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ARCHAEAN AND PALAEOPROTEROZOIC GEOLOGY OF THE NARRYER TERRANE (YILGARN CRATON) AND THE SOUTHERN GASCOYNE COMPLEX (CAPRICORN OROGEN) WESTERN AUSTRALIA — A FIELD GUIDE

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Geological Survey of Western Australia

GEOLOGICAL SURVEY OF WESTERN AUSTRALIA

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S. A. Occhipinti[1], S. Sheppard[2], J. S. Myers[3], I. M. Tyler[2], and D. R. Nelson[2]

[1] Now at Tectonics Special Research Centre, Curtin University of Technology, Western Australia
[2] Geological Survey of Western Australia
[3] Now at Department of Earth Sciences, Memorial University of Newfoundland, Canada

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Information Centre
Department of Mineral and Petroleum Resources
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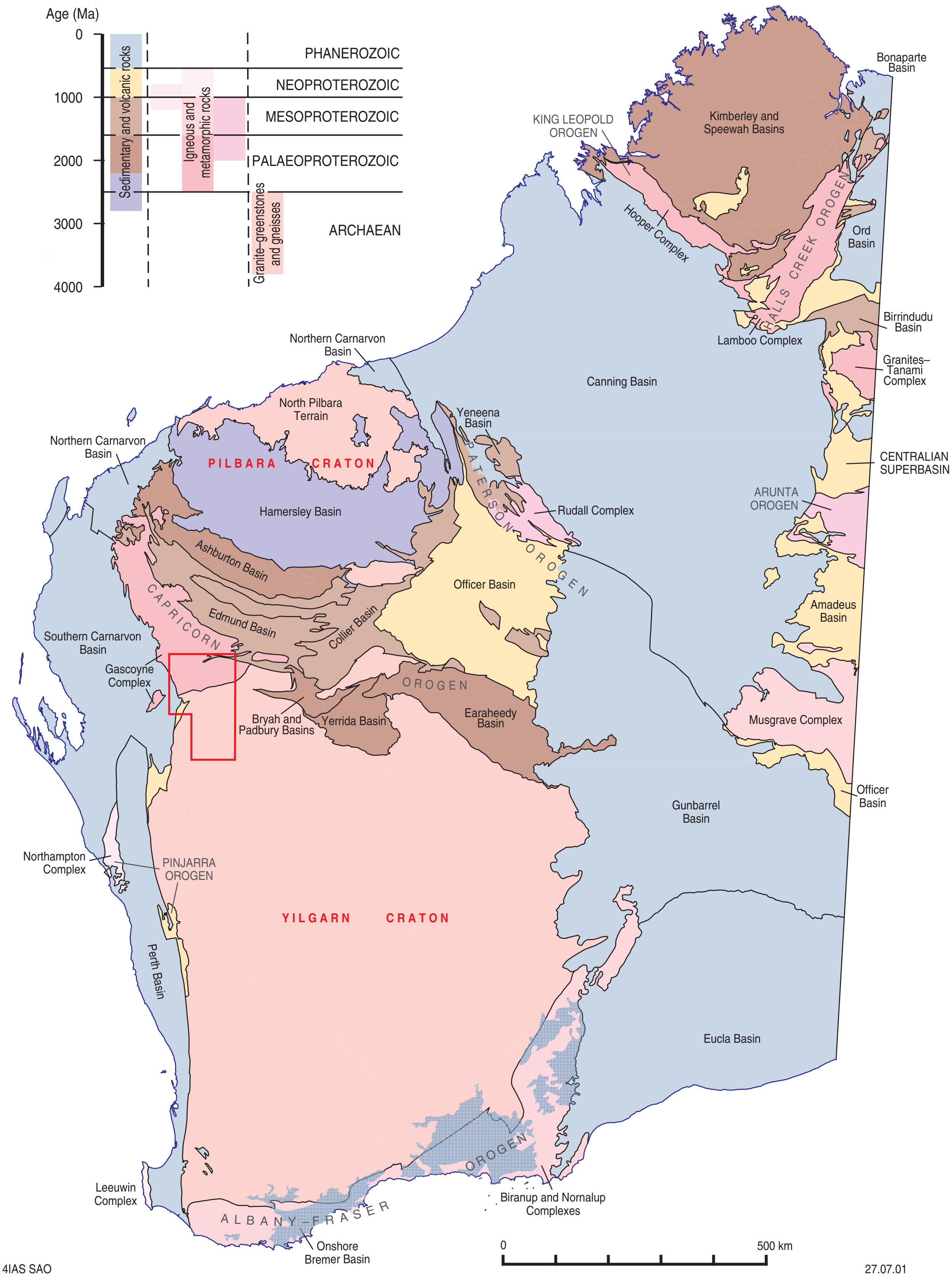

Archaean and Palaeoproterozoic geology of the Narryer Terrane (Yilgarn Craton) and the southern Gascoyne Complex (Capricorn Orogen), Western Australia — a field guide

by

S. A. Occhipinti[1], S. Sheppard[2], J. S. Myers[3], I. M. Tyler[2], and D. R. Nelson[2]

Introduction

by S. Sheppard and S. A. Occhipinti

This field guide is for an excursion of the 4th International Archaean Symposium (4IAS) to the northwest Yilgarn Craton and adjacent southern Gascoyne Complex. The excursion will examine:

- ancient gneisses and metasedimentary rocks of the Narryer Terrane;
- the nature of the boundary between the Archaean Yilgarn Craton and the latest Archaean to Palaeoproterozoic Gascoyne Complex;
- rocks of the newly identified 2540–1970 Ma Glenburgh Terrane of the Gascoyne Complex;
- the nature of metamorphism, deformation, and magmatism during the 2000–1960 Ma Glenburgh Orogeny and the 1830–1780 Ma Capricorn Orogeny.

The field excursion has three main geological components: the Narryer Terrane, the Errabiddy Shear Zone (southern margin of the Gascoyne Complex and the northern margin of the Narryer Terrane), and the Glenburgh Terrane (southern Gascoyne Complex).

The first part of the excursion visits localities in the Mount Narryer region to examine the oldest known rocks in Australia, and metasedimentary rocks containing detrital zircons that are remnants of the most ancient terrestrial rocks (>4000 Ma) on Earth. The second and third parts of the excursion illustrate the results of a recent Geological Survey of Western Australia (GSWA) program integrating regional mapping, geochronological, structural, metamorphic, and geochemical studies in the latest Archaean to Palaeoproterozoic Gascoyne Complex. This work has identified:

- latest Archaean granitic basement in the Gascoyne Complex that is younger than dated basement rocks in either the Yilgarn or Pilbara Cratons;
- a terrane of Palaeoproterozoic (2005–1975 Ma) igneous and metamorphic rocks, which has few time equivalents in Australia;
- a high-grade orogenic event more than 150 million years older than the widely recognized Capricorn Orogeny.

[1] Now at Tectonics Special Research Centre, Curtin University of Technology, Western Australia
[2] Geological Survey of Western Australia
[3] Now at Department of Earth Sciences, Memorial University of Newfoundland, Canada

Location, access, physiography, and climate

The area covered by the field excursion is centred in the Gascoyne Region about 300 km east-southeast of Carnarvon (population 9000*; Fig. 1). A network of unsealed, well-maintained public roads services the area. Station tracks provide year-round access to most parts of the region away from the main roads, although access may be difficult in areas of rough terrain. The only permanent settlements in the region are scattered homesteads on cattle and sheep stations (Fig. 2). Access to Mullewa (population 1100) to the south and Gascoyne Junction (population 35) to the west, is provided by the Carnarvon–Mullewa Road. Meekatharra (population 1100) to the southeast is reached via the Glenburgh – Dalgety Downs – Landor and Landor–Meekatharra roads, whereas Cue (population 500) to the southeast may be reached via Erong Springs Homestead.

The physiography of the excursion area has been briefly described by Williams et al. (1983a) and Williams et al. (1983c). The southern part of the region, which is underlain by the Narryer Terrane, consists of extensive areas of ferruginous duricrust and sandplain, and broad alluvial valleys and low-gradient sheetwash plains. The latter pass into low rocky hills or breakaways of ferruginous duricrust. The highest ridges and hills, such as Mount Narryer, are underlain by resistant quartzite and rise up to 200 m above the surrounding plains (Williams and Myers, 1987). The central part of the region consists of an easterly trending belt of low hills and rugged uplands that form the drainage divide between the Gascoyne River to the north and the Wooramel and Murchison rivers to the south. The northern part of the region, which is underlain by rocks of the Gascoyne Complex and the northern edge of the Narryer Terrane, consists of low rocky hills and uplands and rugged strike ridges of quartz-rich sedimentary rock, separated by wide valleys and pediment slopes.

The region has an arid climate, with hot dry summers (average daily maximum temperature of about 40°C in January) and mild winters (average daily maximum temperature of 22°C in July)†. The mean annual rainfall is about 210 mm. Rainfall in the summer months (November–April) results from northwesterly rain bearing depressions (representing degraded cyclones) and localized thunderstorms. During winter, rain results from the interaction of strong cold fronts from the southwest with tropical cloud bands that originate from the north-northwest. In the southern part of the area, lesser amounts of winter rain are provided by cold fronts alone. All of the watercourses in the region are ephemeral and the Gascoyne and Murchison rivers flow only after heavy rain. However, these two river systems are vast and may carry large volumes of water for several months after heavy rain.

Regional geology

Tectonic units

The excursion covers part of the Narryer Terrane (which forms the northwestern part of the Archaean Yilgarn Craton), and the southern part of the Palaeoproterozoic Capricorn Orogen (which includes the southern part of the Gascoyne Complex; Figs 1–3).

The Narryer Terrane (Myers, 1990c) is one of the largest fragments of Early Archaean (>3300 Ma) crust on Earth, and was reworked by deformation and metamorphism during the late Archaean (Myers, 1990b). The terrane comprises several groups of gneiss derived from early to late Archaean granites‡ and interleaved

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* Data from the Gascoyne Tourism Association website, 2000.
† Climate data from the Commonwealth Bureau of Meteorology website, 2001.
‡ In this guide the term 'granite' is used to refer to any quartz-bearing plutonic rock. Specific granite types are referred to using recommended IUGS terminology (Streckeisen, 1976).

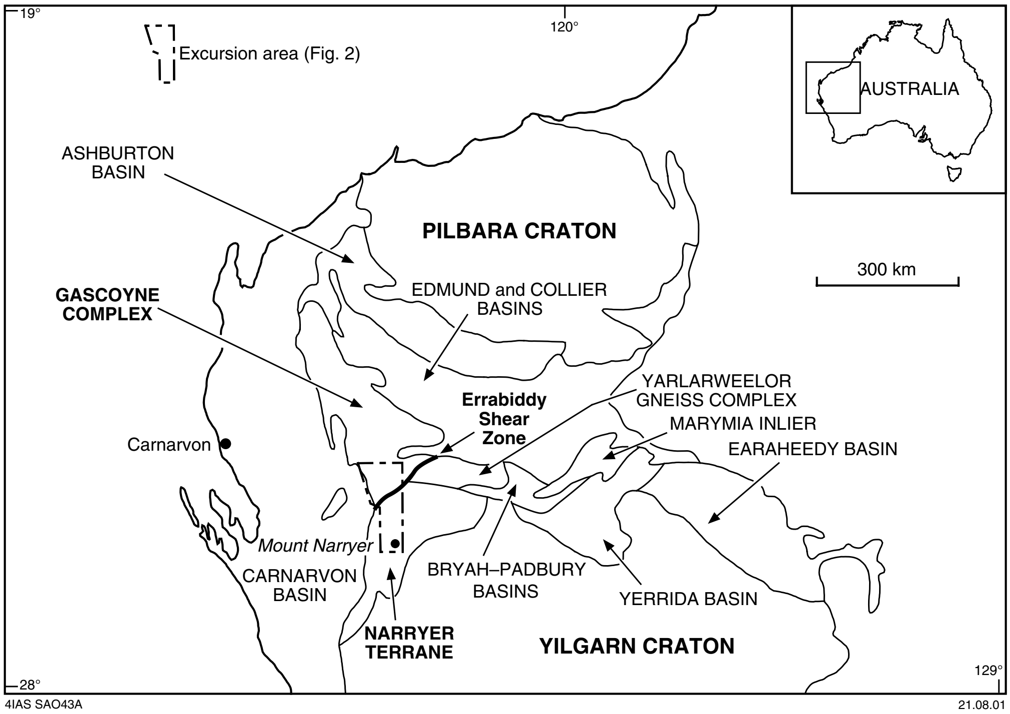


Figure 1. Tectonic units map of northwestern Western Australia. The outlined area broadly marks the area covered by this excursion

metasedimentary and mafic meta-igneous rocks (Williams and Myers, 1987; Nutman et al., 1991). The Capricorn Orogen initially formed in the Palaeoproterozoic during convergence of the Archaean Pilbara and Yilgarn Cratons (Tyler and Thorne, 1990; Krapez, 1999). The orogen also includes some Palaeoproterozoic sedimentary basins farther to the east and northwest, as well as the deformed margins of the Pilbara and Yilgarn Cratons (Tyler and Thorne, 1990; Thorne and Seymour, 1991; Martin et al., 1998; Occhipinti et al., 1998, 1999a).

The Narryer Terrane is separated from the Gascoyne Complex to the north by the Errabiddy Shear Zone (Williams et al., 1983c; Figs 1–3). Within the Errabiddy Shear Zone, granitic gneisses and deformed late Archaean granites of the Narryer Terrane are interleaved with Palaeoproterozoic metasedimentary and mafic and ultramafic meta-igneous rocks, collectively called the Camel Hills Metamorphics. The latter have been subdivided into the Petter Calc-silicate and the Quartpot Pelite (Sheppard and Occhipinti, 2000). Detrital zircon ages from the Petter Calc-silicate suggest that the protoliths were derived from the Yilgarn Craton, whereas the Quartpot Pelite was mainly sourced from Palaeoproterozoic rocks that may have included the Gascoyne Complex (Nelson, 1999, 2000).

