from a number of localities within the central Georgetown Inlier (Fig. 4.1). In addition, detrital zircons were obtained from samples of metasedimentary rocks surrounding the mafic units. Samples containing monazite were primarily sourced from metapelitic schists of the lower Etheridge Group in the areas surrounding Gum Flats and Robinhood stations (Fig. 4.1).

In this chapter, U-Pb dating of igneous zircons from the mafic rocks of the lower Etheridge Group is used to test, in conjunction with available geochemical data, whether they relate to a single magmatic event, or multiple events. It is also an aim of this study to better constrain the timing of this magmatism, as well as that of the main deformation events, in order to establish a realistic estimate for the timing of deposition and deformation of the lower Etheridge Group. The results of this study are also compared to magmatic, detrital, and metamorphic ages obtained from similarly aged terranes in eastern Australia, to investigate regional correlations between the terranes, and what implications this has for the Proterozoic history of the Georgetown Inlier.



Figure 4.1: Geological setting of the central Georgetown Inlier, with locations for analysed samples of zircon and monazite. Geological units as for Figure 2.6.

4.2 Previous work

A lack of suitable rock types has hampered previous efforts to directly measure the depositional age of the lower Etheridge Group. However, these studies have proved more successful in constraining the deformation history of the Etheridge Group, and have placed a minimum age on its deposition. Deposition of the Etheridge Group was thought to have occurred prior to 1570 ± 20 Ma, based on a combination of detailed structural analysis and Rb-Sr whole-rock isochron ages, which suggested that D₁ and D₂ occurred at 1570 ± 20 Ma and 1470 ± 20 Ma, respectively (Black *et al.*, 1979). However, subsequent SHRIMP and conventional multi-grain zircon dating has shown

that at least the younger of these two estimated ages is too young (Black & McCulloch, 1990; Black & Withnall, 1993; Black *et al.*, 1998, 2005).

Recent work has established a minimum age of around 1550 Ma for deposition of the Etheridge Group. Support for this age comes from a conventional multi-grain U-Pb zircon age of 1552 ± 2 Ma (Black & McCulloch, 1990) and a SHRIMP zircon age of 1548 ± 18 Ma (Black & Withnall, 1993), both obtained from the overlying Croydon Volcanic Group. Multi-grain zircon ages for both the Forsayth (1550 ± 7 Ma) and Mistletoe (1544 ± 7 Ma) Granites also support this minimum age (Black & McCulloch, 1990). Similar ages have also been obtained for the Lighthouse, Digger Creek and Mt Hogan Granites (Black & Withnall, 1993). Leucogneiss from the Einasleigh Metamorphics within the Etheridge Group has also returned an age of 1555 ± 3 Ma (Black *et al.*, 2005).

Establishing a maximum age limit for the Etheridge Group has proven difficult due to the absence of exposed basement rocks. Black & McCulloch (1984), using Sm-Nd data from a diverse range of lithologies, arrived at an age of 2490 ± 70 Ma for the initial formation of the Etheridge Group. Subsequent work (Withnall *et al.*, 1988; Black & Withnall, 1993) identified similar-aged zircon in some metasedimentary and igneous rocks, and has since established the ca. 2500 Ma age more likely represents the provenance rather than the age of onset of deposition for the Etheridge Group. This led Black *et al.* (1998) to suggest the 2490 \pm 70 Ma age as an upper age limit for the deposition of the Etheridge Group.

EMPA monazite dating of metasedimentary samples from the Robertson River Subgroup was recently conducted by Cihan *et al.* (2006). They suggested that peak conditions for M₁ and M₂ were reached at 1586 ± 6 Ma and 1542 ± 8 Ma respectively. Based on ages obtained from the oldest monazites analysed, they also suggested that metamorphism leading into M₁ may have begun as early as 1655 ± 2 Ma for some sedimentary components of the lower Etheridge Group. A younger group of monazite ages yielded a weighted mean age of 1512 ± 5 Ma, which they inferred to represent a later stage of hydrothermal or thermal activity in the lower Etheridge Group.

The Etheridge Group is now thought to have been deposited prior to 1656 ± 2 Ma, based on the U-Pb zircon age obtained from a leucogabbro sill intruded into the Lane Creek Formation in the middle of the Group (Black *et al.*, 1998). An age obtained

from granite gneiss intruded into the Einasleigh Metamorphics $(1696 \pm 2 \text{ Ma})$ also indicates that deposition of the older part of the Etheridge Group, below the Lane Creek Formation, took at least 40 million years (Black *et al.*, 1998). The 1707 \pm 8 Ma and 1703 \pm 14 Ma zircon populations from leucogneiss of the Einasleigh Metamorphics may also provide constraints on the maximum depositional age for at least part of the lower Etheridge Group (Black *et al.*, 2005). However, to date no definitive age exists for the onset of deposition across the entire Etheridge Group.

4.3 Objectives of the geochronological study

Apart from the studies of Black *et al.* (1998, 2005), little data exists relating to geochronological studies of the mafic and metasedimentary rocks from the lower Etheridge Group. The analysis of metasedimentary rocks has previously concentrated on areas that reached high grades of metamorphism; consequently, lower-grade areas are poorly constrained. Therefore, the key objectives of this geochronological study are:

- To determine the age of emplacement for the Dead Horse Metabasalt and Cobbold Metadolerite members.
- To determine if the magmatism that led to the emplacement of the Dead Horse Metabasalt and Cobbold Metadolerite was synchronous, and short-lived; or if it was related to a number of 'pulses' that were not temporally associated.
- To provide realistic estimates for the maximum and minimum ages of deposition of the lowermost units of the Etheridge Group based on magmatic and detrital zircon ages.
- To compare mafic and detrital zircon populations from the Etheridge Group to those from sections of the Mt Isa Inlier and Broken Hill Block in eastern Australia, to determine if any regional correlation exists, and to support the comparative geochemical study of Chapter 7.
- To determine the age(s) of metamorphic monazite growth, and where possible relate these to significant structural and metamorphic events in the lower Etheridge Group.

Details of the sampling and analytical techniques used and additional information on the samples analysed are provided below. A general overview of the petrology, mineralogy, and stratigraphic position of samples discussed here is given in Chapter 3 and Appendix 1.

4.4 Method

4.4.1 Sampling

Mafic samples containing zircons were collected from the most fractionated outcrops of both the Dead Horse Metabasalt and Cobbold Metadolerite. These outcrops are commonly found in contact with surrounding metasedimentary units. Metasedimentary samples containing detrital zircons were collected from finegrained, locally massive sandstone and psammitic layers of the Daniel Creek Formation. Where possible, samples of metasedimentary rock were collected from outcrops near fractionated mafic samples. Samples used for monazite analysis were collected from metasedimentary rocks of the lower Etheridge Group at Gum Flats and Robinhood (Fig. 4.1).

4.4.2 Laser ablation ICPMS analysis of zircon

The use of Laser Ablation Inductively-Coupled Plasma Mass-Spectrometry (LA-ICPMS) for the U-Pb dating of zircon has become increasingly popular in recent years due to its availability, low cost, and fast through-put of samples relative to other dating techniques. As such, this technique has been increasingly applied in numerous geochronological studies worldwide (eg. Hirata & Nesbitt, 1995; Horn *et al.*, 2000; Berry *et al.*, 2001; Li *et al.*, 2001; Tiepolo, 2003; Jackson *et al.*, 2004).

The LA-ICPMS technique is best suited to projects requiring a large number of analyses or the analysis of large zircons, due to the relatively large beam diameter (30-50 μ m) of the laser ablation microprobe. As such, the technique is unsuitable for high-resolution analysis of grains, or the analysis of grains containing complex zoning. Difficulties cited in using the laser ablation technique include the induced fractionation of Pb relative to U during the ablation, transport, and ionisation processes, as well as the difficulty in correcting for common-Pb based on ²⁰⁴Pb due to isobaric interference from ²⁰⁴Hg (Jackson *et al.*, 2004). Various methods have been

described in order to overcome these issues (eg. Hirata & Nesbitt, 1995; Horn *et al.*, 2000; Li *et al.*, 2001). However, collecting laser ablation data in time-resolved mode does, at least in part, provide information to recognise these problems where they occur. By using time-resolved analysis of the grains, zones with differing U-Pb ratios can be identified and the data reduced accordingly.

4.4.2.1 Sample preparation

The majority of analysed samples were crushed using a jaw crusher and chrome-steel ring-mill at the University of Tasmania School of Earth Sciences. Individual zircons were then hand picked from panned separates of the appropriate density and mounted in epoxy for analysis. The mounts were polished to expose the zircon grains. Three mafic samples were cut from thin section blocks and prepared as 1-inch round mounts. Limited cathodoluminescence (CL) imaging on several samples was performed on the Cameca SX-100 Electron Micro-Probe (EMP), and the FEI Quanta 600 Scanning Electron Microscope (SEM) at the University of Tasmania Central Science Laboratory. CL imaging was conducted to reveal the internal structures of the zircons for the purpose of interpreting zircon growth and metamorphic histories, as well as choosing appropriate sites for LA-ICPMS analysis on some grains. Backscattered electron (BSE) imaging was also carried out on the University of Tasmania SEM. This was designed to reveal any structures not apparent in the CL images, as well as to identify included phases and surface features for the purpose of choosing appropriate sites for analysis. Internal structures of zircons revealed using CL imaging were also apparent, though generally less distinct. The Mineral Liberation Analyser (MLA) attached to the SEM at the University of Tasmania was used for rapid, accurate location of zircons in mounts cut from sample blocks. Metamorphic and/or igneous rims identified from the imaging techniques on some individual igneous zircons were too thin for analysis by LA-ICPMS.

4.4.2.2 Laser ablation ICPMS operating parameters

Samples were analysed using the HP 4500Plus ICPMS coupled with a Merchantek 266 nm frequency-quadrupled Nd:YAG Laser Ablation system (LA-ICPMS) at the University of Tasmania School of Earth Sciences. A combination of Ar and He gas

was used to transport the ablated material to the ICPMS for isotopic quantification. Isotopic masses collected using the LA-ICPMS system were ⁹⁶Zr, ¹⁴⁶Nd, ¹⁷⁸Hf, ²⁰²Hg, ²⁰⁴Pb, ²⁰⁶Pb, ²⁰⁷Pb, ²⁰⁸Pb, ²³²Th, and ²³⁸U. Zircons were analysed with a 30 µm diameter beam, a laser repetition rate of 5 Hz, and a power setting of 0.2 mJ. The ICPMS was operated in a time-resolved mode, collecting one point per mass per sweep of the mass range with a dwell time of 10 msec/mass. Immediately prior to analysis grains were ablated at 1 Hz for 3-4 sec in order to remove any surface Pb contamination. Each analysis then consisted of a measurement of average background intensities (counts per second) on each mass for 30 sec prior to ablation, followed by 30 sec of ablation (Fig. 4.2). This was followed by an approximate two minute washout period during which the signal intensity returned to background level before the next zircon grain was analysed. Sample runs consisted of 12 unknowns bracketed by 6 standard analyses. Zircons from the international standards Temora and 91500 were analysed as unknowns to monitor drift and data quality during the period of the analyses (Black *et al.*, 2003; Wiedenbeck *et al.*, 2004).



Figure 4.2: Examples of count rates for element isotopes of single zircon grains analysed by LA-ICP-MS; (a) grain 1, and (b) grain 5, both from Sample 6.

4.4.2.3 Data reduction and presentation

Raw data produced by the ICPMS was collected and reduced using the LAMTRACE program (van Achterbach *et al.*, 2001) with the assistance of Dr. Sebastien Meffre. Time-resolved spectra from each analysis were then examined for changes in the ratio of ²⁰⁷Pb/²⁰⁶Pb, which may indicate zones of Pb-loss or common-Pb. Net count rates for each mass were obtained by subtracting average background count rates from

average signal intensities taken ~30 sec after beginning ablation once the signal had stabilised (Fig. 4.2). Homogeneous standard grains with low common-Pb produced near-steady signals with no resolvable change in the signal intensities of measured isotopic masses (e.g. Fig. 4.2a).

Analysed isotopic masses were used to obtain the isotopic ratios ²⁰⁷Pb/²⁰⁶Pb, ²⁰⁶Pb/²³⁸U, ²⁰⁷Pb/²³⁵U, and ²⁰⁸Pb/²³²Th. The ²⁰⁷Pb/²⁰⁶Pb ratio was used for the majority of age estimates, with the remaining three isotopic ratios used to estimate discordance and identify grains containing high common-Pb. Subsequently, individual grains containing zones with widely differing U-Pb isotopic ratios illustrated in their time-resolved spectra were discarded before calculating sample weighted mean ages. This was done to limit the presence of common-Pb affected grains in weighted mean age calculations, as the presence of common-Pb can significantly raise the ²⁰⁷Pb/²⁰⁶Pb and ²⁰⁸Pb/²³²Th ages of a sample, in addition to greatly increasing a sample's ²⁰⁸Pb/²³²Th age relative to its ²⁰⁶Pb/²³⁸U age (Griffin *et al.*, 2006). The ²⁰⁷Pb/²⁰⁶Pb ratio is used to represent the age of individual grains, as it is the least susceptible to common-Pb contamination and laser induced fractionation of the U-Pb isotopic system, and has been successfully used to represent U-Pb zircon ages in numerous other studies of eastern Australian Proterozoic terranes (e.g. Page & Sun, 1998; Page *et al.*, 2005a; Griffin *et al.*, 2006).

Sample ages were calculated from populations of ²⁰⁴Pb-corrected analyses. In many samples the MSWD (Mean Square of Weighted Deviates) was sufficiently high, and the probability of equivalence was sufficiently low that simple weighted mean ages were unrepresentative of the crystallisation age. High MSWD and low probabilities of equivalence may suggest the data represent the amalgamation of zircons from more than one age population. Therefore, to calculate a more realistic weighted mean age, analyses with high common-Pb, and analyses with ²⁰⁷Pb/²⁰⁶Pb ages significantly different from the sample mean were excluded until the remaining data were within analytical uncertainty of one another.

In addition to common-Pb, zircons in a number of igneous samples experienced appreciable Pb-loss, recognisable by the skewing of analyses towards younger ages in the U-Pb concordia diagrams (Ireland *et al.*, 1998). Consequently, analysis of zircons that have been affected by Pb-loss will return values that represent the minimum age of their growth. As zircon is an effectively closed system below 900 °C, Pb-loss will

generally only occur as a result of radiation damage, intense metamorphism, alteration, or hydrothermal fluid activity (Mezger & Krogstad, 1997). Although no distinct measurable overgrowths were observed in this study, cumulative probability diagrams in some samples show small groupings of grains with ages below ca. 1600 Ma, possibly relating to periods of regional metamorphism and/or late stage alteration.

Ablation- and transport-related elemental fractionation of U relative to Pb was encountered during the analysis of individual grains. Therefore corrections were made by matching peak and background signal intensities between sample and standard analyses with identical time intervals. A calibration standard (NIST-612) was also used to correct for elemental fractionation, as well as the mass bias between the U-Pb and U-Th isotopic systems.

In addition to the calculated weighted mean age method outlined above, age populations and provenance were also determined using cumulative probability diagrams, as they illustrate a spectrum of zircon-forming events which have affected a sample population. The diagrams used in this study were constructed with *Isoplot* version 3.0 (Ludwig, 2003). For this study, the cumulative probability diagrams were primarily used to ascertain provenance ages from detrital zircons in metasedimentary samples. Probability diagrams are also presented for samples of igneous rocks, in order to show that the data are either normally distributed, or that in some cases the samples contain inherited zircons from earlier events. For populations of detrital zircons, peaks on the probability diagrams approximate the time of zircon-forming events. Therefore, approximate ages were taken from discrete peaks, rounded to the nearest 10 Ma, and quoted as ca. ages. As probability diagrams do not allow an exact age determination or the estimation of error, ages within 2σ (up to 30 Ma) of the ca. age were considered to belong to the zircon-forming event. In addition to these limitations, Raetz et al. (2002) also cautioned against interpretation of subtle 'shoulders' occurring on peaks in probability diagrams. These 'shoulders' are associated with either the overlapping of normally distributed populations producing artificial distributions, or may represent younger Pb-loss.

The ²⁰⁷Pb/²⁰⁶Pb and ²³⁸U /²⁰⁶Pb ratios for individual analyses are listed in Tables 4.1, 4.2, and 4.4. The remaining isotopic ratios are listed in Appendix 2. Individual zircon analyses for each sample are illustrated in inverse-concordia (Tera & Wasserburg, 1972) diagrams constructed using *Isoplot* version 3.0 (Ludwig, 2003).

Label	U (ppm)	Th/U	$^{238}\mathrm{U}/^{206}\mathrm{Pb}$	$\pm 1 \sigma$	$^{207}Pb/^{206}Pb$	$\pm 1 \sigma$	²⁰⁷ Pb/ ²⁰⁶ Pb Age (Ma)	$\pm 1 \sigma$	Label	U (ppm)	Th/U	$^{238}U/^{206}Pb$	$\pm 1 \sigma$	$^{207}Pb/^{206}Pb$	$\pm 1 \sigma$	²⁰⁷ Pb/ ²⁰⁶ Pb Age (Ma)	$\pm 1 \sigma$
Sample	e: 6 (North Or	rtona)							Sample	: 161 (South	Ortona)						
- 1.1	1380	0.55	3.43854	0.02169	0.10082	0.00139	1639	26	-	928	1.31	3.31022	0.02297	0.10428	0.00243	1702	43
∼ 20	786	0.45	3.32941	0.02348	0.10133	0.00171	1649	31	2	975	1.21	3.39684	0.01872	0.10514	0.00260	1717	45
∽ ¹⁴ ₽	592	0.45	3.51061	0.02082	0.10091	0.00149	1641	27	б	1178	1.39	3.27110	0.02101	0.10473	0.00251	1710	4
ہ h-4	763	0.50	3.43964	0.04360	0.10501	0.00363	1714	64	4	721	0.81	3.15846	0.02300	0.11359	0.00309	1858	49
و د01	1160	0.51	3.49649	0.02286	0.09939	0.00145	1613	27	5	728	0.85	3.38442	0.06632	0.10659	0.00614	1742	106
о те	393	0.39	3.52829	0.03040	0.10455	0.00235	1706	41	9	867	1.11	3.36499	0.04053	0.11082	0.00545	1813	89
= cte	1242	0.52	3.38635	0.02229	0.10061	0.00154	1635	28	7	691	0.99	3.37809	0.02424	0.10608	0.00293	1733	51
12 be	1440	0.59	3.47525	0.05986	0.10227	0.00243	1666	4	8	934	1.10	3.38314	0.02499	0.10231	0.00298	1666	54
U-									10	832	0.96	3.37840	0.02458	0.10147	0.00259	1651	47
4 Sampl	e: 8 (North OI	rtona)							11	594	0.99	3.19677	0.03367	0.10409	0.00311	1698	55
- is	907	0.52	3.50303	0.06825	0.09979	0.00316	1620	59	12	688	0.85	3.21242	0.07736	0.12024	0.00655	1960	97
r7 Anto	1216	0.46	3.47197	0.05489	0.10138	0.00334	1650	61	13	1071	1.09	3.30011	0.04005	0.10629	0.00244	1729	86
m n	1615	0.52	3.47089	0.02976	0.10143	0.00139	1650	25	14	532	0.82	3.35719	0.05300	0.10990	0.00325	1620	157
4 e	1485	0.64	3.52365	0.03277	0.10060	0.00142	1635	26	18	944	0.98	3.30281	0.06093	0.10638	0.00330	1788	102
∿ ate	321	0.28	3.46139	0.04239	0.09860	0.00228	1598	43	19	516	0.98	3.21390	0.06801	0.12985	0.00484	1923	189
ہ م	1308	0.52	3.47374	0.03050	0.09993	0.00136	1623	25	21	1396	1.27	3.38298	0.02559	0.10300	0.00167	1691	73
2 or	1233	0.51	3.51337	0.03502	0.10076	0.00150	1638	28	23	1120	1.08	3.32767	0.02990	0.10825	0.00252	1640	83
2 Zir	1312	0.50	3.43505	0.03080	0.10163	0.00147	1654	27	24	1509	1.19	3.32550	0.04315	0.12232	0.00383	1726	86
0.0									25	798	1.18	3.28632	0.02875	0.13978	0.00292	1622	136
supl.	e: 159 (South	Ortona)							26	1164	1.03	3.42314	0.03399	0.11085	0.00280	1723	84
∼ fr	763	1.09	3.21678	0.08550	0.10239	0.00335	1668	09	27	823	1.34	3.22662	0.06095	0.11783	0.00401	1740	185
m 0	1353	1.40	3.47188	0.06250	0.10668	0.00304	1743	52	28	1003	1.10	3.37757	0.03255	0.10356	0.00250	1534	101
4	956	1.05	3.14738	0.11802	0.10480	0.00490	1711	86	29	931	1.02	3.33597	0.02634	0.10990	0.00225	1685	91
ی Iaf	1102	0.74	3.30541	0.09539	0.10426	0.00338	1701	09	30	972	1.08	3.29344	0.04555	0.10810	0.00290	1707	91
ہ ic	791	0.99	3.29305	0.07296	0.10374	0.00313	1692	56									
2 581	852	0.85	3.39516	0.06513	0.10679	0.00307	1745	53	Sample	: 172 (South	Ortona) [†]						
⊒ mn	624	0.85	3.35760	0.06367	0.10596	0.00328	1731	57	1	885	1.00	3.45526	0.02272	0.10360	0.00270	1633	12
2 12	907	1.04	3.08833	0.02553	0.10709	0.00178	1675	63	3	729	0.75	3.58031	0.02525	0.10338	0.00271	1578	12
15 15	1168	0.92	3.50227	0.03919	0.10591	0.00194	1716	90	4	1035	1.17	3.67439	0.03857	0.10435	0.00365	1537	17
19 18	1017	1.04	3.29107	0.02902	0.10149	0.00166	1695	65	5	610	0.77	3.34659	0.03465	0.12465	0.00344	1643	18
<u>®</u>	433	0.74	3.14006	0.02897	0.10658	0.00228	1673	134	9	326	0.82	3.45754	0.04375	0.11125	0.00578	1617	23
6 ed	1290	0.78	3.39091	0.02620	0.10706	0.00188	1766	52	7	341	0.95	3.43572	0.04298	0.11080	0.00457	1628	22
50 50	1125	0.67	3.54371	0.04584	0.10108	0.00223	1683	88	8	369	0.99	3.41198	0.03521	0.11793	0.00370	1625	18
23 7	1933	1.61	3.40948	0.02426	0.10291	0.00144	1705	42	6	869	1.15	3.60062	0.03756	0.11336	0.00447	1551	18
1d _{/02} +	b corrected 20t	²⁵⁸ U a	ge used														ĺ

Chapter 4: Geochronology

ICP-MS.

Label	U (ppm)	Th/U	$^{238}U/^{206}Pb$	$\pm 1 \sigma$	$^{207} Pb/^{206} Pb$	$\pm 1 \sigma$	²⁰⁷ Pb/ ²⁰⁶ Pb Age (Ma)	$\pm 1 \sigma$	Label	U (ppm)	Th/U	$^{238}U/^{206}Pb$	$\pm 1 \sigma$	$^{207} Pb/^{206} Pb$	$\pm 1 \sigma$	²⁰⁷ Pb/ ²⁰⁶ Pb Age (Ma)	$\pm 1 \sigma$
Sample 10	172 cont. 871	1.17	3.38377	0.08569	0.14061	0.00782	1594	42	Sample 34 [°]	264 cont. 1011	0.33	3.34757	0:03030	0.10607	0.00207	1680	15
12	789	0.94	3.44792	0.02810	0.11088	0.00323	1623	14	35 [†]	231	1.66	2.91730	0.03002	0.12089	0.00277	1890	20
Sample:	264 (Robin	(pooq)							Sample	: 299 (Gum I	flats) [†]						
1	543	0.69	3.49765	0.02636	0.11173	0.00221	1828	36	-	243	0.51	3.17749	0.12306	0.10572	0.01235	1768	71
2	689	0.81	3.59838	0.03115	0.11066	0.00291	1810	48	7	281	0.52	3.15655	0.18044	0.10157	0.01655	1787	105
3	1001	1.06	3.72763	0.03729	0.10240	0.00267	1668	48	3	326	0.69	3.15224	0.12472	0.10861	0.01093	1776	72
4	942	0.68	3.82156	0.04405	0.09782	0.00251	1583	48	4	305	0.56	3.41636	0.15077	0.10489	0.01159	1649	73
5	576	1.06	3.66790	0.04348	0.10765	0.00330	1760	56	5	258	0.72	3.39601	0.08148	0.10222	0.00685	1664	41
9	1076	0.88	3.49178	0.03690	0.10345	0.00206	1687	37	9	368	0.65	3.38777	0.09773	0.10950	0.00895	1655	49
7	1336	0.50	3.32458	0.03832	0.10005	0.00206	1625	38	6	180	0.31	3.47504	0.14300	0.11721	0.01339	1601	68
8	309	0.57	3.23130	0.04903	0.10460	0.00333	1707	59	10	132	0.50	3.16112	0.13732	0.09654	0.01296	1795	80
6	80	1.01	2.42525	0.03838	0.14085	0.00475	2238	58	12	656	0.94	3.37208	0.08156	0.10219	0.00659	1675	41
12	281	0.52	2.17207	0.04862	0.16538	0.00460	2511	47									
13^{\dagger}	567	0.55	3.51020	0.07069	0.10114	0.00405	1645	74	Sample	: 300 (Gum]	Flats) [†]						
14^{\dagger}	425	0.39	3.55985	0.05578	0.10350	0.00403	1688	72	1	189	0.51	3.38082	0.10398	0.10027	0.00895	1675	53
15^{\dagger}	1033	0.61	3.46529	0.07569	0.10424	0.00395	1701	70	2	340	0.48	3.32990	0.08824	0.10536	0.00789	1690	46
16^{\dagger}	310	1.03	3.61637	0.05296	0.09761	0.00319	1579	61	Э	439	0.81	3.53499	0.08974	0.10434	0.00755	1596	41
17^{\dagger}	471	0.46	3.34630	0.03188	0.11583	0.00262	1662	16	9	233	0.58	3.43074	0.09496	0.10457	0.00892	1643	47
18^{\dagger}	525	0.27	2.16517	0.02274	0.18229	0.00319	2373	29	7	164	0.51	3.49340	0.12892	0.10862	0.01188	1607	61
19^{\dagger}	183	0.48	2.00459	0.02064	0.18315	0.00351	2580	33	8	262	0.47	3.27915	0.09905	0.09746	0.00780	1731	53
20^{\dagger}	622	0.73	3.41448	0.05158	0.11755	0.00321	1627	25	6	488	0.72	3.36752	0.08096	0.10868	0.00732	1665	41
21^{\dagger}	522	0.55	3.29980	0.05100	0.11479	0.00426	1687	27	10	484	0.66	3.31547	0.08072	0.09845	0.00694	1710	43
22^{\dagger}	545	0.85	3.51859	0.03058	0.11706	0.00260	1581	14	11	759	0.88	3.55586	0.25926	0.11906	0.02137	1561	116
23^{\dagger}	887	0.83	3.46150	0.07221	0.11078	0.00421	1618	34									
24^{\dagger}	270	1.70	2.86727	0.02780	0.12284	0.00262	1918	19	Sample	: 318 (Gum I	flats) [†]						
25^{\dagger}	577	0.52	3.34983	0.06589	0.11108	0.00427	1669	33	2	984	0.52	3.15251	0.10253	0.09801	0.00738	1797	59
27^{\dagger}	827	0.39	3.41125	0.04256	0.10635	0.00296	1649	21	ŝ	782	0.14	3.32110	0.17171	0.09596	0.01434	1712	90
28^{\dagger}	1067	0.90	3.44777	0.04220	0.11547	0.00306	1615	20	5	1103	0.06	3.52616	0.14308	0.10262	0.01090	1604	99
29^{\dagger}	603	0.82	3.29492	0.03473	0.11183	0.00249	1695	18	9	995	0.03	3.47630	0.12181	0.09701	0.00865	1636	57
30^{\dagger}	759	0.42	3.47160	0.04808	0.10679	0.00332	1620	23	6	1453	0.00	3.60127	0.15747	0.09618	0.01198	1582	70
31^{\dagger}	280	0.43	2.08527	0.02060	0.17394	0.00312	2500	29	11	1296	0.00	3.47211	0.09984	0.09771	0.00894	1636	48
32^{\dagger}	743	0.78	3.46577	0.03628	0.10436	0.00202	1627	17	12	170	0.86	3.43003	0.18420	0.10515	0.01576	1643	90
33^{\dagger}	943	0.94	3.46982	0.04491	0.10855	0.00203	1618	21									
$+^{20/\text{Pp}}$	corrected 201	Pb/ ²³⁸ U a	ge used														

Table 4.1 cont.

Label	U (ppm)	Th/U	²³⁸ U/ ²⁰⁶ Pb	$\pm 1 \sigma$	²⁰⁷ Pb/ ²⁰⁶ Pb	$\pm 1 \sigma$	²⁰⁷ Pb/ ²⁰⁶ Pb Age (Ma)	$\pm 1 \sigma$
Sample	: 135 (Mount	Hogan Graı	nite)					
1	121	0.60	3.24252	0.07732	0.10131	0.00390	1648	71
3	594	0.92	3.92466	0.10001	0.10208	0.00372	1662	67
6	451	0.51	3.84811	0.15532	0.09760	0.00490	1579	94
7	584	0.68	3.61199	0.06022	0.09428	0.00233	1514	47
8	849	1.29	3.95138	0.23402	0.10420	0.00608	1700	107
9	1575	0.04	3.53854	0.06760	0.09531	0.00244	1534	48
11	429	1.40	3.85101	0.04808	0.09650	0.00252	1557	49
Sample	208 (Ropewa	alk Granite)						
1	513	0.35	3.52801	0.02143	0.09619	0.00188	1552	37
3	331	0.36	3.58848	0.02533	0.09729	0.00186	1573	36
5	550	0.45	3.57387	0.02195	0.09829	0.00202	1592	38
7	693	0.40	3.73633	0.04530	0.09579	0.00405	1544	80
11	469	0.46	3.57227	0.02370	0.09503	0.00206	1529	41
12	218	0.74	3.59033	0.03718	0.09498	0.00267	1528	53
Sample	· 209 (Ronew	alk Granite)						
2	796	0 90	3 72600	0 02379	0.09557	0.00179	1539	35
3	431	1.12	3 59818	0.02415	0.09401	0.00189	1508	38
4	717	1.12	3 79178	0.02233	0.09362	0.00179	1500	36
5	721	1.17	3.70523	0.02290	0.09475	0.00171	1523	34
6	1509	1.70	3.74483	0.02015	0.09446	0.00147	1517	29
8	265	0.89	3.72748	0.03162	0.09407	0.00235	1510	47
10	1063	1.51	3.65562	0.03688	0.09632	0.00291	1554	57
11	351	1.53	3.80165	0.05440	0.09717	0.00454	1571	88
12	399	1.02	3.73943	0.06756	0.09945	0.00436	1614	82
Sample	· 210 (Ronew	alk Granite)						
1	644	0 97	3 57072	0.03651	0.09872	0.00263	1600	50
2	260	0.70	3 63488	0.03318	0.09372	0.00234	1502	47
3	275	1 15	3 49572	0.02400	0.10100	0.00229	1643	42
4	390	0.67	3 60791	0.03002	0.09325	0.00204	1493	41
5	285	2.12	3.72676	0.03045	0.09421	0.00241	1512	48
6	309	0.46	3.54898	0.02856	0.09396	0.00232	1507	47
7	489	0.20	3.77140	0.02887	0.09337	0.00212	1495	43
8	531	0.79	3.74129	0.04016	0.09911	0.00228	1607	43
9	530	0.80	3.61350	0.02486	0.09631	0.00209	1554	41
10	314	2.14	3.68651	0.03227	0.09689	0.00257	1565	50
11	531	0.76	3.72087	0.03112	0.09479	0.00215	1524	43
12	195	1.04	3.68916	0.03940	0.09383	0.00271	1505	55

 Table 4.2: ²⁰⁴Pb-corrected U-Pb isotope data for zircons from granitic samples analysed by LA-ICPMS.

4.4.2.4 Source of Pb-loss and common-Pb in igneous zircons

The presence of significant radiogenic Pb-loss coupled with possible low-T Pb-loss due to alteration or chemical weathering, have contributed to the partial resetting of isotopic compositions in some analyses. This has resulted in the scattering of individual isotopic ages within most igneous samples, and large errors for some individual analyses (Table 4.1). Radiogenic Pb-loss was most likely caused by a combination of radiation damage to the crystal lattice, and to a lesser extent partial

recrystallisation due to metamorphism. Pb-loss due to radiation damage of the zircon crystal lattice (metamictization) is caused by the α -decay of ²³⁸U and fission products (Woodhead *et al.*, 1991; Mezger & Krogstad, 1997). Significantly, the likelihood of radiation damage to the crystal lattice will increase with age, U-concentration, and the length of time below the zircons annealing temperature of about 600-650 °C (Mezger & Krogstad, 1997). Recrystallisation associated with post-depositional orogenesis may have also affected the crystal structure of the igneous zircons (Hoskin & Black, 2000). However, due to the low grade of metamorphism in the study area, metamorphic recrystallisation was unlikely to have been the factor controlling radiogenic Pb-loss in these samples. Regardless of the cause, zircons containing appreciable Pb-loss appear discordant on inverse-concordia diagrams, and will provide an erroneously young age. Consequently, individual isotopic analyses that showed >10% discordance were omitted from the calculated ages, and are not shown here. A full list of analyses, including those analyses considered too discordant for use, is given in Appendix 2.

Common-Pb appears to be a significant problem affecting the igneous zircons collected for this study, and it is unclear whether the problem is caused totally by inheritance and inclusions in individual zircon grains, or by an analytical issue related to Pb being retained in the analytical system between analyses. In the former case, inclusions within small zircon grains similar to those analysed in this study are difficult to avoid using the laser ablation method, as the laser ablates though a significant part of the zircon. However, areas containing inclusions of common-Pb can be avoided during data reduction, as areas free of contamination can be preferentially selected using time-resolved spectra.

The problem of common-Pb is harder to deal with when the contamination comes from within the analytical system itself. This problem was encountered during the second analytical run and required correction that impacted on the accuracy of the calculated ages, resulting in the use of ²⁰⁶Pb/²³⁸U ages instead of preferred ²⁰⁷Pb/²⁰⁶Pb ages (S. Meffre, pers. comm., 2005). As a result, samples analysed over the course of two sessions incorporated these errors into the weighted mean age calculations, resulting in the overcompensation of analytical errors for some crystallisation ages (indicated by MSWD values much less than unity). As I was unable to further correct for this problem, the errors are incorporated into the calculated ages of this study.

However, meaningful results, albeit with slightly greater error bars, were still obtained via careful attention to each spectra for analysed zircon.

4.4.3 EMPA analysis of monazite

As monazite occurs as an accessory phase in many igneous and metamorphic rocks, and contains high amounts of Th and U coupled with low common-Pb, it has become a useful mineral for U-Th-Pb geochronology. Numerous studies over the past decade have used electron microprobe U-Th-Pb dating of monazite as a substitute for techniques such as LA-ICPMS and SHRIMP, as it does not involve the separation of grains from their host rock, and thus can provide ages with some textural context (e.g. Montel *et al.*, 1996, 2000; Cocherie *et al.*, 1998; Williams *et al.*, 1999; Shaw *et al.*, 2001; Williams & Jercinovic, 2002; Kelsey *et al.*, 2003; Dahl *et al.*, 2005; Berry *et al.*, 2005).

One of the key benefits of the EMPA technique for dating monazites is that the relatively small beam size used (commonly 1-3 μ m) allows numerous spot analyses to be done on individual monazite grains of less than 10-20 μ m in size, a population not typically sampled by other techniques. This technique also allows for the analysis of thin sections containing relatively few monazite grains (generally four or greater). By combining EMPA monazite-dated grains with back-scattered electron (BSE) images of the surrounding metamorphic fabric, ages can then be tied to specific structural events and/or mineral assemblages, making this an important tool for deciphering the metamorphic and structural history of igneous and metamorphic terranes.

4.4.3.1 Sample preparation

Spot analyses were obtained from monazite grains in thin section, from metasedimentary schists of the lower Etheridge Group that were collected from areas that had experienced amphibolite facies metamorphism (Fig. 4.1). Samples were prepared as polished thin sections before being coated with graphite. Individual monazite grains were identified and their positions recorded for automated runs by BSE imaging using the FEI Quanta 600 SEM housed at the University of Tasmania Central Science Laboratory. BSE images of the areas immediately surrounding

monazite grains were collected for three samples, in order to relate individual monazites ages to particular structural fabrics in each of the samples.

4.4.3.2 Electron microprobe operating parameters

Samples were analysed using the Cameca SX-100 EMP at the University of Tasmania Central Science Laboratory under the supervision of Mr. David Steele and Dr. Karsten Gömann. Five wavelength dispersive spectrometers were used for each analysis, with a configuration involving one PET, one Large-LIF, one TAP, and two Large-PET crystals. Values for Si, Al, K, Fe, Ca, Y, Th, U, Pb, P and 11 rare earth elements (REEs) were measured on each spot to help facilitate matrix corrections using the PAP method (Pouchou & Pichoir, 1984). Standard operating procedure for the analysis of monazite grains involved an accelerating voltage of 20 kV, 2-3 µm beam diameter, and a 100 nA beam current. Total counting times of 180 s (Pb), 40 s (Th) and 150 s (U) were used. X-ray lines on the EMP were calibrated using standard synthetic oxides, glasses and minerals. In particular, Huttonite (71.59 % Th), UO₂ (88.15 % U), and K227 Pb-glass (73.99 % Pb) were used for Th, U and Pb respectively. The EMP calibration was regularly checked using the monazite standard RGL04B (1563 \pm 3 Ma; Rubatto *et al.*, 2001). Analyses were obtained over the course of two sessions in automated runs. Automation and data collection was performed using software supplied by Cameca.

4.4.3.3 Data reduction and presentation

Age calculations based on measured U, Th and Pb concentrations (ppm) are used to iteratively solve the age equation of Suzuki *et al.* (1991), assuming that there is no uncertainty on the decay constants:

$$0 = \left[a Th \left(e^{\lambda 232Th t} - 1 \right) + b U \left(e^{\lambda 238U t} - 1 \right) + c U \left(e^{\lambda 235U t} - 1 \right) - Pb \right]^2$$
(4.1)

where the values a, b, and c are 'molecular constants' with a = 208/232, $b = (0.9928 \times 206)/238.04$, and $c = (0.0072 \times 207)/238.04$, λ^{x} are decay constants of ²³²Th, ²³⁸U and ²³⁵U, t = the age of the sample in Ma, and Th, U, Pb are the measured concentrations in ppm. By applying equation 4.1 to each analysis, an age can be obtained for each

spot analysed. Minimum age uncertainties were calculated for each analysed spot, derived by the propagation of X-ray counting errors through the age equation, with analytical uncertainties (1σ) being taken as equal to the square root of the peak plus background counts. Individual analyses showing signs of chemical contamination (e.g. K > 0.02 wt%, P > 13.0 wt%, total wt% < 97.0) were omitted from age calculations. Individual spot analyses are listed in Table 4.5, with errors quoted at the 1σ level. Errors in sample weighted mean age calculations are quoted as 95% confidence limits (Table 4.6). Weighted mean ages and cumulative probability diagrams were calculated using Isoplot version 3.0 (Ludwig, 2003). Discrete age populations within samples were identified using a similar method to that used for detrital zircon populations (Section 4.4.2.3). Monazite analyses with similar calculated ages were first identified using the cumulative probability diagrams in Figure 4.15. Approximate ages were taken from discrete peaks, and analyses lying within the range of each peak were then selected for weighted mean age calculations. In determining the weighted mean age, the youngest and/or oldest analysis in each group was omitted until the calculated age was statistically acceptable. As a result, more than one acceptable weighted mean age was calculated for some samples, which may reflect the presence of a number of discrete thermal events within the Etheridge Group.

4.5 LA-ICPMS U-Pb zircon dating results

Complete individual isotopic analyses for zircons from each sample are listed in Appendix 2, and are summarised in Tables 4.1, 4.2, and 4.4. A summary of the weighted mean ages calculated from each mafic and felsic sample are listed in Table 4.3. An initial 'test run' analysis on CMD samples 6 and 8 appears to indicate a relatively simple age population for the mafic rocks. However, the combined affects of common-Pb and Pb-loss appeared to have had a variable response on the zircons and their calculated ages. Subsequent analyses were then used to confirm initial ages and determine whether a more complex age distribution exists, as well as to determine the pattern of detrital zircon ages in the surrounding metasedimentary package, and to test the accuracy of the LA-ICPMS method on samples of various Palaeoproterozoic granitoids in the Etheridge Group against established ages.

U-Pb isotopic analyses were conducted on zircons from a total of 9 mafic, 4 granitic, and 2 metasedimentary samples, from numerous locations within the lower-grade area of the Etheridge Group (Fig. 4.1). Weighted mean ages and associated errors for individual samples of the mafic rocks are relatively large compared to other established dating methods due to the small number of zircons obtained from each mafic sample, coupled with the significant open-system processes that affected the chemical composition of many grains. Errors in weighted mean age calculations are quoted as 95 % confidence limits, although analytical errors for individual analyses in data tables and in the Tera-Wasserburg concordia diagrams (Tera & Wasserburg, 1972) are at the 1σ level.

4.5.1 Mafic rocks

4.5.1.1 Sample 6 (CMD – North Ortona)

Zircons from this sample were collected from a late-stage felsic differentiate of a metadolerite sill near Ortona station, and are all of igneous origin. The zircons are moderately uniform in size and elongation, averaging \sim 130 x 60 µm. Most grains are euhedral, and generally characterised by first-order prismatic faces. Terminations are either sharp on grain edges and broken surfaces, or are only very slightly rounded. Individual grains are morphologically simple, consisting of only a primary igneous growth phase, and are either poorly or strongly luminescent in CL images. The relatively low metamorphic grade (greenschist facies) has precluded the growth of metamorphic rims.

Only 12 zircon grains were recovered from this sample. Of these, eight were essentially free of common-Pb and Pb-loss. These grains have a 204 Pb-corrected 207 Pb/ 206 Pb weighted mean age of 1645 ± 22 Ma, free of significant scatter (MSWD = 0.75, probability of equivalence = 0.63; Fig. 4.3). This age is most likely associated with crystallisation of the late-stage felsic differentiate, and represents the emplacement age for the metadolerite sill.



Figure 4.3: (a) Tera-Wasserburg U-Pb concordia plot, (b) cumulative probability plot, and (c) weighted average plot for zircon analyses from sample 6. Data points in (a) represent analyses used in the determination of weighted mean ages. All error bars are 1σ.

4.5.1.2 Sample 8 (CMD – North Ortona)

Zircons from sample 8 are morphologically similar to those in sample 6, and were also collected from a late-stage felsic differentiate within a metadolerite sill near Ortona station. Individual grains have weak cathodoluminescence, and contain no major rims or inclusions. Grains are relatively uniform in size and shape, but are generally smaller than those of sample 6 (averaging $\sim 100 \times 50 \mu m$) and possess either prismatic or pyramidal faces.

The zircons of sample 8 define a single age distribution (Fig. 4.4b). The 8 coherent zircon analyses from this sample define a 204 Pb-corrected 207 Pb/ 206 Pb weighted mean age of 1637 ± 21 Ma (MSWD = 0.28, probability of equivalence = 0.96; Fig. 4.4c). As with sample 6, the weighted mean age for sample 8 is interpreted as the crystallisation age of the felsic differentiate within the metadolerite sill, and also represents the emplacement age of the metadolerite.



Figure 4.4: (a) Tera-Wasserburg U-Pb concordia plot, (b) cumulative probability plot, and (c) weighted average plot for zircon analyses from sample 8. All error bars are 1σ.

4.5.1.3 Sample 159 (CMD – South Ortona)

Zircons from sample 159 were collected from a felsic differentiate within a metadolerite sill south of Ortona station (Fig. 4.1). All separated zircons have igneous morphologies and are generally short with pyramidal to prismatic and lesser rounded edges (averaging $\sim 100 \times 60 \mu m$). As with samples 6 and 8, metamorphic or recrystallization rims were not detected on the igneous zircons.

Some 24 zircons from sample 159 were analysed in two separate sessions. Although the isotopic data appear to form a single peak in the cumulative probability diagram (Fig. 4.5b), the isotopic ages for all analyses are significantly scattered and several have been affected by a combination of Pb-loss and common-Pb. By omitting two analyses with individual isotopic ages below 1500 Ma, one anomalous analysis at 1703 Ma with an extremely high associated error, one analysis with an isotopic age of 3376 ± 201 Ma, and several discordant analyses, the remaining 14 analyses yield an age of 1713 ± 32 Ma (MSWD = 0.25, probability of equivalence = 0.997).

The calculated age for this sample is significantly older than the previously suggested upper age limit for the onset of sedimentation in the eastern Etheridge Group (ca. 1700 Ma; Black *et al.*, 2005). However, it is within error of the age of another sample from the same sill (see 161 below) and is regarded here as the crystallisation age.



Figure 4.5: (a) Tera-Wasserburg U-Pb concordia plot, (b) cumulative probability plot, and (c) weighted average plot for zircon analyses from sample 159. All error bars are 1σ.

4.5.1.4 Sample 161 (CMD – South Ortona)

Sample 161 was collected 10 m east of sample 159 within the same metadolerite sill and, as such, contains zircons that are morphologically indistinguishable from sample 159. The majority of separated zircons are elongate with prismatic and lesser rounded edges (averaging up to \sim 120 x 80 μ m). Metamorphic zircons occurring either as discrete grains or rims were not detected.

Some 30 zircons were analysed in two separate sessions for sample 161. After omitting analyses containing significant Pb-loss, the remaining 24 analyses form a broad single peak in the cumulative probability diagram (Fig. 4.6b). As with sample 159, the isotopic ages for all analyses are significantly scattered and several have been

affected by either minor Pb-loss or common-Pb. The 24 acceptable analyses combine to give a weighted mean age of 1719 ± 27 Ma (MSWD = 1.12, probability of equivalence = 0.31; Fig. 4.6c).



Figure 4.6: (a) Tera-Wasserburg U-Pb concordia plot, (b) cumulative probability plot, and (c) weighted average plot for zircon analyses from sample 161. All error bars are 1σ.

The correlation in calculated ages for samples 159 and 161 indicate that they represent a real crystallisation event, despite the presence of significant common-Pb and Pb-loss in some analyses affecting the calculated ages. The 1719 ± 27 Ma age is within error of the oldest magmatic zircon ages obtained for the Etheridge Group (Black *et al.*, 2005). Consequently, these zircons may be xenocrystal and derived from the same leucogneiss source as the ca. 1710-1700 Ma zircons analysed by Black *et al.* (2005), or they may represent an earlier phase of emplacement and crystallisation of mafic rocks. If the former is the case, then leucogneiss similar to that found in the Einasleigh Metamorphics must underlie the Etheridge Group metasediments in this area, and mafic material forming the CMD sills would have passed through the leucogneiss prior to emplacement. However, given that the analyses of the euhedral (igneous) zircon grains from both samples 159 and 161 form relatively tight isotopic populations, it is highly unlikely that they were inherited.

4.5.1.5 Sample 172 (CMD – South Ortona)

Zircons from sample 172 were collected from a late-stage felsic differentiate within a metadolerite sill separate but adjacent to the sill containing samples 159 and 161 (Fig. 4.1). Individual grains possess igneous morphologies and are short, with pyramidal to rounded edges (averaging $\sim 100 \times 50 \mu m$). Metamorphic or recrystallised rims were not detected on the zircons in this sample.

The diagram of isotopic age distribution for the zircons of sample 172 defines a single peak with a broad younger shoulder (Fig. 4.7b). Some 12 grains were analysed for this sample, however the oldest and youngest values were omitted from subsequent age calculations due to significant common-Pb contamination. Due to the presence of minor common-Pb contamination coupled with some Pb-loss in the remaining analyses, the use of ²⁰⁴Pb-corrected ²⁰⁷Pb/²⁰⁶Pb ages was not possible, so ²⁰⁷Pb-corrected ²⁰⁶Pb/²³⁸U ages were used instead. The seven analyses defining the dominant peak in Figure 4.7b have a ²⁰⁷Pb-corrected ²⁰⁶Pb/²³⁸U weighted mean age of 1628 ± 13 Ma (MSWD = 0.32, probability of equivalence = 0.930). Although the MSWD is significantly below unity, this value cannot be improved further by the omission or retention of more analyses. This age is younger than any previously dated sample for the Cobbold Metadolerite; however the use of the ²⁰⁷Pb-corrected ²⁰⁶Pb/²³⁸U weighted mean age in addition to the presence of minor Pb-loss in some analyses could indicate that this value is a minimum crystallisation age for the dolerite.

Three zircons analysed from this sample returned ages younger than 1600 Ma, and although two grains suffered minor Pb-loss their resultant analyses are acceptable. The three grains combine to give a weighted mean age of 1561 ± 54 Ma; however they are significantly scattered (MSWD = 2.2). Although the ages of the younger grains are consistent with the established age for metamorphism in the Etheridge Group at ca. 1550 Ma, their clear igneous morphologies suggest they are unlikely to be metamorphic in origin. It is possible that these grains experienced complete resetting of their isotopic ages due to metamorphism, as similar ages have been

obtained from recrystallised igneous zircons in a meta-granitoid from the eastern Etheridge Group (Hoskin & Black, 2000). However, the low-grade of metamorphism experienced by sample 172 makes this unlikely.



Figure 4.7: (a) Tera-Wasserburg U-Pb concordia plot, (b) cumulative probability plot, and (c) weighted average plot for zircon analyses from sample 172. All error bars are 1σ.

4.5.1.6 Sample 264 (DHM – Robinhood)

Sample 264 was collected from an amphibolite-facies outcrop of a Dead Horse Metabasalt flow SW of Robinhood station (Fig. 4.1). The sample contained a number of coarse-grained felsic segregations that appear consistent with the fractionation of a felsic component prior to pervasive metamorphism and re-crystallisation. Zircons collected were all of igneous origin. Small unbroken grains are moderately uniform in size and elongation, averaging ~100 x 50 μ m. A number of larger, broken grains were also found, and vary in dimension up to ~110 x 80 μ m. All grains are euhedral and generally characterised by first-order pyramidal, or lesser prismatic faces. Terminations are either sharp on grain edges and broken surfaces, or are only very slightly rounded (smaller grains) or rounded (larger grains). Individual grains are

morphologically simple, consisting of only a primary igneous growth phase; however larger grains contain some form of compositional zoning. Although the metamorphic grade experienced by sample 264 is significantly higher compared to other mafic samples in this study, no significant metamorphic rims or zoning were detected on the analysed zircons.

Some 35 analyses were conducted on the separated zircons from sample 264 over two separate sessions. Three analyses were omitted due to a combination of significant Pb-loss and common-Pb. As with sample 172, the presence of significant common-Pb contamination coupled with minor Pb-loss in several grains prevented the use of ²⁰⁴Pb corrected ²⁰⁷Pb/²⁰⁶Pb ages, so ²⁰⁷Pb-corrected ²⁰⁶Pb/²³⁸U ages were instead used for selected analyses (Table 4.1). Sample 264 is unique among the amphibolite samples analysed in this study, as it appears to contain a spectrum of isotopic ages that suggest a significant inherited component in the sample. There is a large spread in the isotopic age distribution (Fig. 4.8b), with the majority of analyses scattered between 1600 and 2600 Ma. Based on individual isotopic ages, the analyses falling between 1580 and 1710 Ma can be separated into a number of relatively distinct groups. However, only the most statistically acceptable groupings were used in order to determine weighted mean ages.

The youngest group (Group 1) contains 7 analyses that yield a weighted mean age of 1602 ± 17 Ma (MSWD = 0.74, probability of equivalence = 0.62). This age is within error of the crystallisation age determined for the younger phase of mafic intrusions represented by sample 172. However, the large isotopic errors for some Group 1 analyses and similarities in Th/U values to the Group 2 analyses may indicate that these zircons were partially reset by later metamorphism (Hoskin & Black, 2000). The next oldest group (Group 2) contains 15 analyses that yield a weighted mean age of 1663 ± 13 Ma (MSWD = 1.05, probability of equivalence = 0.40). As this group contains the majority of concordant analyses, the 1663 ± 13 Ma age represents the best estimate for the crystallisation age of this rock. Combining the Group 1 and 2 analyses did not return a statistically acceptable grouping.

The remaining analyses are separated into three groups based on isotopic ages. The two younger groups (Groups 3 and 4) contain 3 and 2 analyses respectively, with weighted mean ages of 1809 ± 50 Ma (MSWD = 0.52, probability of equivalence = 0.60) for Group 3, and 1905 ± 27 Ma (MSWD = 1.02, probability of equivalence =

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0.31) for Group 4. The two groups most likely represent inherited sources, and may reflect the peperitic incorporation of Barramundi-aged material from adjacent metasediments into the magma. The oldest group (Group 5) contains 3 analyses that combine for a weighted mean age of 2531 ± 110 Ma (MSWD = 1.7, probability of equivalence = 0.18). This age may represent zircons inherited from a much older (pre-Barramundi?) source that were subsequently incorporated into the ca. 1660 Ma phase of magmatism. The significance of this age is discussed in more detail below. A further three analyses returned ages between 2200 and 2700 Ma, but do not form a statistically acceptable grouping.



Figure 4.8: (a) Tera-Wasserburg U-Pb concordia plot, (b) cumulative probability plot, (c) weighted average plot for Group 1 and 2 analyses (see text for explanation), and (d) Th/U versus age diagram for analyses from sample 264. All error bars are 1σ.

The age groups outlined above are also distinguished by their Th/U characteristics (Fig. 4.8d). The younger groups (Groups 1, 2, and 3) have fairly distributed Th/U values (0.33-1.06, average = 0.71), whereas Group 4 analyses have Th/U values of 1.66 and 1.70 (average = 1.68). Th/U values for Group 5 also form a distinct group, ranging from 0.43 to 0.52 (average = 0.48). The difference in Th/U values, as well as the range in values for the younger groups, is partially a consequence of the younger

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zircons possessing higher U and Th relative to Group 4 and 5 (averaging 770 ppm U and 541 ppm Th for Groups 1 and 2, compared to averages of 248 ppm U and 119 ppm Th for Group 5). Zircon morphology may also be linked to the distribution in isotopic ages in this sample, as analyses conducted on larger grains returned ages above ca. 1700 Ma.

It is possible that the spectrum of isotopic ages obtained for sample 264 may indicate that eruption of the metabasalt occurred sometime around the crystallisation of the Group 1 zircons (ca. 1600 Ma). However, the weighted mean age obtained for the Group 1 zircons is questionable, as in addition to the possibility of some isotopic resetting occurring within the grains, the majority of analyses contained some common-Pb. Due to these effects, the use of 207 Pb-corrected 206 Pb/ 238 U ages rather than the preferred 207 Pb/ 206 Pb ages was required. Consequently, I conclude that the Group 2 age of 1663 ± 13 Ma is the best estimate of the crystallisation age of this rock.

4.5.1.7 Sample 299 (CMD – Gum Flats)

Sample 299 was collected from an outcrop of the Cobbold Metadolerite south of Gum Flats station (Fig. 4.1). Initial attempts to separate zircons from the sample using the milling method described above did not return an adequate number of grains. Therefore the sample was prepared for analysis as a 1 inch polished round mount, with all zircons analysed *in situ*. Individual zircon morphologies were difficult to distinguish due to the impingement of other minerals; however where visible, morphologies were typical of igneous rocks.

Eleven zircons were analysed for sample 299, with one discordant analysis omitted due to Pb-loss. 207 Pb-corrected 206 Pb/ 238 U isotopic ages were again used in preference to 207 Pb/ 206 Pb ages due to common-Pb contamination (Table 4.1). The isotopic data form a broad single peak in the cumulative probability diagram (Fig. 4.9b), reflecting the higher analytical error associated with in situ analysis (Table 4.1). However, despite this variation the data combine to form a statistically acceptable weighted mean age from 9 grains of 1687 ± 38 Ma (MSWD = 1.04, probability of equivalence = 0.40; Fig. 4.9c). This age is similar to the crystallisation of the Dead Horse Metabasalt unit and associated metadolerite sills and supports correlation of the Cobbold Metadolerite with this unit.



Figure 4.9: (a) Tera-Wasserburg U-Pb concordia plot, (b) cumulative probability plot, and (c) weighted average plot for zircon analyses from sample 299. All error bars are 1σ.

4.5.1.8 Sample 300 (CMD – Gum Flats)

Sample 300 was collected from an outcrop of the Cobbold Metadolerite close to sample 299 (Fig. 4.1). As with sample 299, this sample was prepared for analysis as a 1 inch round mount, and all zircons were analysed *in situ*. Problems encountered in sample 299 in distinguishing individual zircon morphologies were also encountered in this sample, but where visible, morphologies suggested an igneous origin.

12 zircons were analysed for sample 300; however three discordant analyses were omitted (Fig. 4.10). 206 Pb/ 238 U ages were used in preference to 207 Pb/ 206 Pb ages due to minor common-Pb contamination in some analyses. The analyses from 9 grains combine to form a statistically acceptable weighted mean age of 1662 ± 32 Ma (MSWD = 0.95, probability of equivalence = 0.48). This age corresponds well with the crystallisation ages sample 299 and Dead Horse Metabasalt listed above.



Figure 4.10: (a) Tera-Wasserburg U-Pb concordia plot, (b) cumulative probability plot, and (c) weighted average plot for zircon analyses from sample 300. All error bars are 1σ.

4.5.1.9 Sample 318 (Einasleigh Metamorphics – Gum Flats)

Sample 318 was collected from an amphibolite outcrop in the Einasleigh Metamorphics, east of Percyvale station (Fig. 4.1). As with samples 299 and 300, this sample was prepared for analysis as a 1 inch round mount, and all zircons were analysed *in situ*.

12 zircons were analysed for sample 318; however five analyses were omitted due to significant Pb-loss (Fig. 4.11). Due to the presence of common-Pb throughout the analysed zircons, 207 Pb-corrected 206 Pb/ 238 U ages were again used in preference to 207 Pb/ 206 Pb ages. The 7 acceptable analyses form a statistically acceptable weighted mean age of 1658 ± 69 Ma (MSWD = 1.4, probability of equivalence = 0.23), which correlates with the inferred crystallisation age for the Dead Horse Metabasalt and Cobbold Metadolerite samples listed above.



Figure 4.11: (a) Tera-Wasserburg U-Pb concordia plot, (b) cumulative probability plot, and (c) weighted average plot for zircon analyses from sample 318. All error bars are 1σ.

4.5.2 Summary of LA-ICPMS zircon ages from mafic samples

The igneous zircons from mafic rocks of the lower Etheridge Group define a number of key crystallisation ages (Table 4.3). The most significant of these relates to the eruption of the Dead Horse Metabasalt unit and associated feeder dykes and sills of the Cobbold Metadolerite. Four samples (samples 264, 299, 300 and 318), including a sample of Dead Horse Metabasalt amphibolite, returned magmatic ages of 1687 ± 38 , 1663 ± 13 , 1662 ± 32 and 1658 ± 69 Ma (Table 4.3). These ages combine to produce an approximate crystallisation age for the Dead Horse Metabasalt and associated metadolerite of ca. 1665 Ma. Two samples collected from Cobbold Metadolerite sills near 'Ortona' station returned crystallisation ages of 1645 ± 22 and 1637 ± 21 Ma (Table 4.3). These ages do not appear to be associated with the age of eruption of the Dead Horse Metabasalt, and may instead suggest an extended period of magmatism in the Etheridge Group. However the errors associated with the calculated ages do not

Stratigraphic unit	Sample	Age (Ma) $+ 95\%^{1}$	No. of analyses ²	MSWD	Prob. of Equivalence	Comment
Mafic samples		= > 0 > 0	unurjoes		1	
Cobbold Metadolerite	6	1645 ± 22	8	0.75	0.63	Magmatic age
Cobbold Metadolerite	8	1637 ± 21	8	0.28	0.96	Magmatic age
Cobbold Metadolerite	159	1713 ± 32	14	0.25	0.997	Magmatic age?
Cobbold Metadolerite	161	1719 ± 27	24	1.12	0.31	Magmatic age?
Cobbold Metadolerite	172	1628 ± 13	7	0.32	0.93	Magmatic age
		1561 ± 54	3	2.2	0.11	Metamorphic age?
Dead Horse Metabasalt	264	1663 ± 13	15	1.05	0.4	Magmatic age
		1602 ± 17	7	0.74	0.62	Partially re-set or
						metamorphic age
		1809 ± 50	3	0.52	0.6	Inherited age
		1905 ± 27	2	1.02	0.31	Inherited age
		2524 ± 64	3	1.7	0.18	Inherited age
Cobbold Metadolerite	299	1687 ± 38	9	1.04	0.4	Magmatic age
Cobbold Metadolerite	300	1662 ± 32	9	0.95	0.48	Magmatic age
Einasleigh Metamorphics	318	1658 ± 69	7	1.4	0.23	Magmatic age
Granitic samples						
Mt Hogan Granite	135	1565 ± 46	6	1.02	0.4	1549 ± 25 Ma
						(Black & Withnall, 1993)
Ropewalk Granite	208	1558 ± 34	6	0.37	0.87	Magmatic age
	209	1524 ± 27	9	0.34	0.95	Magmatic age
	210	1544 ± 33	12	1.30	0.21	Magmatic age
		1539 ± 16	27	0.83	0.71	Combined 208, 209, & 210 age 1550 +7/-5 Ma (Black & McCulloch, 1990)

provide sufficient resolution to determine if these ages definitively date a second discrete pulse of mafic activity.

¹ Pooled age from *n* analyses calculated using *Isoplot* version 3.0 (Ludwig, 2003)

² Number of analyses in pooled age

Table 4.3: Summary of weighted mean ages calculated from individual LA-ICPMS igneouszircon analyses in Tables 4.1 and 4.2.

Two Cobbold Metadolerite samples from the same sill (samples 159 and 161) returned older crystallisation ages of 1713 ± 32 and 1719 ± 27 Ma (Table 4.3). These ages are problematic, as they are older than the inferred maximum depositional age of the Etheridge Group (ca. 1700 Ma; Black *et al.*, 2005). However, they are within error of the calculated age for the volcanic source of the Einasleigh Metamorphics leucogneiss (1706 \pm 6 Ma; Black *et al.*, 2005). Consequently, it is possible that the age of the leucogneiss represents a volcanic age that is not at the base of the sequence and that some parts of the lower Etheridge Group formed at or before ca. 1715 Ma.

Sample 172 returned a crystallisation age of 1628 ± 13 Ma (Table 4.3). It probably represents the last component of igneous activity within the basin before the onset of metamorphism. However, as noted above this age may be erroneously young, and since it is within error of the ca. 1640 Ma samples may relate to this pulse of magmatic activity within the Etheridge Group.

In addition to their magmatic ages, sample 264 contains zircons with isotopic ages that suggest inheritance from the host lower Etheridge Group metasediments. In addition, samples 172 and 264 contained zircon populations that have been reset to ages of 1561 ± 54 and 1602 ± 17 Ma respectively (Table 4.3), possibly due to post-depositional tectonothermal events. The significance of these ages, and the inherited ages in sample 264, are discussed in more detail below.

4.5.3 Granitoids

4.5.3.1 Sample 135 (Mt Hogan Granite – Gum Flats)

Sample 135 was collected from an outcrop of the Mt Hogan Granite located approximately 10 km southwest of Gum Flats (Fig. 4.1). Relatively few zircons were recovered from this sample, and they are moderately uniform in size and elongation, averaging \sim 150 x 70 µm. The majority of grains have prismatic or pyramidal terminations that are typical of igneous growth. Euhedral zoning is also evident in some larger grains.

The majority of zircons analysed had very discordant isotopic compositions (Fig. 4.12a) due to low-temperature exchange of Pb. All analyses affected in this way returned ages in excess of 1700 Ma, and were omitted before calculating the crystallisation age of the rock. In order to calculate a preferred age for the crystallisation of the granite, the concordant individual ages were combined and their mean age determined. The analyses containing the six youngest isotopic ages yield the most statistically acceptable weighted mean age of 1565 \pm 46 Ma (MSWD = 1.02, probability of equivalence = 0.40). This age is within error of the previously published age for the crystallisation of the Mt Hogan Granite (1549 \pm 25 Ma; Black & Withnall, 1993).

4.5.3.2 Samples 208, 209 and 210 (Ropewalk Granite - Robinhood)

Three samples were collected from outcrops of the Ropewalk Granite, south of Forsayth (Fig. 4.1). Zircons collected from these samples are mineralogically indistinguishable, and are of predominantly igneous origin. A range of zircon shapes is present across all three samples, from elongated grains (typically ~170 x 50 μ m) of igneous origin, to equant grains (~120 μ m) of possible metamorphic origin. The

majority of igneous grains have unmodified crystal faces, and very few possess rounded edges. Equant grains are generally well-rounded, and are less luminescent. As analysis was primarily focussed on determining the age of igneous crystallisation, grains of probable igneous origin were preferentially selected for analysis.



