Earth's Oldest Rocks
Edited by Martin J. Van Kranendonk, R. Hugh Smithies and Vickie C. Bennett
Developments in Precambrian Geology, Vol. 15 (K.C. Condie, Series Editor)

DOI: 10.1016/S0166-2635(07)15032-5

Chapter 3.2

ANCIENT ANTARCTICA: THE ARCHAEAN OF THE EAST ANTARCTIC SHIELD

SIMON L. HARLEY AND NIGEL M. KELLY

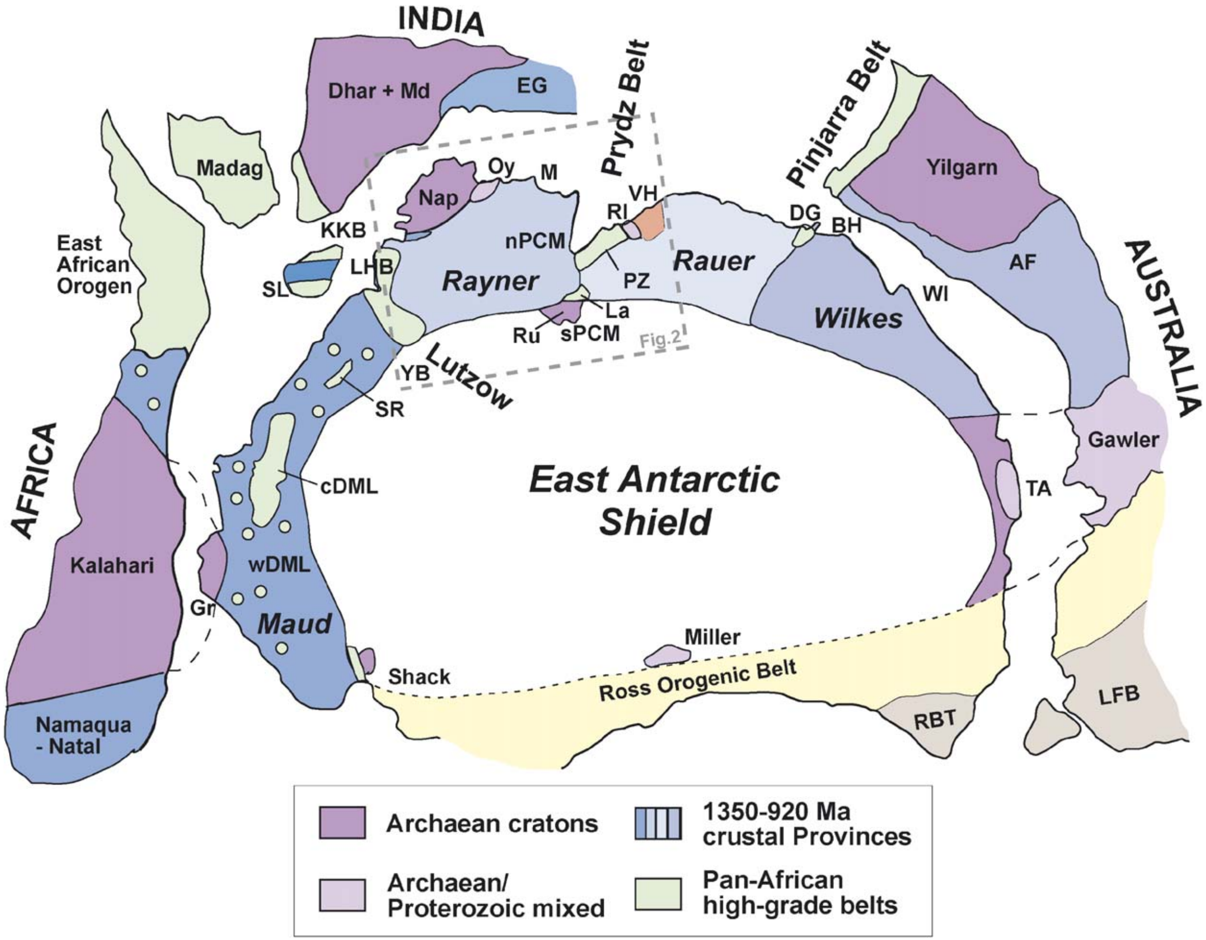
Grant Institute of Earth Science, School of GeoSciences, University of Edinburgh, Kings Buildings, West Mains Road, UK

3.2-1. INTRODUCTION

The East Antarctic Shield comprises most of the main landmass of Antarctica, bounded by the Transantarctic Mountains and the Southern Ocean in the sector from Africa to east Australia (Fig. 3.2-1). Despite less than 0.5% of its land area consisting of exposed rock, the East Antarctic Shield preserves a remarkable record of Earth evolution that spans in time from the earliest Archaean (ca. 3850–4060 Ma) to the Cambrian. The East Antarctic Shield preserves distinct high-grade terrains that were situated adjacent to Africa, India and Australia within Gondwana and in continental reconstructions proposed for earlier time periods (e.g., Dalziel, 1991). As such, it has become a major focus for testing models of supercontinent formation and destruction (Rogers, 1996; Unrug, 1997; Fitzsimons, 2000a), as well as an important region for investigating the nature of high-grade metamorphism

---

Fig. 3.2-1. (*Next page*.) Map of East Antarctica in its reconstructed Gondwana context, modified from Fitzsimons (2000a, 2000b) and Harley (2003) and drawn with latitudes and longitudes referred to the present Antarctic co-ordinates. The Meso- to Neoproterozoic provinces recognised by Fitzsimons (2000a) are labelled (Wilkes, Rayner and Maud), along with a fourth province (Rauer Province) defined by Harley (2003). Latest Neoproterozoic to Cambrian aged belts (Lutzow, Prydz and Pinjarra) are also distinguished. The Maud Province is extensively reworked and reorganised in the Neoproterozoic–Cambrian, in the East Africa–Madagascar–Antarctica belt, shown here by dotted ornament. Specific areas mentioned in the text are labelled with the following abbreviations: AF (Albany-Fraser); BH (Bunger Hills); cDML (central Dronning Maud Land); DG (Denman Glacier); Dhar + Md (Dhawar and Madras); EG (Eastern Ghats); Gr (Grunehogna); KKB (Kerala Khondalite Belt); La (Lambert); LHB (Lützow-Holm Bay); M (Mawson); Madag (Madagascar); Nap (Napier Complex); nPCM (northern Prince Charles Mountains); Oy (Oygarden Islands); PZ (Prydz Bay); RI (Rauer Islands); Ru (Ruker); SL (Sri Lanka); sPCM (southern Prince Charles Mountains); Shack (Shackleton Range); SR (Sør Rondane) TA (Terre Adélie); VH (Vestfold Hills); wDML (western Dronning Maud Land); WI (Windmill Islands); YB (Yamato-Belgica). Dashed box shows the area covered by the map in Fig. 3.2-2.


INDIA
Dhar + Md
EG
Madag
Prydz Belt
Pinjarra Belt
Oy
M
Nap
VH
RI
Yilgarn
KKB
nPCM
DG
BH
East
African
Orogen
SL
LHB
Rayner
PZ
Rauer
AF
AUSTRALIA
WI
La
Wilkes
Ru
sPCM
Fig.2
YB
Lutzow
SR
AFRICA
Gawler
cDML
East Antarctic
Shield
TA
Kalahari
wDML
Gr
Maud
Miller
Shack
Ross Orogenic Belt
RBT
LFB
Namaqua
- Natal
Archaean cratons
1350-920 Ma
crustal Provinces
Archaean/
Proterozoic mixed
Pan-African
high-grade belts

(Harley, 1998, 2003; Boger and Miller, 2004; Kelsey et al., 2003) and the character of early Archaean crust formation processes (Black et al., 1986; Choi et al., 2006). In this contribution the oldest rocks in Antarctica will be described and documented. This follows a brief overview of the Precambrian geology of the East Antarctic Shield that is included to provide a framework for understanding the geological contexts of the earliest Archaean rocks and provinces.

3.2-2. OVERVIEW OF THE GEOLOGY OF THE EAST ANTARCTIC SHIELD

Recent metamorphic, structural and geochronological studies in the East Antarctic Shield (e.g., Tingey, 1991; Fitzsimons, 2000a, 2000b; Harley, 2003; Jacobs et al., 1998, 2003), demonstrate that it consists of a variety of Archaean and Proterozoic to Cambrian high-grade terrains that have distinct crustal histories. These terrains were not finally amalgamated until the Cambrian, most likely in two separate orogenic episodes within the time interval 600–500 Ma (Boger et al., 2001; Boger and Miller, 2004; Cawood, 2005). The broader implications of this, beyond the scope of this work, are that East Gondwana was not finally assembled until the Cambrian (Hensen and Zhou, 1995; Fitzsimons, 2000a, 2000b, 2003).

The major tectonothermal events that are recorded in the East Antarctic Shield occurred in the late Archaean to earliest Proterozoic (<2840 Ma but >2480 Ma), Palaeoproterozoic (2200–1700 Ma), Meso- to Neoproterozoic (1400–910 Ma), and latest Proterozoic to Cambrian (600–500 Ma: "Pan African"). Hence, the basement terrains may be divided into four partially overlapping categories according to their age, their degree of reworking, and the nature, style and ages of the reworking events (Harley, 2003). These categories or groups are:

(1) Regions, belts or areas that preserve extensive evidence for high-grade metamorphism, magmatism and deformation at times within the broad interval 600–500 Ma, the time-span corresponding to the 'Pan-African' and to the major amalgamation events in East Gondwana;

(2) Proterozoic terranes (Provinces) dominated by Meso- to Neoproterozoic tectonic events (1400–910 Ma), affected to a minor or limited extent by overprinting at 600–500 Ma;

(3) Archaean or mixed Archaean/Palaeoproterozoic areas with strongly polyphase histories in which the early record is partially to largely obscured as a result of overprinting by younger tectonic events; and,

(4) Archaean (>2500 Ma) to Palaeoproterozoic (>1690 Ma) terranes little affected by younger overprinting effects, in which largely pristine Archaean crustal histories are preserved.

These terranes, provinces and areas are distinguished in general terms in Fig. 3.2-1. In order to provide a context for the Archaean rocks of Antarctica, the terranes and regions grouped under (1) and (2) are briefly outlined below. The main regions placed in groups (3) and (4) are divided here into those that do not preserve any extensive zircon U-Pb evidence

Table 3.2-1. Summary of protolith and event ages for Archaean terranes in the East Antarctic Shield

Age, Ma	Napier Complex	Kemp Land Coast	Mather Terrane
>3900 Ma	inherited (?) zircons Mt Sones, Gage Ridge[1,2]		
3850	Mount Sones and Gage Ridge orthogneisses[1,2,3]	Hf model ages Oygarden Group and Stillwell Hills[8]	
3650–3620		anatexis and orthogneisses, Oygarden Group[7,8]	
3550–3500	inherited zircons, Rippon Point[8]	minimum ages of older homogeneous orthogneisses on Kemp Land coast[8]	inherited zircon in ca. 2550 Ma TTG sheet; oldest composite layered gneiss component (?)[10]
3490–3420	metamorphism and anatexis, Rippon Point (ca. 3422 Ma)[8]	metamorphism (ca. 3470 Ma). Anatexis and new orthogneisses (?)[7,8]	inherited zircon components in ca. 2550 Ma TTG sheet[10]
3390–3370			
3280–3250	Riiser–Larsen orthogneiss[4]		Short Point orthogneiss[9]
3190–3160	thermal (?) resetting, Rippon Point[8]		inherited zircons in ca. 2550 Ma TTG sheet[10]
3150–3050			
3050–2990	Proclamation orthogneiss[2,3]		
2890			
2840–2800	Dallwitz orthogneiss and dominant tonalitic orthogneisses[2,3]		homogeneous tonalitic orthogneisses; layered mafic complexes[9,10]
2790–2770		granulite facies metamorphism[7,8]	
2650–2640			
2620	Tonagh orthogneiss[5]		
2550–2520	ultrahigh temperature metamorphism[3,6]		trondhjemitic TTG orthogneiss sheets[11]
2520–2460	local granitoids, waning metamorphism[2,3,6]		

(*continued on next page*)

for rocks older than ca. 3000–3050 My old, which are described in general terms, and those that do preserve such evidence and hence are described in more depth, and distinguished in Table 3.2-1. In the final section of this chapter those localities or areas from which the

Table 3.2-1. (*Continued*)

Age, Ma	Ruker Terrane	Vestfold Block	Denman Glacier	Terre Adélie	Grunehogna
>3900 Ma					
3850					
3650–3620					
3550–3500					
3490–3420					
3390–3370					
3390–3370	layered tonalitic orthogneisses[12]				
3280–3250					
3190–3160	dominant granitic ortho-gneisses[12,13]				
3150–3050				orthogneisses[16]	
3050–2990			granitic and tonalitic orthogneisses[15]		granitic gneiss, Annandagstop-pane[17]
2890			granulite facies metamorphism[15]		
2840–2800		inherited zircons in Grace Lake Granodiorite[14]			
2790–2770	amphibolite facies metamorphism[13]				
2650–2640	post-D pegmatite[13]		orthogneisses[15]		
2620					
2550–2520			supracrustals[16]		
2520–2460		tonalitic (ca. 2520 Ma) and dioritic (ca. 2500–2475 Ma) orthogneiss[14]	metamorphism[16]		

References:
[1]Black et al. (1986); [2]Harley and Black (1997); [3]Kelly and Harley (2005); [4]Hokada et al. (2003); [5]Carson et al. (2002); [6]Harley (2003); [7]Kelly et al. (2004); [8]Halpin et al. (2005); [9]Kinny et al. (1993); [10]Harley et al. (1998); [11]Harley, unpubl.; [12]Mikhalsky et al. (2006); [13]Boger et al. (2006); [14]Black et al. (1991); [15]Black et al. (1992); [16]Peucat et al. (1999); [17]Jacobs et al. (1996).

oldest, pre-3400 Ma Antarctic rocks have been documented are described in detail and the implications of their isotopic records for the earliest history of crust formation highlighted.