The Camel Hills Metamorphics are in faulted contact with the Glenburgh Terrane (Sheppard and Occhipinti, 2000) of the Gascoyne Complex. The Glenburgh Terrane comprises c. 2540–2000 Ma granitic rocks of the Halfway Gneiss, 2005–1970 Ma granitic rocks of the Dalgaringa Supersuite (Sheppard et al., 1999b), and metasedimentary rocks of the Moogie Metamorphics. Although the Glenburgh Terrane includes latest Archaean rocks (c. 2540 Ma), they are younger than any dated rocks from the Yilgarn and Pilbara Cratons (Nutman and Kinny, 1994; Occhipinti and

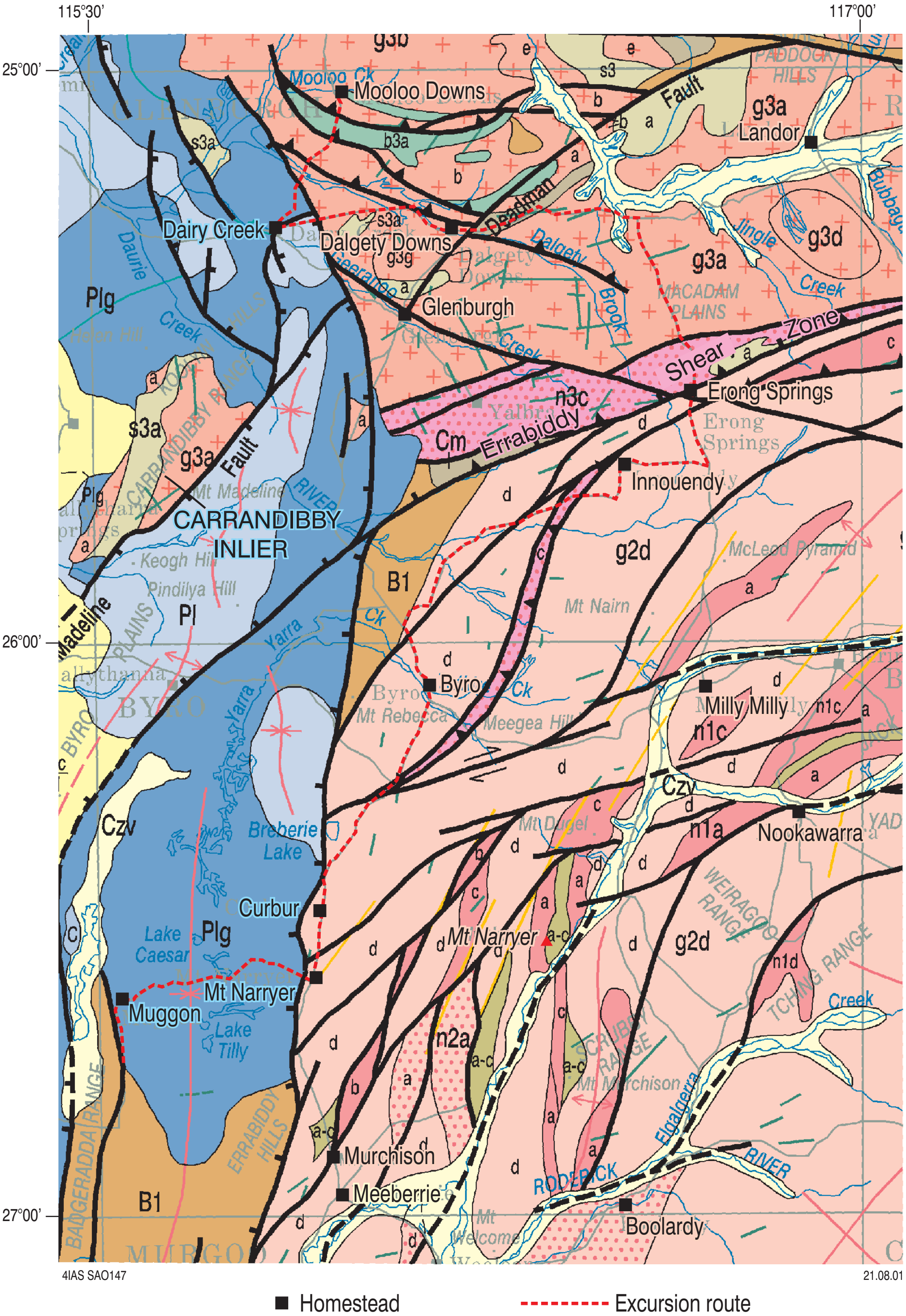


Figure 2. **Portion of the geological map of Western Australia after Myers and Hocking (1998) showing major geological units and their geological relationships, and main access roads used during the excursion**

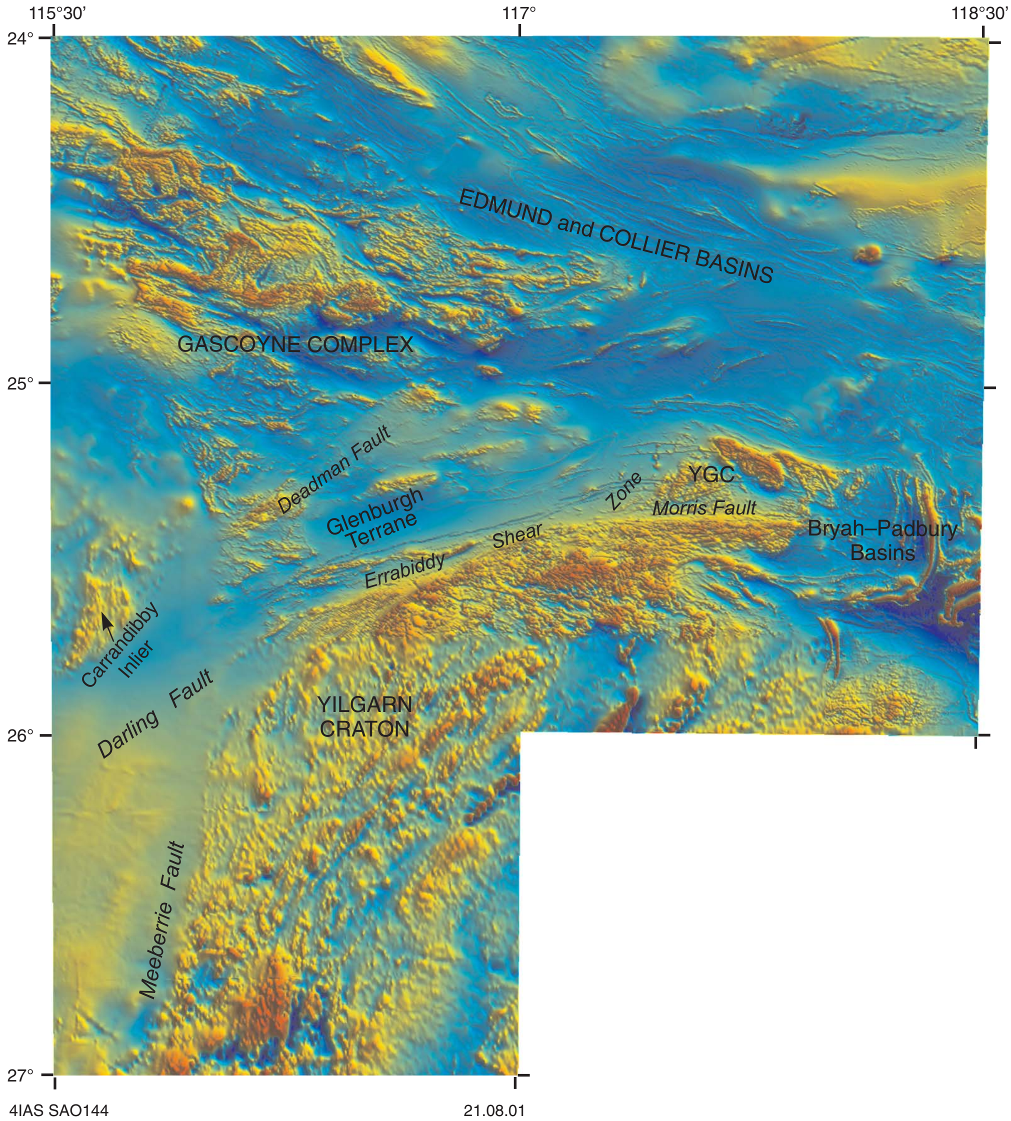


Figure 3. Aeromagnetic image of the northwest Archaean Yilgarn Craton, the Errabiddy Shear Zone, Yarlarweelor Gneiss Complex (YGC), Palaeoproterozoic Bryah and Padbury Basins, and Gascoyne Complex (including the Glenburgh Terrane and Carrandibby Inlier), and the Mesoproterozoic Edmund and Collier Basins

Sheppard, 2001). Moreover, granites of the Dalgaringa Supersuite do not intrude the northwestern margin of the Yilgarn Craton. Therefore, rocks of the Glenburgh Terrane may have formed part of an exotic terrane (Nutman and Kinny, 1994; Sheppard et al., 1999b) incorporated into the Gascoyne Complex during the Palaeoproterozoic. The relationship of the Glenburgh Terrane to the rest of the Gascoyne Complex to the north is unknown.

The Glenburgh Terrane is intruded by large plutons of potassic and silicic granite that belong to the 1830–1780 Ma Moorarie Supersuite. Both the Glenburgh Terrane and the northern edge of the Yilgarn Craton are unconformably overlain by scattered outcrops of the Palaeoproterozoic Coor-de-wandy Formation, the c. 1800 Ma Mount James Formation, and outliers of the Mesoproterozoic Edmund Group (Bangemall Supergroup; Williams et al., 1983c; Drew, 1999a; Martin et al., 1999; Occhipinti and Sheppard, 2001).

Orogenic events

Rocks of the Narryer Terrane were metamorphosed at high grade between 3300 and 3050 Ma, and intruded by granite and pegmatite (Nutman et al., 1991); however, no large-scale structures associated with this metamorphism are preserved. Between 2750 and 2620 Ma, rocks of the Narryer Terrane were multiply deformed and metamorphosed and intruded by late Archaean granite and pegmatite (Myers, 1990b; Nutman et al., 1991).

Two Palaeoproterozoic orogenic events have been identified, largely in rocks of the Gascoyne Complex: the 2000–1960 Ma Glenburgh Orogeny and the 1830–1780 Ma Capricorn Orogeny. During the Glenburgh Orogeny (Occhipinti et al., 1999b), the Glenburgh Terrane, the Camel Hills Metamorphics, and the northwestern edge of the Yilgarn Craton were deformed and metamorphosed at medium to high grade. The Glenburgh Terrane was probably thrust over the Yilgarn Craton from the west or northwest, resulting in the formation of the Errabiddy Shear Zone. The end of the Glenburgh Orogeny was marked by intrusion of 1960–1950 Ma granites into the northwestern margin of the Yilgarn Craton, the Camel Hills Metamorphics, and the southern Glenburgh Terrane. Plutons of the c. 1960 Ma Bertibubba Supersuite (formerly the 'Wooramel suite' of Sheppard et al., 1999a, and 'Bertibubba suite' of Sheppard et al., 2001) intruded into the northwestern margin of the Yilgarn Craton, and 1960–1950 Ma granite dykes intruded into the southern Glenburgh Terrane and Camel Hills Metamorphics.

The Glenburgh Terrane and the Camel Hills Metamorphics were further deformed and metamorphosed at low to medium grade during the Capricorn Orogeny at 1830–1780 Ma, and intruded by granites of the Moorarie Supersuite (Occhipinti et al., 1999a,b; Sheppard and Occhipinti, 2000). Discrete shear zones, which either formed or were reactivated at this time, cut the northwestern margin of the Yilgarn Craton (Sheppard and Occhipinti, 2000). The northeastern part of the Narryer Terrane was deformed, metamorphosed, and intruded by voluminous granite sheets and dykes during the Capricorn Orogeny. This part of the Narryer Terrane is referred to as the Yarlarweelor Gneiss Complex (Occhipinti et al., 1998; Occhipinti and Myers, 1999; Sheppard and Swager, 1999).

Siliciclastic sedimentary rocks of the Coor-de-wandy and Mount James Formations were deposited in a series of small fault-bounded basins on top of the Glenburgh Terrane and Camel Hills Metamorphics, and on the northwestern edge of the Yilgarn Craton. These sedimentary rocks were probably deposited during the latter stages of the Capricorn Orogeny (Occhipinti et al., 1999b). Mesoproterozoic rocks of the Edmund Group were intruded by latest Mesoproterozoic dolerite sills, and then deformed during the Edmundian Orogeny (Halligan and Daniels, 1964) between c. 1020 and c. 750 Ma (Wingate, M. T. D., 1999, written comm.; Sheppard and Occhipinti, 2000; Wingate and Giddings, 2000). During the Edmundian Orogeny, the Bangemall Supergroup and associated dolerite sills formed large-scale dome-and-basin structures. Pre-existing faults and shear zones in basement rocks were probably reactivated at this time. Upper Carboniferous to Lower Permian glacigene rocks of the Carnarvon Basin were deposited on top of all other tectonic units, and locally folded and faulted into northerly trending structures.

Uplift and erosion associated with tectonism of the Narryer Terrane and the southern Gascoyne Complex and substantial vertical movement on the Errabiddy Shear Zone, is indicated by the pattern of Rb–Sr ages of biotite reported by Libby et al. (1999; Fig. 4). They found that Rb–Sr ages on biotite increased from 817–739 Ma in the southern Gascoyne Complex and Yarlarweelor Gneiss Complex, to a minumum age of c. 1544 Ma in the northern part of the Narryer Terrane, closest to the Errabiddy Shear Zone. The Rb–Sr ages mostly progressively increase southwards across the Narryer Terrane. Myers (1990b) suggested that the Rb–Sr age data reflects the extent of substantial southward thrust-stacking of Proterozoic rocks of the southern Capricorn