Figure 4.12: (a) Tera-Wasserburg U-Pb concordia plot, and (b) cumulative probability plot for zircon analyses from sample 135. All error bars are 1σ.

In sample 208, six grains were significantly discordant, and were subsequently omitted from the weighted mean age calculation. The remaining 6 analyses returned a 207 Pb/ 206 Pb weighted mean age of 1558 ± 34 Ma (MSWD = 0.37, probability of equivalence = 0.87). Three discordant analyses were also omitted from sample 209. The 9 acceptable analyses returned a 207 Pb/ 206 Pb weighted mean age of 1524 ± 27 Ma (MSWD = 0.34, probability of equivalence = 0.95). Although this age is younger than that calculated for other granitic bodies crystallising at around this time, it remains within error of the inferred ca. 1550 Ma age for Palaeoproterozoic felsic magmatism in the Etheridge Group (Black & McCulloch, 1990). In sample 210, all 12 analysed grains were free of significant Pb-loss and common-Pb (Fig. 4.13c), and returned a 207 Pb/ 206 Pb weighted mean age of 1544 ± 33 Ma (MSWD = 1.30, probability of equivalence = 0.21).

As the three samples are all assigned to the same granitic unit, I have combined the individual isotopic data in order to produce a more accurate age estimate for the Ropewalk Granite (Fig. 4.13d). Combining the three samples yields a weighted mean age from 27 analyses of 1539 ± 16 Ma (MSWD = 0.83, probability of equivalence = 0.71). As with the Mt Hogan Granite, this age is within error of the ca. 1550 Ma age for felsic magmatism in the Etheridge Group, and the 1550 +7/-5 Ma SHRIMP age calculated for the nearby Forsayth Granite (Black & McCulloch, 1990). Therefore, I
suggest that the ca. 1539 Ma calculated age represents the crystallisation age of the granite and provides some evidence that the Ropewalk Granite is the youngest Mesoproterozoic granitoid in the area.



Figure 4.13: Tera-Wasserburg U-Pb concordia diagrams for samples (a) 208, (b) 209, and (c) 210.
(d) Cumulative probability plot for analyses from all three samples. All error bars are 1σ.

4.5.4 Metasedimentary rocks

Prior to this study, little attention had been given to the provenance of zircons from metasedimentary rocks in the Etheridge Group. The work of Black *et al.* (2005) focussed on the U-Pb SHRIMP dating of zircons collected from high-grade leucogneiss of the Einasleigh Metamorphics in the area near Percyvale (Fig. 4.1). They found detrital zircon populations in a number of samples; however they could define no significant age groupings, and concluded the detrital grains were most likely derived from various sedimentary protoliths.

This study, involving two metasedimentary samples from lower-grade areas of the Daniel Creek Formation, represents the first attempt to directly determine the detrital zircon ages for the metasedimentary unit that directly underlies the Dead Horse

Metabasalt member. The resultant pattern of detrital ages is then compared to detrital populations found in sedimentary units surrounding the key mafic igneous units of the Eastern Mt Isa Succession and Broken Hill Block, in order to determine if any similarities in the detrital zircon patterns exist between the terranes.

Sample: Is4 (North Ortona) Sample: SSample:	Label	²³⁸ U/ ²⁰⁶ Pb	$\pm 1 \sigma$	²⁰⁷ Pb/ ²⁰⁶ Pb	$\pm 1 \sigma$	²⁰⁷ Pb/ ²⁰⁶ Pb Age (Ma)	$\pm 1 \sigma$	Label	²³⁸ U/ ²⁰⁶ Pb	$\pm 1 \; \sigma$	²⁰⁷ Pb/ ²⁰⁶ Pb	$\pm 1 \sigma$	²⁰⁷ Pb/ ²⁰⁶ Pb Age (Ma)	$\pm 1 \sigma$
7 301503 0.00276 2818 157 1 1.86501 0.01597 0.1937 0.0276 2811 23 10 1.99466 0.00372 0.16417 0.01093 2711 97 5 1.96781 0.0220 0.1281 0.00216 2277 32 15 1.92466 0.02707 0.16447 0.00193 2711 33 6 1.82846 0.0210 0.01412 0.00411 2773 35 18 2.4375 0.03448 0.01750 0.00456 2056 17 12 1.7776 0.01735 0.00280 2.887 38 2 2.3554 0.02246 0.17050 0.00403 2.558 40 14 3.2077 0.02830 0.0050 2.696 2.8 2 2.85767 0.07090 0.1731 0.00352 2.981 1.5 1.92504 0.0020 0.1641 0.0272 2.9242 4.6 2 2.85767 0.07090 0.1731 0.0	Samp	le: 184 (Nor	th Ortona))				Samp	le: 255 (Sou	th Head)				
10 199686 0.08728 0.0902 0.01637 2818 124 2 2.7281 0.02425 0.12426 0.00228 2026 32 15 192486 0.00178 0.18641 0.00374 2711 33 6 1.82846 0.02320 0.18321 0.00389 2.877 33 18 2.24737 0.02484 0.07160 0.00466 2067 4.8 1.64644 0.07168 0.00389 2.887 30 6.4 33 0.0133 0.00389 2.887 33 1.6 2.1777 0.0135 0.00366 2.896 32 0.0153 0.0036 2.886 33 1.6 2.46444 0.07025 0.1782 0.00141 2.722 72 2 1.94793 0.01785 0.18163 0.0039 2.253 1.70 2.21576 0.0284 0.1161 0.00232 1.910 5.3 3 2.2577 0.0796 0.1795 0.0039 2.253 1.70 2.2456 0.0284 <td>7</td> <td>3.01503</td> <td>0.05063</td> <td>0.11254</td> <td>0.00977</td> <td>1841</td> <td>157</td> <td>1</td> <td>1.86501</td> <td>0.01597</td> <td>0.19819</td> <td>0.00276</td> <td>2811</td> <td>23</td>	7	3.01503	0.05063	0.11254	0.00977	1841	157	1	1.86501	0.01597	0.19819	0.00276	2811	23
11 100243 0.01412 0.18647 0.00179 211 97 5 1.96781 0.02150 0.18643 0.00373 233 18 2.24737 0.03488 0.17307 0.00456 2097 13 6 1.82846 0.0230 0.19362 0.00133 0.00339 2587 38 19 2.13255 0.02304 0.17306 0.00353 2468 93 13 2.14776 0.01735 0.00360 2686 23 23 2.23554 0.02246 0.01700 0.00030 2558 40 14 3.0777 0.02135 0.1141 0.01031 8866 63 2 2.05373 0.02160 0.1751 0.00362 2668 33 16 2.46464 0.0225 0.1448 0.00321 2422 2.35277 0.0560 0.1751 0.00341 2507 2310 0.2656 0.1384 0.11751 0.0121 230 2.10470 0.01344 0.1679 0.0389<	10	1.99686	0.08782	0.19902	0.01637	2818	134	2	2.78281	0.02425	0.12481	0.00228	2026	32
15 192486 0.0278 0.18643 0.0074 2711 33 6 1.82846 0.0230 0.1301 0.0118 2773 35 18 2.41325 0.0379 0.1766 0.00456 2607 43 8 1.64084 0.01768 0.23173 0.00483 364 33 20 1.33982 0.0455 0.01701 0.00726 2606 175 12 1.77776 0.01735 0.00502 2609 24 2 0.02851 0.0211 0.17351 0.00502 2668 33 16 2.46464 0.0225 0.1428 0.0041 2222 29 32 2.5577 0.0560 0.15751 0.0050 2102 72 17 2.3150 0.0268 0.11619 0.0323 1910 33 3 2.16860 0.01641 0.1624 0.0077 245 38 22 19350 0.0168 0.1191 0.0323 1910 33 3 12	11	1.90243	0.04112	0.18647	0.01093	2711	97	5	1.96781	0.02195	0.18821	0.00361	2727	32
18 2.24737 0.0448 0.17267 0.00486 2.298 2.12700 0.02364 0.1730 0.00486 2.287 3.8 20 1.93982 0.04455 0.17760 0.00756 2066 1.75 1.2 1.64084 0.01756 0.20380 2.08666 2.88 2.3 21 2.26554 0.02246 0.10700 0.00032 2.588 0.4 1.4 3.0277 0.02238 0.00143 1.886 6.3 2 2.05573 0.0216 0.1751 0.0050 2.268 3.1 1.5 1.92504 0.02025 0.1428 0.00241 2.262 2.9 3 2.25277 0.05961 0.1781 0.00402 2.93 1.70 2.0 2.75826 0.0380 0.1161 0.00322 1.9410 0.02322 1.940 0.33 3 2.10470 0.0148 0.1667 0.0382 2.2527 0.0493 0.0031 2.752 2.9 1.9133 0.00213 2.33 1.4	15	1.92486	0.02078	0.18643	0.00374	2711	33	6	1.82846	0.02320	0.19362	0.00411	2773	35
19 2.13325 0.02490 0.07506 0.00456 2607 4.3 8 1.64084 0.01768 0.2138 0.00433 0.00433 0.02483 0.00433 2.04021 0.01735 0.02582 0.0035 2.68 13 2.04021 0.01734 0.01758 0.00350 2.69 2.3 24 2.08654 0.02140 0.17051 0.00535 2.528 4.0 1.4 3.20727 0.0233 0.11541 0.00403 2.782 7 2 2.8577 0.07090 0.11805 0.00503 2.012 7.2 1.7 2.2150 0.02695 0.15702 0.00422 2424 46 3 2.2577 0.05961 0.17541 0.00572 2.451 2.9 1.7 2.02150 0.0238 0.1013 1.7 3 2.10680 0.01761 0.16244 0.00217 2.252 2.6 2.89766 0.02746 0.1174 0.0021 2.33 1 2.10530 0.01234 0.1153 <td>18</td> <td>2.24737</td> <td>0.03448</td> <td>0.17267</td> <td>0.00486</td> <td>2395</td> <td>165</td> <td>7</td> <td>2.12700</td> <td>0.02364</td> <td>0.17301</td> <td>0.00389</td> <td>2587</td> <td>38</td>	18	2.24737	0.03448	0.17267	0.00486	2395	165	7	2.12700	0.02364	0.17301	0.00389	2587	38
20 1.93982 0.04455 0.19781 0.00734 0.01734 0.01734 0.00882 0.00366 2896 28 23 2.23554 0.02241 0.006353 2468 93 13 2.04021 0.01734 0.0734 0.00356 2609 24 24 2.08654 0.0216 0.17351 0.00403 2558 40 14 3.2077 0.05233 0.11541 0.00440 2582 2 2.85767 0.0709 0.12380 0.00503 2012 72 17 2.32150 0.02695 0.15702 0.00422 2424 46 33 2.5277 0.05961 0.17951 0.00430 2583 170 20 2.75826 0.03080 0.11801 0.00323 2714 2 1.04161 0.02312 1910 50 41 2.01535 0.02344 0.16244 0.00210 238 2.18651 0.01626 0.133 37 41 2.165461 0.02316 0.11	19	2.13325	0.02709	0.17506	0.00456	2607	43	8	1.64084	0.01768	0.23173	0.00483	3064	33
23 2.23554 0.00250 0.18004 0.00555 2468 93 13 2.04021 0.01734 0.00250 2600 24 24 2.08654 0.02246 0.17050 0.00430 2558 40 14 3.20727 0.05233 0.11541 0.00403 258 70 29 1.94793 0.01875 0.18163 0.00360 2292 34 15 2.46464 0.02025 0.15702 0.00422 244 46 33 2.25277 0.05961 0.17951 0.00840 2593 170 20 2.75826 0.03080 0.11691 0.00223 1910 50 36 2.10686 0.01741 0.16092 0.00357 2465 38 22 1.90356 0.01622 19118 0.00232 1910 31 31 1.92536 0.0334 0.16692 0.00357 2455 38 22 1.98466 0.0216 1.985 34 27 1.98456 0.01217	20	1.93982	0.04455	0.19781	0.00726	2506	175	12	1.77776	0.01735	0.20882	0.00366	2896	28
24 2.08654 0.0246 0.17005 0.00403 2558 40 14 3.2072 0.05333 0.00401 1886 62 26 2.05373 0.02016 0.17351 0.00362 2668 33 16 1.92544 0.02027 0.20840 0.00304 2722 79 32 2.85767 0.07009 0.12380 0.00503 2012 72 17 2.32150 0.02695 0.1428 0.00232 1910 50 33 2.2577 0.0586 0.1161 0.1624 0.00279 2481 29 21 2.86951 0.02858 0.1128 0.00231 215 21 4 2.01670 0.0216 1.6679 0.0038 232 2.91840 0.02188 0.1228 0.0217 200 33 43 1.92539 0.02355 0.1334 0.00215 1880 34 27 1.98405 0.01740 0.1799 0.0026 22.58 4 2.8384 0	23	2.23554	0.02951	0.18004	0.00535	2468	93	13	2.04021	0.01734	0.17528	0.00250	2609	24
26 205373 0.02016 0.17551 0.00366 2592 34 15 1.92504 0.02025 0.02840 0.00341 2782 70 29 1.94793 0.018163 0.00503 2012 72 17 2.32150 0.02025 0.14288 0.00421 2424 46 33 2.25277 0.05961 0.17941 0.00503 2012 72 17 2.32150 0.02854 0.11891 0.00232 1910 50 36 2.16689 0.01761 0.16244 0.00377 2465 38 22 1.90356 0.01891 0.00232 1913 37 41 2.01535 0.03354 0.1609 0.03151 0.00215 1.885 34 27 1.98405 0.01240 0.1799 0.0266 2.53 2.5 5 3.05080 0.11373 0.0032 2590 29 30 2.12648 0.04905 0.1699 0.0256 2.528 2.6 4 <t< td=""><td>24</td><td>2.08654</td><td>0.02246</td><td>0.17005</td><td>0.00403</td><td>2558</td><td>40</td><td>14</td><td>3.20727</td><td>0.05233</td><td>0.11541</td><td>0.00403</td><td>1886</td><td>63</td></t<>	24	2.08654	0.02246	0.17005	0.00403	2558	40	14	3.20727	0.05233	0.11541	0.00403	1886	63
29 1.94793 0.01875 0.18165 0.00362 2668 33 16 2.46464 0.00265 0.14288 0.00414 2262 29 32 2.5277 0.05906 0.17951 0.00840 2593 170 2.0 2.75826 0.03080 0.11619 0.00323 1940 38 37 2.10470 0.02144 0.16029 0.0037 2465 38 22 1.90356 0.01938 0.00313 2754 27 41 2.01535 0.03346 0.16079 0.00389 228 23 2.91846 0.02188 0.00231 2133 37 43 1.92539 0.02356 0.11534 0.00215 1883 34 27 1.98406 0.0174 0.12388 0.00231 2133 44 2.8984 0.04559 0.11337 0.00260 2577 26 31 2.05717 0.03346 0.16699 0.00266 249 27 45 2.16650 0.04325	26	2.05373	0.02016	0.17351	0.00356	2592	34	15	1.92504	0.02027	0.20840	0.00394	2782	70
32 2.88767 0.07009 0.12380 0.00503 2012 72 17 2.32150 0.03080 0.15702 0.00422 2.424 46 33 2.25277 0.05961 0.17951 0.06340 0.2593 170 20 2.75826 0.03080 0.11691 0.00323 1910 50 34 2.10470 0.02144 0.16244 0.0237 2465 38 22 1.90356 0.02348 0.00217 2000 31 41 2.01535 0.02335 0.11534 0.00210 2380 23 24 2.8546 0.02188 0.02331 2013 33 43 2.95384 0.04235 1.840 72 29 1.69155 0.0174 0.17399 0.00247 749 27 45 3.00580 0.1089 0.0047 1840 72 1.89405 0.01699 0.0026 257 26 31 2.0517 0.0346 0.1689 0.0027 2493 27	29	1.94793	0.01875	0.18163	0.00362	2668	33	16	2.46464	0.02025	0.14288	0.00241	2262	29
33 2.2527 0.05961 0.1764 0.00219 2593 170 20 2.78826 0.03080 0.11691 0.00323 1910 50 36 2.10689 0.01761 0.16244 0.00279 2481 29 21 2.86951 0.02848 0.11891 0.00252 1940 38 41 2.01535 0.03344 0.16692 0.03934 0.1692 0.02316 0.02316 131 42 2.16961 0.02316 0.15304 0.00210 2380 23 2.4 2.85486 0.02718 0.11238 0.00231 2113 33 43 1.9259 0.02345 0.11247 0.00417 1.840 27 1.98405 0.01704 0.1738 0.00266 2532 25 45 3.00580 0.01833 0.01747 0.00477 1.840 2.7 1.98405 0.01704 0.16699 0.00256 2532 2.6 45 3.00580 0.01733 0.00260 2.557 2.6 31 2.05717 0.03434 0.16699 0.00232 1.880	32	2.85767	0.07009	0.12380	0.00503	2012	72	17	2.32150	0.02695	0.15702	0.00422	2424	46
36 2.10689 0.01761 0.16244 0.00279 2481 29 21 2.86951 0.02854 0.11891 0.00252 1940 38 37 2.10470 0.02184 0.16092 0.00389 226 39 23 2.91840 0.02588 0.12296 0.00217 2000 31 42 2.16961 0.02336 0.15304 0.00219 2380 23 24 2.85486 0.02718 0.11717 0.00239 1913 37 43 1.92539 0.0233 0.11357 0.00215 1885 34 277 1.98405 0.01704 0.17099 0.00266 2533 2.57 26 3.8766 0.27164 0.06069 0.0626 253 2.57 26 31 2.05761 0.03355 0.01252 2.889 2.295561 0.03755 0.1053 0.0026 257 26 31 2.05717 0.03346 0.1680 0.00276 253 2.4550 0.03755 0.1150 0.00232 1880 36 47 2.26551 0.04337 0.11733 0.00184 <td>33</td> <td>2.25277</td> <td>0.05961</td> <td>0.17951</td> <td>0.00840</td> <td>2593</td> <td>170</td> <td>20</td> <td>2.75826</td> <td>0.03080</td> <td>0.11691</td> <td>0.00323</td> <td>1910</td> <td>50</td>	33	2.25277	0.05961	0.17951	0.00840	2593	170	20	2.75826	0.03080	0.11691	0.00323	1910	50
37 2.10470 0.02184 0.16092 0.00357 2465 38 22 1.90356 0.01692 0.19138 0.00313 2754 27 41 2.01535 0.03934 0.16679 0.00389 2526 39 23 2.91840 0.02288 0.12286 0.00217 2000 31 43 1.92539 0.02335 0.19354 0.00210 2380 23 2.91846 0.02768 0.12388 0.00231 2103 33 44 2.89384 0.04585 0.11337 0.00210 2380 23 2.91665 0.06546 0.21068 0.00666 210 47 45 3.00580 0.11327 0.00471 2666 43 2.05551 0.16388 0.01679 2463 32 2.95561 0.03785 0.11502 0.0232 1880 36 49 2.77960 0.03437 0.11753 0.00148 1919 29 33 2.15862 0.02469 0.16389 0.0179 2464 19 30 2.00426 0.16080 0.0179 2461 19	36	2.10689	0.01761	0.16244	0.00279	2481	29	21	2.86951	0.02854	0.11891	0.00252	1940	38
41 2.01535 0.03934 0.16679 0.00389 2526 39 23 2.91840 0.02588 0.12296 0.00217 2000 31 42 2.16961 0.02316 0.15354 0.00215 128 24 2.85486 0.02718 0.11717 0.00231 2013 33 44 2.89384 0.04585 0.11535 0.00215 1885 34 27 1.98405 0.01704 0.17380 0.00266 2528 26 45 3.00580 0.01859 0.11737 0.00302 2590 29 1.6155 0.06646 0.21064 0.00667 2493 27 48 2.18789 0.06672 0.11733 0.00302 2557 26 31 2.05717 0.03345 0.11502 0.00221 218 36 50 2.00420 0.03370 0.01732 0.0212 2589 22 35 2.16842 0.04375 0.11753 0.00232 2589 22 35 2.04325 0.17010 0.00233 2561 255 50 2.09473 <t< td=""><td>37</td><td>2.10470</td><td>0.02184</td><td>0.16092</td><td>0.00357</td><td>2465</td><td>38</td><td>22</td><td>1.90356</td><td>0.01692</td><td>0.19138</td><td>0.00313</td><td>2754</td><td>27</td></t<>	37	2.10470	0.02184	0.16092	0.00357	2465	38	22	1.90356	0.01692	0.19138	0.00313	2754	27
42 2.16961 0.02316 0.15304 0.00210 2380 23 24 2.85486 0.02718 0.11717 0.00239 1913 37 43 1.92539 0.02335 0.19334 0.00230 2772 25 26 2.89766 0.02796 0.12388 0.00231 2013 33 44 2.89844 0.04584 0.01659 0.01254 1.885 34 27 1.98405 0.01654 0.0266 2563 252 45 3.00580 0.01690 0.17337 0.00260 2557 26 31 2.05717 0.03346 0.16699 0.00267 2528 26 47 2.26635 0.04325 0.18146 0.00479 2666 44 32 2.95561 0.03346 0.11502 0.00232 1880 36 49 2.7790 0.03437 0.11733 0.00285 1940 37 36 2.05450 0.03322 0.1701 0.00233 2561 13 50 2.09473 0.07834 0.11849 0.00211 2115 28 3	41	2.01535	0.03934	0.16679	0.00389	2526	39	23	2.91840	0.02588	0.12296	0.00217	2000	31
43 1.92539 0.02335 0.19354 0.00293 2.772 25 26 2.89766 0.02796 0.12388 0.00231 2.013 33 44 2.89384 0.04585 0.01535 0.00215 1.885 34 27 1.98405 0.01704 0.07996 0.00266 2532 25 45 3.00580 0.01859 0.01470 0.00302 2590 29 30 2.12648 0.04905 0.16699 0.00256 2528 26 47 2.26635 0.04325 0.1893 0.00270 2557 26 31 2.05717 0.03346 0.16358 0.00232 1880 36 9 2.77960 0.03437 0.11733 0.00181 1919 29 33 2.15862 0.02350 0.1701 0.00233 2625 198 52 2.99473 0.07834 0.1189 0.00295 2589 22 35 2.04325 0.0322 0.1704 0.00233 2561 23 54 2.2529 0.04850 0.17340 0.00299 2648 <t< td=""><td>42</td><td>2.16961</td><td>0.02316</td><td>0.15304</td><td>0.00210</td><td>2380</td><td>23</td><td>24</td><td>2.85486</td><td>0.02718</td><td>0.11717</td><td>0.00239</td><td>1913</td><td>37</td></t<>	42	2.16961	0.02316	0.15304	0.00210	2380	23	24	2.85486	0.02718	0.11717	0.00239	1913	37
44 2.89384 0.04585 0.11535 0.00215 1885 34 27 1.98405 0.01704 0.17999 0.00266 2653 25 45 3.00580 0.10859 0.11247 0.00447 1840 72 29 1.69155 0.06646 0.21064 0.00606 2910 47 46 2.01589 0.04169 0.17337 0.00320 2590 26 31 2.05717 0.03346 0.16699 0.00232 1880 36 49 2.77960 0.0347 0.11753 0.0018 1919 29 33 2.15862 0.02350 0.1701 0.00203 2625 19 50 2.00042 0.02360 0.17732 0.0025 2589 22 35 2.04325 0.03322 0.16080 0.00179 2464 19 51 2.99473 0.07834 0.11889 0.00211 2115 28 37 2.62953 0.04688 0.11823 0.00225 1979 33 52 2.58325 0.04207 0.13124 0.00211 2148 <td< td=""><td>43</td><td>1.92539</td><td>0.02335</td><td>0.19354</td><td>0.00293</td><td>2772</td><td>25</td><td>26</td><td>2.89766</td><td>0.02796</td><td>0.12388</td><td>0.00231</td><td>2013</td><td>33</td></td<>	43	1.92539	0.02335	0.19354	0.00293	2772	25	26	2.89766	0.02796	0.12388	0.00231	2013	33
45 3.00580 0.10859 0.11247 0.00447 1840 72 29 1.69155 0.06546 0.21064 0.00606 2910 47 46 2.01589 0.04169 0.17337 0.00320 2557 26 31 2.05717 0.03346 0.16358 0.00267 2493 27 47 2.26635 0.04337 0.11753 0.00188 1919 29 33 2.15862 0.02460 0.16358 0.00207 2464 19 50 2.00042 0.02360 0.17323 0.00255 2589 22 35 2.04325 0.0236 0.1701 0.00203 2561 23 52 2.99473 0.07834 0.1189 0.0025 2889 22 35 2.04325 0.0322 0.17014 0.00233 2561 23 54 2.2529 0.4850 0.17146 0.0029 2648 28 37 2.62953 0.04868 0.1827 0.00262 1979 33 61 2.07144 0.1838 0.1629 0.00211 2115 28 <td>44</td> <td>2.89384</td> <td>0.04585</td> <td>0.11535</td> <td>0.00215</td> <td>1885</td> <td>34</td> <td>27</td> <td>1.98405</td> <td>0.01704</td> <td>0.17999</td> <td>0.00266</td> <td>2653</td> <td>25</td>	44	2.89384	0.04585	0.11535	0.00215	1885	34	27	1.98405	0.01704	0.17999	0.00266	2653	25
46 2.01589 0.04169 0.17337 0.00302 2590 29 30 2.12648 0.04905 0.16699 0.00256 2528 26 47 2.26635 0.04325 0.16993 0.00260 2557 26 31 2.05717 0.03346 0.16538 0.00252 2493 27 48 2.17860 0.06472 0.18146 0.00479 266 44 32 2.95561 0.03785 0.11600 0.00179 2464 19 50 2.0042 0.02360 0.17323 0.00225 2589 22 35 2.04325 0.02352 0.1701 0.00233 2561 23 51 2.85325 0.04850 0.17946 0.00214 248 28 37 2.62953 0.04868 0.17827 0.00225 1979 33 61 2.07194 0.01833 0.16292 0.0021 248 28 37 2.7528 0.0322 0.11423 0.00142 1868 22 63 2.17455 0.04451 0.17896 0.00311 2413 32<	45	3.00580	0.10859	0.11247	0.00447	1840	72	29	1.69155	0.06546	0.21064	0.00606	2910	47
47 2.26635 0.04325 0.16993 0.00260 2557 26 31 2.05717 0.03346 0.16358 0.00267 2493 27 48 2.18789 0.06672 0.18146 0.00479 2666 44 32 2.95561 0.03785 0.11502 0.00232 1880 36 49 2.77960 0.03437 0.11753 0.00188 1919 29 33 2.15862 0.02359 0.1701 0.00203 225 19 52 2.99473 0.07834 0.11889 0.00245 1940 37 36 2.05450 0.03322 0.1701 0.00233 2561 23 54 2.2529 0.04850 0.17946 0.00299 2648 28 37 2.62953 0.04868 0.1287 0.00266 2074 37 55 2.8825 0.04451 0.17896 0.00211 2115 28 38 2.86244 0.04625 0.1113 0.00144 1952 25 64 1.90561 0.06833 0.1717 0.00497 2569 49<	46	2.01589	0.04169	0.17337	0.00302	2590	29	30	2.12648	0.04905	0.16699	0.00256	2528	26
48 2.18789 0.06672 0.18146 0.00479 2666 44 32 2.95561 0.03785 0.11502 0.00232 1880 36 49 2.77960 0.03437 0.11753 0.00188 1919 29 33 2.15862 0.02496 0.16080 0.00179 2464 19 50 2.09473 0.07834 0.11889 0.00255 2589 22 35 2.04525 0.03322 0.17034 0.00233 2661 33 54 2.2529 0.04850 0.17946 0.00299 2648 28 37 2.62953 0.04868 0.1225 0.00265 1979 33 61 2.07194 0.01833 0.16292 0.00211 2115 28 38 2.86244 0.04625 0.1215 0.00225 1979 33 61 2.07194 0.01833 0.16292 0.00311 2643 32 40 2.70444 0.0284 0.11713 0.00164 1913 25 64 1.90561 0.06671 0.1844 0.0517 2711 4	47	2.26635	0.04325	0.16993	0.00260	2557	26	31	2.05717	0.03346	0.16358	0.00267	2493	27
49 2.77960 0.03437 0.11753 0.00188 1919 29 33 2.15862 0.02496 0.16080 0.00179 2464 19 50 2.00042 0.02360 0.17323 0.00225 2589 22 35 2.04325 0.02359 0.17011 0.00203 2625 19 52 2.99473 0.07844 0.11889 0.00219 2648 28 37 2.6253 0.04868 0.12827 0.00266 2074 37 55 2.58325 0.044207 0.13124 0.00211 2115 28 38 2.86244 0.04625 0.01125 0.00214 1868 22 63 2.17455 0.04451 0.17896 0.00311 2643 32 40 2.79044 0.02844 0.11713 0.00164 1913 25 64 1.90561 0.06733 0.16246 0.00332 2481 37 43 2.87080 0.02174 0.01141 1213 24 68 2.10895 0.05405 0.16389 0.0322 2496 34 4	48	2.18789	0.06672	0.18146	0.00479	2666	44	32	2.95561	0.03785	0.11502	0.00232	1880	36
50 2.0042 0.02360 0.17323 0.00225 2589 22 35 2.04325 0.02359 0.1701 0.00203 2625 19 52 2.99473 0.07834 0.11889 0.00245 1940 37 36 2.05450 0.03322 0.17034 0.00233 2561 23 54 2.2529 0.04850 0.17946 0.00299 2648 28 37 2.62953 0.04868 0.12155 0.00266 2074 37 55 2.58325 0.04451 0.17896 0.00211 2115 28 38 2.86244 0.04625 0.12155 0.00245 1979 33 61 2.07194 0.01431 0.16292 0.00201 2486 21 39 2.97520 0.03232 0.11423 0.00141 1913 25 64 1.90561 0.06983 0.17117 0.00497 2569 49 41 2.7581 0.02724 0.11974 0.00164 1952 25 66 2.18630 0.0671 0.18644 0.00517 2711	49	2.77960	0.03437	0.11753	0.00188	1919	29	33	2.15862	0.02496	0.16080	0.00179	2464	19
52 2.99473 0.07834 0.11889 0.00245 1940 37 36 2.05450 0.03222 0.17034 0.00233 2561 23 54 2.22529 0.04850 0.17946 0.00299 2648 28 37 2.62953 0.04868 0.12827 0.00266 2074 37 55 2.58325 0.04207 0.13124 0.00211 2115 28 38 2.86244 0.04625 0.12155 0.00225 1979 33 61 2.07194 0.01833 0.16292 0.00211 2486 21 39 2.97520 0.03232 0.11423 0.00142 1868 22 63 2.17455 0.04451 0.17896 0.00312 2463 32 40 2.79044 0.02884 0.11713 0.00164 1952 25 66 2.18630 0.06753 0.16246 0.00351 2711 46 44 2.57433 0.02873 0.1212 0.00151 1974 22 67 1.89469 0.06671 0.18644 0.00532 2466	50	2.00042	0.02360	0.17323	0.00225	2589	22	35	2.04325	0.02359	0.17701	0.00203	2625	19
54 2.22529 0.04850 0.17946 0.00299 2648 28 37 2.62953 0.04868 0.12827 0.00266 2074 37 55 2.58325 0.04207 0.13124 0.00211 2115 28 38 2.86244 0.04625 0.12155 0.00225 1979 33 61 2.07194 0.01833 0.16292 0.00201 2486 21 39 2.97520 0.03232 0.11423 0.00142 1868 22 63 2.17455 0.04451 0.17896 0.00341 2643 32 40 2.79044 0.02884 0.11713 0.00164 1913 25 64 1.90561 0.06973 0.16246 0.00353 2481 37 43 2.87008 0.02783 0.1212 0.00181 2131 24 68 2.10895 0.05405 0.16389 0.00332 2496 34 45 1.84628 0.02486 0.19790 0.00247 2809 20 69 1.96970 0.9003 0.17287 0.00332 2068 <	52	2.99473	0.07834	0.11889	0.00245	1940	37	36	2.05450	0.03322	0.17034	0.00233	2561	23
55 2.58325 0.04207 0.13124 0.00211 2115 28 38 2.86244 0.04255 0.12155 0.00225 1979 33 61 2.07194 0.01833 0.16292 0.00201 2486 21 39 2.97520 0.03232 0.11423 0.00142 1868 22 63 2.17455 0.04451 0.17896 0.00341 2643 32 40 2.79044 0.02844 0.11713 0.00164 1913 25 64 1.90561 0.06983 0.17117 0.00497 2569 49 41 2.72581 0.02724 0.11974 0.00164 1952 25 66 2.18630 0.06673 0.16246 0.00353 2481 37 43 2.87008 0.02873 0.12122 0.00181 2131 24 68 2.10895 0.05405 0.16389 0.00322 2496 34 45 1.84628 0.02486 0.19790 0.00217 2646 20 70 2.76208 0.08952 0.1781 0.00322 2068	54	2.22529	0.04850	0.17946	0.00299	2648	28	37	2.62953	0.04868	0.12827	0.00266	2074	37
61 2.07194 0.01833 0.16292 0.00201 2486 21 39 2.97520 0.03232 0.11423 0.00142 1868 22 63 2.17455 0.04451 0.17896 0.00341 2643 32 40 2.79044 0.02884 0.11713 0.00164 1913 25 64 1.90561 0.06983 0.17117 0.00497 2569 49 41 2.72581 0.02724 0.11974 0.00164 1952 25 66 2.18630 0.06753 0.16246 0.00353 2481 37 43 2.87008 0.02873 0.1212 0.00181 2131 24 67 1.89469 0.06671 0.18644 0.00517 2711 46 44 2.57433 0.02486 0.19790 0.00247 2809 200 69 1.96970 0.09030 0.17287 0.00332 2068 46 2.01090 0.02182 0.17923 0.00217 2646 20 70 2.76208 0.0783 0.12450 0.00353 2402 39	55	2.58325	0.04207	0.13124	0.00211	2115	28	38	2.86244	0.04625	0.12155	0.00225	1979	33
63 2.17455 0.04451 0.17896 0.00341 2643 32 40 2.79044 0.02884 0.11713 0.00164 1913 25 64 1.90561 0.06933 0.17117 0.00497 2569 49 41 2.72581 0.02724 0.11974 0.00164 1952 25 66 2.18630 0.06753 0.16246 0.0353 2481 37 43 2.87008 0.02873 0.12122 0.00151 1974 22 67 1.89469 0.06671 0.18644 0.00517 2711 46 44 2.57433 0.02958 0.13246 0.00181 2131 24 68 2.10895 0.05405 0.16389 0.0032 2496 34 45 1.84628 0.02486 0.19790 0.00247 2809 20 69 1.96970 0.90903 0.17287 0.00322 2068 46 48 1.82326 0.02828 0.20844 0.00363 2893 28 71 2.65503 0.0783 0.12781 0.00501 2023 <t< td=""><td>61</td><td>2.07194</td><td>0.01833</td><td>0.16292</td><td>0.00201</td><td>2486</td><td>21</td><td>39</td><td>2.97520</td><td>0.03232</td><td>0.11423</td><td>0.00142</td><td>1868</td><td>22</td></t<>	61	2.07194	0.01833	0.16292	0.00201	2486	21	39	2.97520	0.03232	0.11423	0.00142	1868	22
64 1.90561 0.06983 0.17117 0.00497 2569 49 41 2.72581 0.02724 0.11974 0.00164 1952 25 66 2.18630 0.06753 0.16246 0.00353 2481 37 43 2.87008 0.02873 0.12122 0.00151 1974 22 67 1.89469 0.06671 0.18644 0.00517 2711 46 44 2.57433 0.02958 0.13246 0.00181 2131 24 68 2.10895 0.05405 0.16389 0.00332 2496 34 45 1.84628 0.02486 0.19790 0.00247 2809 20 69 1.96970 0.09033 0.17287 0.00332 2068 46 48 1.82326 0.02828 0.02844 0.00363 2893 28 71 2.65503 0.07883 0.12456 0.00504 2023 72 49 3.03088 0.03731 0.11434 0.00152 1870 24 72 2.23728 0.07181 0.15502 0.00353 2402	63	2.17455	0.04451	0.17896	0.00341	2643	32	40	2.79044	0.02884	0.11713	0.00164	1913	25
66 2.18630 0.06753 0.16246 0.00353 2481 37 43 2.87008 0.02873 0.12122 0.00151 1974 22 67 1.89469 0.06671 0.18644 0.00517 2711 46 44 2.57433 0.02958 0.13246 0.00181 2131 24 68 2.10895 0.05405 0.16389 0.00332 2496 34 45 1.84628 0.02486 0.19790 0.00247 2809 20 69 1.96970 0.09003 0.17287 0.00593 2586 57 46 2.01090 0.02182 0.17923 0.00217 2646 20 70 2.76208 0.08952 0.12781 0.00332 2068 46 48 1.82326 0.02828 0.20844 0.00363 2893 28 71 2.65503 0.07883 0.12456 0.00504 2023 72 49 3.03088 0.03731 0.11434 0.00152 1870 24 72 2.23728 0.07181 0.15502 0.00531 2507	64	1.90561	0.06983	0.17117	0.00497	2569	49	41	2.72581	0.02724	0.11974	0.00164	1952	25
67 1.89469 0.06671 0.18644 0.00517 2711 46 44 2.57433 0.02958 0.13246 0.00181 2131 24 68 2.10895 0.05405 0.16389 0.00332 2496 34 45 1.84628 0.02486 0.19790 0.00247 2809 20 69 1.96970 0.09033 0.17287 0.00593 2586 57 46 2.01090 0.02182 0.17923 0.00217 2646 20 70 2.76208 0.08952 0.12781 0.00332 2068 46 48 1.82326 0.02828 0.20844 0.00363 2893 28 71 2.65503 0.07883 0.12456 0.00504 2023 72 49 3.03088 0.03731 0.11434 0.00152 1870 24 72 2.23728 0.07181 0.15502 0.00353 2402 39 50 2.11984 0.03586 0.16441 0.00283 2502 29 73 2.08899 0.09390 0.16490 0.00531 2507	66	2.18630	0.06753	0.16246	0.00353	2481	37	43	2.87008	0.02873	0.12122	0.00151	1974	22
68 2.10895 0.05405 0.16389 0.00332 2496 34 45 1.84628 0.02486 0.19790 0.00247 2809 20 69 1.96970 0.09003 0.17287 0.00593 2586 57 46 2.01090 0.02182 0.17923 0.00217 2646 20 70 2.76208 0.08952 0.12781 0.00332 2068 46 48 1.82326 0.02828 0.20844 0.00363 2893 28 71 2.65503 0.07883 0.12456 0.00504 2023 72 49 3.03088 0.03731 0.11434 0.00152 1870 24 72 2.23728 0.07181 0.15502 0.00531 2507 54 51 1.93153 0.02150 0.19347 0.00209 2772 18 52 2.52505 0.03390 0.16490 0.00531 2507 54 51 1.93153 0.0312 2109 2703 28 54 1.88531 0.0312 2108 55 2.09824 0.03403 0.176	67	1.89469	0.06671	0.18644	0.00517	2711	46	44	2.57433	0.02958	0.13246	0.00181	2131	24
69 1.96970 0.09003 0.17287 0.00593 2586 57 46 2.01090 0.02182 0.17923 0.00217 2646 20 70 2.76208 0.08952 0.12781 0.00332 2068 46 48 1.82326 0.02828 0.20844 0.00363 2893 28 71 2.65503 0.07883 0.12456 0.00504 2023 72 49 3.03088 0.03731 0.11434 0.00152 1870 24 72 2.23728 0.07181 0.15502 0.00531 2507 54 51 1.93153 0.02150 0.19347 0.00209 2772 18 73 2.08899 0.09390 0.16490 0.00531 2507 54 51 1.93153 0.02162 0.19347 0.00209 2772 18 52 2.52505 0.0338 0.13393 0.00212 2703 28 54 1.88531 0.03612 0.18554 0.0012 2703 28 54 1.88531 0.03403 0.17692 0.00267 <t< td=""><td>68</td><td>2.10895</td><td>0.05405</td><td>0.16389</td><td>0.00332</td><td>2496</td><td>34</td><td>45</td><td>1.84628</td><td>0.02486</td><td>0.19790</td><td>0.00247</td><td>2809</td><td>20</td></t<>	68	2.10895	0.05405	0.16389	0.00332	2496	34	45	1.84628	0.02486	0.19790	0.00247	2809	20
70 2.76208 0.08952 0.12781 0.00332 2068 46 48 1.82326 0.02828 0.20844 0.00363 2893 28 71 2.65503 0.07833 0.12456 0.00504 2023 72 49 3.03088 0.03731 0.11434 0.00152 1870 24 72 2.23728 0.07181 0.15502 0.00353 2402 39 50 2.11984 0.03566 0.16441 0.00283 2502 29 73 2.08899 0.09390 0.16490 0.00531 2507 54 51 1.93153 0.02150 0.19347 0.00209 2772 18 52 2.52505 0.03398 0.13393 0.00202 2150 266 54 1.88531 0.03612 0.18554 0.0012 2703 28 55 2.09824 0.03403 0.17692 0.00267 2624 24 56 1.89127 0.0236 0.18672 0.00240 2713 21 57 2.04890 0.02926 0.16912 0	69	1.96970	0.09003	0.17287	0.00593	2586	57	46	2.01090	0.02182	0.17923	0.00217	2646	20
71 2.65503 0.07883 0.12456 0.00504 2023 72 49 3.03088 0.03731 0.11434 0.00152 1870 24 72 2.23728 0.07181 0.15502 0.00353 2402 39 50 2.11984 0.03586 0.16441 0.00283 2502 29 73 2.08899 0.09390 0.16490 0.00531 2507 54 51 1.93153 0.02150 0.19347 0.00209 2772 18 74 1.8899 0.09390 0.16490 0.00531 2507 54 51 1.93153 0.02150 0.19347 0.00209 2772 18 75 2.08899 0.09390 0.16490 0.00531 2507 54 51 1.93153 0.02150 0.19347 0.00202 2150 26 54 1.88531 0.03612 0.18554 0.00122 2703 28 55 2.09824 0.03403 0.17692 0.00267 2624 24 56 1.89127 0.02366 0.16421 0.00268 <	70	2.76208	0.08952	0.12781	0.00332	2068	46	48	1.82326	0.02828	0.20844	0.00363	2893	28
72 2.23728 0.07181 0.15502 0.00353 2402 39 50 2.11984 0.03586 0.16441 0.00283 2502 29 73 2.08899 0.09390 0.16490 0.00531 2507 54 51 1.93153 0.02150 0.19347 0.00209 2772 18 52 2.52505 0.03398 0.13393 0.00202 2150 26 54 1.88531 0.03612 0.18554 0.00312 2703 28 55 2.09824 0.03403 0.17692 0.00240 2713 21 56 1.89127 0.02366 0.16412 0.00268 2549 27 58 2.12961 0.02396 0.16742 0.00203 2532 20 61 2.10855 0.0279 0.16239 0.00201 2481 21 62 1.96878 0.03215 0.17437 0.00279 2600 27 63 2.87696 0.05598 0.11963 0.00192 1951 29 <td>71</td> <td>2.65503</td> <td>0.07883</td> <td>0.12456</td> <td>0.00504</td> <td>2023</td> <td>72</td> <td>49</td> <td>3.03088</td> <td>0.03731</td> <td>0.11434</td> <td>0.00152</td> <td>1870</td> <td>24</td>	71	2.65503	0.07883	0.12456	0.00504	2023	72	49	3.03088	0.03731	0.11434	0.00152	1870	24
73 2.08899 0.09390 0.16490 0.00531 2507 54 51 1.93153 0.02150 0.19347 0.00209 2772 18 52 2.52505 0.03398 0.13393 0.00202 2150 26 54 1.88531 0.03612 0.18554 0.00312 2703 28 55 2.09824 0.03403 0.17692 0.00257 2624 24 56 1.89127 0.02366 0.18672 0.00240 2713 21 57 2.04890 0.02926 0.16912 0.00268 2549 27 58 2.12961 0.02396 0.16742 0.00203 2532 20 61 2.10855 0.02799 0.16239 0.00201 2481 21 62 1.96878 0.03215 0.17437 0.00279 2600 27 63 2.87696 0.05598 0.11963 0.00192 1951 29	72	2.23728	0.07181	0.15502	0.00353	2402	39	50	2.11984	0.03586	0.16441	0.00283	2502	29
$\begin{array}{cccccccccccccccccccccccccccccccccccc$	73	2.08899	0.09390	0.16490	0.00531	2507	54	51	1.93153	0.02150	0.19347	0.00209	2772	18
541.885310.036120.185540.00312270328552.098240.034030.176920.00257262424561.891270.023360.186720.00240271321572.048900.029260.169120.00268254927582.129610.023960.167420.00203253220612.108550.027990.162390.00201248121621.968780.032150.174370.00279260027632.876960.055980.119630.00192195129								52	2.52505	0.03398	0.13393	0.00202	2150	26
552.098240.034030.176920.00257262424561.891270.023360.186720.00240271321572.048900.029260.169120.00268254927582.129610.023960.167420.00203253220612.108550.027990.162390.00201248121621.968780.032150.174370.00279260027632.876960.055980.119630.00192195129								54	1.88531	0.03612	0.18554	0.00312	2703	28
561.891270.023360.186720.00240271321572.048900.029260.169120.00268254927582.129610.023960.167420.00203253220612.108550.027990.162390.00201248121621.968780.032150.174370.00279260027632.876960.055980.119630.00192195129								55	2.09824	0.03403	0.17692	0.00257	2624	24
572.048900.029260.169120.00268254927582.129610.023960.167420.00203253220612.108550.027990.162390.00201248121621.968780.032150.174370.00279260027632.876960.055980.119630.00192195129								56	1.89127	0.02336	0.18672	0.00240	2713	21
582.129610.023960.167420.00203253220612.108550.027990.162390.00201248121621.968780.032150.174370.00279260027632.876960.055980.119630.00192195129								57	2.04890	0.02926	0.16912	0.00268	2549	27
612.108550.027990.162390.00201248121621.968780.032150.174370.00279260027632.876960.055980.119630.00192195129								58	2.12961	0.02396	0.16742	0.00203	2532	20
621.968780.032150.174370.00279260027632.876960.055980.119630.00192195129								61	2.10855	0.02799	0.16239	0.00201	2481	21
63 2.87696 0.05598 0.11963 0.00192 1951 29								62	1.96878	0.03215	0.17437	0.00279	2600	27
								63	2.87696	0.05598	0.11963	0.00192	1951	29



4.5.4.1 Sample 184 (Daniel Creek Formation – North Ortona)

Zircons collected from sample 184 are generally poorly- to well rounded with many containing pitted exteriors indicating a detrital origin involving varying degrees of abrasion and transportation prior to deposition. Poorly rounded grains are generally elongate, appearing to preserve almost unmodified crystal faces, and may have originally been derived from an igneous source. Overall, the grains are approximately uniform in size, averaging ~100 x 60 μ m.

75 zircons from sample 184 were analysed over three sessions. However, 36 analyses showed significant discordance in their isotopic values due to common-Pb and/or Pb-loss, and were omitted from subsequent analysis. The discarded analyses included one grain which returned a 207 Pb/ 206 Pb age of 1602 ± 32 Ma that may be metamorphic in origin (Appendix 2). A summary of the isotopic values for the concordant analyses is presented in Table 4.4. Inverse U-Pb concordia and relative probability diagrams for sample 184 are illustrated in Fig. 4.14. In the relative probability diagram (Fig. 4.14c) there exists a large spread in individual zircon ages between ca. 1800 and 2800 Ma, with a significant peak between ca. 2400 Ma. This result suggests a largely Archaean source for the Daniel Creek Formation.

4.5.4.2 Sample 255 (Daniel Creek Formation – South Head)

Sample 255 was collected from an outcrop of the Daniel Creek Formation at South Head, approximately 60 km north-west of sample 184 (Fig. 4.1). As with sample 184, zircons collected from sample 255 are generally poorly to well-rounded, with the majority containing pitted exteriors indicating a detrital origin involving varying degrees of abrasion and transportation. Moderately to well-rounded zircons dominate this sample, with the remainder of a similar nature to the poorly-rounded, elongated grains of sample 184. Overall the grains show a slight bimodality in size and shape, ranging from somewhat equant, rounded grains (averaging ~110 x 100 μ m) to poorly-rounded, elongated grains (averaging ~170 x 60 μ m). Numerous zircons of both varieties contain small (<10 μ m) inclusions, the majority of which contain Pb.

63 zircons from sample 255 were analysed, however 15 analyses were discarded due to significant discordance. Isotopic values for the concordant analyses are summarised

in Table 4.4. Inverse U-Pb concordia and relative probability diagrams for sample 255 are illustrated in Fig. 4.14. Significant groupings in the relative probability diagram (Fig. 4.14d) occur at ca. 1900 to 2000, and 2500 to 2800 Ma. Four other analyses, including two at ca. 2270 Ma, one at ca. 2420, and one at ca. 3060 Ma were also recorded but do not form statistically significant groups. The significance of these inheritance ages is discussed in more detail below.

The majority of isotopic values for analyses from sample 255 are much more concordant than for sample 184, with only 15 discordant analyses recorded. However, the removal of these analyses from the relative probability diagram has little effect on the inheritance pattern of this sample.



Figure 4.14: Tera-Wasserburg U-Pb concordia diagrams for analyses from (a) sample 184, and (b) sample 255. Cumulative probability distribution diagrams for (c) sample 184, and (d) sample 255.

4.6 EMPA monazite dating results

Metapelite samples from areas that experienced amphibolite facies metamorphism or higher were preferentially selected for in situ monazite analysis. Initial analysis of monazite grains was conducted on three samples (samples 283, 288, and 308) in order to test the validity of the in situ method for the Etheridge Group metapelites, as well as give an initial estimation as to the metamorphic ages they record. Subsequent analyses were then conducted to expand on the age data collected from the initial analysis. Two samples (304 and 306) from the Robinhood area were also included to compare the EMPA monazite ages of this study to EMPA monazite ages reported by Cihan *et al.* (2006). A small percentage of analyses were discarded from subsequent age calculations based primarily on weight percent oxide totals, and low Pb abundances. Remaining analyses were then used to calculate weighted mean ages for each sample, and where appropriate for distinct age groups within each sample. Complete individual monazite analyses are listed in Appendix 2, and are summarised in Table 4.5. Weighted mean age calculations for each sample are summarised in Table 4.6.

4.6.1 Sample 283 (Einasleigh Metamorphics – Gum Flats)

Sample 283 is a strongly foliated mica schist from the Einasleigh Metamorphics south of Gum Flats (Fig. 4.1). It contains small 10-20 μ m monazite grains that are in textural equilibrium with the surrounding S₁ metamorphic assemblage. 30 analyses were conducted on 14 grains, with the results forming a single distinct group in the corresponding cumulative probability diagram (Fig. 4.15a). Of the 30 analyses, 6 were discarded before final weighted mean age calculations due to low weight percent oxide totals (Appendix 2).

A single analysis returned a calculated age of 1626 ± 14 Ma (Table 4.5). This analysis is not associated with the dominant metamorphic events in the Etheridge Group which are believed to have occurred between ca. 1590 and 1550 Ma, and may instead reflect partial resetting of a detrital monazite. Omitting this analysis, and four younger analyses which do not form part of the prominent peak in the cumulative probability diagram (Fig. 4.15a), the remaining 20 analyses combine to give a weighted mean age of 1537 ± 18 Ma. The MSWD for this age is relatively high (5.1), suggesting that this group may incorporate more than one discrete age population. This age is within error of the crystallisation ages determined for granitoids of the Forsayth Batholith (ca. 1550 Ma), and may relate to this period of elevated heat-flow and associated metamorphism in the Etheridge Group.

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Grain	Spot	Pb (ppm)	Th (ppm)	U (ppm)	Age (Ma)	$\pm 1 \sigma$	Grain	Spot	Pb (ppm)	Th (ppm)	U (ppm)	Age (Ma)	$\pm 1\sigma$
ample: 2	83 (Einas	sleigh Metamo	orphics - Gum	Flats)			Sample 28	38 cont.					
-	1	5089	52457	5260	1548	20	5	-	6803	94588	1630	1473	15
7	1	5408	49031	10206	1385	16	9	-	3429	38713	1871	1632	28
	7	6164	55394	8622	1549	17		7	5788	72701	1602	1598	18
	ŝ	5875	53319	8356	1532	17	7	-	2492	28301	2623	1441	32
Э	1	6696	52446	11549	1548	16	8	1	5505	27473	16011	1423	15
	2	4133	58526	2143	1363	21	6	-	1364	9089	2672	1581	59
4	-	4953	62293	2099	1539	21	10	-	3709	49496	2213	1409	22
	2	5309	55221	6287	1486	18	11	-	4254	51922	2008	1560	22
	ŝ	5360	54667	6345	1505	19	12	-	6365	82547	2152	1530	16
	4	7895	75448	10935	1498	13	13	1	8088	110927	1523	1509	13
5	1	8175	67754	11373	1626	14	14	-	522	15872	4164	397	28
9	-	6602	80115	7828	1338	14							
7	-	6855	55023	10766	1584	16	Sample:	304 (Lane	Creek Forma	tion - Robinho	(poq)		
8	-	5508	49196	8128	1527	18	1	-	3971	31220	6550	1572	23
6	1	2632	27154	2613	1562	36		2	4057	32294	7338	1506	22
10	1	6552	67910	5932	1590	17	2	1	4102	29297	7688	1563	23
11	1	5525	64064	4109	1523	18	б	1	4972	37710	8649	1567	19
	2	5280	55377	5968	1496	18	4	-	4926	39491	8749	1512	19
12	-	5188	55882	5119	1518	19	5	-	4117	34547	7079	1495	21
13	1	5514	55818	6323	1526	18	9	-	1943	18665	3100	1426	39
	2	5732	56016	6533	1564	18	7	-	5351	45922	7879	1563	18
14	1	5075	65591	7819	1198	15	8	-	2914	27393	4652	1444	28
	7	9261	94275	10150	1538	12	6	-	4692	37574	8353	1511	19
	б	6858	59669	11177	1496	15	10	-	4489	33782	8064	1553	21
							11	1	4608	36840	8003	1527	20
Sample: 2	288 (Danie	el Creek Forn	nation - Gum F	lats)			12	1	3163	27057	4958	1532	28
-	-	8140	105243	1916	1573	14		2	3158	25831	5121	1550	28
2	1	2709	26435	2849	1599	36	13	-	2183	20364	3857	1399	34
	7	1714	16390	2316	1509	50	14	1	3563	36056	4546	1482	24
б	1	1550	15452	2901	1320	46	15	1	3840	37206	5295	1488	23
	7	6140	27852	18216	1444	15	16	1	3738	38212	4426	1503	24
	б	6860	29335	20195	1473	14	17	-	4187	33965	7453	1502	21

Table 4.5: U-Th-Pb data and calculated ages for monazite analyses from metasedimentarysamples analysed by EMP.

Grain	Spot	Pb (ppm)	Th (ppm)	U (ppm)	Age (Ma)	$\pm 1 \sigma$	Grain	Spot	Pb (ppm)	Th (ppm)	U (ppm)	Age (Ma)	$\pm 1 \sigma$
Sample:	306 (Corb	tt Formation	1 - Robinhood)				Sample: 3	324 (Eina	sleigh Metamo	orphics - Gum	Flats)		
-	-	1637	30505	2088	958	28			5025	48698	6013	1551	19
	2	2475	25587	2619	1536	34		7	5062	46273	6874	1548	19
2	-	1725	12906	3266	1523	46	2	-	4099	40866	4302	1577	23
	2	1704	13030	3337	1486	45		2	3561	40152	3141	1505	25
ę	-	2655	26010	3343	1518	32	3	-	5085	45301	7266	1546	19
4	-	2990	30989	4152	1424	27		7	5130	43998	7470	1570	19
	2	1821	15729	3111	1476	43		ŝ	5098	44104	7609	1549	19
	ю	2195	19423	3432	1505	37	4	-	4011	46564	2727	1544	23
5	-	3109	27168	4879	1513	28		7	4966	44831	6923	1547	19
9	-	3113	33429	3255	1501	28		ю	3321	38455	2491	1522	27
	7	3095	33881	2950	1513	28		4	4759	43830	6451	1542	20
							5	-	4174	46660	3369	1542	24
Sample:	308 (Dani	el Creek Forn	nation - Gum F	(lats)				7	3844	42677	3290	1532	25
-	-	5714	50089	9780	1464	17		ю	5092	48922	6402	1537	20
2	-	14739	178379	8278	1537	6		4	5990	53494	7730	1595	18
	7	5695	44867	10331	1512	17	9	-	4165	45763	3617	1539	24
З	-	7716	60699	13572	1457	13		2	5355	46923	8558	1501	18
4	1	10056	69389	19957	1548	11	7	-	4035	36852	6064	1499	23
	7	7937	65681	13458	1515	13		7	3847	34123	5643	1536	25
5	-	2415	29820	1500	1492	36	8	1	4075	34996	6488	1520	23
	7	5587	80015	1686	1414	17		7	4262	37577	5864	1575	23
9	-	2545	36769	1182	1355	31		ŝ	4240	44369	4502	1522	23
7	1	8575	69362	14992	1514	13	6	1	4827	37444	7787	1597	21
							10		4303	37460	6143	1567	23

Table 4.5 cont.

Sample	Grains/Analyses	Age (Ma)	Error (95% C.I.)	MSWD	Prob. of Equivalence	Comment
283	13/20	1537	±18	5.1	-	Metamorphic age (M ₂)
288	5/6	1585	± 18	1.06	0.38	Metamorphic age (M ₁)
304	12/16	1527	± 16	2.0	0.012	Metamorphic age (M ₂ ?)
306	6/10	1497	± 20	1.17	0.31	Metamorphic age?
308	4/6	1529	± 17	1.6	0.15	Metamorphic age (M ₂ ?)
324	10/24	1546	± 11	1.6	0.038	Metamorphic age (M ₂)

Table 4.6: Summary of weighted mean ages calculated from EMPA monazite analyses in Table4.5. All weighted mean ages calculated using *Isoplot* version 3.0 (Ludwig, 2003).



Figure 4.15: Cumulative probability diagrams for U-Th-Pb monazite ages from (a) sample 283, (b) sample 288, (c) sample 304, (d) sample 306, (e) sample 308, and (f) sample 324.

4.6.2 Sample 288 (Daniel Creek Formation – Gum Flats)

Sample 288, a muscovite-biotite mica schist from the Daniel Creek Formation near Gum Flats (Fig. 4.1), contains small irregularly shaped monazites that vary from 5 to 30 μ m in size. All identifiable monazite grains in this sample lie parallel to the dominant S₁ fabric. A total of 31 analyses were conducted on 22 grains in this sample, with the analyses forming a broad double-peaked distribution on the cumulative probability diagram (Fig. 4.15b). Of the 31 analyses, 14 were discarded before final weighted mean age calculations due to low Th and total weight percent oxide values (Appendix 2).

When combined, the 17 remaining analyses do not form a statistically acceptable weighted mean age, suggesting that combined the analyses represent a spectrum of mixed ages. Therefore, to determine the pooled age of this sample, analyses were excluded until a statistically acceptable age was obtained. Using this method, the analyses that returned the six oldest ages (on five grains) combined to give a statistically valid weighted mean age of 1585 ± 18 Ma (MSWD = 1.06). This group includes one analysis of 1632 ± 28 Ma which may relate to an earlier stage of monazite growth, but importantly places an upper limit on the onset of metamorphism in this sample. The remaining analyses combine for a weighted mean age of 1472 ± 31 Ma. However, the MSWD for this pooled age is relatively high (5.7). This suggests that either more than one discrete age population is included in this sample, or that at least some monazites were affected by some Pb-loss at a later date.

The ca. 1585 Ma age is of regional significance, as it not only correlates with recent estimates for the age of medium-pressure/temperature metamorphism (M_1) and orogenesis in the Etheridge Group (Black *et al.*, 2005; Cihan *et al.*, 2006), but is also coeval with the age of regional-scale orogenesis and peak metamorphism in the Mt Isa Eastern Succession (Page & Sun, 1998). However, the ca. 1585 Ma age presented here must be treated with some caution, as it is based on a limited dataset. Consequently, it is considered to represent a maximum estimate for M_1 .

4.6.3 Sample 304 (Lane Creek Formation – Robinhood)

Sample 304 is a muscovite-biotite mica schist from the Lane Creek Formation near Robinhood (Fig. 4.1). BSE images reveal that the sample contains small irregularly

shaped monazites which vary from 5 to 30 μ m in size (Fig. 4.16a). In total, 27 analyses were conducted on 23 grains, with the analyses forming one large doublepeaked population with a broad younger shoulder on the cumulative probability diagram (Fig. 4.15c). Eight of the 27 analyses were discarded before final weighted mean age calculations due to low weight percent oxide totals and low Th (Appendix 2).

The 16 oldest analyses form the dominant age population in Figure 4.15c. They combine to give a weighted mean age of 1527 ± 16 Ma (MSWD = 2.0), which is within error of the metamorphic age calculated for sample 283 (Table 4.6). However, the relatively high MSWD suggests that this age may contain a mix of ca. 1550 Ma and ca. 1500 Ma monazites. This hypothesis is further demonstrated by discarding the oldest six analyses from this group, which improves the calculated weighted mean age to 1500 ± 13 Ma (MSWD = 1.16). Regardless, the 1527 ± 16 Ma metamorphic age is preferred as it includes all analyses that lie within the dominant peak of the probability diagram for this sample.

4.6.4 Sample 306 (Corbett Formation – Robinhood)

Sample 306 is a muscovite-biotite-garnet-staurolite schist belonging to the Corbett Formation in the Robinhood area (Fig. 4.1). Monazites in this sample are irregularly shaped but are generally elongated, varying from 10 to 30 μ m in width, and up to 150 μ m in length (Fig. 4.16b). Elongated monazite grains are aligned in the dominant S₃ fabric. In total, 13 analyses were conducted on 8 grains; however 2 analyses were discarded before final weighted mean age calculations due to low weight percent oxide totals and high K values that suggest chemical alteration has occurred (Appendix 2). The remaining 11 analyses form a single peak with a minor young shoulder on the cumulative probability diagram (Fig. 4.15d).

Of the remaining 11 analyses, 10 combine to give a weighted mean age of 1497 ± 20 Ma (MSWD = 1.17, probability of equivalence = 0.31), which is the most statistically acceptable estimate that can be produced from the available data. This age is too young to be linked to the period of granitoid emplacement and associated metamorphism that occurred around ca. 1550 Ma in the lower Etheridge Group. However, monazite analyses returning similar ages were identified in two

metasedimentary samples collected approximately 10 km north of sample 306 by Cihan *et al.* (2006), which they attributed to a thermal pulse or hydrothermal activity occurring after or during the late stages of the ca. 1550 Ma event. It is possible then that the weighted mean age is representative of this event; however the dominant foliation (N-S trend) is consistent with a later stage of folding in the area (F_3 of Cihan *et al.*, 2006).

A single analysis returned a calculated age of 958 ± 28 Ma (Table 4.5). This age is anomalous, as the analysis of another part of the same grain returned an age of $1536 \pm$ 34 Ma (Table 4.5). It is possible that this grain experienced some Pb-loss, as reflected by the younger calculated age for part of the grain.

4.6.5 Sample 308 (Daniel Creek Formation – Gum Flats)

Sample 308 is a strongly foliated, fine-grained mica schist from the Daniel Creek Formation at Gum Flats (Fig. 4.1). As with the other monazite-dating samples described above it contains small, irregularly shaped grains of monazite that are generally less than 15 μ m wide. In this sample, 10 analyses on 7 grains were attempted (Table 4.5). Unfortunately, the majority of analyses from this sample show some signs of alteration, with all but three analyses containing a combination of K > 0.02 wt%, P > 13.00 wt%, and weight percent oxide totals of less than 97 (Appendix 2). In addition broad younger shoulders on the cumulative probability diagram suggest Pb-loss may have affected a number of analyses (Fig. 4.15e).

Despite these limitations, the calculated ages are consistent with ages obtained from other samples, and are therefore retained for the weighted mean age calculations. Of the 10 analyses attempted, the four youngest ages were omitted until a statistically acceptable group of six analyses was formed. The six analyses are spread over four grains, and combine to give a weighted mean age of 1529 ± 17 (MSWD = 1.6). This age is within error of ages obtained from samples 304 and 283, and may therefore relate to the late stages of the high-temperature M₂ event at ca. 1550 Ma. The remaining four analyses are distributed over a large range from ca. 1500 Ma to ca. 1300 Ma corresponding to the broad younger shoulder of the cumulative probability diagram (Fig. 4.15e), and do not form a statistically acceptable group.



Figure 4.16: Backscattered electron (BSE) images of monazite grains and their textural settings for (a) sample 304, (b) sample 306, and (c) sample 324.