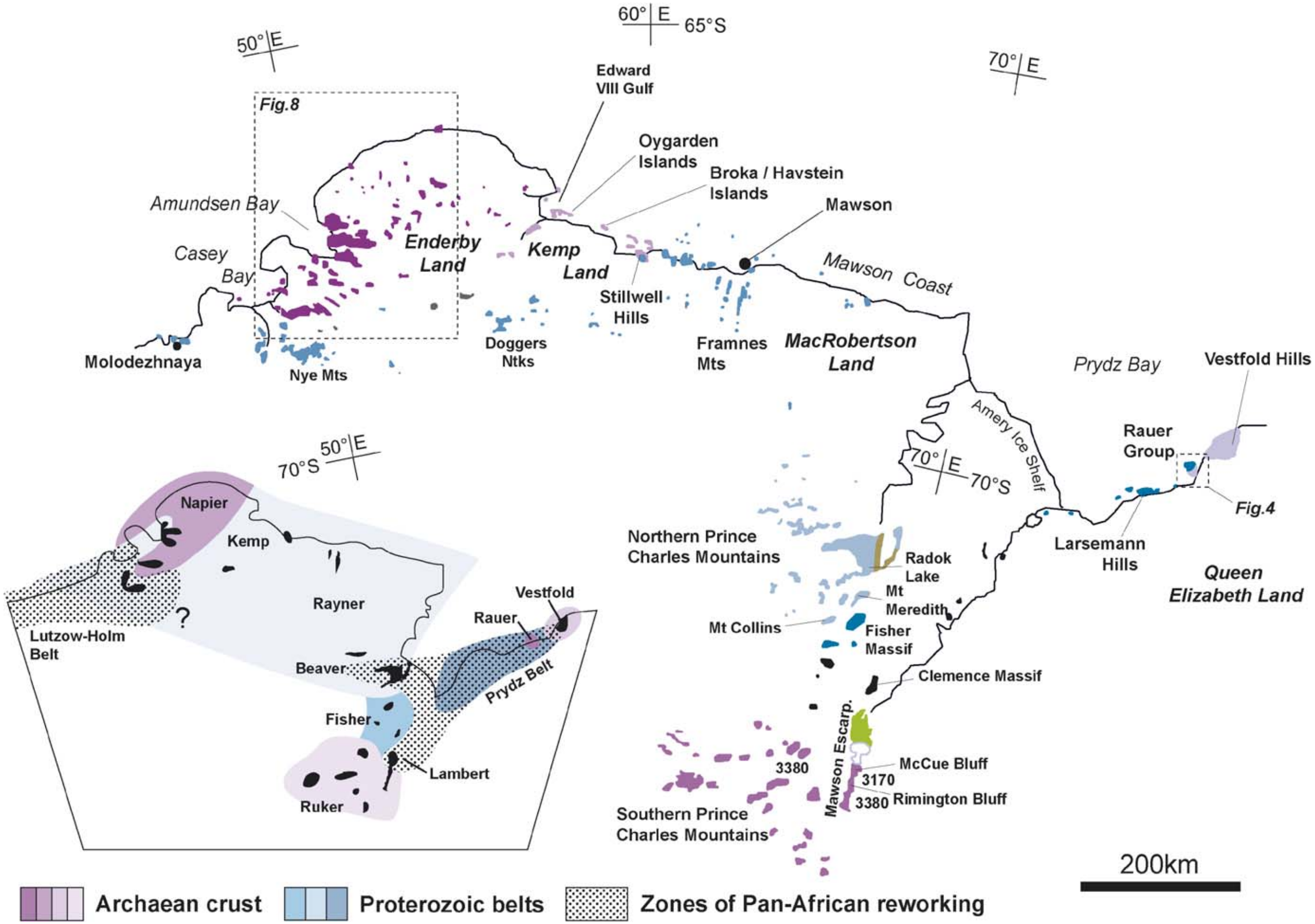
3.2-2.1. Cambrian and Proterozoic Tectonic Belts and Provinces

The recognition of intense 600–500 Ma Pan-African tectonism and high-grade metamorphism in the East Antarctic Shield has been one of the major breakthroughs in Antarctic geology in the past decade. This, coupled with the preservation of distinct earlier crustal histories in intervening terranes, has refuted the notion of a circum-East Antarctic 'Grenville' orogenic belt produced by the ca. 1100–900 Ma collision of a unified East Antarctic Craton with other parts of East Gondwana (Fitzsimons, 2000a).

High-grade Pan-African tectonism at 600–500 Ma is recognised from four potentially distinct regions (Fig. 3.2-1): Dronning Maud Land (East African – Antarctic Orogen: Jacobs et al., 1998), Lützow-Holm Bay (Lützow-Holm Belt), Prydz Bay and Southern Prince Charles Mountains (Prydz Belt) and Denman Glacier region (Pinjarra Belt) (Fitzsimons, 2000b, 2003). In all cases the high-grade metamorphism is associated with intense deformation, melting, and the emplacement of syn- to late-tectonic intrusives (e.g., Shiraishi et al., 1994; Fitzsimons et al., 1997; Jacobs et al., 1998; Boger et al., 2001). Temporal, structural and metamorphic constraints from the best-studied Pan-African areas are consistent with their formation and evolution as collisional belts, with collision terminated either by late-stage extensional collapse or the lateral escape of mid-crustal domains along high-strain zones at ca. 520–500 Ma (Fitzsimons et al., 1997; Jacobs et al., 1998, 2003; Boger et al., 2001; Harley, 2003). These high-grade belts juxtapose distinct Mesoproterozoic and Neoproterozoic crustal provinces (Maud, Rayner, Rauer and Wilkes: Fig. 3.2-1), and have reworked the margins of Archaean cratonic remnants in the Napier Complex and southern Prince Charles Mountains.

Proterozoic terranes dominated by Meso- to Neoproterozoic tectonism (1400–910 Ma) have been grouped into three broad 'provinces' that have distinct geological histories (Fitzsimons, 2000a) – the Wilkes (Wilkes Land coast), Rayner (Enderby, Kemp and MacRobertson Lands: Fig. 3.2-2), and Maud (Dronning Maud Land) provinces, to which Harley (2003) has added a fourth, the Rauer Province (Fig. 3.2-1).

Fig. 3.2-2. Outcrop distribution and simplified time/event information for high-grade terranes and regions in the Antarctica-Indian sector from Lützow-Holm Bay (west) to the Vestfold Hills (east). Latitudes and longitudes are referred to present Antarctic co-ordinates. Inset shows the generalised distribution of Archaean crust (Napier Complex, Kemp Land, Ruker Terrane, Rauer Terrane and Vestfold Hills), areas dominated by Proterozoic tectonothermal events (Rayner, Fisher and Lambert Provinces and Rauer–Prydz Bay basement area) and those dominated by Cambrian tectonism (Prydz and Lützow-Holm belts). Dashed boxes show the areas covered by the maps in Fig. 3.2-4 (Rauer Group) and Fig. 3.2-8 (Enderby Land).



60° E 65°S
50° E
70° E
Edward VIII Gulf
Fig.8
Oygarden Islands
Broka / Havstein Islands
Mawson
Amundsen Bay
Enderby Land
Kemp Land
Casey Bay
Mawson Coast
Stillwell Hills
Molodezhnaya
Nye Mts
Doggers Ntks
Framnes Mts
MacRobertson Land
Prydz Bay
Vestfold Hills
Amery Ice Shelf
Rauer Group
70° E 70°S
Fig.4
Larsemann Hills
Queen Elizabeth Land
Northern Prince Charles Mountains
Radok Lake
Mt Meredith
Mt Collins
Fisher Massif
Clemence Massif
Mawson Escarp.
McCue Bluff
3380
3170
3380 Rimington Bluff
Southern Prince Charles Mountains
200km
70°S 50° E
Napier
Kemp
Rayner
Vestfold
Rauer
?
Lutzow-Holm Belt
Beaver
Prydz Belt
Fisher
Lambert
Ruker
Archaean crust
Proterozoic belts
Zones of Pan-African reworking

The *Wilkes Province* correlates very well with the Albany-Fraser belt of west Australia (Fig. 3.2-1). It lacks evidence for any high-grade tectonothermal events younger than ca. 1130 Ma, is dominated by low-medium pressure granulite metamorphism and the emplacement of charnockites at ca. 1210–1150 Ma, and locally preserves evidence for earlier tectonic events at ca. 1340–1280 Ma (Post et al., 1997).

In the *Rayner Province* (Fig. 3.2-1) older crust (e.g., Archaean Napier Complex, 3800–2500 Ma; 1450–1500 Ma anorthosites: Black et al., 1987; Sheraton et al., 1987) and Proterozoic sediments and younger intrusives (e.g., ca. 960 Ma charnockites: Young and Black, 1991) have undergone granulite facies deformation either at ca. 1000–980 Ma or ca. 930–920 Ma, or in both of these events. In the west and northern Rayner Province the major event is ca. 930–920 Ma in age (Kelly et al., 2000, 2002), whereas the ca. 1000–980 Ma peak metamorphic event dominates in the sub-area, known as the Beaver Terrane (Mikhalsky et al., 2001), exposed in MacRobertson Land and the Northern Prince Charles Mountains (Fig. 3.2-1; Fig. 3.2-2). This age distribution may reflect two-stage collision between the Ruker Terrane and other areas of inland Antarctica (e.g., the Mawson Continent of Fanning et al., 1999), a continent comprising much of eastern India, and a microcontinent that included the Archaean Napier Complex (Kelly et al., 2002).

The *Maud Province* (Fig. 3.2-1) is dominated by arc-related felsic plutonism at ca. 1180–1130 Ma and ca. 1080 Ma, regional high-grade metamorphism and deformation at ca. 1050–1030 Ma (Grantham et al., 1995; Jacobs et al., 1998), and further thrust-related deformation that may be as young as 980 Ma. These events are considered to reflect arc accretion followed by collision between the Kalahari Craton and an Antarctic Craton (Grantham et al., 1995; Jacobs et al., 1998).

The *Rauer Province* (Figs. 3.2-1 and 3.2-2) lies on the eastern margin of the Prydz Belt, and therefore separates the Prydz Belt from the Pinjarra Belt that occurs on the western seaboard of Australia and extends into the Denman Glacier area of Antarctica (Fig. 3.2-1). The Proterozoic tectonic history of the Rauer Province involved partial melting, high-grade tectonism and emplacement of felsic intrusives over the time interval 1080–990 Ma (Kinny et al., 1993; Harley, 2003). This event chronology bears more similarity with that of the Maud Province than with those preserved in the nearest-neighbour provinces, the Wilkes (to the E) which has an older tectonothermal record, and the Rayner (to the W) which has a major event at 930–920 Ma.

3.2-2.2. Archaean and Archaean – Palaeoproterozoic Terranes with Little pre-3000 Ma Record

The *Mawson Block* of Terre Adélie and other areas on the eastern fringe of the East Antarctic Shield (TA: Fig. 3.2-1; Peucat et al., 1999; Fanning et al., 1999) preserves an Archaean to Palaeoproterozoic geological record that is correlated with the Gawler Craton and adjacent areas of Australia (Fig. 3.2-1) and the Nimrod Group of the Miller Range in central Trans Antarctic Mountains (Miller: Fig. 3.2-1; Goodge and Fanning, 1995; Goodge et al., 2001). The basement geology in the Terre Adélie region includes vestiges of ca. 3050–3150 gneisses (Table 3.2-1) and ca. 2560–2450 Ma supracrustals that were metamorphosed in the earliest Proterozoic (ca. 2440 Ma; Peucat et al., 1999). These formed the basement to rift-related sediments and volcanics deposited between ca. 1775 and 1700 Ma (Fanning et al., 1999) and metamorphosed at ca. 1710–1690 Ma in the Kimban tectonothermal event (Goodge and Fanning, 1995). This event progressed under low-pressure granulite facies conditions in Terre Adélie, whereas in the Nimrod Group an event of similar age (ca. 1700–1690 Ma) progressed under eclogite facies conditions (Goodge and Fanning, 1995; Goodge et al., 2001). The correlations in events, and provenance ages in the metasediments, has stimulated the proposal that the Mawson Block may form much of the continental land-mass of East Antarctica underlying the ice, at least as far to the west as the subglacial Lake Vostok (Fanning et al., 1999; Fitzsimons, 2003; Boger and Miller, 2004).

In the *Denman Glacier/Obruchev Hills Area* (DG: Fig. 3.2-1) granitic and tonalitic orthogneiss precursors with ages between ca. 3000 and 2640 Ma have been recognised (Table 3.2-1), with the older orthogneiss affected by a granulite facies metamorphic event at ca. 2890 Ma (Black et al., 1992).

The *Vestfold Block* of eastern Prydz Bay (VH: Figs. 3.2-1 and 3.2-2) experienced its main magmatic accretion events and crust formation in the time interval ca. 2520–2480 Ma (Black et al., 1991; Snape et al., 1997). 2520 Ma tonalitic orthogneisses and pyroxene granulites, and supracrustal rocks including metavolcanics, Fe-rich semipelitic sediments and Mg-Al rich claystones (Oliver et al., 1982, Harley, 1993) were metamorphosed under granulite conditions by 2500 Ma (Table 3.2-1). The broadly syn-tectonic Crooked Lake Gneisses were then emplaced and overprinted in an upper amphibolite to granulite deformation episode at ca. 2500–2480 Ma (Black et al., 1991; Snape et al., 1997). A pre-2530 Ma geological record is preserved only in a group of migmatitic granodioritic orthogneisses, the Grace Lake Granodiorite (Table 3.2-1), which contains inherited zircons up to ca. 2800 Ma in age (Black et al., 1991) and records a T_{DM}(Nd) model age of ca. 3050 Ma based on the data of Kinny et al. (1993).

The *Lambert Province* or *Terrane* (Boger et al., 2001, 2006; Mikhalsky et al., 2001, 2006a, 2006b; Fig. 3.2-2) is dominated by felsic to intermediate orthogneisses derived from granitoids emplaced at ca. 2420 Ma. These were interleaved with mafic amphibolite, calc-silicate gneiss and marble and metamorphosed to upper amphibolite facies at ca. 2065 Ma and possibly at other times in the Palaeoproterozoic (e.g., ca. 1800, ca. 1600 Ma: Boger et al., 2001; Mikhalsky et al., 2006a). Some of the orthogneisses are derived from granitoids that have elevated ε_{Nd} values (to +6.4) and T_{DM} approximating 2400 Ma and hence are considered to be juvenile crustal additions at ca. 2450 Ma. Other orthogneiss protoliths have highly negative ε_{Nd} values at 2400 Ma (to −6.4), and thus have old T_{DM} model ages (ca. 3000–3400 Ma) that imply the involvement of older Archaean crust in their genesis (Mikhalsky et al., 2006b). The Lambert Terrane is strongly overprinted by Cambrian tectonism, manifested in the intrusion and subsequent deformation of abundant granite sheets, veins and vein networks with zircon U-Pb ages between ca. 530 Ma and 495 Ma (Boger et al., 2001; Mikhalsky et al., 2006a).

The Grunehogna Craton of western Dronning Maud Land (Gr: Fig. 3.2-1) is exposed at only one nunatak (Table 3.2-1), where ca. 3000 Ma granitic basement is overlain by ca. 1100–1000 Ma shelfal sediments and volcanics of the Ritscherflya Supergroup (Grantham et al., 1995). The greater lateral extent of this basement, interpreted to be a fragment of the Kalahari Craton of southern Africa left attached to East Antarctica, is supported by geophysical surveying (Jacobs et al., 1996).

3.2-2.3. Archaean and Archaean/Proterozoic Terranes with Pre-3000 Ma Crustal Records

3.2-2.3.1. The Ruker Terrane, Southern Prince Charles Mountains

The Ruker Terrane of the Southern Prince Charles Mountains (sPCM: Figs. 3.2-1 and 3.2-2) is widely regarded as representative of a larger region of the 'inboard' EAS that collided or interacted with outboard terranes (e.g., the Archaean Napier Complex) and other shield areas (e.g., the Dharwar and Bastar cratons of India) at 990–910 Ma and 550–500 Ma (e.g., Fitzsimons, 2000b; Boger and Miller, 2004). It comprises a polyphase Archaean (ca. 3390–3155 Ma) granite gneiss basement, the Mawson Suite (Fig. 3.2-3(b)), tectonically interleaved with deformed metasedimentary and metavolcanic rocks (Fig. 3.2-3(c); Tingey, 1982, 1991; Mikhalsky et al., 2001; Boger et al., 2006; Mikhalsky et al., 2006a).