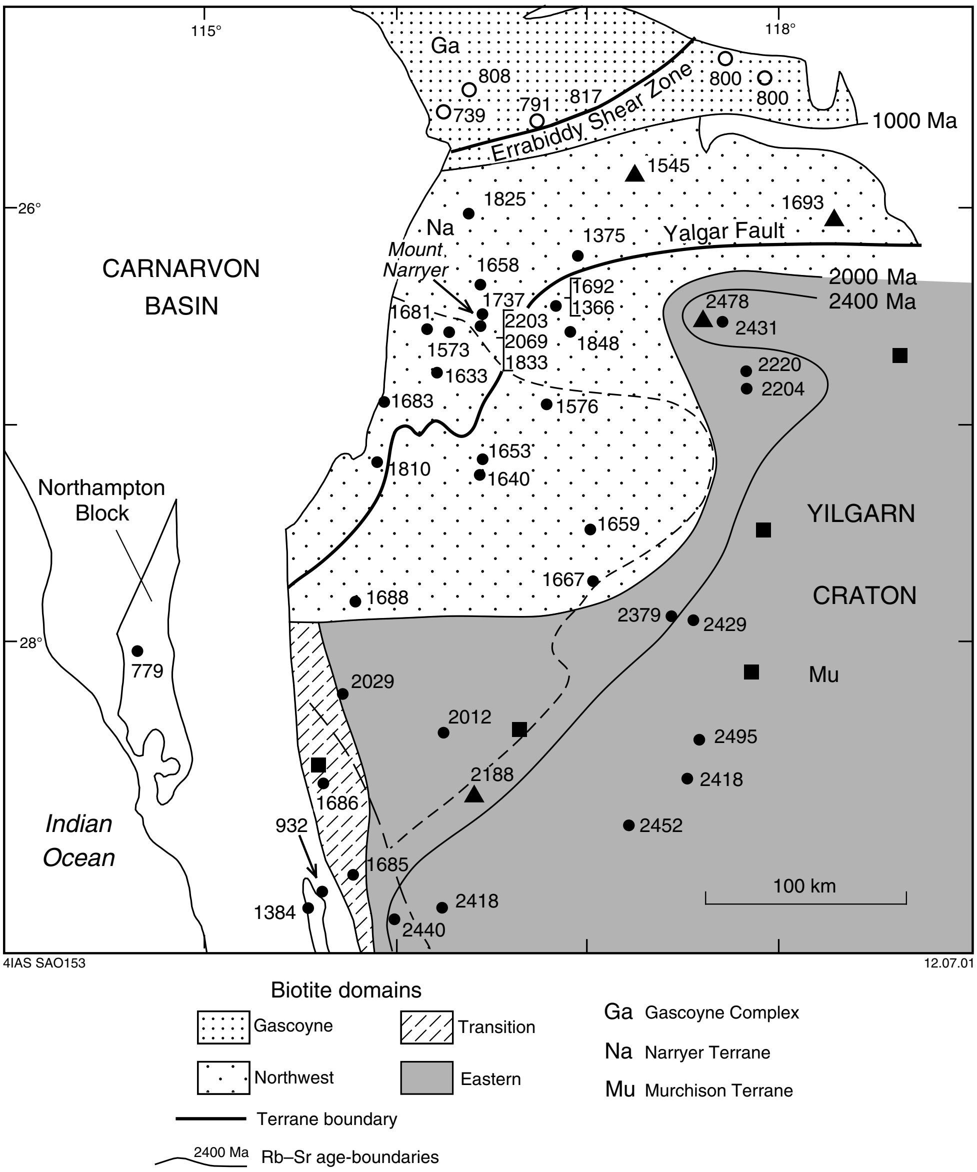


Figure 4. **Rubidium–strontium biotite ages and domains in the Yilgarn Craton and the southern Gascoyne Complex, modified from Libby et al. (1999)**

Orogen over the autochthonous Narryer Terrane. He also suggested the Rb–Sr data supported that c. 1700 Ma reflects a time of cooling, when the Narryer Terrane and its overlying thrust pile was being uplifted and eroded. Myers (1990b) and Libby et al. (1999) suggested that c. 1650 Ma ages in the northwestern part of the Narryer Terrane are related to plutonic and tectonic activity in the 'Gascoyne Province' (herein the Glenburgh Terrane) based on the Rb–Sr ages of plutons within the Gascoyne Complex (Libby et al., 1986). This has not been substantiated from recent regional mapping and U–Pb geochronological work on zircons in the Glenburgh Terrane of the Gascoyne Complex. However, to the northwest and north, within the northeastern part of the Yarlarweelor Gneiss Complex and the north-central part of the Gascoyne Complex, U–Pb zircon ages of c. 1620 Ma (Nelson, 1997) and c. 1670 Ma (Pearson, 1996) have been reported from granite plutons.

Narryer Terrane

by

J. S. Myers and S. A. Occhipinti

The Narryer Terrane forms the northwestern component of the Yilgarn Craton and consists of the Narryer Gneiss Complex (Myers, 1988a; previously called 'early gneiss complex' by Myers and Williams, 1985), and 3000–2700 Ma metamorphosed sedimentary and volcanic rocks. It is of considerable geological significance because it contains siliciclastic metasedimentary rocks with 4400–4000 Ma (Froude et al., 1983; Compston and Pidgeon, 1986; Wilde et al., 2001) detrital zircons, the oldest known remnants of terrestrial rocks. The Narryer Gneiss Complex contains the oldest known rocks in Australia (c. 3730 Ma; Kinny et al., 1988), which are some of the oldest known rocks on Earth. It also contains 4000–4100 Ma xenocrystic zircons within c. 2636 Ma gneissic granite intrusions (Nelson et al., 2000).

Early work

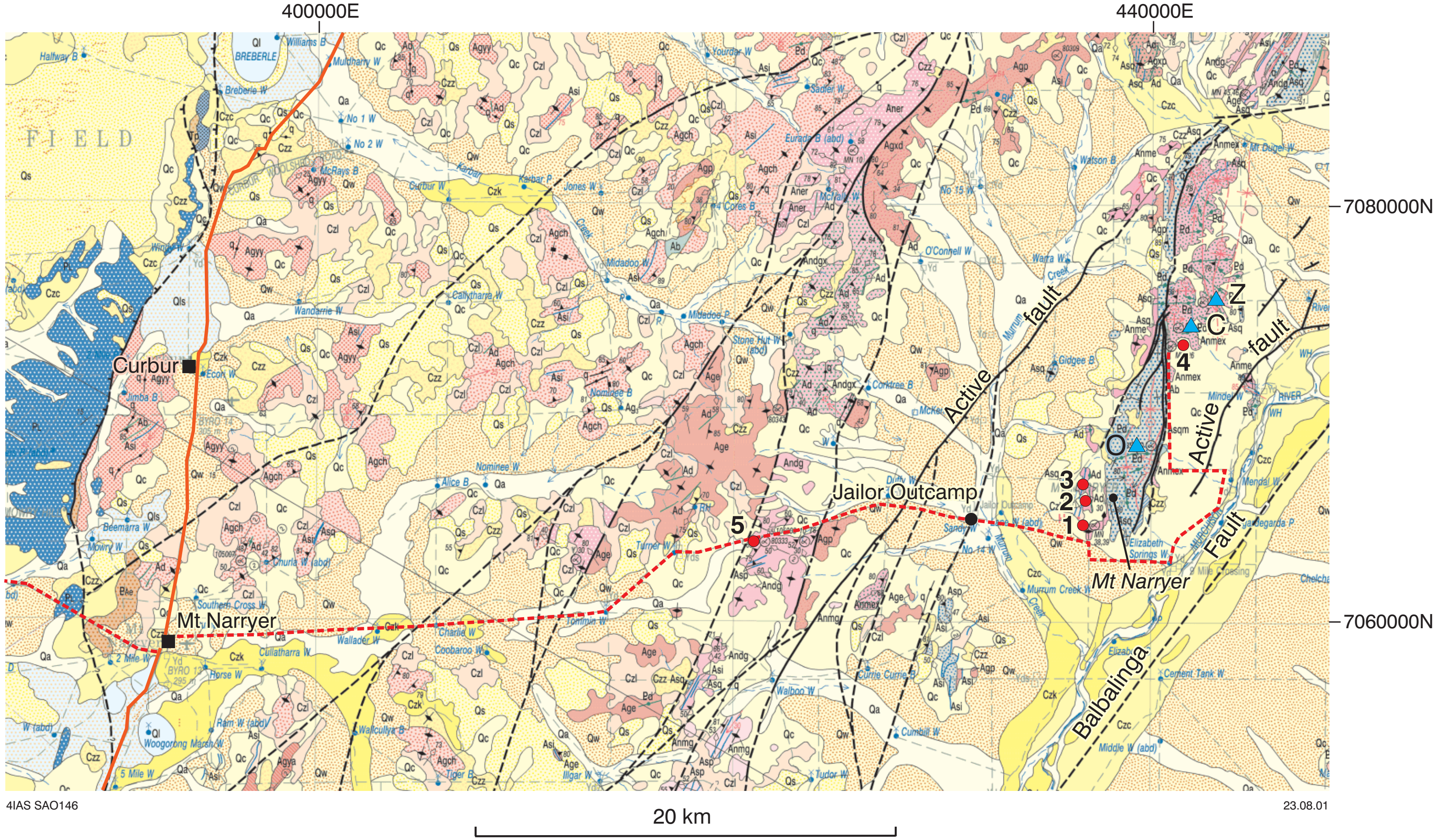
The northwestern part of the Yilgarn Craton and adjacent regions were first systematically mapped in the late 1970s by GSWA (Baxter, 1974; Elias and Williams, 1980; Elias, 1982; Williams et al., 1983a,c). An outcome of this mapping was the identification of gneiss that was older than the widespread late Archaean granites of the Yilgarn Craton.

To further investigate this gneiss, S. J. Williams initiated a Rb–Sr geochronology project in 1978. At a locality northeast of Mount Narryer, the gneiss gave a Rb–Sr whole-rock age of 3348 ± 43 Ma (de Laeter et al., 1981b), which was then the oldest-known age from the Yilgarn Craton. Samples from sites C and Z (de Laeter et al., 1981b) north-northeast of Mount Narryer (Fig. 5) were then chosen for Sm–Nd analysis. They gave Sm–Nd model ages of c. 3630 and c. 3510 Ma respectively, confirming the great antiquity of these rocks (de Laeter et al., 1981a).

Further detailed mapping of the Mount Narryer and adjacent gneisses (at 1:20 000 by I. R. Williams in 1981–82 and J. S. Myers in 1983) also revealed that, in contrast with previous interpretations by Williams et al. (1983c), most of the gneisses were derived from igneous rather than sedimentary precursors (Williams and Myers, 1987). The banding was found to reflect intense deformation and attenuation of different igneous components, rather than sedimentary bedding. The two different units of granitic gneiss, known as the Meeberrie Gneiss and the younger Dugel Gneiss, were found to be temporally and compositionally distinct (Myers and Williams, 1985).

Northeast of Mount Narryer, Myers and Williams (1985) found abundant inclusions of amphibolite, ultramafic rocks, and metamorphosed megacrystic anorthosite and leucogabbro within the Dugel Gneiss. These inclusions were interpreted as parts of a major layered intrusion called the Manfred Complex that had been disrupted by concordant, sheet-like intrusions of granite that now forms the Dugel Gneiss. The Sm–Nd isotope systematics of two samples of the Manfred Complex indicated model ages of 3630 ± 50 Ma (de Laeter et al., 1981a).

The early work summarized above identified some of the main geological units of the Narryer Gneiss Complex, then informally called the 'early gneiss complex' by Myers and Williams (1985), and provided an outline of its tectonic and metamorphic history.



■ Homestead 3 ● Excursion locality ▲ Geochronology sample site Road Track

Figure 5. Portion of Byro 1: 250 000 scale geological map (after Myers, 1997). Localities 1–5 and the route for the Narryer section of the excursion have been highlighted. Note the locations of geochronological sites C, Z, and O

Regional geological setting of the Narryer Terrane

The Yilgarn Craton may have formed at c. 2650 Ma by the amalgamation of several raft-like fragments of continental crust with diverse geological histories (Myers, 1995). These crustal fragments have been referred to as terranes and superterranes (Witt et al., 1998). Both the Yilgarn Craton and its constituent terranes and superterranes are bounded by faults.

The Narryer Terrane forms the northwestern part of the Yilgarn Craton. It is bounded to the southeast by the Balbalinga Fault, along which it is juxtaposed against the Murchison Terrane, another heterogeneous unit of continental crust with a different geological history. The Murchison Terrane consists of c. 3280 Ma granitic basement (Bearra Gneiss; Myers, 1997), overlain by largely c. 3000 Ma basaltic volcanic rocks and sheets of c. 2900 Ma granite intrusions. The northern part of the Narryer Terrane is truncated by the Errabiddy Shear Zone, which forms the southern boundary of the Palaeoproterozoic Capricorn Orogen, whereas to the west the Darling Fault separates the Narryer Terrane from the eastern margin of the Mesoproterozoic and Neoproterozoic Pinjarra Orogen. The latter was last active between 430 and 130 Ma (Playford et al., 1976) as the eastern margin of the Perth Basin, which formed part of a rift zone along which greater India separated from Australia.

The suture zone between the Narryer and Murchison Terranes remains a major zone of crustal weakness and continued tectonic movement. The most recent movements produced extensive fault scarps (Mount Narryer East and Mount Narryer West Faults), probably in 1885 (Williams, 1979), and the largest earthquake ever recorded on the Australian mainland (Meeberrie Earthquake in 1941), with a Richter magnitude of ML 7.2 (Everingham, 1968). During the 1980s, the Murchison Shire Office and adjacent roadhouse (the only settlement in the region besides homesteads) was unfortunately established precisely along the projection of the longest fault scarp, 17 km southwest of the Meeberrie Earthquake's epicentre.

The Narryer Gneiss Complex (Myers, 1988a) is located in the northwestern corner of the Archaean Yilgarn Craton (Fig. 1) and forms a major part of the Narryer Terrane (Table 1). It comprises several heterogeneous units of granite, basic and ultrabasic intrusions, and sedimentary rocks, which have been repeatedly deformed under high-grade metamorphic conditions (Fig. 5). The rocks range in age from 3730 to 3000 Ma, but the main tectonic and metamorphic features formed during two major episodes of continental collision. The first deformation episode resulted from the collision of the Narryer Terrane (Narryer Gneiss Complex and late Archaean granite intrusions) with the Murchison Terrane at 2750–2650 Ma, along the Balbalinga Fault. Other deformation events resulted from Palaeoproterozoic orogenic activity.

Components of the Narryer Terrane

A geochronological study by Kinny (1987) provided the SHRIMP U–Pb zircon ages of 10 samples from around Mount Narryer. These comprise: two samples of the Manfred Complex with ages of 3757–3730 Ma; three samples of Meeberrie Gneiss giving ages of 3678–3673 Ma; a sample of Dugel Gneiss with an age of c. 3380 Ma; a sample of gneiss from Eurada Bore with an age of c. 3483 Ma (subsequently called Eurada Gneiss by Nutman et al., 1991); granites with ages of c. 3300 and c. 2920 Ma, and metasedimentary rocks with detrital zircons ranging from 4200–3100 Ma. Zircon rims with ages of c. 3300 and 2680 Ma were interpreted as the age of high-grade metamorphism. Fletcher et al. (1988) determined similar but less precise Sm–Nd and Pb–Pb ages of 3680 ± 70 and 3689 ± 146 Ma respectively, for samples from the Manfred Complex. In addition, the Sm–Nd analyses indicated substantial disturbance

Table 1. Sequence of main rock units and tectonic events in the Narryer Terrane

Age[a] (Ma)	*Rock unit[a]*	*Deformation[b]*	*Peak metamorphism[b]*	*Zircon history[a]*
Plutonic episode 3				
2650–2600	granite sheets	F_3 upright folds with north–south axes (D_3)	amphibolite facies	igneous zircons
		F_2 upright folds with east–west axes (D_2)	granulite facies	metamorphic rims and euhedral metamorphic zircons in melt patches
2750–2650	granite and gabbro sheets	D_1 widespread intense deformation, tectonic interleaving by thrusting and recumbent isoclinal folding		igneous zircons
3280–2700	metasedimentary rocks (including Narryer and Jack Hills)		detrital zircons	
Plutonic episode 2				
3300	monzogranite, Dugel Gneiss	?regional deformation	amphibolite–granulite facies	metamorphic rims and igneous zircons
3385	syenogranite and monzogranite, Dugel Gneiss			igneous zircons
	metasedimentary rocks (including Mindle meta-sedimentary rocks)			no zircons found
		regional deformation	amphibolite (–granulite) facies	
Plutonic episode 1				
3650–3620	monzogranite, Meeberrie Gneiss			igneous zircons
3730	Manfred Complex, tonalite–granodiorite, Meeberrie Gneiss			igneous zircons
4400–3750				age of detrital zircons in Narryer and Jack Hills metasedimentary rocks

NOTES: (a) Based on references in the text
(b) Based on field evidence

at c. 3320 Ma, and whole rock Rb–Sr data indicated metamorphism at c. 2700 Ma. Further work has refined some of these ages.