4.6.6 Sample 324 (Einasleigh Metamorphics – Gum Flats)

Sample 324, a coarse-grained muscovite-biotite mica schist from the Einasleigh Metamorphics north of Gum Flats (Fig. 4.1), contains moderately sized irregularly shaped monazites varying from 20 to 50 μ m in width. All analysed monazite grains in

this sample show corroded margins (Fig.4.16c); however, for the most part their chemistry remains largely unaffected. Some 27 analyses were conducted on 11 grains, with the analyses combining to form a single, symmetric peak on the cumulative probability diagram (Fig. 4.15f). Of the 27 analyses, 3 were discarded before final weighted mean age calculations, due to low weight percent oxide totals, and high K values (Appendix 2).

The 24 acceptable analyses combine to give a weighted mean age of 1546 ± 11 Ma (MSWD = 1.6). This value is consistent with the established age for the high-temperature metamorphic event at ca. 1550 Ma (M₂) that is also associated with significant felsic magmatism in the lower Etheridge Group (Black & McCulloch, 1990; Black & Withnall, 1993).

4.7 Discussion

4.7.1 Significance of a younger (ca. 1500 Ma) age population recorded in the metamorphic monazites of the lower Etheridge Group

The occurrence of a distinct monazite age population at ca. 1500 Ma in sample 306, coupled with mixed age populations occurring in other samples, may suggest that a relatively weak metamorphic or hydrothermal event occurred at this time. Little geochronological evidence concerning the presence of a metamorphic event around ca. 1500 Ma has been uncovered in previous studies of the Etheridge Group. Black *et al.* (1979) presented Rb-Sr total rock isochron data that dated M₂ in the Etheridge Group to 1469 \pm 20 Ma. However, subsequent U-Pb zircon dating (e.g. Black & Withnall, 1993) suggested that this age is probably related to cooling and may not have geological significance. Black *et al.* (2005) presented U-Pb zircon data from leucogneiss of the Einasleigh Metamorphics that gave metamorphic zircon ages ranging from 1560-1550 Ma to as low as 1500 Ma. The cause of these relatively young zircon ages in their study was not considered to be diagnostic of a younger (ca. 1500 Ma) event, but rather was attributed to the incomplete isotopic resetting of the zircons by partial recrystallization (Hoskin & Black, 2000).

The ca. 1500 Ma monazite age recorded in mica schist from the Corbett Formation (sample 306), and in mixed age populations containing ca. 1500 Ma analyses (samples 283 and 288; Table 4.5) has also been identified previously through

individual monazite ages in metasediments from the Robertson River Subgroup near Robinhood (Cihan *et al.*, 2006). The possible regional extent of this age suggests that a thermal pulse or foliation forming event may have affected the lower Etheridge Group at this time. However, this event has a limited spatial distribution and is not chronologically related to any dated intrusive body. Consequently, the existence and significance of this event await clarification by future work.

Significantly, similar activity is recorded in other Proterozoic basins in northern Australia, including the Mt Isa Eastern Succession. In this region, younger plutons of the Williams and Naraku Batholiths are associated with major regional metasomatism and record U-Pb zircon ages ranging from ca. 1510 to 1490 Ma (Page & Sun, 1998; Wyborn, 1998). The prevalence of the ca. 1510-1490 Ma igneous and/or metamorphic ages in the Mt Isa Inlier and its existence in the Georgetown Inlier, suggests that a thermal event of regional significance occurred at this time, producing I-type granitoids in the Mt Isa Eastern Succession, but only crystallising or recrystallising monazites and developing a localised N-S trending structural fabric in the metasediments of the lower Etheridge Group in the Georgetown Inlier.

In summary, the majority of metamorphic monazites analysed in this study lie within the dominant S_1 foliation, which is observable throughout the lower Etheridge Group in the central region of the Georgetown Inlier. The monazites record a single ca. 1540 Ma event that would appear to be related to S_1 in some samples (e.g. samples 283 and 288 were sampled far from granitoids), or is perhaps younger than S_1 as indicated by the lack of aligned monazite grains in other samples (e.g. samples 304, 308, and 324: all sampled near granitoids suggesting late heating). Based on this evidence, I prefer the interpretation that the monazites recrystallised as a result of later heating associated with M_2 and the emplacement of the Forsayth Batholith at ca. 1550 Ma, as they do not appear to be associated with a distinct post- S_1 foliation. A localised foliation-forming event (D_3) is also proposed for the ca. 1500 Ma monazites. The implications of these results are discussed further in Chapter 5.

4.7.2 Significance of detrital zircon populations in metasedimentary samples, and comparison to other Proterozoic terranes of northeastern Australia

The age distributions of detrital zircon populations in metasediments of the Etheridge Group have previously been reported by Black & McCulloch (1990), Black & Withnall (1993), and Black *et al.* (2005). The inheritance patterns in samples 184, 255, and 264 presented here provide a useful comparison to these published detrital zircon populations, as well as to the inheritance patterns observed in the metasediments of other Palaeoproterozoic basins in eastern Australia. These broad similarities are discussed in more detail below.

Leucogneiss from the Einasleigh Metamorphics north of Gum Flats analysed by Black et al. (2005) contained inherited zircons that these authors suggested were most likely derived from a number of sedimentary protoliths. They defined a maximum age limit of ca. 1700 Ma for the leucogneiss and its equivalents, based on the presence of substantial detrital populations at ca. 1700 Ma in some samples of leucogneiss. Four samples from their study contained inherited zircon populations that have some similarities to the inheritance patterns of the Daniel Creek Formation samples (Fig 4.17). Three leucogneiss samples contained no ages younger than 1700 Ma, and their detrital patterns were thought to reflect the contribution of many different source terranes. The remaining sample contained a significant concentration of younger ages which they related to at least three episodes of crystallisation and recrystallization between 1570 and 1500 Ma. The most obvious similarities between the two lithologies are the presence of a major ca. 2600 Ma component and ca. 1800-2000 Ma component, coupled with the relative lack of ages between 2000 and 2400 Ma in three samples (Fig 4.17). The similarity in the inheritance patterns between the Daniel Creek Formation samples of this study and the Einasleigh Metamorphics leucogneiss of Black et al. (2005) indicates that both lithologies have shared a similar provenance.

Inherited zircons were also reported in samples of Proterozoic granite from the Georgetown Inlier by Black & McCulloch (1990) and Black & Withnall (1993). As with the metasedimentary samples, the inherited zircons present a similar age distribution. Demonstrating this relationship is a sample of the Lighthouse Granite analysed by Black & Withnall (1993), which contains a number of inherited grains forming a similar inheritance pattern to that of sample 264, a sample of Dead Horse Metabasalt amphibolite (Fig. 4.18). The authors suggested that their inheritance

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patterns reflected the input of material from numerous sources into the source magma of the Lighthouse Granite. This hypothesis may also account for the pattern observed in sample 264, indicating the passage of mafic magmas though similar sedimentary rocks.



Figure 4.17: (a) Cumulative probability diagrams for four leucogneiss samples from the Einasleigh Metamorphics (after Black *et al.*, 2005). Also shown for comparison in (b) are detrital zircon age data for Daniel Creek Formation samples 184 and 255 (this study).

An alternative hypothesis that may account for the detrital populations between 2500 and 2000 Ma is that the inherited grains were sourced from now unexposed late-Archaean/early-Proterozoic basement rocks underlying the Etheridge Group, similar to that proposed for inherited populations in the Mt Isa Eastern Succession (Giles & Nutman, 2003). This would account for the regional extent of the ca. 2500-2000 Ma ages that have been obtained from the U-Pb zircon studies of numerous lithologies



across the region (e.g. Black & McCulloch, 1990; Black & Withnall, 1993; Black *et al.*, 2005; and this study).

Figure 4.18: Comparison of cumulative probability diagrams for (a) a sample of the Lighthouse Granite (Black & Withnall, 1993) and (b) sample 264 from the Dead Horse Metabasalt (this study).

The inheritance patterns of the Etheridge Group metasediments share a number of traits common to rocks of both the Broken Hill Group and Mt Isa Eastern Succession, including: (i) a significant late-Archaean to earliest Palaeoproterozoic component (ca. 3100-2500 Ma); (ii) an absent or relatively insignificant early Palaeoproterozoic component (ca. 2400-2100 Ma); and (iii) a significant mid-Palaeoproterozoic component (ca. 2100-1800 Ma). In the Broken Hill Group of the Willyama Inlier (Willis *et al.*, 1983) the spectrum of inheritance patterns listed above has been recognised in numerous samples of metasedimentary and gneissic rocks, including lithologies associated with and hosting the Pb-Zn-Ag orebody at Broken Hill (Fig. 4.19; Page & Laing, 1992; Raetz *et al.*, 2002; Page *et al.*, 2005a). However, the absence in the Daniel Creek Formation of the ca. 1800-1700 Ma detrital population

present in both the Broken Hill Group and Mt Isa Eastern Succession (Fig. 4.19 and 4.20 respectively), may also have important implications for the prospectivity of the Etheridge Group. In particular, that the Dead Horse Metabasalt-Daniel Creek Formation horizon within the Etheridge Group did not share the same provenance as the otherwise lithologically similar terranes further west.



Figure 4.19: Compilation of U-Pb SHRIMP detrital zircon analyses for four samples of the Hores Gneiss in the Broken Hill Group (Page & Laing, 1992). The dominant peak at ca. 1680 Ma represents the inferred crystallisation age of the host rock.

The metasediments of the Broken Hill Group were also found to contain a large abundance of detrital zircons aged between ca. 1870 and 1780 Ma (Page *et al.*, 2005a). These zircons were thought to have been sourced from terranes adjacent to the Willyama Inlier that contained rocks formed during the Barramundi Orogeny (ca. 1870-1850 Ma; Scott *et al.*, 2000; Page *et al.*, 2005a), and between ca. 1820 and 1780 Ma (Page *et al.*, 2005a). As with the metasediments of the Etheridge Group, the late-Archaean component present in the metasediments of the Broken Hill Group has been attributed to incorporation of detrital zircons from unexposed basement material into the metasediments (Page *et al.*, 2005b).

The analysis of metasedimentary and meta-igneous rocks from the Mt Isa Eastern Succession by Page & Sun (1998) and Giles & Nutman (2003) revealed detrital zircon age populations consistent with the inheritance patterns described above (Fig 4.20). In both cases their analyses showed broad similarities across the length of the terrane, as well as correlations with the detrital populations in metasediments from the rest of the

Mt Isa Inlier (Scott *et al.*, 2000). In particular, Giles & Nutman (2003) described two significant detrital populations occurring at ca. 2600-2300 Ma and ca. 1750-1720 Ma (Fig. 4.20). The older population of detrital ages overlaps with and may relate to the population thought to represent late-Archaean basement inheritance in both the Etheridge Group and Broken Hill Group. The younger detrital population is believed to have originated locally, as metasedimentary and magmatic rocks of ca. 1780-1720 Ma age are widely exposed in the Kalkadoon-Leichhardt Belt further west (Giles & Nutman, 2003). The absence of a detrital population at ca. 2000 Ma is significant, as this peak is present in the metasediments of the Etheridge Group. However, its absence could be accounted for if the source for this population was localised to the Etheridge Group, in a similar way to the localised sourcing of Barramundi-age material into the Willyama Inlier and Mt Isa Eastern Succession.



Figure 4.20: Cumulative probability diagrams for selected metasedimentary samples from the Mt Isa Eastern Succession. Data sources: (a) and (b) Giles & Nutman (2003), (c) and (d) Page & Sun (1998).

From the overview of inheritance patterns in the Palaeoproterozoic terranes of northeastern Australia outlined above, it is clear that a small but significant component that occurs throughout is the detrital population of late-Archaean age (between 3000 and 2500 Ma). This component has been inferred to represent the

addition of material from unexposed or eroded Archaean rocks that were basement to the Palaeoproterozoic terranes. A similar inheritance pattern was identified in the Belt Supergroup of western USA and southern Canada by Blewett *et al.* (1998), and may have important implications for reconstructions of the Rodinia supercontinent involving north Queensland (e.g. Ross *et al.*, 1992; Brookfield, 1993). The presence of this Archaean component throughout the Proterozoic basins of northeastern Australia, including the Georgetown Inlier, indicates that these terranes may have shared a common Archaean basement, and would therefore all lie within the Proterozoic framework of the North Australian Craton.

4.8 Conclusions

4.8.1 Magmatic history of the lower Etheridge Group

New igneous zircon ages obtained from mafic rocks of the lower Etheridge Group, combined with previously reported data, provide important geochronological constraints on the early depositional and magmatic history of the Georgetown Inlier. The mafic rocks were emplaced in at least two discrete pulses, with the most significant period of emplacement related to the eruption of the Dead Horse Metabasalt member and associated doleritic sills at ca. 1670 Ma.

The ca. 1670 Ma crystallisation and emplacement age of the Dead Horse Metabasalt and related Cobbold Metadolerite intrusions is based on the combination of pooled zircon ages in four samples (Samples 264, 299, 300, and 318). This age is of regional significance, as it closely resembles established ages for similar Fe-rich tholeiitic intrusive units in the Mt Isa Eastern Succession (Toole Creek Volcanics, 1658 \pm 8 Ma; Page & Sun, 1998) and felsic volcanics spatially and temporally associated with Fe-rich tholeiitic rocks in the Broken Hill Group (Hores Gneiss, 1685 \pm 3 Ma; Page *et al.*, 2005a). The significance of the association between these mafic units is discussed in more detail in subsequent chapters. Mafic magmatism in the Etheridge Group may have continued until ca. 1620 Ma, although evidence for this younger age is less robust than for the main ca. 1670 Ma event. Subsequent magmatic activity in the Palaeoproterozoic was limited to the intrusion of felsic granitoids associated with the Forsayth Batholith at ca. 1550 Ma. The oldest mafic rocks found in the basin crystallised at about 1715 Ma (1713 \pm 32 Ma and 1719 \pm 27 Ma). They were closely followed by volcanism recorded in zircons of a leucogneiss from the Einasleigh Metamorphics at 1705 Ma (Black *et al.*, 2005). Alternatively, these zircons may have been derived from the same juvenile volcanic source that provided zircons to the sedimentary precursor of the leucogneiss dated by Black *et al.* (2005). However, the tight grouping of isotopic ages for these clearly igneous zircons suggests they are unlikely to be inherited.

Late-stage felsic differentiates in two Cobbold Metadolerite sills yielded ages of ca. 1640 Ma. A late-stage felsic differentiate in a dolerite sill also crystallised at 1630 Ma and may form the last stage of this cycle of magmatism. The range of ages suggests a period of mafic magmatism spanning roughly 80 million years. It is not clear if this was a continuous event or was made up of a number of magmatic pulses of activity, however based on the distribution of pooled ages and the stratigraphic distribution of the mafic rocks the latter is most likely. The reported ages strongly support the inclusion of the lower Etheridge Group with the Calvert Superbasin (ca. 1730-1670 Ma) (Betts & Giles 2006). The ca. 1670 Ma Dead Horse Metabasalt may reflect the rejuvenation of the basin at the start of the Isa Superbasin cycle.

Isotopic zircon ages calculated for samples of the Mt Hogan and Digger Creek Granites yielded ages of 1565 ± 46 Ma and 1539 ± 16 Ma respectively. These ages are consistent with the established ages for both granites (Black & McCulloch, 1990; Black & Withnall, 1993), and place them within the timeframe of the emplacement of the regionally extensive Forsayth Batholith, and the associated low-pressure thermal event (M₂).

Inherited zircons were found in one mafic sample, revealing the presence of two significant age populations which may have been derived from adjacent metasedimentary rocks. The inherited sources include Barramundi-aged material (ca. 1800-1900 Ma) and a late Archaean-earliest Proterozoic component (ca. 2600 Ma). These provenance ages are of regional significance, as they are also found in metasediments from both the Etheridge Group and other Proterozoic basins of eastern Australia. No zircons were found that were coeval with deposition implying a very low component of volcanic activity during deposition of the Daniels Creek Formation.

4.8.2 Metamorphic history of the lower Etheridge Group

In the metapelites of the lower Etheridge Group, metamorphic monazite grew in response to a number of relatively distinct metamorphic and/or thermal events. A possible later stage of monazite growth was minor in comparison, and does not appear to be associated with a metamorphic fabric. There is little evidence for crystallisation of metamorphic monazite at ca. 1585 Ma, the previously proposed age for D_1 in these rocks. Only one sample analysed had a significant monazite population at this age and that sample also contained a much larger population of a younger age. The ca. 1585 Ma age is of regional importance, as it approximates the age of similar medium pressure-temperature metamorphism in the Mt Isa Eastern Succession and Broken Hill Block that occurred between ca. 1600 and 1580 Ma (Page & Sun, 1998; Page *et al.*, 2005a; Cihan *et al.*, 2006; Sayab, 2006). However, more work is required to constrain the age of this event in the lower Etheridge Group.

The majority of samples analysed in this study contain monazite formed at ca. 1530 Ma (1546 ± 11 Ma, 1537 ± 18 Ma, 1529 ± 17 Ma, and 1527 ± 16 Ma). These ages lie within error of the published ca. 1550 Ma age of high-temperature, low-pressure metamorphism (M₂) which is associated with the intrusion of voluminous granitoids throughout the lower Etheridge Group. Consequently they may represent crystallisation ages associated with this event. In some samples this age is related to monazite that is younger than S₁ but in at least two samples the monazites are aligned in the local S₁ and appear to be related to the earliest foliation in the area.

The results of this study, and to a lesser extent in Cihan *et al.* (2006), suggest a discrete younger event (D_3) dated by monazite at ca. 1500 Ma based on observed structural relationships in a small area west of Robinhood Station. The occurrence of this age may relate to the intrusion of a granitic body younger than the ca. 1550 Ma Forsayth Batholith, but this hypothesis remains to be tested. The ca. 1500 Ma age is coeval with granitoid magmatism in the Mt Isa Eastern Succession at ca. 1510-1500 Ma (Page & Sun, 1998), and may suggest a further link in the post-depositional histories of the two terranes.

4.8.3 The timing of deposition, and provenance ages in the lower Etheridge Group: implications for the regional tectonic history of the Georgetown Inlier

Based on the combination of new igneous and detrital zircon ages, and metamorphic monazite ages from a number of sources, the timing and duration of sediment deposition, and major periods of magmatism and metamorphism in the Etheridge Group can be identified. Deposition of the earliest sedimentary units had commenced by ca. 1715 Ma. At ca. 1670 Ma, significant volumes of tholeiitic mafic magmas were intruded into, and erupted onto the existing sedimentary package in response to continued or reinvigorated rifting. The tholeiitic magmatism continued as a probable series of discrete pulses until ca. 1630 Ma. The ca. 1670 Ma age of peak tholeiitic magmatism in the lower Etheridge Group is also broadly temporally equivalent to similar episodes of tholeiitic magmatism in the Mt Isa Eastern Succession and Broken Hill Group, which may suggest a regional correlation between the terranes at this time.

Regional-scale orogenesis involving the lower Etheridge Group is generally regarded to have reached its peak (M_1) at ca. 1580 Ma, coeval with medium pressure/temperature metamorphism in the Mt Isa Eastern Succession. However this was not the most common metamorphic age detected in this study. In response to the orogenesis and associated crustal thickening, voluminous S-type granitoids were emplaced at ca. 1550 Ma, forming a part of the regionally extensive Forsayth Batholith. The majority of crystallisation ages from monazites found in this study were between ca. 1550 and ca. 1525 Ma, and correlate with the period of hightemperature thermal metamorphism (M_2) reported previously. Later periods of elevated heat flow led to the development of monazites at ca. 1500 Ma, and appear restricted to west of Robinhood Station.

Detrital zircon populations in metasedimentary and mafic rocks of the lower Etheridge Group define a number of provenance ages, including two significant groups at ca. 2700-2500 Ma and ca. 2000-1800 Ma. Detrital patterns similar to those observed in the Etheridge Group metasediments are also found in metasedimentary rocks from the Mt Isa Eastern Succession and Broken Hill Group, and suggest that these sedimentary units had a similar provenance. The similarity in these provenance ages, coupled with the similar age for major periods of tholeiitic magmatism and

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metamorphism across the Georgetown Inlier, Mt Isa Eastern Succession, and Broken Hill Group suggest that all three terranes were proximal to one another, and most likely shared a similar tectonic history from ca. 1700 to 1500 Ma. This implies that the Georgetown Inlier was a part of the North Australian Craton during the Palaeoproterozoic.

Chapter 5.

Thermobarometric conditions of early- to mid-Proterozoic metamorphic events in the lower Etheridge Group

5.1 Introduction

Proterozoic rocks of the Georgetown Inlier display a range of metamorphic assemblages indicating significant variation in metamorphism across the inlier. In the central Georgetown Inlier, there is a general eastward increase in metamorphic grade, from lower greenschist facies through to upper amphibolite and locally granulite facies (Withnall, 1996). In areas of higher metamorphic grade, such as those of the amphibolite and granulite facies, the mineral assemblages are particularly suited to the application of geothermometry and geobarometry techniques.

In this chapter, quantitative determinations of the pressure and temperature conditions at the time of formation are presented for the major Mesoproterozoic metamorphic events that affected the lower Etheridge Group. In previous studies (e.g. Boger & Hansen, 2004; Cihan *et al.*, 2006) it has been suggested that the lower Etheridge Group experienced a period of high-temperature, medium-pressure, mid- to upperamphibolite facies metamorphism (M_1), followed by a period of retrogression, and then a low-pressure thermal event (M_2). The results of this study, coupled with EMPA metamorphic monazite dating (Chapter 4) will be used to test this hypothesis.

Later metamorphic events affecting the Georgetown Inlier have less well constrained ages, but were also less intense and generally formed only local crenulations in the dominant metamorphic fabric (Withnall, 1996). Metamorphic assemblages for these later events were not observed, and their thermobarometric characteristics are not presented here.

5.2 Previous work

Previous studies describe at least six phases of deformation that have affected the Etheridge Group. Of these, only the first two were considered to have occurred in conjunction with the major mid-Proterozoic prograde metamorphic events (Davis, 1995; Withnall, 1996; Boger & Hansen, 2004). The first event (M_1) has previously

been considered to involve medium-pressure, high-temperature conditions, ranging from lower greenschist facies in the west to upper amphibolite facies and locally granulite facies in the east of the terrane (Bell & Rubenach, 1983; Withnall, 1996). This event was followed by a low-pressure thermal event (M_2) that was associated with the emplacement of Mesoproterozoic granites, and also locally graded into upper amphibolite facies (Withnall, 1996). However, recent metamorphic and geochronological studies have subsequently questioned the separation of these two events (Boger & Hansen, 2004; Black *et al.* 2005).

The pressure conditions during mid-Proterozoic metamorphism in the Etheridge Group were considered to be intermediate between those of the classic low-pressure (Buchan) and medium-pressure (Barrovian) facies series (Withnall, 1996). Withnall (1996; and references therein) observed numerous features in the lower Etheridge Group that are characteristic of a medium-pressure terrane. These included the existence of a distinct biotite isograd in the metapelites, the co-existence of albite and hornblende in mafic rocks, and co-existing quartz and clinozoisite in amphibolite facies calc-silicate rocks.

A first attempt at determining the *P*-*T* conditions for M_{1-2} in the Etheridge Group was made by Withnall (1996) using a petrogenetic grid constructed from the plotting of experimental equilibrium curves and stability fields of minerals in pelitic rocks. Using this method, amphibolite facies *P*-*T* estimates for the Etheridge Group were estimated to range from lower amphibolite facies conditions of 3.5 kbar and 530 °C, to upper amphibolite-granulite facies conditions as high as 5.0 kbar and 820 °C (McNaughton & Wilson, 1983; Withnall, 1996).

More recent work has attempted to constrain the *P*-*T* conditions during the peak of metamorphism for both M_1 and M_2 . Boger & Hansen (2004), using mineral textures coupled with thermodynamic modelling, indicated that the M_1 and M_2 events were separate and unrelated. They determined peak *P*-*T* conditions for M_1 of 8.0-9.0 kbar and 750-800 °C from metapelites at Einasleigh. However, they noted that rocks of the Robertson River area only reached 7.0-8.0 kbar and 600-650 °C during M_1 . They also regard M_2 as a low-pressure thermal event, caused by the emplacement of the Forsayth Batholith and associated granitoids. *P*-*T* conditions at the time of M_2 in rocks of the Robertson River area were determined to be 3.0-4.0 kbar and 600-650 °C.

Further constraints on the *P*-*T* conditions for M_1 and M_2 were presented by Cihan *et al.* (2006) from four samples of amphibolite grade Robertson River Metamorphics metapelites near Robinhood. They analysed FIA (Foliation Intersection Axis) trends in garnet and staurolite porphyroblasts, as well as average *P*-*T* conditions and monazite ages from the metapelites. Their results produced *P*-*T* estimates for M_1 that correlate with those of Boger & Hansen (2004); however in addition to the two established prograde metamorphic events, they described a period of retrogressive metamorphism immediately following M_1 . Based on monazite ages, they determined that this metamorphic 'event' occurred over an approximate 45 Ma period, over which time the thickened orogenic pile was exhumed and weakened, forming crustal rocks amenable to granite intrusion at ca. 1550 Ma.

5.3 Objectives of the thermobarometric study

Although previous studies over the past two decades have touched on the Proterozoic metamorphic history of the Georgetown Inlier, few authors have directly studied the pressure and temperature conditions at the time of formation for mineral assemblages in metapelites of the lower Etheridge Group. Therefore, it is the objective of this chapter to:

- Determine the pressure and temperature conditions of metamorphism for a number of amphibolite facies-grade samples from the central Georgetown Inlier.
- Where possible, correlate these conditions with ages determined from U-Th-Pb metamorphic monazite dating in Chapter 4.
- Synthesize the early- to mid-Proterozoic thermobarometric history of the lower Etheridge Group.

Details of the sampling and analytical techniques used and additional petrographic and mineralogical information on samples analysed is provided below. A detailed overview of the petrology, mineralogy, stratigraphic, and textural relationship of the metasedimentary units referred to in this chapter is given in Chapter 3.

5.4 Method

Electron Microprobe (EMP) analysis of metamorphic mineral assemblages combined with a number of directly calibrated geothermometers and geobarometers are used to determine the pressure-temperature conditions of the earliest two metamorphic events to have affected rocks from higher grade areas of the lower Etheridge Group. EMP analysis has proved successful in determining the metamorphic conditions of a number of terranes that have reached varying levels of metamorphic grade, though it is generally best suited to terranes that have experienced metamorphic conditions of at least amphibolite facies, and where clear equilibrium assemblages can be obtained (e.g. Clarke *et al.*, 2000; Boger & Hansen, 2004; Fitzherbert *et al.*, 2004).

5.4.1 Sample selection

Metasedimentary samples from the lower Etheridge Group were selected on the basis that they contained mineral assemblages that appeared representative of metamorphic events. First preference was given to samples that also contained dated metamorphic monazite grains. Also analysed were metasedimentary samples of similar grade from the same localities that also possess mineral assemblages in apparent textural equilibrium. The locations of all samples analysed are presented in Figures 5.1 and 5.2.

5.4.2 Analytical technique

In situ microprobe analyses of individual minerals were obtained using the Cameca SX-100 EMP housed at the University of Tasmania. Data was acquired using five wavelength dispersive spectrometers (one TAP, one PET, two LPET, and one LLIF) at operating conditions of 15 kV accelerating voltage, 20 nA electron beam and a beam diameter of 5 μ m. PAP data reduction was performed using software supplied by Cameca. Analyses with total weight-percent oxide in the range 98-101 % were accepted as representing the true chemical composition of anhydrous minerals. Minerals with high Fe³⁺ content (e.g. epidote) and minerals with high contents of hydroxide, water, or volatile elements (e.g. biotite, hornblende, epidote, phengitic white mica, and chlorite) were tested against appropriate compositional ranges.



Figure 5.1: Simplified geological map of the central Georgetown Inlier. Phyllite/schist transition boundary in Etheridge Group metasediments is after Withnall (1996).

5.4.3 Data reduction and presentation

Cations per formula unit for each mineral analysed were calculated from weightpercent oxide totals obtained from the microprobe (Table 5.1). A number of directly calibrated geothermometers and geobarometers were then used in the quantitative determination of *P*-*T* conditions for each mineral assemblage. The methods involve cationic exchange and net transfer equilibria, as well as the average *P*-*T* method of THERMOCALC (Powell & Holland, 1988). The average *P*-*T* method involves the determination of all end-member reactions that can be calculated for a mineral assemblage, based on the internally consistent thermodynamic dataset of Holland & Powell (1998), and the calculation of the most favourable conditions of formation within a specified temperature or pressure window. An advantage of using THERMOCALC is that error propagation is applied to all calculations, thereby providing the user with uncertainties on all *P*-*T* estimations. THERMOCALC also incorporates the largest possible number of phases, which becomes useful when temperature estimates vary significantly within an individual sample. All errors quoted for THERMOCALC outputs are at 95% confidence. Representative THERMOCALC outputs and results are listed in Appendix 3.



Figure 5.2: Detailed geological map of the area surrounding Gum Flats station (modified after Withnall *et al.*, 2003a).

The computer program AX (Holland & Powell, 2000) was used to estimate ferric iron content for each of the microprobe analyses used in thermobarometric calculations. This involved the recalculation of Fe^{2+} as determined by microprobe analysis to achieve perfect stoichiometry. Mineral end-member activities required for average *P*-*T* calculations using THERMOCALC were also determined by AX. The chemistry of individual minerals is detailed below, and the analyses of all minerals used in

22.26 27.14 11.02 0.14 chl 0.05 0.05 0.34 0.02 0.51 0.01 0.03 86.3 6.6 0.0 5.69 D.01 4.93 0.06 3.57 0.01 82 18.96 2.82 42.58 ilm 3.23 0.00 1.45 0.41 0.01 0.06 0.76 100.3 0.16 1.80 0.16 0.00 1.74 0.06 0.03 0.00 0.01 0.05 3.99 9 51.91 0.49 43.64 3.34 0.00 0.10 00.1 0.00 1.83 0.14 0.00 0.00 0.00 ilm 0.00 0.01 0.03 .96 0.03 0.01 0.61 0.01 9 5 58 0.31 34.89 9.78 93.66 0.00 0.14 0.00 nm 0.02 0.02 0.70 00.0 0.92 556 00.0 0.16 0.24 l.69 3.99 1.44 0.03 5 33.94 94.36 nm 0.30 0.01 0.00 9.39 0.00 0.00 0.15 0.00 0.23 3.91 2.07 0.76 0.03 0.88 30 0.03 5.36 0.23 l .61 5 20.43 24.20 10.11 38.80 0.08 0.19 0.03 0.26 3.39 5.44 i9 0.08 0.03 2 0.02 3.94 0.00 3.31 0.04 2.47 0.01 0.01 0.71 19.28 25.72 245 0.10 0.01 1.91 7.90 0.45 0.02 2.27 90.11 0.00 3.48 0.26 1.91 0.08 0.4715.14 3.68 0.01 Ē 5 53 0.01 21.06 gt[r] 35.86 0.02 33.06 0.06 00.6 0.18 0.12 0.05 0.66 1.32 0.02 0.00 2.26 0.53 0.01 00.0 8.06 .48 12 .93 00.0 2.03 35.98 21.01 0.02 33.06 0.19 ಹ 0.07 7.81 .54 .30 0.02 0.00 100.8 2 .93 00.0 203 0.00 2.25 0.54 0.11 0.00 0.00 3.05 21.17 101.2 36.03 0.02 32.92 8.10 0.04 0.00 0.56 0.20 0.10 0.06 1.66 1.20 0.00 2.03 0.00 2.24 0.01 0.00 8.06 ы 2 2.93 Garnet traverse gt[c] 21.02 35.76 0.07 0.02 00 0 7.86 0.03 0.00 00.6 0.00 0.54 0.21 0.11 0.00 1.73 1.24 6 00.0 2.02 2.25 0.00 8.07 12 gt[c] 35.71 20.97 33.33 0.50 0.20 0.01 0.03 7.19 1.65 1.26 0.03 0.01 00.2 2 2.93 00.0 2.03 0.00 2.29 0.11 0.01 0.00 8.06 36.17 20.80 0.11 0.04 3 51 0.480.19 0.00 0.01 6.95 1.57 1.24 0.02 0.03 00.3 2 00.0 2.00 0.00 2.29 0.00 20.8 ಕ gt[r] 35.51 20.90 0.01 0.03 3.56 0.50 0.18 0.12 7.20 1.48 0.05 0.03 2.03 0.00 2.31 0.01 0.00 8.07 1.33 00.1 2 00.0 35.61 20.85 34.01 2.33 0.200.12 0.00 0.04 0.01 6.75 2 1.39 0.00 0.01 00.3 12 2.92 0.00 2.02 0.00 0.47 0.00 3.07 ы 36.22 20.93 32.26 0.04 0.00 00.6 288 gt 3.06 12. 1.31 0.04 0.01 2 .95 00.0 0.01 0.00 2.20 0.56 0.21 0.11 0.01 0.00 8.05 5 30 32.74 10.30 0.22 0.43 92.27 0.02 0.12 4.02 0.01 2.08 0.14 1.02 0.03 0.02 5.33 00.0 0.24 0.21 0.01 1.82 nm 2 44.80 33.88 10.70 93.44 4.09 nu 0.44 0.02 2.16 0.50 0.17 0.00 0.13 60.0 0.00 0.00 0.25 580 22 5.47 0.01 1.87 15.32 18.90 15.10 3.52 79.03 2.15 0.16 18. 0.01 0.16 g 0.40 0.46 0.00 2.86 0.03 0.08 0.81 ġ 2 5.51 0.25 3.26 29.45 17.84 \$5.75 1.85 0.01 25.51 0.28 0.19 0.23 2.63 3.60 0.00 3.65 0.04 1.98 0.03 0.08 0.58 5.24 ā TT.T 5 5.04 0.24 gt[r] 36.19 21.04 33.09 99.25 0.01 0.00 0.02 0.39 0.26 0.10 0.00 0.01 5.65 2.17 1.08 0.00 0.00 8.02 2 00.0 2.03 2.27 6. gt[c] 36.11 20.93 33.55 99.37 0.00 0.00 5.53 2.20 .06 0.00 0.00 0.00 2.30 0.38 0.27 0.09 0.00 0.00 12 .96 00.0 203 .03 35.75 20.73 0.00 0.04 32.95 5.91 2.25 1.01 0.01 0.00 98.66 2 26.95 00.0 2.02 0.00 2.28 0.41 0.28 0.09 0.00 0.00 8.03 50 20.99 36.02 0.00 0.00 32.86 5.80 0.03 98.95 0.40 2.31 0.95 0.01 00.0 2.03 0.00 2.26 0.280.08 0.00 0.00 3.03 283 0 .96 50 61.76 0.00 22.55 0.00 0.08 0.00 0.00 0.15 7.61 0.00 0.00 0.00 0.00 0.19 0.80 9.05 80 00.0 1.20 5.00 Ъ 4.01 œ 0.01 21.98 51.15 0.12 95.96 0.18 0.00 0.00 0.21 0.02 0.02 3.61 8.85 00.0 1.19 0.00 0.01 0.00 0.00 0.79 0.01 1.99 Ы × 2.81 33.11 0.55 91.56 3 00 0.16 0.00 4.04 nu 0.63 0.02 2.17 0.03 0.00 0.28 0.00 0.25 0.00 0.07 .88 5 C0.0 6.45 3 70 19.34 90.19 1.75 9.06 7.53 5.38 0.02 8.12 0.20 0.00 0.03 1.93 0.04 0.06 ġ 0.24 0.24 52 5.38 0.21 3.64 2.55 1.53 18.61 89.23 33.21 1.94 0.00 21.21 0.32 0.23 5.89 2.87 0.03 1.83 0.05 0.07 1.22 0.25 7.58 52 .38 24 3.56 0.00 ā gt[r] 35.95 20.96 99.10 32.43 00.0 0.05 5.36 0.00 0.44 0.29 0.00 2.35 96°C 10.C 12 :95 00.0 2.03 0.00 2.23 0.09 0.00 analysis; [r], rim analysis 99.12 gt[c] 20.76 36.04 33.11 5.79 2.40 0.03 0.01 0.40 0.29 0.08 0.00 0.00 0.01 0.01 0.97 00.0 2.01 0.00 2.27 2 36.00 20.67 0.00 0.00 100 0.00 0.26 0.00 0.00 5.60 1.04 10.C 0.00 0.46 0.09 8.02 ಹ 12 10.3 2.22 20.62 282 35.75 0.00 0.00 LC C 6.71 11 0.96 0.03 0.00 98.51 12 00.0 2.01 0.00 2.23 0.47 0.27 0.09 0.00 0.00 8.04 ಕ್ಷ ox. [c], core a Mineral Sample number FeO^{tota} Cr_2O_3 Al,O, MnO MgO NaO K_2O Total SiO₂ ΠÖ, CaO Total

geothermometry and geobarometry calculations in this chapter are listed in Table 5.1. Mineral abbreviations are after Kretz (1983), and Holland & Powell (1998).

Table 5.1: Representative electron microprobe analyses.

5.5 Metamorphic Petrology

Etheridge Group sedimentary rocks that have been metamorphosed to amphibolite facies conditions are located in the areas surrounding Gum Flats and Robinhood stations, and include the Einasleigh Metamorphics, Bernecker Creek Formation, Daniel Creek Formation, and Corbett Formation members (Fig. 5.1 and 5.2). These rocks contain variable amounts of quartz, biotite, muscovite, garnet, and more locally plagioclase feldspar, epidote, staurolite, and calcite. Aluminosilicate minerals were not observed, most likely due to the period of retrogression following peak metamorphism rather than unfavourable bulk rock compositions. Chlorite was observed in numerous samples as a retrograde mineral pseudomorphing large porphyroblasts, possibly of either an aluminosilicate mineral or staurolite.

Timing relationships between minerals are implied from the presence of inclusions, the growth of one mineral on or around another, or via the formation of overgrowths/coronas of finer-grained minerals along the margins of coarser-grained ones. Minerals found as inclusions, or minerals partially or wholly enveloped by other minerals, are implied to precede those that enclose them. The notation of metamorphic events and their associated structural fabrics (i.e. M_1 , S_1 , etc.) follows the notation used by Withnall (1996) and Withnall *et al.* (1997), except where indicating the retrograde event that followed M_1 , which is denoted M_{1R} .

Sample 306, a muscovite-biotite-garnet-staurolite schist belonging to the Corbett Formation was collected from the Robinhood area (Fig. 5.1). The sample contains small aligned plates of biotite and muscovite which together with elongate quartz grains define a dominant schistosity that trends NW-SE and corresponds to the D_3 event of Cihan *et al.* (2006). Porphyroblasts of garnet and staurolite are enclosed by the foliation, forming large strain shadows containing biotite, muscovite and granoblastic quartz (Fig. 5.3a). Porphyroblasts of both garnet and staurolite are inclusion-rich and of variable size, however staurolite porphyroblasts are generally larger and occasionally enclose or partly enclose small garnet porphyroblasts (Fig. 5.3b). Garnet porphyroblasts are anhedral to subhedral in shape and have diameters ranging up to 1 mm. Within garnet, small elongate quartz inclusions define an earlier (S₁) foliation that is commonly orthogonal to the external schistosity (Fig. 5.3c). Examples where the internal foliation of the quartz inclusions trail morphology is

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more difficult to recognise in staurolite due to the larger size of the quartz inclusions (Fig. 5.3b), however S_1 is still partially visible within some larger grains. The staurolite porphyroblasts are generally subhedral to euhedral, and are between 2 and 5 mm in diameter. Monazite grains lying within S_3 assemblage minerals record a metamorphic age of 1497 ± 20 Ma, which I infer to date this event (see Chapter 4).



Figure 5.3: Textural relationships preserved in sample 306 of the Corbett Formation from the Robinhood station area. (a) Photomicrograph of biotite-muscovite-quartz strain shadows on garnet and truncation by external foliation. (b) Photomicrograph of staurolite porphyroblast containing small garnet inclusion. (c) Photomicrograph of garnet porphyroblast containing inclusion trails of quartz. Note the high angle of the inclusion trails to the external schistosity.

All remaining samples were collected from the Gum Flats area (Fig. 5.2). They are compositionally similar, with relative variations in the abundance of the main rockforming minerals, and the level of metamorphism attained being their key differences. Samples 282 and 283 are biotite-muscovite-garnet-bearing schists from the Einasleigh Metamorphics. The rocks contain small aligned plates of biotite and muscovite which define the main layer-parallel schistosity (S_1). In both samples, quartz, along with minor amounts of variably sericitised plagioclase constitute the remainder of the matrix. Garnet porphyroblasts are more common in sample 282, whereas muscovite is more common and garnet less common in sample 283. Garnet is the only porphyroblast-forming mineral in both samples, and is commonly anhedral to subhedral with diameters ranging up to 1 mm. Small, randomly aligned laths of biotite, blocky plagioclase, and quartz commonly surround garnet (Fig. 5.4a). No strain shadows surrounding garnet porphyroblasts, or inclusion trails within porphyroblasts, are recognised in either sample. Monazite grains lying between and within S_{1-2} assemblage minerals in sample 283 record a metamorphic age as reported in Chapter 4. The primary metamorphic age was determined from 13 grains as being 1537 ± 18 Ma, which lies at the younger end of the inferred age spectrum for deformation and associated peak metamorphism in the Etheridge Group (c. 1580-1550 Ma; Withnall, 1996; Black *et al.*, 2005).



Figure 5.4: Textural relationships preserved in metapelite samples from the Gum Flats area. (a) Photomicrograph of matrix biotite, muscovite, and quartz surrounding garnet porphyroblast (sample 282). (b) Photomicrograph of inclusion trail in garnet porphyroblast, orthogonal to the external schistosity; surrounding and replacing garnet are low-strain biotite plates (sample 288).

(c) Photomicrograph of chlorite porphyroblast surrounded by biotite rims and containing inclusion trails of quartz. Note the high angle of the inclusion trails to the external schistosity (sample 288).
 (d) Photomicrograph of garnet porphyroblast with poikiloblastic texture in Bernecker Creek Formation calc-silicate (sample 293).

Sample 288 is a biotite-muscovite-garnet-bearing schist from the Daniel Creek Formation. It contains small aligned laths of biotite and muscovite which again define the dominant layer-parallel schistosity (S_1) . Small flattened grains of quartz constitute the remainder of the matrix. Porphyroblasts of garnet and chlorite are enclosed by this foliation, and form large symmetrical strain shadows containing biotite and blocky quartz (Fig. 5.4b and 5.4c). The garnet porphyroblasts are anhedral to subhedral, and are less than 1 mm in diameter. They are also commonly enclosed by thin coronas of biotite. Inclusion trails of quartz are found within most garnet porphyroblasts. Their orientation varies from sub-parallel, to more commonly orthogonal to the external schistosity. Other large porphyroblasts occurring throughout the sample were of a prograde metamorphic mineral, possibly staurolite, which has subsequently been pseudomorphed by retrograde chlorite. The porphyroblasts are commonly tabular, anhedral, and range up to 3 mm in diameter. They contain inclusion trails of quartz which commonly define a foliation orthogonal to the matrix foliation (Fig. 5.4c). Six analyses on five monazite grains lying within the dominant S_1 schistosity record a metamorphic age of 1585 ± 18 Ma (Chapter 4), which I associate with the pervasive M₁ event.

Sample 293 is a biotite-muscovite-garnet-bearing schist from the Bernecker Creek Formation. In this rock, biotite again defines the dominant foliation (S_1); however the foliation is weaker relative to previous samples. Small, blocky to elongated grains of plagioclase, and lesser quartz and calcite define the remainder of the matrix, suggesting the sedimentary precursor of this rock may have been calc-silicate in composition. Fine-grained epidote is accessory. Coarse-grained garnet forms inclusion-rich porphyroblasts of variable size surrounded by fine-grained biotite and blocky quartz grains. The garnet porphyroblasts are anhedral to subhedral and have diameters ranging between 1 and 3 mm. Within garnet, numerous small elongate quartz inclusions define a foliation that is generally consistent with the matrix foliation, giving the garnet a poikiloblastic texture (Fig. 5.4d).

Sample 312 is a garnet-hornblende-bearing calc-silicate gneiss from the Daniel Creek Formation. The sample contains garnet-hornblende-quartz-plagioclase as the peak metamorphic assemblage along with calcite, with secondary actinolite and accessory clinozoisite. Small, elongated grains of hornblende define a weak layer-parallel foliation, and are the most common form of hornblende in this sample. Fine-grained

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blocky quartz and plagioclase up to 1 mm in length constitute the remainder of the matrix. Both garnet and hornblende form porphyroblasts in this sample, with garnet more numerous. Garnet porphyroblasts have diameters up to 4 mm, are anhedral to subhedral, and contain large inclusions of quartz. Hornblende porphyroblasts have diameters up to 2 mm, are commonly anhedral, and also contain numerous large quartz inclusions similar to garnet.

5.6 Mineral Chemistry

This section describes the mineral chemistry for each sample analysed. Iron values for all analyses were obtained as Fe^{2+} , except for epidote group minerals in which all iron was assumed to be Fe^{3+} . Chemical compositions of minerals analysed by electron microprobe and selected for thermobarometric analysis are shown in Table 5.1 and are summarized in Table 5.2.

	282	283	288	293	306	312
Garnet						
X _{alm}	0.75-0.72	0.76-0.74	0.77-0.71	0.43-0.38	0.78-0.71	0.44-0.29
X _{spss}	0.17-0.13	0.15-0.13	0.19-0.12	0.37-0.29	0.13-0.06	0.23-0.15
X _{gr}	0.04-0.03	0.03	0.04-0.03	0.25-0.20	0.08-0.06	0.49-0.37
X_{py}	0.10-0.07	0.10-0.08	0.09-0.05	0.03	0.10-0.06	0.03-0.02
X _{Fe}	0.92-0.87	0.90-0.89	0.93-0.89	0.93-0.92	0.92-0.88	0.96-0.93
Staurolite						
X_{Fe}					0.88-0.84	
Hornblende						
X _{Ca}						1.00-0.86
X_{Al}						0.88-0.09
Biotite						
X_{Fe}	0.64-0.54	0.65-0.55	0.64-0.55	0.57-0.54	0.64-0.46	
Muscovite						
X_{Fe}	0.61	0.59-0.52	0.87-0.54	0.68-0.57	0.63-0.47	
X_{K}	0.96	0.95-0.93	0.88-0.85	0.98-0.97	0.85-0.84	
Plagioclase						
X _{alb}	0.83-0.80		0.92			
X _{an}				0.72-0.62	0.67	0.70-0.42

Table 5.2: Summary of electron microprobe data listed in Table 5.1.

5.6.1 Silicates

5.6.1.1 Garnet

A single generation of garnet porphyroblast growth is recognised in sample 306 from the Robinhood area. Porphyroblasts from this sample form an almandine-rich solid solution. Individual porphyroblasts are weakly zoned (Fig. 5.5a), with average values of Alm₇₁Spss₁₃Py₈Gr₈ for cores, and Alm₇₆Spss₁₀Py₇Gr₇ for rims (Alm = almandine, Spss = Spessartine, Py = Pyrope, Gr = Grossular). End-member components between individual porphyroblasts show small but significant variation, with X_{alm} [Fe/(Fe + Mg + Mn + Ca)] varying from 0.71 to 0.78, X_{spss} varying from 0.06 to 0.13, and X_{py} varying from 0.06 to 0.10. The ratio of Fe:Mg [X_{Fe} = Fe/(Fe + Mg)] remains almost constant (X_{Fe} c. 0.89-0.91) across all analysed garnet grains implying that the grains grew over a relatively small temperature range, with rim temperatures (X_{Fe} = 0.89) generally cooler than core temperatures (X_{Fe} = 0.91; Fig. 5.5a). Relatively rapid decrease in X_{py} and associated increase in X_{alm} along some garnet rims (Fig. 5.5a), suggests diffusion zoning occurred, possibly involving adjacent biotite.

Garnet porphyroblasts from mica schists of the Einasleigh Metamorphics and Daniel Creek Formation in the south of the Gum Flats area (samples 282, 283, and 288) also possess similar almandine-rich compositions to those of the Robinhood area. In samples 282 and 283, a single generation of porphyroblast growth was recognised, with end-member compositions remaining relatively consistent between grains ($X_{alm} = 0.73-0.76$, $X_{spss} = 0.13-0.17$, $X_{py} = 0.07-0.10$, $X_{gr} = 0.03-0.04$). Minor zoning was observed across individual garnet porphyroblasts, with end-member component values of Alm₇₄Spss₁₃Py₁₀Gr₃ for cores and Alm₇₃Spss₁₅Py₉Gr₃ for rims. Garnet porphyroblasts in sample 288 are relatively more Mn-rich ($X_{spss} = 0.12-0.19$) than those of the Einasleigh Metamorphics samples, but otherwise show limited variation between grains in two end-member components ($X_{alm} = 0.71-0.76$, $X_{gr} = 0.03-0.04$), and greater variation in Mg ($X_{py} = 0.05-0.09$), suggesting growth occurred over a limited temperature range at near constant pressure. Individual porphyroblasts are weakly zoned with high X_{spss} cores (Fig. 5.5b), and high X_{alm} rims. X_{Fe} is also relatively constant ($X_{Fe} c. 0.91-0.92$) across all analysed grains.

Garnet compositions in the two remaining samples (samples 293 and 312) differ significantly from those described above. In both cases, the garnet porphyroblasts are significantly enriched in both Mn and Ca relative to Fe, suggesting that both samples were possibly derived from calcareous sediments, or that some amount of calcareous material was present in the sedimentary precursors of the rocks. The composition of garnet porphyroblasts in sample 293 varies from $X_{alm} = 0.38$ to 0.43, with variation mostly observed within individual grains, rather than between grains. Within-grain variation is observed between other end-members, particularly spessartine ($X_{spss} =$ 0.29-0.37), and grossular ($X_{gr} = 0.22$ -0.25). Significant enrichment in the grossular component in sample 293 ($X_{gr} = 0.22$ -0.25) relative to other metapelite samples detailed above ($X_{gr} = 0.03$ -0.04) also suggests a calc-silicate origin for the rock. Similar compositional variation is observed in sample 312, with X_{alm} varying from 0.29 to 0.44. The dominant compositional variation in this sample is again in the form of core to rim zoning (Fig. 5.5c), with representative end-member compositions of Alm₄₄Spss₁₅Py₃Gr₃₈ from a core analysis and Alm₂₉Spss₂₃Py₂Gr₄₆ from a rim analysis. X_{Fe} is also consistent across all analyses ($X_{Fe} c. 0.93$ -0.95).



Figure 5.5: Compositional zoning across garnet porphyroblasts. (a) Plot of values for traverse of garnet porphyroblast from sample 306. (b) Plot of values for traverse of garnet porphyroblast

from sample 288. (c) Plot of values for almandine and grossular components in traverse of garnet
porphyroblast from sample 312. (d) Plot of values for spessartine and pyrope components in
traverse of garnet porphyroblast from sample 312. In all plots: Alm = Almandine, Spss =
 Spessartine, Py = Pyrope, and Gr = Grossular.

5.6.1.2 Feldspar

Plagioclase is found in samples 282, 288, 293, and 312. In samples 282 and 288 plagioclase is an albite-rich solid solution with $X_{alb} = 0.81-0.83$ [$X_{alb} = Na/(Na + Ca)$] in sample 282, and $X_{alb} = 0.92$ in sample 288. Plagioclase in samples 293 and 312 consists of an anorthite-rich solid solution, which again suggests a calc-silicate origin for the rocks. Plagioclase compositions vary between $X_{an} = 0.62$ to 0.72 [$X_{an} = Ca/(Na + Ca)$] in sample 293, and $X_{an} = 0.42$ to 0.70 in sample 312.

5.6.1.3 Epidote group minerals

Epidote group minerals only occur in samples 293 and 312. Epidote in sample 293 is relatively Fe-rich, with $X_{\text{Fe}}^{3+} = 0.25 \ [X_{\text{Fe}}^{3+} = \text{Fe}^{3+}/(\text{Fe}^{3+}+\text{Al})]$. In sample 312, greater variation in Fe³⁺ and Al is observed, with X_{Fe}^{3+} ranging from 0.15 to 0.23.

5.6.1.4 Staurolite

Staurolite is only identified in sample 306. Compositional variation is limited within staurolite, with $X_{Mg} [X_{Mg} = Mg/(Mg + Fe)]$ for all analyses falling within the range of 0.84 to 0.88. Mg also shows some minor variation between core ($X_{Mg} = 0.16$) and rim analyses ($X_{Mg} = 0.14$), and may reflect some compositional heterogeneity or zoning. Zn values were not measured in this study, but are likely to be low given very low Na values. Chlorite porphyroblasts in sample 288 may have originally been staurolite, but were completely pseudomorphed by subsequent retrogression.

5.6.1.5 Amphibole

Sample 312 is the only analysed sample containing amphibole. Two compositionally distinct Ca-rich amphiboles are identified in this sample. 'Common' hornblende is found as large phenocrysts throughout, often associated with garnet with which it defines the prograde metamorphic mineral assemblage. In this sample, hornblende

shows some compositional variation between individual grains, with compositions ranging between magnesio-hornblende ($X_{Ca} = 0.97-0.88$ [$X_{Ca} = Ca/(Ca + Na)$]; $X_{Al} = 0.17-0.50$ [$X_{Al} = Al/(Al + Mg)$]), ferro-hornblende ($X_{Ca} = 0.89$; $X_{Al} = 0.49$), pargasite ($X_{Ca} = 1.00$; $X_{Al} = 0.88$) and ferro-pargasite ($X_{Ca} = 0.86$; $X_{Al} = 0.54$). Secondary amphibole in the form of actinolite was also identified, and has a narrow compositional range ($X_{Ca} = 0.98-0.97$; $X_{Al} = 0.15-0.09$). In addition to the hornblende porphyroblasts, both hornblende and actinolite occur as small, elongate grains within the matrix of the sample.

5.6.1.6 Biotite

Biotite occurs in all samples with the exception of sample 312. Despite its occurrence in all mica schist samples, it has a moderate compositional range. In the Gum Flats area, X_{Fe} values for biotite lie within the range of 0.54 to 0.65 for all samples, and it is possible that more than one stage of biotite growth in present. In sample 306 from the Robinhood area, biotite shows a large compositional range, with X_{Fe} falling between 0.46 and 0.64. As with the samples from Gum Flats, it is possible that more than one stage of biotite growth is present in this sample. However, Fe-Mg exchange between biotite and adjacent garnet ($X_{\text{Fe}} = 0.92$ -0.88) may in part account for the compositional variation. TiO₂ values for biotite across all samples vary significantly, ranging from less than 1 wt% in sample 288 through to up to 3 wt% in samples 283 and 293.

5.6.1.7 White Mica

White mica in all samples is muscovite, with Fe and Mg constituting up to 0.4 cations per formula unit based on 22 oxygens. Si mostly constitutes 6.0-6.3 cations per formula unit (22 oxygens); however phengitic compositions are not observed, as the ratio of Si to Al is typically close to 6:5. Fe in most samples is in excess of Mg with measured X_{Fe} values typically between 0.5 and 0.7. The exception is two analyses in sample 306 which have X_{Fe} values of 0.47. Substitution of Na for K is low in most samples with X_{K} [$X_{\text{K}} = \text{K}/(\text{K} + \text{Na})$] typically between 0.93 and 0.98. Muscovite in samples 288 and 306 both contain lower K:Na, with X_{K} varying between 0.85 and 0.88 in sample 288, and between 0.84 and 0.85 in sample 306. Na typically constitutes less than 0.15 cations per formula unit (22 oxygens) in samples with high K:Na ratios (greater than 0.90), but varies between 0.2 and 0.3 cations per formula unit in samples 288 and 306.

5.6.1.8 Titanite (Sphene)

Sphene is only found in sample 312. All analyses contain between 3.49 and 3.69 Ti, 3.95 and 3.99 Ca, and between 0.35 and 0.61 Al cations per formula unit (20 oxygens).

5.6.1.9 Chlorite

Chlorite was identified and analysed in samples 288 and 306. The composition of chlorite in both samples lies within the range of ripidolite (after Deer *et al.*, 1992), with X_{Fe} values of 0.50 and 0.58, and 5.14 to 5.40 Si cations per formula unit (28 oxygens).

5.6.2 Non-silicates

5.6.2.1 Ilmenite

Ilmenite is only found in sample 288. Individual analyses are compositionally indistinguishable within the sample, containing on average 2 cations per formula unit (4 oxygens) of both Ti and Fe.

5.6.2.2 Rutile

Rutile is only found in sample 306. Although essentially consisting of TiO_2 , analysed rutile contains minor amounts of both Al (up to 2 wt%) and Fe (up to 3 wt%).

5.7 Thermobarometry results

Temperature estimates were made using five geothermometer calibrations. These included the average *P-T* (pressure-temperature) method of THERMOCALC (Powell & Holland, 1988), the garnet-biotite exchange thermometers of Ferry & Spear (1978), Perchuck & Lavrent´eva (1983), and Bhattacharya *et al.* (1992), and the garnet-hornblende exchange thermometer of Graham & Powell (1984). Across all analysed

samples containing biotite, reasonable correlation between garnet-biotite thermometer methods was observed. However, temperature and to a lesser extent pressure estimates obtained using the average *P*-*T* method of THERMOCALC returned significant differences. THERMOCALC temperature estimates were in the order of 50-100 °C higher than those obtained using other methods (Table 5.3). However, as all THERMOCALC temperature estimates were obtained from independent sets of reactions and contain low associated errors, the estimates obtained are considered reliable, and may therefore reflect maximum recorded temperature conditions. Of the three remaining garnet-biotite thermometers, the temperature estimates obtained using the method of Perchuck & Lavrent´eva (1983) are preferred, based on a recent assessment of the reliability of numerous garnet-biotite thermometer calibrations (Wu & Cheng, 2006). Subsequently, only temperature estimates produced using this method are discussed below and presented in Table 5.3, along with estimates obtained from THERMOCALC average *P-T* calculations.

Pressure estimates using net-exchange barometers were limited due to unfavourable mineral assemblages in biotite-grade samples. Subsequently only two geobarometers were used. These were the average *P-T* method of THERMOCALC, and for sample 312 the net-transfer barometer of Kohn & Spear (1990). Pressure estimates using THERMOCALC were obtained, with varying degrees of success, from all samples except sample 283. Pressure estimates obtained using THERMOCALC for samples containing biotite (all samples except 283 and 312) have relatively high errors (Table 5.3). For sample 312, where more than one method for estimating pressure was used, the THERMOCALC estimates also had relatively high calculated errors. However, in this sample the net-transfer barometer estimates lie within error, or correlate with THERMOCALC estimates (Table 5.3), and are therefore considered reliable.

P-T estimates in metapelites of lower- to mid-amphibolite grade from both the Gum Flats area (samples 282, 283, 288, and 293) and the Robinhood area (sample 306) are limited by the dominant mineral assemblage garnet-biotite-muscovite-quartz. Application of four calibrations of the garnet-biotite thermometer (Perchuk & Lavrent`eva, 1983; Ferry & Spear, 1978; Bhattacharya *et al.*, 1992) to these samples was reasonably successful, as estimates for M₁ were made by pairing almandine-rich garnet porphyroblast cores with adjacent S₁ biotite grains (e.g. Fig. 5.3a). In samples 282 and 283, temperature estimates for M₁ conditions calculated using the method of Perchuk & Lavrent'eva (1983) vary between 522 and 558 °C, with assumed pressures between 4 and 6 kbar (Table 5.3). Slightly higher temperature estimates were obtained for S₁ assemblages in sample 288, at an assumed pressure of 5.5 kbar (566 to 568 °C). Temperature estimates for mineral assemblages in sample 288 that returned average *P-T* conditions analogous to M₂ were obtained using the garnet-biotite thermometer of Perchuk & Lavrent'eva (1983), and returned estimates of 483 and 484 °C at an assumed pressure of 3 kbar (Table 5.3).

			Assumed			ted Result	
Sample	Lithology (GDA94 grid co-ordinates)	Mineral Assemblage	P (kbar)	T (°C)	<i>T</i> (°C)	P (kbar)	Method
282	Einasleigh Metamorphics	gt-bi-mu-pl-qz	-	-	$726 \pm 45, 768 \pm 75$	$6.9 \pm 1.5, 7.3 \pm 2.3$	1
	0806720 7880664	(M1 assemblage)	4	-	551, 558		2a, 2b
	(Easting Northing)		6	-	522, 522		2a, 2b
283	Einasleigh Metamorphics 0806700 7880664	gt-bi-mu-pl-qz (M1 assemblage)	4.5	-	557, 552		2a, 2b
288	Daniel Creek Formation	gt-bi-mu-pl-qz	-	-	642 ± 31	7.6 ± 1.3	1
	0805845 7881526	(M1 assemblage)	5.5	-	568, 566		2a, 2b
		gt-bi-mu-chl-qz	-	-	557 ± 117	3.1 ± 1.7	1
		(M2 assemblage)	3		484, 483		2a, 2b
293	Bernecker Creek Formation 0803330 7884267	gt-bi-ep-pl-mu-qz (M1 assemblage)	-	-	624 ± 69	6.3 ± 1.8	1
		gt-bi-ep-pl-mu-qz	-	-	559 ± 35	5.1 ± 1.0	1
		(M1R assemblage)	5.5	-	426, 424		2a, 2b
			6.5	-	418, 419		2a, 2b
306	Corbett Formation	gt-bi-mu-st-pl-qz	-	-	574 ± 18	6.3 ± 2.5	1
	0766403 7908517	(M1 assemblage)	6.5		523, 526		2a, 2b
312	Daniel Creek Formation	gt-hbl-pl-qz	-	-	649 ± 69, 631 ± 69	$7.1 \pm 1.8, 6.4 \pm 1.7$	1
	0800551 7888325	(M1 assemblage)	-	650		6.5	4
			6	-	659		3
		gt-act-pl-cz-qz	-	-	$562 \pm 51, 564 \pm 60$	$5.3 \pm 1.4, 5.5 \pm 1.6$	1
		(M1R assemblage)	-	550		5.2	4
			5.5	-	512		3

Table 5.3: Pressure and temperature estimates calculated from a variety of thermobarometric techniques to the mineral assemblages of selected samples from the lower Etheridge Group metasediments. Methods: 1- THERMOCALC average *P-T* method (Powell & Holland, 1988); 2a-Perchuk & Lavrent`eva (1983) (Perchuk *et al.*, 1981); 2b- Perchuk & Lavrent`eva (1983) (Hewitt & Wones, 1975); 3- Graham & Powell (1984); 4- Kohn & Spear (1990). Mineral abbreviations after Kretz (1983). THERMOCALC output files are listed in Appendix 3.

The application of the THERMOCALC average *P*-*T* method to S₁ and retrograde mineral assemblages in sample 282 and 288 was successful, and is illustrated best by the clear separation of *P*-*T* estimates for both M₁ and M₂ in sample 288 (Table 5.3). Subsequently, estimates made using THERMOCALC for these samples are considered to best represent the maximum *P*-*T* conditions reached during each event. For sample 282, THERMOCALC returns average *P*-*T* conditions of 6.9 ± 1.5 kbar

and 726 ± 45 °C, and 7.3 ± 2.3 kbar and 768 ± 75 °C (Table 5.3). These values are significantly higher than those of the garnet-biotite thermometers, and therefore represent maximum estimates for $M_1 P$ -*T* conditions in this sample.

In sample 288, THERMOCALC returned conditions of 7.6 \pm 1.3 kbar and 642 \pm 31 °C from an independent set of five reactions for an M₁ assemblage (Table 5.3; Appendix 3). This temperature estimate is noticeably higher than the calculated garnet-biotite exchange thermometer estimates, and represents a maximum estimate for the temperature conditions during M₁. A THERMOCALC average *P*-*T* estimate was also obtained from a later assemblage containing chlorite, and returned conditions of 3.1 \pm 1.7 kbar and 557 \pm 117 °C from an incomplete independent set of three reactions (Table 5.3; Appendix 3). These conditions are low enough that they are analogous to calculated *P*-*T* conditions for M₂ in the Robertson River Subgroup near Robinhood (Cihan *et al.*, 2006); therefore I suggest that this estimate is more likely to represent early M₂ rather than retrograde M₁ conditions. Temperature estimates for M₂ using garnet-biotite exchange thermometers returned values lower than the THERMOCALC temperature estimate; however they remain within error of this value (Table 5.3).

Temperature conditions estimated for sample 293 using garnet-biotite exchange thermometers returned average temperatures approximately 100 °C below that of samples 282/283 and 288 (Table 5.3). They vary between 418 and 426 °C for assumed pressures of 5.5-6.5 kbar, and may represent M_{IR} temperatures. Two estimates of average *P*-*T* conditions were calculated from equilibrium assemblages using THERMOCALC. In both cases estimates produced an independent set of reactions (Appendix 3). Maximum M_1 conditions experienced by this rock are best represented by the average *P*-*T* estimates of 6.3 ± 1.8 kbar and 624 ± 69 °C (Table 5.3). Although of lower pressure, this estimate corresponds to the average *P*-*T* conditions estimate of 5.1 ± 1.0 kbar and 559 ± 35 °C (Table 5.3), which is significantly lower than the inferred maximum M_1 conditions. This assemblage may have been influenced by later retrogression related to M_{IR} , as the calculated *P*-*T* estimates obtained from M_{IR} assemblages in sample 312.