The deformed metasedimentary and metavolcanic rocks of the Ruker Terrane have been divided into three tectonostratigraphic units, the Menzies, Ruker and Sodruzhestvo Series (Mikhalsky et al., 2001). The Menzies Series shares the same Archaean deformational history as the Mawson Suite and its correlatives (Boger et al., 2006), and hence is at least late Archaean in depositional age. Quartzites, pelitic and calcareous metasediments and amphibolites of the Menzies Series exhibit medium-pressure Barrovian-style metamorphism (staurolite + kyanite ± garnet) and may preserve evidence for two distinct metamorphic events (Boger et al., 2006). The Ruker Series includes mafic to felsic metavolcanic rocks and associated metadolerite sills, metapelitic schist, slate, phyllite, and banded ironstones. The Sodruzhestvo Series consists of calcareous schist, pelite, phyllite and slate, and minor marble, quartzite, and conglomerate, and is thought to be the younger of the two metasedimentary series (Mikhalsky et al., 2001). Both the Ruker and Sodruzhestvo Series post-date the Archaean tectonics and intrusive events that have affected the Mawson Suite and Menzies Series, can be considered to represent Proterozoic cover sequences to the latter lithological groups (Boger et al., 2006), and are now metamorphosed to greenschist facies conditions.

An Archaean age for the Ruker Terrane basement orthogneiss (e.g., Mawson Suite granites) was initially established on the basis of Rb-Sr whole rock isochrons, which gave ages in the range ca. 2700–2760 Ma (Tingey, 1982), and an age of ca. 2589 Ma for a cross-cutting muscovite-bearing pegmatite. These minimum ages have been complemented by conventional multigrain zircon U-Pb data that indicate a crystallisation age of 3005±57 Ma for a granite at Mount Ruker (Mikhalsky et al., 2001), and by Sm-Nd whole rock isochron ages of 3124 ± 130 Ma and 3176 ± 140 Ma on granite gneisses from the Mawson Escarpment (Figs. 3.2-3(a,b)). Whilst indicative of the presence of older crust, these age data have large inherent errors and so do not allow detailed resolution of the age-event history of the Ruker Terrane. However, the ages of Archaean orthogneisses in the Ruker Terrane have recently been clarified by zircon U-Pb SHRIMP dating (Table 3.2-1).

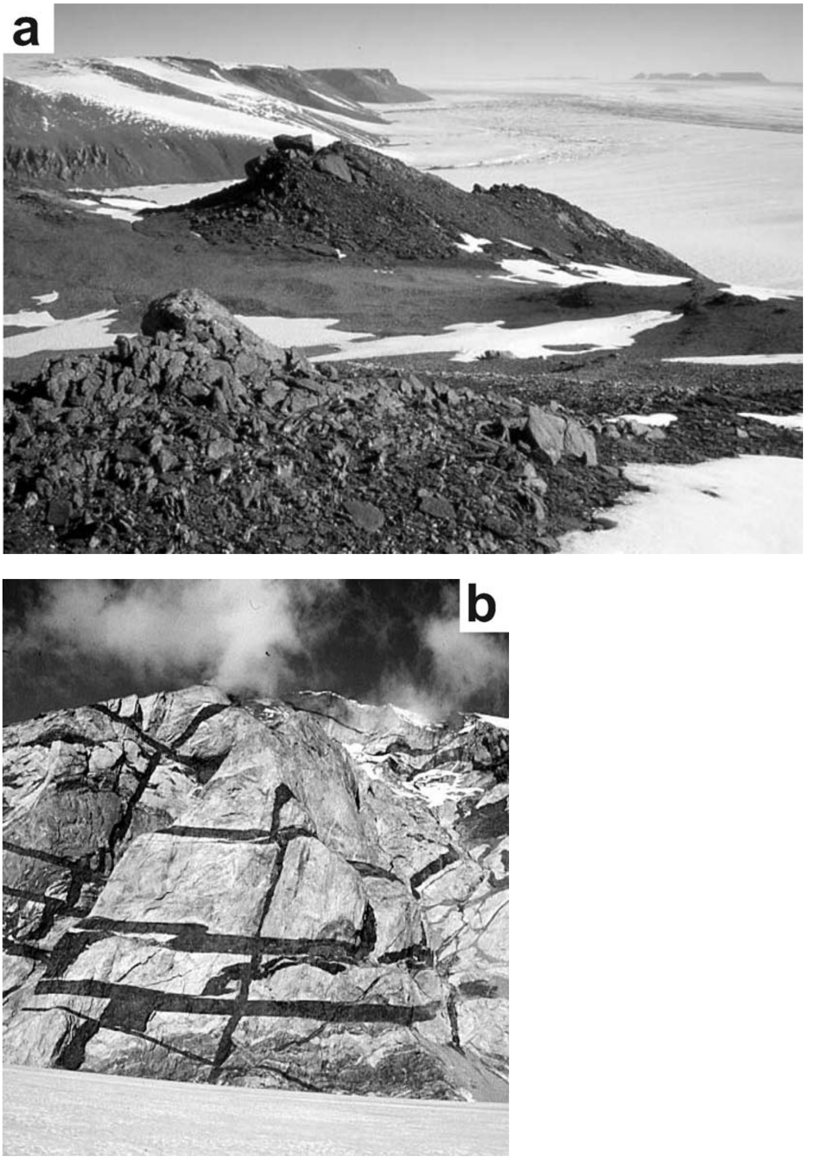


Fig. 3.2-3. Geological features and relations in the Ruker Terrane, Southern Prince Charles Mountains (all photos courtesy of Dr. S. Boger, University of Melbourne). (a) View of the Lambert Glacier and surroundings, looking south along the Mawson Escarpment towards Mount Menzies, some 100 km distant. (b) View of the 1500 metre high cliff face of McCue Bluff. Typical Archaean (ca. 3160 Ma) granitic orthogneiss of the Mawson Suite is cut by several orientations of Proterozoic mafic dykes.

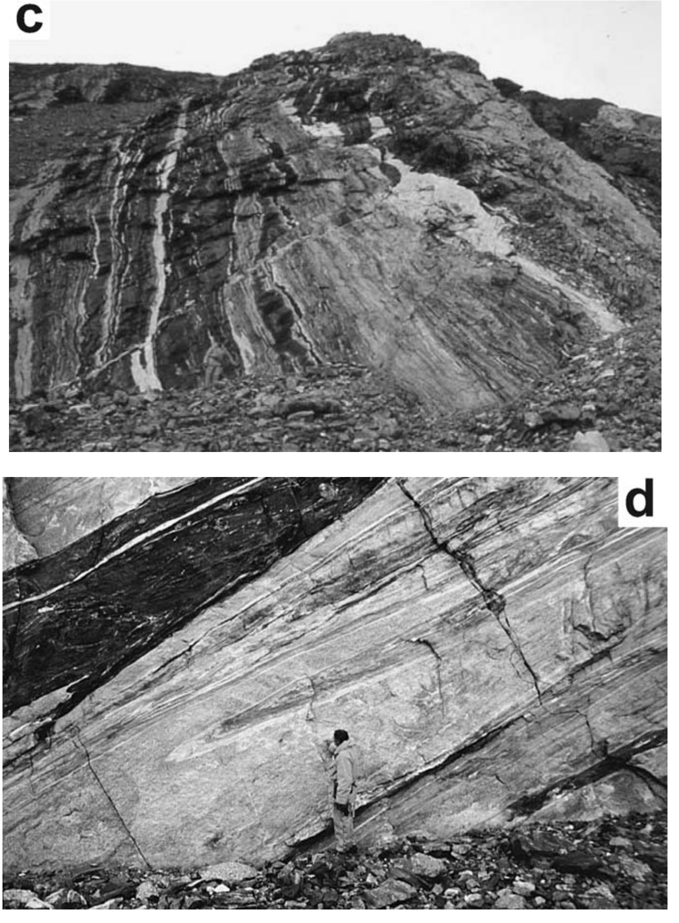


Fig. 3.2-3. (*Continued.*) (c) Menzies Series layered paragneisses, including mafic and felsic/intermediate schists interpreted as former volcanics, at Rimmington Bluff (note geologist for scale). (d) Isoclinal F_2 fold in Menzies Series layered schists and gneisses at Rimmington Bluff (note geologist for scale). (e) Boudinaged mafic unit in layered paragneiss of the Menzies Series at McCue Bluff. Syn-D3 leucosome from these boudin necks constrain an age of ca. 2772 Ma for this event (Boger et al., 2006).

Homogeneous/massive gneissic granites (Mawson Suite) that crop out in the Southern Mawson Escarpment have yielded oscillatory zoned magmatic zircons that record crystallisation over the time interval ca. 3185–3155 Ma, with specific granitoids yielding population ages of 3182 ± 9 Ma (Mikhalsky et al., 2006a), 3177 ± 6 Ma, 3174 ± 9 Ma and 3160 ± 6 Ma (Boger et al., 2006). One of these granitoids preserved evidence for older crustal sources, in the form of xenocrystic zircon cores with a weighted population

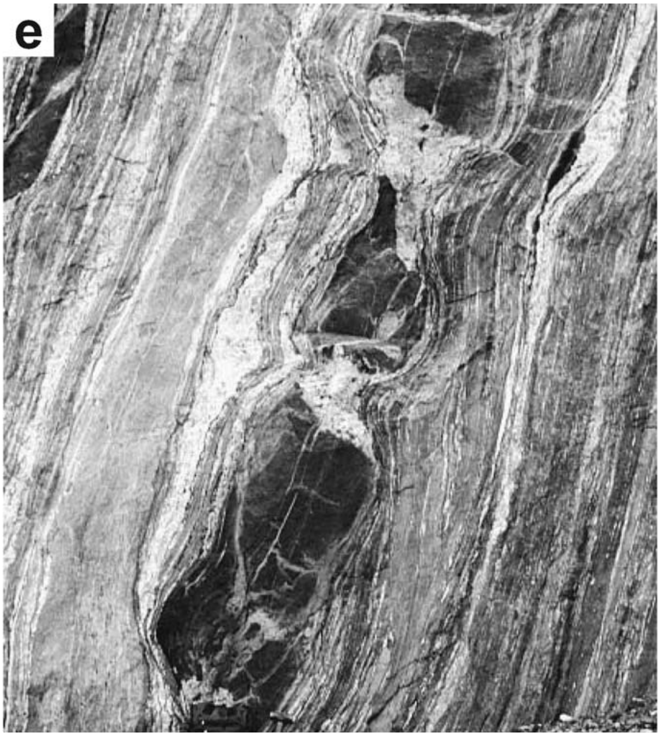


Fig. 3.2-3. (*Continued.*)

concordia age of 3370 ± 11 Ma. This age is consistent with the incorporation of crustal sources similar to a Y-depleted banded trondhjemitic gneiss and a porphyritic tonalite cobble preserved in Ruker Series metasediments and dated by Mikhalsky et al. (2006a), which yielded zircon U-Pb concordia ages of 3392 ± 6 Ma and 3377 ± 9 Ma respectively. These results demonstrate that the Ruker Terrane contains crustal components at least as old as ca. 3390 Ma, and on the basis of Sm-Nd model ages (T_{DM}) of ca. 3200–3900 Ma (Mikhalsky et al., 2006b) it is possible that even more ancient early Archaean rocks may be present.

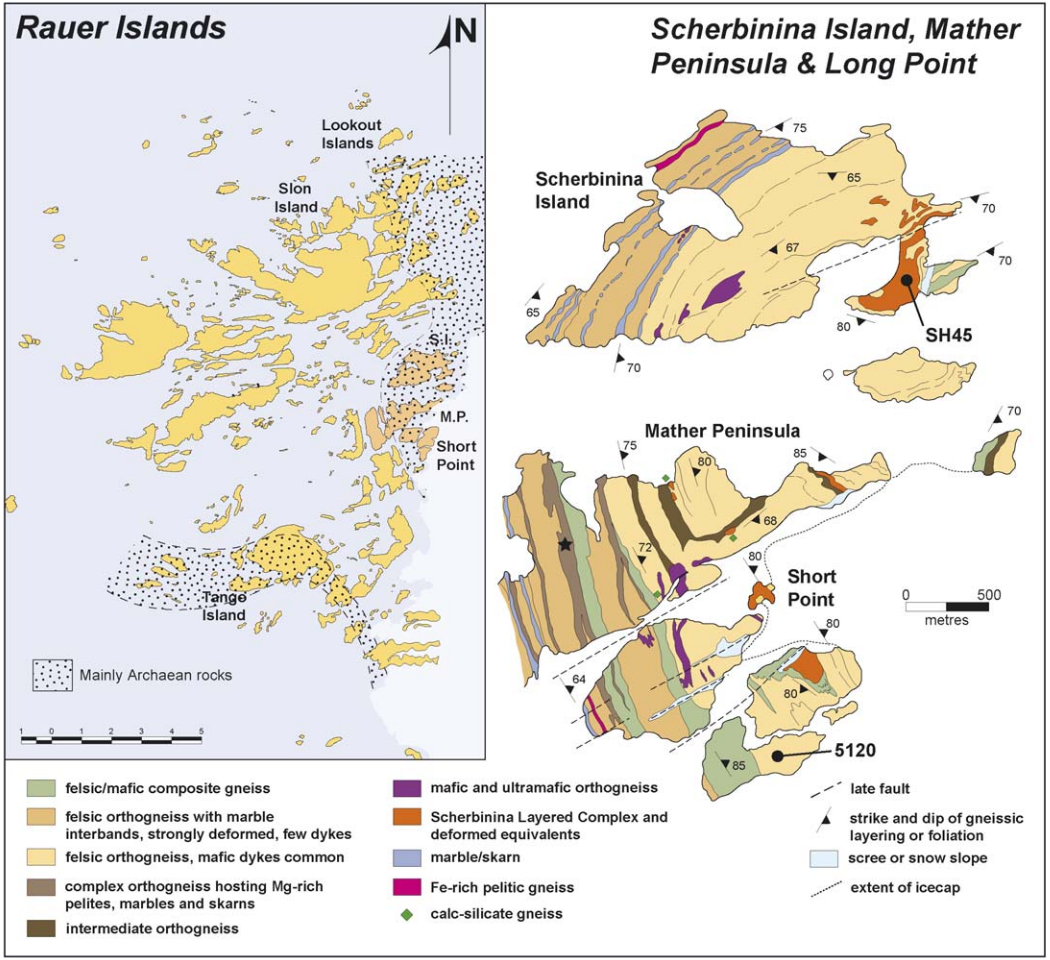
The ages of principal Archaean regional deformation events (Fig. 3.2-3(d)) that have affected the Ruker Terrane have until recently only been constrained to pre-date cross-cutting pegmatite that has a zircon U-Pb age of ca. 2650 Ma (Boger et al., 2001, 2006). Mikhalsky et al. (2006a) inferred a ca. 3145 Ma thermal/metamorphic event based on the age of a concordant high-U zircon rim formed in one of the ca. 3380 Ma orthogneisses, but was not able to relate this to any deformation or tectonic event. However, recent zircon U-Pb dating of structurally-constrained samples (Boger et al., 2006; Fig. 3.2-3(e)) provides tight brackets on the Archaean deformation and metamorphism. Leucosomes formed between the first and second (D_1–D_2) folding events, or segregated into inter-boudin necks in more localised D_3 high strain zones, were formed over the interval ca. 2790–2770 Ma. These ages are consistent with a single, relatively short-lived, Archaean tectonothermal event and imply Rb-Sr re-equilibration on the whole rock scale during the associated upper amphibolite facies regional metamorphism (Rb-Sr ages ca. 2760–2700 Ma: Tingey, 1982).