Manfred Complex (3730 Ma)

The Manfred Complex (Myers, 1988b) is the oldest-known unit within the Narryer Gneiss Complex. It is a minor, although widespread, gneissic component of the complex and forms as layers, lenses, and trains of fragments in the younger Dugel Gneiss and

Meeberrie Gneiss. The Manfred Complex comprises layered anorthosite, leucogabbro, gabbro, melanogabbro, and ultramafic rocks that are heterogeneously deformed and metamorphosed. Igneous layering and relict coarse-grained igneous textures are well preserved. Large igneous plagioclase, pyroxene, and olivine crystals have also partly survived, mostly enclosed by finer grained aggregates of metamorphic plagioclase, amphibole, and serpentine.

The isotopic history of the Manfred Complex has been determined by Sm–Nd, Pb–Pb, and Rb–Sr studies (Fletcher et al., 1988), and by SHRIMP U–Pb analyses of zircon inclusions within igneous plagioclase megacrysts (Kinny, 1987; Kinny et al., 1988). Rocks of the Manfred Complex can be seen at Locality 4 (Figs 5, 6, and 7) enclosed by Dugel Gneiss.

Fletcher et al. (1988) found that the very high single-stage μ-value and essentially chondritic Nd ε-value in the same rocks indicated that the parent magma was contaminated by isotopically evolved older crustal material. Zircon inclusions recognized by W. G. Libby (GSWA) within plagioclase megacrysts from anorthosite were found to be 3730 ± 6 Ma (Kinny et al., 1988), substantially older than the Meeberrie Gneiss (Kinny et al., 1988).

Meeberrie Gneiss (3730–3620 Ma)

The main component of the Meeberrie Gneiss is a pegmatite-layered monzogranite gneiss composed of quartz, plagioclase, and biotite (Myers and Williams, 1985). This gneiss can be followed into regions of low deformation, where it comprises uniform porphyritic monzogranite containing a heterogeneously developed network of thin pegmatite veins. With increasing deformation, the pegmatite veins became subparallel and attenuated, and the feldspar phenocrysts of the gneiss became augen. These were then streaked-out with the pegmatite veins to form the tectonically induced pegmatite-layering (e.g. Locality 1; Fig. 6). Pegmatite-layered monzogranitic Meeberrie Gneiss can also be seen at Localities 1 and 2 (Fig. 6).

The Meeberrie Gneiss contains fragments of the Manfred Complex and is locally interleaved with a pegmatite-layered tonalitic–granodioritic gneiss from which Nutman et al. (1991) determined a SHRIMP U–Pb zircon age of c. 3730 Ma. In addition, Kinny (1987) and Kinny et al. (1988) dated zircons from widespread localities of the monzogranitic component within the Meeberrie Gneiss, including Localities 1 and 4, which gave igneous ages of 3680–3600 Ma. These ages confirmed the first U–Pb zircon analyses on the Meeberrie Gneiss (Compston, W., 1982, written comm.) which determined that zircon cores gave ages of c. 3600 Ma and their rims gave ages of c. 3300 Ma.

The c. 3300 Ma zircon rims were thought to have grown during granulite facies metamorphism associated with D_2 deformation (Myers and Williams, 1985; Williams and Myers, 1987; Myers, 1988a; Kinny et al., 1990). However, mapping found that the D_2 deformation (see **Structure and metamorphism**) and associated granulite-facies metamorphism affected the widespread late Archaean granites that cut the gneisses. Thus, the zircon rims grew during an earlier high-grade metamorphic event, long before the regional metamorphic and tectonic fabrics were recorded in the main mineral assemblages. Kinny et al. (1990) reported on isotopic studies that included U–Th–Pb on zircon, and K–Ar and $^{40}Ar/^{39}Ar$ on other minerals such as hornblendes and biotites, and showed isotopic evidence of high-grade metamorphism at c. 2700 Ma, interpreted to be coeval with the D_2 deformation.

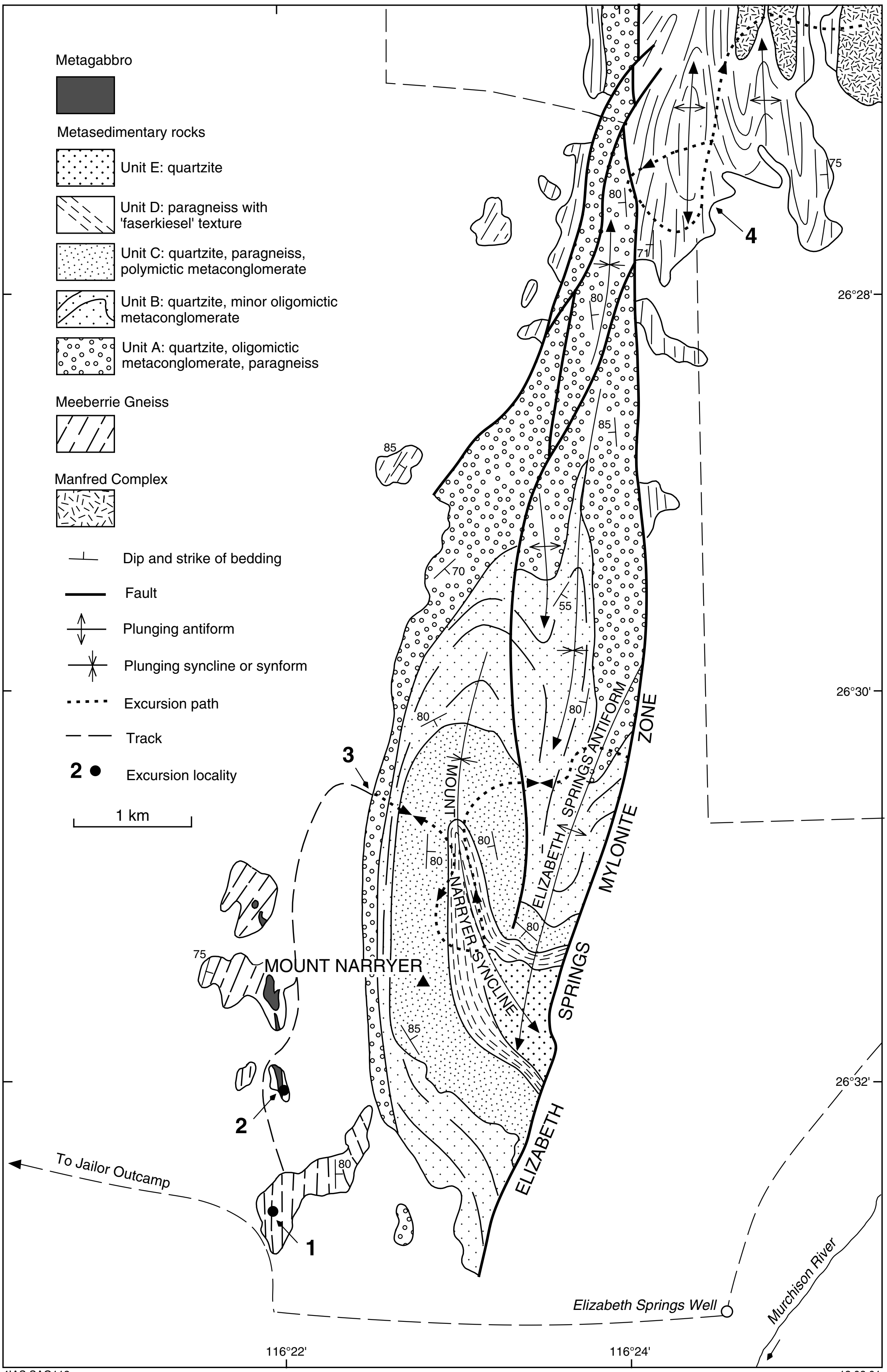


Figure 6. Simplified geological map of the Mount Narryer region, showing Localities 1–4 (after Myers et al., 1990)

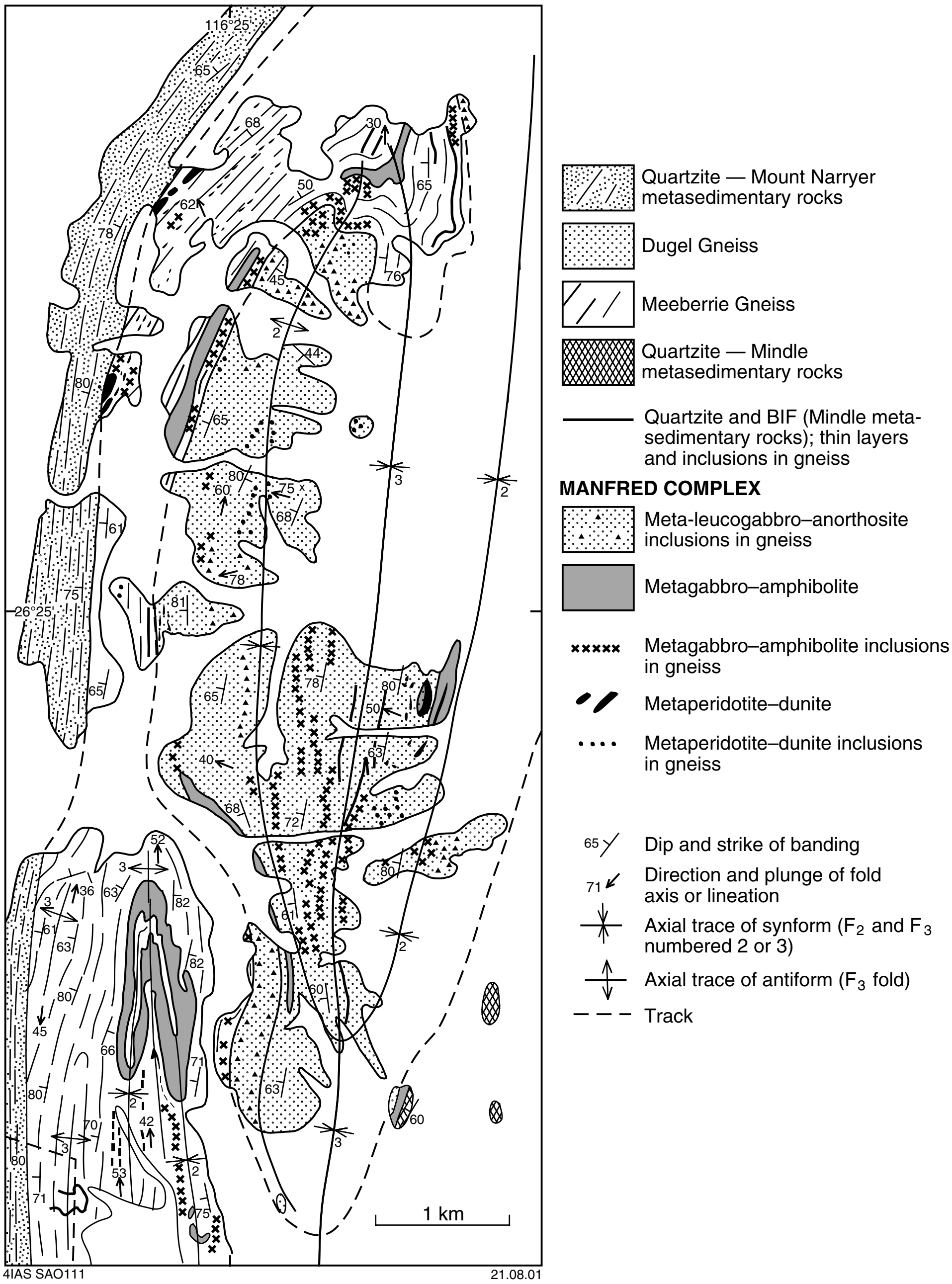


Figure 7. Simplified geological map of the Manfred Complex, Locality 4 (after Myers et al., 1990)

Dugel Gneiss (3385–3300 Ma)

The Meeberrie Gneiss was extensively veined by granite and pegmatite between 3400 and 3300 Ma. These intrusions were deformed and metamorphosed at high grade and comprise a complex of gneisses called Dugel Gneiss (Myers and Williams, 1985; Myers, 1988a). The main component of the Dugel Gneiss (Myers and Williams, 1985) is a leucocratic, pegmatite-layered syenogranitic gneiss, composed of quartz, K-feldspar, and plagioclase, with minor biotite or garnet, or both. The gneiss can be traced into regions of low deformation, where it is seen to have been derived from a medium-grained leucocratic syenogranite with a network of thin pegmatite veins (see **Locality 5**). In some places the Dugel Gneiss is more heterogeneous and contains a number of phases including a fine-grained biotite-rich component veined by leucocratic pegmatite.

Components of the Dugel Gneiss range from monzogranite to syenogranite, and are between 3400 and 3300 Ma (Compston, W., 1982, written comm.; Kinny, 1987; Kinny et al., 1988). SHRIMP U–Pb zircon analyses by W. Compston (1986, written comm.) gave an age of c. 3400 Ma on a little-deformed sample of Dugel Gneiss from west of Mount Narryer (see **Locality 1**; Fig. 5). Kinny et al. (1988) determined an age of 3381 ± 22 Ma for Dugel Gneiss, which is within error of the earlier result. The intrusion of the youngest granitic components of the Dugel Gneiss were contemporaneous with the growth of metamorphic rims on zircons in the Meeberrie Gneiss at c. 3300 Ma (Kinny, 1987; Kinny et al., 1988; Nutman et al., 1991), and may coincide with an episode of high-grade regional metamorphism (Kinny et al., 1990).

On a regional scale, the leucocratic syenogranitic gneiss of the Dugel Gneiss contains abundant dismembered sheets of metagabbro–dolerite and metaperidotite. They post-date, or are contemporaneous with, the Dugel Gneiss, and pre-date the D_2 deformation (see **Structure and metamorphism**). Their concentration in the Dugel Gneiss, rather than being equally distributed in both the Dugel and Meeberrie Gneisses, suggests that they are contemporaneous with the syenogranite protolith of the Dugel Gneiss.

A widespread younger phase of the Dugel Gneiss is derived from a coarse, even-grained or porphyritic monzogranite composed of quartz, K-feldspar, plagioclase, biotite, and a network of pegmatite veins. An example of this rock can be seen at Locality 4 (Fig. 6), where it contains igneous zircons with an age of c. 3302 Ma (Kinny, 1987). The monzogranite phase of the Dugel Gneiss is mostly pegmatite layered gneiss, similar to the main component of the Meeberrie Gneiss.