Estimated P-T conditions for an M_1 mineral assemblage containing garnet and staurolite in sample 306 are close to those previously inferred to represent the peak

 M_1 conditions in the area (Boger & Hansen, 2004; Cihan *et al.*, 2006). The average *P*-*T* method of THERMOCALC gives conditions of 6.3 ± 2.5 kbar and 574 ± 18 °C from an independent set of five reactions (Table 5.3; Appendix 3). As with previous samples, temperature estimates using the garnet-biotite thermometers of Perchuk & Lavrent`eva (1983) returned estimates lower than the average *P*-*T* method. Estimates for paired garnet and M_1 biotite are 523 and 526 °C for an assumed pressure of 6.5 kbar (Table 5.3).

For sample 312, the methods of Kohn & Spear (1990) and Graham & Powell (1984) were used in addition to the average *P*-*T* method of THERMOCALC for *P*-*T* estimates, due to the absence of biotite and muscovite and the presence of hornblende. Close correlation is observed between the THERMOCALC estimates and the exchange method estimates (Table 5.3). Due to the presence of both M₁ hornblende and retrograde M_{1R} actinolite, *P*-*T* estimates for both M₁ and M_{1R} were calculated. For M₁, THERMOCALC returned average *P*-*T* conditions of 7.1 ± 1.8 kbar and 649 ± 69 °C, and 6.4 ± 1.7 kbar and 631 ± 69 °C. The garnet-hornblende exchange thermometer of Graham & Powell (1984) gave slightly lower temperature estimates for M₁ ranging up to 659 °C at an assumed pressure of 6 kbar from several garnet-hornblende pairs. Pressure conditions calculated by the garnet-hornblende-plagioclase-quartz Fe and Mg exchange barometers of Kohn & Spear (1990) gave estimates of up to 6.5 kbar for an assumed temperature of 650 °C (Table 5.3).

Average *P-T* conditions calculated by THERMOCALC from minerals considered representative of M_{1R} in sample 312, including S_2 actinolite, returned estimates below that calculated for M_1 . Using the average *P-T* method, two independent sets of reactions were produced giving conditions of 5.3 ± 1.4 kbar and 562 ± 51 °C, and 5.5 ± 1.6 kbar and 564 ± 60 °C (Table 5.3). As with the estimates produced for M_1 , the garnet-hornblende exchange thermometer of Graham & Powell (1984) gave a slightly lower temperature estimate for M_{1R} of 512 °C at an assumed pressure of 5.5 kbar. Pressure conditions calculated using the garnet-hornblende-plagioclase-quartz Fe and Mg exchange barometer of Kohn & Spear (1990) correlate well with the average *P* estimate, returning a calculated pressure of 5.2 kbar at an assumed temperature of 550 °C (Table 5.3).

5.8 Discussion

5.8.1 Limitations on the interpretation of *P*-*T* estimates using the exchange and net transfer methods.

The use of exchange and net transfer calculations in thermobarometric analysis require that the compositions analysed are in equilibrium at some point in their evolution. Typically, compositions in equilibrium are assumed, often conveniently, to represent peak metamorphic assemblages. However, in terranes that have experienced multiple recrystallising events, or relatively long cooling histories, equilibrium assemblages are often difficult to identify due to compositional heterogeneities in the phases of interest. The compositional heterogeneity within crystals is caused by a combination of exchange and net transfer reactions (hence their use in thermobarometric studies), and is often represented by chemical zoning in minerals (Spear, 1993). Zoning can occur during the prograde history of a mineral ('growth zoning'), and during its cooling history ('diffusion zoning').

The compositional zoning within crystals can have significant effects on *P-T* estimates, especially if the correct mineral compositions are not selected for particular stages of mineral growth (Spear, 1993). Garnet porphyroblasts from samples analysed in this study demonstrate the effects of diffusion and possibly also growth zoning (Fig. 5.5). The limited effects of diffusion on the garnet porphyroblasts analysed are visible on garnet rims in Figure 5.5. The diffusive zoning is a consequence of the Fe-Mg equilibrium exchange between garnet and adjacent biotite. However, net transfer reactions involving the consumption/production of metastable phases such as garnet and biotite, have also played an important role. This is demonstrated by the observed range in biotite compositions (e.g. Sample 306), and the compositional variation across the garnet porphyroblast in Sample 306 (Fig. 5.5a).

Consequently, there is significant risk in the compositions selected for analysis not being in equilibrium. Therefore, it is reasonable to suggest that P-T estimates obtained from reactions which involve garnet and other potentially zoned minerals, such as the garnet-biotite exchange thermometer used in this study, be treated with some caution.

5.8.2 Widespread retrograde assemblages in Etheridge Group metasediments from Gum Flats: evidence of an extended period of retrogression following peak metamorphism?

Widespread retrogression of prograde (M_1) metamorphic minerals, most notably staurolite to chlorite, and associations of other low-grade metamorphic minerals coexisting with higher-grade minerals, notably actinolite and hornblende, suggests a period of retrograde metamorphism immediately followed M_1 in the Etheridge Group metasediments near Gum Flats. These relationships are also supported by average *P*-*T* calculations from two samples (samples 293 and 312; Table 5.3), involving inferred retrograde assemblages that produced estimates consistent with a decrease in both pressure and temperature from M_1 .

Numerous outcrops of Daniel Creek Formation metasediments from both the Gum Flats and Ortona areas contain chlorite porphyroblasts that I interpret to have formed via complete replacement of M₁ staurolite. In sample 288, the pairing of chlorite to an adjacent garnet rim, muscovite, biotite and quartz located in the strain shadow of the porphyroblast (Fig. 5.4c), allowed for the calculation of average *P*-*T* conditions of 3.1 \pm 1.7 kbar and 557 \pm 117 °C. I infer that these conditions mark the start of M₂ metamorphism, as well as the end of the decreasing *P*-*T* path representing M_{1R}. However, it is also possible that these conditions were reached later, between ca. 1550-1500 Ma. An M₁ assemblage from the same sample involving a garnet core, adjacent S₁ biotite, muscovite and plagioclase returned a maximum average *P*-*T* estimate for M₁ of 7.6 \pm 1.3 kbar and 642 \pm 31 °C, which also provides a tying point between maximum M₁ *P*-*T* conditions and initial *P*-*T* conditions for M₂ (Fig. 5.6).

In sample 312 coexisting M_1 hornblende and actinolite may suggest progressive metamorphism, with hornblende forming at higher grades. However, using the same method as for sample 288, actinolite when paired with an adjacent garnet rim, plagioclase and epidote returned average *P*-*T* conditions consistent with the retrograde M_{1R} event (5.5 ± 1.6 kbar, 564 ± 60 °C). given that these conditions are significantly different from assumed M_1 estimates made by pairing hornblende with garnet cores (7.1 ± 1.8 kbar, 649 ± 69 °C), it is reasonable to assume that the co-existence of hornblende and actinolite may not indicate immiscibility but rather a later stage of retrogressive metamorphism, with M_1 hornblende replaced by actinolite and possibly clinozoisite.



Figure 5.6: *P-T-t* diagram for the lower Etheridge Group metasediments of the Gum Flats area (in red). *P-T-t* path in blue represents metamorphic history for Robertson River Subgroup metasediments near Robinhood (after Cihan *et al.*, 2006). Yellow star indicates maximum average *P-T* estimate obtained for M₁ from mica schist of the Einasleigh Metamorphics at Gum Flats (Sample 282; Table 5.3).

As the number of samples successfully analysed in this study was relatively small, and their mineral assemblages relatively limited, only a small number of geothermometer and geobarometers calibrations were successfully applied. Despite these limitations, the average *P-T* conditions calculated for representative mineral assemblages, and to a lesser extent the garnet-biotite and garnet-hornblende exchange thermometers used, provide an indication of the differing metamorphic conditions that were experienced during both the M₁ and M_{1R} events. From the results obtained, it appears that retrogression following M₁ in the Gum Flats area involved an approximate decrease in pressure and temperature of up to 4 kbar and 200 °C, possibly associated with uplift and extrusion immediately following M₁.

Indirect evidence for the presence of a retrograde M_{1R} event in the metasediments of the Etheridge Group was also recently reported by Cihan *et al.* (2006) using EMPA ages obtained from M_1 (1586 ± 6 Ma) and inferred M_2 (1542 ± 8 Ma) monazites and

associated average *P*-*T* estimates. Based on these weighted-average monazite ages, in addition to microstructural data including the retrogression of M_1 staurolite to andalusite, they inferred that an extended period of retrogression may have lasted for as long as ~45 Ma. Although *P*-*T* estimates from their study were produced from analysis of metasediments of the Robertson River Subgroup at Robinhood, their *P*-*T*-*t* path is broadly analogous to the path *P*-*T*-*t* path for Gum Flats (Fig. 5.6). The correlation of *P*-*T* estimates was also confirmed by the analysis of M₁ minerals in sample 306, collected from the same locality, which returned average *P*-*T* estimates that lay upon the prograde path of M₁ from their study (Fig. 5.6).

5.8.3 Summary and timing of *P-T* conditions during mid-Proterozoic prograde and retrograde metamorphism in the lower Etheridge Group metasediments, and regional implications

Quantitative estimates obtained through the use of several directly calibrated barometers and thermometers and the average *P*-*T* method of THERMOCALC support the premise that the mid-Proterozoic (ca. 1600-1550 Ma) metamorphic evolution of the metapelites from the southern extensions of the lower Etheridge Group began with burial and metamorphism to mid-crustal pressures and temperatures (M₁). This was followed by a period of retrogression (M_{1R}), then a low-pressure, high-temperature thermal overprint (M₂). Although sample 306 was overprinted by a later metamorphic fabric (S₃), I was unable to obtain a reliable *P*-*T* estimate for this event. In the Gum Flats area, M₁ in Bernecker Creek Formation and Daniel Creek Formation metasediments resulted in *P*-*T* conditions of up to 7.6 kbar and 650 °C, and facilitated the development of a weak to strong layer-parallel schistosity (S₁). The garnet-hornblende gneiss (sample 312) reached similar peak metamorphic conditions for M₁, recording conditions of up to 7.1 ± 1.8 kbar and 649 \pm 69 °C.

Estimates obtained for M_1 in samples of the Einasleigh Metamorphics (samples 282 and 283) are consistent with this unit experiencing on average higher metamorphic conditions than those experienced in the Bernecker Creek Formation and Daniel Creek Formation. For sample 282, M_1 average *P*-*T* estimates of up to 7.3 ± 2.3 kbar and 768 ± 78 °C were calculated (Table 5.3). A gradual decrease from the peak metamorphic conditions of M_1 is indicated in a number of samples by the replacement

of prograde metamorphic minerals. Examples of this include garnet by biotite and staurolite porphyroblasts pseudomorphed by chlorite. Initial *P*-*T* conditions at the onset of M_2 in the metasediments at Gum Flats are best represented by the average *P*-*T* estimate of 3.1 ± 1.7 kbar and 557 ± 117 °C obtained from sample 288 (Table 5.3).

The average *P*-*T* estimate for M_1 from sample 288 appears anomalously high when compared to the average *P*-*T* estimate for sample 293, given their relative stratigraphic positions (Fig. 5.2; Table 5.3). However, it is possible that prior to displacement along the Gilberton Fault during the Late Ordovician or Early Silurian (Withnall, 1989), the rocks surrounding sample 288 were in proximity to biotitegneiss of the (higher-grade) Einasleigh Metamorphics, at a similar stratigraphic position to that now observed for sample 312. Removing the suggested ~50 km of dextral displacement along the Gilberton Fault (Withnall *et al.*, 1980) would place the Daniel Creek Formation rocks surrounding sample 288 in proximity to Einasleigh Metamorphics biotite-gneiss and layered migmatites that now lie ~20 km southwest of Gilberton (Withnall *et al.*, 2003a). Consequently, this suggests that outcrops of the Bernecker Creek Formation surrounding Gum Flats Station are of a relatively lower metamorphic grade than those of the Daniel Creek Formation schists proximal to schist/gneiss of the Einasleigh Metamorphics further north (Fig. 5.2), and that this difference may be related to the orogenic event that promoted M₁.

The major metamorphic events of the lower Etheridge Group (M_1 and M_2) have previously been estimated to have occurred within a ~30 million year period between 1580 and 1550 Ma (Withnall *et al.*, 1997; Black *et al.*, 2005). However, metamorphic monazite dating from Cihan *et al.* (2006) and this study infer that M_1 may have occurred slightly earlier (ca. 1590-1580 Ma). In sample 288, monazite that had grown within an M_1 assemblage returned a weighted mean age of 1585 ± 18 Ma, and correlates well with the estimated age of 1586 ± 6 Ma for M_1 from Cihan *et al.* (2006). This age is also coeval with the onset of major orogenesis in the Mt Isa Eastern Succession (Page & Sun, 1998; Sayab, 2006), as well as D_1 in the Yambo and Coen Inliers (Blewett & Black, 1998; Blewett *et al.*, 1998; Fig. 1.1). As discussed in Chapter 4, monazites recording M_2 conditions yielded ages between ca. 1550 Ma and 1525 Ma, suggesting that the grains may have been affected by partial resetting during a later event. For example, sample 283 recorded a weighted mean age within a mixed M_{1-2} assemblage of 1537 ± 18 Ma. This age is within error of the proposed age for M_2 from Cihan *et al.* (2006) of 1542 ± 8 Ma. However, due to the younger weighted mean age and spread of individual U-Th-Pb ages from individual grains, I suggest that the 1537 ± 18 Ma age may represent a partially reset M₂ age, possibly influenced by a later (ca. 1500 Ma) event.

The metamorphic history for the lower Etheridge Group metasediments at Gum Flats is illustrated in the *P*-*T*-*t* diagram shown in Fig. 5.6. Also shown is the *P*-*T*-*t* path for metasediments of the Robertson River Subgroup at Robinhood as proposed by Cihan *et al.* (2006). From the diagram, it is apparent that all rocks of the lower Etheridge Group experienced a similar 'clockwise' metamorphic history, beginning with a prograde metamorphic path leading to medium *P*-*T* metamorphism (M₁), followed by an extended period of retrograde metamorphism (M_{1R}), and then low-*P*, high-*T* metamorphism (M₂). The key difference between the two areas relates to level of metamorphism attained, with the metasediments at Gum Flats generally reaching higher temperatures for a given pressure during M₁ (Fig. 5.6). The highest M₁ conditions are recorded by the metasediments of the Einasleigh Metamorphics (Samples 282 and 283), reflecting their evolution at different crustal conditions before faulting along the Gilberton Fault led to their juxtaposition alongside the Daniel Creek Formation in the Palaeozoic.

The clockwise *P-T-t* path suggested here is not unique to the lower Etheridge Group. A similar metamorphic history has been proposed for the rocks of the Mt Isa Eastern Succession based on FIA trends and thermobarometric estimates (Sayab, 2006). The metamorphic history of this area involved similar mid-Proterozoic medium pressure-temperature conditions for M_1 that were followed by decompression, retrogression, and a later thermal overprint. In the case of Mt Isa this thermal overprint was associated with the emplacement of the Williams and Naraku Batholiths between ca. 1550-1500 Ma (Page & Sun, 1998). The emplacement of these batholiths may correlate with a late-stage 'thermal pulse' which was suggested by Cihan *et al.* (2006), and may have also led to the development of new metamorphic monazite and the partial resetting of M_2 monazite ages in this study. If this is the case, the evidence presented here, along with the growing collection of geochronological data indicating a coeval M_1 event across terranes (e.g. Blewett & Black, 1998; Blewett *et al.*, 1998; Page & Sun, 1998; Cihan *et al.*, 2006), would suggest that the Proterozoic cratons of

Northeastern Australia may have shared a similar metamorphic history spanning the period between ca. 1600 and ca. 1500 Ma.

Chapter 6.

Geochemical, Sm-Nd isotopic characteristics and petrogenesis of Palaeoproterozoic mafic rocks from the lower Etheridge Group, Georgetown Inlier, north Queensland

6.1 Introduction

Crustal blocks of Palaeoproterozoic – Mesoproterozoic age in eastern and northern Australia include the Curnamona Province in western New South Wales, the Mt Isa Inlier in western Queensland, and the Georgetown Inlier in north Queensland. Rock sequences in these inliers, including products of prominent and widespread mafic magmatism, are generally considered to have formed in major intra-cratonic basins developed in rift settings (e.g., Scott *et al.* 2000). These inliers, and several others further west in Northern Territory and central Australia, have been considered once to have formed a contiguous crustal block known as the North Australian Craton (Myers *et al.*, 1996; Tyler *et al.*, 1998; Scott *et al.*, 2000; Betts & Giles 2006).

Whether the assembly and subsequent dispersion of these Palaeoproterozoic – Mesoproterozoic crustal blocks involved dominantly intraplate processes with limited or no lateral accretion (e.g., Etheridge *et al.* 1987), or processes analogous to those occurring via plate tectonics on the modern Earth (e.g. Myers *et al.* 1996; McDonald *et al.* 1997; Scott *et al.* 2000; Dawson *et al.* 2002; Betts *et al.* 2002; Giles *et al.* 2002, 2004; Betts & Giles 2006) is still debated. Giles, Betts and co-workers proposed that plate margin processes may affect basin evolution within continental blocks, and as summarised in Betts & Giles (2006), these authors have argued that the North Australian Craton was bordered on its southern margin by a S-dipping subduction zone from 1800-1600 Ma, and on its eastern margin by W-dipping subduction zone from 1600-1500 Ma.

Knowledge of the timing and composition of mafic magmatism in each of the inliers that are considered to have comprised the North Australian Craton is vital to developing a clearer understanding of the possible temporal and spatial links between these inliers, and their tectonic setting at the time of the mafic magmatism. To that end, I report here a detailed geochemical study of the widespread mafic rocks of late Palaeoproterozoic age in the Georgetown Inlier of north Queensland.

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6.2 Previous Work

Previous geochemical studies of mafic rocks within the lower Etheridge Group include the work of McNaughton & Wilson (1983), Black & McCulloch (1984), Withnall (1985) and Knutson & Stephenson (1989). McNaughton & Wilson (1983) analysed 13 mafic granulites from the Einasleigh Metamorphics near Einasleigh, and concluded that the immobile element concentrations of the rocks were virtually indistinguishable from those of modern MORB, and that the igneous precursors of the mafic rocks were probably derived from a mantle source similar to that which underlies modern mid-ocean ridges.

As part of a broader study of the Nd isotopic compositions of Palaeoproterozoic mafic rocks in central and northern Australia, Black & McCulloch (1984) made a reconnaissance study of the Sm-Nd isotope characteristics of mafic rocks from the Georgetown Inlier. Based on the analysis of T_{DM} ages from several Cobbold Metadolerites, they concluded that the mafic igneous rocks of the lower Etheridge Group were emplaced at 2490 ± 70 Ma, and were derived from partial melting of a depleted upper mantle source ($\varepsilon_{Nd} = +4.0 \pm 0.5$).

The studies of both Withnall (1985) and Knutson & Stephenson (1989) incorporated major and some trace element analyses of the mafic rocks. Withnall (1985) presented analyses of 52 mafic samples collected from the lower Etheridge Group. Knutson & Stephenson (1989) reported compositions for 11 mafic samples collected from the lower Etheridge Group. Lithologies analysed in both studies included greenschist and amphibolite facies metabasalt and metadolerite, amphibolite and 'mafic granulite'. Withnall (1985) concluded that the mafic rocks within the lower Etheridge Group are basaltic, and compare closely with K-poor Tertiary continental tholeiites of Greenland and Baffin Island, and relatively incompatible-element depleted Karoo mafic rocks of southern Africa, both suites being generated during extension of continental crust.

6.3 Objectives of the geochemical study

As there has been no comprehensive geochemical study of the mafic rocks of the lower Etheridge Group to date, and conflicting interpretations (MORB-like vs. continental tholeiites) exist for the geochemical character or likely tectonic setting of eruption of these igneous rocks, the key objectives of this geochemical study are:

- To investigate the geochemical character of the mafic rocks of the lower Etheridge Group, and if possible, to infer the likely tectonic setting of their formation.
- To assess whether any significant differences exist between the various mafic rocks of the lower Etheridge Group.
- To investigate the petrogenesis and chemical characteristics of the parental magmas to the mafic rocks in order to make a comparative study with similarly aged mafic rocks from the Mt. Isa and Broken Hill Inliers (Chapter 7).

Details of the sampling and analytical techniques used and additional petrographic information on samples analysed are provided below. A detailed overview of the petrology, mineralogy, stratigraphic, and textural relationships between the DHM and CMD is given in Chapter 3.

6.4 Methods

6.4.1 Sampling

In areas of semi-continuous outcrop, mafic rocks were collected along traverses with an approximate sample spacing of 100 m, except where detailed sampling to monitor geochemical variations across several thick sills used a much closer (20 m) sample spacing. Thin sections were prepared for all samples, and least altered samples were chosen for whole rock analysis on the basis of geographic spread. Samples showing a foliation, oxidation, veining, or more than 1-2 modal% carbonate or sulphides were excluded from analysis.

For ease of reference, mafic rocks from the lower Etheridge Group are separated into five domains based on their location (Fig. 6.1), and to a lesser extent their grade of metamorphism. The 'Ortona', 'South Head' and 'Stockyard' domains contain samples of greenschist facies DHM and CMD, whereas the 'Robinhood' and 'Gum Flats' domains contain samples of amphibolite facies CMD. Some 88 samples were analysed by XRF, 25 by ICPMS, and 10 for Nd-isotopes by MC-ICPMS using the methods described below. A full list of analyses and sample locations is given in Tables 6.1 and 6.3 along with average compositions for each domain (Table 6.2).



Figure 6.1: Simplified geological map of the central Georgetown Inlier showing key sampling locations, and the location of sample domains.

6.4.2 Major, trace, and rare earth element analytical techniques

Samples were prepared for analysis using a jaw crusher and tungsten-carbide ringmill. For XRF analysis, samples were prepared as glass discs for major element-, and pressed powder discs for some trace element analysis. All samples were analysed using a Phillips PW1480 XRF system at the University of Tasmania School of Earth Sciences. Sample preparation, instrument operating conditions, and calibration procedures of the analyses follow the method of Norrish & Hutton (1969).

A representative collection of samples was also analysed for low-level trace elements and rare earth elements (REE) using the HP 4500Plus Inductively Coupled Plasma Mass Spectrometer (ICPMS) at the School of Earth Sciences, University of Tasmania. Samples for ICPMS analysis were digested using a low-pressure HF/HNO₃ Savillex digestion. For both XRF and ICPMS analysis, in-house standards TASBAS, BHVO-1, Tafahi and BIR-1 were analysed as unknowns in order to monitor instrumental drift and data quality during the analytical period. Precision for both methods (determined by repeated analysis of the standards and presented as relative standard deviation [% RSD] for the period of analysis) is better than 5% except for Li, Sc, V, and Zn (5-14%). Analytical accuracy (presented as expected value/average measured value) is better than 5%. Conservative estimation of the detection limits is less than 50 ppb for all elements analysed by ICPMS. Total analytical blank is less than 10 ppb for most elements except for Li, Sc, V, Zn, Mo, and Pb (21-215 ppb), making blank corrections mostly less than 0.1% for mafic samples. Both XRF and ICPMS methods were used for 14 trace elements (Sc, V, Cr, Mn, Ni, Cu, Zn, Sr, Y, Zr, Nb, Ba, Ce, and Nd) on 25 samples. The difference between the two methods is generally less than 10% for all elements, except for Zr and Cr in some samples (up to 30% lower for ICPMS) due to incomplete digestion of chromite and zircon. Samples which returned ICPMS Zr values that varied by more than 15% from their XRF values were re-analysed using a high-pressure HF/H₂SO₄ PicoTrace digestion, in order to ensure complete digestion of zircon.

6.4.3 Sm-Nd isotope analytical technique

All Sm-Nd analyses were carried out on the NU Instruments MC-ICPMS in the School of Earth Sciences at the University of Melbourne by Dr. Roland Maas. Sample preparation followed Maas *et al.* (2005), and involved the dissolution of whole rock sample powders in Krogh-type high-pressure bombs. Initial digestions with HF-HNO₃ (48 hrs) were followed by digestion with 6M HCl (12 hrs). A mixed tracer solution spike of ¹⁴⁹Sm-¹⁵⁰Nd was added to each sample, and samples were prepared for mass spectrometry using a combination of cation exchange and resin methods (Pin *et al.*, 1994; Maas *et al.*, 2005).

Sm and Nd concentrations in spike and unknown samples were adjusted to yield a mean of 6.59 and 28.8 ppm respectively, based on 3 analyses of international standard BCR-1 run in the same batch. Total laboratory blanks of Sm and Nd for dissolutions were lower than 50 pg. Uncertainties in ¹⁴⁷Sm/¹⁴⁴Nd and ¹⁴³Nd/¹⁴⁴Nd are considered to be better than $\pm 0.2\%$ (2 σ) and $\pm 0.004\%$ (2 σ) respectively (R. Maas, pers. comm., 2006). ¹⁴³Nd/¹⁴⁴Nd ratios were normalised to 0.7219 for ¹⁴⁶Nd/¹⁴⁴Nd, and are reported relative to La Jolla Nd = 0.511860.

Sample:	6	8	22	23	60	72	74	80	82	83	85	86
Unit:	CMD	CMD	CMD	CMD	CMD	DHM	DHM	DHM	DHM	DHM	CMD	CMD
Rock type:	Dolerite	Dolerite	Dolerite	Dolerite	Dolerite	Basalt	Basalt	Basalt	Basalt	Basalt	Dolerite	Dolerite
Domain:	Ortona	Ortona	Ortona	Ortona	Ortona	Ortona	Ortona	Ortona	Ortona	Ortona	Ortona	Ortona
Method	ICPMS	XRF	ICPMS	XKF	XRF	XRF	ICPMS	XRF	XKF	XRF	XRF	XRF
SiO.	53 64	40.18	47 14	46 73	48 50	48 21	48 73	45 31	46.27	47.66	49.15	47.62
TiO ₂	1 20	1 70	1.65	1 92	1 13	1 47	1 47	1 64	1 44	1.52	1 71	2.05
Al ₂ O ₃	12.95	13.58	13.15	13.41	15.35	13.75	14.25	15.46	14.05	14.90	12.67	12.48
Fe ₂ O ₃ *	13.13	6.81	15.82	18.31	13.32	15.79	15.27	17.52	15.49	16.08	17.40	17.75
MnO	0.21	0.33	0.23	0.25	0.19	0.23	0.30	0.31	0.22	0.25	0.25	0.24
MgO	5.18	4.43	7.99	5.90	6.93	6.53	7.74	7.49	6.87	6.09	5.12	5.51
CaO	7.69	11.17	9.09	8.57	9.59	9.36	5.01	4.60	11.18	5.29	8.32	9.48
Na ₂ O	3.53	5.65	1.86	2.14	2.73	2.14	3.64	3.84	1.54	4.21	2.74	2.99
K ₂ O	0.46	0.82	0.15	0.13	0.11	0.36	0.05	0.05	0.03	0.03	0.10	0.06
P ₂ O ₅	0.11	0.14	2.69	0.18	0.10	2.00	0.12	0.14	2.04	0.14	0.14	0.15
Total	99.76	99.82	99.91	100.07	100.33	100.07	99.97	99 94	100.15	100.20	100.07	100.00
Total	<i>)).</i> (0	<i>>></i> .02	,,,,,	100.07	100.55	100.07	,,,,,		100.10	100.20	100.07	100.00
Cs	0.36		0.38				0.08					
Rb	19.85	21.8	6.50	4.2	3.4	16.5	0.61	1.1			1.5	
Ba	1.82	166	39	69	147	91	40	20	55	22	85	85
In II	0.61		0.36	2			0.59					
Nb	5.7	7	7.1	7	4	6	6.20	5	5	7	5	6
Та	0.58		0.57				0.40					
La	10.09	8	8.58	7	4	5	5.21	10	2	2	4	5
Ce	23.26	18	20.38	16	10	14	13.63	18	13	7	13	16
Pb	8.95	4	9.24	12	1	8	1.74	2		3	6	2
Pr	3.17	0.5	2.92				2.16					
Sr	131.78	95	1/2.14	148	453	151	33.75	/6	80	39	208	192
Sm	3 70	9.0	3.68	15.8	7.8	11.5	3 /1	12.2	10.4	1.5	9.7	10.0
Zr	136	78	94	100	56	93	96	92	86	99	87	94
Hf	2.69		2.10				1.52					
Eu	1.17		1.03				0.69					
Gd	4.50		4.02				4.31					
Tb	0.83		0.68				0.77					
Dy	5.27	0.0	4.04	20.2	10.7	20.2	4.84	20.2	25.0	22.0	20.2	20.7
Y He	30.89	9.9	20.74	29.2	19.7	28.2	25.55	30.2	25.8	22.0	28.2	29.7
Er	3 38		2 28				2.84					
Tm	0.51		0.31				0.41					
Yb	3.28		1.98				2.42					
Lu	0.48		0.28				0.34					
V	403.80	412	462.71	486	303	405	351.66	524	418	363	553	624
Cr	72	121	84	59	189	137	125	184	135	145	35	3
N1	60 270	39.9	/4	94.2	103.2	/1.8	82	90.3	86.6	84./	220	210
Zn	106.75	17	135.81	109	100	149	192	139	190	80 111	130	132
Ga	18.04	1,	21.26	155	100	100	20.30	150	120		150	152
Zr/Nb	23.91	11.00	13.18	14.13	15.44	16.02	15.53	19.25	17.45	13.93	17.38	15.23
Y/Nb T:/W	5.35	1.39	2.83	4.11	5.47	4.86	4.03	6.29	5.27	3.10	5.64	4.79
11/V 7r/V	20.87	24.71	4.66	25.05	22.45	21.08	3.85	3.06	20.01	25.08	18.57	3.18
Ti/Zr	52.59	130.46	105 72	114 70	122.34	94 60	91 33	106 48	100 80	92.11	118.08	129.96
Ti/Y	235.02	1029.17	492.30	394.00	345.28	311.66	351.78	325.77	334.05	414.09	363.86	413.08
Th/Nb	0.32		0.18	0.23			0.10					
Nb/La	0.57	0.93	0.83	1.03	0.88	1.29	1.19	0.48	2.13	3.55	1.25	1.32
Ta/Yb	0.18		0.29				0.17					
(La/Sm)n [†]	1.67		1.47				0.96					
(Gd/Yb)n ^T	1.11		1.64				1.44					
(Eu/Eu*)n'	0.80		0.82				0.55					
^ Method use	d to obtain t	trace and ra	re earth eler	nent compo	sitions, all m	naior elemer	nt data obtai	ned bv XRF				

† Normalising values after Boynton (1984)

Table 6.1: Major (wt%), trace (ppm) and rare earth element (ppm) compositions for mafic rocks of the lower Etheridge Group.

Sample: Unit: Rock type: Domain: Method^:	87 DHM Basalt Ortona XRF	88 CMD Dolerite Ortona XRF	89 DHM Basalt Ortona ICPMS	90 DHM Basalt Ortona XRF	100 DHM Basalt Ortona ICPMS	101 CMD Dolerite Ortona ICPMS	102 DHM Basalt Ortona XRF	103 DHM Basalt Ortona ICPMS	104 DHM Basalt Ortona XRF	106 CMD Dolerite Ortona ICPMS	107 DHM Basalt Ortona XRF	108 CMD Dolerite Ortona ICPMS
SiO ₂	50.63	50.70	50.54	48.57	49.40	48.85	46.40	49.27	47.01	46.11	42.02	48.49
TiO ₂	1.41	1.42	1.43	1.67	1.54	2.07	1.70	1.42	1.42	2.31	2.23	1.41
Al_2O_3	12.97	13./1	13.08	12.97	14.89	15.01	14.57	14.04	14.11	11.78	11.64	15.98
MnO	0.23	0.22	0.22	0.25	0.22	0.23	0.25	0.24	0.24	0.23	0.41	0.22
MgO	6.81	6.25	5.43	6.22	5.80	5 35	6.31	6.87	6 44	4 92	2.83	6.00
CaO	7.67	7 42	8.65	8 51	6 59	8 47	6.01	6.36	9.09	8 83	11 47	9.62
Na ₂ O	2.76	3.91	2.74	1.85	4.35	3.25	3.91	4.01	2.98	2.38	1.75	2.38
K ₂ O	0.05	0.06	0.07	0.10	0.09	0.09	0.04	0.06	0.05	0.20	0.01	0.17
P_2O_5	0.13	0.12	0.14	0.16	0.15	0.22	0.19	0.15	0.16	0.18	0.22	0.13
LOI	2.27	1.84	2.52	2.44	2.08	1.94	3.41	2.11	2.71	4.29	9.99	2.06
Total	100.00	100.05	99.79	100.08	100.00	100.38	99.51	100.13	99.88	99.61	99.80	99.88
Cs			0.16		0.31	1.45		0.07		1.11		0.31
Rb			1.60	2.3	0.91	1.68		0.45		9.64		5.92
Ba	27	30	39	36	40	128	14	26	23	468	5	141
Th	2		0.94		0.72	1.09		0.63		1.06		0.59
U			0.31	-	0.29	0.38		0.21		0.32		0.20
Nb T-	6	5	5.9	7	7.5	9	8	5.6	6	6.7	8	4.8
Ia			6.12	4	0.48	0.01	2	0.41	4	0.55	0	0.44
La	0	Ŷ	15.28	12	4.07	9.57	5	4.54	4	0.39 20.50	25	12.09
Ph	3	0	2 64	4	1 89	24.24	4	0.33	10	20.30	23 4	1 1 9
Pr	5		2.04	7	2.07	3 76	7	1.68		3.15	4	2 20
Sr	37	38	88.52	121	97.64	99.00	39	29.95	126	338.48	85	173.38
Nd	6.0	7.9	11.22	12.4	10.59	17.74	5.8	7.89	10.2	14.96	18.0	10.59
Sm			3.45		3.62	5.46		2.50		4.55		3.29
Zr	90	82	87	110	115	152	111	91	91	113	152	84
Hf			2.05		1.72	3.32		1.83		2.84		1.87
Eu			1.09		1.16	1.68		0.57		1.48		1.21
Gd			4.27		4.71	6.80		3.14		5.59		4.32
Тъ			0.75		0.86	1.21		0.60		0.99		0.76
Dy	22.0	22.4	4.73	22.4	5.38	7.82	22.0	3.83	27.1	6.16	42.1	4.89
Y Ho	23.0	23.4	25.85	33.4	29.55	42.27	22.9	18.90	27.1	1 20	42.1	27.42
Fr			2.82		3.27	1.04		2.25		3.60		3.06
Tm			0.40		0.47	0.69		0.34		0.53		0.47
Yb			2.56		2.95	4.35		2.26		3.39		2.90
Lu			0.36		0.41	0.62		0.34		0.50		0.42
V	358	379	373.04	448	398.26	365.45	410	417.17	379	686.83	476	400.41
Cr	138	127	82	70	146	98	64	135	132	1	3	82
Ni	73.7	71	61	49.2	73	55	66.2	79	79.7	50	24	69
Cu	123	139	181	182	114	156	230	25	95	241	95	206
Zn Ga	125	107	166.26 17.57	137	155.92 15.50	111.91 12.31	134	119.50 19.55	123	146.20 21.84	185	111.65 19.39
Zr/Nb	15 72	15.67	14 76	15.07	15.27	16.96	14.05	16 20	16.45	16 70	10 77	17.52
V/Nb	13.72	4.50	4.70	13.97	3.02	1 83	2 90	3 30	10.45	5.01	5 47	5.83
Ti/V	23 57	22 37	21.60	22 37	21.73	26.21	2.50	22.06	22.53	23.86	28.06	21.03
Zr/Y	3.90	3.48	3.35	3.30	3.89	3.49	4.85	4.81	3.34	3.35	3.62	3.00
Ti/Zr	94.18	104.13	98.40	90.93	80.75	81.90	91.74	93.28	94.31	123.12	87.81	100.20
Ti/Y	366.90	362.68	329.64	300.03	314.47	285.62	444.67	448.25	314.93	412.23	317.47	300.96
Th/Nb	0.33		0.16		0.10	0.12		0.11		0.16		0.12
Nb/La			0.96	1.97	1.60	0.96	2.47	1.23	1.57	0.80	0.91	0.87
Ta/Yb			0.15		0.16	0.14		0.18		0.16		0.15
(La/Sm)n [†]			1.12		0.81	1.08		1.14		1.16		1.06
(Gd/Yb)n [†]			1.35		1.29	1.26		1.12		1.33		1.20
(Eu/Eu*)n'			0.87		0.86	0.84		0.62		0.90		0.98

Sample: Unit: Rock type: Domain: Method^:	109 CMD Dolerite Ortona ICPMS	110 CMD Dolerite Ortona XRF	112 CMD Dolerite Ortona ICPMS	113 CMD Dolerite Ortona ICPMS	114 CMD Dolerite Ortona XRF	115 CMD Dolerite Ortona ICPMS	120 CMD Dolerite Ortona XRF	121 CMD Dolerite Ortona ICPMS	128 CMD Dolerite Ortona XRF	129 CMD Dolerite Ortona XRF	130 CMD Dolerite Ortona ICPMS	148 DHM Basalt Ortona ICPMS
SiO ₂	49.82	48.28	50.91	46.92	49.20	46.38	47.23	47.71	49.06	49.21	50.50	46.60
TiO ₂	1.70	1.54	1.20	0.93	1.49	0.90	1.08	1.08	1.13	1.24	1.78	2.41
Al_2O_3	12.40	12.37	14.34	15.81	13.23	14.93	14.06	14.46	13.65	13.35	12.91	12.78
Fe ₂ O ₃ *	16.86	17.15	13.49	13.22	16.63	13.34	14.72	14.44	15.06	15.08	17.50	18.72
MnO	0.27	0.26	0.20	0.18	0.24	0.19	0.21	0.21	0.22	0.20	0.20	0.30
MgO G-O	4.76	5.99	0.29	8.45	5.76	9.87	8.63	/.55	7.37	0.51	3.92	6.49
Va Na O	9.34	9.38	2.23	9.99	2.05	0.97	9.44	0.00	2.64	8.34 2.40	2.84	1.81
K ₂ O	0.14	0.05	0.43	0.20	0.05	0.30	0.17	0.10	1 19	1.77	1 37	0.06
P ₂ O ₅	0.14	0.05	0.45	0.09	0.03	0.09	0.09	0.10	0.10	0.10	0.21	0.00
LOI	2.55	2.11	2.40	2.78	2.30	3.12	2.78	2.79	2.24	1.58	2.38	4.10
Total	100.49	99.86	100.32	100.42	99.95	99.89	100.04	100.04	99.69	99.79	100.43	99.54
Cs	3.87		1.13	0.75		0.74		0.26			0.36	0.18
Rb	2.34		18.02	7.98		12.93	8	3.43	42	59	44.02	1.07
Ba	110	83	217	116	34	216	87	47	118	73	58	17
Th	0.68	2	2.13	0.35		0.36		0.46			1.12	1.34
U	0.23		0.61	0.13		0.13		0.16			0.36	0.47
Nb	6.2	5	5.7	3.5	5	3.8	4	4.3	5	4	6.5	10.3
1a Lo	0.46	7	0.41	0.25	2	0.24		0.30	2		0.65	0.67
La	0.19	12	0.00	5.59 8.66	11	5.42 9.77	0	4.15	5 10	10	9.74	26.8
Ph	4 00	29	2 95	0.70	5	0.79	,	0.82	2	10	3 36	3 56
Pr	2 44	27	2.75	1 37	5	1 40		1 72	-	-	3 79	4 18
Sr	142.46	121	158.83	145.56	170	120.67	101	119.96	87	95	107.34	83
Nd	11.83	10.4	11.86	6.72	10.2	6.68	5.5	8.29	9.8	6.8	18.29	20.2
Sm	3.71		3.29	2.14		2.15		2.64			5.48	6.15
Zr	91	84	88	54	86	54	53	69	58	66	106	165
Hf	2.33		2.07	1.38		1.38		1.63			3.04	4.36
Eu	1.42		1.05	0.80		0.78		0.93			1.78	2.10
Gd	4.71		3.93	2.68		2.70		3.39			6.83	7.46
1b De	0.85		0.67	0.49		0.49		0.61			1.23	1.38
Dy V	5.55 30.35	20.2	4.23	3.10	27.2	3.05 16.74	17.5	3.79 20.75	20.1	22.7	/./9	8.44 48.0
Но	1 19	2).2	0.91	0.66	27.2	0.65	17.5	0.83	20.1	22.1	1.65	1.81
Er	3.46		2.54	1.87		1.86		2 30			4 68	5.25
Tm	0.51		0.38	0.28		0.27		0.34			0.68	0.79
Yb	3.25		2.35	1.82		1.73		2.13			4.32	4.90
Lu	0.48		0.34	0.26		0.25		0.31			0.63	0.74
V	469.85	418	319.42	264.66	463	248.05	307	307.95	346	348	518.94	521
Cr	21	53	166	92	55	95	127	92	103	58	5	112
Ni	41	92.2	86	193	63.8	242	191.1	150	87.8	74.3	44	59
Cu	242	171	157	120	225	106	137	137	280	263	256	122
Zn Ga	20.28	207	106.28	83.79 16.91	135	84.29 16.40	112	95.42 17.63	82	68	95.40 22.86	183
7r/Nb	14.68	16.76	15 35	15.40	16.78	14.24	13.84	16.12	12 01	15.42	16 35	16.85
Y/Nh	4 79	5 84	4 35	5.00	5 33	4 32	4 61	4 84	4 47	5 28	6 12	4 83
Ti/V	22.39	22.04	21.93	20.97	19.33	21.93	21.08	21 41	19.55	21 40	23.08	27 70
Zr/Y	3.06	2.87	3.53	3.08	3.15	3.30	3.01	3.33	2.89	2.92	2.67	3.49
Ti/Zr	112.10	109.83	81.99	103.86	104.46	99.93	123.17	93.82	116.36	112.17	100.25	88.45
Ti/Y	343.46	315.20	289.28	319.88	328.76	329.64	370.22	312.59	336.35	327.60	267.76	308.64
Th/Nb	0.11	0.30	0.37	0.10		0.09		0.11			0.17	0.13
Nb/La	1.00	0.72	0.65	1.03	1.65	1.11		1.04	1.41		0.67	0.96
Ta/Yb	0.14		0.17	0.14		0.14		0.14			0.15	0.14
(La/Sm)n [†]	1.05		1.68	0.99		1.00		0.99			1.12	1.09
(Gd/Yb)n ^T	1.17		1.35	1.19		1.26		1.29			1.28	1.23
(Eu/Eu*)n'	1.04		0.89	1.05		0.98		0.90			0.89	0.95

Sample: Unit: Rock type: Domain: Method^:	151 DHM Basalt Ortona XRF	152 CMD Dolerite Ortona XRF	153 DHM Basalt Ortona XRF	154 DHM Basalt Ortona XRF	172 CMD Dolerite Ortona XRF	174 CMD Granophyre Ortona XRF	175 CMD Dolerite Ortona XRF	179 CMD Dolerite Ortona XRF	181 CMD Dolerite Ortona ICPMS	188 CMD Dolerite Ortona XRF	192 CMD Dolerite Stockyard XRF	194 CMD Dolerite Stockyard XRF
SiO ₂ TiO ₂	47.34 1.38	49.35 1.40	43.68 2.15	47.63 1.38	57.80 0.87	59.89 1.06	49.61 1.63	48.31 1.01	47.28 0.84	48.17 1.55	48.40 1.57	49.15 1.40
Al_2O_3	13.95	11.08	14.69	13.86	13.14	12.96	13.06	14.07	15.31	13.57	13.57	14.07
Fe ₂ O ₃ *	15.36	15.26	19.81	16.89	10.59	6.73	14.49	12.66	11.89	14.54	15.33	13.80
MnO	0.22	0.24	0.29	0.26	0.18	0.16	0.23	0.20	0.15	0.21	0.29	0.20
MgO	6.71	7.30	6.89	7.29	4.69	1.84	5.93	7.64	8.49	5.41	6.42	6.54
CaO Na O	2 41	11.87	3.06	2 72	7.89	4.48	9.55	9.50	10.54	1 01	2 35	2.28
K ₂ O	0.08	0.09	0.06	0.05	0.23	0.36	0.46	0.62	0.59	0.02	0.29	0.20
P ₂ O ₅	0.13	0.11	0.21	0.11	0.11	0.20	0.15	0.09	0.08	0.15	0.15	0.15
LOI	2.22	2.04	3.43	2.70	1.73	6.12	3.21	2.88	3.01	3.74	0.80	0.82
Total	99.90	99.88	99.95	100.03	99.94	99.86	99.69	99.45	99.99	100.32	99.74	99.65
Cs									1.41			
Rb		2	1		10	18	27	27	30.3		5	2
Ba	65	119	13	43	77	112	112	61	74	12	184	11
III II	2					14	2		0.44			
Nb	5	2	8	5	7	13	8	3	2.45	5	6	5
Та									0.25			
La	4	4	7	5	9	34	10		3.25	4	9	5
Ce	10	9	10	7	21	71	21	13	7.71	15	19	14
Pb	2	4	2	3	9	10	17	6	2.91	3	5	3
Pr	65	250	24	C1	121	107	205	10.4	1.17		100	176
Sr Nd	65 7.5	359	24	51	131	126	295	184	5.62	007	190	1/6
Sm	1.5	0.7	11.0	7.5	14	32.9	14.4	0.0	1.81	9.7	11./	11.2
Zr	80	70	129	87	163	199	104	52	42	94	78	75
Hf									1.18			
Eu									0.79			
Gd									2.29			
Tb									0.41			
Dy V	24.1	22.0	20.0	22.0	22.0	24.0	<u></u>	10.1	2.65	27.4	247	22.0
Но	24.1	23.0	29.9	22.0	33.9	34.9	22.2	19.1	0.57	27.4	24.7	22.9
Er									1.65			
Tm									0.24			
Yb									1.55			
Lu									0.23			
V	405	387	447	427	213	174	390	340	311	414	415	382
Cr	156	142	/8 66 /	140	40	2	39 52	122.2	16/	106	11/	13/
Cu	156	190	62	164	40.2	119	85	132.2	92	108	132	121
Zn	109	123	164	132	124	26	161	76	45	118	161	100
Ga									15.76			
7./\]	16 10	20.55	16 29	17.40	22.50	15 14	12.80	15.14	12.29	19.63	14.05	14.54
Y/Nb	4 87	10 11	3 78	4 4 5	4 70	2 65	2 74	5 52	4 66	5 42	4 46	4 44
Ti/V	20.43	21.68	28.81	19 39	24 47	36.45	25.04	17.82	16.21	22.44	22.67	21.99
Zr/Y	3.30	3.02	4.31	3.93	4.81	5.70	4.68	2.74	2.85	3.43	3.15	3.27
Ti/Zr	103.91	120.73	100.05	95.62	31.97	31.91	94.12	115.52	123.70	98.93	120.95	112.18
Ti/Y	343.19	364.82	430.97	375.95	153.81	182.04	440.06	316.93	352.06	339.66	380.96	367.21
Th/Nb	0.34					1.03	0.22		0.18			
Nb/La	1.37	0.53	1.18	1.08	0.82	0.38	0.84		0.75	1.17	0.60	0.97
Ta/Yb									0.16			
(La/Sm)n'									1.13			
$(Ga/YD)n^{\prime}$									1.19			
(EWEW.)II												

Sample: Unit:	195 CMD	196 CMD	197 CMD	199 CMD	200 CMD	201 CMD	202 CMD	221 CMD	222 CMD	224 CMD	226 CMD	227 CMD
Domain: Method^:	Stockyard XRF	Stockyard ICPMS	Stockyard XRF	Stockyard XRF	Stockyard XRF	Stockyard XRF	Stockyard XRF	South Head XRF				
SiO ₂	49.28	49.31	48.63	48.51	49.13	49.14	49.20	45.39	49.12	47.16	48.85	45.41
TiO ₂	1.51	1.44	1.20	1.56	1.17	1.30	1.06	1.29	1.51	1.85	1.55	1.29
Al_2O_3	13.25	14.01	14.23	13.86	13.83	14.02	13.58	13.82	13.24	14.16	13.54	14.43
Fe ₂ O ₃ *	14.31	14.07	12.79	14.77	13.25	13.69	13.75	15.37	16.32	18.58	15.72	15.78
MnO	0.21	0.22	0.20	0.22	0.27	0.21	0.21	0.25	0.24	0.42	0.24	0.34
MgO	6.37	6.56	7.47	6.28	7.86	7.02	7.38	6.03	5.97	6.41	6.24	7.64
CaO Na O	2 42	2 21	2.06	2 45	9.57	2 45	2 26	9.08	8.38	2.79	2.41	8.88 2.05
K ₂ O	0.42	0.25	0.23	0.32	0.39	0.26	0.23	0.32	0.18	0.10	0.15	2.03
P ₂ O ₅	0.15	0.14	0.11	0.15	0.10	0.20	0.09	0.13	0.14	0.17	0.15	0.00
LOI	0.42	0.81	0.77	0.73	0.94	0.48	0.49	4.89	1.48	2.36	0.87	4.28
Total	99.47	100.38	99.59	99.64	99.77	100.01	99.69	99.53	99.65	99.63	99.83	100.30
Cs	10	0.55							10			
KD Do	13	4.04	2	8	16	1	122	129	12	2	4	2
Th	238	0.68	2	150	2	157	155	158	528	47	15	155
U		0.21										
Nb	5	5.43	4	5	4	4	4	6	4	7	4	3
18	5	0.52	6	5	4	3	5	6	3	0	8	8
Ce	14	14.7	13	14	16	13	12	13	16	16	9	13
Pb	4	3.05	4	5	6	4	11	12	2	5	2	3
Pr		2.25										
Sr	189	178	235	180	223	149	150	256	395	39	111	160
Nd	11.1	10.9	6.6	11.4	10.8	9.2	7.3	10.1	10.9	13.5	10.6	9.5
Sm		3.22										
Zr	80	76	67	81	59	61	57	80	90	116	95	77
En		2.00										
Gd		3.89										
Tb		0.68										
Dy		4.25										
Y	24.6	22.7	20.4	24.3	19.5	19.8	20.5	27.3	28.7	35.9	30.7	24.3
Но		0.90										
Er		2.58										
Thi Vh		2.28										
Lu		0.34										
V	405	393	349	411	355	384	364	498	461	564	473	389
Cr	82	136	242	145	272	180	106	89	73	78	142	128
Ni	51.9	65	92.2	62.5	97	83.3	71.5	52.3	63.2	55.4	62.7	138.5
Cu	105	112	107	145	91	59	74	234	215	159	169	65
Zn Ga	120	106 18.67	101	127	108	103	139	160	136	155	130	128
7r/Nib	16.52	15 04	17.01	15 70	16.10	15 20	14 46	14.27	20 77	16.05	21 01	24.42
Y/Nh	5 08	4 87	5 43	4 73	5 33	5.01	14.40 5.18	4 93	20.77 6.60	5 26	21.81	24.42 7.68
Ti/V	22.37	21.94	20.62	22 74	19 75	20.30	17 47	15 52	19.61	19.67	19.65	19.87
Zr/Y	3.26	3.26	3.30	3.34	3.02	3.06	2.79	2.92	3.15	3.22	3.09	3.18
Ti/Zr	112.98	114.77	106.87	115.14	119.05	128.78	111.07	97.13	100.11	95.83	97.89	100.02
Ti/Y	367.89	373.62	352.55	384.76	359.61	393.51	309.90	283.21	315.33	308.85	302.60	318.17
Th/Nb		0.12	0.51		0.49							
Nb/La	0.90	0.92	0.68	1.05	1.02	1.24	0.86	0.88	1.67	0.78	0.54	0.42
Ta/Yb		0.23										
(La/Sm)n'		1.10										
$(Gu/YD)n^{\dagger}$		1.00										
(EwEd.)II												

Sample:	228	229	230	231	232	233	234	236	237	240	241	242
Unit: Rock type:	DHM	DHM	CMD Dolorito	DHM	CMD	CMD Dolorito						
Domain ⁻	South Head	South Head	South Head	South Head	South Head	South Head						
Method^:	XRF	XRF	XRF	XRF	XRF	XRF	ICPMS	XRF	XRF	XRF	ICPMS	XRF
SiO ₂	47.58	48.83	49.03	46.10	47.69	47.19	51.53	48.48	48.60	49.16	48.66	49.91
TiO ₂	2.18	2.45	1.19	1.29	1.13	1.15	1.18	1.13	1.15	1.20	1.46	1.46
Al_2O_3	13.42	12.58	13.70	14.83	13.84	14.08	12.97	13.95	14.06	14.22	13.49	13.07
Fe ₂ O ₃ *	18.43	19.50	14.18	15.14	14.21	14.79	14.56	14.17	14.37	13.57	15.61	14.54
MnO	0.27	0.27	0.20	0.22	0.21	0.23	0.24	0.22	0.22	0.21	0.22	0.19
MgO G-O	5.85	6.34	0.51	0.64	/.40	8.02	/./0	10.20	/.51	/.06	7.40	/.12
Na.O	2.76	1.00	0.96	1 75	0.48	2 32	2.57	2 33	9.20	1 08	3.00	0.22 2.54
K ₂ O	0.42	0.10	0.03	0.03	0.48	0.08	0.09	0.10	0.10	0.14	0.08	0.05
P ₂ O ₅	0.24	0.25	0.12	0.12	0.11	0.10	0.10	0.10	0.10	0.11	0.14	0.13
LOI	1.60	2.36	1.73	1.85	2.57	1.80	2.74	2.10	2.27	2.20	2.70	2.21
Total	99.59	100.27	100.22	99.96	100.08	99.80	100.23	100.19	99.99	100.34	100.23	99.44
Cs							0.22				0.24	
Rb	30	3	10	25	1	01	1.16	2	2	4	1.83	21
Da Th	520	33	19	23	17	91	0.37	14/	50	108	0.70	21
U							0.13				0.70	2
Nb	8	9	3	4	3	3	3.53	3	3	4	6.00	5
Та							0.37				0.48	
La	14	10	6	2	5	4	3.44	7	5	6	6.07	6
Ce	25	18	10	10	9	5	9.08	8	8	9	15.4	13
Pb	4	3	2				0.45				1.21	
Pr							1.46		405	1.50	2.41	
Sr	54	137	313	163	270	110	103	134	105	179	57	63
Sm	17.5	14.2	0.0	/.0	1.5	0.7	2.44	0.7	0.0	1.1	3.66	10.0
Zr	166	151	64	68	57	64	60	56	60	69	98	91
Hf	100	101	0.	00	27	0.	1.69	50	00	0,	2.74	<i>,</i>
Eu							0.58				1.15	
Gd							3.12				4.52	
Tb							0.57				0.80	
Dy							3.63				4.97	
Y	46.3	41.3	20.9	23.3	20.3	20.3	20.0	19.8	20.9	22.1	27.3	24.6
Но							0.78				1.06	
Er							2.24				5.05	
Yh							2.12				2.88	
Lu							0.32				0.44	
v	445	494	410	437	381	414	360	374	374	366	411	418
Cr	86	75	160	172	190	207	193	205	216	168	117	102
Ni	70.5	57.8	104.3	108.4	125.4	115.9	104	104.2	104.2	91.8	79	70.6
Cu	198	179	173	190	158	154	132	151	168	146	150	182
Zn Ga	192	224	105	124	109	117	126 11.24	108	113	101	123 18.32	116
7r/Nib	20.02	15.00	24 77	16 70	17.24	22.78	10.64	20.00	22.22	17.86	10.20	18.00
ZI/IND V/Nib	20.92	15.90	24.//	10.79	6.22	25.78	19.64	20.90	23.33	5 73	19.39 5.30	18.00
Ti/V	29.36	29.70	17.38	17.68	17 79	16.64	19.63	18.09	18 42	19.65	21.31	20.95
Zr/Y	3.57	3.66	3.05	2.92	2.79	3.13	2.76	2.82	2.87	3.12	3.60	3.69
Ti/Zr	78.95	97.24	111.97	113.53	119.66	108.54	130.00	121.37	114.87	104.38	89.47	96.37
Ti/Y	282.20	355.54	341.25	331.83	333.62	339.53	359.00	342.05	329.78	325.44	321.71	355.71
Th/Nb							0.11				0.12	0.32
Nb/La	0.58	1.00	0.45	1.93	0.73	0.65	1.03	0.40	0.56	0.62	0.99	0.83
Ta/Yb							0.17				0.17	
(La/Sm)n [†]							0.89				1.04	
(Gd/Yb)n ^T							1.19				1.27	
(Eu/Eu*)n'							0.04				0.80	

Sample:	243	249	250	251	253	256	260	261	262	268	269	270
Unit:	DHM	CMD	DHM	DHM	DHM	DHM	DHM	DHM	DHM	DHM	DHM	DHM
Rock type:	Basalt	Dolerite	Basalt	Basalt	Basalt	Basalt	Amphibolite	Amphibolite	Amphibolite	Amphibolite	Amphibolite	Amphibolite
Domain:	South Head	Robinhood	Robinhood	Robinhood	Robinhood	Robinhood	Robinhood					
Method^:	XRF	ICPMS	ICPMS	XRF	XRF	XRF	XRF	XRF	XRF	XRF	ICPMS	XRF
SiO	50.37	49 30	47.65	49.23	49 42	49 32	47 72	50.28	47 91	46 33	50.04	49 64
TiO ₂	1 44	1 89	1.62	1.51	1.52	1 92	1 33	1 40	1.80	2.28	1 43	2.03
Al ₂ O ₃	13.18	12.61	12.36	13.10	13.48	11.61	14.91	14.03	13.06	12.02	13.14	12.87
Fe ₂ O ₃ *	14.44	16.23	16.71	15.71	15.58	16.03	14.67	14.95	17.16	19.84	15.45	17.81
MnO	0.20	0.28	0.29	0.24	0.22	0.21	0.23	0.21	0.24	0.26	0.25	0.25
MgO	6.69	5.61	5.08	6.66	5.74	4.49	7.04	6.17	6.73	6.05	6.59	5.53
CaO	9.36	8.76	11.42	8.58	9.35	6.30	9.80	8.31	9.86	10.13	9.65	9.66
Na ₂ O	2.24	2.87	1.71	2.66	1.54	0.67	2.77	3.93	1.85	1.95	2.54	0.86
K_2O	0.04	0.34	0.13	0.05	0.05	0.05	0.15	0.11	0.28	0.17	0.23	0.46
P_2O_5	0.14	0.18	0.17	0.13	0.14	0.19	0.11	0.13	0.19	0.14	0.12	0.21
LOI	2.10	2.06	2.71	2.56	3.19	8.85	1.09	0.29	0.51	0.52	0.64	0.41
Total	100.20	100.13	99.85	100.43	100.23	99.64	99.82	99.81	99.59	99.69	100.08	99.73
Cs		0.77	0.19								2.29	
Rb	10	13.2	2.99	20	12	2	7	1	9	9	10.8	12
Ba ть	13	146	8/	30	43	22	118	109	146	160	124	198
		0.72	0.75						2	2	0.40	
Nh	5	7 79	6.42	5	5	8	4	5	6	4	4 70	7
Та	5	0.63	0.53	5	5	0	-	5	0	-	0.39	/
La	7	7.65	6.41	7	10	9	4	7	8	6	4.32	11
Ce	13	19.4	16.7	15	19	19	13	19	17	14	11.3	24
Pb		2.53	1.99		3	4	9	3	7	6	1.14	28
Pr		3.07	2.63								1.79	
Sr	48	139	43	168	290	85	245	310	153	343	185	310
Nd	9.2	15.2	13.1	9.3	13.3	12.5	9.8	11.9	16.2	8.7	9.10	17
Sm		4.60	4.12								3.04	
Zr	93	124	110	82	98	130	79	91	98	82	73	129
Ht		3.36	3.00								2.03	
Eu		5.86	5.20								2.02	
Th		1.05	0.93								0.73	
Dv		6.41	5.81								4.53	
Y	25.2	35.6	32.9	27.3	28.6	35.4	25.7	27.3	32.6	27.5	25.6	39.1
Но		1.36	1.27								1.01	
Er		3.84	3.66								2.91	
Tm		0.58	0.55								0.44	
Yb		3.55	3.52								2.71	
Lu		0.53	0.51								0.42	
V	432	495	503	481	412	546	422	413	475	999	456	538
Cr	98	/1	54 46	/9	67 2	30 40.2	2/5	104	107	047	104	52.0
NI Cu	197	179	40	199	122	40.5	04.4	162	109.1	94.7 186	202	152.9
Zn	120	142	129	124	123	134	144	112	165	167	114	152
Ga	120	20.47	20.73	124	127	154	144	117	150	107	18.21	152
Zr/Nb	18 16	19 43	16 65	16 96	18 28	17 35	21 10	19 49	15 20	19 35	16 79	18 16
Y/Nb	4.90	5.43	4.92	5.63	5.36	4.71	6.84	5.87	5.07	6.47	6.21	5.49
Ti/V	19.99	22.88	19.32	18.83	22.10	21.06	18.89	20.33	22.74	13.68	18.79	22.60
Zr/Y	3.71	3.58	3.38	3.01	3.41	3.68	3.09	3.32	3.00	2.99	2.70	3.31
Ti/Zr	92.40	92.09	88.03	110.10	93.34	88.25	100.52	92.61	110.42	166.04	120.04	94.10
Ti/Y	342.48	329.29	297.83	331.50	318.53	325.07	310.17	307.36	330.93	496.91	324.64	311.17
Th/Nb		0.09	0.11						0.26	0.47	0.08	
Nb/La	0.73	1.02	1.00	0.75	0.56	0.88	0.87	0.70	0.82	0.77	1.09	0.63
Ta/Yb		0.18	0.15								0.14	
(La/Sm)n ^T		1.05	0.98								0.89	
(Gd/Yb)n'		1.55	1.19								1.1/	
(Eu/Eu*)n'		0.93	0.93								0.97	

Sample: Unit:	296 CMD	299 CMD	300 CMD	301 CMD
Rock type:	Amphibolite	Amphibolite	Amphibolite	Amphibolite
Domain:	Gum Flats	Gum Flats	Gum Flats	Gum Flats
Method^:	XRF	ICPMS	XRF	ICPMS
SiO.	48 99	47.96	48 50	50.27
TiO ₂	0.84	2.85	2.66	1.56
Al ₂ O ₂	14.50	12.37	13.03	13.64
Fe ₂ O ₃ *	11.59	21.11	20.20	13.43
MnO	0.22	0.35	0.33	0.20
MgO	8.04	3.28	3.18	6.95
CaO	12.26	7.60	7.76	10.53
Na ₂ O	1.62	2.76	2.78	1.98
K ₂ O	0.64	0.58	0.57	0.39
P_2O_5	0.09	0.28	0.33	0.16
LOI	1.59	0.54	0.42	0.83
Total	100.38	99.68	99.76	99.94
Cs		1.54		2.60
Rb	42	15.2	23	20.2
Ba	76	81	122	119
Th		1.39	3	1.08
U		0.29		0.22
Nb	4	11.6	8	5.82
Ta		0.67	0	0.54
La	4	7.75	8	/.69
Ce	10	20.3	26	17.4
Pb De	21	/./5	10	14.2
ri Sr	150	3.10	172	2.79
Nd	58	192	16.3	13.2
Sm	5.6	5 28	10.5	3.86
Zr	41	124	118	91
Hf	41	3 45	110	2.53
Eu		2.17		1.35
Gd		6.54		4.39
Tb		1.19		0.78
Dy		7.24		4.64
Y	19.5	39.4	38.8	24.0
Но		1.57		0.96
Er		4.52		2.68
Tm		0.65		0.40
Yb		4.18		2.34
Lu		0.61		0.35
v	327	84	55	382
Cr	164	8	8	127
N1	104.9	5	4.7	97
Cu Zu	123	/8	24	190
Zn Ga	1/8	260	204	18.49
7.01	10.47	11.27	15.00	17.40
ZI/IND V/Nib	10.47	2.80	15.28	17.40
Ti/V	4.93	203 50	288.81	4.30
Zr/Y	2 12	203.39	3 04	3.81
Ti/Zr	121.61	146.12	135.07	102.18
Ti/Y	258 18	427.03	410.89	389 57
Th/Nb	200.10	0.12	0.43	0.18
Nb/La	1.13	1.50	0.93	0.76
Ta/Yb		0.16		0.23
(La/Sm)n [†]		0.92		1.25
(Gd/Yb)n [†]		1.26		1.51
(E (E*)†		1.13		1.00

		Ortona		South	n Head	Stockyard	Robinhood	Gum Flats
No. of analyses	metabasalt 17	metadolerite 28	granophyre 1	metabasalt 14	9	metadolerite 9	DHM amphibolite 6	CMD amphibolite
SiO ₂	47.37	48.67	59.89	48.68	48.04	48.97	48.65	48.93
TiO ₂	1.63	1.43	1.06	1.49	1.49	1.36	1.71	1.98
Al_2O_3	13.88	13.49	12.96	13.38	13.60	13.82	13.34	13.39
Fe ₂ O ₃ *	16.38	14.81	6.73	15.50	15.84	13.97	16.65	16.58
MnO	0.26	0.22	0.16	0.23	0.27	0.23	0.24	0.28
MgO	6.40	6.36	1.84	6.54	6.67	6.88	6.35	5.36
CaO	7.57	9.09	4.48	9.48	8.52	10.69	9.57	9.54
Na ₂ O	2.92	2.50	6.06	1.78	2.56	2.74	2.32	2.29
K ₂ O	0.07	0.36	0.36	0.10	0.16	0.29	0.23	0.55
P_2O_5	0.16	0.13	0.20	0.14	0.14	0.13	0.15	0.22
LOI	3.29	2.94	6.12	2.74	2.57	0.70	0.58	0.85
Total	99.94	100.00	99.86	100.07	99.86	99.77	99.79	99.94
Cs	0.16	1.01		0.21	0.51	0.55	2.29	2.07
Rb	3	16	18	6	8	6	8	25
Ba^	34	109	112	79	115	188	142	100
Th	1	1	14	1	1	1	1	2
U Nb^	0.30	0.30	13	0.19	0.27	0.21	0.17	0.25
Та	0.47	0 44	15	0.45	0.56	0.52	0 39	0.60
La	5	6	34	7	6	5	7	7
Ce	13	15	71	13	14	14	16	19
Pb	3	6	10	3	4	5	9	13
Pr	2.48	2.49		2.05	2.74	2.25	1.79	2.98
Sr	72	163	126	150	147	185	258	169
Nd	11	11	33	10	11	10	12	13
Sm	3.83	3.50		3.28	4.13	3.22	3.04	4.57
Zr^	104	86	199	90	92	70	92	94
Hf	2.30	2.15		2.34	3.05	2.06	2.03	2.99
Eu	1.12	1.18		1.01	1.38	1.16	1.10	1.76
Gd	4.78	4.31		4.16	5.19	3.89	3.92	5.47
Tb	0.87	0.77		0.75	0.92	0.68	0.73	0.98
Dy	5.44	4.86		4.72	5.69	4.25	4.53	5.94
Y	28	25	35	27	28	22	30	30
Но	1.16	1.04		1.02	1.21	0.90	1.01	1.26
Er	3.29	2.96		2.95	3.45	2.58	2.91	3.60
Tm	0.48	0.43		0.45	0.52	0.37	0.44	0.53
Yb	3.02	2.76		2.82	3.21	2.28	2.71	3.26
Lu	0.44	0.40		0.42	0.48	0.34	0.42	0.48
V	419	398	174	433	454	384	550	212
Cr^	117	87	2	126	113	157	115	77
Ni^	72	91	6	83	77	73	80	53
Cu^	139	174	119	171	169	105	235	104
Zn	140	111	26	131	134	118	142	188
Ga	18.35	18.50		15.98	19.39	18.67	18.21	20.62
Zr/Nb	16.31	16.35	15.14	18.95	19.83	15.61	18.35	13.64
Y/Nb	4.43	4.92	2.65	5.98	6.15	4.95	5.99	4.61
Ti/V	22.87	21.91	36.45	20.52	19.76	21.09	19.50	133.07
Zr/Y	3.77	3.51	5.70	3.22	3.25	3.16	3.07	2.97
Ti/Zr	94.40	104.50	31.91	104.13	98.20	115.75	113.95	126.28
Ti/Y	354.85	362.22	182.04	330.42	318.29	365.56	346.86	371.42
Th/Nb	0.18	0.19	1.03	0.11	0.18	0.37	0.27	0.24
Nb/La	1.50	0.95	0.38	0.81	0.85	0.91	0.82	1.08
Ta/Yb	0.16	0.16		0.16	0.17	0.23	0.14	0.19
(La/Sm)n [†]	1.02	1.20		0.93	1.05	1.16	0.89	1.09
(Gd/Yb)n [†]	1.28	1.27		1.19	1.30	1.38	1.17	1.39
(Eu/Eu*)n [†]	0.77	0.95		0.80	0.91	1.00	0.97	1.06

^ XRF values preferred

† Normalising values after Boynton (1984)

Table 6.2: Average geochemical compositions for mafic rocks by domain within the centralGeorgetown Inlier. Domains are illustrated in Fig. 6.1.