3.2-2.3.2. The Archaean Mather Terrane, Rauer Province, Prydz Bay

The Rauer Province, which outcrops in the Rauer Group of islands on the eastern side of Prydz Bay (RI; Fig. 3.2-1; Rauer, Fig. 3.2-2), includes both Archaean and Mesoproterozoic crustal components metamorphosed and deformed at ca. 530–500 Ma during intense shearing and mylonitisation, and amphibolite to granulite facies high-strain zones associated with Pan-African tectonism in the adjacent Prydz Belt (Kinny et al., 1993; Sims et al., 1994; Harley, 2003). Here we distinguish the *Mather Terrane* as the sub-area of the Rauer Group of islands that is dominated by Archaean protoliths (Fig. 3.2-4).

The *Mather Terrane* is dominated by >3270 Ma and ca. 2820–2800 Ma tonalitic orthogneisses (Fig. 3.2-5(a); Table 3.2-1; Kinny et al., 1993), accompanied by minor supracrustal units rich in marbles and magnesian-aluminous pelites and quartzites (Mather Paragneiss: Harley et al., 1998). The oldest orthogneiss components, which contain zircons up to ca. 3500 Ma in age, will be described in depth in a later section. The ca. 2820–2800 Ma orthogneisses have a T_{DM}(Nd) model age of ca. 3250–3350 Ma and are interpreted to be derived from the remelting of older protoliths (Table 3.2-1; Kinny et al., 1993). These orthogneisses intrude or enclose ca. 2840 Ma layered igneous complexes that preserve original igneous features despite the effects of later tectonothermal events (Harley et al., 1998). The extensively dyked and reworked Archaean orthogneisses and related supracrustal packages (Harley et al., 1992; Sims et al., 1994) are interleaved with Mesoproterozoic supracrustals and ca. 1030–1000 Ma felsic to mafic intrusives (Kinny et al., 1993) that experienced high-grade metamorphism and partial melting at ca. 1000 Ma and probably 510 Ma (Harley et al., 1998). Despite extensive texturally constrained geochronology, it is not clear as to whether the Archaean gneisses were reworked during the Mesoproterozoic at ca. 1000 Ma, or only affected by Archaean tectonothermal events and subsequently overprinted during interleaving with the Mesoproterozoic gneisses in the Prydz Belt event (Hensen and Zhou, 1995). This ambiguity exists because the zircon U-Pb data obtained from Archaean gneisses does not record any Mesoproterozoic disturbance (Kinny et al., 1993; Harley et al., 1998). Moreover, the polyphase reactivation of earlier gneissic fabrics in the youngest, ca. 500 Ma ductile deformation events has resulted in the Archaean and Mesoproterozoic lithological packages being rotated into parallelism (Sims et al., 1994; Harley et al., 1998).

Fig. 3.2-4. Geological map of the Mather Peninsula–Scherbinina Island area in the Rauer Group, showing the locations of the Short Point orthogneiss sample 5120 (Sheraton et al., 1984) and SH45 (Harley et al., 1998). Inset is a map of the Rauer Group as a whole, showing the general distribution of Archaean rocks ascribed to the Mather Terrane. See Fig. 3.2-2 for location of the Rauer Group and this map area within East Antarctica.


Rauer Islands
N
Lookout Islands
Slon Island
S.I.
M.P.
Short Point
Tange Island
Mainly Archaean rocks
1 0 1 2 3 4 5
Scherbinina Island, Mather Peninsula & Long Point
Scherbinina Island
SH45
Mather Peninsula
Short Point
0 500
metres
5120
75
65
70
67
70
65
80
70
70
75
80
85
68
72
80
80
64
80
85
felsic/mafic composite gneiss
felsic orthogneiss with marble interbands, strongly deformed, few dykes
felsic orthogneiss, mafic dykes common
complex orthogneiss hosting Mg-rich pelites, marbles and skarns
intermediate orthogneiss
mafic and ultramafic orthogneiss
Scherbinina Layered Complex and deformed equivalents
marble/skarn
Fe-rich pelitic gneiss
calc-silicate gneiss
late fault
strike and dip of gneissic layering or foliation
scree or snow slope
extent of icecap

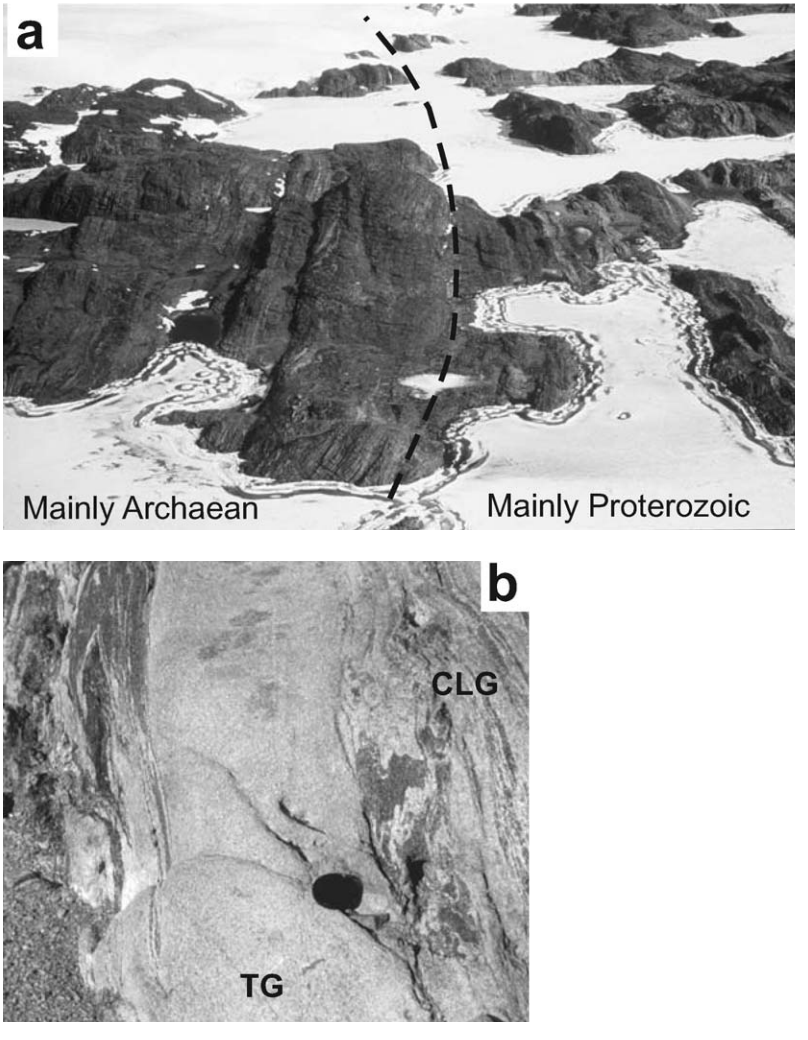


Fig. 3.2-5. Archaean geological features and relations in the Mather Terrane, Rauer Group of islands, Prydz Bay. (a) View, looking south, of the Mather Peninsula (foreground) and Short Point area. The eastern part of the area is dominated by Archaean (ca. 2820 Ma and >3200 Ma) protoliths. In the western part Archaean and Proterozoic protoliths are interleaved on tens to hundreds of metres scales. (b) 40 centimetre wide sheet of locally discordant homogeneous tonalitic orthogneiss (TG) cutting composite layered gneiss (CLG) characterised by folded mafic-felsic banding on centimetre to decimetre scales. All rocks are metamorphosed at granulite facies conditions. Scherbinina Island, Rauer Group. (c) Folded leucotonalitic–trondhjemitic orthogneiss (TrG) cutting composite layered gneiss (CLG) that preserves earlier folds in mafic-felsic banding. Scherbinina Island, Rauer Group.

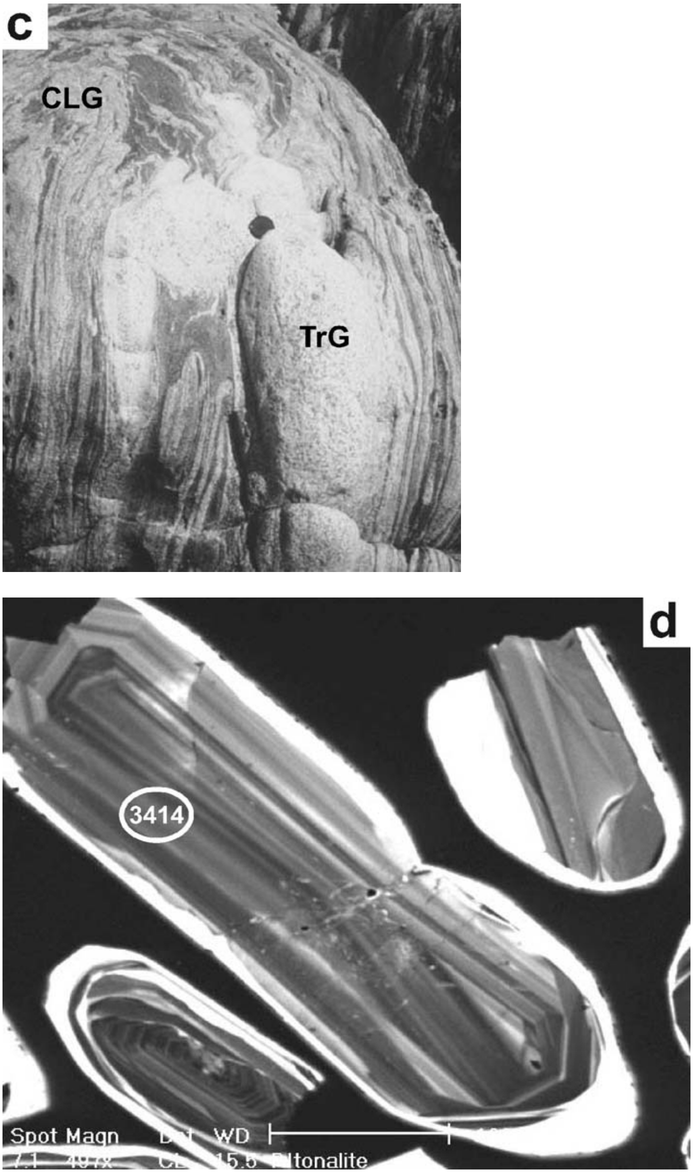


Fig. 3.2-5. (*Continued*.) (d) Cathodoluminescence (CL) image of an elongate, oscillatory zoned zircon of the type that dominates the magmatic zircon population in sample SH45 from the Rauer Group. The ca. 3414 Ma $^{207}Pb/^{206}Pb$ age of this concordant zircon indicates that it is inherited – much older than the ca. 2550 Ma age of the trondhjemite sheet itself. Bright CL outer rims and rounded terminations reflect zircon recrystallisation and dissolution-reprecipitation at ca. 510 Ma (Harley et al., 1998).

3.2-2.3.3. The Kemp Land Coast Region

In the Oygarden Group of Kemp Land (Oy: Figs. 3.2-1 and 3.2-2) initial deformation in the Rayner Structural Episode at ca. 930 Ma involved the deep-crustal ductile reworking and thrusting of Archaean gneisses that range in age up to ca. 3650 Ma (Kelly et al., 2002, 2004). Given the proximity of this region, and the broad first-order similarities in the age spectra of the Archaean crustal components, the Rayner Province in this area has been regarded as the reworked equivalent of the adjacent Napier Complex, described below.

Kelly et al. (2004) reported SHRIMP U-Pb zircon data for a number of structurally constrained orthogneisses from the Oygarden Group, and interpreted the quite complex isotopic results in the light of these structural constraints and detailed cathodoluminescence (CL) and back scattered electron (BSE) imaging of the separated zircons. In addition to the isotopic disturbance of zircons in the Rayner Structural Episode at ca. 930 Ma, Kelly et al. (2004) were able to identify thermal events at ca. 1600 Ma and 2400 Ma, and define a late Archaean history involving the metamorphism of early Archaean layered composite orthogneisses and intrusive homogeneous felsic orthogneisses at $\geqslant$2780 Ma (Table 3.2-1). The pre-2780 Ma events recorded in these orthogneisses and similar felsic orthogneisses from areas elsewhere in Kemp Land (Halpin et al., 2005) are considered in detail in the subsequent section on the oldest rocks.

3.2-2.3.4. The Archaean Napier Complex, Enderby Land

The Archaean Napier Complex (Nap: Figs. 3.2-1 and 3.2-2) of Enderby Land contains the oldest rocks yet recorded from Antarctica (Table 3.2-1). Whilst some granitic and tonalitic gneiss precursors, to be described in detail below, range in age up to ca. 3850 Ma (Williams et al., 1984; Black et al., 1986; Harley and Black, 1997; Kelly and Harley, 2005), the dominant granulite facies orthogneisses were derived from the metamorphism of granites, tonalites and granodiorites with mid- to late-Archaean ages of ca. 3270 Ma, 3070–2970 Ma and 2840–2800 Ma (Sheraton et al., 1987; Harley and Black, 1997; Hokada et al., 2003; Kelly and Harley, 2005). These and the rarer ancient orthogneiss precursors together with distinctive supracrustal packages, were strongly deformed and interleaved under ultrahigh temperature metamorphic conditions in the late Archaean.

The Napier Complex records the highest grades of ultrahigh temperature (UHT) crustal metamorphism worldwide (Harley, 2003). The lithological diversity of the layered paragneisses, which include mafic pyroxene granulites, metaironstones, pelites and both Fe- and Mg-rich metaquartzites, has enabled the deduction of key indicator mineral assemblages that imply temperatures of metamorphism of between 1000 and 1120 °C at depths of 20–35 kilometres. These UHT indicators, summarised by Harley (1998, 2003), include sapphirine and quartz, with orthopyroxene or garnet, aluminous orthopyroxene (9–11 wt% Al_2O_3) coexisting with sillimanite and quartz, the K-Mg-Al mineral osumilite coexisting with sapphirine or garnet, ternary mesoperthitic feldspars in pelitic gneisses, and inverted magnesian pigeonite in metamorphosed ironstones.

Following many years of controversy (Grew and Manton, 1979; Black et al., 1983b; Harley and Black, 1997; Grew, 1998), the age of the UHT metamorphism in the Napier Complex is now constrained to be younger than 2626 ± 28 Ma, which is the protolith age of a granodioritic orthogneiss from Tonagh Island (Carson et al., 2002). U-Pb ages of zircons present in syn-metamorphic pegmatitic 'sweats' and of the oldest zircon rims formed on older grains within paragneisses further constrain the age of metamorphism to between ca. 2590 Ma and 2510 Ma (Hokada and Harley, 2004; Kelly and Harley, 2005). The ubiquitous zircon age clusters at ca. 2480–2450 Ma determined in many gneisses from the Napier Complex reflect fabric formation and fluid access subsequent to the UHT event, and are interpreted to relate to final melt crystallisation and fluid expulsion that occurred as the Complex cooled, at depth in the crust, through 700–750 °C (Kelly and Harley, 2005).

Unlike many other high grade terrains, in which variations in the intensity of deformation and the presence of low-strain domains allows original rock features and relationships to be observed, any original lithological features and contact relationships in the Napier Complex have been obliterated by the intense ductile deformation and recrystallisation that was associated with the UHT metamorphism. This deformation produced pervasive flat-lying gneissic fabrics and early isoclinal folds that have been refolded and reoriented about several later phases of folds and high-strain zones. Intense down-dip linear fabrics defined by the orientations of the highest-grade minerals (e.g., sillimanite, sapphirine, orthopyroxene, quartz) are common, as is boudinage and megaboudinage of competent units during layer-parallel extension (e.g., Black and James, 1983; Black et al., 1983a, 1983b; Sheraton et al., 1987; Harley, 2003). Later deformation events are discrete, localised to pseudotachylites and metre- to tens of metres width high-strain and mylonitic zones, probably of Cambrian and Proterozoic ages (Sheraton et al., 1987).