Eurada Gneiss (3500–3400 Ma)

The Eurada Gneiss is derived from tonalite and monzogranite ranging in age from c. 3500 to 3400 Ma (Nutman et al., 1991). Using the U–Pb SHRIMP technique, Kinny (1987) analysed zircons from a sample of gneiss at Eurada Bore and obtained an age of c. 3483 Ma. The gneiss was subsequently called Eurada Gneiss (Nutman et al., 1991). It is found in fault-bounded slices interleaved with the Meeberrie Gneiss and the Dugel Gneiss and, according to Nutman et al. (1991), does not contain any evidence of metamorphic zircon overgrowths at 3300 Ma or veins of Dugel Gneiss. It is unknown whether the protoliths of the Eurada Gneiss were intruded into the Meeberrie Gneiss, or formed elsewhere and were tectonically juxtaposed against the Meeberrie and Dugel gneisses during the Middle or Late Archaean. Nutman et al. (1991) considered that the Eurada Gneiss was tectonically juxtaposed with the Meeberrie and Dugel Gneisses at 3300–3280 Ma, whereas Nutman et al. (1993) suggested that this event took place much later, at 2750–2720 Ma.

Mount Narryer and Jack Hills metasedimentary rocks (3280–2700 Ma)

Metasedimentary rocks and mafic metavolcanic or intrusive rocks form thin layers within the granitic gneisses of the Narryer Terrane, and layers and trains of inclusions within younger granites. They are lithologically similar to greenstones of the lower grade parts of the Yilgarn Craton that were deposited on sialic crust. The metasedimentary rocks comprise mainly quartzite with minor cordierite, garnet, sillimanite, heavy mineral layers, quartz–magnetite–pyroxene banded iron-formation (BIF), quartz-pebble and polymictic metaconglomerate, and pelitic rocks (commonly biotite-rich schist or granoblastic quartz–cordierite–garnet–biotite–sillimanite rocks). Sedimentary structures, including graded bedding and small-scale cross-bedding, are best preserved at Mount Narryer and Jack Hills, and can be seen at Locality 3 (Figs 6 and 8). Eriksson et al. (1988) briefly discussed the stratigraphy of the metasedimentary rocks at Mount Narryer.

The age of the metasedimentary rocks is poorly constrained. However, it is older than the c. 2700 Ma granite that intrudes the Jack Hills metasedimentary rocks (Wilde and Pidgeon, 1990), and younger than both the 3280 Ma (the youngest detrital zircons at Mount Narryer; Kinny et al., 1990), and 3000–3100 Ma (the youngest detrital zircons from the Jack Hills; Maas and McCulloch, 1991) dates.

Most detrital zircons analysed from these localities have ages of about 3750 and 3500 Ma, but some zircons have ages of about 4400–4100, 3350–3300, and 3100 Ma (Froude et al., 1983; Compston and Pidgeon, 1986; Kober et al., 1989; Kinny et al., 1990; Wilde et al., 2001). The most recent work (SHRIMP U–Pb analysis) on detrital zircons from Jack Hills found the age of part of a zircon to be 4404 ± 8 Ma (Wilde et al., 2001). The authors suggested that a high $\delta^{18}O$ value and micro-inclusions of SiO_2 within the zircon are indicative that the zircon grew from a granitic melt.

From the REE patterns of the metasedimentary rocks at Mount Narryer and Jack Hills, Maas and McCulloch (1991) concluded that the sediments were derived from mature continental crust. Whole-rock geochemical studies of the metasedimentary rocks by Leake (1996) support this conclusion. The interpretation of Mass and McCulloch (1991) is consistent with the range of very old detrital zircons in the same sequence of sedimentary rocks (Froude et al., 1983; Compston and Pidgeon, 1986). Maas and McCulloch (1991) also noted that the metasedimentary rocks and the surrounding granitic gneisses have contrasting REE patterns, and that the detrital zircon ages differ in detail from the ages of the gneisses. Therefore, they suggested that the sediments were not derived from these gneisses but were tectonically juxtaposed. Alternatively, the precursor sedimentary rocks may have been deposited as a cover sequence on the Narryer Terrane, and not derived from the surrounding gneisses.

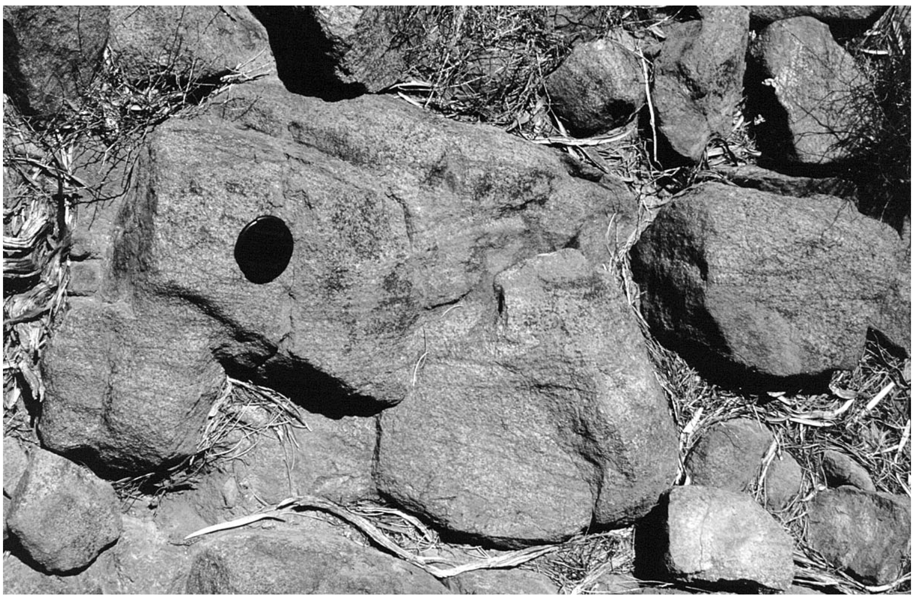


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Figure 8. Cross-bedding in high-grade Mount Narryer metasedimentary rocks at Locality 3b

Contacts between the metasedimentary rocks and gneisses are intensely deformed and in some places, such as the eastern side of Mount Narryer (Figs 2, 5, and 6), are zones of tectonic transport. The relative ages of most metasedimentary rocks and the adjacent Meeberrie and Dugel Gneisses are mostly unclear from field evidence. There are at least two groups of metasedimentary rocks: the Mindle metasedimentary rocks that predate the 3385 Ma Dugel Gneiss, and the Mount Narryer and Jack Hills metasedimentary rocks that contain detrital zircons as young as 3280 and 3100 Ma, respectively. The Jack Hills metasedimentary rocks may also comprise at least two components of different age. However, as these rocks form typically small, widely scattered outcrops, their relative age is unknown (Fig. 2).

Metamorphosed granite and gabbro sheets (2750–2600 Ma)

Sheets of granite and gabbro were intruded throughout the Narryer Gneiss Complex during and before the juxtaposition of the Narryer and Murchison Terranes (Figs 5–7). The intrusive rocks have diverse textures and grain sizes that range from even grained to porphyritic, and from coarse to fine. Their intrusion spanned a major episode of deformation, and so the rocks range from concordant layers of intensely deformed, pegmatite-layered gneiss (similar to the Meeberrie Gneiss and the Dugel Gneiss) to discordant sheets with little-deformed igneous textures. The older intrusions were metamorphosed under amphibolite- or granulite-facies conditions, whereas the younger intrusions were only metamorphosed in the greenschist facies. Contemporaneous rocks also increase in metamorphic grade northwards, reflecting the greater subsequent uplift and exposure of deeper crustal levels that resulted from a combination of Late Archaean and Proterozoic tectonism.

Diverse samples of these metamorphosed granites from a wide region of the Narryer Terrane and adjacent part of the Murchison Terrane contain zircons with igneous ages between 2735 and 2610 Ma (Nelson, 1996; Myers, 1997). Examples of the metamorphosed granites can be seen at Locality 1 (Fig. 5).

Sheets of metagabbro are less abundant than those of metamorphosed granite, and no geochronological data are available. Most are thin and consist of massive metagabbro, such as at Locality 2 (Fig. 5), but some are well preserved, thicker and layered, and others are schistose amphibolite.

Structure and metamorphism

Deformation and metamorphism of the Narryer Gneiss Complex took place mainly during the late Archaean, and was associated with the juxtaposition of the Narryer and Murchison Terranes between 2750–2650 Ma. The most prominent deformation events, D_2 and D_3 (Table 1), are common to both terranes (Myers and Watkins, 1985; Myers and Williams, 1985).

The first major episode of late Archaean deformation, D_1, involved the development of mainly flat-lying structures, but began at different times in the Narryer and Murchison Terranes. The Narryer Terrane contains a variety of older tectonic fabrics and structures, but no large-scale folds associated with these have been recognized. The D_2 event formed folds with subvertical, easterly trending axial surfaces and subhorizontal fold axes. The D_3 event resulted in the formation of folds with subvertical, northerly trending axial surfaces, and led to fold-interference structures with F_2 folds on scales of a few centimetres to tens of kilometres. Deformation was especially strong in the steep northerly trending limbs of F_3 folds. Many of the fold limbs acted as ductile shear zones in which previous layering was attenuated and rotated, and in places was transposed into a new tectonic fabric.

Metamorphic grade associated with both D_2 and D_3 increases northwestwards from greenschist facies in most of the Murchison Terrane, to amphibolite facies in the vicinity of the Balbalinga Fault Zone, and granulite facies in much of the Narryer Terrane. Within the Narryer Terrane, granulite-facies mineral assemblages formed during and after D_2, and were extensively retrogressed to amphibolite-facies assemblages during D_3.

Examples of F_2 folds with an axial-planar flaser foliation and syn- and post-D_2 metamorphic mineral assemblages are present at Locality 5 (Fig. 9) in an area of relatively low D_3 deformation. Examples of D_2–D_3 fold-interference structures can be seen at Locality 4 (F_2–F_3; Fig. 7), as well as a range of D_3 structures in both the cores and limbs of major and minor folds. Although the Narryer Syncline (see **Locality 3**; Fig. 6) is the first fold in the metasedimentary rocks, it is a D_2 structure refolded by the F_3 Elizabeth Springs Antiform (Fig. 6). These folds are truncated by the late-D_3 Elizabeth Springs Mylonite Zone (Fig. 6).

The major Balbalinga and Eurada Fault Zones (Fig. 5) formed during D_1, and were regions of intense deformation during D_2 and D_3 (Table 1). Granite sheets were emplaced into both zones and were intensely deformed at about 2700–2650 Ma. The Balbalinga Fault Zone is the main boundary between the Narryer and Murchison Terranes, and probably represents the suture zone between what were two plates of sialic crust. The relative original orientation of the suture may have been easterly trending, with the Murchison Terrane dipping northwards under the Narryer Terrane. In the Narryer Terrane, the first major phase of late Archaean deformation, D_1, caused tectonic interleaving of the gneissic granitic basement with a cover of sedimentary rocks (Narryer and Jack Hills metasedimentary rocks; Table 1). Deformation may have been

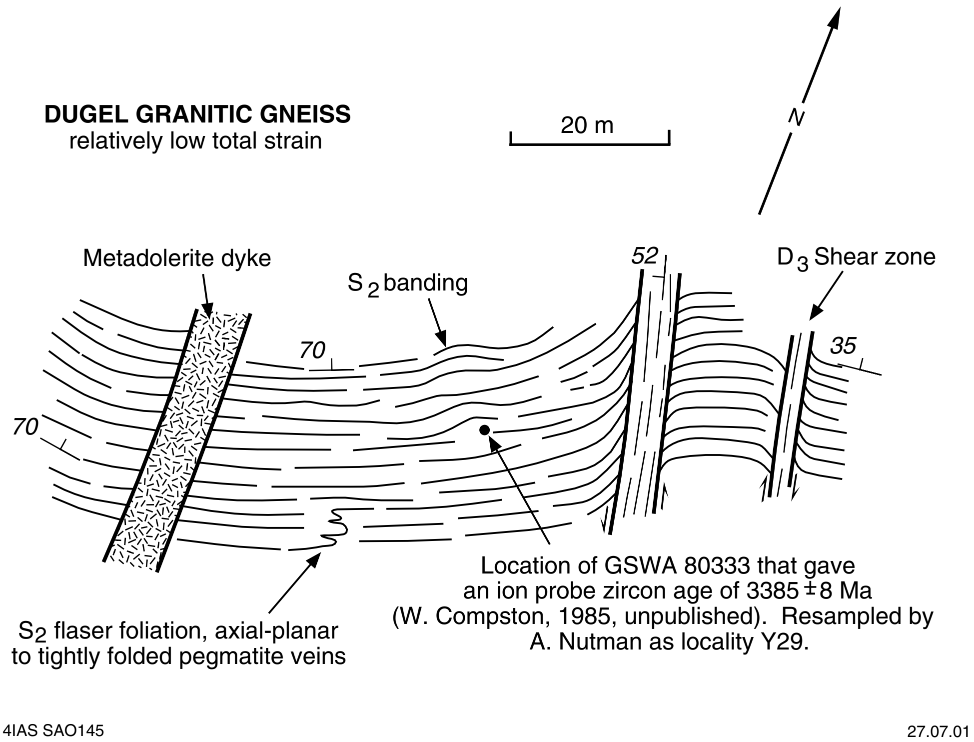

Figure 9. Sketch map of the geology around the site of geochronology sample GSWA 80333 at Locality 5

associated with thrusting that stacked and thickened the Narryer Terrane on top of the edge of the Murchison Terrane to the south.

Myers (1990b) suggested that the Narryer Terrane was affected by another major episode of continental collision at c. 1800 Ma, when the Pilbara and Yilgarn Cratons collided, forming the Capricorn Orogen. However, the nature of this collision is enigmatic and it is now understood that an earlier continental collision took place at 2000–1960 Ma (Occhipinti et al., 1999b; see **Development of the Errabiddy Shear Zone**) along the northern margin of the Narryer Terrane. As a consequence of these tectonic movements, deeper levels of the Narryer Terrane are exposed towards the northwest. The Narryer Gneiss Complex appears to be a tabular package of tectonically stacked and tightly folded units of lower to middle Archaean gneiss overlying late Archaean granite and granite–greenstone terrain.

Excursion localities — Narryer Terrane

Day 1

Travel from Perth to Mount Narryer.

Day 2

Today we will look at the Meeberrie Gneiss of the Narryer Gneiss Complex and younger metagranite and metagabbro that intrudes it. In addition, we will walk through parts of the Mount Narryer metasedimentary rocks, where detrital zircons between 4200 and 4100 Ma have been dated.

The following descriptions for Localities 1 to 3 are largely from Myers et al. (1990).