6.5 Results

6.5.1 Affinities and classification of the Etheridge Group mafic rocks

This study shows that Etheridge Group mafic rocks are a sequence of comagmatic metabasalts and metadolerites that resemble modern, relatively evolved, low-K tholeiites matching those generated during the early stage of continental rifting.

Mobility of the alkali elements during metamorphism compromises classification using the total alkalis – SiO_2 diagram, yet the Etheridge Group mafic rocks (excluding several granophyres) cluster in the subalkaline basalt field (Fig. 6.2a). On the Zr/TiO₂ versus Nb/Y diagram of Winchester & Floyd (1977) (Fig. 6.2b), in which Zr/TiO₂ effectively proxies for variations in SiO₂, and Nb/Y reflects alkalinity, both the DHM and CMD show a tight and overlapping distribution, and they are confidently assigned as subalkaline mafic suites. Further, the pronounced clustering of the DHM and CMD samples, despite the observed variation in crystallisation ages (Chapter 4) and spatial distribution, suggests that these units may be comagmatic. Most DHM and CMD have 5-8% MgO and are thus relatively evolved, in keeping with the apparent absence of olivine phenocrysts in samples from either group.



Figure 6.2: (a) Total Alkalis-Silica (TAS) volcanic rock classification diagram (Le Bas *et al.*, 1986). (b) Zr/TiO₂ versus Nb/Y volcanic rock classification diagram (Winchester and Floyd, 1977).

Strong Fe-, Ti- and V-enrichment with advancing fractionation (Figs. 6.3 and 6.4) shows clearly that both the DHM and CMD suites have tholeiitic affinities. Consistently low K₂O contents in both CMD and DHM (average values for K₂O in both the DHM and CMD 0.08 and 0.31 wt% respectively for greenschist facies samples; Table 2) suggest probable low-K tholeiite affinities. This is strongly supported by the Ti/Zr values that cluster between 80 and 120 for the majority of samples (Fig. 6.5a), a range typical of low-K tholeiites (Pearce, 1983).

Total iron (as Fe₂O₃) contents for both the DHM and CMD vary significantly, but predominantly lie within the range of 13 to 18 wt% (Table 6.1), defining a strong Feenrichment trend between 4 and 9 wt% MgO (Fig. 6.3). Four samples of the CMD do not follow this trend, and include three felsic differentiates and one granophyre sample, all from Ortona. However, one of the felsic differentiates has an LOI value



>15 wt%, and was probably subject to greater than average element mobility (Sample 8; Table 6.1).

Figure 6.3: Major element compositions of the lower Etheridge Group mafic rocks plotted against MgO. Additional Etheridge Group mafic rock data from Withnall (1985).



Figure 6.4: Selected trace element compositions (in ppm) of the lower Etheridge Group mafic rocks plotted against MgO. Symbols as for Figure 6.3.



Figure 6.5: Selected immobile trace element diagrams for mafic rocks of the lower Etheridge Group. Additional Etheridge Group mafic rock data from Withnall (1985). Average N-MORB and E-MORB values after Sun and McDonough (1989).

Despite the regionally extensive spread of samples collected for this study, variation diagrams for major and trace elements within the DHM and CMD show significant overlap between the DHM and CMD (Figs. 6.3 and 6.4). The majority of samples have basaltic SiO₂ contents (45-53 wt%), with granophyres from CMD sills having higher SiO₂ contents (to 60 wt%). MgO ranges from 10 - 2 wt%, and most elements/oxides regarded as immobile during alteration show good correlations against MgO (e.g. TiO₂, Al₂O₃, Fe₂O₃, P₂O₅, Cr, Ni, V, Zr, Y, Nb; Figs. 6.3 and 6.4). The pronounced overlap on all immobile element – MgO diagrams (Fig. 6.4) is also consistent with the DHM and CMD having being derived from chemically similar parent magmas.

6.5.2 Trace- and rare earth element characteristics of the Etheridge Group mafic rocks

MORB-normalised trace-element diagrams and chondrite-normalised REE diagrams are presented in Figures 6.6 and 6.7 respectively for 25 representative samples of the DHM and CMD (Table 6.1). Although Rb, Ba, Pb, K and Sr were all likely to have been mobile to variable extents during post-magmatic alteration (as shown by the

scatter of points on element-MgO diagrams), K, Pb and Sr are included on the N-MORB-normalised plots to emphasise a number of points regarding element budgets during low-grade metamorphism.



Figure 6.6: MORB-normalised trace element diagrams for representative samples of the DHM and CMD. Normalising values after Sun and McDonough (1989).

N-MORB-normalised multi-element patterns for eight well preserved DHM (Fig. 6.6a) show similar shapes, with the most notable differences being for the mobile elements Sr and Eu, both reflecting albitisation of plagioclase (see below). A striking feature of all patterns is the significant negative K anomalies. I believe these reflect chlorite-albite-silica alteration of glass and glassy mesostasis in the basalts, a feature not seen in the dolerites, which consequently show generally higher K₂O contents (Fig. 6.6b), probably close to their primary values. Six of the eight basalts show strong positive Pb anomalies, a feature also characteristic of the majority of CMD patterns (Fig. 6.6b), and these are considered to be primary. These patterns show slight enrichment in LILE (Th, U, and La) relative to the similarly incompatible high field strength elements Ta and Nb.


Figure 6.7: Chondrite-normalised REE diagrams for representative samples of the DHM and CMD. Normalising values after Boynton (1984).

Chondrite-normalised REE patterns for the DHM and CMD (Fig. 6.7) range from weakly LREE-enriched to moderately LREE-depleted (avg. $(La/Sm)_n = 1.11$; range 0.81-1.68), with flat to weakly HREE-enriched HREE patterns (avg. $(Gd/Yb)_n = 1.28$; range 1.11-1.64). Chondrite-normalised REE diagrams display the expected upward shift to higher total REE values with increased fractionation, as shown in Figure 6.8, in which all 25 representative samples are grouped in order of their MgO contents. DHM and CMD samples from both the South Head and Ortona areas (Fig. 6.8) have similar REE patterns; however, the metabasalts show more pronounced negative Euanomalies (avg. $(Eu/Eu^*)_n = 0.80$ for DHM and 0.96 for CMD samples; Fig. 6.7a; Table 6.1). There is no significant difference in extent of fractionation between the Ortona DHM and CMD for which REE were analysed. Thus the significant negative Eu anomalies in the metabasalts relative to compositionally similar (in terms of major elements and other REE apart from Eu) metadolerites presumably reflects increased leaching of plagioclase-hosted and thus relatively mobile Eu (compared to the

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immobile +3 valence state adjacent REE) during albitisation. Plagioclase is always albitised in the basalts, whereas calcic plagioclase is often preserved in the dolerites.



Three CMD samples from Ortona (Samples 6, 22, and 112) and one CMD amphibolite from Gum Flats (Sample 301) show distinct LREE-enriched, slightly HREE-depleted patterns ((La/Sm)_n = 1.67, 1.47, 1.68, and 1.25 respectively). These samples have among the highest La/Nb and Th/Nb values for Etheridge Group mafic rocks, which may suggest (Pearce, 1983; James *et al.*, 1987) that the LREE-enrichment reflects some crustal contamination.

A simple mixing relationship between the Etheridge Group mafic rocks and the host metasediments of the lower Etheridge Group could account for this variation. In Figure 6.9, a clear trend towards increasing degrees of contamination by the host metasediments is apparent, with the most contaminated dolerite sill-margin samples showing up to 20% assimilation of sedimentary material. However, the majority of samples contain less than 10% sedimentary material. A granophyric sample from the margin of a CMD sill (sample 174) shows approximately 40% assimilation of sedimentary material, and also lies within the field of the lower Etheridge Group



metasediments (Fig. 6.9). This relationship suggests that the lower Etheridge Group metasediments may have been the only contaminant of the mafic rocks.

Figure 6.9: Th/Nb versus Nb/La diagram for mafic and metasedimentary rocks of the lower Etheridge Group showing possible mixing curve between the lithologies. The mixing curve was calculated using the following parameters for the uncontaminated Etheridge Group mafic rocks (Th/Nb = 0.096, Nb/La = 1.189, Th = 1 ppm, Nb = 6 ppm, and La = 5 ppm), and for the Etheridge Group metasediments (Th/Nb = 1.889, Nb/La = 0.273, Th = 17 ppm, Nb = 9 ppm, and La = 33 ppm). Additional Etheridge Group mafic data after Withnall (1985). Lower Etheridge Group metasedimentary data available from the Geoscience Australia OZCHEM Database. Average N-MORB and E-MORB values after Sun and McDonough (1989). Average continental crust value after Weaver and Tarney (1984).

Fractionation-independent Zr/Nb values mainly range from 12-20 and show no significant regional or stratigraphic variation, with CMD overlapping the compositional field of the DHM. The same applies to a variety of immobile element ratios shown in Figure 6.5, strongly supporting the suggestion from major and trace element data and REE patterns that the DHM are comagmatic with the CMD, and that the latter were likely feeders to the basalts. Dolerite sills emplaced after eruption of the DHM into the overlying Lane Creek Formation are compositionally identical to those emplaced during the main magmatic episode.

6.5.3 Compositional variations across-sill for CMD, and across stratigraphic for DHM

Wholerock geochemical data were measured for a series of samples taken systematically across the outcropping width of a DHM flow near 'Ortona' (650 m thick), and a thick CMD sill at Stockyard Creek (395 m thick). Detailed sampling was also carried out across the full width of the best exposed basalt-dolerite pile outcrops along the Gilbert River at South Head (~ 3000 m true thickness).

Figure 6.10 illustrates the chemical variations across the sills. Marginal samples have MgO contents < 7%, in keeping with the generally evolved nature of the magmas that formed the CMD and DHM. There appears to have been no significant accumulation of olivine in the lower sections of any sill sampled, in accordance with the absence of olivine in the relatively evolved marginal variants of each sill, although each sill shows generally higher MgO contents in its lower third to lower half. TiO₂ and Fe₂O₃ are generally higher in the upper half of both sills. The antithetic relationship between the more compatible (e.g. Ni) and more incompatible (e.g. Zr) elements is as expected from limited in-situ fractionation of a single magma batch.

Relative to many mafic sills 500-1000 m thick that have crystallised essentially insitu, the sampled sills in the Lower Etheridge Group show surprisingly limited internal fractionation, with only relatively rare and small volumes of more felsic granophyres with ~60 wt% SiO₂. This may indicate that the sills were essentially open system conduits in which continuous provision of magma essentially added to and thickened the sills, rather than having been static, closed system bodies.

The thick (>3000 m) interlayered basalt-dolerite sequence at South Head (Fig. 6.10a and d) shows a slight decrease in MgO from the base to the top of the sequence, and sharp increase in Fe_2O_3 and TiO_2 in basalts from near the top of the pile, the latter probably reflecting expulsion of relatively late stage fractionation products from a deeper stagnating magma chamber.

The compositional variation between CMD sills emplaced stratigraphically below, within, and above the DHM is insignificant (Table 6.2; Figure 6.11). The CMD sill emplaced within the Lane Creek Formation at Stockyard Creek, above the level of the DHM, is compositionally similar (including REE) to CMD sills within and beneath the DHM, suggesting no significant change in the source mantle or partial melting/fractionation processes over the time interval represented by the stratigraphic

section from the DHM to the Lane Creek Formation (keeping in mind that many sills in the Bernecker Creek Formation and Daniel Creek Formation *below* the DHM may have been emplaced pre-, syn- or post-DHM).



Figure 6.10: Major oxide (wt %) and trace element (ppm) variation with stratigraphic height for selected CMD sills and DHM layers. (a) and (d): interlayered CMD and DHM sequence along Gilbert River at South Head; (b) and (e): CMD sill at Stockyard Creek; (c) and (f): CMD sill at Ortona.



Figure 6.11: MgO versus Fe₂O₃* showing geochemical variation in CMD sills emplaced stratigraphically below, within and above the level of the DHM.

6.5.4 Sm-Nd radiogenic isotope geochemistry of representative Etheridge Group mafic rocks

The relative immobility of both Sm and Nd compared to Rb, Sr, U and Pb makes the Sm-Nd isotopic system particularly useful for determining both the petrogenetic characteristics and age of older (Proterozoic to Achaean) terranes that may have been affected by multiple stages of deformation (e.g. Black & McCulloch, 1984; Zhao & McCulloch, 1995; Blewett *et al.*, 1998). In general, concentrations of both Sm and Nd in igneous rocks increase with increasing degrees of differentiation; however the Sm/Nd ratio decreases with fractionation, since Nd is more incompatible than Sm. Similarly, tholeiitic basalts that have assimilated crustal material (which is always LREE-enriched) will have lower Sm/Nd than unmodified magmas from the same source. Thus Sm-Nd isotopic systematics are useful for evaluating crustal interaction in mafic magmas.

For this study, three samples of DHM and seven samples of CMD were analysed for their Sm and Nd isotopic compositions (Table 6.3). Initial isotopic compositions for the individual samples are listed in the form of $\varepsilon_{Nd}(T)$ values, which were calculated for each sample assuming a magmatic age of 1.67 Ga, and the chondritic uniform reservoir (CHUR) to have a present-day ¹⁴⁷Sm/¹⁴⁴Nd value of 0.1967 and a ¹⁴³Nd/¹⁴⁴Nd value of 0.512638 (Wasserburg *et al.*, 1981).

Sample No.		Sm (ppm)	Nd (ppm)	147Sm/144Nd	143Nd/144Nd	εNd(t)	$f_{\rm Sm/Nd}$	T _{DM} (Ga)	La/Sm
MAFIC									
Dead Horse Metabasalt (DHM)		εNd(1.67 Ga)							
100		3.52	10.43	0.2042	0.512937	4.2	0.04	3.39	1.33
234		2.39	7.49	0.1926	0.512798	4.0	-0.02	2.53	1.44
269		2.94	9.15	0.1941	0.512824	4.2	-0.01	2.52	1.47
Cobbold Metadolerite (CMD)									
6		3.87	14.58	0.1605	0.512388	2.9	-0.18	2.17	2.60
113		2.10	6.70	0.1895	0.512781	4.3	-0.04	2.31	1.61
181		1.76	5.67	0.1879	0.512717	3.4	-0.04	2.55	1.84
196		3.19	11.03	0.1749	0.512590	3.8	-0.11	2.19	1.86
249		4.68	15.46	0.1828	0.512692	4.0	-0.07	2.25	1.64
299		5.19	16.20	0.1936	0.512737	2.6	-0.02	3.11	1.49
301		3.73	13.25	0.1701	0.512618	5.3	-0.14	1.85	2.06
Black and McCulloch (1984)									
73303006	CMD	2.48	8.52	0.1757	0.512492	1.7		2.63	
79300232	CMD	1.76	5.59	0.1905	0.512807	4.6		2.24	
82303067	Qtz diorite	45.75	247.70	0.1116	0.5116	-2.0		2.30	
82303068	Granophyre	4.74	27.43	0.1045	0.511358	-5.2		2.49	
80303065	Mafic granulite	2.48	7.62	0.1971	0.51287	4.5		2.56	
80303066	Mafic granulite	2.11	6.52	0.1958	0.512815	3.7		2.84	
Knutson and Sun (1997)									
92836540	Amphibolite	3.67	11.96	0.1857	0.512718	4.0		2.34	
92836504B	CMD	5.93	29.04	0.1235	0.511674	-3.0		2.48	
FELSIC									
Black and McCulloch (1984)					εNd(1.685 Ga)				
73303024	Einasleigh Metamorphics	7.52	42.77	0.1063	0.511376	-5.1		2.51	
73303001	Einasleigh Metamorphics	7.56	43.02	0.1064	0.511364	-5.4		2.52	
79300058	Candlow Formation	1.19	5.76	0.1251	0.511916	1.4		2.11	
79300059	Candlow Formation	1.14	5.33	0.1291	0.511965	1.5		2.13	
79300060	Candlow Formation	1.62	6.84	0.1434	0.511885	-3.2		2.73	
79300060	Candlow Formation	1.66	7.06	0.1420	0.511897	-2.6		2.65	
79300062	Candlow Formation	1.33	6.85	0.1171	0.51175	-0.1		2.20	
79300062	Candlow Formation	1.28	6.67	0.1162	0.511721	-0.5		2.22	
Knutson and Sun (1997)									
92836520	Corbett Formation schist	9.29	54.22	0.1036	0.511363	-4.8		2.46	7.23
92836536	Townley Formation schist	8.83	48.30	0.1105	0.511719	0.7		2.11	8.61

$$\begin{split} f_{SmNM} &= [(^{147} \text{ Sm}/^{144} \text{ Nd})_{sample/}('^{147} \text{ Sm}/^{144} \text{ Nd})_{CHUR}]\text{-}1 \\ CHUR (chondrite uniform reservoir): $^{147} \text{ Sm}/^{144} \text{ Nd} = 0.1967, $^{143} \text{ Nd}/^{144} \text{ Nd} = 0.512638 \\ DM (depleted mantle): $^{147} \text{ Sm}/^{144} \text{ Nd} = 0.2137, $^{143} \text{ Nd}/^{144} \text{ Nd} = 0.51315 \\ \lambda &= decay \ constant \ of $^{147} \text{ Sm} = 6.54 \times 10^{-12} \end{split}$$

Table 6.3: Sm-Nd isotopic data and depleted mantle model ages for rocks of the lower Etheridge Group. Sm and Nd ppm values for mafic rocks of this study taken from MC-ICPMS analyses.

The analysed Etheridge Group mafic rocks show only a limited range in initial Nd isotopic compositions, with 7 analyses having $\varepsilon_{Nd}(T)$ values between +3.4 and +4.3. Samples 6 and 299 have isotopic compositions of +2.9 and +2.6 respectively, whereas sample 301 had the highest $\varepsilon_{Nd}(T)$ value of +5.3 (Table 3). The $\varepsilon_{Nd}(T)$ values for these samples are identical to those for samples of the CMD and related rocks reported by Black & McCulloch (1984), and clearly indicate that the majority of these Georgetown Inlier mafic rocks are broadly co-magmatic and were derived from partial melting of a depleted upper mantle source.

Sm-Nd isochron age for mafic igneous rocks of the lower Etheridge 6.5.5 Group

The ¹⁴⁷Sm/¹⁴⁴Nd and ¹⁴³Nd/¹⁴⁴Nd values for the Etheridge Group igneous rocks show a relatively good linear correlation on the Sm-Nd isochron diagram (Fig. 6.12). However, contamination by crustal material in a number of samples (see below), as well as the relatively short isochron, have affected the determination of a reliable Sm-Nd isochron age. For this reason (i.e. crustal contamination in mafic samples), mafic samples from the Etheridge Group analysed by Black & MuCulloch (1984) were not used in constructing the isochron in Figure 6.12. The isochron produced by the ten Etheridge Group samples corresponds to a poorly constrained age of 1691 ± 370 Ma (MSWD = 16 based on the Model 3 solution of *Isoplot* (Ludwig, 2003) and an initial ¹⁴³Nd/¹⁴⁴Nd of 0.51065 ± 0.00046 ($\varepsilon_{Nd}(T) = +4.0$).



Figure 6.12: ¹⁴⁷Sm/¹⁴⁴Nd-¹⁴³Nd/¹⁴⁴Nd whole rock isochron of the Etheridge Group mafic rocks. Model 3 solution isochron age and initial ¹⁴³Nd/¹⁴⁴Nd ratio calculated using *Isoplot* version 3.0 (Ludwig, 2003). Data point error ellipses are 2σ.

A number of factors have contributed to the large error in the age calculation; the first and possibly most significant being the relatively limited spread in Sm/Nd across the sample range. Second, data points in Figure 6.12 that lie furthest from the Sm-Nd isochron correspond to those samples considered to have been significantly modified by crustal contamination prior to crystallisation. Unless all analysed samples contain crustal contaminants from a similar source, it would be expected that the Sm/Nd values for those contaminated samples would shift from the Sm-Nd isochron, compared to samples that were produced by simple fractionation from the depleted mantle-derived, unmodified parental magma. Nevertheless, despite the large error associated with the Sm-Nd isochron, the calculated age approximates the inferred crystallisation age of the DHM.

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Nd depleted mantle model ages (T_{DM}) were calculated for the Etheridge Group mafic rocks using the equation of Faure (1986), a depleted mantle reservoir with a presentday ¹⁴⁷Sm/¹⁴⁴Nd = 0.2137, and ¹⁴³Nd/¹⁴⁴Nd = 0.51315 (Goldstein *et al.*, 1984). The Nd model ages give an estimate of the average crustal residence time of the source material of a particular sample, and hence the time at which the sample is believed to have separated from its mantle source (either depleted mantle or CHUR). Knowledge of this reference reservoir is essential in order to calculate the correct model age, as is the assumption that Sm/Nd has not been modified following extraction from the source. Subsequently, the Nd model age can also give an indication as to the probable time of crust formation. As the Etheridge Group mafic samples all have $\varepsilon_{Nd}(T) > 0$, a depleted mantle source for the mafic rocks is inferred, and as such all model ages were calculated in reference to the depleted mantle.

The T_{DM} ages for each sample are listed in Table 3. T_{DM} ages range from 3.39 to 1.85 Ga for all samples, however the majority of ages lie between 2.17 and 2.55 Ga (average = 2.36 Ga). The majority of T_{DM} ages from this study cover a similar range to those calculated from CMD and other Etheridge Group rocks by Black & McCulloch (1984) and Knutson & Sun (1997) (Fig. 6.13; Table 6.3), and mostly likely represent the time at which the mantle source last underwent melt extraction that imposed its depleted characteristics.



Figure 6.13: T_{DM} model ages versus ¹⁴⁷Sm/¹⁴⁴Nd for Etheridge Group mafic rocks. Additional samples (grey triangles) from Black and McCulloch (1984) and Knutson and Sun (1997). T_{DM} ages and ¹⁴⁷Sm/¹⁴⁴Nd values listed in Table 6.3.

In the T_{DM} versus 147 Sm/ 144 Nd diagram (Fig. 6.13), younger T_{DM} ages correspond to a general decrease in ¹⁴⁷Sm/¹⁴⁴Nd, possibly relating to the contamination of the Sm-Nd system by small amounts of crustal material. Those samples that may have involved some contamination by crustal material also deviate slightly from the evolution curve. The evolution of the Nd isotopic composition in the mafic samples is illustrated in Figure 6.14. The evolution lines for the majority of samples cross near $\varepsilon_{Nd}(T) = +4$ and ~1.7 Ga, again indicating that the samples are geochemically related and share a near isochronous relationship. Samples that have been contaminated by crustal material would show a decrease in their ε_{Nd} values over time, relative to the uncontaminated mafic rocks that were simply derived from a depleted mantle source. Sample 299 is an example of this, as it contains U-Pb dated zircons that have returned a weighted mean age of 1687 \pm 38 Ma. However, the $\varepsilon_{Nd}(T)$ value and T_{DM} age for this sample produce a Nd evolution line that lies below the cross-over point for the Etheridge Group mafic rocks at ca. 1.7 Ga (Fig. 6.14). Contamination by material with a lower 147 Sm/ 144 Nd and hence a lower ε_{Nd} than uncontaminated Etheridge Group mafic rocks, such as Archaean crustal rocks or the metasedimentary rocks of the lower Etheridge Group, could account for this difference.



Figure 6.14: Nd isotope evolution diagram for Etheridge Group mafic rocks. Model ages indicate the source was early Palaeoproterozoic. Evolution lines for uncontaminated samples cross at ~1.7 Ga, indicating an isochronous relationship. Depleted mantle curve (DM) assumes a linear evolution of the depleted mantle from CHUR ($\varepsilon_{Nd} = 0$) at 4.56 Ga to a present-day value of +10 (Goldstein *et al.*, 1984; Jahn & Condie, 1995; Blewett *et al.*, 1998). T_{DM} ages and $\varepsilon_{Nd}(T)$ values

listed in Table 6.3.

The distribution of $\varepsilon_{Nd}(T)$ values is partially reflected in the ¹⁴⁷Sm/¹⁴⁴Nd values for each sample, with a relative decrease in ¹⁴⁷Sm/¹⁴⁴Nd corresponding to a decrease in $\varepsilon_{Nd}(T)$ (Fig. 6.15). This relationship is consistent with limited crustal contamination by a typical upper crustal component with ¹⁴⁷Sm/¹⁴⁴Nd <~1.50, and a $\varepsilon_{Nd}(T)$ value below +2.5 (Fig. 6.16). Importantly, Sample 6, which has the lowest $\varepsilon_{Nd}(T)$ and ¹⁴⁷Sm/¹⁴⁴Nd values, is a felsic differentiate from a CMD sill at Ortona. Sample 6 has the highest La/Sm of analysed CMD (2.60; Table 6.3), indicating that crustal contamination was a possible cause of the relatively lower ¹⁴⁷Sm/¹⁴⁴Nd, as addition of crustal material to the host magma generally leads to higher La/Sm values (Marsh, 2004; Lightfoot & Keays, 2005). The effect of this contamination is illustrated in Figure 6.16, where the representative Etheridge Group mafic rocks follow a simple mixing trend towards increasing degrees of contamination by the host metasediments of the lower Etheridge Group.



Figure 6.15: ¹⁴⁷Sm/¹⁴⁴Nd versus $\varepsilon_{Nd}(T)$ for mafic rocks of the lower Etheridge Group. Symbols as in Figure 6.13.



Figure 6.16: La/Sm versus ¹⁴⁷Sm/¹⁴⁴Nd diagram showing simple mixing curve for the Etheridge Group mafic rocks. The mixing curve was calculated using the following parameters for the uncontaminated Etheridge Group mafic rocks (La/Sm = 1.33, ¹⁴⁷Sm/¹⁴⁴Nd = 0.2042, La = 4.67 ppm, Sm = 3.52 ppm, and Nd = 10.43 ppm), and for the lower Etheridge Group metasediments (La/Sm = 7.23, ¹⁴⁷Sm/¹⁴⁴Nd = 0.1036, La = 67.95 ppm, Sm = 9.29 ppm, and Nd = 54.22 ppm).
Etheridge Group metasedimentary data from Knutson and Sun (1997). La/Sm and ¹⁴⁷Sm/¹⁴⁴Nd data from Table 6.3.

6.5.6 Fractionation of the depleted mantle source of the Etheridge Group mafic rocks modelled using $\varepsilon_{Nd}(T)$ values

The fractionation factor $f^{\text{Sm/Nd}}$ is a measure of the ¹⁴⁷Sm/¹⁴⁴Nd ratio of a given reservoir relative to CHUR, and is therefore the ¹⁴⁷Sm/¹⁴⁴Nd equivalent to the ε_{Nd} parameter (Rollinson, 1993). Consequently, $f^{\text{Sm/Nd}}$ can be used to estimate the level of enrichment or depletion of a given sample relative to CHUR, as well as estimate the fractionation history of the source. The $f^{\text{Sm/Nd}}$ value of an individual sample is calculated by the equation of DePaolo (1988):

$$f^{Sm/Nd} = \frac{\binom{147}{Sm} Sm^{144}Nd}{\binom{147}{Sm} Sm^{144}Nd} - 1$$
(6.1)

The $f^{\text{Sm/Nd}}$ values for individual Etheridge Group mafic samples are listed in Table 6.3. The calculated $\varepsilon_{\text{Nd}}(T)$ and $f^{\text{Sm/Nd}}$ values for the Etheridge Group mafic rocks can be used to estimate the ¹⁴⁷Sm/¹⁴⁴Nd value of the depleted mantle source, and hence the amount of depletion of the source relative to CHUR at the time of crystallisation. Using the average value for $\varepsilon_{Nd}(T)$ of +4.0, derived from the Sm-Nd isochron of Figure 6.12, the $f^{\text{Sm/Nd}}$ value of the source can be estimated using the equation of DePaolo (1988):

$$f_{Sm/Nd}^{source} = \frac{\varepsilon_{Nd}(T_X)}{Q(T_S - T_X)}$$
(6.2)

where $Q = 25.09 \text{ Ga}^{-1}$ (Rollinson, 1993), $T_S = \text{model}$ age of the magma source = 4.5 Ga (Goldstein *et al.*, 1984), and $T_X = \text{crystallisation}$ age of the Etheridge Group mafic rocks = 1.67 Ga. Based on these values, the $f^{\text{Sm/Nd}}$ value of the source was calculated to be +0.06, which equates to an initial ${}^{147}\text{Sm}/{}^{144}\text{Nd}$ for the source of 0.2085, calculated using the following equation (DePaolo, 1988):

$$f_{source}^{Sm/Nd} = \frac{\left({}^{147}Sm/{}^{144}Nd\right)_{source} - \left({}^{147}Sm/{}^{144}Nd\right)_{CHUR}}{\left({}^{147}Sm/{}^{144}Nd\right)_{CHUR}}$$
(6.3)

rearranged to solve for ¹⁴⁷Sm/¹⁴⁴Nd:

$$\left(\frac{{}^{147}Sm}{{}^{144}Nd}\right)_{source} = \left[f_{source}^{Sm/Nd} \times \left(\frac{{}^{147}Sm}{{}^{144}Nd}\right)_{CHUR}\right] + \left(\frac{{}^{147}Sm}{{}^{144}Nd}\right)_{CHUR}$$
(6.4)

The calculated $\int^{\text{Sm/Nd}}$ value indicates that the depleted mantle source of the Etheridge Group mafic rocks had a Sm/Nd ratio that was ~6 % higher than CHUR at the time of crystallisation of the mafic rocks.

The $f^{\text{Sm/Nd}}$ values obtained for both the source and sampled rocks can then be used to estimate the amount of depletion in the Sm/Nd value from the crystallised rocks relative to their depleted mantle source. The depletion factor ($\alpha_{\text{Sm/Nd}}$) is calculated from the following equation (DePaolo, 1988):

$$\alpha_{Sm/Nd} = \frac{1 + f_{sample}^{Sm/Nd}}{1 + f_{sample}^{Sm/Nd}}$$
(6.5)

Using the values of $f^{\text{Sm/Nd}} = +0.06$ for the depleted mantle source, and $f^{\text{Sm/Nd}} = -0.06$ for the Etheridge Group mafic samples (average of the 10 samples, Table 3), a value of 0.890 was calculated for $\alpha_{\text{Sm/Nd}}$. This indicates that the Sm/Nd value of the mafic rocks decreased by ~11 % from that of the depleted mantle source, which was originally depleted in Sm/Nd by ~6 % relative to CHUR (Fig. 6.17). A relatively low degree of partial melting coupled with significant fractional crystallisation of a tholeiitic parental magma could account for the observed depletion (DePaolo, 1988). This evidence, in addition to the calculated T_{DM} and crystallisation ages, indicates that the Etheridge Group mafic rocks had a relatively short residence time in the upper crust during the Palaeoproterozoic, but did assimilate some crustal material from host metasediments during emplacement.



Figure 6.17: ¹⁴⁷Sm/¹⁴⁴Nd versus $\varepsilon_{Nd}(T)$ diagram showing the effects of Sm/Nd fractionation (α) associated with partial melting for the Etheridge Group mafic rocks (after DePaolo, 1988). For explanation of $\alpha_{Sm/Nd}$, $\varepsilon_{Nd}(T)$, and ¹⁴⁷Sm/¹⁴⁴Nd values of source see text and Figure 6.12. $\alpha_{Sm/Nd}$ values calculated for a present day depleted mantle ¹⁴⁷Sm/¹⁴⁴Nd ratio of 0.2317 (DePaolo, 1988).

Shaded areas represent tholeiitic MORB and tholeiitic basalt fields of DePaolo (1988).

6.6 Similarities to modern continental and oceanic tholeiitic flood basalt provinces: implications for the tectonic setting of the Etheridge Group mafic rocks

The tectonic setting for the eruption and emplacement of the Etheridge Group mafic rocks is reasonably well constrained to a continental-rift setting leading to break-up. This is based partially on regional geological and lithological evidence, and also on the geochemical evidence presented above and by previous authors (e.g. Withnall, 1985). Despite this well constrained setting, at present little work has been done in order to determine the processes that affected the evolution of the magma. To address this issue, in this section the mafic rocks of the Etheridge Group are compared to a number of major Phanerozoic tholeiitic flood basalt provinces (Table 6.4). A number of these provinces (e.g. Karoo, East Greenland) have been associated with Phanerozoic continental rifting, and as such, may provide a close modern analogue to the processes that affected the genesis of the Etheridge Group mafic rocks.

The relatively primitive, Fe-enriched nature of the Etheridge Group tholeiites, coupled with the apparent lack of contamination by subduction-related material, precludes the possibility that the tholeiites formed in a back-arc basin setting proximal to a volcanic arc. For example, contemporaneous Mg-rich and Fe-rich metatholeiites found in the Olary Domain, western Curnamona Province display geochemical characteristics typical of back-arc basin basalts. These include relatively elevated LREE/HREE, Nb/Y, Zr/Y, and more evolved Nd-isotope values ($\varepsilon_{Nd} < +2$) compared to more primitive tholeiites (Rutherford *et al.*, 2006; Chapter 7). It has been proposed that the geochemically distinct character of the Olary Domain metatholeiites is a consequence of refertilisation of the lithospheric mantle by a subduction-related component (Rutherford *et al.*, 2006); a feature common to tholeiites formed in back-arc basins (Ishizuka *et al.*, 2006; Haraguchi & Ishii, 2007).

In contrast, otherwise geochemically-related Fe-rich tholeiites of the Broken Hill Domain (Rutherford *et al.*, 2006), and contemporaneous Etheridge Group tholeiites are relatively primitive, and do not show the same level of LREE or isotopic enrichment. The geochemical characteristics and relationship between the mafic suites is expanded upon in Chapter 7. Although the geochemical characteristics of the Etheridge Group tholeiites precludes formation in a back-arc basin setting, a tectonic model involving a more distal back-arc setting (Giles *et al.*, 2002), and thus not

involving fluids from a subducting slab effecting the geochemistry of the mafic rocks, may form a realistic interpretation for the development of the mafic rocks. This model is discussed in more detail in Chapter 8.

In general, the mafic rocks of the Etheridge Group appear relatively Fe-rich when compared to modern examples of rift-related tholeiites (Table 6.4). Although possessing lower Nb/Zr values than N-MORB (Galapagos suite in Fig. 6.18), the range of Nb/Zr values for the Etheridge Group rocks and their trend of increasing Ti values with advancing fractionation are typical of basalts from volcanic rifted passive margins (e.g. E Greenland) and of tholeiitic basalts erupted during plume-head related Large Igneous Province eruptions such as those basalts forming the Siberian Traps and Ontong Java Plateau (Fig. 6.18). Significantly, the basalts of all of these three terranes were formed by processes that involved the presence of a mantle plume source (Neal *et al.*, 1997; Saunders *et al.*, 1997; Sharma, 1997).

In the Zr/Nb versus Zr/Y diagram (Fig. 6.19) the Etheridge Group mafic rocks plot predominately within the field of continental within-plate basalts erupted in Large Igneous Provinces, for example, central Karoo (Marsh *et al.*, 1997) and the Ontong Java Plateau. In this diagram, values for continental within-plate basalts erupted in rift settings, such as for the central Karoo basalts, plot between the values for average N-and E-MORB, and above average Primitive Mantle.

An intra-cratonic LIP or volcanic passive margin setting for the emplacement of the Etheridge Group tholeiites is further supported by the close correlation in their incompatible element ratios to tholeiitic rocks emplaced both onshore and offshore during plume-triggered rifting of East Greenland-Faeroe Islands margin in the North Atlantic Igneous Province (Fig. 6.20; Saunders *et al.*, 1997). The onshore and offshore lavas of East Greenland show a clear transition from continental lithosphere-influenced to MORB-type oceanic magmatism, illustrated by changes in the incompatible element compositions across the region (Fig. 6.21; Fitton *et al.*, 1995, 2000). Older pre- to syn-rift lavas are represented by onshore lava suites from both East Greenland and the Faeroe Islands (Fig. 6.20a), and additionally by ODP drillsites which preferentially sampled lower series lavas from the seaward-dipping reflector sequences (SDRS) along the East Greenland continental margin (Sites 989 and lower portion of 917; Fig. 6.20c). The rift-margin lavas show enriched compositions (Zr/Nb < 20), consistent with evolution involving fractional crystallisation and assimilation of

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crustal material in deep crustal magma reservoirs beneath the continental lithosphere (Fitton *et al.*, 2000). Younger lavas of the middle and upper SDRS (Sites 990 and upper portion of 917) represent later stages of rifting leading to continental break-up and eruption of oceanic crust, and preferentially have compositions more consistent with N-type MORB (Zr/Nb > 20; Fig. 6.21a).

No. of analyses	Karoo tholeiite 24	East Greenland tholeiite 22	Galapagos tholeiitic MORB 75	Oceanic tholeiite 74	Continental tholeiite 44
SiO ₂	50.50	48.20	49.69	49.26	50.18
TiO ₂	1.07	2.84	2.34	1.34	1.35
Al_2O_3	14.05	13.21	14.87	14.53	16.06
Fe ₂ O ₃ *	11.89	13.88	13.57	12.54	12.08
MnO	0.17	0.19	0.16	0.20	0.18
MgO	7.46	8.00	6.82	6.95	6.83
CaO	10.07	10.55	8.34	12.08	9.96
Na ₂ O	2.30	2.36	2.45	2.28	2.29
K ₂ O	0.67	1.72	0.20	0.33	0.76
P_2O_5	0.12	0.28	0.37	0.25	0.20
LOI	-	1.50	-	0.38	2.33
Nb	14	17	13	5	11
Zr	109	210	305	79	134
Sr	205	292	62	148	266
Cr	349	318	102	178	192
Ba	235	88	73	18	-
Sc	35	30	44	49	32
V	295	373	358	332	228
La	-	15	11	5	15
Ce	-	36	24	12	35
Nd	-	23	23	9	19
Y	21	30	96	27	25
Rb	30	5	2	10	15
Pb	-	2	14	-	-
Zn	-	110	229	-	106
Cu	-	198	1592	-	89
Ni	95	194	43	98	106
K (ppm)	5527	14309	1675	2706	6268
Ti (ppm)	6435	17040	14040	8028	8114
P (ppm)	538	1234	1625	1085	861

Table 6.4: Average geochemical compositions for various unaltered tholeiitic suites. Karoo analyses from Marsh *et al.* (1997); East Greenland analyses from Peate *et al.* (2003); Galapagos MORB analyses from Schilling *et al.* (1976), Fisk *et al.* (1982), Embley *et al.* (1988), Ridley *et al.* (1994), and Perfit *et al.* (1999); oceanic tholeiite (Ontong Java Plateau) analyses from Mahoney *et al.* (1993); continental tholeiite (Siberian Traps) analyses from Lightfoot *et al.* (1990).



Figure 6.18: Ti (ppm) versus Nb/Zr for Etheridge Group mafic rocks and selected Phanerozoic flood basalt provinces (data sources as for Table 6.4). CMD = pink squares; DHM = blue diamonds. PM = Primitive Mantle; K = Karoo Igneous Province; SB = Siberian Traps; EG = East Greenland; G = Galapagos spreading centre; OJ = Ontong Java plateau. Average N-MORB, E-MORB, and Primitive Mantle values after Sun and McDonough (1989). Fractional crystallisation on this diagram will produce horizontal vectors.

In Figure 6.21, incompatible trace element ratios from the suite of Etheridge Group tholeiites show a clear association with the older, onshore and lower series lavas from the East Greenland margin and Faeroe Islands (Zr/Nb = 10-30). This suggests that the differentiation processes which influenced the evolution of the Etheridge Group tholeiites were similar to those of East Greenland, and that emplacement/eruption of the mafic rocks occurred in a similar tectonic setting, in a developing rift in attenuating continental crust along a volcanic passive margin.



Figure 6.19: Zr/Nb versus Zr/Y for Etheridge Group mafic rocks and selected Phanerozoic flood basalt provinces (data sources as for Table 6.4). OIB = Oceanic Island Basalt. Average OIB value after Sun and McDonough (1989). All other symbols and fields as in Figure 6.18.



Figure 6.20: (A) Location of East Greenland and Faeroe Islands lavas erupted during Palaeogene plume-triggered rifting in the North Atlantic. (B) Location of ODP drillsites along Southeast Greenland margin (after Larsen *et al.*, 1999). (C) Section across the Southeast Greenland margin



at 63°N showing location of ODP drillsites (after Larsen *et al.*, 1999). Analyses from underlined sites in (C) represent East Greenland offshore lavas.

Figure 6.21: Zr/Nb versus Ti/Zr for mafic rocks of the Etheridge Group, and comparison to (a)
East Greenland offshore lavas, and (b) East Greenland and Faeroe Island onshore lavas. Blue arrows in (a) denote progressive fractionation from more primitive to more evolved compositions. Data for East Greenland offshore lavas from Larsen *et al.* (1998), Larsen *et al.* (1999), Fitton *et al.* (2000), and Philipp *et al.* (2001). Data for East Greenland onshore lavas from Noe-Nygaard and Pedersen (1974), Brooks *et al.* (1976), Thirlwall *et al.* (1994), Hansen and Nielsen (1999), and Momme *et al.* (2002). Data for Faeroe Islands lavas from Waagstein *et al.* (1984), and Hald and Waagstein (1991).

6.7 Conclusions

The extrusive (DHM) and intrusive (CMD) mafic igneous rocks of the lower Etheridge Group share a common geochemical history, were likely derived from the same parental magmas, and possess geochemical characteristics typical of relatively evolved, low-K, Fe-rich continental tholeiites and tholeiites during strong extension of continental crust leading to continental break-up. The mafic rocks were most likely emplaced over a relatively short timeframe, from melts derived directly from a depleted convecting asthenospheric mantle source, and limited assimilation of upper crustal material from the host lower Etheridge Group metasediments. Trace and REE data indicate that the rocks were generated by partial melting of a mantle source similar to that which provides modern MORB, and evolved along a tholeiitic trend of Fe-enrichment that produced rare evolved samples with >20 wt% Fe₂O₃^{total}. A poorly constrained Sm-Nd isochron age, coupled with $\varepsilon_{Nd}(T)$ and T_{DM} values indicate that the mafic rocks crystallised at ~1.7 Ga, which correlates with crystallisation ages determined from U-Pb igneous zircons analysed by LA-ICPMS. T_{DM} model ages also indicate that the source of the mafic rocks may have formed during the early Palaeoproterozoic (ca. 2.36 Ga).

Despite their relatively widespread spatial distribution, the Etheridge Group mafic rocks show only limited geochemical variation. Little, if any, geochemical variation is observed between CMD sills emplaced below, within, or above the level of the DHM. Individual sills of the CMD (<1000 m thick) show little variation in MgO, indicating that individual sills were emplaced relatively rapidly.

In several samples of the mafic rocks, REE values show distinct LREE-enriched HREE-depleted chondrite-normalised patterns and LILE enrichment relatively to similarly incompatible Nb, consistent with small degrees of contamination by the host metasediments (Fig. 6.7; Fig. 6.9). The presence of this contamination is also illustrated by ¹⁴⁷Sm/¹⁴⁴Nd isotopic values which show a trend towards increasing degrees of contamination by an upper crustal component (Fig. 6.16).

A continental rifting environment at a volcanic passive margin provides the appropriate conditions for the emplacement of the Etheridge Group mafic rocks. During opening of the rift system that contained the lower Etheridge Group metasediments, the underlying depleted mantle melted due to decompression and tholeiitic magma was introduced into the upper crust as sills and lava-sill packages.

Similar relatively Fe-rich tholeiitic basalts of similar age to the lower Etheridge Group mafic rocks are recorded from the Eastern Succession of the Mt Isa Inlier, and the from the Broken Hill Inlier. The following chapter investigates further these correlations and their implications for the geological evolution of eastern Australia in the Palaeoproterozoic.

Chapter 7.

Geochemical comparison of Fe-rich tholeiites from the lower Etheridge Group, Broken Hill Group and Mt Isa Eastern Succession: their genesis and implications for the eastern margin of Palaeoproterozoic Australia

7.1 Introduction

The Palaeoproterozoic basins of northeastern and eastern Australia host large amounts of variably deformed and metamorphosed Fe-rich tholeiitic rocks. Their origin has been the focus of considerable debate in recent years (e.g. Phillips *et al.*, 1985; Withnall, 1985; Etheridge et al., 1987; Williams, 1998b; Rutherford et al., 2006); however at present the most accepted hypothesis is that the rocks were derived from partial melts of widespread asthenospheric convection accreting material to the base of Archaean and earliest Palaeoproterozoic lithosphere (Scott et al., 2000; Giles et al., 2002). Emplacement and/or extrusion of the mafic rocks occurred during one or more episodes of crustal thinning associated with the second cycle of Palaeoproterozoic regional extension and sedimentation across northeastern Australia (ca. 1730-1650 Ma; Etheridge et al., 1987; Betts et al., 2002; Giles et al., 2002). The widespread occurrence of the high-Fe tholeiites, coupled with complementary geochronological, sedimentary, and geophysical evidence, has led several authors to imply a shared evolution for the terranes of Palaeoproterozoic northeastern Australia (e.g. Black et al., 1998; Williams, 1998b; Giles et al., 2002; Betts & Giles, 2006; MacCready et al., 2006). However to date, none of these studies have provided a comparative geochemical study of the Fe-rich tholeiitic rocks found in each terrane.

Both the Broken Hill Block and Mt Isa Eastern Succession host world-class basemetal ore deposits. Therefore, the geochemical comparison of Fe-rich tholeiites from northeastern Australia is not only of regional significance, but may also be of significant economic importance for the Georgetown Inlier, as mineralisation is often spatially associated with areas containing significant tholeiitic magmatism. Subsequently, by using the new geochemical data from the mafic rocks of the Georgetown Inlier presented in Chapter 6, along with previously published data from mafic rocks of the Mt Isa Eastern Succession and Broken Hill Block, I provide a detailed geochemical overview of the similarities and differences between the

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Georgetown, Mt Isa, and Broken Hill Fe-rich tholeiites. The implications for basemetal mineralisation in the Georgetown Inlier will also be discussed.

The first section of this chapter reviews the Palaeoproterozoic Broken Hill and Mt Isa Eastern Succession volcano-sedimentary basins, providing detailed summaries of the Fe-rich tholeiitic rocks found in each terrane. This is followed by a geochemical comparison of the ca. 1690-1660 Ma Fe-rich tholeiites that occur in the Georgetown Inlier, Mt Isa Eastern Succession, and Curnamona Province. The results of this study are then combined with similar work that has focussed on comparing the geochronology and lithostratigraphy of the Proterozoic terranes, with the aim of proposing a regional tectonic model for the formation and evolution of the Proterozoic sedimentary basins of eastern Australia.

7.2 Objectives

The objectives of the comparative geochemical study of the Fe-rich tholeiitic rocks from the Georgetown Inlier, Mt Isa Eastern Succession, and Broken Hill Block are:

- To compare the geochemistry of the mafic igneous rocks; and evaluate possible implications for regional constructions of the Proterozoic margin of eastern Australia, as well as the Palaeoproterozoic position of the Georgetown Inlier.
- To further constrain the petrogenesis and tectonic setting of the Fe-rich tholeiites from the Georgetown Inlier, as described in Chapter 6.
- To determine whether the Proterozoic rocks of the Georgetown Inlier were also a suitable host for major base-metal mineralisation based on their relationships with the Mt Isa Eastern Succession and Broken Hill Block.

7.3 Regional Geology of the Broken Hill Block and Mt Isa Eastern Succession: a review

The Broken Hill Block of western New South Wales, and the Mt Isa Eastern Succession of western Queensland, form disjunct parts of regionally extensive Proterozoic volcano-sedimentary sequences in eastern Australia (Fig. 7.1). Both sequences share several key features, including the occurrence of ca. 1690-1660 Ma Fe-rich tholeiites (Phillips *et al.*, 1985; James *et al.*, 1987; Williams, 1998b), and the

presence of major 'Broken Hill-type' Pb-Zn-Ag deposits associated with zones of high metamorphic grade (Walters, 1998; Williams, 1998a). To aid the geochemical study, a review of the regional geology of the Broken Hill Block and Mt Isa Eastern Succession is presented below. This is followed by a detailed description of the location and petrology of significant Fe-rich mafic-igneous units from each terrane.



Figure 7.1: Palaeoproterozoic terranes of eastern Australia. Palaeoproterozoic location of the Curnamona Province after Giles *et al.* (2004). Figure adapted after Myers *et al.* (1996) and Betts & Giles (2006).

7.3.1 Broken Hill Block

The Broken Hill Block of the Willyama Inlier in western New South Wales represents the easternmost part of the regionally extensive Curnamona Province (Fig. 7.1; Willis *et al.*, 1983), and has been the focus of considerable research for the past 100 years (e.g. Mawson, 1912; Vernon, 1969; Willis *et al.*, 1983), due in large part to the world-class stratiform Pb-Zn-Ag orebodies at Broken Hill. The Broken Hill Block is

composed of metamorphosed and multiply-deformed volcano-sedimentary rocks intruded by bimodal igneous volcanics. This package is thought to be approximately 6-7 km in thickness (Fig. 7.2; Willis *et al.*, 1983). The sequence is unconformably overlain by Neoproterozoic sediments of the Adelaide Fold Belt, and is thought to have been deposited onto Archaean continental crust (James *et al.*, 1987).



Figure 7.2: Simplified time-space diagram showing tectonostratigraphic evolution of eastern Australian Proterozoic terranes between 1800 and 1600 Ma. BAS: Basso Suite; LLS: Lady Louise Suite; RG: Redan Gneiss; TCG: Thorndale Composite Gneiss; TG: Thackaringa Group; BHG: Broken Hill Group; RRG: Rasp Ridge Gneiss; BV: Benagerie Volcanics; MMG: Mundi Mundi Granite; MGQ: Mount Guide Quartzite; ECV: Eastern Creek Volcanics; MSG: Myally Subgroup; FCV: Fiery Creek Volcanics; SCF: Surprise Creek Formation; SYB: Sybella Granite;

MLMB: Magna Lynn Metabasalt; LCF: Llewellyn Creek Formation; MNQ: Mount Norna Quartzite; TCV: Toole Creek Volcanics; TCB: Tommy Creek Beds; RRSG: Robertson River

Subgroup; CMD: Cobbold Metadolerite; DHM: Dead Horse Metabasalt; FB: Forsayth Batholith; FHS: Forest Home Suite. Data summarised from Withnall *et al.* (1988a), Page *et al.* (2005b), Betts *et al.* (2006), Betts & Giles (2006), and Rutherford *et al.* (2006).

The lithostratigraphy of the Broken Hill Block has been defined by Willis et al. (1983), and revised by Raetz et al. (2002) (Fig. 7.2). The lowermost sequence comprises paragneiss (Redan and Farmcote Gneiss) belonging to the Redan Geophysical Zone (Raetz et al., 2002). The Redan Geophysical Zone is overlain by migmatites (Clevedale Migmatite) and feldspathic metasedimentary composite gneiss (Thorndale Composite Gneiss) that are thought to represent interlayered immature siliciclastic sediment and felsic volcanics deposited in a shallow shelf environment (Willis et al., 1983). The overlying Thackaringa Group is the main outcropping lithology in the Broken Hill region and is characterised by quartzo-feldspathic granite gneiss and albite-quartz rocks (Himalaya Formation), psammo-pelitic rocks (Cues Formation), and lesser amounts of amphibolites and basic gneiss (Willis et al., 1983; Raetz et al., 2002). The basal sequence of the overlying Broken Hill Group comprises turbiditic, thinly-bedded metasediments and mafic gneiss (Allendale Formation), overlain by mafic and felsic gneiss (Parnell Formation), pelitic metasediments (Freyers Metasediments), and mafic and felsic gneisses and pelitic metasediments (Silver King Formation) (Willis et al., 1983; Page & Laing, 1992). The uppermost member of the group is the Hores Gneiss, comprising felsic gneiss ('Potosi gneiss'; Page & Laing, 1992), mafic gneiss and rare pelite, which hosts to the major Pb-Zn-Ag orebody at Broken Hill (Willis et al., 1983). The Thackaringa and Broken Hill Groups are thought to represent a period of deepening shelf sedimentation proximal to a propagating continental-rift environment, during which there was widespread bimodal magmatism, with most of the mafic granulites being regarded as sills (Willis et al., 1983; James et al., 1987). No stratabound magmatic rocks of either mafic or felsic nature are identified above the level of the Broken Hill Group (Fig. 7.2; James et al., 1987). Pelitic and psammo-pelitic metasediments are the primary lithologies in the overlying Sundown and Paragon Groups, and were thought to have been deposited in a continental slope to deep water environment (Willis et al., 1983).

Numerous studies involving zircon U-Pb SHRIMP analyses have been published over the past two decades in order to constrain an approximate depositional age for the Broken Hill Group (e.g. Stevens, 1986; Page & Laing, 1992; Donaghy *et al.*, 1998; Nutman & Ehlers, 1998; Page *et al.*, 2000; Raetz *et al.*, 2002; Page *et al.*, 2005a). At present, near-consensus exists that deposition had commenced by ca. 1690 Ma (e.g. Page & Laing, 1992; Raetz *et al.*, 2002; Page *et al.*, 2005a), although Nutman & Ehlers (1998) argued for deposition between ca. 1720-1690 Ma. Page & Laing (1992) analysed samples of the Hores Gneiss at a number of locations across the terrane, and were able to define a major zircon population at 1690 ± 5 Ma which they ascribed to deposition of the volcano-sedimentary precursor to the gneiss.

The rocks of the Broken Hill Group are believed to have undergone at least three periods of deformation and metamorphism in the early Proterozoic (James et al., 1987; Raetz et al., 2002; Gibson & Nutman, 2004). The first two deformation events involved extension and subsequent folding, with the latter coinciding with peak lowpressure and high-temperature metamorphic conditions in the upper amphibolite to granulite facies (James et al., 1987; Page & Laing, 1992; Gibson & Nutman, 2004). The third deformation event also produced folding, but resulted in the formation of retrograde assemblages where the intensity of this event was greatest (Gibson & Nutman, 2004). As with the depositional age of the Broken Hill Group, there is also debate as to the age (and relevance) of the tectono-metamorphic events that affected the terrane during the Proterozoic. Page & Laing (1992) obtained a metamorphic age of 1600 ± 8 Ma which they ascribed to M₂, determined from the analysis of metamorphic overgrowths surrounding the magmatic cores of zircons from the granulite facies rocks surrounding Broken Hill. They also inferred that M₁ preceded this event by a short time interval, and that both events were compressional. This age also corresponds to the early stages of the inter-cratonic Olarian Orogeny (Raetz et al., 2002). Based on the U-Pb zircon analysis of three amphibolites and one granulite from the Clevedale Migmatite and Thorndale Composite Gneiss, Gibson & Nutman (2004) suggested the first tectono-metamorphic event occurred between ca. 1690 Ma and 1670 Ma, some 70 million years earlier than the M_1 age suggested by Page & Laing (1992). However, it is possible that the zircons analysed by Gibson & Nutman (2004) were magmatic rather than metamorphic in origin, and consequently date the crystallisation age of the rocks. Evidence for the third deformation event is best preserved in retrograde amphibolite facies assemblages in areas of high D₃ strain (Gibson & Nutman, 2004). ⁴⁰Ar/³⁹Ar hornblende and Pb-Pb apatite ages indicate

metamorphism associated with this event occurred between 1590-1560 Ma (Harrison & McDougall, 1981; Gulson, 1984).

7.3.1.1 Metamorphosed Fe-rich tholeiites of the Broken Hill Group

Mafic or 'basic' gneisses in the Broken Hill Block are confined to the lower half of the sequence, below the Sundown Group (Fig. 7.2; James et al., 1987; Raetz et al., 2002). They commonly occur as tabular or lenticular bodies 0.5-200 m in thickness and 1-2000 m in length, but are generally up to 1000 m long and up to 40 m wide within the Broken Hill Group (Phillips et al., 1985). They are mostly conformable to bedding, and are generally considered to be sills, although discordant examples also exist (James et al., 1987). High-Fe (>10 wt% Fe₂O₃^{total}) tholeiitic basic gneisses are common in the Broken Hill and Thackaringa Groups, with the most Fe-rich examples occurring proximal to areas of mineralisation (Phillips et al., 1985), including along the main Broken Hill line of lode. The exceptionally high-Fe contents from these areas (up to 25 wt% $Fe_2O_3^{total}$) were thought to be a product of hydrothermal alteration that occurred synchronous with mineralisation and prior to regional deformation in the Broken Hill Block (Phillips et al., 1985). However, a more detailed geochemical study by James *et al.* (1987) suggested that even the most extreme Fe-rich rocks may have evolved by a process of fractional crystallisation along a strong Fe-enrichment trend.

The basic gneisses, including the Fe-rich tholeiites, are commonly composed of plagioclase and hornblende with or without quartz, epidote, garnet, ilmenite and clinopyroxene, and orthopyroxene is a common phase in the granulites. In the Broken Hill area, their mineralogy is indicative of amphibolite to low-granulite facies metamorphism (James *et al.*, 1987). Due to the high level of recrystallization, most primary igneous features have been generally obliterated; however, Stroud *et al.* (1983) and James *et al.* (1987) indicated that some igneous textures and contacts may have been preserved, particularly in the lower grade northwest part of the Broken Hill Block.

7.3.2 Mt Isa Eastern Succession

The Mt Isa Eastern Succession represents the easternmost part of the regionally extensive Mt Isa Inlier in western Queensland (Fig. 7.1). It is interpreted to have formed as a product of Palaeoproterozoic intracratonic rifting, in a similar tectonic environment to that proposed for the Broken Hill and Georgetown regions (O'Dea et al., 1997; Page & Sun, 1998; Williams, 1998b). The rocks of the Eastern Succession, in particular those belonging to the informally named Maronan Supergroup (Beardsmore et al., 1988), have come under increasing scrutiny over the past two decades, due to the discoveries of major base-metal deposits within the succession. These include the Broken Hill-type Pb-Zn-Ag orebody at Cannington, smaller Pb-Zn deposits at Pegmont and Fairmile, and major Cu-Au deposits such as Ernest Henry (Williams, 1998a). The succession is composed of a number of lithologically distinct units including metamorphosed siliciclastic volcano-sedimentary and mafic volcanic rocks of the Maronan Supergroup, metasedimentary and bimodal volcanic rocks of the Tommy Creek Beds, the Argylla, Corella and Doherty Formations (collectively termed the Mary Kathleen Group), and younger granitic rocks of the Williams and Naraku Batholiths (Fig. 7.2; Beardsmore et al., 1988; Page & Sun, 1998).

Using U-Pb SHRIMP zircon ages and crustal Sm-Nd isotope signatures calculated from a diverse range of lithologies, Page & Sun (1998) inferred that the Mt Isa Eastern Succession was most likely deposited as two tectonostratigraphic packages over a period of ~170 million years, between ca. 1770-1600 Ma. The older sequence, deposited between ca. 1770-1725 Ma, included sediments and bimodal volcanics belonging to the Mary Kathleen Group. The younger sequence included metasediments and bimodal volcanics of the Maronan Supergroup, and metasediments of the Tommy Creek Beds deposited between ca. 1720-1600 Ma (Page & Sun, 1998). Deposition of sediments and the emplacement of bimodal volcanics, including Fe-rich tholeiites belonging to the Soldiers Cap Group of the Maronan Supergroup, are thought to have commenced ca. 1670 Ma. This age approximates inferred or calculated ages for the deposition of other key volcano-sedimentary sequences containing base-metal mineralisation in northeastern Australia including the Broken Hill Group (ca. 1690 Ma; Page et al., 2005a), and the Mt Isa Group of the Mt Isa Western Succession (ca. 1655 Ma; Page & Sun, 1998), as well as metasedimentary rocks of the lower Etheridge Group (ca. 1700-1650 Ma) (Fig. 7.2).

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At least three major episodes of deformation are thought to have affected the Mt Isa Eastern Succession (Laing, 1998). There is general consensus that the last two events were compressional, the latter corresponding to the later stages of the Isan Orogeny between ca. 1540 and 1500 Ma (Page & Sun, 1998; Betts et al., 2000; Betts & Giles, 2006). Peak metamorphic conditions within the Mt Isa Inlier are suggested to have occurred during the earlier stage of the Isan Orogeny between ca. 1600 and 1570 Ma, which led to metamorphism ranging from amphibolite to granulite facies in the Soldiers Cap Group (Page & Bell, 1986; Page & Sun, 1998). The estimated age of the first deformation event is controversial. Oliver et al. (1991) and Oliver (1995) proposed that the first event occurred between ca. 1750 and 1730 Ma and involved extension coupled with magmatic underplating, leading to the emplacement of the Eastern Creek Volcanics in the Mt Isa Western Succession. However, a metamorphic age of 1584 ± 17 Ma obtained from overgrowths on the rims of zircons sampled from felsic gneisses that formed by partial melting at the base of the Soldiers Cap Group led Page & Sun (1998) to suggest that this age may represent metamorphism associated with the first deformation event. This age is also broadly coincident with the ca. 1600 Ma timing of high-grade metamorphism at Broken Hill (Page & Laing, 1992).