3.2-3. THE OLDEST ROCKS: >3400 MA

Having outlined the general geological context of the Archaean of Antarctica and described the Archaean and reworked Archaean terranes themselves in more detail, we now go on to consider specific examples of the oldest rocks so far documented from the continent (Table 3.2-1). For convenience we will consider the evidence for crust, or crustal precursors, in the age range ca. 3400 Ma to 4000 Ma and further restrict the discussion to those cases where the evidence is principally provided or strongly supported by zircon U-Pb microanalysis. These restrictions limit us to consider three areas or terranes: the Mather Terrane, the Kemp Land sub-area of the Rayner Province, and the Napier Complex. Whilst crustal material as old as ca. 3390 Ma is reported from the Ruker Terrane (Table 3.2-1), as there is no zircon evidence for older crust in this terrane we will not consider it further here.

3.2-3.1. Early Archaean Orthogneisses of the Mather Terrane

The oldest gneisses and gneiss precursors in the Mather Terrane are located in the Mather Peninsula area (Fig. 3.2-4), where Kinny et al. (1993) reported an age of 3269 ± 9 Ma for oscillatory zoned magmatic zircons extracted from a charnockitic-felsic gneiss at Short

Point (Table 3.2-1). Sm-Nd data from this reasonably homogeneous orthogneiss (sample 5120: Sheraton et al., 1984) indicate that its magmatic protolith was generated from evolved crustal material ($\varepsilon_{Nd}(3270) = -2.5$) that may have been sourced from depleted mantle in the earliest Archaean ($T_{DM}(Nd)$ = ca. 3750 Ma).

The high strain associated with the latest high-grade ductile deformation episodes has led to early structures being transposed into near-parallelism with younger ones across much of the Mather Peninsula area. Archaean and Mesoproterozoic lithologies now form horizons or packages with unit-parallel boundaries, coaxial folds and refolding, and coincident SSE-plunging mineral elongation lineations on the macro- to meso-scale (e.g., Fig. 3.2-5(a)). Within low-strain zones Harley et al. (1998) have demonstrated that the oldest gneiss components of the Rauer Terrane are represented by composite layered orthogneisses on the basis of cross-cutting relationships and relative structural chronologies (Figs. 3.2-5(b,c)). These dominantly tonalitic to granodioritic gneisses are migmatitic, containing rafts and schlieren of mafic granulite. They are cut by metre to tens of metres wide sheets of homogeneous felsic gneiss similar to those dated at ca. 2820–2800 Ma by Kinny et al. (1993) (Fig. 3.2-5(b)), or by lenses and veins of leucocratic tonalitic to trondhjemitic gneiss (Fig. 3.2-5(c)) that may also cut homogeneous tonalitic gneisses (Harley et al., 1998). The composite layered orthogneisses, homogeneous tonalitic gneisses and tonalitic-trondhjemitic gneisses are all cut by several generations of metamorphosed and deformed mafic dykes (e.g., Sims et al., 1994), including most of the suites of meta-tholeiite dykes that cut the Scherbinina Layered Complex (Harley et al., 1998).

This composite layered orthogneiss unit has not been dated directly. However, Harley et al. (1998) reported zircon U-Pb age data from a tonalitic-trondhjemitic orthogneiss sheet intruded into the ca. 2840 Ma Scherbinina Layered Complex. This sheet (SH45; Fig. 3.2-4) yielded a complex pattern of zircon U-Pb data that could be divided into a several distinct Archaean populations, each defining discordia with lower intercepts at ca. 510 Ma (the Prydz Belt event; Fig. 3.2-6(a)). With the exception of one concordant analysis at ca. 2550 Ma, all of the Archaean zircon arrays defined concordia ages older than the Scherbinina Layered Complex itself. The oldest linear array yielded a concordia intercept age of 3470 ± 30 Ma (Harley et al., 1998).

The unambiguous geological observation that this tonalitic-trondhjemitic sheet has intruded into the ca. 2840 Ma Scherbinina Layered Complex (Table 3.2-1) leads to the conclusion that the majority of the oscillatory zoned Archaean zircon in the tonalite sample is inherited from an older source rock or range of source rocks. Detailed CL and BSE imaging (e.g., Fig. 3.2-5(d)) has now been used to establish a zircon growth stratigraphy, and so texturally constrain those zircon growth zones that might be related to its crystallisation in the tonalite (Harley, unpubl.). SHRIMP analysis of these specifically targetted outer oscillatory zones yield a linear array of concordant to variably discordant zircon U-Pb analyses that define an upper intercept age of ca. 2550 Ma (Table 3.2-1), interpreted to be the crystallisation age of the tonalitic-trondhjemitic sheet itself. The 3300–3470 Ma zircon U-Pb ages are invariably obtained from elongate weakly to moderately oscillatory zoned and planar to sector zoned, elongate magmatic cores that are overgrown and trun-

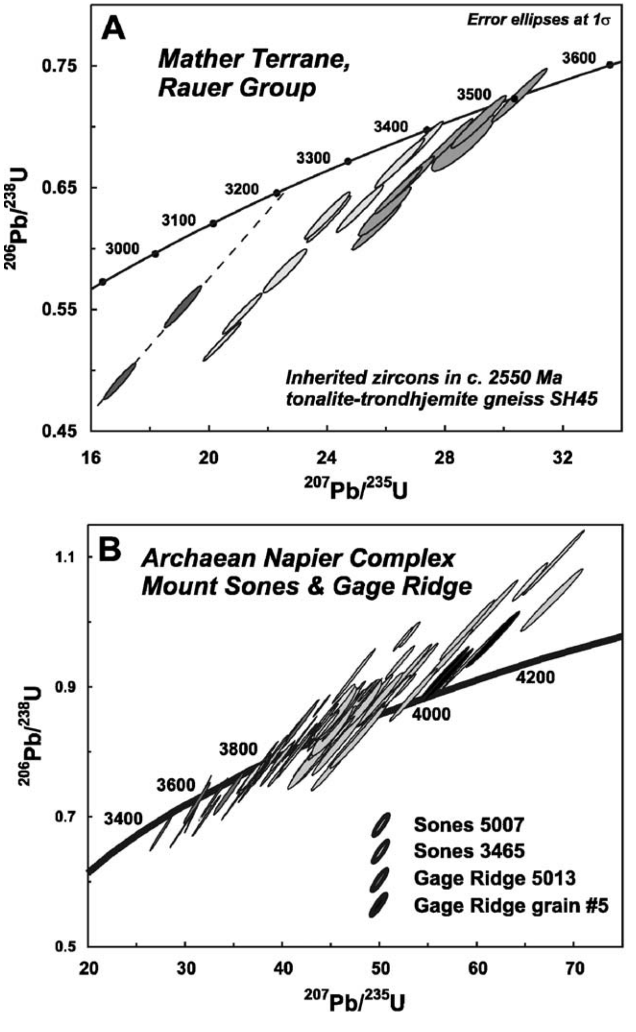


Fig. 3.2-6. (a) Concordia diagram illustrating the U-Pb systematics of inherited zircons (pre-2550 Ma) in the tonalitic-trondhjemitic orthogneiss sheet SH45 (location shown in Fig. 3.2-4). Note the presence of at least 3 sub-parallel alignments, distinguished by different shadings, controlled by Pb-loss at ca. 500 Ma from zircons with initial ages of ca. 3500–3480 Ma, ca. 3420–3400 Ma and ca. 3200 Ma. (b) Concordia diagram illustrating the collective U-Pb features of pre-3400 Ma zircon populations in the Mount Sones and Gage Ridge orthogneisses (locations shown in Fig. 3.2-8). Zircon U-Pb data (all at one sigma uncertainty) are from Williams et al. (1984), Black et al. (1986) and Harley and Black (1997). Note the presence of highly reverse discordant zircons in each of the three samples analysed, and the occurrence of distinctive near-concordant zircons that lie at older ages (>3950 Ma) than the bulk of the shared analytical population, which lies at ca. 3850 Ma. The three U-Pb analyses from grain #5 in the Gage Ridge orthogneiss are distinguished by the darkest shading.

cated either by the ca. 2550 Ma oscillatory zoned zircon or by bright CL and homogeneous rims (ca. 500 Ma).

The older zircon U-Pb analytical alignments have been re-examined in the light of the CL and BSE imaging of internal growth zones in the zircons using ISOPLOT/EX (Ludwig, 1999). Linear fits to these alignments, anchored at 500±20 Ma to account for the important isotopic effects of the Prydz event, yield concordia upper intercept ages of 3488 ± 13 Ma, 3421 ± 21 Ma and ca. 3200 Ma (Fig. 3.2-6(a)). The oldest concordant zircon core, with a $^{207}Pb/^{206}Pb$ age of 3501 ± 3 Ma, provides a lower limit for the age of the oldest local crustal sources available to be melted or incorporated in the tonalitic-trondhjemitic sheet, which are most likely to be the composite layered orthogneiss that hosts such sheets in the Scherbinina area. The spectrum of inherited zircon ages from ca. 3500 Ma to 3200 Ma is consistent with the multi-phase character of the composite layered orthogneiss, the ca. 3250–3350 Ma T_{DM} estimated for the later Archaean tonalites of the Mather Terrane, and the ca. 3270 Ma age of the Short Point orthogneiss described above (Table 3.2-1). As the least CL responsive, strongly metamict and spongiform zircon cores have been avoided in SHRIMP analysis of sample SH45, it is possible that older crustal components or sources may be present in the Mather Terrane, perhaps as old as the ca. 3750 Ma Nd model age obtained for sample 5120 (Fig. 3.2-4).

3.2-3.2. Reworked Early Archaean Crust in the Rayner Province of Kemp Land

The oldest Archaean material identified in the Oygarden Group of western Kemp Land (Fig. 3.2-2) is a layered composite orthogneiss (Figs. 3.2-7(a,b)), which in domains only weakly affected by the Rayner Structural Episode preserves a banding defined by the alignment of leucocratic segregations (Fig. 3.2-7(b)). Zircon domains in OG615, an example of the layered composite orthogneiss, are of three main types: core domains with oscillatory zoning or with modified (blurred/patchy) oscillatory zoning; core domains characterised by radial, sector and firtree zoning; and rims with homogeneous luminescence. The oldest preserved cores in this case were those with radial and sector zoning, with concordia intercept ages up to 3655 ± 15 Ma (Table 3.2-1). This age is interpreted to record the partial melting of the protolith to produce the leucocratic segregations, implying that the protolith is even older (Kelly et al., 2004). These cores form an analytical array along concordia down to an age of 3469±13 Ma, which is interpreted to approximate the age of a thermal/metamorphic event. A younger homogeneous felsic orthogneiss (OG525: Fig. 3.2-7(b)) intruded the layered composite orthogneiss either prior to, or synchronous with metamorphism and deformation at ca. 2840–2780 Ma (Table 3.2-1; Kelly et al., 2004).

Halpin et al. (2005) have presented zircon U-Pb age and Hf isotope data from zircons in several orthogneisses from the Kemp Land and MacRobertson Land coastal region, from the edge of the Napier Complex itself and into the Rayner Province as far as Mawson station. Four of their samples, analysed using laser ablation microprobe ICP-MS methods, yielded highly discordant zircon arrays with significant contributions from Archaean zircons that have $^{206}Pb/^{207}Pb$ ages between ca. 3000 Ma and 3600 Ma (Table 3.2-1). These gneisses included a felsic orthogneiss from Rippon Point (sample OG235), which lies

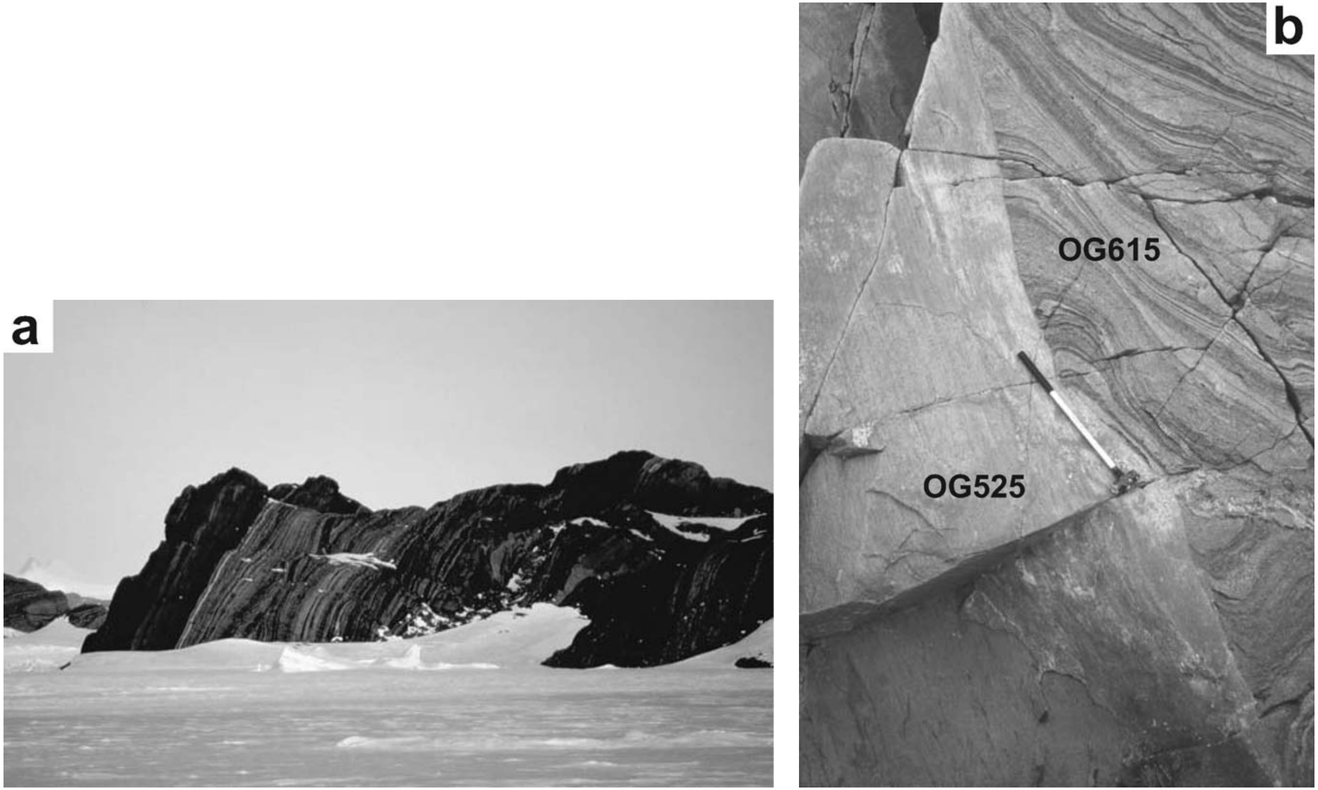


Fig. 3.2-7. Palaeoarchaean geological features and relations in the Oygarden Group of islands, Kemp Land Coast: (a) Composite layered gneisses, Shaula Island. This package includes paragneisses and banded mafic-felsic gneisses, all deformed into tight to isoclinal early D_1 folds and later intruded by homogeneous tonalitic orthogneiss similar to that depicted in Fig. 3.2-7(b) below (OG525) (the cliffs are ca. 80–100 metres high); (b) Palaeoarchaean composite orthogneiss OG615, which has yielded magmatic zircons up to ca. 3655 Ma old, intruded by ca. 2840–2780 Ma homogeneous orthogneiss OG525 (see text for details).

within the Napier Complex (Sheraton et al., 1987; Harley and Black, 1997), a duplicate sample of the Oygarden Group layered composite orthogneiss OG615 (Halpin et al., 2005, sample OG614), and orthopyroxene-bearing felsic orthogneisses from Broka-Havstein Islands and the Stillwell Islands to the east.