Locality 1: Meeberrie Gneiss southwest of Mount Narryer (MGA 437388E 7064016N)

From Mount Narryer Homestead take the track that leads to the woolshed and then runs between it and the shearers' quarters. Take the track to the northeast, right of the airstrip, through a gate in an easterly trending fenceline. Follow this well-used track east for 22.8 km to Tommin Well, then northeast for 4.4 km to Turner Well, before continuing east-northeast for 10.6 km to Duffy Well. Between Turner Well and Duffy Well, prominent small hills of granitic Dugel Gneiss outcrop on either side of the track. From Duffy Well follow the track east for 3.9 km to a Y-junction. Take the left fork for 500 m to Jailor Outcamp. From Jailor Outcamp take the easterly trending track south of the yard for 1 km to a fence junction. From the junction travel towards the northeast for 100 m then turn right. Follow this track east through mulga woodland for 800 m to a fence corner, then follow the track east along the fence for 4.3 km. Cross the fenceline through a gate (at MGA 437191E 7063620N). Turn left after the gate, following the track through open country towards Mount Narryer.

About 500 m from the gate there is good outcrop of granite and granitic gneiss, particularly on the western side of the track. This is Locality 1.

At Locality 1 (Fig. 6), pegmatite-layered Meeberrie Gneiss is intruded by sheets of late Archaean granite. The Meeberrie Gneiss consists of alternating layers of biotite monzogranite and pegmatite that have been strongly deformed and recrystallized in amphibolite facies. The gneissic fabric is northerly trending and dips vertically.

Although apparently simple in composition and structure on the scale of the outcrop, the gneiss is heterogeneous in detail. The gneiss ranges from portions that were originally homogeneous with scattered K-feldspar phenocrysts, and with either few or many pegmatite veins, to portions that contain remnants of primary nebulitic compositional layering and wispy biotite-rich schlieren. The layering is isoclinally folded and, with increasing intensity of deformation, the rocks in the fold limbs became finer grained and the layering thinner. These structures were refolded by small asymmetric folds and shear zones that indicate mainly dextral movements. The axial planes and shear zones are subvertical and strike north–south.

The Meeberrie Gneiss was intruded by sheets of fine-grained granite that are now deformed and metamorphosed, and are mainly subparallel to the layering of the gneiss. Crosscutting relations can be seen in places, as well as a few angular xenoliths of gneiss in the granite. The granite sheets contain pegmatite veins and nebulitic igneous layering. They were strongly deformed by the D_3 deformation; some pegmatite veins are ptygmatically folded and the granite has an S_3 foliation.

Isotopic studies on whole rocks from the vicinity were reported by de Laeter et al. (1985). The Meeberrie Gneiss gave a Sm–Nd (T_{CHUR}) model age of 3620 ± 40 Ma, and a 12-point Rb–Sr isochron of 3302 ± 65 Ma with an initial ratio of 0.7004 ± 0.0023 (Model 3, MSWD 23.5). Granite sheets north of Locality 2 gave a Sm–Nd (T_{CHUR}) model age of 3120 ± 30 Ma and a 6-point Rb–Sr isochron of 2579 ± 122 Ma with an initial ratio of 0.7143 ± 0.0092.

Kinny and Nutman (1996) suggested that the Meeberrie Gneiss is best described as a complex migmatite with resolved ages that range from 3730–3300 Ma, which records high-grade metamorphic events with in situ anatexis and granite intrusion during the Early Archaean. They reported two groups of zircon ages: one at 3650–3600 Ma and one at c. 3300 Ma from site MN39 (2 km south of Mount Narryer; Kinny and Nutman, 1996), which is just east of Locality 1. They interpreted the older age as an igneous crystallization age of precursor granite to the gneiss, and the younger c. 3300 Ma age as either an igneous crystallization age of a younger granite component to the gneiss or the age of in situ partial melt. Rims on the cores of both populations of zircons gave ages of c. 3300 Ma, which would support the presence of in situ melt, or a combination of in situ melt with 3300 Ma granite veins into 3650–3600 Ma granite or gneiss (Kinny and Nutman, 1996).

Zircon analyses from homogenous fine-grained granite at Locality 1 (site MN38) were reported by Myers et al. (1990). They found that 30% of the analysed grains had distinct cores, which had uniform $^{207}Pb/^{206}Pb$ ratios corresponding to an age of 2919 ± 12 Ma and apparently igneous rims that gave a $^{207}Pb/^{206}Pb$ age of 2638 ± 11 Ma. The latter age was interpreted as the igneous crystallization age of the granite.

Locality 2: Metagabbro intrusion in Meeberrie Gneiss southwest of Mount Narryer (MGA 437178E 7065190N)

From Locality 1 travel in a northerly direction for 1.2 km to a hill of metagabbro and granitic gneiss on the eastern side of the track.

At Locality 2 (Figs 5 and 6), pegmatite-layered Meeberrie Gneiss, similar to that at Locality 1, contains a younger sheet of metagabbro. Both the metagabbro and the gneiss have been deformed and boudinaged during D_3. The intensity of deformation in the gneiss increases towards the contact with the more competent metagabbro. The metagabbro has a thin, fine-grained schistose margin, but internally shows little deformation and its igneous texture is well preserved. In thin section large igneous grains of plagioclase and hypersthene are enclosed by metamorphic hypersthene, clinopyroxene, and brown hornblende. The metagabbro is probably late Archaean, but an absolute age has not been determined.

Locality 3: Mount Narryer metasedimentary rocks (start at MGA 438648E 7067581N)

From Locality 2 travel north for about 600 m to a track junction. Take the left-hand northeasterly trending track . Continue to the north for 500 m to another junction, and again take the left track. After another 500 m bear left through scattered low outcrops. Follow this track for 1.5 km into a creek system, then follow the main creek to the east-northeast, towards the prominent ridge. Park about 400 m from where you entered the creek system, and walk up the creek to the east to Locality 3.

At Locality 3 (Figs 5 and 6), a complete section is preserved through the Mount Narryer metasedimentary rocks. For the 1990 Mount Narryer excursion (Myers et al., 1990), Locality 3 included a traverse following a creek southeast across the western limb of the Mount Narryer Syncline. From the core of the syncline the route continued northeast through a low pass and downwards to the core of the Elizabeth Springs Antiform. It then went east down a creek to the Elizabeth Springs Mylonite zone, which forms the tectonic eastern contact of the metasedimentary rocks with Meeberrie Gneiss. The total return distance is about 10 km.

For this excursion, Locality 3 has been shortened and divided into three parts (3a, 3b, and 3c), which contain the main elements and the best outcrops of Locality 3 from the 1990 excursion guide. However, for completeness the following descriptions include all that was covered in the 1990 Mount Narryer excursion guide.

The metasedimentary rocks strike north–south and dip steeply to the east or west. The Mount Narryer Syncline is a D_2 structure (although it is the first deformation structure within the metasedimentary rocks) that is refolded by the F_3 Elizabeth Springs Antiform (Fig. 6). Both structures plunge steeply southwards and the associated cleavage is subvertical. The metasedimentary rocks are relatively little deformed and sedimentary features are exceptionally well preserved. The rocks were recrystallized by prograde amphibolite- to granulite-facies metamorphism, with local retrogression to amphibolite and greenschist facies. They are divided into five lithostratigraphic units designated A to E (from oldest to youngest), with a minimum total thickness of 2250 m. The stratigraphy is described in detail by Williams and Myers (1987, p. 9–13, and table 4 on p. 25–26).

Locality 3a: Pelite and metaconglomerate of the Mount Narryer metasedimentary rocks (MGA 438648E 7067581N)

The western contact between the metasedimentary sequence and the Meeberrie Gneiss is not exposed at Locality 3. Units A and B, encountered in the first part of the traverse, consist mainly of quartzite and quartz-rich gneiss. About 500 m east of the first outcrops, unit C is well exposed where the creek swings north-northeast along a major dolerite dyke of late Archaean or Palaeoproterozoic age. The dyke cuts the main ridge formed by unit C, which comprises a sequence of cordierite–garnet–sillimanite quartzite, polymictic metaconglomerate, and grey-green cordierite–biotite–garnet–sillimanite gneiss. Some of the rocks contain cross-bedding and graded bedding. Between sites H and I (Williams and Myers, 1990), sedimentary structures in polymictic metaconglomerate and the cleavage–bedding relations of S_2 are well exposed on a smooth rock surface in the creek bed. A detailed measured section (209 m) for part of unit C exposed in this creek is presented in Williams and Myers (1987).

Locality 3b: Mount Narryer metasedimentary rocks (MGA 438424E 7067243N)

The creek swings to the southeast and becomes broad and open in unit D. Here the traverse leaves the creek and heads south up the ridge formed by unit C, towards the summit of Mount Narryer. On the east flank of the ridge crest, north of Mount Narryer summit, small-scale cross-bedding (Fig. 8) can be seen in garnetiferous quartzite, and there is also a good view of the Mount Narryer Syncline.

In the 1990 Mount Narryer excursion, the route continued southeast from Locality 3b across unit D, comprising quartz–sillimanite gneiss with quartz–sillimanite (fibrolite) clots (faserkiesal texture), to the massive quartzite of unit E in the core of the Mount Narryer Syncline. A number of north-trending amphibolite dykes also outcrop in the vicinity.

Locality 3c: Quartzite of the Mount Narryer metasedimentary rocks (MGA 439899E 7068287N)

From Locality 3b, walk down the hill back towards Locality 3a and then across the creek into a northerly trending valley. At the head of the valley is an easterly trending dolerite dyke. Follow this east towards Locality 3c (around MGA 439930E 7068383N) over a low ridge into an easterly trending valley. The dyke cuts through unit C. To the east, Locality 3c is located in quartzite at the head of the next easterly trending creek system to the north. Site O of Williams and Myers (1990) is situated close to a spectacular vertical face of pelite with folded pegmatite veins (located at MGA 439899E 7068287N).

Locality 3c (Site O) is located near the base of unit B (cordierite–garnet–sillimanite gneiss), and consists of grey quartzite. Detrital zircons from this quartzite were dated using the U–Pb SHRIMP technique as greater than 4000 Ma (4200–4100 Ma; GSWA 71932; Froude et al., 1983).

Unit A comprises oligomictic quartz-pebble metaconglomerate and quartzite in which graded bedding is developed in places. The intensity of deformation and retrograde metamorphism increases rapidly to the east, and bedding is obliterated and replaced by a pronounced tectonic foliation as the Elizabeth Springs Mylonite Zone is approached. The asymmetry of small folds in the quartzite mylonites indicates dextral movements. The contact with the gneiss to the east is not exposed.

Day 3

The granitic Meeberrie and Dugel Gneisses of the Narryer Gneiss Complex and the metamorphosed gabbro and anorthosite of the Manfred Complex will be our first stop today. Time permitting, on the way out of this first part of the excursion we will stop to look at the Dugel Gneiss, west of Mount Narryer (Locality 5). Then we will travel north to the Errabiddy Shear Zone, which marks the northwestern boundary of the Archaean Yilgarn Craton. At Locality 6 we will look at a granitic gneiss, which is a part of the Narryer Terrane that was reworked in the Errabiddy Shear Zone during the Palaeoproterozoic.

The following descriptions for Localities 4 and 5 are largely from Myers et al. (1990).

Locality 4: Meeberrie Gneiss, Dugel Gneiss, and Manfred Complex northeast of Mount Narryer

From the Elizabeth Springs Mylonite Zone, return to the starting point, retracing part of the route southwest across the Elizabeth Springs Antiform and eastern limb of the Mount Narryer Syncline, then down the main creek towards the northwest. Return past Locality 1 to the gate in the fenceline, but don't go through it. Turn left and follow the track to the southeast for 900 m to a fenceline. Follow the fence to Elizabeth Springs Well (4.2 km from gate). At Elizabeth Springs Well take the northeasterly track on the western side of the tank. Follow this for 5.3 km to Mendell Well, then take a track on *the west side of the well and head north for 500 m to a gate. Go through the gate and turn left. Travel west along the fence for 2.9 km then north for 3.9 km to a fence junction. Continue north through the narrow gate for 3 km to Locality 4.*

At this locality metal stakes mark the locations of samples collected for whole-rock Rb–Sr geochronology by I. R. Williams in 1980. The rocks gave a Rb–Sr whole-rock isochron of 3348 ± 43 Ma (de Laeter et al., 1981b), which was then the oldest known age from the Yilgarn Craton. Determination of the Sm–Nd (T_{CHUR}) model age of a whole-rock sample (60735) from site C as 3630 ± 40 Ma (de Laeter et al., 1981a) gave the first indication that these rocks were of early Archaean age. Myers et al. (1990) reported SHRIMP U–Pb analyses of zircons from sites A and C. At site C they found that the main population of zircons from the Meeberrie Gneiss gave a SHRIMP U–Pb age of c. 3620 Ma with a smaller population giving an age of c. 3740 Ma. At site A they reported a single population of zircons with an age of 3302 ± 6 Ma, which is consistent with the rock being Dugel Gneiss.

The Meeberrie Gneiss at this locality is a pegmatite-layered monzogranite gneiss in the core of a major fold-interference structure produced by the superposition of F_2 and F_3 antiforms. The gneiss resembles that at Localities 1 and 2 and is superficially simple; however, in detail it is heterogeneous in composition and structure. Besides gneissic layering, the most prominent structures are small tight folds with upright north–south axial surfaces. They fold a flaser foliation associated with D_2 deformation, and some small-scale fold-interference structures may be seen. The limbs of many F_3 folds are attenuated, and acted as ductile shear zones in which an S_3 foliation was intensely developed and older structures were transposed into the north–south D_3 tectonic fabric. The same phenomenon can be seen on a larger scale by thinning of individual layers and streaking out of surviving augen of K-feldspar.

The contrast between massive, little-deformed porphyritic granite and previously gneissose granite was intensified by D_3, which enhanced the tectonic fabric in the previously strongly deformed rocks relative to the more competent, previously little-deformed rocks. Sheets of pale-green, metamorphosed, fine-grained late Archaean granite, like those at Locality 1, crosscut the older gneisses and were deformed along with the gneisses by the D_3 deformation.

Much of the gneiss appears to be Meeberrie Gneiss, which is locally little deformed and can be seen to have been derived from a porphyritic monzogranite. Some of the least deformed porphyritic monzogranite contains xenoliths of diffusely layered gneiss. Zircons from similar rocks northwest of Jack Hills have igneous ages of about 3650 Ma. However, zircons from a little-deformed porphyritic granite near the fenceline track at Locality 4 have an igneous age of 3300 Ma (Kinny et al., 1990). Monzogranite and monzogranite gneiss of this age, which closely resemble the main unit of the Meeberrie Gneiss, are a widespread component of the Narryer Gneiss Complex. However, as at Locality 4, it is commonly difficult to distinguish the 3650 and 3300 Ma monzogranite gneisses.