7.3.2.1 Metamorphosed mafic igneous rocks of the Mt Isa Eastern Succession

Bimodal volcanic rocks occur throughout the Mt Isa Eastern Succession, and locally dominate parts of the ca. 1670 Ma Soldiers Cap Group (Fig. 7.2; Beardsmore *et al.*, 1988; Page & Sun, 1998; Hatton & Davidson, 2004). The lithostratigraphy of the Soldiers Cap Group was originally proposed by Derrick *et al.* (1976), and comprises a basal sequence of turbiditic pelite and lesser metadolerite (Llewellyn Creek Formation) conformably overlain by massively bedded quartzite, psammo-pelite, pelite and intercalated metadolerite, and lesser metabasalt (Mt Norna Quartzite). The Mt Norna Quartzite is conformably overlain by a 2800 m-thick sequence of metadolerite and metabasalt, and minor interbedded quartzite, pelite, and ironstone termed the Toole Creek Volcanics (Derrick *et al.*, 1976; Hatton & Davidson, 2004).

Williams (1998b) defined amphibolites of the Soldiers Cap Group as high-Fe metatholeiites, as they commonly contain between 11 and 19 wt% $Fe_2O_3^{total}$. The Fe-enriched nature of these tholeiites has led a number of authors to suggest an intra-

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cratonic rift setting for the rocks (Glikson & Derrick, 1978; Williams, 1998b). The amphibolites are commonly composed of hornblende and plagioclase with lesser biotite, locally developed garnet and quartz, and accessory titanite or ilmenite (Williams, 1998b). Despite the high levels of recrystallization experienced by the mafic rocks, igneous features indicative of a syn-sedimentary to diagenetic depositional origin are preserved, including basalt pillows and peperitic contacts (Hatton & Davidson, 2004).

7.3.2.2 Other mafic igneous rocks of the Mt Isa Inlier

Mafic igneous rocks of comparable geochemistry to the Fe-rich tholeiites of the Toole Creek Volcanics occur within the Eastern Creek Volcanics and Fiery Creek Volcanics of the Mt Isa Western Succession, and the Marraba Volcanics and Magna Lyn Metabasalt of the lower Eastern Succession (Fig. 7.2; Page & Sun, 1998; Giles *et al.*, 2002). Due to their age and stratigraphic location, the mafic rocks of the Eastern Creek Volcanics are unlikely to have been associated with the ca. 1670 Ma magmatism that generated the Fe-rich tholeiites of the Soldiers Cap Group, being some 100 million years older. However, their geochemistry and close spatial relationship to the large Cu and Pb-Zn-Ag orebody at Mt Isa are noteworthy, and as such, a representative suite of samples from this unit is included in the comparative geochemistry section.

The Eastern Creek Volcanics are a suite of meta-igneous, sedimentary, and volcaniclastic rocks deposited between ca. 1780 and 1750 Ma in an intra-continental rift setting (Gulson *et al.*, 1983; Wyborn, 1987). Although their age suggests that they are unlikely to be related to post-1700 Ma tholeiitic suites from other terranes, the high-Fe Eastern Creek Volcanics are comparable in age to other volcano-sedimentary packages including the Nungbalgarri, Bartalumba and Seigal Volcanics of the Palaeoproterozoic McArthur Basin of northern Australia, and the Magna Lyn and Marraba Volcanics of eastern-central Mt Isa (Giles *et al.*, 2002).

The Magna Lyn Metabasalt and Marraba Volcanics also contain Fe-rich metabasalt and amphibolite, which may be equivalent to the Fe-rich tholeiites of the Eastern Creek Volcanics (Giles *et al.*, 2002). Both units contain basalt with >10 wt% Fe₂O₃^{total} (Ozchem, 2000; Hatton, 2004). Metabasalts from the ca. 1710 Ma Fiery Creek Volcanics (Page & Sun, 1998) also display the high-Fe values characteristic of Proterozoic tholeiitic volcanism, including some mafic rocks that are extremely Ferich (>15 wt% Fe₂O₃^{total}; Ozchem, 2000; Hatton, 2004). However, representative samples from this unit have <3 wt% MgO and were likely affected by some degree of crustal assimilation.

7.4 Geochemical comparison of Fe-rich tholeiites and basic gneisses from the Etheridge Group, Mt Isa Eastern Succession, and Broken Hill Group.

7.4.1 Data sources used for comparative geochemistry

Geochemical data from mafic rocks of the Soldiers Cap Group and other mafic units of the Mt Isa Inlier were obtained from Davidson (1994), Williams (1998b), Scott *et al.* (2000), Hatton (2004) and the Geoscience Australia Ozchem Database (Ozchem, 2000). Geochemical data for basic gneisses from the Broken Hill Group were obtained from the Geological Survey of New South Wales Broken Hill whole rock geochemical database (1998), and A.J. Crawford (unpublished data). Sample numbers where listed for samples from the Ozchem database are those provided by Ozchem. Analyses obtained from the Ozchem Hill and Mt Isa mafic rocks in Appendix 5.

7.4.2 Comparative major, trace, and rare earth element geochemistry

The Fe-rich tholeiites of the Etheridge Group possess whole-rock chemical compositions remarkably similar to contemporaneous Fe-rich mafic rocks from both the Broken Hill Group of the Willyama Inlier and Soldiers Cap Group of the Mt Isa Eastern Succession. A possible geochemical link between the mafic rocks of the Broken Hill and Soldiers Cap Groups has been postulated in some previous studies (e.g. Williams, 1998b; Scott *et al.*, 2000; Giles *et al.*, 2002). Subsequently, a significant geochemical correlation between the mafic rocks from these terranes has important implications for the location, timing of deposition, and regional tectonic setting of the Georgetown Inlier during the Palaeoproterozoic.

In order to accurately demonstrate the geochemical similarities that exist between the mafic rocks of the Broken Hill Group (BHG), Soldiers Cap Group (SCG; including the Toole Creek Volcanics), and Etheridge Group (EG), their key major and trace

element values are illustrated in Figures 7.3 and 7.4. Representative samples from the Eastern Creek Volcanics (ECV) are also provided for comparison (Fig. 7.5; Scott *et al.*, 2000; Ozchem, 2000). In comparing magmas from these sequences, I have concentrated on rocks with >4 wt% MgO to avoid sample compositions compromised by strong assimilation and/or fractionation. Based on their sampling location, mafic gneiss and amphibolite from the BHG are separated into two distinct groups: BHG-L, for rocks sampled along the line of lode at Broken Hill; and BHG-R, for 'regional' BHG mafic rocks (i.e. samples collected away from areas of significant mineralisation; Crawford, A.J., pers. comm., 2006).

From the variation diagrams, it is apparent that the EG mafic rocks possess major and trace element compositions virtually indistinguishable from those of the BHG-R and SCG. These three suites are also relatively Fe-rich and Ti-poor when compared with a suite of modern MORB (Fig. 7.3b, 7.3c, 7.4b, and 7.4c), suggesting that the mantle source for the mafic suites may have been slightly more Fe-rich than the N-MORB source. Compositional variation between the suites, where visible, is limited to samples of BHG-L, which on average possess higher Fe₂O₃^{total} and TiO₂, and generally have lower MgO values relative to the other suites (Fig. 7.3c). Samples of the SCG are also marginally depleted in Fe₂O₃^{total} relative to the other suites (Fig. 7.4c), however this effect may be partially a consequence of the sampling criterion, as only 'regional' SCG samples were included (i.e. samples collected away from areas of mineralisation and extreme alteration). Samples from the ECV correlate reasonably well with the other tholeiitic suites in the majority of major element variation diagrams (Fig. 7.5a-d), as would be expected for tholeiitic mafic rocks. The ECV samples follow the same major geochemical trends as the other mafic suites, but generally show a greater variation in some major and trace element values as they move towards more evolved (lower MgO) compositions (Fig. 7.5).



Figure 7.3: Selected major element oxides (in weight per cent) versus MgO and selected trace element variation diagrams for mafic rocks of the Broken Hill Group (BHG-L and BHG-R) and Etheridge Group (EG). Etheridge Group mafic rocks defined by 88 samples of the Dead Horse Metabasalt and Cobbold Metadolerite (this study; Appendix 4). For Broken Hill data source see text; data listed in Appendix 5. MORB field defined by 200+ samples from the East Pacific Rise (23°N-23°S; numerous sources; Appendix 5).



Figure 7.4: Selected major element oxides (in weight per cent) versus MgO and selected trace element variation diagrams for mafic rocks of the Soldiers Cap Group (SCG) and Etheridge Group (EG). For SCG data source see text; data listed in Appendix 5. Etheridge Group and MORB data sources as in Fig. 7.3.



Figure 7.5: Selected major element oxides (in weight per cent) versus MgO and selected trace element variation diagrams for mafic rocks of the Eastern Creek Volcanics (ECV) and Etheridge Group (EG). For ECV data source see text; data listed in Appendix 5. Etheridge Group and MORB data sources as in Fig. 7.3.
Trace element variation diagrams also support a geochemical relationship between the EG, SCG and BHG mafic rocks (Fig. 7.3e-h, 7.4e-h). Significantly, the tight correlation in immobile incompatible element values across the compositional range (denoted by MgO values) is consistent with the mafic rocks experiencing very similar fractionation histories, and derivation from similar parental magmas and mantle sources. The immobile incompatible element values for the mafic suites (e.g. Fig. 7.3e and 7.4e) are also consistent with a less depleted mantle source than for typical N-MORB. Zr values for the ECV are relatively higher compared to the other suites at the same stage of their evolution (e.g. Fig. 7.5e-h), suggesting that the composition of the parental magma for this suite may have differed from those of the other suites, an assertion strongly supported by REE data (see later). Variable assimilation of crustal material is the most likely reason for variation in Th/Nb values for each mafic suite (Fig. 7.3h, 7.4h, and 7.5h). In Chapter 6 the observed trend in Th/Nb values for the EG tholeiites was consistent with a local, upper-crustal contaminant. However, the contaminants of the Broken Hill and Mt Isa mafic suites are more difficult to evaluate, although the observed trend for the SCG suggests this suite may have assimilated a similar late-stage upper-crustal contaminant to the EG tholeiites (Fig. 7.4h).

The strong similarities in the immobile incompatible trace elements from the mafic rocks of the EG, SCG (Toole Creek Volcanics), BHG-R, and BHG-L are also illustrated by the MORB-normalised multi-element diagram (Fig. 7.6). The MORB-normalised profiles are broadly similar, with the majority of the averaged patterns appearing near-identical across the EG, SCG, BHG-R and BHG-L. The majority of incompatible trace element values for both the ECV and FCV are elevated relative to the other suites, again suggesting a different source. The average values for mafic rocks from the BHG-R, BHG-L and SCG suites also lie with the observed compositional range of the EG tholeites.

Similarities in both the geochemistry and fractionation history across the mafic suites are also reflected in their chondrite-normalised REE patterns (Fig. 7.7). Average REE values for the Dead Horse Metabasalt (DHM) and Cobbold Metadolerite (CMD) units lie between average values for the BHG and SCG tholeiites, and these values also lie within the range of all analysed EG tholeiites (indicated by the dashed line). Average REE values for both the ECV and FCV show significant enrichment in LREE relative to the other suites, suggesting more enriched parental magmas and a more enriched

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source composition. Comparable fractionation histories and parent magmas for the EG, SCG, and BHG tholeiites are suggested by similar levels of LREE enrichment (avg. $(La/Sm)_n = 1.11, 1.15, and 1.06$), and HREE depletion (avg. $(Gd/Yb)_n = 1.27, 1.32, and 1.16$).



Figure 7.6: MORB-normalised trace element diagram for average values of mafic rocks from the Broken Hill Group (BHG-R and BHG-L), Etheridge Group (CMD and DHM), and Toole Creek,

Eastern Creek and Fiery Creek Volcanics (Mt Isa). Data from individual Etheridge Group samples from this study lie within the field denoted by the dashed line. Normalising values after Pearce (1983).



Figure 7.7: Chondrite-normalised REE diagram for average values of mafic rocks from the Broken Hill Group, Etheridge Group, and Toole Creek, Eastern Creek and Fiery Creek Volcanics (Mt Isa). Symbols and fields as in Figure 7.6. Normalising values after Boynton (1984).

Although the MORB-normalised trace element and chondrite-normalised REE patterns suggest near-geochemically identical tholeiitic parent magmas for the EG, SCG and BHG mafic suites, they cannot implicitly indicate that the rocks were derived from a common source. Therefore, a number of immobile trace element ratio diagrams are presented to better discriminate between the mafic suites (Fig. 7.8). The close association of the mafic suites in the Zr/Nb versus Ti/Zr and Zr/Nb versus Zr/Y variation diagrams (Fig. 7.8) again suggests that the mafic rocks may have been sourced from similar parental magmas, as Zr/Nb is a fractionation-insensitive ratio that is believed to reflect mantle source values (Pearce, 1983). The large variation in Ti/Zr values relative to Zr/Nb for the BHG tholeiites is probably fractionation-related due to removal of Fe-Ti oxides with advancing fractionation. In summary, the Palaeoproterozoic mafic suites show remarkable similarity in their key immobile incompatible trace element and REE contents. In the following section, the Sm-Nd isotope geochemistry of these suites is compared.



Figure 7.8: (a) Zr/Nb versus Ti/Zr, and (b) Zr/Nb versus Zr/Y for mafic rocks of the Broken Hill Group, Etheridge Group, and Toole Creek, Eastern Creek and Fiery Creek Volcanics (Mt Isa).
Data sources as in Figures 7.3, 7.4, and 7.5. Average N-MORB and E-MORB values after Sun & McDonough (1989). Average Continental Crust values after Weaver & Tarney (1984).

Sm-Nd isotope geochemistry 7.4.3

Comparison of the Sm-Nd whole-rock isotope characteristics for tholeiitic rocks of the BHG, SCG, and EG is compromised to some extent by the paucity of available data. However, some conclusions can still be drawn from the available analyses. Only one Sm-Nd isotope analysis was available from the mafic rocks of the SCG, that being for a garnetiferous mafic rock in the Mt Norna Quartzite member (Page & Sun, 1988; Fig. 7.2). No Sm-Nd isotope analyses were available for mafic rocks of the BHG; however a recent study of tholeiitic metabasalts from the ca. 1685 Ma Lady Louise Suite (LLS) of the Curnamona Province (Conor & Fanning, 2001) by Rutherford et al. (2006) provided Sm-Nd analysis of 24 samples. Based on their age and geochemical characteristics, the metabasalts of the LLS have been correlated to those of the BHG (Page et al., 2005b; Rutherford et al., 2006), and are subsequently included here for comparison. The Sm-Nd data for both the SCG and LLS are summarised in Table 7.1, and individual analyses are listed in Appendix 5.

Sample	No. of analyses	Sm (ppm)	Nd (ppm)	Age	147Sm/144Nd	¹⁴³ Nd/ ¹⁴⁴ Nd	$\epsilon_{\rm Nd}(T)$	T _{DM} (Ga)	Data Source
Etheridge Group	10	3.3	11.0	1.67	0.1850	0.512708	3.9	2.49	Table 6.3
Soldiers Cap Group Mt Norna Quartzite	1	6.9	26.1	1.67	0.1603	0.512265	0.5	2.51	Page & Sun (1998)
Plumbargo	2	3.4	13.1	1.685	0.1687	0.512339	0.2	2.77	Rutherford et al. (2006)
Outalpa	9	3.4	12.4	1.685	0.1651	0.512327	0.8	2.63	
Waukaloo	2	2.9	9.9	1.685	0.1766	0.512509	1.8	2.62	
Mingary	9	2.1	6.7	1.685	0.1916	0.512663	1.6	2.98	
Mundi	2	4.8	14.5	1.685	0.2023	0.512747	0.9	n/a	

CHUR (chondrite uniform reservoir): 147 Sm/ 144 Nd = 0.1967, 143 Nd/ 144 Nd = 0.512638 DM (depleted mantle): 147 Sm/ 144 Nd = 0.2137, 143 Nd/ 144 Nd = 0.51315

 $\lambda = decay \ constant \ of \ ^{147} \ Sm = 6.54 \ x \ 10^{-12}$

Table 7.1: Summary of Sm-Nd isotope data for mafic rocks of the Etheridge Group, Soldiers Cap Group, and Lady Louise Suite (Curnamona Province). Individual analyses listed in Table 6.3 (Etheridge Group); for all other analyses see Appendix 5.

¹⁴⁷Sm/¹⁴⁴Nd ratios are variable within each terrane but mostly lie between 0.16 and 0.22, consistent with their tholeiitic nature (Fig. 7.9a). Samples from the LLS with lower ¹⁴⁷Sm/¹⁴⁴Nd ratios may have been contaminated at some stage during their genesis or reflect source heterogeneity, as they plot along a trend towards decreasing (negative) ε_{Nd} values (Fig. 7.9a; Rutherford *et al.*, 2006). The correlation in the range of 147 Sm/ 144 Nd ratios between the EG and LLS samples is also reflected in their T_{DM} ages (Fig. 7.9b).



Figure 7.9: (a) ¹⁴⁷Sm/¹⁴⁴Nd versus $\varepsilon_{Nd}(T)$, and (b) T_{DM} model ages versus ¹⁴⁷Sm/¹⁴⁴Nd for samples from the Etheridge Group (EG), Soldiers Cap Group (SCG), and Lady Louise Suite (LLS). Data sources as for Table 7.1.

Initial ε_{Nd} values [$\varepsilon_{Nd}(T)$] for metabasalts from the LLS were calculated for an assumed crystallisation age of 1685 Ma (Rutherford *et al.*, 2006), and the SCG sample for an assumed age of 1670 Ma (Page & Sun, 1998). $\varepsilon_{Nd}(T)$ values for the EG tholeiites (5.3-2.6; Table 6.3) are higher on average than those of both the LLS (1.8-0.2) and SCG (0.5) (Table 7.1), indicating that the EG tholeiites experienced little contamination by crustal material relative to the other suites (Chapter 6). Although some form of crustal contamination or late-stage assimilation of crustal material can account for the differing $\varepsilon_{Nd}(T)$ and 147 Sm/¹⁴⁴Nd values between the EG and LLS samples (Fig. 7.9), Rutherford *et al.* (2006) suggested that the observed variation in LLS geochemistry could not be adequately explained by crustal contamination, and was rather an effect of limited source heterogeneity. They demonstrated that some of the most isotopically primitive samples from the LLS (i.e. those samples with $\varepsilon_{Nd}(T)$ values > 0) were also displaced towards higher La/Sm values (Fig. 7.10).

Rutherford *et al.* (2006) proposed that tholeiites showing this relationship were most likely derived from a geochemically enriched, isotopically depleted lithospheric mantle source beneath the Olary Domain (western Curnamona Province) that had incorporated a subduction-related component. Their hypothesis is also supported by distinct, elevated REE profiles for the most enriched LLS samples (Fig. 7.11). They concluded that the enriched magma produced in the Olary Domain of the Curnamona Province during the ca. 1.7-1.67 Ga period of lithospheric thinning was absent, or at



least less obvious from the more depleted magmas emplaced within the Broken Hill Block to the east.

Figure 7.10: $\varepsilon_{Nd}(T)$ versus La/Sm for mafic rocks of the Etheridge Group and Lady Louise Suite. DM = Depleted Mantle ε_{Nd} value at 1.67 Ga, derived from DePaolo (1981). All other data sources and symbols as in Figure 7.9.



Figure 7.11: Chondrite-normalised REE compositions for representative tholeiitic samples from the Etheridge Group (sample 241) and Lady Louise Suite (samples M07, M12 and M27; data from Rutherford *et al.*, 2006). Samples M07 and M12 are 'enriched' tholeiites from the Olary Domain, western Curnamona Province. Sample M27 is a 'depleted' tholeiite from the Broken

Hill Domain, eastern Curnamona Province. Normalising values after Boynton (1984).

Conversely, the EG tholeiites of this study possess relatively flat REE profiles similar to the BHG, as well as trace element and isotopic values consistent with the late-stage assimilation of relatively minor amounts of crustal material into a geochemically and isotopically depleted magma (Chapter 6). Compositional variation between the most geochemically primitive and evolved EG tholeiites was also shown to be the result of magmatic fractionation. There is no strong evidence for the existence of a relatively enriched mantle source.

Despite the limited Sm-Nd isotopic datasets for the mafic rocks of the Curnamona Province and SCG, it is still possible to draw a meaningful relationship from the available data. The Nd isotope evolution diagram (Fig. 7.12) shows that the tholeiitic magmas which produced the mafic rocks in each terrane were derived from a relatively depleted mantle source over a similar time interval, but each evolved to unique isotopic characteristics based on the extent of crustal contamination and/or prior enrichment of the depleted mantle source. Variably enriched to depleted tholeiitic magma which forms the LLS in the western Curnamona Province produced mafic rocks with T_{DM} ages ranging from 3.31 to 2.26 Ga (Fig. 7.12; Rutherford *et al.*, 2006). These model ages overlap with those of both the EG (3.39 to 1.85 Ga; Table 6.3), and the single analysed mafic igneous rock from the SCG (2.51 Ga; Page & Sun, 1998). The overlap in T_{DM} ages for these terranes coupled with similar crystallisation ages of their mafic rocks is significant, as it suggests that the processes which led to the emplacement of tholeiitic magma in each terrane occurred over a similar timescale, producing geochemically similar intrusive and/or extrusive igneous rocks.

Although the dominant Nd isotope evolution trajectories for each mafic suite differ significantly (Fig. 7.12), there is some inter-regional correlation between individual analyses. A number of processes have been suggested which together may account for the observed variation. Page & Sun (1998) suggested that alteration may have affected the Sm-Nd systematics of some samples from the SCG, including the mafic sample from the Mt Norna Quartzite (Table 7.1). However, without access to additional Sm-Nd data from this area it is impossible to predict whether the Nd isotope evolution trajectory of this sample is representative of the SCG mafic rocks. Alternatively, for the LLS metabasalts an anomalous enriched component in the source is proposed, with little or no additional input from crustal material (Rutherford *et al.*, 2006). Despite these differences, the majority of the available geochemical and isotopic

evidence suggests that tholeiitic mafic rocks emplaced ca. 1.69-1.67 Ga throughout the eastern Australian Proterozoic terranes are geochemically linked, and may have even been derived from a geochemically similar, relatively depleted upper-mantle source. The significance of this relationship, for both the Georgetown Inlier and the North Australian Craton, is discussed below.



Figure 7.12: Nd isotope evolution diagram for mafic rocks of the Etheridge Group, Soldiers Cap Group (SCG), and Lady Louise Suite (Curnamona Province; Fig. 7.1). Depleted mantle curve (DM) assumes a linear evolution of the depleted mantle from CHUR (ε_{Nd} = 0) at 4.56 Ga to a present-day value of +10 (Goldstein *et al.*, 1984; Jahn & Condie, 1995; Blewett *et al.*, 1998). Data

sources as for Table 7.1.

7.5 Discussion

7.5.1 Variation in the level of Fe-enrichment: the effect of source or setting?

The generation of ca. 1670 Ma Fe-rich tholeiitic melts with remarkably similar geochemistry within Broken Hill Block, Mt Isa Eastern Succession, and Georgetown Inlier has important implications for lithospheric evolution on the eastern margin of the North Australian Craton during the Palaeoproterozoic. However, before this issue can be addressed the origin and evolution of the individual tholeiitic melts must be considered.

Strongly Fe-rich tholeiitic basalts (>15 wt% $Fe_2O_3^{total}$), such as those found in the Proterozoic terranes of eastern Australia, are rare and generally restricted to extensional tectonic settings (Brooks *et al.*, 1991). Numerous conditions have been

suggested to account for the development of Fe-enriched assemblages during the fractionation of tholeiitic melts, including the composition of the upper mantle source, low oxygen fugacity (fO_2) and closed system fractionation in the melt, and the relatively high density of Fe-rich liquids (Sinton *et al.*, 1983; James *et al.*, 1987; Brooks *et al.*, 1991). Continental rift settings are highlighted as favourable environments, as they are thought to contain high-level magma chambers that are well removed from their mantle sources and thus are more likely to develop as relatively closed systems (Sinton *et al.*, 1983; Brooks *et al.*, 1991).

It is beyond the scope of this study to assess the viability of each of these conditions in influencing the general level of Fe-enrichment in tholeiitic melts. However, based on the combination of geochemistry and lithostratigraphic evidence, reasonable assumptions can be made regarding the possible contributions from each of these conditions to Fe-enrichment in the tholeiitic rocks of this study. Brooks et al. (1991) suggested that the trend towards strong Fe-enrichment of a tholeiitic liquid was a direct result of differentiation in a closed system at a relatively low fO_2 state. For this to occur, the low fO_2 state would need to have been inherited from the source or magma chamber. Additionally, the low oxidation state would be maintained so long as little to no replenishment of the magma took place, essentially implying fractionation as a closed system. If replenishment of the magma did take place, equally low fO₂ magma would be required to maintain the reduced state. If these conditions were imposed on the BHG, SCG, and EG suites, then based on the nearidentical nature of the Fe-enrichment trends for each terrane (Fig. 7.3 and 7.4), similar low fO_2 conditions coupled with little to no replenishment of the source magmas would have been required to produce the trends observed.

Based on the conditions for Fe-enrichment discussed above, I prefer a scenario whereby the Fe-rich tholeiites of the BHG, SCG, and EG were derived from a geochemically similar upper-mantle source, and evolved along a unique tholeiitic fractionation trend towards strong Fe-enrichment (up to 20 wt% $Fe_2O_3^{total}$). The Fe-enrichment observed in each terrane was accommodated by fractionation within a closed system at a relatively low oxidation state, which suppressed magnetite crystallisation until relatively late in the fractionation sequence.

7.5.2 Base metal mineralisation in the lower Etheridge Group: is there a worldclass Broken Hill-type deposit in the Georgetown Inlier?

The geochronological, geochemical, and lithostratigraphic similarities between the Etheridge Group, Mt Isa Eastern Succession, and Broken Hill Block suggest that the Proterozoic rocks of the Georgetown Inlier were also a suitable host for base-metal mineralisation. Numerous small stratabound and/or stratiform base metal deposits are found within the lower Etheridge Group, and are hosted by the Einasleigh Metamorphics (Bain *et al.*, 1990). Further west in the Robertson River Subgroup, a zone of high Zn, Pb and Cu has been reported in the lowermost part of the Daniel Creek Formation, 8 km northwest of Gilberton (Onley, 1978). However, no other significant base metal deposits have been found in this part of the inlier (Withnall *et al.*, 1997). Despite the presence of numerous small-scale deposits, world-class Broken Hill-type base metal deposits of a similar size to those found at Broken Hill and in the Mt Isa Eastern Succession (e.g. Cannington deposit) are not found in the Etheridge Group are discussed below.

Although debate concerning the exact lithological, geochemical, and regional relationships responsible for the development of Broken Hill-type deposits is ongoing, reasonable constraints can be placed on their genesis based on the relationships observed in existing deposits. Identification of these criteria is essential for determining the economic prospectivity of exploration sites, including the Georgetown Inlier. A comprehensive summary of the features associated with Broken Hill-type deposits was presented by Walters (1998); however, only the key geologic criteria are mentioned here.

Both the Broken Hill and Cannington deposits are hosted by quartzo-feldspathic sequences, with the host rock dominated by garnet quartzite that has been metamorphosed to upper amphibolite-granulite facies. The quartzite lies at the transition between a lower sequence consisting predominantly of quartzo-feldspathic units, and a dominantly psammopelitic to pelitic upper stratigraphy (Walters, 1998). Occurring proximal to the ore lodes at both Broken Hill and Cannington are concentrations of amphibolite and felsic igneous rocks (Fig. 7.13). At Broken Hill, the felsic igneous rocks are termed 'Potosi' gneiss (Stevens *et al.*, 1988), whereas at Cannington the felsic igneous rocks are pegmatites (Williams & Smith, 2003). The

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amphibolites are generally described as sills emplaced during the intrusion of tholeiitic magmas, introduced during rifting of the depositional basins. The magnetic response of the amphibolites has a variable association with mineralisation, as demonstrated by the poor magnetic response produced by magnetite in amphibolite in the Broken Hill main lode, whereas at Cannington magnetite has produced a direct magnetic target (Walters, 1998).



Figure 7.13: Simplified cross-sections of (a) the Broken Hill Main Lode (in the region of the Zinc Corp main shaft; after Walters, 1998), and (b) Cannington deposit (southern zone, through 4800N; after Williams & Smith, 2003).

In the Etheridge Group, base metal deposits hosted by the Einasleigh Metamorphics fall into four categories: stratabound Fe-Cu \pm Ag sulphide deposits (e.g. Einasleigh), Fe-Zn sulphide deposits (e.g. Eveleigh), Fe-Pb-Zn-Cu \pm Ag sulphide deposits (e.g. Mount Misery), and Cu/Fe/Si/Ba gossans in the Werrington area (Withnall *et al.*, 1997). These deposits have features that suggest an affinity to Broken Hill-type mineralisation, as they are commonly associated with quartzite and quartzo-feldspathic units, and are concentrated at the transition between a lower calcareous quartzo-feldspathic sequence, and an upper psammopelitic sequence (Withnall *et al.*, 1997). Concentrations of amphibolite (mafic gneiss) are also found in near-ore

sequences (Withnall *et al.*, 1997), and quartzo-feldspathic gneiss associated with the deposits may also be of possible volcanic origin (Johnston, 1978; Oversby, 1981).

In the south-western extensions of the lower Etheridge Group (Fig. 2.6), the transition from lower quartzo-feldspathic sequences to upper psammopelitic to pelitic sequences are located at the contact between the Bernecker Creek and Daniel Creek Formations. Metadolerite sills belonging to the Cobbold Metadolerite member are locally intruded; however apart from sparse and volumetrically minor granophyric bodies representing late-stage felsic differentiates in the metadolerite, no other concentrations of contemporaneous felsic igneous or volcanic rocks are observed. Peak metamorphism in the area is also generally restricted to greenschist facies assemblages.

Taking these characteristics into consideration, lower Etheridge Group rocks in the south-western and western extensions of the Georgetown Inlier are an unlikely host for large Broken-Hill-type orebodies, as a number of key lithologic and metamorphic conditions have not been met. However, higher-grade equivalents represented by the Einasleigh Metamorphics further east appear conducive to mineralisation of this type, as exemplified by the numerous small base metal deposits already identified. It is therefore possible that the majority of mineralisation in the lower Etheridge Group was restricted to the eastern half of the succession, with the presence of possible volcaniclastic sequences (Black *et al.*, 2005), and the influx of oxidised ore fluids restricted to this area alone. The prospectivity of the Einasleigh Metamorphics is further enhanced by the presence of elevated Na contents relative to other Etheridge Group rocks further west. This suggests a regional Na metasomatic event, perhaps associated with mineralisation, may have affected the Einasleigh Metamorphics during the Proterozoic (Black *et al.*, 2005).

The absence of an exposed volcaniclastic sequence, coupled with low metamorphic grade in the central and western extensions of the lower Etheridge Group does not necessarily imply that mineralisation is absent from these areas. Leucogneiss in the Einasleigh Metamorphics dated by Black *et al.* (2005) contains zircons which they inferred were of possible volcaniclastic origin. If this volcaniclastic source is located within the Einasleigh Metamorphics, it is possible that this sequence may also occur within un-exposed equivalents of the Bernecker Creek and/or Daniel Creek Formations further west (Withnall *et al.*, 1988a). Subsequently, if higher-grade equivalents of the Bernecker Creek and Daniel Creek Formations are located at depth

in the central and western parts of the inlier, they may provide a viable target for further exploration. High magnetic responses are also recorded in the central and south-western parts of the Etheridge Group, and are associated with exposed or partially exposed amphibolite (Withnall *et al.*, 2003a), or amphibolite under cover (Withnall *et al.*, 2003b). As these magnetic highs also occur proximal to exposed contacts between the Daniel Creek and Bernecker Creek Formations, they may also provide viable exploration targets.

7.6 Conclusions

The Fe-rich tholeiites of the lower Etheridge Group are hosted by late Palaeoproterozoic metasedimentary rocks, and show strong geochemical affinities with similarly aged mafic rocks from the Soldiers Cap Group and Broken Hill Group in the Mt Isa Eastern Succession and Broken Hill Block respectively. These affinities include the correlation of major and incompatible trace element values, and variation diagram trends that are virtually indistinguishable (e.g. Fig. 7.3 and 7.4). Compositional variation, where present, is limited to samples of Broken Hill Group mafic gneiss collected proximal to the main line-of-lode at Broken Hill. These rocks generally have higher Fe₂O₃^{total} and TiO₂ compositions relative to other samples at a given stage of their evolution, and their unusual geochemistry may be related to processes responsible for the mineralisation (Phillips *et al.*, 1985).

Similarities in the fractionation histories of the mafic suites are also supported by the near-identical trends in compatible and incompatible trace elements including Ni and Zr. In addition, similar REE and incompatible trace element patterns are also suggestive of similar fractionation histories for the mafic rocks (Fig. 7.6 and 7.7). They also imply the mafic rocks followed the same or very similar tholeiitic liquid lines of decent. By comparison, tholeiitic basalts from the Mt Isa Western Succession (represented by the Eastern Creek and Fiery Creek Volcanics) have higher incompatible trace element and LREE values, suggesting slightly more alkaline parent magmas than for the Broken Hill, Soldiers Cap, and Etheridge Group tholeiites. The correlation in immobile trace element ratios (e.g. Fig. 7.8) also suggests the Broken Hill, Soldiers Cap, and Etheridge Group tholeiites possessed similar initial geochemical compositions. This correlation, coupled with the observed similarities in

trace and rare earth element patterns, implies that the rocks were most likely derived from chemically similar parent magmas.

In addition to major, trace, and rare earth element geochemistry, the Etheridge Group tholeiites possess ¹⁴⁷Sm/¹⁴⁴Nd ratios and T_{DM} model ages similar to the ca. 1685 Ma Lady Louise Suite, a suite of variable enriched tholeiitic basalts in the central-western Curnamona Province that have been correlated to the mafic rocks of the Broken Hill Group. In addition, a contemporaneous mafic igneous rock in the Soldiers Cap Group possesses a similar T_{DM} age to those of the Etheridge Group tholeiites. The similarities in Sm-Nd systematics between the terranes suggest that separation of their parental magmas from a depleted mantle source occurred over a similar time-span, providing a further link for the correlation of the tholeiitic rocks from each terrane.

The geochemistry and geochemical similarities between similarly aged Fe-enriched tholeiites from the Broken Hill, Soldiers Cap and Etheridge Groups suggest they each formed from Fe-rich parental melts, with strong Fe-enrichment most likely occurring as a result of closed system fractionation under generally reduced conditions (low fO_2). Additional Fe-enrichment up to 25 wt% Fe₂O₃^{total}, observed in the tholeiites along the line-of-lode at Broken Hill, may be linked to mineralisation processes.

The geochemical similarities observed between the Fe-rich tholeiites of the Broken Hill, Soldiers Cap and Etheridge Groups are significant; and if Fe-rich tholeiites from both Mt Isa Eastern Succession and Broken Hill are associated with the development of significant mineralisation, may suggest that ore-forming processes also occurred within the lower Etheridge Group. The correlation of these Palaeoproterozoic mafic suites suggests that they were derived from similar depleted mantle sources. This relationship, coupled with other lithostratigraphic, geochronological, and structural evidence, suggests that the Etheridge Group was deposited in an extensional basin not far removed from the Mt Isa and Broken Hill terranes. This has important implications for the Palaeoproterozoic location of the Georgetown Inlier, as it implies that the terrane formed an integral part of the North Australian Craton during this time.

Chapter 8.

Summary and conclusions

8.1 Palaeoproterozoic evolution of the lower Etheridge Group

Sedimentary rocks of the lower Etheridge Group were deposited in a subsiding volcanic passive continental margin setting between ca. 1720 Ma and 1650 Ma. Deposition of the earliest sedimentary units had commenced by ca. 1715 Ma, determined from crystallisation ages for two samples of the Cobbold Metadolerite that were intruded into the lower Etheridge Group. This is the preferred maximum depositional age for the Etheridge Group, as it is within error of zircon SHRIMP ages obtained from samples of interbedded leucogneiss in the underlying Einasleigh Metamorphics, which have previously placed an upper limit on the depositional age of the Etheridge Group (Black *et al.*, 2005).

Voluminous Fe-rich tholeiitic igneous rocks were emplaced into the lower Etheridge Group from around ca. 1670 Ma. The mafic rocks are assigned to two units: the extrusive Dead Horse Metabasalt (DHM), a 1000m-thick series of fine- to coarsegrained basalts and pillow lavas; and the intrusive Cobbold Metadolerite (CMD), a multitude of sills and lesser dykes. At least one later pulse of magmatic activity is observed where CMD sills are intruded into the Lane Creek Formation, the upper unit of the Robertson River Subgroup. These uppermost sills were not dated in this study due to an absence of granophyric differentiation; however CMD sills intruding the Bernecker Creek Formation, which is lower in the stratigraphic sequence, returned pooled ages of ca. 1645 Ma that may date the later generation of sills.

Following deposition, the lower Etheridge Group sedimentary rocks underwent an extended period of regional-scale orogenesis and metamorphism. At least two discrete tectonothermal events are recorded, with the intensity of each event generally increasing from west to east. EMPA dating of metamorphic monazite from this and a previous study (Cihan *et al.*, 2006) suggests that the earliest orogenic event occurred at around ca. 1585 Ma. This event corresponds to the D₁ event of Withnall *et al.* (1988a), and was associated with regional-scale folding, and the development of a pervasive metamorphic fabric. In the area around 'Gum Flats', the metasediments of the Robertson River Subgroup record peak metamorphic conditions of up to 650 °C

and 7.5 kbar associated with this event. Samples of Einasleigh Metamorphics mica schist that are in fault-contact with Robertson River Subgroup metasediments in this area record P-T conditions of up to 750 °C and 7.5 kbar also associated with this event.

The second regionally significant tectonothermal event was associated with the emplacement of voluminous S-type granitoids, and corresponds to the D_2 event of Withnall *et al.* (1988a). LA-ICPMS analysis of igneous zircons collected from granitoid samples, coupled with EMPA monazite analysis of lower Etheridge Group metasediments indicate that this event occurred between ca. 1550 Ma and 1530 Ma. An estimate of the *P-T* conditions associated with this event was limited to a sample of mica schist from the Daniel Creek Formation that returned a retrograde M₁-early M₂ *P-T* estimate of 550 °C and 3 kbar. EMPA monazite analysis of a sample containing a later (S₃) foliation suggests a possible weak metamorphic event occurred at ca. 1500 Ma, and may date the previously unconstrained D₃ event of Withnall (1996).

8.2 Geochemical evolution of Fe-rich tholeiites within the lower Etheridge Group

The DHM and CMD are shown to be co-magmatic and share a common geochemical history. They possess geochemical characteristics typical of relatively evolved, low-K, Fe-rich continental tholeiites and basalts erupted in the early stages of development of volcanic passive margins, typified by the early Tertiary E Greenland margin. Trace element, REE and isotopic data indicate that the rocks were generated by partial melting of a depleted mantle source, and that the magmas continued to evolve along a trend of Fe-enrichment involving fractional crystallisation within a closed system. As a result, some samples show extreme Fe-rich compositions (>20 wt% Fe₂O₃^{total}). $\varepsilon_{Nd}(T)$ values average +4.0, suggesting the magma was originally derived from a depleted mantle source. A poorly constrained Sm-Nd isochron age, coupled with $\varepsilon_{Nd}(T)$ and T_{DM} values, indicate that the mafic rocks crystallised at ~1.69 Ga. This correlates with U-Pb zircon crystallisation ages obtained from some of the mafic samples. T_{DM} model ages also indicate that the mantle source of the mafic rocks formed during the early Palaeoproterozoic (ca. 2.36 Ga).

In several samples, REE values show distinct LREE-enriched patterns consistent with small degrees of crustal contamination. Trace element ratio diagrams (e.g. Fig. 6.14, 6.21) indicate that the crustal contaminant was derived from the host metasediments of the lower Etheridge Group, with the majority of samples showing less than 10% assimilation of sedimentary material.

Based on their geochemistry and geological setting (i.e. large volumes of sills), a continental rifting environment provides the appropriate conditions for the emplacement of the Etheridge Group mafic rocks. During opening of the rift-system and deposition of the lower Etheridge Group metasediments, tholeiitic magmas derived from convecting upper mantle melted due to decompression, and tholeiitic magma was introduced into the upper crust. Differentiation of the tholeiitic magma to the most Fe-rich (13-20 wt% Fe₂O₃^{total}) compositions occurred without the significant input of crustal material via processes associated with fractional crystallisation including partial melting, until emplacement within the overlying lower Etheridge Group sediments as the CMD member. During the initial period of sustained magmatism, batches of melt that reached the surface were erupted to form the DHM member.

8.3 Implications for the Palaeoproterozoic location of the Georgetown Inlier (ca. 1700-1500 Ma)

The lithological, geochemical, and geochronological similarities between the Georgetown Inlier, the Broken Hill Block, and the Mt Isa Eastern Succession have important implications for the Palaeoproterozoic position of the Georgetown Inlier. Such similarities, when found to occur over significant intervals of time, imply that the terranes must have been spatially associated. Both the Broken Hill Block and Mt Isa Eastern Succession have previously been included as integral parts within the North Australian Craton (NAC; Myers *et al.*, 1996; Giles *et al.*, 2002). Subsequently, the observed similarities suggest that the Georgetown Inlier also developed along the eastern-most section of the craton. Therefore, plate tectonic models for the evolution of the Australian Plate between ca. 1700 and 1500 Ma should incorporate the Georgetown Inlier in their construction.

The evolution of the eastern margin of the NAC between 1800 Ma and 1500 Ma was summarised by Betts & Giles (2006), following the studies of Wingate & Evans (2003) and Giles *et al.* (2004). Their model proposed that plate tectonic processes played an integral part in the development of the craton, and its constituent extensional basins. In particular, their model suggested that subduction processes along the southern margin of the NAC promoted the development of several extensional basins within the plate interior, corresponding to the sedimentary successions now located within the Broken Hill, Mt Isa, and Georgetown Inliers (Giles *et al.*, 2002). Following deposition, the basins experienced similar, but not identical deformation histories corresponding to a number of regional- and local-scale orogenic events (Betts & Giles, 2006). A summary of this history for the eastern margin of the NAC is provided below; however particular attention is given to the location and tectonic processes responsible for the development of the Georgetown Inlier and its associated terranes.

8.3.1 Onset of deposition in the Georgetown Inlier, and renewed deposition in the Broken Hill Block and Mt Isa Eastern Succession (ca. 1740-1600 Ma)

By ca. 1740 Ma, a number of extensional basins had developed within the NAC in response to earlier periods of lithospheric extension (Betts & Giles, 2006). These basins included the McArthur basin, and both the Western and Eastern Successions of the Mt Isa Inlier (Fig. 8.1a). At ca. 1740 Ma, a depositional hiatus was recorded in these basins, and additionally in the Curnamona Province. This hiatus has been linked to the onset of collision between the Mawson continent (comprising the Gawler Craton and associated rocks in East Antarctica) and the NAC (Fig. 8.1a; Scott *et al.*, 2000; Giles *et al.*, 2002; Betts & Giles, 2006).

Sedimentation in the Mt Isa Eastern Succession and Curnamona Province had recommenced by ca. 1720 Ma (Page & Sun, 1998; Page *et al.*, 2005a). By this time, deposition of the oldest exposed sedimentary units had also commenced within the Georgetown Inlier. This period of basin development and sedimentation was linked to the retreat (roll-back) of a north-dipping subduction zone along the southern margin of the NAC (Fig. 8.1a; Giles *et al.*, 2002; Betts & Giles, 2006). The subsequent intrusion and eruption of geochemically related, Fe-rich tholeiitic rocks into the extensional basins between ca. 1680 Ma and 1660 Ma occurred as a result of

lithospheric thinning beneath the NAC. However, the amount of lithospheric thinning was not enough to induce the development of new oceanic crust.



Figure 8.1: Interpreted tectonic evolution of the eastern North Australian Craton between ca.
1740 and 1500 Ma. Regional scale tectonic processes, orogenic events, and their location follow the interpretation of Betts & Giles (2006), MacCready *et al.* (2006), and references therein.
Figures modified after Betts & Giles (2006).

A significant change in the style of sedimentation is recorded in the Mt Isa, Broken Hill and Georgetown extensional basins after ca. 1660 Ma, with sedimentary lithologies progressively changing from continental shelf-type sands and silts to deeper marine sediments including turbidites, muds and carbonaceous sediments (Willis *et al.*, 1983; Withnall, 1996; Betts *et al.*, 2006). This transition was co-incident with the start of the Isa Superbasin cycle, and was thought to coincide with the opening of an oceanic basin to the east of the Georgetown Inlier, with the locus of lithospheric extension shifting from beneath the NAC to the new spreading centre (Fig. 8.1b; Betts *et al.*, 2003; Betts & Giles, 2006).

8.3.2 Orogenesis along the eastern margin of the North Australian Craton (ca. 1600-1500 Ma)

A series of regional-scale orogenic events are recorded within the extensional basins along the eastern margin of the NAC between ca. 1600-1500 Ma. The earliest regional-scale event occurred between ca. 1600-1580 Ma, and was associated with a period of contemporaneous north-south oriented thin-skinned deformation in the Broken Hill Block (Olarian Orogeny), Mt Isa Inlier (Isan Orogeny) and Georgetown Inlier (Page *et al.*, 2005a; Betts & Giles, 2006; Giles *et al.*, 2006). This event also produced peak metamorphic conditions in the Georgetown Inlier (D₁), and the development of upper amphibolite facies assemblages in the Mt Isa Eastern Succession (D₂; Giles *et al.*, 2006; Sayab, 2006). Betts & Giles (2006) related this event to tectonic activity occurring along the southern margin of the NAC (Fig. 8.1c).

Subsequent orogenesis in the extensional basins of eastern Australia recorded a change from generally north-south oriented thin-skinned deformation to east-west oriented thick-skinned deformation, commencing at ca. 1565 Ma and continuing until ca. 1500 Ma (Betts & Giles, 2006; MacCready *et al.*, 2006). In contrast to the ca. 1600-1580 Ma event, the timing of metamorphism and felsic magmatism associated with the later event differed between the terranes, suggesting that this event may have been partly diachronous across the craton. In the Georgetown Inlier, voluminous felsic magmatism dated to ca. 1550 Ma is associated with this event. In the Mt Isa Eastern Succession magmatism of this age is restricted to the Marramungee Granite (Page & Sun, 1998). This led MacCready *et al.* (2006) to suggest that thin-skinned deformation occurring in the Mt Isa Eastern Succession before ca. 1540 Ma may have been linked

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to deep-crustal thickening and thick-skinned deformation beneath the Georgetown Inlier at ca. 1550 Ma.

Betts & Giles (2006) suggested that the onset of thick-skinned orogenesis and associated felsic magmatism in the Georgetown Inlier was related to westward subduction of oceanic crust along the eastern margin of the NAC, and the subsequent collision of the Laurentia continent with eastern Australia at ca. 1540 Ma (Fig. 8.1d). However, the absence of unambiguous arc-related rocks or igneous rocks with subduction signatures along the eastern margin of the NAC questions this interpretation. Regardless of the cause, the subsequent locus of crustal shortening shifted, and was accommodated by thick-skinned deformation in the Mt Isa Inlier (Betts *et al.*, 2006; MacCready *et al.*, 2006). This phase of deformation was later accompanied by the intrusion of ca. 1520-1500 Ma granitoids in the Mt Isa Eastern Succession (Williams and Naraku Granites; Page & Sun, 1998), and is also recorded by S₃ metamorphic monazite ages in the Georgetown Inlier (Cihan *et al.*, 2006; and this study).

8.4 Conclusions

New geological data and interpretations presented in this thesis are combined with data from previous studies of the lower Etheridge Group to make the following important statements on the geologic characteristics and history of the Georgetown Inlier:

- The mafic rocks hosted by metasediments of the lower Etheridge Group form two texturally distinct units: the extrusive Dead Horse Metabasalt, a 1000 mthick series of fine- to coarse-grained basalts and pillow lavas; and the intrusive Cobbold Metadolerite, a multitude of sills and lesser dykes.
- ii) The DHM and CMD are co-magmatic, and possess geochemical characteristics typical of relatively evolved, low-K, Fe-rich continental tholeiites and tholeiites erupted along volcanic passive margins such as the E Greenland margin during rifting and the early stages of ocean opening. The rocks evolved along a trend of strong Fe-enrichment, resulting in some samples containing in excess of 20 wt% Fe₂O₃^{total}.

- iii) The mafic rocks possess Sm-Nd isotopic values consistent with derivation from a depleted mantle parent (initial ε_{Nd} values between +2.6 and +5.3). T_{DM} crustal residence ages range from 3.39 Ga to 1.85 Ga. Variation in the Sm/Nd isotopic ratio and observed trends in trace element values are consistent with a volumetrically minor amount of contamination by host metasediments of the lower Etheridge Group which occurred during the emplacement of the mafic rocks. As expected, those samples showing the strongest LREEenrichment also show the lowest initial ε_{Nd} values, recording this crustal contamination.
- iv) The Etheridge Group mafic rocks show strong geochemical affinities to contemporaneous Fe-rich tholeiites from the Soldiers Cap Group and Broken Hill Group of the Mt Isa Eastern Succession and Broken Hill Block respectively. This implies that the mafic rocks of each terrane may have been derived from a similar depleted mantle source during breakup of continental crust along the eastern margin of the NAC.
- v) The mafic rocks of the Etheridge Group crystallised between ca. 1715 Ma and ca. 1630 Ma in at least two distinct pulses. The most significant was related to the eruption of the DHM and associated CMD sills, occurring at ca. 1670 Ma. A later pulse was related to the intrusion of CMD sills into the upper Robertson River Subgroup, and occurred at ca. 1640 Ma. The crystallisation age of the DHM member is significant, as it approximates the inferred crystallisation ages of mafic rocks in both the Broken Hill Group (ca. 1685 Ma) and Soldiers Cap Group (ca. 1670-1650 Ma).
- vi) The S-type Mt Hogan and Ropewalk Granites, part of the voluminous Forsayth Batholith, were intruded into the lower Etheridge Group at ca. 1550 Ma.
- vii) Inheritance patterns produced from detrital zircon analyses of metasediments from the lower Etheridge Group, Broken Hill Group and Mt Isa Eastern Succession share some common components. These include a late-Archaean component (ca. 3100-2500 Ma), which may represent a common Archaean basement, and a significant mid-Palaeoproterozoic component (ca. 2100-1800 Ma). The similarities in inheritance patterns suggest a shared provenance for the sediments.

- viii) Prograde metamorphism in the lower Etheridge Group had commenced by ca. 1600 Ma, and had reached peak medium *P-T* conditions at around ca. 1585 Ma. This age corresponds to inferred orogenic ages in both the Broken Hill Block and Mt Isa Eastern Succession. Following a period of retrogression, a second medium-*T*, low-*P* event occurred at ca. 1550-1530 Ma, associated with the intrusion of voluminous S-type felsic igneous rocks. A third, relatively weak metamorphic event is recorded at ca. 1500 Ma, and may relate to a similar period of thermal activity and granitoid emplacement in the Mt Isa Eastern Succession occurring between ca. 1520 and ca. 1500 Ma.
- ix) Metasedimentary rocks of the Robertson River Subgroup near Gum Flats record peak metamorphic conditions of up to 650 °C and 7.5 kbar associated with M₁. Retrogressive metamorphism produced a clockwise *P*-*T* path for the metasediments, leading to initial *P*-*T* conditions for M₂ of 550 °C and 3 kbar.
- x) The geochemical affinities of the Etheridge Group mafic rocks, coupled with geochronological and lithostratigraphic evidence, suggests that the Georgetown Inlier was proximal to the Mt Isa and Broken Hill Inliers during the Palaeoproterozoic. This further implies that the Georgetown Inlier formed the easternmost part of the NAC during the Proterozoic.

The combination of geochemical, geochronological, and tectonostratigraphic evidence presented in this thesis provides strong support to the hypothesis that the Georgetown Inlier was proximal to the sedimentary basins of northeastern Australia, in particular the Broken Hill Block and Mt Isa Eastern Succession, during the Palaeoproterozoic. Additionally, this relationship implies that the Georgetown Inlier formed the easternmost extension of the NAC during this time, and that the Georgetown Inlier is an integral component of the Proterozoic history of the Australian continent.

8.5 Suggestions for future work

Although compelling evidence is presented here and by other sources documenting the geochemical nature and affinities of the Etheridge Group Fe-rich tholeiites, additional work is required to resolve some remaining issues. In particular, a detailed isotopic analysis of the Fe-rich tholeiites from both the Mt Isa Eastern Succession and Broken Hill Block would categorically determine if these mafic rocks shared the same depleted mantle source as the mafic rocks of the Etheridge Group, and would additionally determine if the rocks shared the same or similar crustal residence ages. Additional isotopic dating of the Fe-rich tholeiites from both the Soldiers Cap and Broken Hill Groups would provide an exact crystallisation age for the mafic rocks, and would also enable a more accurate comparison to the crystallisation age of the Etheridge Group mafic rocks.

Significant work is also required to resolve the tectonothermal history of the lower Etheridge Group, in particular the variation in thermobarometric conditions and its control on mineralogy across the group. Although evidence presented here and in previous studies indicates a clockwise *P*-*T*-*t* path for the Etheridge Group involving medium *P*-*T* metamorphic conditions, contemporaneous *P*-*T*-*t* paths proposed for the Mt Isa Inlier and Broken Hill Block involve anticlockwise *P*-*T*-*t* paths, with peak metamorphism reached during high-*T*, low-*P* conditions. Further detailed geochronological work is required to constrain the ages of the M₁ and M₃ events in the Etheridge Group, as only limited data for these two events was obtained over the course of this study. In particular, the ca. 1500 Ma monazite age has possible implications to a contemporaneous period of thermal activity in the Mt Isa Eastern Succession. Further investigation of these issues should determine if the orogenic history of the Georgetown Inlier approximated that of the Mt Isa Inlier and Broken Hill Block.

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APPENDIX

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Appendix 1: Representative Thin Section Descriptions

The thin sections described below are representative of the lithologies discussed in Chapters 2 and 3. Where heterogeneities in the lithologies were observed more than one sample description is provided. Thin sections of samples used in geochemical, geochronological, and thermobarometric analysis are also described here. Sample locations including grid references are presented in Appendix 6.

Einasleigh Metamorphics

282: Einasleigh Metamorphics mica schist collected from outcrop south of Gilberton Fault near Gum Flats. Sample contains small abundant laths of quartz, fine-grained muscovite and biotite surrounding garnet porphyroblasts. Garnet porphyroblasts are generally small (up to 1 mm diameter), and subhedral to anhedral.

283: Very strongly foliated muscovite schist from same outcrop as sample 282. Sample contains quartz, abundant muscovite, anhedral biotite. Garnet porphyroblasts (up to 1 mm diameter) also occur, but are less common in this sample.

317: Biotite gneiss from outcrop near Percyvale. Contains coarse-grained subhedral biotite laths, quartz, andesine to albite, calcite, small garnet porphyroblasts, and accessory sphene and Fe-Ti oxide (possibly haematite). Thin discontinuous layers of retrogressive sericite have also replaced a prograde phase, possibly sillimanite.

318: Einasleigh Metamorphics amphibolite containing foliated euhedral but rounded olive hornblende, albite, and lesser quartz, clinopyroxene and Fe-Ti oxide. Sample is near granulite facies or high-grade amphibolite facies, but is probably equivalent to amphibolites from lower grade areas.

324: Coarse-grained muscovite-biotite-quartz mica-schist. Sample contains large clots of muscovite and biotite, in addition to quartz and lesser albite. Metamorphic monazite is accessory.

Bernecker Creek Formation

7: Psammitic rock with 60-70% fine-grained quartz. Abrupt change to pelitic texture containing microgranular quartz. Small percentage of calcite and Fe-Ti oxide also present in psammitic section.

10: Fine-grained psammitic rock containing muscovite and quartz as the dominant matrix phases, along with accessory fine-grained albite, epidote and biotite.

13: Pelite containing fine-grained epidote occurring as bands of aligned prismatic grains. Some chlorite and lesser Fe-Ti oxide also associated with the epidote bands. Surrounding matrix comprises fine-grained quartz.

124: Psammo-pelitic rock containing up to 50% fine-grained quartz, and lesser fine-grained calcite, muscovite, and accessory Fe-Ti oxide.

292: Fine-grained mica schist containing layer-parallel foliation. Dominant phase is calcite, with lesser fine-grained biotite, muscovite, quartz, and feldspar also present.

293: Moderately to strongly foliated biotite-muscovite-garnet-bearing schist. Small elongated grains of plagioclase, quartz, and lesser calcite comprise the remainder of the matrix. Inclusion-rich garnet porphyroblasts are anhedral to subhedral and vary from 1 to 3 mm in diameter. The inclusions are

mainly quartz, and define a foliation that is generally consistent with the matrix foliation. The garnet porphyroblasts are surrounded by strain-shadows comprising quartz and fine-grained biotite.

307: Fine-grained biotite schist containing equigranular anhedral biotite, quartz, and lesser muscovite, feldspar, and accessory Fe-Ti oxide.

Daniel Creek Formation

97: Strongly foliated, very fine-grained muscovite schist, also containing quartz, biotite, and lesser calcite and albite. Small porphyroblasts of a prograde metamorphic mineral (possibly staurolite) have been pseudomorphed by chlorite. Muscovite-rich/quartz-poor and muscovite-poor/quartz-rich layering is visible consistent with small-scale changes to the clastic composition of the sedimentary precursor.

184: Low-grade, very fine-grained quartz-muscovite sandstone. Quartz dominates (~80 %) and is surrounded by sericitic muscovite, and accessory zircon. Patches of an opaque mineral (possibly haematite) are scattered throughout, comprising \sim 5 % of the sample.

255: Fine-grained sandstone containing ~80 % quartz surrounded by sericitic muscovite, and accessory zircon.

288: Moderately- to strongly-foliated biotite-muscovite-garnet-bearing schist. Small aligned laths of biotite and muscovite define the dominant layer-parallel schistosity (S_1) . Small elongate quartz constitutes the remainder of the matrix. Garnet porphyroblasts are anhedral to subhedral, and are generally less than 1 mm in diameter. The other porphyroblast-forming phase has been completely pseudomorphed by retrograde chlorite. These porphyroblasts are commonly tabular, anhedral, and range up to 3 mm in diameter. Inclusion trails of quartz in these porphyroblasts commonly define a foliation orthogonal to the matrix foliation.

308: Strongly foliated, fine-grained biotite-muscovite-quartz mica schist. Biotite and muscovite define the layer-parallel schistosity, while quartz and lesser plagioclase comprise the remainder of the matrix. Metamorphic monazite is accessory.

311: Mica-schist containing fine- to coarse-grained biotite, muscovite, fine-grained quartz, small (up to 1 mm diameter) garnet porphyroblasts, and lesser feldspar and chlorite. A second small porphyroblast-forming phase (possibly of an aluminosilicate) has been sericitised.

312: Weakly foliated, garnet-hornblende-bearing calc-silicate gneiss. Peak metamorphic assemblage consists of garnet, hornblende, and fine-grained quartz and plagioclase. Lesser calcite, actinolite, and zoisite also occur. Small elongate hornblende defines a weak layer-parallel foliation, and is the most common form of hornblende in this sample. Garnet porphyroblasts have diameters up to 4 mm, are anhedral to subhedral, and contain large inclusions of quartz. Hornblende porphyroblasts have diameters up to 2 mm, are commonly anhedral, and also contain numerous large quartz inclusions.

Dead Horse Metabasalt

72: Foliated, low-greenschist facies metabasalt containing fine-grained actinolite, albite, chlorite, epidote, and accessory Fe-Ti oxide. A thin (less than 1 mm wide) quartz-calcite vein cuts the foliation. Igneous texture and some relic igneous clinopyroxene are locally preserved.

74: Relatively undeformed greenschist facies metabasalt containing sparse, fine-grained actinolite, albite, chlorite, relic igneous clinopyroxene and lesser plagioclase, and calcite. Accessory Fe-Ti oxide also common.

80: Greenschist facies metabasalt containing weak foliation denoted by alignment of elongate, finegrained feldspar laths. Dominant minerals include chlorite, actinolite, albite, some relic igneous clinopyroxene, and accessory Fe-Ti oxide. Thin section contains one large relic clinopyroxene grain containing alteration rim of chlorite and Fe-Ti oxide.

82: Aphyric, greenschist facies metabasalt containing actinolite, albite, chlorite, epidote, accessory Fe-Ti oxide, and minor relic igneous clinopyroxene. Relic igneous texture is relatively well preserved.

83: Weakly foliated, aphyric, greenschist facies rock containing albite, actinolite, chlorite, epidote, and sparse very fine-grained Fe-Ti oxide. One, possibly two relic clinopyroxene grains are preserved, and outlines of larger replaced clinopyroxene grains are visible.

87: Weakly-foliated greenschist facies metabasalt. Aphyric, but well preserved sample containing chlorite, albite, fibrous actinolite, epidote/zoisite, and accessory Fe-Ti oxide.

89: Weakly foliated, low-greenschist facies metabasalt containing fine-grained albite, actinolite, chlorite, epidote, and accessory Fe-Ti oxide. Epidote has grown proximal to larger actinolite grains.

90: Upper-greenschist facies aphyric metabasalt containing pale-green actinolite, albite, clinozoisite, sphene, and accessory chlorite and Fe-Ti oxide.

100: Weakly-foliated greenschist facies metabasalt containing well preserved igneous texture. Contains actinolite, albite, chlorite, epidote, minor relic clinopyroxene, and accessory Fe-Ti oxide. Rare assemblage of igneous plagioclase and an altered olivine phenocryst is also observed.

102, 103: Fine-grained, weakly-foliated, greenschist facies metabasalts containing well preserved igneous texture. Rare olivine and clinopyroxene phenocrysts are replaced by epidote and chlorite respectively. Dominant phases are chlorite, albite, and actinolite, with lesser relic igneous minerals, epidote, and Fe-Ti oxide.

104: Greenschist facies metabasalt with well preserved igneous texture, including rare, largely replaced relic olivine and clinopyroxene phenocrysts. Mineral assemblage is albite, actinolite, clinopyroxene, olivine, chlorite, epidote, calcite, and lesser Fe-Ti oxide. Medium-grained calcite and chlorite vein cuts sample normal to section orientation.

107: Weakly foliated, fine-grained, lower-greenschist facies metabasalt containing medium-grained quartz, actinolite, chlorite, albite, calcite, and lesser relic clinopyroxene, and Fe-Ti oxide.

111: Low- to mid-greenschist facies metabasalt containing fine-grained actinolite, albite, quartz, chlorite, and minor epidote and Fe-Ti oxide. Relic igneous clinopyroxene and plagioclase form sub-ophitic texture.

148: Fine-grained greenschist facies metabasalt with near-aphyric texture containing fibrous actinolite and albite, and very fine-grained chlorite, quartz, and minor Fe-Ti oxide.

151, 153, 154: Aphyric greenschist facies metabasalts containing fine-grained actinolite, albite, chlorite, and relic clinopyroxene.

228: Aphyric greenschist facies metabasalt containing fine-grained actinolite and albite. Accessory minerals are biotite, quartz, and minor Fe-Ti oxide.

229, 230, 236, 240: Aphyric greenschist facies metabasalts containing fine-grained actinolite and albite. Accessory phases include quartz and minor chalcopyrite. Hematite and ilmenite (with possible limonite) form concentric growth bands around chalcopyrite grains.

232, 243: Coarser-grained aphyric metabasalts containing actinolite, albite, relic igneous clinopyroxene, and accessory chalcopyrite.

233: Well preserved, quenched pillow basalt metamorphosed to greenschist facies containing finegrained actinolite, chlorite, and albite.

234: Fine-grained greenschist facies metabasalt, originally forming basalt pillow rim. Relic igneous plagioclase and olivine phenocrysts are almost completely replaced by quartz, calcite, and chlorite. Groundmass altered to anhedral chlorite and very fine-grained actinolite.

250, **251**: Fine-grained, greenschist facies sub-ophitic metabasalts containing actinolite, albite, quartz, minor calcite and Fe-Ti oxide, and relic clinopyroxene and plagioclase altered to chlorite and epidote.

253: Aphyric greenschist facies metabasalt containing fine-grained actinolite, albite, chlorite, calcite, quartz, and accessory Fe-Ti oxide.

256: Aphyric lower-greenschist facies metabasalt containing fine-grained calcite, chlorite, albite, and accessory Fe-Ti oxide.

260, 269: Thoroughly recrystallised medium-grained, foliated amphibolite containing elongate hornblende in an albite-quartz groundmass.

261: Weakly foliated, fine-grained amphibolite containing hornblende, quartz, albite, and minor Fe-Ti oxide. Accessory chalcopyrite without Fe-oxide alteration rim also identified.

262: Medium- to coarse-grained amphibolite containing hornblende, albite, quartz, and Fe-Ti oxide.