The zircon U-Pb data obtained on OG614 are consistent with those reported by Kelly et al. (2004). Homogeneous orthogneiss from the Broka-Havstein Islands preserves zircon cores as old as ca. 3540 Ma, a possible minimum age for its protolith, but the effects of later metamorphic events between ca. 2800 Ma and 2400 Ma are severe and hamper further interpretation. The Stillwell Hills Orthogneiss has a minimum protolith age of ca. 3460 Ma, based on the oldest surviving concordant oscillatory zoned zircon core analysis. The orthogneiss records extensive resetting to ca. 2970 Ma as well as the profound effects of recent Pb-loss on most zircons formed or reset between ca. 3460 Ma and 3000 Ma (Table 3.2-1; Halpin et al., 2005). In contrast to the Broka-Havstein orthogneiss the Stillwell Hills orthogneiss does not record any evidence in its zircon populations for resetting or regrowth at ca. 2400–2800 Ma.

The homogeneous felsic orthogneiss OG235 from Rippon Point, within the Napier Complex, contains zircons with relatively homogeneous to radial and sector zoned cores, as well as rare oscillatory zoned zircons. The latter, with concordia ages of ca. 3525 Ma, have been interpreted as inherited or xenocrystic as they are distinct in Th/U and age from the main zircon core population (Halpin et al., 2005), though it is possible that these record a minimum protolith age instead. A minimum age of 3422 ± 9 Ma for the earliest metamorphic/anatectic event (Table 3.2-1) is interpreted from the oldest zircon core preserving radial/sector zoning, whilst a second resetting event younger than 3173 ± 9 Ma has been invoked to explain the scatter of the oldest zircon analyses along concordia (Halpin et al., 2005). This sample also records a strong isotopic event affecting the zircons at 2468 ± 7 Ma, consistent with zircon U-Pb data obtained from all other localities in the Napier Complex (e.g., Black and James, 1983; Black et al., 1986; Harley and Black, 1997).

Zircon Hf isotope data from these orthogneisses is reported by Halpin et al. (2005) in terms of T_{DM}(Hf), ε_{Hf}(zircon age) and using a plot of zircon $^{207}Pb/^{206}Pb$ age against $^{176}Hf/^{177}Hf$ measured on the same grain domains. For the purposes of this discussion the details of the decay constant used and the depleted mantle model applied are not critical, as the key interpretations do not depend strongly on these but rather on how younger zircon was produced (new vs recrystallised), and what Lu/Hf is applied for ^{176}Hf ingrowth in the crust following its initial formation, if that proceeded zircon crystallisation by a significant time.

The Oygarden Group layered composite orthogneiss OG615/614, and most probably the Broka-Havstein orthogneiss (despite marked outliers in the Hf isotope data) have T_{DM}(Hf) similar to, or only slightly older than, their probable maximum crytallisation ages of ca. 3550–3650 Ma (Oygarden: ca. 3570 Ma; Broka: ca. 3600 Ma). These model ages are slightly older, near 3700 Ma, if the zircons are considered to have crystallised from a melt with average crustal Lu/Hf, in which some ^{176}Hf ingrowth would have occurred since initial crust formation. Older source ages are also possible if the mantle was more depleted than the DM model at ca. 3700–4300 Ma, as suggested by the data of Choi et al. (2006).

The Rippon Depot (Napier Complex) orthogneiss OG235 and Stillwell orthogneiss SW268 have, on average, lower or less evolved ^{176}Hf/^{177}Hf for their zircon ^{207}Pb/^{206}Pb age and so produce T_{DM}(Hf) model ages that are older than the most ancient U-Pb ages recorded in their zircons, and older than the samples noted above. T_{DM}(Hf) is ca. 3720–3860 Ma based on the most concordant grains (Table 3.2-1), and as old as ca. 3900–4050 Ma if ^{176}Hf ingrowth is assumed in the crust prior to the melting event that produced the oldest zircons. For these orthogneisses it is entirely feasible that ancient Napier Complex (Gage Ridge type) crust (>3850 Ma depleted sources) was remelted to produce the >3460–3530 Ma orthogneiss precursors.

In detail, the Archaean event record determined from the Oygarden Group and Kemp Land is not entirely comparable to that preserved in the adjacent Napier Complex itself. Whilst it is feasible and likely, given the Hf isotope data summarised above, that some of the layered composite orthogneisses include components as old as ca. 3850 Ma, the age of the earliest Napier Complex orthogneiss precursors, the ca. 3650 Ma partial melting/intrusive event and proposed metamorphism/thermal overprint at ca. 3450 Ma have not, as yet, been recorded in the Napier Complex of the Amundsen Bay region (Table 3.2-1; Fig. 3.2-8). On the other hand, the post-3200 Ma similarities are clear (Table 3.2-1). The ca. 2780 Ma age for the thermal event in the Oygarden Group is a minimum age, loosely constrained by the available data, and could potentially correspond to the ca. 2840 Ma low-pressure metamorphic episode inferred for the Napier Complex by Kelly and Harley (2005).

3.2-3.3. The Napier Complex

3.2-3.3.1. Fyfe Hills

Interest in Antarctica, and specifically the Napier Complex (Fig. 3.2-8), as a site where the most ancient rocks might be exposed was stimulated in the 1970s by the report of ca. 4000 Ma ages from the Fyfe Hills by Soviet scientists (Sobotovich et al., 1976). The Fyfe Hills (Figs. 3.2-8 and 3.2-9(a)) outcrop over an area of about 15 km^2 as two 5 km long ridges and smaller nunataks just inland from Khmara Bay in the southwest of Enderby Land (Fig. 3.2-8). Some 5–10 km west of the Fyfe Hills lie McIntyre Island and Zircon Point, sites that have also been the subjects of extensive and detailed zircon geochronology (Black et al., 1983b).

The old ages reported by Sobotovich et al. (1976) were derived from the slope of an alignment of whole rock analyses on an ^{207}Pb/^{204}Pb vs ^{206}Pb/^{204}Pb diagram, with enderbite and gneiss yielding apparent total rock ages of 4100 ± 100 Ma and 3700 ± 200 Ma respectively. Irrespective of the validity of the actual reported ages, the high ^{207}Pb with respect to ^{206}Pb in these rocks certainly indicate a prolonged residence of the sources for the rocks in a high U/Pb environment prior to metamorphism in the late-Archaean events at ca. 2840 Ma and ca. 2550–2480 Ma that characterise the Napier Complex (Kelly and Harley, 2005).

The geology of the Fyfe Hills Soviet sample site was described in detail by Black et al. (1983a), and complemented by a more regional description and analysis of the

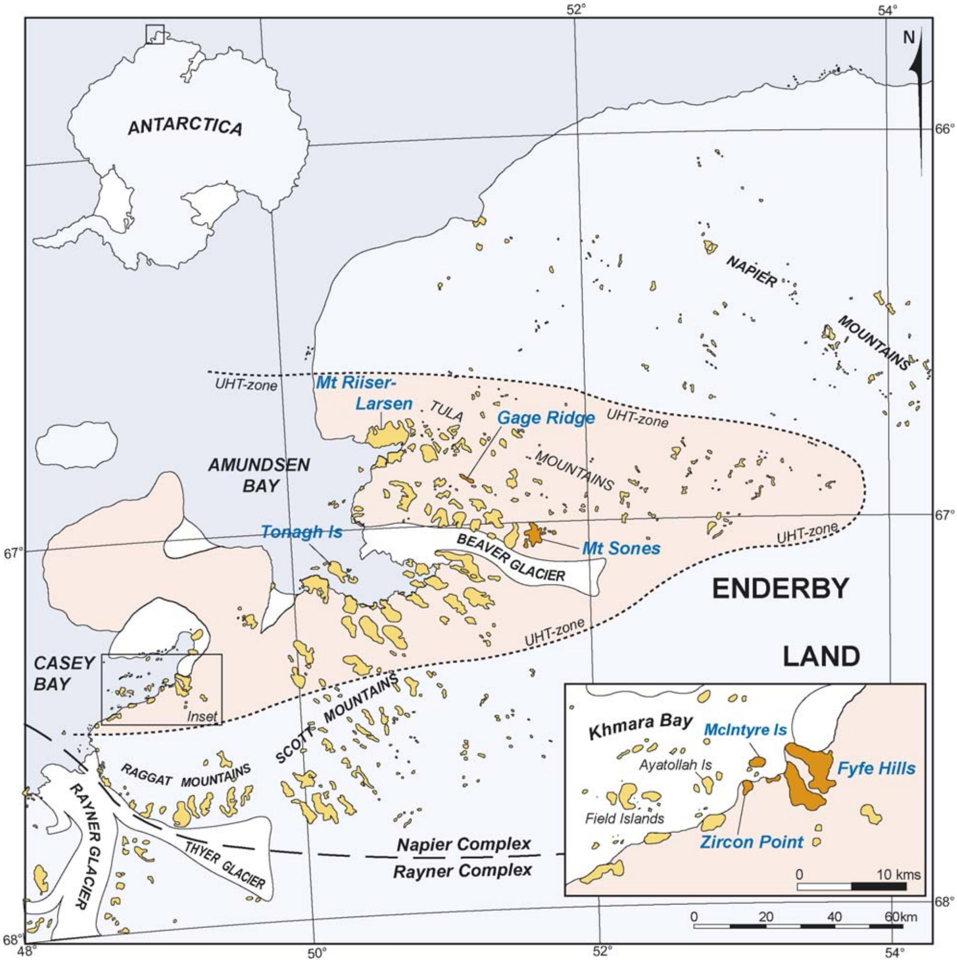


Fig. 3.2-8. Map of the western part of the Napier Complex, Enderby Land, showing the geographical positions of the localities discussed in the text (e.g., Mount Sones, Gage Ridge, Fyfe Hills, Mount Riiser-Larsen, Tonagh Island). Regional extent of ultrahigh-temperature mineral assemblages such as sapphirine + quartz is denoted by the shaded field enclosed by the dashed boundary. The Napier Complex is bound southwards by the Proterozoic Rayner Complex, along an EW trending zone that preserves Cambrian high-grade metamorphism. Area of main map is shown by the small box on the map of Antarctica, and in Fig. 3.2-2. Inset box shows the localities in the Casey Bay region (marked by a small outline box) referred to in the text.

Fyfe Hills by Sandiford (1985). The Soviet Sample site largely comprises a moderately to steeply NE-dipping (60/045) isoclinally folded charnockitic orthogneiss typified by a strong modal layering and planar fabric defined by ribbon quartz, mesoperthitic K-feldspar

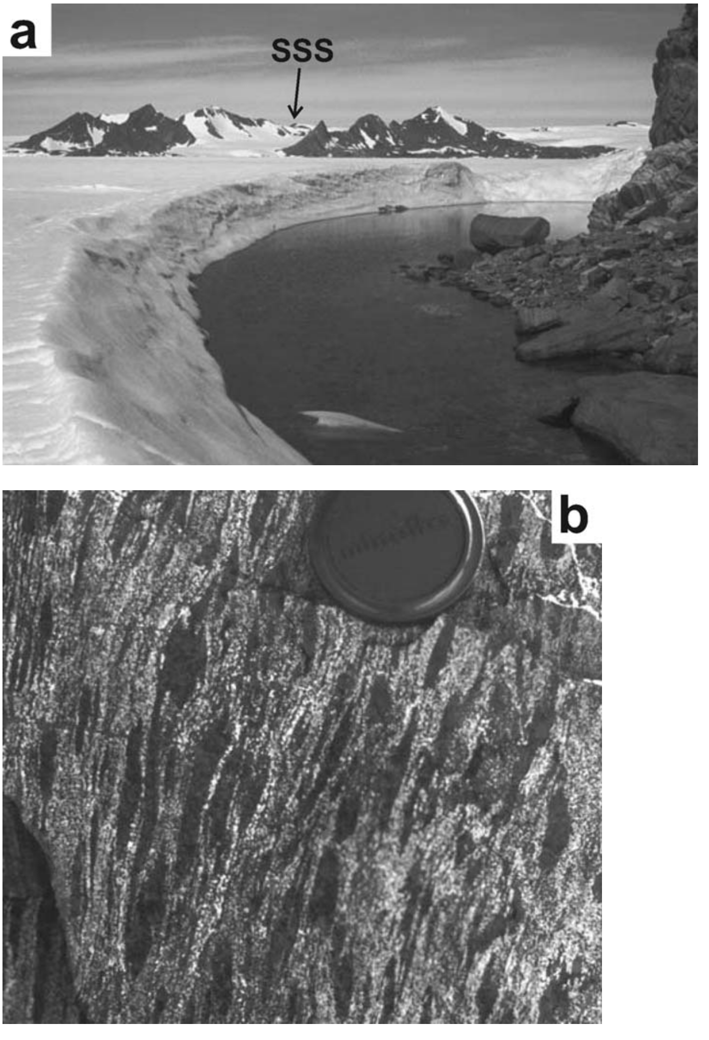


Fig. 3.2-9. Archaean geological features and relations at Fyfe Hills, Enderby Land. See inset map in Fig. 3.2-8 for localities: (a) View of Fyfe Hills, looking east from Khmara Bay. The Soviet Sample Site (SSS) is denoted by the arrow; (b) F_1 isoclinal fold hinges and strong lensoidal fabric defined by trails and clusters of clinopyroxene alternating with plagioclase in a mafic granulite, Fyfe Hills Soviet Sample Site. This rock was originally interpreted by Black et al. (1983a) as a metagabbro, but subsequent geochemistry by Sandiford (1985) suggested a marl/altered basic volcanic protolith; (c) Isoclinal F_1 fold in granodioritic orthogneiss, Fyfe Hills Soviet Sample Site. A strong S_1 fabric axial planar to the fold is defined by the alignment of ribbons of quartz (dark) in the more granitic layers; (d) Lance Black contemplating the layered charnockitic gneiss sampled for Rb-Sr, Sm-Nd and zircon U-Pb from the Soviet Sample Site, and reported in Black et al. (1983a), McCulloch and Black (1984), and Black and McCulloch (1987).