The emplacement of monzogranites at 3300 Ma coincided with amphibolite- to granulite-facies metamorphism, during which there was widespread metamorphic overgrowth on older igneous zircons, and new euhedral zircons grew in local melt patches.

Layers of BIF can be seen by walking north near the fence and track. The BIF is part of a major east–west F_2 fold (Fig. 7). The contact relations between the BIF and adjacent gneisses, and the effects of various amounts of deformation on the BIF show that with increasing intensity of deformation, the individual layers thinned and the grain size decreased.

From the first geochronology sample sites marked by metal stakes, where the fence and track turn from north–south to northwest–southeast, more good outcrops of Meeberrie Gneiss can be seen by walking north-northeast. The pegmatite-layered gneiss contains abundant small-scale, tight north–south folds superimposed on east–west D_2 folds. The valley to the north-northeast contains scattered outcrops and rubble of Meeberrie Gneiss, as well as ultramafic rocks.

A large scale F_2–F_3 fold-interference structure is well exposed on the ridges to the north-northeast. It is outlined by a sheet of amphibolite that is part of the 3730 Ma Manfred Complex (Fig. 7). The amphibolite is within pegmatite-layered gneiss comprising the Meeberrie Gneiss and the Dugel Gneiss, interlayered on all scales. The amphibolite locally passes into lenses of less-deformed metagabbro (from which it appears to be derived) and contains layers of metapyroxenite, metaperidotite, and serpentinite. The metagabbro contains relict igneous texture and remnants of igneous feldspars, orthopyroxene, and olivine. The metaperidotite has remnants of igneous orthopyroxene, clinopyroxene, and olivine.

Most mineral assemblages reflect retrograde amphibolite-facies metamorphism during and after D_3 deformation. Remnants of syn-D_2 metamorphism can be seen to the north-northeast in gneisses within the core of the major F_3 fold. The gneiss is mainly pegmatite-layered leucocratic Dugel Gneiss. It contains isoclinal F_2 folds of an older gneissose layering, with an S_2 axial-planar flaser fabric overgrown by garnets that are commonly obliterated by D_3 deformation and associated metamorphism.

Across a broad valley to the east, metamorphosed leucogabbro and anorthosite of the 3730 Ma Manfred Complex is found as rubble and fragments in the Dugel Gneiss. Relict igneous textures are visible in both the leucogabbro and anorthosite as well as some igneous layering and remnants of igneous plagioclase megacrysts, but most of the rocks are now recrystallized to zoisite and clinozoisite. The Dugel Gneiss and Manfred Complex are cut by undeformed dykes of both metadolerite and aplite that may be late Archaean or Palaeoproterozoic in age.

Ion-microprobe zircon analyses of anorthosite and leucogabbro of the Manfred Complex and a sample from the Dugel Gneiss, which gave an age of c. 3375 Ma from this locality, are described by Kinny et al. (1988), and summarized by Myers et al. (1990). Whole-rock samples of the Manfred Complex, ranging from anorthosite to ultramafic rock, with relict igneous textures and minerals have given a Sm–Nd isochron of 3680 ± 70 Ma with ε_{Nd} of –0.2 ± 0.7 (Fletcher et al., 1988).

Locality 5: Dugel Gneiss west of Mount Narryer (MGA 421500E 7063800N; time permitting)

Return to Mount Narryer Homestead following the routes described for Day 2. Locality 5 is on the south side of the easterly trending track between Jailor Outcamp and Turner Well.

The main leucocratic phase of the Dugel Gneiss can be seen at this locality in a zone where D_3 deformation is minimal (Figs 5 and 9). It is a pale pegmatite-layered syenogranite gneiss with pegmatite layers of various thicknesses and spacing in an otherwise uniform, even-grained leucocratic granite. There is a flaser foliation in much of the gneiss, mostly parallel with the layering, but in fold hinges it is axial planar to F_2 isoclinal folds. Thin pegmatite veins are parallel to the flaser fabric and the axial planes of F_2 folds. Quartz and feldspar in S_2 flaser fabrics are dark grey and appear waxy, and have been little affected by metamorphism during D_3. Small garnets grew in association with, and overprint, D_2 fabrics. These features reflect the syn- to post-

D_2 peak metamorphism in granulite facies. In most of the region they are obliterated by syntectonic recrystallization at amphibolite-facies conditions during D_3.

At Locality 5, the dominant regional north–south D_3 tectonic trend is developed only locally in shear zones and fold limbs, where older pegmatite layering and D_2 fabrics are rotated and attenuated (Fig. 5). The sample of syenogranite gneiss selected for SHRIMP U–Pb zircon dating (GSWA 80333; Fig. 5) has an igneous crystallization age of 3385 ± 8 Ma (Myers et al., 1990). It has a moderately well preserved igneous texture with coarse- and fine-grained components, partly overprinted by the D_2 tectonic fabrics, but unaffected by D_3. The granitic gneiss contains small garnets, and quartz and feldspar are dark and appear waxy. West of the sample site, a metadolerite dyke crosscuts the D_2 structures and fabrics. A few hundred metres to the southeast, similar dolerite dykes are cut by the D_3 shear zones and are converted locally to amphibolite schist.

In the hills south of Locality 5, there are marked variations in the intensity of D_2 deformation in similar syenogranite of the Dugel Gneiss. The gneiss contains layers of pelitic rocks, mainly quartz–cordierite–garnet rocks with granoblastic textures.

Errabiddy Shear Zone

by

I. M. Tyler, S. A. Occhipinti, S. Sheppard, and D. R. Nelson

The Errabiddy Shear Zone defines the boundary between the northwestern part of the Yilgarn Craton (early to late Archaean Narryer Terrane), the 3300–1800 Ma Yarlarweelor Gneiss Complex (reworked Narryer Terrane), and the southern part of the Gascoyne Complex (2540–1970 Ma Glenburgh Terrane; Figs 2, 3, and 10). The shear zone contains components of the:

- Narryer Terrane — granitic gneiss derived from early to late Archaean rocks, and interleaved metasedimentary and mafic to ultramafic igneous rocks of the Narryer Terrane, heterogeneously deformed and metamorphosed during the late Archaean (Myers, 1990b);
- Warrigal Gneiss — well-foliated to banded late Archaean granites (2700–2600 Ma) of the Narryer Terrane;
- Yarlarweelor Gneiss Complex — the part of the Narryer Terrane that was deformed, metamorphosed, and intruded by voluminous granite sheets and dykes during the Palaeoproterozoic Capricorn Orogeny (Occhipinti et al., 1998; Occhipinti and Myers, 1999; Sheppard and Swager, 1999);
- Camel Hills Metamorphics — dominantly Palaeoproterozoic metasedimentary rocks, which are confined to the Errabiddy Shear Zone, and includes the Petter Calc-silicate and the Quartpot Pelite;
- Bertibubba Supersuite — c. 1960 Ma granite intruded into the northwestern Yilgarn Craton along the Errabiddy Shear Zone. Granite dykes (c. 1950 Ma) that cut the Dalgaringa Supersuite in the southernmost Glenburgh Terrane (Figs 2, 3, and 11) may also be related to the Bertibubba Supersuite;
- Coor-de-wandy Formation — >1800 Ma low-grade micaceous metasedimentary schists;
- ?Mount James Formation — <1800 Ma low-grade metasiliclastic rocks, largely composed of quartz sandstone and quartzite.

All components have fault-bounded contacts with each other except for some of the granites of the Bertibubba Supersuite, which intrude granitic gneiss of the Narryer Terrane. The Coor-de-wandy Formation (Drew, 1999a,b), which is unconformably overlain by the ?Mount James Formation, may now be sheared. Components of the Glenburgh Terrane have not been found within the Errabiddy Shear Zone.

Development of the Errabiddy Shear Zone

The Errabiddy Shear Zone (Figs 2, 3, and 11) initially developed during the 2000–1960 Ma Glenburgh Orogeny (Occhipinti et al., 1999b). This orogeny reflects the east- or southeast-directed convergence and accretion of the latest Archaean to Palaeoproterozoic Glenburgh Terrane onto the passive western or northwestern margin of the Yilgarn Craton (Occhipinti et al., 1999b; Occhipinti and Sheppard, 2001; Table 2). During the Glenburgh Orogeny, rocks of the Camel Hills Metamorphics, Warrigal Gneiss, and Narryer Terrane were tectonically interleaved. Subhorizontal or gently dipping faults and folds and a gently dipping or subhorizontal foliation developed within and between these units, and have been correlated with the regional D_{2g} deformation (Table 2). Rocks caught up in the Errabiddy Shear Zone at this time were metamorphosed at medium to high grade, with metamorphic mineral assemblages in the Camel Hills Metamorphics, which were also locally migmatized, forming during D_{2g} (Table 2). Granite of the Bertibubba Supersuite intruded into the northwestern margin of the Yilgarn Craton at or towards the end of the Glenburgh Orogeny.

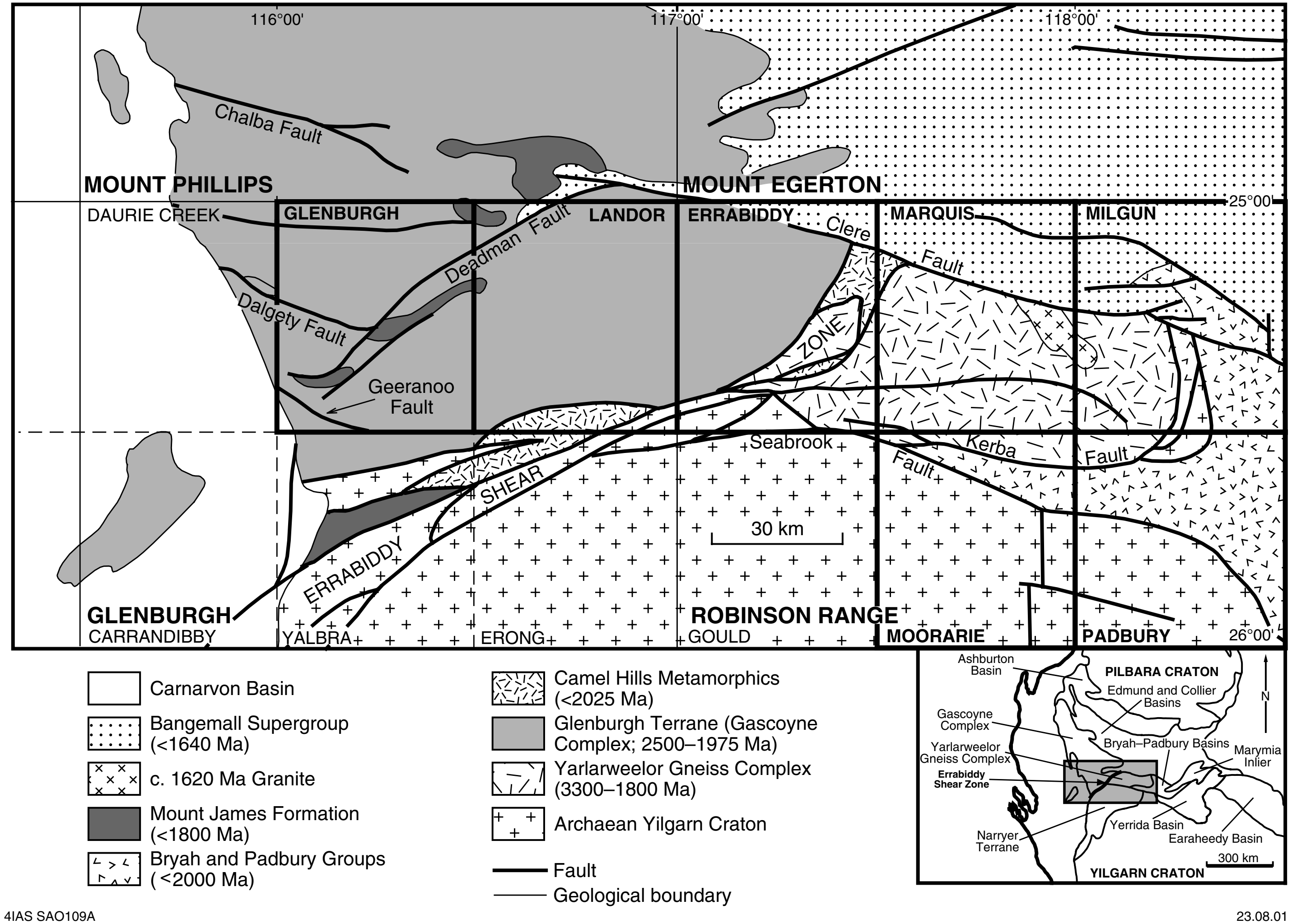

Figure 10. Simplified geological map of the southern part of the Gascoyne Complex (Glenburgh Terrane) and part of the Errabiddy Shear Zone. Bold outlined areas represent 1:100 000-scale maps recently published by the GSWA (1996–2000). Thin solid lines represent 1:250 000-scale maps

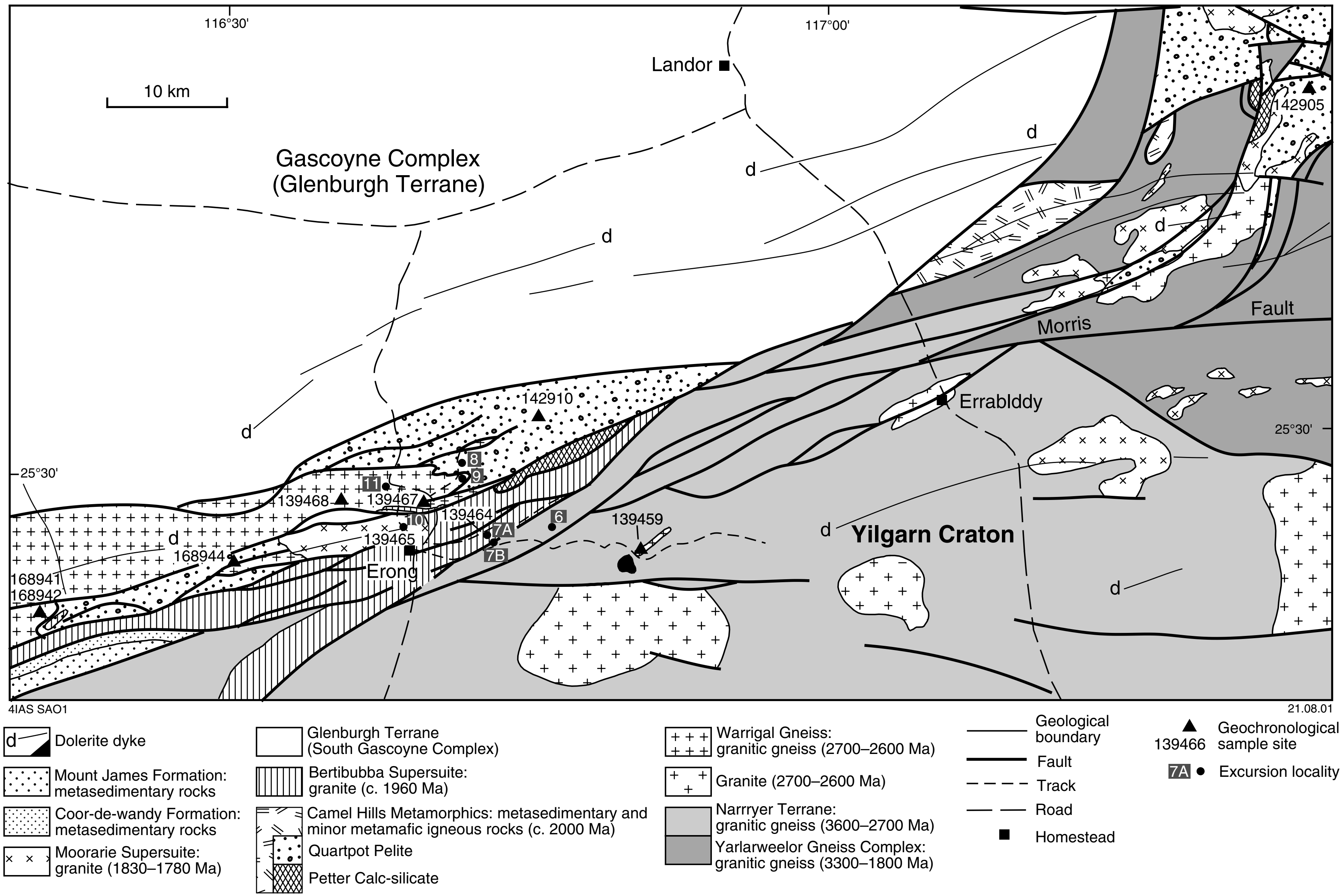

Figure 11. Simplified geological map of the Errabiddy Shear Zone showing main rock units, geochronological sites, and Localities 6–11