264: Foliated, medium-grained amphibolite containing hornblende, albite, quartz, and accessory Fe-Ti oxide. Sample is cut by thin (10 mm thick) felsic vein of the same mineralogy as the host rock, however coarser-grained albite and quartz dominate.

268: Fine-grained, weakly-foliated amphibolite. Hornblende is the dominant phase, with minor phases including quartz, Fe-Ti oxide, and albite.

270: Strongly foliated amphibolite containing fine-grained hornblende, quartz, albite, and minor Fe-Ti oxide.

Corbett Formation

278B: Weakly foliated phyllite containing very fine-grained quartz, muscovite, lesser biotite, and accessory Fe-Ti oxide. Alternating muscovite-rich/quartz-poor and muscovite-poor/quartz-rich layers are common.

306: Foliated muscovite-biotite-schist collected near Robinhood, containing phenocrysts of staurolite after garnet. Small aligned laths of biotite and muscovite define the dominant S_3 foliation. Both garnet and staurolite porphyroblasts are inclusion-rich and variable in size; however staurolite is generally larger (2 to 5 mm in diameter), and partially encloses some garnet grains. Garnet porphyroblasts are small (up to 1 mm in diameter), anhedral to subhedral, and contain small inclusion trails of quartz that with staurolite define an earlier (S_1) foliation that is generally orthogonal to the external schistosity.

Lane Creek Formation

245: Chloritoid-bearing meta-siltstone containing very fine-grained quartz and fibrous muscovite. Lesser biotite and accessory Fe-Ti oxide are also present within the matrix.

304: Fine- to medium-grained mica-schist containing foliated, elongate muscovite, fine-grained blocky quartz, and lesser biotite. Large porphyroblasts of a prograde metamorphic mineral have been pseudomorphed by blocky quartz, muscovite and chlorite.

Cobbold Metadolerite

6: Mid-greenschist facies metadolerite containing actinolite, interstitial quartz, albite, epidote, sericitic muscovite, and accessory zircon. Sample is cut by late-stage tonalitic vein (10-20 mm wide) containing anhedral quartz, albite, and lesser actinolite and epidote.

8: Unusual syenite/quartz-free diorite containing plagioclase, K-feldspar, calcite, interstitial chloriteactinolite, sphene, Fe-Ti oxide, and accessory zircon.

22: Low-greenschist facies metadolerite containing medium- to fine-grained actinolite, epidote, interstitial quartz, and lesser albite, and sericite.

23: Low- to mid-greenschist facies metadolerite containing abundant blue-green hornblende in addition to albite, epidote, interstitial quartz, and lesser sericitic muscovite.

60: Mid-greenschist facies metadolerite containing large (cm-size) actinolite, albite, and fine-grained epidote, chlorite, and Fe-Ti oxide.

85, 86: High-greenschist facies metadolerites containing medium- to coarse-grained hornblende, plagioclase-albite, and fine-grained epidote, quartz, and altered Fe-Ti oxide.

88: Fine-grained, weakly foliated metadolerite containing actinolite, albite, Fe-Ti oxide, and lesser epidote and chlorite.

101: Coarse-grained, well-preserved greenschist facies metadolerite containing blue-green actinolite, albite, and lesser epidote, chlorite, and Fe-Ti oxide. Several small grains of igneous clinopyroxene are also visible.

106: Low- to mid-greenschist facies metadolerite containing medium-grained brown hornblende rimmed by fibrous green actinolite, albite, quartz and apatite needles, and accessory Fe-Ti oxide.

108, 109, 110, 112: Coarse-grained greenschist facies metadolerites containing blue-green actinolite, albite, interstitial quartz, epidote, chlorite, and large altered Fe-Ti oxide.

113, 114, 115: Largely ophitic, coarse-grained low-greenschist facies metadolerites containing rare altered olivine, actinolite, albite, chlorite, epidote, interstitial quartz, and accessory Fe-Ti oxide.

120, 121: Low-greenschist facies metadolerites with coarse-grained actinolite overprinting clinopyroxene, albite, epidote, chlorite, and accessory Fe-Ti oxide. Preserved igneous texture is near-ophitic in places.

128, 129: Greenschist facies, well-preserved metadolerites containing actinolite after clinopyroxene, albite, interstitial quartz, and lesser epidote.

130: Greenschist facies metadolerite with pegmatoidal segregation showing myrmekitic texture. Sample contains fine- to medium-grained actinolite, albite, and lesser quartz, epidote, and Fe-Ti oxide.

152: Medium- to fine-grained greenschist facies metadolerite containing actinolite after clinopyroxene, albite, chlorite, interstitial quartz, and minor Fe-Ti oxide.

159, 161: Fine- to medium-grained greenschist facies metadolerites containing minor felsic segregations showing myrmekitic texture. Samples contain actinolite, albite, quartz, chlorite, lesser Fe-Ti oxide and igneous clinopyroxene.

172: Fine- to medium-grained greenschist facies metadolerite containing late-stage felsic differentiate with quartz, albite, chlorite, epidote, and accessory zircon. Mafic section contains medium-grained actinolite, chlorite, albite, and quartz.

174: Fine- to medium-grained granophyric sample showing myrmekitic texture. Sample contains albite, quartz, calcite, chlorite, and lesser biotite.

175: Fine- to medium-grained greenschist facies metadolerite containing actinolite, albite, quartz, and lesser chlorite and sericite.

178: Fine- to medium-grained greenschist facies metadolerite from CMD outcrop containing spheroidal clinozoisite-epidote xenoliths. Doleritic section contains actinolite, albite, and lesser igneous clinopyroxene.

179: Greenschist facies metadolerite containing actinolite after clinopyroxene, albite with sericitic alteration, chlorite, quartz, and lesser Fe-Ti oxide.

181, 188: Coarse-grained greenschist facies metadolerites containing actinolite after clinopyroxene, Fe-Ti oxide, and finer-grained albite, quartz and chlorite.

192, 194, 195, 196, 197, 199, 200: Mid- to high-greenschist facies metadolerites containing mediumgrained actinolite, albite, quartz, and lesser calcite and Fe-Ti oxide. Fe-Ti oxide in sample 195 is mostly altered to sphene.

201: High-greenschist facies metadolerite, transitional to amphibolite. Sample contains fine-grained actinolite, albite, quartz, and lesser Fe-Ti oxide.

202: Very fine-grained feldspar-poor greenschist facies metadolerite from sill margin. Sample contains actinolite, albite, quartz, and accessory Fe-Ti oxide.

215: Amphibolite facies metadolerite containing clear, euhedral, medium-grained hornblende, albite with sericitic alteration, quartz, and accessory Fe-Ti oxide.

221, 222: Greenschist facies metadolerites containing coarse-grained actinolite, albite, chlorite, quartz, and accessory Fe-Ti oxide altered to sphene.

224, 226: Fine-grained greenschist facies metadolerites containing actinolite, albite, chlorite, quartz, and minor Fe-Ti oxide.

227: Fine-grained greenschist facies metadolerite or coarse metabasalt containing actinolite, albite, chlorite, and accessory Fe-Ti oxide.

237: Blocky holocrystalline greenschist facies metabasalt or fine-grained metadolerite containing actinolite, albite, sericite, and accessory Fe-Ti oxide.

241: Fine-grained greenschist facies metadolerite containing actinolite after clinopyroxene, albite, chlorite, calcite, minor igneous plagioclase, and accessory haematite surrounding chalcopyrite.

242, **249**: Fine-grained greenschist facies metadolerites containing actinolite after clinopyroxene, albite, chlorite, calcite, and accessory Fe-Ti oxide and chalcopyrite.

291: Texturally well-preserved greenschist facies metadolerite containing fine-grained actinolite after clinopyroxene, albite, some igneous plagioclase, sericite, and accessory Fe-Ti oxide.

296: Greenschist facies metadolerite containing abundant, subhedral to anhedral actinolite after clinopyroxene, albite, and accessory Fe-Ti oxide.

299, 300: Strongly recrystallised amphibolite facies metadolerites containing blue-green hornblende, albite, quartz, lesser sericite, and small clusters of Fe-Ti oxide.

301: Fine-grained sub-ophitic strongly recrystallised amphibolite facies metadolerite containing hornblende after clinopyroxene, albite, quartz, lesser sericite, and accessory Fe-Ti oxide.

Ropewalk Granite

208, **209**, **210**: Medium- to coarse-grained biotite-muscovite granites. Biotite is the dominant mica, and accessory phases include quartz, sphene (absent in sample 208), alkali feldspar (accessory in sample 208), lesser plagioclase, and accessory zircon. All samples are weakly foliated; however the foliation is most intense in sample 210.

Mount Hogan Granite

135: Unfoliated medium- to coarse-grained granite containing quartz, microcline, fine-grained muscovite, and accessory zircon.

APPENDIX 2:

Zircon LA-ICPMS and EMPA Monazite data

Table 1. Zircon U-Pb LA-ICPMS data for mafic samples from the lower Etheridge Group.

Sample	/analysis	U ppm	Th/U	206Pb/238U	±	207Pb/235U	±	207Pb/206Pb	±	207Pb/206Pb age (Ma)	±
6	1	1380	0.55	0.2908	0.0018	4.027	0.056	0.1008	0.0014	1639	26
	2	786	0.45	0.3004	0.0021	4.158	0.067	0.1013	0.0017	1649	31
	3	592	0.45	0.2849	0.0017	3.956	0.061	0.1009	0.0015	1641	27
	4	397	0.39	0.2968	0.0025	4.563	0.091	0.1124	0.0022	1838	35
	5	763	0.50	0.2907	0.0037	4.067	0.145	0.1050	0.0036	1714	64
	6	1160	0.51	0.2860	0.0019	3.901	0.059	0.0994	0.0015	1613	27
	7	481	0.44	0.2849	0.0043	4.130	0.164	0.1105	0.0043	1807	72
	8	749	0.46	0.2848	0.0020	4.184	0.079	0.1068	0.0019	1745	32
	9	393	0.39	0.2834	0.0024	4.045	0.087	0.1046	0.0023	1706	41
	10	511	0.30	0.2638	0.0019	4.086	0.078	0.1131	0.0021	1850	33
	11	1242	0.52	0.2953	0.0019	4.074	0.058	0.1006	0.0015	1635	28
	12	1440	0.59	0.2877	0.0050	3,892	0.108	0.1023	0.0024	1666	44
8	1	907	0.52	0.2855	0.0056	3.831	0.140	0.0998	0.0032	1620	59
	2	1216	0.46	0.2880	0.0046	4.136	0.149	0.1014	0.0033	1650	61
	3	1615	0.52	0.2881	0.0025	3.991	0.059	0.1014	0.0014	1650	25
	4	1485	0.64	0.2838	0.0026	3,917	0.061	0.1006	0.0014	1635	26
	5	321	0.28	0.2889	0.0035	3.859	0.090	0.0986	0.0023	1598	43
	6	983	0.46	0.2702	0.0036	3.806	0.070	0.1025	0.0017	1670	31
	7	393	0.15	0.3542	0.0046	12.094	0.394	0.2467	0.0068	3164	44
	8	1146	0.57	0.2602	0.0091	3 583	0 173	0 1034	0.0046	1686	82
	9	1308	0.52	0.2879	0.0025	3 941	0.057	0.0999	0.0014	1623	25
	10	1233	0.51	0.2846	0.0028	3 901	0.059	0 1008	0.0015	1638	28
	11	831	0.43	0.2528	0.0026	3 505	0.057	0 1015	0.0015	1651	28
	12	1312	0.50	0.2011	0.0026	4 071	0.061	0 1016	0.0015	1654	27
	12	1012	0.00	0.2011	0.0020	4.071	0.001	0.1010	0.0010	1004	21
159	1	1919	0.96	0.3436	0.0271	13,946	2.554	0.2823	0.0363	3376	201
	2	763	1 09	0.3109	0.0083	4 488	0 179	0 1024	0.0033	1668	60
	3	1353	1 40	0.2880	0.0052	4 391	0.136	0 1067	0.0030	1743	52
	4	956	1.05	0.3177	0.0119	4 816	0.267	0 1048	0.0049	1711	86
	5	1187	0.74	0.2624	0.0060	4 321	0.189	0 1171	0.0047	1913	72
	6	1102	0.74	0.3025	0.0087	4 506	0 192	0 1043	0.0034	1701	60
	7	1512	1 19	0.2798	0.0074	4 349	0.266	0 1104	0.0056	1806	92
	8	1257	0.99	0.2669	0.0054	4 289	0.160	0 1127	0.0039	1844	63
	9	791	0.99	0.3037	0.0067	4 456	0 161	0 1037	0.0031	1692	56
	10	852	0.85	0 2945	0.0057	4 524	0 151	0 1068	0.0031	1745	53
	11	624	0.85	0.2978	0.0056	4 492	0 152	0.1060	0.0033	1731	57
	12	741	0.00	0.2816	0.0067	4 511	0.102	0.1000	0.0037	1847	59
	13	2025	1 33	0.2669	0.0046	3.834	0.114	0.1082	0.0025	1720	62
	14	907	1.03	0.2003	0.0040	1 A 77	0.111	0.1002	0.0023	1675	63
	15	1169	0.02	0.3255	0.0027	3 0.08	0.077	0.1071	0.0010	1716	00
	16	1017	1.04	0.2000	0.0032	3 975	0.003	0.1035	0.0013	1695	65
	17	1468	0.94	0.2743	0.0027	3 821	0.007	0.1013	0.0036	1303	148
	18	433	0.34	0.3185	0.0079	4 351	0.096	0.1066	0.0023	1673	134
	10	1200	0.74	0.0100	0.0023	4.001	0.050	0.1000	0.0023	1766	50
	19	1290	0.76	0.2949	0.0023	4.071	0.074	0.1071	0.0019	1693	5Z 99
	20	722	0.07	0.2022	0.0036	3.020	0.003	0.1011	0.0022	1561	121
	21	1033	0.00	0.2940	0.0030	3,901	0.104	0.1022	0.0027	1705	121
	22	1933	1.01	0.2933	0.0021	3.093	0.059	0.1029	0.0014	1/03	42 177
	23	2115	0.89	0.2270	0.0003	3.219	0.103	0.1120	0.0000	1407	177
1	24	1965	0.77	0.2817	0.0071	4.170	0.257	0.1198	0.0066	1703	329

Sample/	/analysis	U ppm	Th/U	206Pb/238U	±	207Pb/235U	±	207Pb/206Pb	±	207Pb/206Pb age (Ma)	±
161	1	928	1.31	0.3021	0.0021	4.532	0.115	0.1043	0.0024	1702	43
	2	975	1.21	0.2944	0.0016	4.455	0.117	0.1051	0.0026	1717	45
	3	1178	1.39	0.3057	0.0020	4.602	0.116	0.1047	0.0025	1710	44
	4	721	0.81	0.3166	0.0023	5.170	0.150	0.1136	0.0031	1858	49
	5	728	0.85	0.2955	0.0058	4.821	0.304	0.1066	0.0061	1742	106
	6	867	1.11	0.2972	0.0036	4.477	0.221	0.1108	0.0054	1813	89
	7	691	0.99	0.2960	0.0021	4.490	0.118	0.1061	0.0029	1733	51
	8	934	1.10	0.2956	0.0022	4.259	0.122	0.1023	0.0030	1666	54
	9	1537	0.62	0.2720	0.0040	4.266	0.204	0.1014	0.0050	1650	91
	10	832	0.96	0.2960	0.0022	4.320	0.112	0.1015	0.0026	1651	47
	11	594	0.99	0.3128	0.0033	4.688	0.136	0.1041	0.0031	1698	55
	12	688	0.85	0.3113	0.0075	5.742	0.342	0.1202	0.0065	1960	97
	13	1071	1.09	0.3030	0.0037	4.048	0.088	0.1063	0.0024	1729	86
	14	532	0.82	0.2979	0.0047	4.070	0.109	0.1099	0.0032	1620	157
	15	1070	0.67	0.2864	0.0038	4.097	0.109	0.1104	0.0029	1805	85
	16	1346	1.10	0.2845	0.0025	3.748	0.073	0.1042	0.0019	1741	68
	17	779	1.05	0.3087	0.0048	5.066	0.185	0.1231	0.0039	1722	150
	18	944	0.98	0.3028	0.0056	4.223	0.122	0.1064	0.0033	1788	102
	19	516	0.98	0.3111	0.0066	5.033	0.230	0.1298	0.0048	1923	189
	20	846	1.04	0.2839	0.0038	4.071	0.143	0.1091	0.0034	1620	137
	21	1396	1.27	0.2956	0.0022	3.933	0.064	0.1030	0.0017	1691	73
	22	1899	1.04	0.2766	0.0032	3.696	0.094	0.1067	0.0027	1680	64
	23	1120	1.08	0.3005	0.0027	4.148	0.093	0.1082	0.0025	1640	83
	24	1509	1.19	0.3007	0.0039	4.704	0.144	0.1223	0.0038	1726	86
	25	798	1.18	0.3043	0.0027	5.761	0.124	0.1398	0.0029	1622	136
	26	1164	1.03	0.2921	0.0029	4.258	0.108	0.1108	0.0028	1723	84
	27	823	1.34	0.3099	0.0059	4.710	0.178	0.1178	0.0040	1740	185
	28	1003	1.10	0.2961	0.0029	3.977	0.101	0.1036	0.0025	1534	101
	29	931	1.02	0.2998	0.0024	4.368	0.090	0.1099	0.0022	1685	91
	30	972	1.08	0.3036	0.0042	4.257	0.123	0.1081	0.0029	1707	91
172*	1	885	1.00	0.2894	0.0019	4.320	0.110	0.1036	0.0027	1633	12
	2	556	0.85	0.4308	0.0064	21.936	1.196	0.3683	0.0193	1652	76
	3	729	0.75	0.2793	0.0020	4.182	0.108	0.1034	0.0027	1578	12
	4	1035	1.17	0.2722	0.0029	4.227	0.154	0.1043	0.0036	1537	17
	5	610	0.77	0.2988	0.0031	5.378	0.147	0.1246	0.0034	1643	18
	6	326	0.82	0.2892	0.0037	4.839	0.252	0.1113	0.0058	1617	23
	7	341	0.95	0.2911	0.0036	4.791	0.201	0.1108	0.0046	1628	22
	8	369	0.99	0.2931	0.0030	4.991	0.161	0.1179	0.0037	1625	18
	9	869	1.15	0.2777	0.0029	4.721	0.178	0.1134	0.0045	1551	18
	10	871	1.17	0.2955	0.0075	6.474	0.419	0.1406	0.0078	1594	42
	11	1073	0.82	0.2400	0.0046	5.342	0.326	0.1449	0.0081	1296	28
	12	789	0.94	0.2900	0.0024	4.616	0.135	0.1109	0.0032	1623	14
264	1	567	0.55	0.2849	0.0057	4.081	0.173	0.1011	0.0041	1645	74
	2	425	0.39	0.2809	0.0044	4.031	0.186	0.1035	0.0040	1688	72
	3	1033	0.61	0.2886	0.0063	4.380	0.171	0.1042	0.0040	1701	70
	4	310	1.03	0.2765	0.0040	3.900	0.141	0.0976	0.0032	1579	61
	5	543	0.69	0.2859	0.0022	4.545	0.101	0.1117	0.0022	1828	36
	6	689	0.81	0.2779	0.0024	4.315	0.121	0.1107	0.0029	1810	48
. 207	7	1001	1.06	0.2683	0.0027	3.956	0.126	0.1024	0.0027	1668	48
* - Pb corr	ected ""Pb	/~~~U used									

Sample/analysis	U ppm	Th/U	206Pb/238U	±	207Pb/235U	±	207Pb/206Pb	±	207Pb/206Pb age (Ma)	±
264 8	942	0.68	0.2617	0.0030	3.670	0.101	0.0978	0.0025	1583	48
9	576	1.06	0 2726	0.0032	4 217	0 128	0 1077	0.0033	1760	56
10	1076	0.88	0.2864	0.0030	4 232	0.094	0 1035	0.0021	1687	37
11	1336	0.50	0.2004	0.0035	4 345	0.004	0.1000	0.0021	1625	38
11	200	0.50	0.0000	0.0000	4.660	0.104	0.1001	0.0021	1702	50
12	309	0.57	0.3095	0.0047	4.003	0.179	0.1046	0.0033	1707	59
13"	80	1.01	0.4123	0.0065	8.212	0.286	0.1409	0.0048	2238	58
14*	392	0.63	0.3797	0.0053	8.886	0.204	0.1643	0.0034	2500	35
15*	404	0.98	0.4170	0.0137	10.525	0.422	0.1781	0.0045	2636	42
16*	281	0.52	0.4604	0.0103	10.919	0.372	0.1654	0.0046	2511	47
17*	471	0.46	0.2988	0.0028	4.596	0.104	0.1158	0.0026	1662	16
18*	525	0.27	0.4619	0.0049	11.162	0.214	0.1823	0.0032	2373	29
19*	183	0.48	0.4989	0.0051	12.168	0.236	0.1832	0.0035	2580	33
20*	622	0.73	0.2929	0.0044	4.566	0.129	0.1176	0.0032	1627	25
21*	522	0.55	0.3030	0.0047	4.551	0.175	0.1148	0.0043	1687	27
22*	545	0.85	0.2842	0.0025	4.381	0.090	0.1171	0.0026	1581	14
23*	887	0.83	0.2889	0.0060	4.160	0.158	0.1108	0.0042	1618	34
24*	270	1.70	0.3488	0.0034	5.657	0.119	0.1228	0.0026	1918	19
25*	577	0.52	0 2985	0.0059	4 486	0.178	0 1111	0.0043	1669	33
26*	1138	1.04	0.2000	0.0040	6 112	0 179	0 1/107	0.0035	1621	22
20	927	0.30	0.2031	0.0040	4 106	0.113	0.1064	0.0030	1621	22
21	1067	0.39	0.2931	0.0037	4.100	0.113	0.1004	0.0030	1645	21
20	1007	0.90	0.2900	0.0035	4.410	0.121	0.1155	0.0031	1015	20
29-	603	0.82	0.3035	0.0032	4.540	0.094	0.1118	0.0025	1695	18
30"	759	0.42	0.2881	0.0040	4.163	0.130	0.1068	0.0033	1620	23
31*	280	0.43	0.4796	0.0047	11.153	0.201	0.1739	0.0031	2500	29
32*	743	0.78	0.2885	0.0030	4.019	0.085	0.1044	0.0020	1627	17
33*	943	0.94	0.2882	0.0037	4.177	0.079	0.1085	0.0020	1618	21
34*	1011	0.33	0.2987	0.0027	4.199	0.081	0.1061	0.0021	1680	15
35*	231	1.66	0.3428	0.0035	5.583	0.115	0.1209	0.0028	1890	20
299* 1	243	0.51	0.3147	0.0122	4.782	0.523	0.1057	0.0123	1768	71
2	281	0.52	0.3168	0.0181	4.747	0.753	0.1016	0.0165	1787	105
3	326	0.69	0.3172	0.0126	4.729	0.485	0.1086	0.0109	1776	72
4	305	0.56	0.2927	0.0129	4.175	0.469	0.1049	0.0116	1649	73
5	258	0.72	0.2945	0.0071	4.329	0.279	0.1022	0.0069	1664	41
6	368	0.65	0.2952	0.0085	4.416	0.361	0.1095	0.0090	1655	49
7	285	0.67	0 2604	0.0138	3 699	0.467	0 1018	0.0125	1478	78
8	180	0.31	0.2878	0.0118	4 526	0 534	0 1172	0.0134	1601	68
۵ ۵	132	0.50	0 3163	0.0137	4 302	0.556	0.0965	0.0130	1795	80
10	340	0.00	0.2804	0.0000	3 808	0.307	0.0000	0.0108	1594	57
10	540	0.75	0.2004	0.0039	4 295	0.397	0.1040	0.0066	1675	11
· · ·	050	0.94	0.2900	0.0072	4.200	0.290	0.1022	0.0000	1075	41
200* 4	190	0.51	0.2059	0.0001	2 001	0.255	0 1002	0.0000	1675	E2
JUU" 1	189	0.51	0.2958	0.0091	3.991	0.355	0.1003	0.0090	10/5	53
2	340	0.48	0.3003	0.0080	4.348	0.330	0.1054	0.0079	1690	46
3	439	0.81	0.2829	0.0072	4.095	0.297	0.1043	0.0075	1596	41
4	481	0.59	0.2476	0.0081	3.377	0.271	0.0989	0.0075	1412	46
5	369	0.54	0.2454	0.0106	3.191	0.304	0.1016	0.0092	1396	60
6	233	0.58	0.2915	0.0081	4.388	0.395	0.1046	0.0089	1643	47
7	164	0.51	0.2863	0.0106	3.986	0.438	0.1086	0.0119	1607	61
8	262	0.47	0.3050	0.0092	3.901	0.315	0.0975	0.0078	1731	53
9	488	0.72	0.2970	0.0071	4.488	0.300	0.1087	0.0073	1665	41

* 207Pb corrected 206Pb/238U used

Sample/a	analysis	U ppm	Th/U	206Pb/238U	±	207Pb/235U	±	207Pb/206Pb	±	207Pb/206Pb age (Ma)	±
300*	10	484	0.66	0.3016	0.0073	4.104	0.297	0.0984	0.0069	1710	43
	11	759	0.88	0.2812	0.0205	4.010	0.722	0.1191	0.0214	1561	116
	12	300	0.53 0.2429 0.0		0.0073	3.243	0.276	0.0974	0.0079	1389	42
318*	1	1335	0.06	0.2622	0.0160	3.523	0.601	0.1037	0.0160	1485	91
	2	984	0.52	0.3172	0.0103	4.211	0.317	0.0980	0.0074	1797	59
	3	782	0.14	0.3011	0.0156	3.723	0.552	0.0960	0.0143	1712	90
	4 504		0.01	0.2295	0.0085	3.068	0.281	0.0998	0.0098	1312	49
	5	1103	0.06	0.2836	0.0115	3.786	0.399	0.1026	0.0109	1604	66
	6	995	0.03	0.2877	0.0101	3.653	0.332	0.0970	0.0087	1636	57
	7	1673	0.01	0.2588	0.0140	3.508	0.355	0.1024	0.0105	1469	78
	8	1347	0.00	0.2294	0.0118	3.260	0.417	0.1055	0.0140	1304	67
	9	1453	0.00	0.2777	0.0121	3.424	0.414	0.0962	0.0120	1582	70
	10	1441	0.00	0.2455	0.0080	3.270	0.278	0.1016	0.0088	1397	46
	11	1296	0.00	0.2880	0.0083	3.727	0.332	0.0977	0.0089	1636	48
	12	770	0.86	0.2915	0.0157	3.875	0.559	0.1052	0.0158	1643	90

* ²⁰⁷Pb corrected ²⁰⁶Pb/²³⁸U used

Sample/a	analysis	U ppm	Th/U	206Pb/238U	±	207Pb/235U	±	207Pb/206Pb	±	207Pb/206Pb age (Ma)	±
135	1	121	0.60	0.3084	0.0074	4.404	0.174	0.1013	0.0039	1648	71
	2	442	0.59	0.1432	0.0023	2.199	0.072	0.1099	0.0034	1799	56
	3	594	0.92	0.2548	0.0065	3.692	0.149	0.1021	0.0037	1662	67
	4	635	1.86	0.2345	0.0058	3.379	0.183	0.1002	0.0048	1627	89
	5	618	0.69	0.2420	0.0058	3.948	0.178	0.1128	0.0049	1844	79
	6	451	0.51	0.2599	0.0105	3.483	0.232	0.0976	0.0049	1579	94
	7	584	0.68	0.2769	0.0046	3.689	0.103	0.0943	0.0023	1514	47
	8	849	1.29	0.2531	0.0150	3.823	0.332	0.1042	0.0061	1700	107
	9	1575	0.04	0.2826	0.0054	3.797	0.103	0.0953	0.0024	1534	48
	10	698	0.38	0.2110	0.0030	3.411	0.095	0.1117	0.0028	1827	45
	11	429	1.40	0.2597	0.0032	3.557	0.094	0.0965	0.0025	1557	49
	12	1068	2.92	0.1487	0.0034	3.251	0.104	0.1560	0.0044	2413	48
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208	1	513	0.35	0.2834	0.0017	3.879	0.078	0.0962	0.0019	1552	37
	2	696	0.56	0.2522	0.0016	3.481	0.078	0.0968	0.0022	1564	42
	3	331	0.36	0.2787	0.0020	3.911	0.087	0.0973	0.0019	1573	36
	4	734	0.56	0.2543	0.0028	3.665	0.109	0.0997	0.0028	1619	53
	5	550	0.45	0.2798	0.0017	3.895	0.082	0.0983	0.0020	1592	38
	6	298	0.51	0.2306	0.0029	3.359	0.142	0.1014	0.0038	1650	69
	7	693	0.40	0.2676	0.0032	3.730	0.160	0.0958	0.0041	1544	80
	8	995	0.96	0.2285	0.0051	3.358	0.211	0.1017	0.0048	1656	88
	9	876	0.41	0.2350	0.0020	3.346	0.066	0.1003	0.0018	1630	33
	10	711	0.40	0.2581	0.0046	3.678	0.141	0.0981	0.0037	1589	71
	11	469	0.46	0.2799	0.0019	3,785	0.088	0.0950	0.0021	1529	41
	12	218	0.74	0.2785	0.0029	3.732	0.099	0.0950	0.0027	1528	53
209	1	854	0.37	0.2567	0.0016	3.482	0.074	0.0952	0.0020	1532	39
	2	796	0.90	0.2684	0.0017	3.635	0.072	0.0956	0.0018	1539	35
	3	431	1.12	0.2779	0.0019	3.701	0.078	0.0940	0.0019	1508	38
	4	717	1.22	0.2637	0.0016	3.487	0.072	0.0936	0.0018	1501	36
	5	721	1.17	0.2699	0.0017	3.627	0.070	0.0948	0.0017	1523	34
	6	1509	1.70	0.2670	0.0014	3.582	0.059	0.0945	0.0015	1517	29
	7	530	1.05	0.2503	0.0019	3.390	0.090	0.0957	0.0025	1541	49
	8	265	0.89	0.2683	0.0023	3.573	0.092	0.0941	0.0024	1510	47
	9	506	0.37	0.3465	0.0023	5,146	0.108	0.1043	0.0020	1702	36
	10	1063	1.51	0.2736	0.0028	3.830	0.126	0.0963	0.0029	1554	57
	11	351	1.53	0.2630	0.0038	3,584	0.159	0.0972	0.0045	1571	88
	12	399	1.02	0.2674	0.0048	3.741	0.179	0.0995	0.0044	1614	82
210	1	644	0.97	0.2801	0.0029	3.941	0.108	0.0987	0.0026	1600	50
	2	260	0.70	0.2751	0.0025	3.663	0.094	0.0937	0.0023	1502	47
	3	275	1.15	0.2861	0.0020	4.106	0.097	0.1010	0.0023	1643	42
	4	390	0.67	0.2772	0.0023	3.687	0.084	0.0932	0.0020	1493	41
	5	285	2.12	0.2683	0.0022	3.628	0.099	0.0942	0.0024	1512	48
	6	309	0.46	0.2818	0.0023	3.729	0.092	0.0940	0.0023	1507	47
	7	489	0.20	0.2652	0.0020	3.531	0.087	0.0934	0.0021	1495	43
	8	531	0.79	0.2673	0.0029	3.780	0.099	0.0991	0.0023	1607	43
	9	530	0.80	0.2767	0.0019	3.814	0.086	0.0963	0.0021	1554	41
	10	314	2.14	0.2713	0.0024	3.760	0.105	0.0969	0.0026	1565	50
	11	531	0.76	0.2688	0.0022	3.645	0.092	0.0948	0.0022	1524	43
	12	195	1.04	0.2711	0.0029	3.638	0.107	0.0938	0.0027	1505	55

Table 2. Zircon U-Pb LA-ICPMS data for samples of the Mount Hogan Granite (135) and Ropewalk Granite (208, 209, 210).

Table 3. Zircon U-Pb LA-ICPMS data for metasediments of the Daniel Creek Formation, lower Etheridge Grou	ıp.
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Sample/analysis	U ppm	Th/U	206Pb/238U	±	207Pb/235U	±	207Pb/206Pb	±	207Pb/206Pb age (Ma)	±
184 1	1724	1.19	0.1820	0.0038	6.322	0.161	0.2613	0.0037	3085	38
2	764	0.87	0.3882	0.0062	10.542	0.350	0.2097	0.0069	2730	89
3	231	0.81	0.5195	0.0056	12.773	0.252	0.1864	0.0037	2711	33
4	800	0.35	0.3608	0.0041	7.771	0.182	0.1643	0.0038	2438	72
5	552	1.21	0.3283	0.0091	7.968	0.355	0.1887	0.0077	2308	229
6	290	0.41	0.4450	0.0068	9.918	0.293	0.1727	0.0049	2395	165
7	476	0.28	0.4688	0.0060	10.723	0.280	0.1751	0.0046	2607	43
8	277	1.01	0.5155	0.0118	14.036	0.557	0.1978	0.0073	2506	175
9	824	0.47	0.3726	0.0071	9.894	0.440	0.2012	0.0093	2743	95
10	4015	0.30	0.0993	0.0035	3.229	0.173	0.2369	0.0100	2532	215
11	716	0.75	0.4473	0.0059	10.503	0.329	0.1800	0.0054	2468	93
12	186	0.69	0.4793	0.0052	10.735	0.258	0.1701	0.0040	2558	40
13	1259	0.59	0.2485	0.0022	5.022	0.087	0.1508	0.0025	1987	69
14	175	0.32	0.4869	0.0048	11.173	0.264	0.1735	0.0036	2592	34
15	1502	1.13	0.2194	0.0038	6.103	0.152	0.2070	0.0031	2485	61
16	1085	0.86	0.2724	0.0061	6.958	0.205	0.1913	0.0042	2589	95
17	238	0.25	0.5134	0.0049	12.367	0.262	0.1816	0.0036	2668	33
18	1279	1.31	0.2106	0.0044	5.972	0.152	0.2117	0.0039	2525	70
19	828	1.13	0.3567	0.0067	9.616	0.296	0.2000	0.0065	2452	110
20	263	1.32	0.3499	0.0086	5.611	0.253	0.1238	0.0050	2012	72
21	385	0.72	0.4439	0.0117	10.327	0.423	0.1795	0.0084	2593	170
22	1689	1.07	0.1917	0.0020	5.025	0.092	0.1959	0.0034	1985	92
23	2233	0.78	0.1257	0.0040	3.141	0.104	0.1901	0.0046	1841	146
24	440	0.16	0.4746	0.0040	10.267	0.175	0.1624	0.0028	2481	29
25	125	0.38	0.4751	0.0049	10.322	0.225	0.1609	0.0036	2465	38
26	512	0.48	0.3219	0.0089	7.817	0.241	0.1843	0.0035	2556	101
27	934	1.24	0.2643	0.0040	7.592	0.163	0.2124	0.0031	2645	49
28	874	0.84	0.2254	0.0103	5.089	0.297	0.1640	0.0080	2497	83
29	3530	0.58	0.0756	0.0015	1.718	0.071	0.1649	0.0071	2507	72
30	504	0.65	0.2640	0.0058	4,729	0.200	0.1301	0.0075	2100	101
31	918	0.72	0 2011	0.0044	4 363	0 184	0 1576	0.0073	2430	79
32	271	0.64	0.4641	0.0085	14,263	0.578	0.2232	0.0094	3004	68
33	1215	1.69	0.1405	0.0043	3.175	0.151	0.1641	0.0087	2499	89
34	199	0.39	0.3317	0.0056	5.139	0.205	0.1125	0.0098	1841	157
35	612	0.56	0.3178	0.0052	7 289	0.289	0 1666	0.0069	2523	69
36	856	1 25	0.2261	0.0081	4 851	0.247	0 1558	0.0104	2411	113
37	80	0.47	0.5008	0.0220	13 723	0.781	0 1990	0.0164	2818	134
38	149	1 45	0.5256	0.0114	13 496	0.568	0 1865	0.0109	2711	97
39	1110	0.96	0 1919	0.0060	3 930	0.188	0 1487	0.0087	2331	100
40	602	0.45	0.3138	0.0037	5 556	0.080	0 1303	0.0016	2101	22
41	185	0.91	0 4962	0.0097	11 080	0.264	0 1668	0.0039	2526	39
42	150	0.43	0.4609	0.0049	9 587	0.133	0 1530	0.0021	2380	23
43	280	0.46	0.5194	0.0040	13 266	0.224	0.1000	0.0021	2772	25
40	193	0.00	0.3456	0.0055	5 333	0.109	0.1000	0.0020	1885	34
44	37	1 35	0.3327	0.0120	4 872	0.240	0 1125	0.0045	1840	72
45	189	0.34	0.4961	0.0120	11 537	0.240	0.1734	0.0040	2590	29
40	183	0.75	0.4412	0.0103	10 112	0.230	0.1699	0.0026	2557	26
47	469	0.54	0.4571	0.0004	10.885	0.268	0.1815	0.0020	2666	44
40	405	1 95	0.4571	0.0139	5 686	0.000	0.1015	0.0040	1010	20
49	231	0.53	0.3390	0.0044	11 737	0.092	0.1732	0.0013	2580	23
50	440	1 31	0.4335	0.0059	7 806	0.105	0.17.32	0.0023	2009	24
51	440	1.01	0.0102	0.0000	7.000	0.154	0.1000	0.0025	2000	41

Sample/	/analysis	U ppm	Th/U	206Pb/238U	±	207Pb/235U	±	207Pb/206Pb	±	207Pb/206Pb age (Ma)	±
184	52	298	0.31	0.3339	0.0087	5.442	0.164	0.1189	0.0025	1940	37
	53	298	0.76	0.4087	0.0112	10.088	0.325	0.1798	0.0046	2651	42
	54	384	0.46	0.4494	0.0098	10.933	0.238	0.1795	0.0030	2648	28
	55	212	0.39	0.3871	0.0063	6.879	0.126	0.1312	0.0021	2115	28
	56	2521	0.40	0.1161	0.0027	2.856	0.065	0.1836	0.0028	2686	25
	57	1255	0.95	0.1624	0.0039	4.187	0.113	0.1883	0.0028	2727	24
	58	2290	0.71	0.1251	0.0031	3.233	0.079	0.1907	0.0021	2748	18
	59	773	0.80	0.2713	0.0046	6.890	0.129	0.1918	0.0030	2758	26
	60	209	1.35	0 2753	0.0044	3 702	0.069	0.0988	0.0017	1602	32
	61	185	0.51	0.4826	0.0043	10.678	0.135	0.1629	0.0020	2486	21
	62	148	0.88	0 2486	0.0084	4 048	0 147	0 1199	0.0039	1955	58
	63	514	0.46	0.4599	0.0094	11 149	0.251	0.1790	0.0034	2643	32
	64	237	0.40	0.4000	0.0004	12 089	0.201	0.1712	0.0050	2569	49
	65	237	0.74	0.3724	0.0205	10.286	0.512	0.2066	0.0000	2879	54
	66	406	0.74	0.5724	0.0205	0.067	0.316	0.1625	0.0005	2491	37
	67	203	0.30	0.5279	0.0196	13 210	0.310	0.1023	0.0053	2701	46
	69	250	0.34	0.3278	0.0100	10.219	0.478	0.1630	0.0032	2/11	40
	00	209	0.44	0.4742	0.0122	10.555	0.270	0.1039	0.0055	2490	54
	69	228	0.76	0.5077	0.0232	11.000	0.524	0.1729	0.0059	2580	57
	70	360	0.08	0.3620	0.0117	0.297	0.215	0.1278	0.0033	2008	40
	71	195	0.26	0.3766	0.0112	0.212	0.205	0.1246	0.0050	2023	72
	72	68	1.43	0.4470	0.0143	9.243	0.271	0.1550	0.0035	2402	39
	73	473	0.41	0.4787	0.0215	10.379	0.379	0.1649	0.0053	2507	54
	74	180	0.87	0.4511	0.0148	11.029	0.394	0.1792	0.0046	2646	43
	75	677	0.69	0.3956	0.0166	9.987	0.386	0.1897	0.0053	2740	46
055		700	0.44	0 5000	0.0040	44.055	0.010	0.4000	0.0000	0014	00
255	1	722	0.44	0.5362	0.0046	14.255	0.216	0.1982	0.0028	2811	23
	2	439	0.04	0.3593	0.0031	5.992	0.110	0.1248	0.0023	2026	32
	3	1910	0.26	0.2199	0.0027	11.043	0.190	0.3751	0.0051	2030	140
	4	1773	0.48	0.2085	0.0029	6.527	0.110	0.2355	0.0046	1932	101
	5	200	0.53	0.5082	0.0057	12.625	0.258	0.1882	0.0036	2727	32
	6	101	0.83	0.5469	0.0069	14.152	0.294	0.1936	0.0041	2773	35
	7	139	1.29	0.4701	0.0052	10.883	0.246	0.1730	0.0039	2587	38
	8	111	0.62	0.6094	0.0066	18.919	0.407	0.2317	0.0048	3064	33
	9	1140	0.15	0.2731	0.0027	5.493	0.090	0.1508	0.0022	1962	53
	10	773	0.27	0.3747	0.0033	9.672	0.151	0.1934	0.0030	2444	55
	11	2377	0.11	0.1440	0.0019	4.341	0.099	0.2244	0.0037	2188	78
	12	235	0.88	0.5625	0.0055	15.753	0.305	0.2088	0.0037	2896	28
	13	792	0.47	0.4901	0.0042	11.458	0.174	0.1753	0.0025	2609	24
	14	56	1.30	0.3118	0.0051	4.751	0.163	0.1154	0.0040	1886	63
	15	242	1.30	0.5195	0.0055	14.335	0.261	0.2084	0.0039	2782	70
	16	461	0.31	0.4057	0.0033	7.731	0.141	0.1429	0.0024	2262	29
	17	95	0.76	0.4308	0.0050	9.017	0.245	0.1570	0.0042	2424	46
	18	897	0.78	0.2494	0.0038	5.060	0.086	0.1536	0.0034	1920	108
	19	212	0.75	0.3789	0.0054	10.020	0.209	0.2010	0.0043	2684	109
	20	145	0.79	0.3625	0.0040	5.619	0.151	0.1169	0.0032	1910	50
	21	304	1.15	0.3485	0.0035	5.513	0.121	0.1189	0.0025	1940	38
	22	408	1.27	0.5253	0.0047	13.539	0.224	0.1914	0.0031	2754	27
	23	625	0.26	0.3427	0.0030	5.609	0.104	0.1230	0.0022	2000	31
	24	444	0.38	0.3503	0.0033	5.449	0.110	0.1172	0.0024	1913	37
	25	428	0.17	0.4863	0.0045	11.434	0.205	0.1749	0.0028	2605	27
	26	683	0.37	0.3451	0.0033	5.673	0.111	0.1239	0.0023	2013	33

Sample	/analysis	U ppm	Th/U	206Pb/238U	±	207Pb/235U	±	207Pb/206Pb	±	207Pb/206Pb age (Ma)	±
255	27	770	0.58	0 5040	0.0043	12 140	0 172	0 1800	0 0027	2653	25
200	28	338	0.58	0.5635	0.0324	14 008	0.768	0.1823	0.0069	2910	47
	29	190	0.74	0.5912	0.0229	16.831	0.623	0.2106	0.0061	2910	47
	30	706	0.36	0.4703	0.0108	10.572	0.264	0.1670	0.0026	2528	26
	31	498	0.04	0.4861	0.0079	10.688	0.231	0.1636	0.0027	2493	27
	32	77	0.86	0.3383	0.0043	5.246	0.104	0.1150	0.0023	1880	36
	33	723	0.02	0.4633	0.0054	10.106	0.134	0.1608	0.0018	2464	19
	34	893	0.74	0.3780	0.0051	11.079	0.157	0.2155	0.0025	2948	19
	35	608	0.33	0.4894	0.0057	11.817	0.155	0.1770	0.0020	2625	19
	36	327	0.23	0.4867	0.0079	11.329	0.182	0.1703	0.0023	2561	23
	37	278	0.39	0.3803	0.0070	6.407	0.142	0.1283	0.0027	2074	37
	38	359	1.54	0.3494	0.0056	5.790	0.116	0.1216	0.0022	1979	33
	39	476	0.25	0.3361	0.0037	5.205	0.071	0.1142	0.0014	1868	22
	40	285	1.00	0.3584	0.0037	5.714	0.081	0.1171	0.0016	1913	25
	41	213	0.74	0.3669	0.0037	5.975	0.086	0.1197	0.0016	1952	25
	42	588	0.60	0.4793	0.0088	11.990	0.266	0.1830	0.0033	2681	30
	43	392	0.83	0.3484	0.0035	5.750	0.076	0.1212	0.0015	1974	22
	44	261	0.19	0.3884	0.0045	7.021	0.110	0.1325	0.0018	2131	24
	45	160	0.72	0.5416	0.0073	14.584	0.222	0.1979	0.0025	2809	20
	46	194	0.71	0.4973	0.0054	12.154	0.151	0.1792	0.0022	2646	20
	47	407	0.34	0.3468	0.0032	6.682	0.090	0.1416	0.0019	2247	23
	48	289	0.67	0.5485	0.0085	15.213	0.292	0.2084	0.0036	2893	28
	49	531	0.26	0.3299	0.0041	5.122	0.075	0.1143	0.0015	1870	24
	50	514	0.34	0.4717	0.0080	10.649	0.224	0.1644	0.0028	2502	29
	51	408	0.82	0.5177	0.0058	13.633	0.181	0.1935	0.0021	2772	18
	52	159	0.66	0.3960	0.0053	7.203	0.123	0.1339	0.0020	2150	26
	53	435	0.62	0.4571	0.0064	11.833	0.181	0.1887	0.0031	2731	27
	54	183	1.02	0.5304	0.0102	13.249	0.276	0.1855	0.0031	2703	28
	55	330	0.40	0.4766	0.0077	11.529	0.197	0.1769	0.0026	2624	24
	56	192	0.61	0.5287	0.0065	13.477	0.196	0.1867	0.0024	2713	21
	57	154	0.47	0.4881	0.0070	11.234	0.204	0.1691	0.0027	2549	27
	58	275	0.47	0.4696	0.0053	10.641	0.137	0.1674	0.0020	2532	20
	59	393	0.31	0.3963	0.0093	8.403	0.187	0.1597	0.0033	2452	35
	60	947	0.45	0.35/6	0.0051	11.199	0.187	0.2299	0.0029	3051	21
	01	307	0.51	0.4/43	0.0063	10.477	0.147	0.1624	0.0020	2481	21
	62	251	0.90	0.5079	0.0083	12.017	0.223	0.1744	0.0028	2600	27
	03	000	0.01	0.3476	0.0068	5.599	0.113	0.1196	0.0019	192.1	29

Table 4	. Mon	azite EMPA	data for	metase	dimentary	samples	from the	lower E	theridge G	roup.
					2		-		<u> </u>	

Grain/ar	nalysis	Si	Al	К	Fe	Ca	Sr	Pb	Y	Th	U	La	Ce	Pr	Nd	Sm	Eu	Gd	Dy	Yb	Er	Р
Sample 2	83: Einas	sleigh Me	tamorphics	s (Gum Flats	5)																	
1	1	0.12	0.00	0.01	0.06	1.04	-0.01	0.61	2.08	5.23	1.11	10.65	21.24	2.47	9.05	1.45	0.14	1.15	0.58	0.07	0.16	12.59
	2	0.13	0.00	-0.01	0.06	0.96	-0.01	0.56	2.48	5.31	0.76	10.43	21.08	2.48	9.06	1.45	0.07	1.31	0.67	0.01	0.14	12.45
	3	0.11	0.00	0.02	0.04	0.92	-0.02	0.51	2.40	5.25	0.53	10.81	22.47	2.55	9.11	1.44	-0.01	1.23	0.66	0.04	0.15	12.56
2	1	0.08	0.00	-0.01	0.04	0.96	-0.01	0.55	1.88	4.74	0.97	10.86	21.74	2.47	8.94	1.46	0.13	1.17	0.56	0.08	0.16	12.48
	2	0.09	-0.01	-0.01	0.07	1.01	-0.01	0.54	1.97	4.90	1.02	11.43	23.14	2.50	9.03	1.42	0.15	1.14	0.58	0.09	0.18	12.60
	3	0.08	0.00	0.00	0.06	1.04	-0.01	0.62	1.52	5.54	0.86	11.02	22.52	2.53	9.25	1.42	0.18	1.11	0.50	0.04	0.11	12.45
	4	0.08	0.00	0.01	0.06	1 00	-0.01	0.59	1 43	5.33	0.84	11 43	22.91	2 44	9 16	1.39	0.14	1 12	0 49	0.07	0.12	12 51
3	1	0.22	0.00	-0.01	0.04	1.00	-0.01	0.61	1 44	7 25	0.38	10.46	21.46	2 42	8.95	1 40	0.15	1.00	0.45	0.04	0.15	12.16
Ũ	2	0.08	0.01	-0.01	0.04	1 15	0.00	0.67	1 72	5.24	1 15	10.90	22.12	2.50	9.07	1 47	0.10	1 16	0.56	0.06	0.14	12.76
	3	0.00	0.01	0.00	0.07	0.72	0.01	0.42	1 17	5.85	0.21	11 90	24.09	2.55	9.25	1 47	0.15	1.00	0.49	0.05	0.07	12.39
4	1	0.10	0.01	0.00	0.06	0.97	0.00	0.57	2.26	5 38	0.21	10.76	22.20	2.00	8.01	1.47	-0.01	1.00	0.45	0.06	0.16	12.00
-	2	0.10	0.00	-0.01	0.00	0.37	0.00	0.50	1.06	6.23	0.73	11 10	22.20	2.50	9.16	1.48	0.01	1.24	0.05	0.00	0.10	12.01
	2	0.01	0.00	-0.01	0.00	0.74	0.01	0.50	2.57	5.52	0.21	10.93	22.54	2.02	0.16	1.40	0.10	1.00	0.43	0.02	0.07	12.23
	1	0.14	0.01	0.00	0.10	0.90	0.00	0.55	2.57	5.52	0.03	10.05	22.03	2.52	9.10	1.31	0.01	1.34	0.72	0.05	0.10	12.77
	5	0.14	0.00	-0.01	0.07	1 42	-0.01	0.34	2.55	7.54	1.00	10.40	21.07	2.49	9.10	1.44	0.00	1.34	0.74	0.03	0.10	12.00
-	4	0.05	0.00	-0.03	0.00	1.42	0.00	0.79	1.72	6.70	1.09	10.05	22.14	2.37	0.00	1.50	0.17	1.07	0.54	0.04	0.12	12.02
5	1	0.11	0.02	-0.01	0.12	1.39	-0.01	0.62	2.13	0.70	0.79	10.27	21.00	2.30	0.93	1.01	0.10	1.24	0.05	0.09	0.10	13.00
7	1	0.21	0.00	0.01	0.14	1.40	0.11	0.00	1.04	6.01 E E0	1.09	10.35	21.00	2.31	0.72	1.44	-0.01	1.12	0.56	0.03	0.11	12.72
<i>'</i>		0.14	-0.01	0.01	0.04	1.10	0.01	0.00	1.00	5.50	1.00	10.64	22.95	2.40	0.90	1.40	-0.05	1.15	0.00	0.00	0.11	12.04
8	1	0.03	-0.01	0.00	0.02	0.36	0.02	0.12	2.34	0.94	0.22	11.36	25.50	2.91	10.38	1.75	0.02	1.56	0.72	0.04	0.17	12.70
~	2	0.11	0.00	0.01	0.02	0.93	0.00	0.55	2.29	4.92	0.81	11.65	22.86	2.45	8.93	1.45	-0.13	1.12	0.61	0.13	0.16	12.54
9	1	0.11	0.00	0.01	0.13	0.39	0.00	0.26	0.84	2.72	0.26	13.70	24.64	2.58	9.87	1.65	0.04	1.31	0.40	0.02	0.01	12.49
10	1	0.15	0.00	0.01	0.10	1.23	0.01	0.65	1.41	6.79	0.59	11.67	22.88	2.44	8.36	1.23	-0.03	1.02	0.49	0.06	0.14	12.63
11	1	0.20	0.00	0.02	0.25	0.91	0.00	0.55	1.39	6.41	0.41	11.87	23.63	2.48	8.76	1.36	-0.06	1.10	0.51	0.03	0.06	12.39
	2	80.0	0.00	0.02	0.22	1.05	0.00	0.52	1.92	5.54	0.60	11.21	22.88	2.51	9.07	1.39	-0.09	1.25	0.69	0.08	0.15	12.60
12	1	0.12	0.00	0.01	0.08	0.99	0.01	0.51	2.00	5.59	0.51	10.79	22.53	2.47	9.19	1.49	-0.09	1.22	0.60	0.09	0.13	12.54
13	1	0.18	0.00	0.01	0.04	1.00	-0.01	0.55	2.24	5.58	0.63	10.78	22.33	2.46	9.09	1.49	-0.15	1.36	0.67	0.07	0.11	12.52
	2	0.16	0.00	0.01	0.02	1.00	-0.01	0.57	2.27	5.60	0.65	10.59	22.06	2.44	9.08	1.43	-0.18	1.28	0.68	0.03	0.13	12.54
14	1	0.33	0.00	0.02	0.14	1.07	0.02	0.50	1.51	6.56	0.78	10.98	22.51	2.42	8.90	1.52	-0.04	1.26	0.59	0.03	0.11	12.27
	2	0.11	0.00	0.01	0.02	1.77	-0.01	0.92	1.87	9.43	1.02	9.21	19.86	2.19	8.24	1.47	-0.06	1.30	0.63	0.09	0.15	12.58
	3	0.09	0.00	0.01	0.07	1.29	0.00	0.68	1.86	5.97	1.12	10.37	21.75	2.49	8.91	1.56	-0.11	1.35	0.66	0.06	0.16	12.54
Sample 2	288: Danie	O 40	Formation (Gum Flats)	0.00	4.05	0.04	0.00	1.05	10.50	0.40	0.40	10.07	0.05	0.00	1 70	0.00	4 40	0.40	0.00	0.00	11.00
	1	0.48	0.03	0.01	0.23	1.35	0.01	0.82	1.05	10.52	0.19	9.19	19.37	2.35	9.00	1.70	0.33	1.42	0.40	0.02	0.06	11.90
2	1	0.09	0.01	-0.01	0.29	0.33	0.01	0.27	1.10	2.04	0.28	13.42	25.45	2.75	9.76	1.53	0.28	1.00	0.39	0.03	0.06	12.05
~	2	0.11	0.03	0.02	0.38	0.28	0.01	0.18	0.89	1.64	0.23	13.28	25.67	2.88	10.00	1.49	0.34	0.80	0.35	0.04	0.11	12.60
3	1	0.07	0.01	0.00	0.41	0.34	0.00	0.16	0.98	1.55	0.29	13.17	24.72	2.82	10.04	1.50	0.32	0.96	0.35	0.05	0.06	12.47
	2	0.06	0.01	0.01	0.47	0.84	0.02	0.62	0.95	2.79	1.82	11.89	23.23	2.69	9.52	1.59	0.44	1.21	0.38	0.05	0.05	12.82
	3	0.05	0.01	0.00	0.39	0.41	0.02	0.69	1.10	2.93	2.02	12.53	22.46	2.62	9.43	1.45	0.38	1.09	0.25	0.02	0.06	12.29
4	1	0.26	0.10	0.00	0.85	1.53	0.07	0.09	0.76	3.17	0.19	12.81	20.03	2.97	10.35	1.60	0.27	0.90	0.32	0.02	0.08	10.89
	2	0.13	0.05	-0.02	0.68	0.64	0.12	0.13	0.98	1.50	0.21	14.26	24.59	3.38	12.22	1.94	0.39	1.07	0.32	0.01	0.08	12.62
	3	0.14	0.03	-0.01	0.51	1.24	0.00	0.13	0.93	1.16	0.32	13.47	23.94	2.76	10.07	1.58	0.33	0.90	0.36	0.07	0.07	13.29
	4	0.23	0.05	0.05	0.48	0.64	0.05	0.10	0.82	0.93	0.21	13.81	24.76	2.91	10.15	1.52	0.30	0.85	0.34	0.03	0.03	12.51
	5	0.11	0.01	-0.01	0.30	0.92	0.03	0.28	1.53	1.23	0.71	12.45	24.96	2.91	10.20	1.62	0.30	1.05	0.52	0.01	0.10	13.21
5	1	0.39	0.08	0.00	1.45	0.45	0.04	0.06	1.11	2.73	0.20	13.29	24.94	2.65	9.44	1.35	0.21	0.89	0.32	0.07	0.09	11.98
6	1	0.28	0.09	0.01	0.39	0.36	0.00	0.13	0.96	0.72	0.28	13.59	26.69	2.90	10.08	1.54	0.31	0.91	0.34	0.06	0.07	12.80
7	1	0.19	0.02	0.11	0.14	0.62	0.01	0.45	0.94	5.20	0.27	11.81	23.02	2.64	9.40	1.40	0.25	0.85	0.37	0.03	0.07	12.28
	2	0.30	0.03	0.09	0.12	0.95	0.01	0.50	1.03	7.05	0.22	11.23	21.76	2.41	8.98	1.43	0.30	0.81	0.43	0.01	0.07	12.25
8	1	0.36	0.01	0.02	0.11	1.17	0.03	0.68	1.02	9.46	0.16	10.12	20.65	2.31	8.97	1.70	0.34	1.58	0.48	0.02	0.08	11.97
9	1	0.22	0.02	0.01	1.32	0.69	0.01	0.34	0.95	3.87	0.19	12.49	23.94	2.59	9.30	1.54	0.13	1.30	0.47	0.02	-0.01	12.17
	2	0.32	0.01	0.01	0.98	1.10	0.03	0.57	0.97	7.27	0.16	10.90	21.73	2.41	8.91	1.63	0.10	1.39	0.49	0.03	0.05	11.96

Grain/analysis		S	As	0	Total (wt%)	Pb (ppm)	Th (ppm)	U (ppm)	Age (Ma)	±1σ
Sample	283: Ein	asleigh Me	tamorphics	s (Gum Flats)						
1	1	0.00	0.01	26.26	96.08	6087	52270	11105	1444	16
	2	0.00	0.03	26.08	95.52	5548	53099	7602	1502	18
	3	0.01	0.02	26.43	97.23	5089	52457	5260	1548	20
2	1	0.01	0.02	26.01	95.31	5510	47424	9659	1465	17
	2	0.00	0.02	26.59	98.48	5408	49031	10206	1385	16
	3	0.00	0.00	26.19	97.06	6164	55394	8622	1549	17
	4	0.00	0.01	26.30	97.42	5875	53319	8356	1532	17
3	1	0.00	0.01	25.76	95.44	6091	72468	3765	1537	17
	2	0.00	0.01	26.59	97.52	6696	52446	11549	1548	16
	3	0.00	0.01	26.44	98.57	4133	58526	2143	1363	21
4	1	0.00	0.01	26.40	96.97	5712	53753	7855	1515	18
	2	0.00	0.02	26.04	96.60	4953	62293	2099	1539	21
	3	0.00	0.02	26.98	99.29	5309	55221	6287	1486	18
	4	0.00	0.00	26.58	97.37	5360	54667	6345	1505	19
	5	0.00	0.01	26.95	99.74	7895	75448	10935	1498	13
5	1	0.00	0.00	27.35	100.08	8175	67754	11373	1626	14
6	1	0.01	0.01	26.89	99.05	6602	80115	7828	1338	14
7	1	0.01	0.00	26.87	98.62	6855	55023	10766	1584	16
8	1	0.00	0.02	26.62	97.82	1297	9359	2236	1612	69
	2	0.00	0.01	26.50	98.06	5508	49196	8128	1527	18
9	1	0.00	0.01	26.30	97.74	2632	27154	2613	1562	36
10	1	0.00	0.01	26.66	98.53	6552	67910	5932	1590	17
11	1	0.00	0.01	26.49	98.83	5525	64064	4109	1523	18
	2	0.00	0.00	26.58	98.37	5280	55377	5968	1496	18
12	1	0.00	0.02	26.40	97.30	5188	55882	5119	1518	19
13	1	0.00	0.00	26.46	97.57	5514	55818	6323	1526	18
	2	0.01	0.01	26.38	96.95	5732	56016	6533	1564	18
14	1	0.01	0.01	26.35	97.90	5075	65591	7819	1198	15
	2	0.00	0.00	26.38	97.21	9261	94275	10150	1538	12
	3	0.00	0.02	26.37	97.35	6858	59669	11177	1496	15
0 1 -	000 D-									
Sample	288: Da	niel Creek i	-ormation (Gum Flats)	00.25	0140	105040	1010	1570	4.4
1	1	0.00	0.02	20.67	90.35	0140	105243	1910	15/3	14
2	1	0.00	0.02	20.00	99.01	2709	20435	2049	1599	30
2	2	0.01	0.03	20.50	97.65	1714	16390	2310	1209	50
3	2	0.00	0.01	20.14	90.47	6140	10402	10016	1320	40
	2	0.00	0.01	20.07	96.14	6140	27652	10210	1444	13
	3	0.00	0.03	23.77	90.00	0000	29335	20195	1473	14
4	1	0.24	0.02	24.72	92.25	1000	31009	1910	1240	20
	2	0.07	0.03	27.45	102.69	1202	14976	2144	1249	52
	3	0.50	0.02	20.37	100.24	1225	0070	3102	109	53
	4	0.09	0.02	20.08	97.53	921	9273	2102	1216	68
-	5	0.04	0.02	27.58	100.08	2808	12328	7092	1604	33
5	1	0.03	0.03	26.37	98.16	568	27343	2015	374	31
6	1	0.00	0.01	27.10	99.67	1289	/159	2/03	1027	70
1	1	0.00	0.02	25.92	96.01	4416	51969	2051	1558	23
•	2	0.01	0.03	26.00	96.00	4931	/04/5	2185	13/4	18
8	1	0.01	0.02	25.86	97.11	0803	94588	1030	14/3	15
э	1	0.00	0.01	20.27	91.01	3429	30/13	10/1	1032	20 10
	2	0.00	0.01	25.95	96.99	5788	(2/01	1602	1598	18