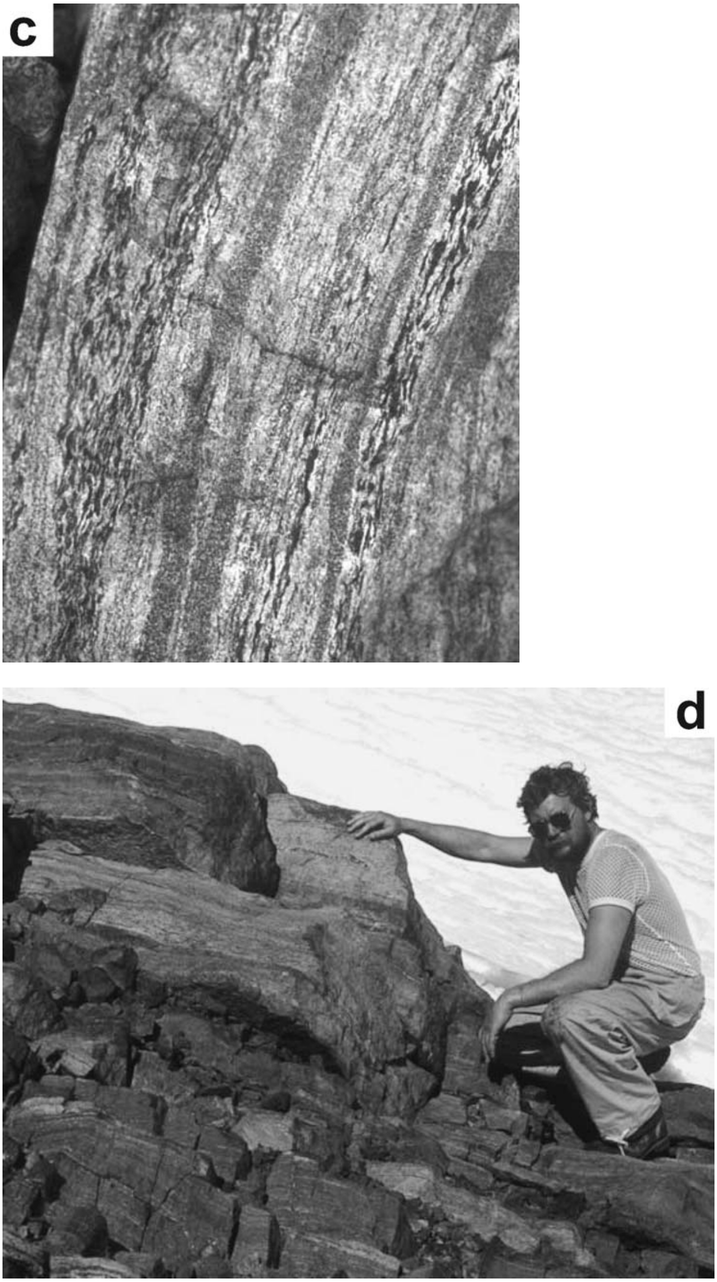


Fig. 3.2-9. (*Continued.*)

and orthopyroxene (Fig. 3.2-9(b,c)). Minor interlayered quartzitic gneiss, garnet-quartz gneiss, pyroxene-magnetite-quartz metaironstone, pyroxene granulite and a clinopyroxene-plagioclase-titanite gneiss, derived from a meta-calcareous or marly precursor, are also present (Fig. 3.2-9(d)) (Sandiford, 1985). This suite of compositionally diverse gneisses, including Mn-Fe rich rocks, are considered to represent a supracrustal package deposited prior to at least ca. 3100 Ma. Leucocratic and melanocratic metre to several metre scale boudins of pyroxene-plagioclase gneisses also occur, and are interpreted by Black et al. (1983a) to be metamorphosed equivalents of layered mafic-anorthositic bodies. All rocks are intensely folded to produce isoclinal structures on decimetre to metre scales (Fig. 3.2-9(c)). Strong platey and ribbon fabrics defined by quartz, feldspars or aggregates of pyroxenes, and NE plunging lineations were produced in the first recognised UHT deformation (termed D_1; Fig. 3.2-9(c)). These flat-lying to moderately dipping structures and layering are locally refolded about close/isoclinal F_2 folds that formed also under granulite facies conditions but are distinguished by their re-orientation of the earlier fold closures and folding of the earlier-formed axial planar fabrics (e.g., Black and James, 1983; Black et al., 1983b; Sheraton et al., 1987). Overprinting both events are recrystallisation seams and deformation bands associated with a third regional deformation that domes the previously flat-lying to reclined gneiss suite on km to tens of km scales (D_3 in Black et al., 1983a, 1983b) and is responsible for the current moderate to steep orientation of the gneisses.

Black et al. (1983a) re-evaluated the earlier age data from the Soviet Sample site and presented new Rb-Sr analyses from charnockitic gneiss, leuconoritic gneiss and a variety of mafic granulites, and multi-grain zircon U-Pb TIMS analyses from a charnockitic leucogneiss. The key Rb-Sr result relevant to establishing the presence or otherwise of early Archaean crustal precursors were that large blocks of charnockitic gneiss sampled from an approximately 1 m^3 volume yielded an imprecise errorchron with an age of 3120 (+230, −180) Ma, and an initial ratio ($^{87}Sr/^{86}Sr$) of 0.725 ± 0.006. The high initial ratio associated with this analytical alignment yields a T_{UR} Sr model age of ca. 3900 Ma, consistent with the presence of earliest Archaean crustal precursors. The zircon U-Pb age data obtained by Black et al. (1983a) defined arrays approximating discordia between ca. 2450 Ma and ca. 3100 Ma. One clear zircon fraction, which yielded a much higher $^{207}Pb/^{235}U$ ratio than all other fractions and gave a $^{207}Pb/^{206}Pb$ upper intercept age of ca. 2900 Ma, suggested the presence of significantly older material but was also the most discordant of the analyses. Black et al. (1983a) concluded that, whilst early Archaean crust may exist at the Fyfe Hills, the effects of intense deformation and metamorphic events at ca. 3100 Ma and younger ages largely obliterated the evidence for initial crust formation.

DePaolo et al. (1982) reported Sm-Nd, Rb-Sr and Pb-Pb isotopic data on a variety of gneisses from the Soviet Sample site, provided by Soviet workers. These gneisses included the dominant quartzofeldspathic orthogneiss (charnockitic gneiss), garnet-bearing granulites, pyroxene granulite and a metaironstone. Like Black et al. (1983a) and McCulloch and Black (1984), these workers documented late Archaean disturbance of the Rb-Sr and U-Pb systems, and potentially some disturbance of the Sm-Nd whole rock system also. If we consider only charnockitic orthogneisses and the pyroxene granulites the T_{DM} of the dominant rock suites at Fyfe Hills lie in the range ca. 3200–3500 Ma, implying a significant crustal pre-history and consistent with the conclusions of Black et al. (1983a) that no ca. 3900–4000 Ma crust occurs in the Fyfe Hills, and possibly not in the entire area of the Napier Complex to the south of Amundsen Bay. This suggestion is consistent with the available zircon U-Pb data obtained from corroded zircon cores in metasedimentary paragneisses from Khmara Bay (Harley and Black, 1997; Kelly and Harley, 2005) and Mt Cronus (Harley et al., unpubl data). Inherited cores in all cases yield concordant or near-concordant ages no older than ca. 2970 Ma. If any early Archaean crust was present in this region it cannot have been available as a source of detritus for the sedimentary precursors of these paragneisses.

3.2-3.3.2. Mt Sones and Gage Ridge

Mount Sones lies in the Tula Mountains, east of Amundsen Bay (Figs. 3.2-8 and 3.2-10). The essential geology and structure of Mount Sones is typical of that observed in the Tula Mountains as a whole (Harley, 1986). Two distinctive gneiss suites or packages are present. Much of the northern and northwestern parts of Mount Sones consists of a

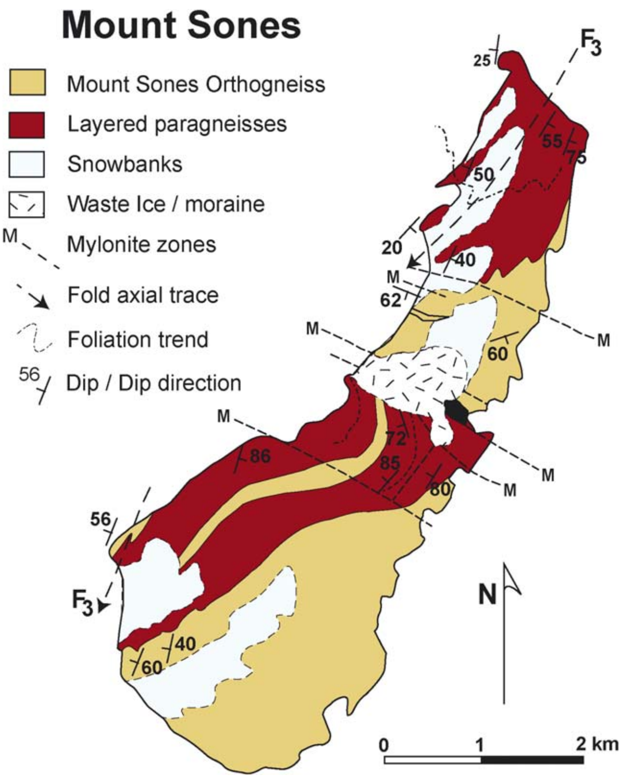


Fig. 3.2-10. Geological map of Mount Sones, modified after Harley (1986). See Fig. 3.2-8 for the location of Mount Sones within Enderby Land.

strongly deformed and compositionally diverse package of paragneisses and orthogneisses (Fig. 3.2-10), whereas the southeastern and southern sides of the mountain comprise a homogeneous, orthopyroxene-bearing, tonalitic to granodioritic orthogneiss – the Mount Sones Orthogneiss (Fig. 3.2-11(a)).

The paragneiss package (Fig. 3.2-11(b,c)) is composed of quartzo-feldspathic leucogneiss, pyroxene granulite, garnet quartzite, garnet-sillimanite pelitic gneiss, boudinaged ultramafic gneisses (diopsidites, bronzitites) and sapphirine-spinel-pyroxenite (Harley, unpubl.), and a rare sapphirine-cordierite-garnet-sillimanite-mesoperthite gneiss described in detail by Harley (1986). These are interleaved and layered on decimetre to metre scales, and characterised by strong SW-plunging mineral elongation lineations (Fig. 3.2-11(d)). Foliation and compositional banding in the layered gneisses and Mount Sones Orthogneiss is generally steeply SE- or NW-dipping over much of Mount Sones (Fig. 3.2-10), but SW-dipping in the hinge region of an open, kilometre-scale SW-plunging antiformal fold (F_3 of Black and James, 1983; Fig. 3.2-10) exposed on the northern and western flanks of the mountain. All gneisses and the F_3 folds in them, are cut by later ESE-trending upright mylonite zones (Fig. 3.2-11(b,c)) that may be up to 100 m wide, and which preserve amphibolite facies mineral assemblages (e.g., garnet-hornblende-plagioclase). Near the middle of Mount Sones, the layered gneiss package and Mount Sones Orthogneiss are brought into contact along a NE-trending mylonite zone that appears to be cut by the ESE-trending generation (Figs. 3.2-10 and 3.2-11(c)).

The Mount Sones Orthogneiss preserves extensive zircon U-Pb evidence for an early Archaean protolith age (Table 3.2-1; Black and James, 1983; Black et al., 1986; Harley and Black, 1997). Based on conventional multi-grain TIMS analyses of zircons in sample 78285007, a strongly HREE-depleted orthogneiss (Sheraton et al., 1987), Black and James (1983) proposed that the orthogneiss precursor may include a component as old as ca. 3700 Ma. This work was subsequently complemented by detailed SHRIMP ion microprobe analysis of zircon grain separates, with the classification of zircon type and microtexture controlled by polarised light observations (Williams et al., 1984; Black et al., 1986). These SHRIMP studies confirmed the presence of zircon overgrowths and structureless grains with ages typical of many in the Napier Complex (ca. 2950–2850 Ma and ca. 2450–2550 Ma). They also demonstrated the presence of zircon cores and moderately to weakly oscillatory-zoned zircon euhedra that produced near-concordant ages ranging between ca. 3500 and 4000 Ma. The U-Pb isotopic data and zircon features were very complex and open to interpretation in terms of the precise age significance of populations because of evidence for ancient movement of radiogenic Pb with respect to parent U and Th (Williams et al., 1984). However, Black et al. (1986) were able to infer, from the intersection with concordia of the discordia defined by the zircon U-Pb array, an age of 3927 ± 10 Ma as their best estimate of the age of the zircon cores and hence of the crustal precursor to the Mount Sones Orthogneiss. Until the discovery and dating of the Acasta Gneiss in Canada, this age gave the Mount Sones Orthogneiss sample 78285007 the status of the oldest terrestrial rock.

One of the outstanding features of the Mount Sones Orthogneiss zircon data presented by Black et al. (1986) was the large number of core analyses that were reversely discordant,

Fig. 3.2-11. Archaean geological features and relations at Mount Sones, Enderby Land: (a) View, looking SSE from its central moraine area, of the summit ridge of Mount Sones. The highest crag on the E edge of the ridge (elevation ca. 850 metres) consists of the Mount Sones Orthogneiss, which also crops out as the dark horizon/lens forming the nearer summit of the ridge. Layered paragneisses occur either side of this 100–200 metre thick horizon, and make up the exposures nearest to the tent; (b) View, looking east from near the campsite, of steeply N-dipping 10–50 metres wide mylonite and ultramylonite zones (highlighted by dashed lines and M symbols) cutting layered quartzitic, quartzofeldspathic and semi-pelitic layered paragneisses in central Mount Sones. Ridge is approximately 250 metres high; (c) Aerial view, looking west, of the eastern flank of Mount Sones, showing the extensive layered paragneiss package in contact with the Mount Sones Orthogneiss. Both are cut by mylonite zones, highlighted by dashed lines and M symbols. Cliff face in the central foreground is ca. 350 metres high; (d) SSW-plunging mineral elongation lineation in garnet-bearing quartzofeldspathic gneiss, Mount Sones. The strong lineation associated here with D_1 structures is defined by quartz rods and ribbons, and by the elongation of feldspar aggregates. The photo is taken looking onto the S_1 gneissosity plane.

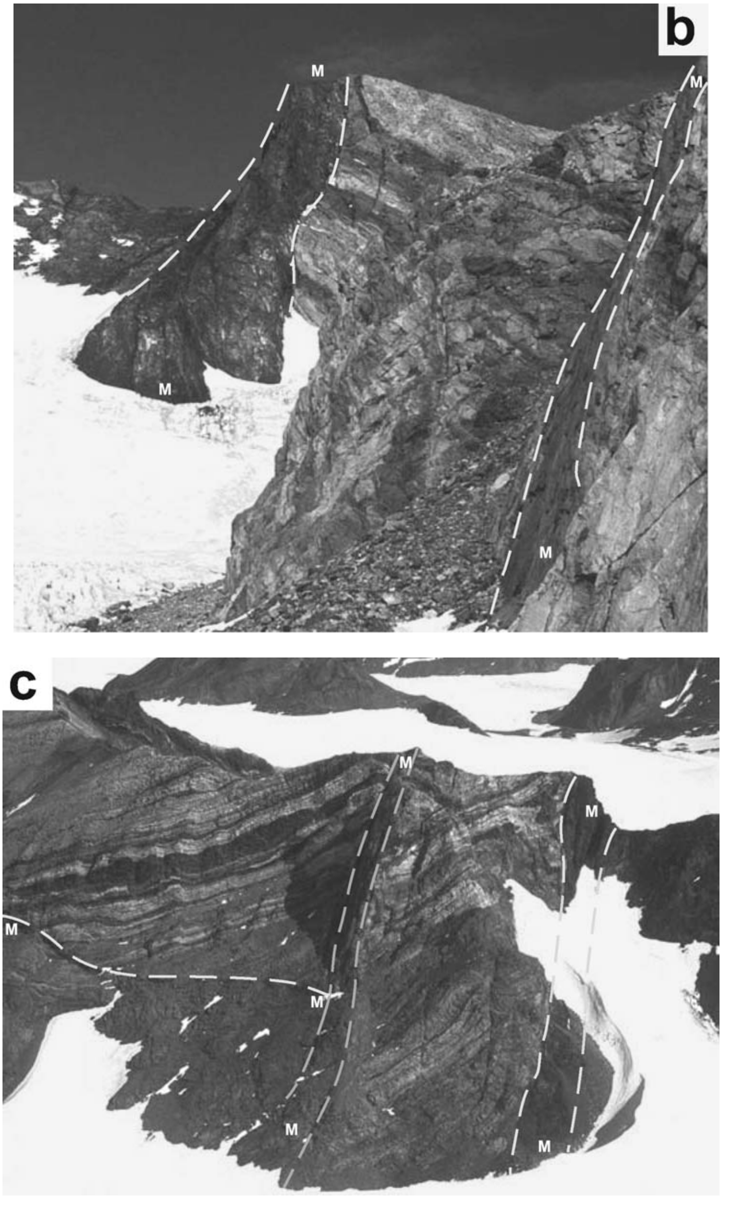


Fig. 3.2-11. (*Continued.*)