Grain/ana	alysis	Si	Al	K	Fe	Ca	Sr	Pb	Y	Th	U	La	Ce	Pr	Nd	Sm	Eu	Gd	Dy	Yb	Er	Р
10	1	0.09	0.00	0.01	0.28	0.41	0.01	0.24	0.91	2.83	0.26	12.88	25.17	2.71	9.59	1.50	0.13	1.04	0.41	0.04	0.04	12.48
11	1	0.06	0.01	0.01	0.45	0.85	0.03	0.55	0.88	2.75	1.60	12.20	23.92	2.71	9.52	1.54	0.17	1.33	0.43	0.03	0.06	12.60
12	1	0.06	0.01	0.02	0.24	0.23	0.01	0.13	0.97	0.91	0.27	13.64	26.75	2.87	9.98	1.51	0.12	0.92	0.44	0.05	0.10	12.58
13	1	0.03	0.00	0.01	0.20	0.36	0.00	0.27	1 39	0.69	0.92	13.40	26.15	2 77	9.51	1 4 1	0.13	0.89	0.45	0.03	0.09	12 50
10	2	0.00	0.00	0.00	0.18	0.00	0.00	0.26	1.00	0.68	0.87	13.62	26.56	2.75	9.69	1.36	0.10	0.00	0.40	0.00	0.05	12.00
	2	0.07	0.00	0.00	0.10	0.23	0.01	0.20	1.10	0.00	0.07	14.05	20.00	2.75	0.70	1.30	0.13	0.31	0.42	0.01	0.03	12.07
4.4	3	0.02	0.00	0.01	0.20	0.23	0.00	0.10	1.02	0.00	0.40	14.05	20.94	2.04	9.70	1.55	0.12	0.79	0.35	0.00	0.03	12.03
14	1	0.19	0.05	0.04	1.01	0.83	0.04	0.37	0.99	4.95	0.22	11.90	23.65	2.60	9.66	1.00	0.20	1.10	0.42	0.04	0.03	12.23
15	1	0.19	0.01	0.03	0.87	0.79	0.03	0.42	1.02	5.19	0.20	12.09	23.07	2.47	9.22	1.60	0.12	1.32	0.46	0.01	0.04	12.31
16	1	0.30	0.00	0.02	0.62	1.09	0.02	0.63	1.10	8.25	0.22	10.34	20.86	2.38	9.00	1.68	0.16	1.43	0.45	0.06	0.07	11.93
17	1	0.57	0.00	0.01	0.92	1.28	0.04	0.80	0.91	11.09	0.15	9.62	20.09	2.29	9.09	1.72	0.11	1.47	0.46	0.03	0.06	11.69
18	1	0.14	0.02	0.01	0.80	0.60	0.02	0.30	0.91	3.40	0.22	11.67	22.53	2.37	8.68	1.36	0.14	1.02	0.36	0.04	0.05	11.60
19	1	0.15	0.00	0.00	0.20	0.13	0.01	0.02	0.82	0.80	0.16	14.05	26.80	2.85	9.90	1.31	0.12	0.93	0.32	0.06	0.03	12.40
	2	0.39	0.16	0.01	0.34	0.36	0.02	0.05	1.33	1.59	0.42	13.26	25.38	2.74	9.72	1.46	0.06	1.13	0.52	0.05	0.05	12.37
Sample 30)4: Lane	e Creek F	ormation (R	lobinhood)																		
1	1	0.03	0.00	0.01	0.10	0.75	0.00	0.39	1.54	3.12	0.66	11.91	24.33	2.64	9.67	1.42	-0.08	1.20	0.54	0.08	0.10	12.74
	2	0.02	0.00	0.00	0.06	0.35	0.00	0.06	1.60	0.63	0.14	13.59	25.79	2.71	9.80	1.65	0.00	1.37	0.63	0.06	0.07	12.93
	3	0.03	0.01	0.01	0.05	0.74	0.00	0.40	1.56	3 23	0.73	12 15	24 57	2.63	9.51	1.51	-0.07	1 18	0.57	0.05	0.13	12 77
2	1	0.02	0.00	0.01	0.07	0.32	-0.01	0.13	1.50	0.81	0.20	13 30	25.52	2.00	9.96	1.67	-0.05	1.46	0.64	0.06	0.03	12.80
-	2	0.02	0.00	0.03	0.04	0.62	-0.01	0.10	1.57	2 93	0.20	12.03	24.58	2.68	9.61	1.55	-0.08	1.40	0.54	0.00	0.00	12.00
2	4	0.05	0.00	0.03	0.07	0.03	-0.01	0.41	1.57	2.33	0.77	11.40	27.50	2.00	0.60	1.55	-0.00	1.20	0.54	0.03	0.11	12.01
3	1	0.05	0.00	0.03	0.07	0.87	-0.02	0.49	1.50	3.77	0.86	11.42	23.05	2.04	9.69	1.51	-0.09	1.20	0.62	0.07	0.12	12.92
4	1	0.06	0.01	0.04	0.08	0.88	0.00	0.49	1.05	3.95	0.87	11.25	23.40	2.03	9.55	1.53	-0.10	1.10	0.54	0.02	0.17	12.73
5	1	0.07	0.00	0.01	0.09	0.76	-0.01	0.41	1.40	3.45	0.71	12.09	24.36	2.63	9.35	1.47	-0.10	1.09	0.51	0.08	0.10	12.59
6	1	0.16	0.04	0.09	0.52	0.95	0.01	0.49	1.70	4.19	0.80	11.14	23.21	2.54	9.30	1.54	-0.13	1.26	0.58	0.02	0.09	12.62
7	1	0.11	0.00	0.03	0.06	0.94	0.00	0.53	1.77	4.59	0.79	11.03	22.95	2.55	9.34	1.47	-0.10	1.26	0.64	0.05	0.13	12.73
8	1	0.06	0.00	0.03	0.08	0.57	0.01	0.19	1.45	1.87	0.31	12.35	25.00	2.82	9.74	1.50	-0.03	1.11	0.56	0.03	0.10	12.59
9	1	0.94	0.22	0.08	1.66	0.83	0.08	0.25	1.45	3.45	0.39	10.83	22.55	2.58	9.40	1.45	-0.07	1.08	0.48	0.06	0.07	12.00
10	1	0.12	0.03	0.07	0.58	0.59	0.01	0.29	1.53	2.74	0.47	11.83	24.84	2.78	9.83	1.57	-0.10	1.38	0.65	0.04	0.07	12.78
11	1	0.05	0.01	0.08	0.09	0.84	0.00	0.46	1.66	3.76	0.84	11.61	23.50	2.50	9.28	1.47	-0.12	1.20	0.56	0.04	0.14	12.61
12	1	0.11	0.00	0.02	0.06	0.76	0.00	0.44	1.69	3.38	0.81	11.51	23.96	2.70	9.52	1.52	-0.14	1.22	0.61	0.05	0.12	12.81
13	1	0.12	0.03	0.07	0.27	0.66	0.00	0.28	1 97	2 77	0.45	11.54	24 12	2 72	9.58	1.54	-0.09	1.30	0.67	0.08	0.07	12 84
14	1	0.07	0.00	0.02	0.04	0.83	-0.02	0.46	1.63	3.68	0.80	11.55	23.96	2 59	9.37	1.55	-0.09	1 19	0.58	0.03	0.13	12 65
15	1	0.06	0.00	0.03	0.18	0.60	0.00	0.31	1.63	2 71	0.50	12.02	24.77	2 70	9.66	1 51	-0.06	1 30	0.50	0.05	0.05	12.80
15	2	0.00	0.00	0.03	0.10	0.00	0.00	0.31	1.00	2.71	0.50	12.02	24.80	2.70	0.57	1.01	-0.00	1.50	0.53	0.05	0.05	12.01
10	4	0.00	0.01	0.03	0.55	0.00	0.00	0.31	1.50	2.30	0.51	12.04	24.00	2.05	9.57	1.49	-0.04	1.15	0.51	0.00	0.00	12.70
10	1	0.04	0.01	0.08	0.13	0.65	0.00	0.29	1.54	2.38	0.56	11.82	23.97	2.73	9.64	1.52	-0.08	1.23	0.58	0.04	0.07	12.53
17	1	0.04	0.00	0.03	0.12	0.53	-0.01	0.21	1.73	2.04	0.39	11.39	24.56	2.88	9.95	1.70	-0.04	1.28	0.65	0.06	0.14	12.80
18	1	0.06	0.00	0.05	0.26	0.70	0.00	0.35	1.60	3.61	0.45	11.60	23.99	2.78	9.71	1.51	-0.11	1.25	0.51	0.06	0.13	12.53
19	1	0.06	0.04	0.02	0.20	0.19	-0.01	0.04	1.37	0.54	0.09	12.99	27.00	2.91	10.27	1.51	0.00	1.11	0.55	0.04	0.08	12.65
20	1	0.38	0.00	0.01	0.12	0.75	0.00	0.38	1.55	3.72	0.53	11.71	23.92	2.73	9.62	1.54	-0.15	1.20	0.60	0.07	0.09	12.46
21	1	0.06	-0.01	0.03	0.07	0.74	0.00	0.37	1.65	3.82	0.44	11.53	23.81	2.79	9.64	1.57	-0.12	1.27	0.57	0.05	0.11	12.60
22	1	0.02	0.00	0.01	0.11	0.77	-0.01	0.41	1.45	3.40	0.75	11.92	23.61	2.62	9.24	1.41	-0.05	1.10	0.51	0.04	0.08	12.57
23	1	0.03	0.00	0.03	0.14	0.89	-0.01	0.50	1.55	4.19	0.80	11.07	22.77	2.61	9.34	1.49	-0.09	1.18	0.56	0.03	0.09	12.55
Sample 30	6: Cork	oett Form	ation (Robi	nhood)																		
1	1	0.34	0.01	0.02	0.13	0.24	-0.01	0.16	0.87	3.05	0.21	12.42	26.15	2.86	10.14	1.56	-0.01	0.91	0.37	0.03	0.05	12.15
	2	0.07	0.00	0.03	0.09	0.45	0.01	0.24	1.18	2.56	0.26	12.38	25.85	2.79	9.80	1.56	0.02	1.04	0.46	0.06	0.09	12.55
2	1	0.04	0.01	0.05	0.04	0.26	-0.01	0.17	1.20	1.29	0.33	13.06	26.37	2.87	9.88	1.49	-0.02	1.06	0.43	0.02	0.08	12.57
	2	0.04	0.00	0.03	0.04	0.29	-0.01	0.17	1 18	1.30	0.33	13 20	26.47	2.92	9 70	1 43	-0.02	1 07	0.49	0.04	0.08	12.57
3	1	0.07	0.00	0.01	0.06	0.53	-0.01	0.26	1.30	2 60	0.33	12 44	25.41	2 79	9.51	1.39	-0.02	1.00	0.51	0.03	0.09	12.52
4	1	0.10	0.00	0.01	0.06	0.55	0.00	0.20	1 18	3 10	0.00	12.50	25.02	2.60	9.57	1 4 2	0.05	1.00	0.30	0.00	0.00	12.52
7	2	0.10	0.00	0.01	0.00	0.33	0.00	0.25	1.10	1.57	0.42	12.00	20.02	2.05	9.57	1.42	0.05	1.00	0.35	0.03	0.00	12.58
	2	0.04	0.00	0.02	0.09	0.33	-0.02	0.10	1.10	1.07	0.31	12.90	20.21	2.04	9.00	1.42	0.00	0.00	0.44	0.04	0.00	12.04
1	3	0.06	0.00	0.03	0.12	0.40	0.01	0.21	1.11	1.94	0.34	12.88	25.80	2.13	9.54	1.47	-0.02	0.99	0.45	0.07	0.11	12.71

Grain/a	analysis	S	As	0	Total (wt%)	Pb (ppm)	Th (ppm)	U (ppm)	Age (Ma)	±1σ
10	1	0.00	0.01	26.25	97.32	2492	28301	2623	1441	32
11	1	0.00	0.01	26.51	98.21	5505	27473	16011	1423	15
12	1	0.01	0.02	26.50	98.31	1364	9089	2672	1581	59
13	1	0.00	0.02	26.32	97.54	2748	6874	9188	1504	28
	2	0.00	0.02	26.43	98.03	2679	6790	8717	1531	30
	3	0.00	0.01	26.49	98.13	1616	6580	4807	1482	45
14	1	0.02	0.02	26.43	98.61	3709	49496	2213	1409	22
15	1	0.01	0.01	26.30	97.77	4254	51922	2008	1560	22
16	1	0.01	0.02	25.75	96.39	6365	82547	2152	1530	16
17	1	0.01	0.03	26.03	98.49	8088	110927	1523	1509	13
18	1	0.00	0.02	24.58	90.83	3020	33958	2231	1560	29
19	1	0.00	0.01	26.22	97.28	217	7981	1566	372	43
	2	0.00	0.00	26.69	98.11	522	15872	4164	397	28
Sample	304.1 a	he Creek Fo	rmation (R	obinhood)						
1	1	0.01	0.02	26 60	97 86	3971	31220	6550	1572	23
•	2	0.01	0.01	26.87	98.34	698	6259	1365	1374	91
	3	0.00	0.02	26.71	98.57	4057	32294	7338	1506	22
2	1	0.00	0.00	26.65	98.00	1322	8066	2888	1556	60
	2	0.01	0.01	26.73	98.44	4102	29297	7688	1563	23
3	1	0.00	0.02	26.83	98.39	4972	37710	8649	1567	19
4	1	0.00	0.02	26.57	97.68	4926	39491	8749	1512	19
5	1	0.00	0.02	26.41	97.60	4117	34547	7079	1495	21
6	1	0.02	0.00	26.69	97.96	4931	41945	7970	1519	19
7	1	0.00	0.01	26.58	97.56	5351	45922	7879	1563	18
8	1	0.02	0.02	26.35	96.73	1943	18665	3100	1426	39
9	1	0.08	0.00	26.88	96.81	2531	34515	3889	1157	24
10	1	0.01	0.01	26.96	99.19	2914	27393	4652	1444	28
11	1	0.00	0.00	26.36	97.05	4692	37574	8353	1511	19
12	1	0.00	0.01	26.73	98.05	4489	33782	8064	1553	21
13	1	0.01	0.02	26.85	97.95	2893	27728	4455	1445	28
14	1	0.00	0.02	26.51	97.68	4608	36840	8003	1527	20
15	1	0.00	0.01	26.74	98.24	3163	27057	4958	1532	28
	2	0.00	0.01	26.73	98.28	3158	25831	5121	1550	28
16	1	0.00	0.02	26.12	95.93	2957	23751	5621	1469	27
17	1	0.00	0.02	26.56	97.06	2183	20364	3857	1399	34
18	1	0.00	0.02	26.37	97.54	3563	36056	4546	1482	24
19	1	0.00	0.02	26.61	98.27	490	5369	938	1242	104
20	1	0.00	0.01	26.62	98.02	3840	37206	5295	1488	23
21	1	0.00	0.00	26.42	97.57	3738	38212	4426	1503	24
22	1	0.00	0.02	26.17	96.22	4187	33965	7453	1502	21
23	1	0.00	0.00	26.11	95.93	5028	41937	8032	1542	19
<u> </u>										
Sample	306: Co	rbett Forma	ition (Robii	1hood) 26.10	07.97	1637	30505	2088	059	29
1	2	0.00	0.02	20.19	97.07	2475	30303	2000	900	20
2	2 1	0.00	0.01	20.40	97.90	1725	12006	3266	1523	34 46
4	2	0.00	0.01	20.33	97.50	1704	12000	3237	1486	40
з	2 1	0.00	0.02	20.30	97.10	2655	26010	3343	1518	40
4	1	0.00	0.02	20.30	97.10	2000	20010	4152	1424	27
4	2	0.00	0.03	20.41	97.31	1821	15720	3111	1424	43
	3	0.00	0.07	26.50	97.50	2195	19423	3432	1505	37
	5	0.00	0.02	20.00	31.50	2100	10720	0702	1000	51

Grain/ana	alysis	Si	Al	K	Fe	Ca	Sr	Pb	Y	Th	U	La	Ce	Pr	Nd	Sm	Eu	Gd	Dy	Yb	Er	Р
5	1	0.07	0.00	0.02	0.24	0.60	0.01	0.31	1.91	2.72	0.49	10.93	23.77	2.86	10.67	1.93	-0.02	1.37	0.60	0.04	0.14	12.54
6	1	1.14	0.65	0.04	0.30	0.63	0.31	0.10	0.73	3.08	0.11	12.37	23.55	2.60	9.29	1.61	0.11	1.05	0.33	0.04	0.02	11.63
7	1	3.51	3.63	0.36	1.34	0.52	0.07	0.12	0.61	2.11	0.16	11.26	21.81	2.63	9.66	1.31	0.03	0.84	0.25	0.04	0.08	10.51
8	1	0.13	0.00	0.00	0.03	0.56	-0.01	0.31	1.30	3.34	0.33	12.67	24.58	2.70	9.65	1.55	0.02	1.22	0.45	0.04	0.06	12.52
	2	0.15	0.00	0.01	0.04	0.55	-0.01	0.30	1.24	3.39	0.30	12.55	24.35	2.65	9.78	1.62	0.00	1.17	0.45	0.03	0.05	12.49
Sample 30	08: Dan	iel Creek I	ormation (Gum Flats)																		
1	1	0.31	0.01	0.04	0.08	1.54	-0.02	0.57	1.60	5.01	0.98	11.03	21.81	2.67	9.75	1.55	0.23	1.08	0.54	0.04	0.09	13.50
2	1	0.34	0.02	0.04	0.08	2.96	-0.01	1.48	1.69	17.84	0.83	8.28	16.47	1.86	6.45	1.07	0.30	1.03	0.49	0.04	0.08	12.48
	2	0.12	0.10	0.05	0.09	1.05	-0.01	0.57	1.69	4.49	1.03	11.72	23.50	2.58	9.39	1.52	0.10	1.18	0.57	0.06	0.09	13.02
3	1	0.30	0.04	-0.01	0.17	1.44	0.00	0.77	1.61	6.69	1.36	10.36	20.24	2.41	8.49	1.28	0.18	1.05	0.49	0.03	0.10	12.21
4	1	0.04	0.00	0.03	0.03	1.59	-0.01	1.01	1.80	6.94	2.00	10.07	19.84	2.33	8.39	1.25	0.18	1.02	0.50	0.06	0.14	12.49
	2	0.04	0.01	0.04	0.05	1.36	0.00	0.80	1.42	6.57	1.35	10.62	20.48	2.45	8.72	1.36	0.25	1.06	0.45	0.09	0.12	12.41
5	1	0.18	0.01	0.09	0.25	0.33	0.00	0.25	0.88	2.98	0.15	11.66	23.82	2.87	10.33	1.68	0.28	1.08	0.40	0.04	0.10	12.34
	2	0.54	0.02	0.13	0.22	0.38	0.01	0.56	0.85	8.00	0.17	10.64	21.03	2.60	9.81	1.51	0.21	0.94	0.38	0.03	0.03	11.29
6	1	0.44	0.01	0.02	0.22	0.41	-0.01	0.26	0.99	3.68	0.12	12.26	24.60	2.73	9.88	1.56	0.21	1.06	0.42	0.03	0.00	12.16
7	1	0.09	0.03	-0.02	0.18	1.49	-0.02	0.86	1.79	6.94	1.50	10.43	20.91	2.34	8.67	1.33	0.21	1.10	0.53	0.04	0.16	12.98
Sample 32	24: Eina	asleiah Me	tamorphics	Gum Flats	5)																	
1	1	0.12	0.00	0.02	0.18	0.88	0.01	0.50	1.65	4.87	0.60	12.04	23.03	2.44	8.99	1.43	-0.07	1.20	0.58	0.05	0.12	12.39
	2	0.07	0.01	0.02	0.14	0.94	0.01	0.50	1.82	4.63	0.69	11.97	22.93	2.50	9.10	1.40	-0.10	1.23	0.59	0.09	0.18	12.60
	3	0.40	0.00	0.03	0.20	0.76	0.01	0.59	1.60	6.85	0.39	11.90	22.72	2.40	8.58	1.38	-0.08	1.17	0.57	0.06	0.13	11.98
2	1	0.14	0.00	0.01	0.10	0.61	0.01	0.41	1.58	4.09	0.43	12.44	23.80	2.63	9.18	1.40	-0.11	1.15	0.54	0.05	0.08	12.28
-	2	0.18	0.21	0.01	0.16	0.57	0.01	0.35	1 48	4 02	0.31	12.46	24.05	2.60	9.22	1 45	-0.04	1.07	0.47	0.02	0.11	12 18
3	1	0.07	-0.01	0.01	0.14	0.90	0.00	0.50	1.83	4.53	0.73	11.82	22 77	2.58	9.03	1.39	-0.11	1 16	0.58	0.10	0.12	12.43
U U	2	0.10	0.00	0.00	0.10	0.87	0.01	0.51	1 79	4 40	0.75	12.02	23.37	2.50	9 10	1.34	-0.11	1 13	0.60	0.09	0.15	12.56
	3	0.08	0.00	0.01	0.09	0.89	0.00	0.51	1.86	4 4 1	0.76	11.96	23 10	2.50	9.16	1 41	-0.08	1 22	0.58	0.08	0.14	12.00
4	1	0.20	0.00	0.03	0.06	0.67	0.00	0.37	1.53	4 07	0.32	12 34	23.97	2.59	9.25	1 44	-0.05	1 13	0.50	0.06	0.09	12.25
5	1	0.25	0.00	0.01	0.04	0.57	0.01	0.40	1.51	4 66	0.27	12.50	23.56	2.55	9.25	1 47	-0.07	1 18	0.53	0.03	0.00	12 21
-	2	0.09	0.00	0.01	0.02	0.86	0.02	0.49	1 79	4 48	0.69	12 15	23 13	2.66	8.95	1 43	-0.08	1 14	0.59	0.09	0.09	12 47
	3	0.19	0.01	0.01	0.05	0.52	0.02	0.33	1 43	3 85	0.25	13 21	24 52	2.59	9 24	1.39	-0.08	1 02	0.46	0.05	0.13	12 17
	4	0.06	0.00	0.02	0.14	0.91	0.02	0.00	1 75	4.38	0.65	12 10	23.23	2.58	9.01	1.36	-0.10	1 18	0.57	0.07	0.14	12.54
6	1	0.26	0.00	0.01	0.15	0.58	0.01	0.41	1.48	4.67	0.34	12.27	23.40	2.63	9.36	1.46	-0.06	1.15	0.48	0.06	0.09	12.29
°	2	0.24	0.00	0.01	0.09	0.52	0.00	0.38	1 45	4 27	0.33	12.34	23.93	2.65	9.53	1 48	-0.07	1 14	0.55	0.05	0.12	12.26
	3	0.17	0.00	0.01	0.00	0.80	0.02	0.50	1.83	4 89	0.64	11.64	22 73	2.54	9.20	1 41	-0.12	1 15	0.56	0.06	0.11	12.39
	4	0.18	0.00	0.01	0.17	0.93	0.01	0.59	1.00	5.35	0.77	11.01	22.28	2 4 9	8.98	1.35	-0.05	1 18	0.55	0.10	0.13	12.55
7	1	0.10	0.00	0.02	0.03	0.00	0.01	0.00	1.54	4.58	0.36	12 41	23.84	2.40	9.32	1.50	-0.08	1.10	0.49	0.10	0.10	12.00
•	2	0.09	0.01	0.01	0.02	1.00	0.00	0.53	2 4 9	4 69	0.86	11.50	22.59	2.58	9 19	1 41	-0.23	1.30	0.70	0.11	0.21	12.53
8	1	0.09	0.00	0.00	0.00	0.68	0.00	0.40	1.56	3.69	0.61	12 38	23.62	2.63	9.41	1 4 9	-0.07	1.05	0.53	0.04	0.08	12.00
0	2	0.07	0.00	0.00	0.00	0.65	0.00	0.38	1.00	3.41	0.56	12.58	24.03	2.68	9.35	1.40	-0.06	1.00	0.50	0.07	0.00	12.47
	3	0.07	0.00	0.00	0.07	0.65	0.01	0.00	1.46	3 38	0.53	12.50	23.70	2.58	9.18	1 41	-0.07	1.00	0.02	0.05	0.11	12.47
q	1	0.07	0.00	0.00	0.02	0.03	0.00	0.07	1.73	3.50	0.65	12.00	23.53	2.50	9.36	1.35	-0.07	1.12	0.43	0.03	0.10	12.20
5	2	0.08	0.00	0.00	0.01	0.74	0.00	0.42	1.65	3.76	0.59	12.21	24.14	2.50	9.31	1 41	-0.07	1 16	0.52	0.07	0.10	12.50
	3	0.00	0.00	0.01	0.01	0.77	0.01	0.42	1 78	4 44	0.33	12.01	23.01	2.57	9.16	1 37	-0.07	1 12	0.52	0.07	0.11	12.33
10	1	0.05	0.00	0.00	0.01	0.78	0.00	0.48	1.81	3 74	0.78	12.00	23.57	2.55	9.15	1 39	-0.14	1 12	0.50	0.00	0.12	12.40
11	1	0.00	0.00	0.01	0.13	0.69	0.00	0.43	1 75	3 75	0.61	12.35	23.78	2.58	9 44	1 44	-0.08	1 12	0.55	0.00	0.10	12.00
	•	0.00	0.00	0.0.	0.0.	0.00	0.0.	00		00	0.0.		-00				0.00	··· · -	0.00	0.0.	0	

Grain/a	analysis	S	As	0	Total (wt%)	Pb (ppm)	Th (ppm)	U (ppm)	Age (Ma)	±1σ
5	1	0.00	0.03	26.43	97.67	3109	27168	4879	1513	28
6	1	0.01	0.00	26.60	96.29	1001	30813	1075	646	27
7	1	0.00	-0.12	30.03	100.88	1237	21070	1574	1028	36
8	1	0.00	0.00	26.41	97.88	3113	33429	3255	1501	28
	2	0.00	0.03	26.33	97.48	3095	33881	2950	1513	28
Sample	308: Da	niel Creek F	Formation (Gum Flats)						
1	1	0.00	0.00	28.01	100.45	5714	50089	9780	1464	17
2	1	0.01	0.02	26.98	100.84	14739	178379	8278	1537	9
	2	0.00	0.01	27.35	100.28	5695	44867	10331	1512	17
3	1	0.00	0.03	25.93	95.19	7716	66909	13572	1457	13
4	1	0.00	0.02	26.04	95.79	10056	69389	19957	1548	11
	2	0.00	0.02	25.90	95.55	7937	65681	13458	1515	13
5	1	0.00	0.02	25.91	95.64	2415	29820	1500	1492	36
	2	0.00	0.01	24.85	94.20	5587	80015	1686	1414	17
6	1	0.00	0.02	26.20	97.27	2545	36769	1182	1355	31
7	1	0.00	0.01	26.98	98.57	8575	69362	14992	1514	13
Sample	324: Ein	asleigh Me	tamorphics	Gum Flat	s)					
1	1	0.00	0.01	26.24	97.37	5025	48698	6013	1551	19
	2	0.00	0.01	26.50	97.92	5062	46273	6874	1548	19
	3	0.00	0.02	26.07	97.81	5911	68474	3939	1552	17
2	1	0.00	0.01	26.07	97.01	4099	40866	4302	1577	23
	2	0.00	0.03	26.17	97.13	3561	40152	3141	1505	25
3	1	0.00	0.01	26.18	96.91	5085	45301	7266	1546	19
	2	0.00	0.01	26.46	97.85	5130	43998	7470	1570	19
	3	0.00	0.01	26.29	97.51	5098	44104	7609	1549	19
4	1	0.00	0.02	26.11	97.02	3706	40721	3246	1536	25
5	1	0.00	0.01	26.10	97.22	4011	46564	2727	1544	23
	2	0.00	0.01	26.29	97.44	4966	44831	6923	1547	19
	3	0.00	0.01	26.06	97.50	3321	38455	2491	1522	27
-	4	0.00	0.00	26.38	97.55	4759	43830	6451	1542	20
6	1	0.00	0.01	26.20	97.31	4174	46660	3369	1542	24
	2	0.00	0.02	26.18	97.54	3844	42677	3290	1532	25
	3	0.00	0.01	26.21	96.97	5092	48922	6402	1537	20
_	4	0.00	0.02	26.46	97.48	5990	53494	7730	1595	18
7	1	0.00	0.01	26.36	98.18	4165	45763	3617	1539	24
	2	0.01	0.01	26.57	98.41	5355	46923	8558	1501	18
8	1	0.00	0.01	26.22	96.95	4035	36852	6064	1499	23
	2	0.00	0.02	26.22	97.15	3847	34123	5643	1536	25
	3	0.00	0.02	25.88	95.92	3759	33789	5286	1546	25
9	1	0.00	0.00	26.31	96.95	4075	34996	6488	1520	23
	2	0.00	0.03	26.58	98.56	4262	37577	5864	1575	23
40	3	0.00	0.03	26.21	96.71	4240	44369	4502	1522	23
10	1	0.00	0.02	26.75	98.47	4827	37444	//8/	1597	21
11	1	0.00	0.03	20.40	97.70	4303	3/400	6143	1507	23

APPENDIX 3:

Representative THERMOCALC outputs

Sample 282 Average P-T

Average P-T

Activities and their uncertainties

Activities an	d their u	uncertain	nties				
a 0 sd(a)/a 0	ру .00121 3 .71841 (gr 3.50e-5).91713	alm 0.340 0.15000	0.0353 0.40638	ann 0.0530 0.35554	east 0.0390 0.39308	mu 0.720 0.10000
a sd(a)/a 0	pa 0.296 .14868 (cel 0.0210).47619	an 0.280 0.15552	ab 0.820 0.05000) q 1.00) 0	H2O 1.00	
Independent s 1) gr + 2pa 2) 3east + 6 3) phl + eas 4) ann + 3an 5) phl + 3an	et of rea + 3q = 3a q = py + t + 6q = = gr + a = py + g	actions an + 2ab phl + 2r py + 2ce alm + mu gr + mu	+ 2H2O nu 21				
Calculations	for the :	independe	ent set	of reacti	ons		
$\begin{array}{c} (101 \ x(120) \ - \ P(T) \\ 1 \ 4.6 \\ 2 \ 5.1 \\ 3 \ 7.9 \\ 4 \ 5.7 \\ 5 \ 4.8 \end{array}$	sd(P) 0.97 3.14 2.43 1.15 1.43	a 142.97 13.93 59.19 -34.73 9.80	sd(a) 0.70 1.24 1.09 1.16 0.71	k -0.26761 0.01303 0.03692 0.12438 0.11083	8.562 -3.575 -4.185 -7.402 -7.110	ln_K 8.479 -0.985 -7.856 -4.911 -10.143	sd(ln_K) 1.076 1.453 1.320 1.104 1.323
Average PT (f	or x(H2O) = 1.0)					
a lsq 6	vP sd .9 1.5	avT 726	sd 45 0.	cor fit 718 1.13			
py 7. gr 7. alm 6. ph1 6. east 6. mu 6. pa 6. cel 6. an 7. ab 6. q 6.	P sd(P 46 1.3% 63 1.7% 75 1.4% 98 1.4% 92 1.5% 98 1.5% 90 1.5% 35 1.4% 19 1.6% 92 1.4%	T 748 731 730 730 730 730 730 730 730 730 730 730 730 730 730 730 730 730 730 723 726 728 726 726 726 728 726 726	sd(T) 43 42 46 46 47 45 46 59 40 44 44 46	cor 0.753 0.675 0.739 0.721 0.781 0.711 0.730 0.659 0.719 0.695 0.705 0.718	fit 0.90 -1. 1.06 -0. 1.10 0. 1.12 0. 1.13 0. 1.13 -0. 1.13 -0. 1.13 -0. 1.13 -0. 1.13 0. 1.13 -0. 1.13 0. 1.13 -0. 1.13 0. 1.13 0. 1.13 0. 1.13 0.	<pre>hat 13 0.20 53 0.50 30 0.05 29 0.07 71 0.26 01 0.09 13 0.04 05 0.39 20 0.17 27 0.13 02 0.04 0 0 0 0 0 0 0 0 0 0 0 0 0 0 0 0 0 0</pre>	

T = 726°C, sd = 45, P = 6.9 kbars, sd = 1.5, cor = 0.718, sigfit = 1.13

a sd(a)/a	0.0	ру 0146 3 0738 0	gr .40e-5 .92604	alm 0.370 0.15000	n ph 0.027 0.4377	nl 77 0. 77 0.4	ann 0220 6630	east 0.0520 0.35756	mu 0.720 0.10000
a sd(a)/a	0 0.1	pa .296 4868 C	cel 0.0210 .47619	ar 0.330 0.13467	0.80 0.80 0.0500	ab)0)0	q 1.00 0	H2O 1.00	
Independen 1) 3east 2) 2east 3) phl + 4) ann + 5) py + 2	t set + 6q + 6q 3an = 3an = pa +	of rea = py + = py + py + g gr + a 3cel =	nctions phl + 2r mu + cel pr + mu llm + mu 3east +	nu 1 2ab + 9	9q + 2H20)			
Calculatio	ns for (1) = 1	r the i	ndepende	ent set	of react	cions			
P(T) - 1	sd(P)	a	sd(a)		b	с	ln K	sd(ln K)
1 3	.2	2.83	13.91	1.24	0.0128	39 -3	.547	-1.903	1.372
2 6	.1	2.12	36.56	1.12	0.0248	34 -3	.856	-4.808	1.117
3 4	.5	1.34	9.11	0.71	0.1117	71 -7	.128	-10.235	1.313
4 5	. 6	1.11	-35.56	1.16	0.1254	10 -7	.418	-4.469	1.127
5 5	.5	1.51	57.33	2.02	-0.2186	59 9	.460	11.238	1.947
Average PT	(for	x(H2O)	= 1.0)						
	avP	sd	avT	sd	cor fi	lt			
lsq	7.3	2.3	768	75 0.	754 1.6	54			
	P	sd(P)	Т	sd(T)	cor	fit	e*	hat	
ру	8.56	1.89	817	64	0.801	1.22	-1.71	0.26	
gr	7.96	2.80	772	75	0.703	1.60	-0.46	0.52	
alm	7.13	2.29	759	76	0.767	1.59	0.43	0.03	
phl	7.39	2.34	770	79	0.758	1.64	0.15	0.09	
ann	6.23	2.14	721	74	0.808	1.36	-1.33	0.28	
east	7.56	2.36	770	74	0.748	1.61	0.53	0.10	
mu	7.45	2.34	773	78	0.765	1.62	-0.22	0.03	
pa	7.30	2.35	762	95	0.695	1.64	-0.10	0.34	
cel	6.61	2.07	759	64	0.749	1.39	1.56	0.17	
an	7.53	2.47	769	75	0.733	1.63	0.20	0.10	
ab	7.34	2.32	767	78	0.743	1.64	0.03	0.04	
q	7.34	2.31	. 768	75	0.754	1.64	0	0	
H2O	7.34	2.31	. 768	75	0.754	1.64	0	0	

 $T = 768^{\circ}C, sd = 75,$

P = 7.3 kbars, sd = 2.3, cor = 0.754, sigfit = 1.64

Sample 288

Average P-T (M1) Activities and their uncertainties alm phl ann east qr ру 0.000680 6.60e-5 0.350 0.0188 0.0540 0.0370 а sd(a)/a 0.74867 0.82873 0.15143 0.48501 0.35355 0.40013 н20 pa cel an ab q 0.660 0.0150 0.145 0.910 1.00 1.00 а sd(a)/a 0.10061 0.66667 0.21931 0.05011 0 Independent set of reactions 1) phl + east + 6q = py + 2cel 2) 2east + 6q = py + mu + cel 3) phl + 3an = py + gr + mu4) ann + 3an = gr + alm + mu 5) py + 2pa + 3cel = 3east + 2ab + 9g + 2H2O Calculations for the independent set of reactions (for x(H2O) = 1.0)P(T) sd(P) sd(a) b С ln K sd(ln K) а 1 7.7 2.77 59.16 1.09 0.03650 -4.108 -8.422 2 5.3 2.33 36.53 1.12 0.02455 -3.804 -5.330 3 6.9 1.34 8.20 0.71 0.11276 -7.126 -7.583 4 6.9 1.05 -36.59 1.16 0.12658 -7.414-2.395 5 5.3 1.86 56.79 2.02 -0.21602 9.130 10.644 Average PT (for x(H20) = 1.0) avP sd cor fit sd avT lsq 7.6 1.3 642 31 0.716 1.02 P sd(P) Т sd(T) cor fit e* hat 8.08 1.36 34 0.750 0.84 -0.95 0.19 656 ру 7.20 1.65 640 31 0.669 0.99 0.29 0.45 qr alm 7.55 1.34 639 33 0.738 1.01 0.17 0.06 phl 7.59 1.29 638 32 0.711 0.98 -0.46 0.07 ann 7.29 1.44 629 39 0.783 0.97 -0.40 0.34 east 8.07 1.33 646 31 0.713 0.73 1.20 0.08 mu 7.73 1.34 645 33 0.734 1.01 -0.16 0.05 7.63 1.33 640 38 0.678 1.02 -0.06 0.22 ра cel 7.51 1.33 641 31 0.712 0.99 0.44 0.11 an 7.33 1.56 640 31 0.677 1.00 -0.23 0.28 ab 7.64 1.32 642 33 0.703 1.02 0.03 0.06 q 7.65 1.31 642 31 0.716 1.02 0 н20 7.65 1.31 642 31 0.716 1.02 0

 $T = 642^{\circ}C$, sd = 31, P = 7.6 kbars, sd = 1.3, cor = 0.716, sigfit = 1.02 Average P-T (M2)

mu

0.650

1.653

1.287

1.388

1.130

2.460

0

0

0.10000

Activities and their uncertainties

a sd(a)/a	0.000	ру 410 125 0.	alm 0.340 15000	ph] 0.0348 0.40826	L ar 3 0.053 5 0.3508	nn 30 0. 36 0.3	east 0640 3482 (clin 0.0105 0.54900	daph 0.0520 0.35756				
a sd(a)/a	a 0.0 0.48	imes 110 074 0.3	mu 0.730 15000	cel 0.0180 0.55556	L) 1.(0 0 9	H2O 1.00						
••INCOMPLE 1) 2east 2) py + 2 3) 10east	TE•• i + 6q = 2ann + : + 3da	ndepende py + m 3ames + ph + 30e	ent set u + ce] 6q = 2 q = 5a]	c of rea L 2alm + 3 Lm + 3cl	actions Bclin + 2 Lin + 5mu	2mu 1 + 5ce	1						
Calculatic (for x(H2C	ons for)) = 1.	the ind 0)	depende	ent set	of react	cions							
P (T) s	d(P)	а	sd(a)		b	C	ln_K	sd(ln_K)				
1 3	1.4	1.97	35.53	1.12	0.0261	16 -3	.871	-6.634	1.172				
2 3	5.0	5.20 -	68.25	2.50	0.0109	91 -3	.082	10.748	2.462				
3 2	.6	1.69 -	41.96	7.87	0.1540	ol -19	.191	-4.365	4.890				
Average PI	Average PT (for $x(H20) = 1.0$)												
	avP	sd	avT	sd	cor f	it							
lsq	3.1	1.7	557	117 0	.189 0.4	12							
	P	sd(P)	т	sd(T)	cor	fit	e*	hat					
ру	2.97	2.07	539	204	0.541	0.40	0.06	0.79					
alm	3.09	1.75	556	122	0.141	0.42	-0.01	0.04					
ann	3.05	1.74	557	117	0.185	0.36	-0.14	0.01					
east	3.39	2.45	557	117	0.142	0.38	0.11	0.42					
clin	2.93	1.85	573	133	-0.013	0.33	0.18	0.28					
daph	3.12	1.74	550	125	0.135	0.39	0.09	0.07					
ames	3.01	1.75	558	117	0.180	0.27	-0.28	0.02					
mu	3.06	1.80	557	117	0.178	0.41	0.03	0.03					
cel	3.30	2.28	557	117	0.150	0.39	-0.09	0.29					
q	3.09	1.13	557	117	0.189	0.42	0	U					
T = 557°C,	sd =	117,											

P = 3.1 kbars, sd = 1.7, cor = 0.189, sigfit = 0.42

Sample 293

Average P-T (M1)

Activities and their uncertainties

qr alm mu pa cel an ру 0.000165 0.0170 0.0510 0.163 0.0590 0.510 0 800 а 0.80398 0.49667 0.35625 0.15000 0.21332 0.26564 0.05025 sd(a)/a ab phl ann east сz ep fep 0.460 0.0350 0.0770 0.0310 0.240 0.710 0.0560 а 0.08748 0.39650 0.31321 0.42337 0.17328 0.10000 0.34963 sd(a)/a H2O q 1.00 1.00 а sd(a)/a Ο Independent set of reactions 1) 4gr + 5pa + 6q = 5ab + 6cz + 2H202) 2east + 6q = py + mu + cel 3) 3an + phl = py + gr + mu4) cel + 3an + east = py + gr + 2mu 5) 2py + 3mu + 24cz = 5qr + 33an + 3east + 12H2O 6) 3an + ann = gr + alm + mu7) alm + mu + 8ep + q = 8an + ann + 4fep + 2H2O Calculations for the independent set of reactions (for x(H2O) = 1.0)P(T) sd(P) а sd(a) b С ln_K sd(ln_K) 1 7.8 3.11 66.08 2.59 -0.23594 5.952 12.923 2.521 2 5.2 2.30 36.78 1.12 0.02446 -3.834 -5.266 1.207 3 0.71 0.11179 -7.118 -9.436 5.4 1.07 8.98 1.047 4 32 1.20 -13.64 0.76 0.09984 -6.813 -7.157 1 121 0.57 1148.36 7.62 -1.85662 71.345 15.532 5.533 5 5.0 -7.408 0.72 -35.67 1.16 0.12546 -4.490 0.719 6 5.6 7 0.77 331.35 4.67 -0.41406 17.430 -9.490 1.733 57 Average PT (for x(H20) = 1.0) avP sd avT sd cor fit 624 69 0.929 2.01 lsq 6.3 1.8 P sd(P) Т sd(T) fit cor e* hat 6.71 73 0.941 1.95 -1.06 0.19 1.93 641 ру 6.39 1.91 628 71 0.934 2.00 -0.40 0.12 qr 6.17 621 77 0.943 2.01 0.14 0.18 alm 2.08 62 0.932 mu 6.74 1.64 640 1.76 -1.46 0.03 6.66 1.76 651 72 0.921 1.88 1.27 0.29 pa 593 46 0.934 1.31 3.33 0.07 cel 5.41 1.23 6.44 1.84 623 68 0.889 1.98 0.48 0.22 an 631 70 0.926 1.97 -0.52 0.05 6.37 1.80 ab 67 0.937 1.84 -1.48 0.09 1.77 602 phl 5.68 76 0.942 2.05 621 2.01 -0.12 0.14 6.18 ann 69 0.929 6.30 1.81 625 1.99 0.56 0.01 east 68 0.894 сz 6.43 1.82 622 1.97 -0.75 0.36 6.29 1.83 626 71 0.929 2.01 0.16 0.04 ep fep 6.31 1.84 627 72 0.928 2.00 -0.27 0.11

q 6.26 1.82 69 0.929 2.01 0 0 624 6.26 1.82 69 0.929 0 0 Н2О 624 2 01 $T = 624^{\circ}C$, sd = 69, P = 6.3 kbars, sd = 1.8, cor = 0.929, sigfit = 2.01 Average P-T (M1R) Activities and their uncertainties vα gr alm phl ann east сz 0.000178 0.0120 0.0490 0.0320 0.0890 0.0270 0.280 а sd(a)/a 0.80159 0.53508 0.36184 0.41925 0.29541 0.44100 0.15552 ep fep an ab mu pa cel 0.680 0.0470 0.800 0.450 0.610 а 0.239 0.0310 0.10000 0.36805 0.10000 0.09075 0.10000 0.17374 0.32258 sd(a)/a н20 q 1.00 1.00 а Ω sd(a)/a Independent set of reactions 1) 2east + 6q = py + mu + cel 2) py + gr + mu = phl + 3an3) py + gr + east + cel = 2phl + 3an 4) 5py + 12cz + 15cel = 8gr + 15east + 57g + 6H2O 5) alm + 3east + 6q = 2py + ann + 2mu 6) py + 2pa + 3cel = 3east + 2ab + 9q + 2H2O 7) 2py + 12gr + 18fep + 19pa + 6cel = 6east + 36ep + 19ab + 10H20 Calculations for the independent set of reactions (for x(H2O) = 1.0)P(T)sd(P) sd(a) b ln_K sd(ln_K) а С 2.00 35.45 1.12 0.02569 -3.766 -5.378 1.239 1 5.8 0.93 0.71 -0.11470 7.142 2 4.6 -6.79 9.439 1.098 3 1.15 -29.42 0.80 -0.12665 7.446 12.589 1.421 6.1 11.03 -0.53000 32.877 4 5.0 1.90 94.13 20.989 10.249 5 5.2 4.08 57.91 1.55 -0.00043 -3.177 -6.823 2 140 2.02 -0.21643 6 4.6 1.29 58.68 8.841 9.485 1.866 2.02 -40.61 21.58 -1.25471 34.993 122.689 11.192 7 6.0 Average PT (for x(H2O) = 1.0) avP avT sd cor fit sd 35 0.889 1.18 559 lsq 5.1 1.0 т sd(T) fit P sd(P) cor ~ م hat 38 0.908 ру 5.17 1.07 562 1.17 -0.21 0.20

5.08

5.25

1.02

1.12

gr

alm

559

564

36 0.895

1.18

39 0.916 1.17 -0.26 0.18

0.02 0.09
phl	4.77	0.91	547	32	0.900	1.04	-1.04	0.08
ann	5.21	1.09	563	38	0.911	1.17	0.21	0.12
east	5.12	0.97	560	34	0.889	1.16	0.44	0.02
CZ	5.17	1.00	559	34	0.861	1.16	-0.31	0.19
ep	5.07	0.98	558	35	0.890	1.18	-0.11	0.02
fep	5.06	0.99	557	36	0.891	1.17	0.20	0.07
an	5.42	1.05	559	33	0.788	1.13	0.50	0.52
ab	5.18	0.97	565	35	0.885	1.15	-0.39	0.07
mu	5.24	0.94	565	33	0.894	1.11	-0.55	0.02
pa	5.33	0.94	576	36	0.882	1.09	0.74	0.27
cel	4.70	0.85	548	30	0.893	0.79	2.01	0.06
q	5.09	0.98	559	35	0.889	1.18	0	0
H2O	5.09	0.98	559	35	0.889	1.18	0	0

T = 559°C, sd = 35, P = 5.1 kbars, sd = 1.0, cor = 0.889, sigfit = 1.18

Sample 306

Average P-T

Activities and their uncertainties

a sd(a)/a	phl 0.0427 0.37765	ann 0.0180 0.49008	east 0.0680 0.32788	ру 0.000920 0.73345	alm 0.380 0.15000	clin 0.0240 0.45567	daph 0.0260 0.44574			
a sd(a)/a	ames 0.0320 0.40454	mu 0.780 0.15000	cel 0.0130 0.76923	mst 0.000330 1.10096	fst 0.560 0.20000	9 1.00 0	H2O 1.00			
<pre>Independent set of reactions 1) 27phl + 4mst = 21east + 13py + 2clin + 6cel 2) phl + east + 6q = py + 2cel 3) 16east + ames + 39q = 6phl + 10cel + 2mst 4) 5phl + 13cel + 2mst = 18east + 46q + 4H20 5) 2phl + mu + 6q = py + 3cel 6) 97phl + 12fst = 16ann + 63east + 39py + 6clin + 18cel 7) 53phl + 10ann + 12mst = 45east + 39py + 6daph + 18mu 8) ann + 3east + 6q = 2phl + alm + 2mu</pre>										
Calculation	s for the $= 1 0$	independ	ent set (of reactio	ons					
P(T) sd(P)	a	sd(a)	b	С	ln K	sd(ln K)			
1 4.	3 32.44	664.67	9.57	-0.29716	3.566	-63.643	16.847			
2 5.	3 2.97	59.16	1.09	0.03650	-4.108	-9.835	1.776			
3 5.	2 2.41	55.14	8.82	0.37536	-28.142	-31.928	9.841			
4 5.	3 2.40	109.29	10.40	-0.68364	34.386	39.869	11.969			
5 6.	7 3.95	81.78	1.14	0.04845	-4.413	-13.464	2.541			
6 3.	8 33.96	2910.55	29.62	-1.14294	10.798	-293.988	53.416			
7 4.	2 21.53	1576.22	29.44	-1.01367	12.981	-116.486	40.615			

	8	5.4	2.52	-30.89	1.55	0.02643	-3.787	4.311	1.375
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Average PT (for x(H20) = 1.0)

	avP	sd	avT	sd	cor	fit
lsq	6.3	2.5	574	18 0	.297	1.45

	P	sd(P)	Т	sd(T)	cor	fit	e*	hat
phl	6.26	2.50	575	19	0.282	1.45	-0.30	0.06
ann	4.97	2.10	571	15	0.318	1.15	-2.19	0.12
east	8.26	2.76	572	16	0.177	1.30	1.33	0.43
ру	6.66	2.46	579	19	0.344	1.39	-1.10	0.16
alm	5.65	2.57	567	20	0.409	1.39	0.68	0.16
clin	6.75	2.38	571	17	0.243	1.35	-1.21	0.10
daph	6.46	2.45	581	21	0.313	1.41	0.71	0.28
ames	6.33	2.43	578	19	0.294	1.40	0.86	0.09
mu	6.12	2.78	574	18	0.192	1.45	0.13	0.15
cel	6.11	2.68	574	18	0.318	1.45	0.28	0.17
mst	5.98	2.49	571	18	0.335	1.41	0.89	0.11
fst	6.11	2.49	573	18	0.304	1.43	-0.41	0.01
q	6.28	2.51	574	18	0.297	1.45	0	0
Н2О	6.28	2.51	574	18	0.297	1.45	0	0

T = 574°C, sd = 18, P = 6.3 kbars, sd = 2.5, cor = 0.297, sigfit = 1.45

Sample 312

Average P-T (M1)

Activities and their uncertainties

	an	tr	fact	ts	CZ	ep	fep				
a	0.650	0.122	0.00310	0.000100	0.300	0.430	0.0280				
sd(a)/a	0.05000	0.25406	0.70389	2.00000	0.14700	0.10000	0.41600				
	ру	gr	alm	р	H2O						
a	6.90e-5	0.0570	0.0580	1.00	1.00						
sd(a)/a	2.40772	0.34057	0.33810	0							
Independent set of reactions											
1) $2cz + py + q = 2an + ts$											
2) $5ts + 12cz + 14q = 28an + 3tr + 8H20$											
3) 6cz + 5	ipy + 4gr	+ 15q = 1	8an + 3t	r							
4) 6cz + 4	lgr + 5alm	1 + 15g =	18an + 31	fact							
5) 12ep +	4gr + 5al	.m + 15g =	18an + 3	3fact + 6f	ep						
Calculation	Calculations for the independent set of reactions										
(for x(H2O)	= 1.0)										
P(I) sd(P)	a	sd(a)	b	С	ln_K	sd(ln_K)				
1 8.	4 4.06	17.73	2.22	-0.08961	5.969	1.917	3.145				

2 4.2 1.61 737.93 11.31 -1.37312 49.440 42.126 10.279 3 4.3 2.39 21.84 2.99 -0.64277 39.258 52.524 12.205 4 7.5 0.63 327.07 6.19 -0.75590 41.415 7.836 3.280 5 7.3 0.81 477.43 9.00 -0.75528 41.258 -10.713 4.202	3 3.9 0.99 759.03 7.21 -1.74575 79.578 54.330 11.410 4 5.6 0.56 310.32 6.19 -0.73863 41.698 9.651 3.264 5 5.4 0.74 461.51 9.00 -0.73916 41.570 -11.403 4.267
Average PT (for x(H2O) = 1.0)	Average PT (for x(H2O) = 1.0)
avP sd avT sd cor fit lsg 7.1 1.8 649 690.942 1.37	avP sd avT sd cor fit lsq 6.4 1.7 631 69 0.936 1.41
P sd(P) T sd(T) cor fit e* hat an 7.17 1.81 648 69 0.913 1.37 0.05 0.18 tr 7.17 1.75 649 69 0.942 1.36 0.22 0.00 fact 6.44 2.15 627 78 0.955 1.31 -0.46 0.31 ts 7.29 1.29 652 51 0.942 0.98 -1.90 0.01 cz 7.14 1.77 651 72 0.889 1.37 0.12 0.51 ep 7.10 1.81 646 72 0.943 1.37 -0.13 0.04 fep 7.03 1.85 643 74 0.946 1.36 0.27 0.17 py 7.54 1.63 663 64 0.946 1.22 -1.21 0.09 gr 7.13 2.35 648 96 0.967 1.37 0.01 0.38 alm 6.66 2.05 633 76 0.952 1.33 0.37 0.20 q 7.15 1.77 649 69 0.942 1.37 0 0 H20 7.15 1.77 649 69 0.942 1.37 0 0	P sd(P) T sd(T) cor fit e* hat an 6.46 1.75 631 69 0.906 1.40 0.06 0.17 tr 6.46 1.70 632 68 0.936 1.39 0.22 0.00 fact 5.59 1.99 605 74 0.950 1.30 -0.61 0.29 ts 6.57 1.24 634 50 0.936 1.01 -1.92 0.01 cz 6.45 1.71 627 72 0.877 1.40 -0.18 0.52 ep 6.48 1.74 634 71 0.938 1.40 0.12 0.04 fep 6.54 1.78 637 74 0.940 1.40 -0.27 0.17 py 6.83 1.62 646 65 0.941 1.27 -1.14 0.12 gr 6.10 2.25 616 95 0.964 1.39 0.20 0.38 alm 5.83 1.93 612 73 0.947 1.33 0.49 0.19 q 6.44 1.72 631 69 0.936 1.41 0 0 H20 6.44 1.72 631 69 0.936 1.41 0 0 T = 631°C, sd = 69, P = 6.4 kbars, sd = 1.7, cor = 0.936, sigfit = 1.41
	Average P-T (M1R)
Average P-T (MI)	Activities and their uncertainties
Activities and their uncertainties an tr fact ts cz ep fep a 0.700 0.131 0.00340 0.000100 0.290 0.450 0.0209 sd(a)/a 0.05000 0.25038 0.69188 2.00000 0.15123 0.10000 0.43906	py gr alm an cz ep fep a 0.000133 0.0580 0.0600 0.720 0.400 0.560 0.0244 sd(a)/a 0.81046 0.33810 0.33326 0.15000 0.10800 0.10000 0.42719 tr fact g H20
py gr alm q H2O a 8.70e-5 0.0570 0.0580 1.00 1.00 sd(a)/a 2.14423 0.34057 0.33810 0	a 0.145 0.00350 1.00 1.00 sd(a)/a 0.25034 0.68808 0
<pre>Independent set of reactions 1) 2cz + py + q = 2an + ts 2) 6cz + 5py + 4gr + 15q = 18an + 3tr 3) 22cz + 5py + 19q = 38an + 3tr + 8H20 4) 6cz + 4gr + 5alm + 15q = 18an + 3fact </pre>	<pre>Independent set of reactions 1) 5py + 4gr + 6cz + 15q = 18an + 3tr 2) 21an + 6tr = 10py + 11gr + 27q + 6H2O 3) 4gr + 5alm + 6cz + 15q = 18an + 3fact 4) 4gr + 5alm + 12ep + 15q = 18an + 6fep + 3fact</pre>
<pre>5) 12ep + 4gr + 5alm + 15q = 18an + 3fact + 6fep Calculations for the independent set of reactions (for x(H2O) = 1.0)</pre>	$ \begin{array}{c} \mbox{Calculations for the independent set of reactions} \\ (for x(H20) = 1.0) \\ P(T) & sd(P) & a & sd(a) & b & c & ln_K \; sd(ln_K) \\ 1 & 4.3 & 0.89 & 4.04 & 2.99 & -0.62442 & 39.558 & 49.807 & 5.150 \\ 2 & 4.0 & 1.34 \; 558.15 & 4.74 & 0.40786 & -49.102 & -102.087 & 9.576 \\ 3 & 5.8 & 0.68 \; 310.32 & 6.19 & -0.73863 & 41.698 & 8.076 & 4.071 \\ 4 & 5.6 & 0.84 \; 461.51 & 9.00 & -0.73916 & 41.570 & -12.743 & 4.916 \\ \end{array} $

Average PT (for x(H20) = 1.0)

Average PT (for x(H20) = 1.0)

avP sd avT sd cor fit

lsq	avP 5.3	sd 1.4	avT 562	sd 51 0.	cor 795 1	fit .47				
py	P 6.22	sd(P) 1.10	Т 596	sd(T) 41	cor 0.854	fit 0.97	e* -1.39	hat 0.35		
gr	4.81	1.62	538	66	0.876	1.38	0.47	0.41		
alm	4.96	1.19	552	43	0.805	1.21	0.84	0.06		
an	5.47	1.74	561	51	0.539	1.46	-0.15	0.74		
C2 en	5.34	1 38	559	52	0.735	1 46	-0.15	0.15		
fep	5.27	1.39	559	53	0.803	1.45	0.26	0.02		
tr	5.39	1.36	564	50	0.801	1.43	0.26	0.01		
fact	4.85	1.12	550	40	0.808	1.12	-1.04	0.09		
q	5.32	1.37	562	51	0.795	1.47	0	0		
Н2О	5.32	1.37	562	51	0.795	1.47	0	0		
T = 562°C, sd = 51, P = 5.3 kbars, sd = 1.4, cor = 0.795, sigfit = 1.47										
Average P-T (M1R)										
Activities and their uncertainties										
a sd(a)/a	0.000 0.81	ру 107 0. 657 0.3	gr 0570 4057	alm 0.0600 0.33326	0. 0.15	an 700 000 0.3	cz 0.410 10443	ep 0.550 0.10000	fep 0.0330 0.40181	
a sd(a)/a	0. 0.25	tr 137 0.0 036 0.6	fact 0320 9978	q 1.00 0	1	H2O .00				
<pre>Independent set of reactions 1) 5py + 4gr + 6cz + 15q = 18an + 3tr 2) 5py + 22cz + 19q = 38an + 3tr + 8H20 3) 4gr + 5alm + 6cz + 15q = 18an + 3fact 4) 4gr + 5alm + 12ep + 15q = 18an + 6fep + 3fact</pre>										
Calculation (for x(H20	ons for)) = 1.	the ind 0)	epende	nt set	of rea	ctions				
P	(T) s	d(P)	a	sd(a)		b	С	ln_K	sd(ln_K)	
1 4	4.8	1.01 2	1.84	2.99	-0.64	277 3	9.258	50.138	5.174	
2	5.3	0.72 82	6.57	7.21	-1.82	119 7	9.288	45.812	7.416	
3	/.6 7 5	0.77 32	7.07	6.19	-0.75	590 4	1.415 1.250	7.221	4.089	
т	1.5	0.93 4/	/.45	9.00	-0.75	JZO 4.	1.200	-11,422	4.000	

lsq	5.5	1.6	564	60	0.800	1.74
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	P	sd(P)	Т	sd(T)	cor	fit	e*	hat
ру	6.56	1.28	605	47	0.857	1.15	-1.64	0.35
gr	5.10	2.01	546	82	0.878	1.70	0.35	0.41
alm	5.10	1.50	555	54	0.810	1.51	0.86	0.06
an	5.62	2.08	563	61	0.545	1.73	0.11	0.75
CZ	5.47	1.65	565	62	0.744	1.74	0.06	0.14
ep	5.39	1.61	559	60	0.804	1.68	-0.43	0.02
fep	5.30	1.57	555	59	0.809	1.62	0.85	0.09
tr	5.56	1.63	567	60	0.806	1.70	0.30	0.01
fact	4.98	1.45	552	51	0.813	1.43	-1.08	0.09
q	5.48	1.64	564	60	0.800	1.74	0	0
Н2О	5.48	1.64	564	60	0.800	1.74	0	0

 $T = 564^{\circ}C$, sd = 60,

P = 5.5 kbars, sd = 1.6, cor = 0.800, sigfit = 1.74

Appendix 6: Sample Locations

			GDA 94 arid	co-ordinate				GDA 94 arid	co-ordinate
Sample #	Domain	Unit	Easting	Northing	Sample #	Domain	Unit	Easting	Northing
6	Ortona	Cobbold Metadolerite	0764910	7876412	197	Stockvard	Cobbold Metadolerite	0735282	7941168
7	Ortona	Bernecker Creek Formation	0764782	7876443	199	Stockyard	Cobbold Metadolerite	0735252	7941109
8	Ortona	Cobbold Metadolerite	0764822	7876539	200	Stockyard	Cobbold Metadolerite	0735260	7941052
10	Ortona	Bernecker Creek Formation	0764782	7876533	201	Stockyard	Cobbold Metadolerite	0735261	7941003
13	Ortona	Bernecker Creek Formation	0764531	7876765	202	Stockyard	Cobbold Metadolerite	0735229	7940955
22	Ortona	Cobbold Metadolerite	0764081	7877964	208	Robinhood	Ropewalk Granite	0766287	7952017
23	Ortona	Cobbold Metadolerite	0763573	7879147	209	Robinhood	Ropewalk Granite	0769766	7938112
60	Ortona	Cobbold Metadolerite	0766289	7893968	210	Robinhood	Ropewalk Granite	0769256	7936256
72	Ortona	Dead Horse Metabasait	0780166	7866980	215	Robinnood	Cobbold Metadolerite	0766619	7936146
74 80	Ortona	Dead Horse Metabasalt	0758021	7883022	221	South Head	Cobbold Metadolerite	0745832	7904657
82	Ortona	Dead Horse Metabasalt	0758968	7882795	224	South Head	Cobbold Metadolerite	0745760	7904752
83	Ortona	Dead Horse Metabasalt	0758990	7882593	226	South Head	Cobbold Metadolerite	0745706	7904811
85	Ortona	Cobbold Metadolerite	0759485	7882209	227	South Head	Cobbold Metadolerite	0745698	7904896
86	Ortona	Cobbold Metadolerite	0759644	7882155	228	South Head	Dead Horse Metabasalt	0745690	7904962
87	Ortona	Dead Horse Metabasalt	0759676	7881475	229	South Head	Dead Horse Metabasalt	0745600	7905105
88	Ortona	Cobbold Metadolerite	0759735	7881315	230	South Head	Dead Horse Metabasalt	0745525	7905294
89	Ortona	Dead Horse Metabasalt	0759838	7881284	232	South Head	Dead Horse Metabasalt	0745399	7905478
90	Ortona	Dead Horse Metabasalt	0759889	7881014	233	South Head	Dead Horse Metabasalt	0745299	7905707
97	Ortona	Daniel Creek Formation	0760433	7879591	234	South Head	Dead Horse Metabasalt	0745128	7906067
100	Ortona	Dead Horse Metabasalt	0761691	7880234	236	South Head	Dead Horse Metabasalt	0744865	7906543
101	Ortona	Cobbold Metadolerite	0761714	7880144	237	South Head	Cobbold Metadolerite	0744716	7906815
102	Ortona	Dead Horse Metabasalt	0761755	7880084	240	South Head	Dead Horse Metabasalt	0744501	7907232
103	Ortona	Dead Horse Metabasalt	0761806	7879995	241	South Head	Cobbold Metadolerite	0744406	7907452
104	Ortona	Dead Horse Metabasalt	0761837	7879961	242	South Head	Cobbold Metadolerite	0744302	7907651
106	Ortona	Cobbold Metadolerite	0762360	7879578	243	South Head	Dead Horse Metabasalt	0744217	7907752
107	Ortona	Cabbald Matadalarita	0762299	7879600	245	South Head	Cabbald Matadalarita	0737343	7911863
100	Ortona	Cobbold Metadolerite	0762155	7079024	249	South Head	Cobbold Metadolenite	0741346	7900095
103	Ortona	Cobbold Metadolerite	0762008	7879755	250	South Head	Dead Horse Metabasalt	0741270	7906880
110	Ortona	Dead Horse Metabasalt	0761929	7879749	253	South Head	Dead Horse Metabasalt	0741010	7907251
112	Ortona	Cobbold Metadolerite	0765456	7879961	255	South Head	Daniel Creek Formation	0740889	7907568
113	Ortona	Cobbold Metadolerite	0765188	7879846	256	South Head	Dead Horse Metabasalt	0740889	7907568
114	Ortona	Cobbold Metadolerite	0765157	7879755	260	Robinhood	Dead Horse Metabasalt	0765214	7911170
115	Ortona	Cobbold Metadolerite	0765216	7879873	261	Robinhood	Dead Horse Metabasalt	0765269	7911303
120	Ortona	Cobbold Metadolerite	0764776	7879884	262	Robinhood	Dead Horse Metabasalt	0765424	7911326
121	Ortona	Cobbold Metadolerite	0764439	7879949	264	Robinhood	Dead Horse Metabasalt	0765535	7911389
124	Ortona	Bernecker Creek Formation	0760673	7877719	268	Robinhood	Dead Horse Metabasalt	0765747	7911590
128	Ortona	Cobbold Metadolerite	0764030	7876727	269	Robinhood	Dead Horse Metabasalt	0765675	7911505
129	Ortona	Cobbold Metadolerite	0764012	7876729	270	Robinhood	Dead Horse Metabasalt	0765625	7911442
130	Ortona	Cobbold Metadolerite	0764012	7876729	278B	Ortona	Corbett Formation	0763205	7881161
135	Gum Flats	Mt Hogan Granite	0798215	7881440	282	Gum Flats	Einasleigh Metamorphics	0806726	7880850
148	Ortona	Dead Horse Metabasalt	0764368	7871563	283	Gum Flats	Einasleign Metamorphics	0806706	7880850
151	Ortona	Cobbold Motodolorito	0764327	7071479	200	Gum Flats	Cobbold Motodolorito	0803645	7001020
152	Ortona	Dead Horse Metabasalt	0764252	7871356	202	Gum Flats	Bernecker Creek Formation	0803612	7885150
154	Ortona	Dead Horse Metabasalt	0764232	7871333	292	Gum Flats	Bernecker Creek Formation	0803330	7884267
159	Ortona	Cobbold Metadolerite	0765618	7869966	296	Gum Flats	Cobbold Metadolerite	0803182	7882632
161	Ortona	Cobbold Metadolerite	0765618	7869966	299	Gum Flats	Cobbold Metadolerite	0803317	7881548
172	Ortona	Cobbold Metadolerite	0765141	7870249	300	Gum Flats	Cobbold Metadolerite	0803334	7881548
174	Ortona	Cobbold Metadolerite	0763625	7878064	301	Gum Flats	Cobbold Metadolerite	0803598	7881489
175	Ortona	Cobbold Metadolerite	0763688	7878004	304	Robinhood	Lane Creek Formation	0766619	7936146
178	Ortona	Cobbold Metadolerite	0763728	7878008	306	Ortona	Corbett Formation	0766403	7908517
179	Ortona	Cobbold Metadolerite	0763728	7878008	307	Gum Flats	Bernecker Creek Formation	0802540	7885152
181	Ortona	Cobbold Metadolerite	0763628	7878039	308	Gum Flats	Daniel Creek Formation	0801829	7887568
184	Ortona	Daniel Creek Formation	0763245	7882389	311	Gum Flats	Daniel Creek Formation	0800748	7888073
188	Ortona	Cobbold Metadolerite	0763367	7881600	312	Gum Flats	Daniel Creek Formation	0800551	7888325
192	Stockyard	Cobbold Metadolerite	0735451	7941350	317	Gum Flats	Einasleigh Metamorphics	0800000	7892233
194	Stockyard	Cobbold Metadolerite	0735421	7941321	318	Gum Flats	Einasleigh Metamorphics	0799901	7892058
195	Stockyard	Cobbold Metadolerite	0735378	7941237	324	Gum Flats	Einasleigh Metamorphics	0798867	7893315
196	Stockyard	CODDOID Metadolerite	0735320	7941215					