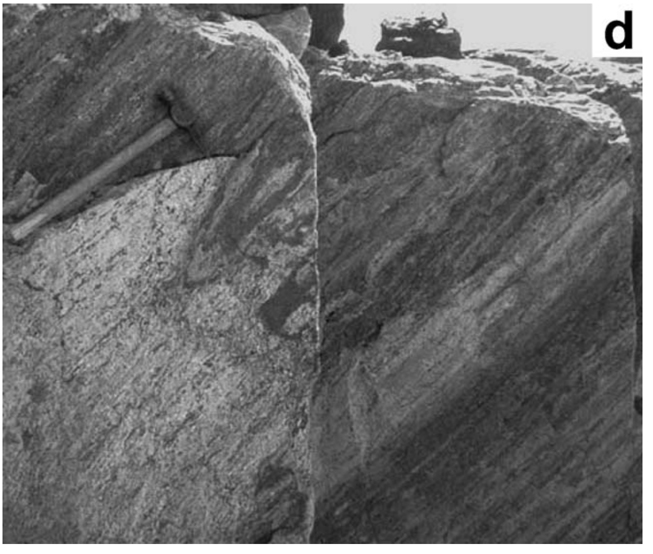


Fig. 3.2-11. (*Continued.*)

lying on a discordia that intersected the concordia curve at ca. 3930 Ma. As noted above, this feature was interpreted as resulting from ancient movement of Pb relative to its parent elements so that on the scale of an ion microprobe analysis pit (20–30 microns diameter) some zircon domains had unsupported radiogenic Pb in them (Williams et al., 1984). In the light of this, further zircon U-Pb analyses were undertaken on the Mount Sones Orthogneiss sample 78285007, another sample of orthogneiss from Mount Sones (sample 77283465), as well as of an equally ancient tonalitic orthogneiss from Gage Ridge (sample 78285013), some 10 km to the west (Harley and Black, 1997). In the case of sample 78285007, Harley and Black (1997) produced an age of 3800 (+50/−100) Ma by fitting all their new zircon analyses to a line with a lower intercept at 2950 Ma (±350 Ma). Despite the use of improved calibrational procedures their analyses also include a significant number that are reversely discordant, even though the common Pb is very low and the U contents not high (53–340 ppm). In other words, their data were similar to that reported in Black et al. (1986) and the range of U-Pb analyses must be an inherent feature of the old zircons themselves; that is, the reverse discordance is not linked to any measured chemical signal (e.g., U content, Th/U).

The Mount Sones Orthogneiss zircon datasets of Black et al. (1986) and Harley and Black (1987) have been re-examined here using ISOPLOT/EX (Ludwig, 1999). These data have been fitted by regressing analyses that are older than 3500 Ma and <5% discordant (Fig. 3.2-6(b)). A population age of 3844 ± 81 Ma is obtained based on the Black et al. (1986) data, using only the pre-3500 Ma grains, excluding the most discordant analyses, and with the lower intercept anchored at 2500 Ma. This anchored age is chosen as it approximates the age of the most ubiquitous isotopic resetting event recorded in the Napier Complex. The concordia population age is marginally older, but within error (3854 ± 90 Ma), when the lower intercept of the regression allowed to be a free parameter (1705 ± 860 Ma). By the same procedure, the Harley and Black (1997) data yields a population age of 3863 ± 53 Ma when the lower intercept is anchored at 2500 Ma. The anchored 'age' determined from all pre-3500 Ma analyses in the dataset produced by combining both studies is 3850 ± 50 Ma, identical to the free regression age of 3846 ± 48 Ma (with a highly imprecise lower intercept of 2501 ± 350 Ma) derived from the same data. The general result is that the main zircon population in the orthogneiss precursor, presumably recording the magmatic protolith age, formed at ca. 3850±50 Ma (Table 3.2-1). Older near-concordant grains some ca. 3980–4050 Ma old may be inherited, especially given the Sm-Nd systematics of sample 78285007 (Black and McCulloch, 1987), which has an average ε_{Nd} of 0 at 3840 Ma (T_{CHUR} of 3840 Ma) and hence a model age (T_{DM}) close to ca. 3960–4040 Ma (Table 3.2-1).

Limited zircon data obtained from a more layered tonalitic gneiss (sample 77283465) from the northwestern flank of Mount Sones is consistent with that described above for sample 78285007 (Fig. 3.2-6(b)), and yielded a concordia intercept age of 3773 (+13/−11) Ma in the analysis of Harley and Black (1997). Regression of this data using ISOPLOT-E/X (Ludwig, 1999) produces the same intercept age, but with a larger error (3773 ± 83 Ma). Anchoring the lower intercept at 2500 Ma following the logic applied to sample 78285007 produces a concordia upper intercept age of 3805 ± 65 Ma.

The granitic Gage Ridge Orthogneiss (sample 78285013) has been dated using SHRIMP zircon U-Pb analysis by Harley and Black (1997) and re-evaluated in the light of CL and BSE textural observations and REE analysis by Kelly and Harley (2005). The zircon U-Pb data presented by Harley and Black (1997) followed two alignments that were regressed separately, although there were no obvious differences in the morphologies or U and Th contents of the zircons placed into each alignment. The older alignment, which included reversely discordant analyses with apparent $^{206}Pb/^{238}U$ ages of >4000 Ma, yielded a concordia intercept age of 3840 (+30/−20) Ma, taken as the age of the orthogneiss protolith. Kelly and Harley (2005) distinguished elongate cores with variably preserved oscillatory zoning, traversed in some grains by networks of low luminescence healed fractures. These cores were interpreted as magmatic based on their oscillatory zoning, steep chondrite-normalised HREE patterns with Yb/Gd near 20, large negative Eu anomalies, and Th/U ratios of greater than 0.05. Undisturbed core analyses together with cores exhibiting minor to moderate textural modification (e.g., weakening or blurring of oscillations, presence of healed microfractures) define a discordant array that produces a concordia upper intercept age of 3851±62 Ma and lower intercept of 2740±110 Ma (Kelly and Harley, 2005). This same data array of 30 analytical points yields a concordia age of 3877 ± 62 Ma when a regression anchored at 2500 Ma is used, and 3889 ± 75 Ma if only the oldest grains (17 analyses at >3500 Ma) from the population are used in the anchored regression.

In detail the Gage Ridge zircon array includes four analyses (Fig. 3.2-6(b)) that are distinct in being older than ca. 3940 Ma based on both $^{206}Pb/^{238}U$ and $^{206}Pb/^{207}Pb$ ratios. The three 'oldest' and reversely discordant of these distinct analyses are all from one

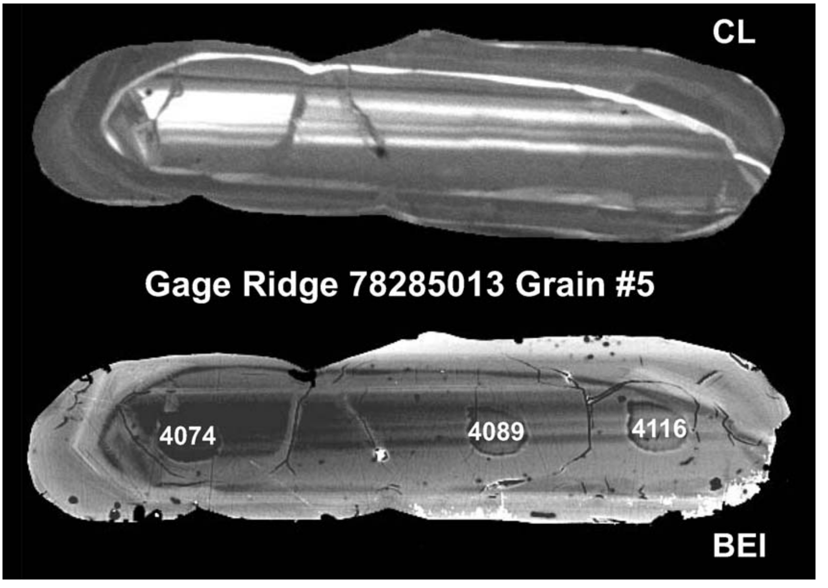


Fig. 3.2-12. Hadean zircon? The oldest zircon grain in Antarctica, Grain #5 from Gage Ridge orthogneiss sample 78285013. The averaged $^{207}Pb/^{206}Pb$ age of ca. 4090 Ma for this grain, based on the three SHRIMP U-Pb analyses (Harley and Black, 1997), when coupled with the Hf isotopic measurements of Choi et al. (2006), gives an ε_{Hf} of 8.3. This implies either that the source for the magma in which this zircon crystallised was ultra-depleted mantle, or that the U-Pb ages are too old despite their proximity to concordia.

grain (grain #5; Fig. 3.2-12), whilst a concordant zircon with an age of ca. 3940 Ma is derived from another zircon grain (#28). Exclusion of these grains from the regressed array of analyses with $^{206}Pb/^{238}U$ ratios consistent with an age >3500 Ma yields a concordia upper intercept at 3810 ± 110 Ma, compatible with the accepted ca. 3850 Ma age of the orthogneiss protolith and similar to that of the Mount Sones orthogneiss sample 78285007 (Table 3.2-1; Fig. 3.2-6(b)).

Choi et al. (2006) have analysed the Hf isotopic composition of four zircon grains from the Gage Ridge Orthogneiss, including the oldest grains #5 and #28. For a protolith age of ca. 3850 Ma the zircons yield ε_{Hf} in the range 2.5 ± 0.3 to 5.6 ± 0.4 (with $^{176}Hf/^{177}Hf$ calculated for 3850 Ma using $\lambda^{176}Lu$ of 1.865×10^{-11} yr and CHUR parameters of $^{176}Lu/^{177}Hf = 0.0332$ and $^{176}Hf/^{177}Hf = 0.282772$). These highly positive ε_{Hf} values imply juvenile sourcing of the orthogneiss protolith at ca. 3850 Ma from mantle that was highly depleted relative to CHUR, and which had been depleted from early in the Earth's history. In order to generate the high ε_{Hf} values the time integrated Lu/Hf fractionation, $f_{Lu/Hf}$ $(=[^{176}Lu/^{177}Hf]_{mantle} / [^{176}Lu/^{177}Hf]_{CHUR} - 1)$ is required to be as high as 0.51, implying the presence of strongly depleted early Archaean mantle irrespective of the baseline (e.g., CHUR) parameters used (Choi et al., 2006).

Grain #5, labelled with the $^{207}Pb/^{206}Pb$ ages obtained on each analysis spot, is illustrated in Fig. 3.2-12. If this grain and grain #28 are inherited, and not related to crystallisation of the granite protolith at ca. 3850 Ma, then their ε_{Hf} would be even higher than the values given above, using the same decay constant and CHUR parameters. For an averaged $^{207}Pb/^{206}Pb$ age of ca. 4090 Ma this grain would have ε_{Hf} of 8.3, and for its $^{207}Pb/^{206}Pb$ age of ca. 3940 Ma the near-concordant grain #28 would have an initial ε_{Hf} of 7.7. These extreme ε_{Hf} values would imply the existence of an ultra-depleted early mantle source and hence the complementary existence of early, highly enriched, continental crust since the early Hadean as proposed in other studies (e.g., Bennett et al., 1993; Blichert-Toft et al., 1997; Bizzarro et al., 2003). This tantalising speculation requires further investigation through Hf isotope analysis of more of these earliest Archaean zircons from the Napier Complex, especially those that are concordant or have negligible reverse discordance.

3.2-4. CONCLUSIONS

The Archaean event histories described above are summarised on a terrane, province or area basis in Table 3.2-1. Even though the ice cap covers most of the East Antarctic Shield, limiting the total area of exposed of Archaean rocks to coastal strips and two regions of inland nunataks, it is clear from this summary that there is considerable potential for Archaean crust to form a major part of the unexplorable interior of the continent. This is particularly the case for the vast region bounded by Terre Adélie and the Miller range in the east and southern Prince Charles Mountains in the west. However, it is not likely, given the differences between the geology and age-event history of the Ruker Terrane and the Archaean preserved in the 'Mawson Block' that these represent one large craton (Table 3.2-1; Boger et al., 2006).

The oldest rocks in Antarctica, represented by early Archaean tonalites, granodiorites and granites with emplacement ages from ca. 3850 Ma to 3400 Ma, all occur in terranes that lie in the sector to the south of India in Gondwana reconstructions, bounding the Rayner Province and adjacent Prydz Belt. Although some similarities exist in the age-event chronologies recorded for each terrane, the only areas that share similar early Archaean events are the Napier Complex and the adjacent part of the Rayner Province in Kemp Land (Table 3.2-1), where an anatectic/metamorphic event at ca. 3420–3450 Ma is locally preserved. However, even in this instance there are important differences in the ages of the oldest preserved rocks: in Kemp Land these are arguably no older than ca. 3650 Ma whereas the evidence for ca. 3850 Ma orthogneiss precursors in the Napier Complex of Enderby Land is robust.

The key conclusion to emerge from this analysis is that there is no compelling reason to correlate across and between the early Archaean terranes and vestiges that now lie situated around the Rayner Province (Table 3.2-1). Based on our present state of knowledge it is likely that the Mather and Ruker terranes were always separate in the Archaean, not sharing any geological events until being reworked in either the latest Mesoproterozoic or, more certainly, in the Cambrian. The Rayner Province in Kemp Land contains some components of reworked Napier Complex, but also appears to contain Archaean crust not yet identified in the Napier Complex proper. Given this, and despite T_{DM}(Nd) model ages suggesting initial extraction of crust at ca. 3700–3900 Ma, it is highly unlikely that crustal rocks of similar antiquity to the ca. 3850 Ma Mount Sones and Gage Ridge orthogneisses will be found in East Antarctica outside of the Napier Complex, where such old rocks only form a minor component of the orthogneiss suites in any case. Hence, the Napier Complex must remain the prime target of efforts in East Antarctica to find the earliest Archaean to Hadean rocks that can be used to evaluate models for Archaean tectonics, crust production and geochemical evolution of the crust-mantle system.

ACKNOWLEDGEMENTS

Discussions over the years on various aspects of Antarctic crustal evolution with numerous Antarctic petrologists and geochronologists have influenced this synthesis, and we thank all of these for their contributions to Antarctic geology. Particular thanks go to Lance Black, John Sheraton, Mark Fanning, Peter Kinny, Ian Fitzsimons, Ian Snape, Chris Carson, Steve Boger, Ed Grew, Tomokazu Hokada, Yoichi Motoyoshi, Kazuyaki Shiraishi, Joachim Jacobs, Bas Hensen and Dan Dunkley for their insights and discussions. We are indebted to Steve Boger for providing us with the photographs of Ruker Terrane geology used in Fig. 3.2-3. Microanalytical and SEM imaging work contributing to this paper (SH45, 78285013) has been supported by a NERC large grant to SLH. NMK was supported by a Royal Society of Edinburgh Fellowship for part of the work that is included in this chapter. We thank Martin Van Kranendonk for his editorial work on our behalf.