During the 1830–1780 Ma Capricorn Orogeny, the Errabiddy Shear Zone, northwestern Yilgarn Craton (including the Yarlarweelor Gneiss Complex), and Glenburgh Terrane were deformed. Rocks that already formed part of the Errabiddy Shear Zone were variably folded, faulted, and foliated during this D_{1n} deformation event (Table 2). These folds are commonly tight and upright and contain shallow to steeply plunging fold axes. They were refolded during the latter stages of the Capricorn Orogeny by east-southeasterly trending dextral strike-slip movement and north–south compression, which appears to have accompanied a transition from ductile to brittle conditions in the region. The Yarlarweelor Gneiss Complex, which largely originated between 1820 and 1800 Ma by pervasive reworking of a tectonic slice of Narryer Terrane and intrusion of voluminous granite and pegmatite (Occhipinti et al., 1998), was uplifted, possibly in a restraining bend, at this time. Moderate to gentle south-dipping faults, which developed in discrete narrow zones within the northwestern part of the Yilgarn Craton, probably also formed during the Capricorn Orogeny. These faults may reflect backthrusting of parts of the Yilgarn Craton over the Errabiddy Shear Zone (Occhipinti et al., in prep.).

Excursion localities — Errabiddy Shear Zone

Locality 6: Narryer Gneiss (MGA 479249E 7174670N)

From Mount Narryer Homestead turn right onto the Carnarvon–Mullewa road. Travel north for 59 km to the junction with the Beringarra–Byro road. Turn right and travel east for 52 km to a junction with Erong Road. Turn left towards Erong Springs Homestead. After 24.5 km bear left at a Y-junction (right to Yundra Outcamp). After a further 25.3 km go straight across a grid at a T-junction (left to Innouendy Homestead) then continue for 15.6 km to Erong Springs Homestead.

Continue past Erong Springs Homestead for 500 m then go through a gate in the fence on the south side of the road. Turn left and travel east along the fence. After 11.2 km bear left at a Y-junction for 300 m to Bunty Well. At Bunty Well go through the yard and follow a derelict fence to the northeast for 1.9 km. Leave the fence and head northwest for 200 m to a creek. Cross the creek (at MGA 478823E 7173859N). From here, head northeast for 900 m towards low rocky hills and park (at MGA 479125E 7174551N).

At this locality (Fig. 11) pavements and small tors of banded granitic gneiss are present. The gneiss comprises several phases including a porphyritic medium-grained biotite monzogranite, an even-textured leucocratic granite, and pegmatite. Hooked refolds can be seen on a pavement on the southeastern side of the outcrop, with early, isoclinal, layer-parallel folds refolded about steeply plunging, upright, westerly trending folds.

The banding in the granitic gneiss is defined by the porphyritic biotite monzogranite and the leucocratic granite. It is cut by two generations of pegmatites: an early one that is folded by the upright folds, and a later one intruded subparallel to the axial surfaces of the upright folds.

Pods and lenses of amphibolite ranging from less than one metre to several hundreds of metres in length are widespread in the granitic gneiss, and are within the banding and folded with it. A prominent outcrop of metapyroxenite is present to the west of the main granitic gneiss outcrop. Elsewhere, remnants of greenstones consisting of BIF, metagabbro, calc-silicate gneiss, pelite, and psammite are also interlayered and folded with the granitic gneiss. The relationship between these and the granitic gneiss is uncertain. However, as at Mount Narryer they may have been tectonically interleaved during early deformations, which also produced the gneissic banding and layer-parallel folds. Some of the amphibolites may represent dykes intruded into the original granites.

Table 2. Summary of geological history of the Errabiddy Shear Zone and the Glenburgh Terrane

		Domain	
Age (Ma)	*Deformation event*	*Errabiddy Shear Zone*	*Glenburgh Terrane*[(a)]
PRE-GLENBURGH OROGENY (>2000)			
c. 2500			Crystallization of some granite protoliths to the Halfway Gneiss
?			Deposition of protoliths of Moogie Metamorphics
GLENBURGH OROGENY (2000–1960)			
			Intrusion of Dalgaringa Supersuite
>2000	D_{1g}, S_{1g} M_{1g}; medium to high grade		**SGT** — Foliation in gneissic and foliated granite of the Dalgaringa Supersuite
?< 2025–1960		Deposition of precursor sedimentary rocks of Camel Hills Metamorphics onto ?northern Yilgarn Craton and ?south Gascoyne Complex	**NGT** — Gneissosity in Halfway Gneiss
c. 1990	Continuation of M_{1g}; up to granulite facies		**SGT** — Metamorphism and local folding of c. 2000 Ma granitic rocks and coeval intrusion of pegmatite, quartz diorite, and monzogranite
c. 1975			**SGT** — Intrusion of the Nardoo Granite into older granitic rocks
c. 1960	D_{2g}, F_{2g}, S_{2g} M_{2g}, high grade D_{2g}, F_{2g}, S_{2g} M_{2g}; epidote–amphibolite facies	Metamorphism, deformation, and local migmatization of Camel Hills Metamorphics; local recumbent to subhorizontal folds, formation of subhorizontal fabric; intrusion of biotite granite of the Bertibubba Supersuite into northwestern margin of Yilgarn Craton	**SGT** — Deformation of Nardoo Granite; local retrogression of high-grade assemblages in Dalgaringa Supersuite; folding of 2000–1990 Ma foliated and gneissic granites and mafic gneisses of the Dalgaringa Supersuite
?	?S_{2g} Medium- to high-grade metamorphism		**NGT** — Juxtaposition of the Halfway Gneiss and the Mumba Pelite; formation of flat faults and mylonite zones; subhorizontal folds in Moogie Metamorphics and Halfway Gneiss; foliation in Moogie Metamorphics
POST-GLENBURGH OROGENY			
c. 1950			**SGT** — Intrusion of dykes of leucocratic granodiorite, biotite trondhjemite, and monzogranite

Table 2. (continued)

Age (Ma)	*Deformation event*	*Domain* *Errabiddy Shear Zone*	 *Glenburgh Terrane*[(a)]
CAPRICORN OROGENY (1830–1780)			
c. 1825	D_{ln}, F_{ln}, S_{ln}	Intrusion of plugs, sheets, and veins of granitic rocks of Moorarie Supersuite	
c. 1810	M_{ln}; greenschist-facies metamorphism		**NGT** — Intrusion of Dumbie Granodiorite; upright folding in Moogie Metamorphics and Halfway Gneiss **SGT** — Local foliation in c. 1825 Ma Moorarie Supersuite
c. 1800			**NGT; ?SGT** — Intrusion of Scrubber Granite; local deformation of the Scrubber Granite
?<1800	?Local extension		**NGT; SGT** — Deposition of Mount James Formation sediments in possible elongate 'strike-slip pull-apart' structures
>1640		Coplanar deformation with D_{ln} in the Glenburgh Terrane[(b)], possibly into tight asymmetric folds; deformation of Mount James Formation	
<1640		Deposition of Edmund Group in intracratonic sag, mainly over the Gascoyne Complex (including Glenburgh Terrane)	
c. 1465		Extension; intrusion of dolerite sills into the Edmund Group	
EDMUNDIAN OROGENY (1020–750)			
<c. 1020	D_{le}, F_{le}, S_{le}; subgreenschist metamorphism	Deformation of Edmund Group	
>750	D_{2e}, S_{2e}	Deformation of Edmund Group	

NOTES: (a) SCT: Southern Domain of the Glenburgh Terrane; NGT: Northern Domain of Glenburgh Terrane
(b) Alternatively, the Mount James Formation may have been deformed during the Edmundian Orogeny

Ten kilometres to the southeast of Locality 6, around Black Hills Bore and outside of the Errabiddy Shear Zone, granitic gneiss containing similar greenstone remnants outcrops. There, a weakly foliated, sparsely porphyritic biotite monzogranite with a SHRIMP U–Pb zircon age of 2738 ± 5 Ma (GSWA 139459; Nelson, 2000) intrudes the granitic gneiss. This indicates a minimum age of Archaean for the gneisses and for the gneissic banding, consistent with them belonging to the Narryer Terrane.

Across the valley to the southeast of Locality 6, a prominent quartz vein has intruded a major fault zone. The quartz vein is locally mylonitized and the adjacent granitic gneiss is intensely deformed and mylonitic parallel to the axial surfaces of the upright folds, which are correlated regionally with D_{1n} (Table 2). Although the regional shear sense is difficult to ascertain, locally a dextral strike-slip shear sense is indicated by S–C fabrics and delta- and sigma-tails around feldspar phenocrysts.

Day 4

Today, components of the Errabiddy Shear Zone will be visited. These are the granitic gneiss of the Archaean Warrigal Gneiss, metasedimentary rocks of the Palaeoproterozoic Camel Hills Metamorphics, and the c. 1800 Ma Erong Granite. By the end of the day we will have travelled north of the Errabiddy Shear Zone into the Glenburgh Terrane of the Gascoyne Complex.

Locality 7: Granite of the Bertibubba Suite and mylonite zone

From Locality 6 follow the vehicle tracks back to the creek crossing. Cross the creek and return to the fence, following it back to Bunty Bore. Return to Erong Road east of Erong Springs Homestead. Turn right onto Erong Road and head east for 600 m. At a Y-junction bear right onto a station track and travel east for 6.2 km (to MGA 472926E 7173960N). Turn to the southeast following vehicle tracks cross-country to a fence. Follow the fence to a geochronology sample site (at MGA 473633E 7173767N, 900 m from the track). This is Locality 7.

At this locality (Fig. 11) monzogranite of the c. 1960 Ma Bertibubba Supersuite is present, which was intruded into the northwestern margin of the Yilgarn Craton at the end of the Glenburgh Orgeny. A traverse will be made southwards from an area of low strain in the monzogranite (Locality 7a) to a prominent ridge formed by a mylonite zone that comprises mylonitized granite and quartz-vein material (Locality 7b).

Locality 7a: Porphyritic biotite monzogranite (MGA 473646E 7173731N)

At this locality, variably foliated porphyritic biotite monzogranite is veined by pegmatite. An even-textured, medium-grained biotite monzogranite is also present.

A SHRIMP U–Pb zircon age of 1958 ± 5 Ma has been interpreted as the age of igneous crystallization for the porphyritic biotite monzogranite, with an older group of zircons dated at 2619 ± 8 Ma interpreted as xenocrysts (GSWA 139464; Nelson, 2000; Fig. 12).

Walk south through the porphyritic biotite monzogranite towards a quartz-vein ridge (Locality 7b; see below).

The porphyritic granite has been heterogeneously deformed, is locally well foliated, and contains a strong, subhorizontal mineral lineation. In places, narrow shear zones are present, subparallel to the foliation. Rotated feldspar phenocrysts and S–C fabrics indicate both sinistral and dextral movements.

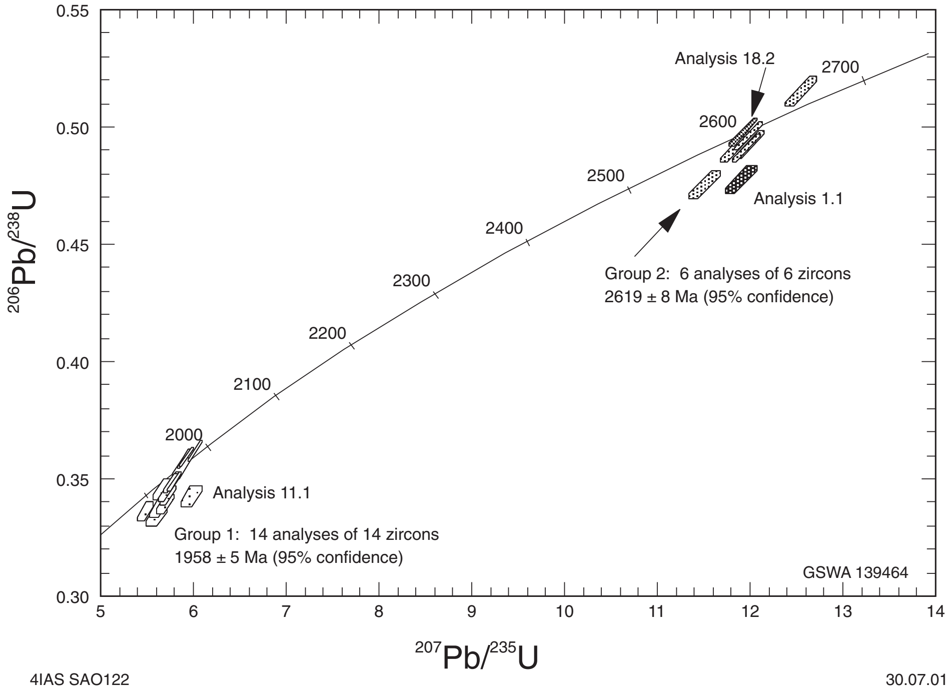

Figure 12. Concordia plot for porphyritic biotite monzogranite of the Bertibubba Supersuite, 3 km southwest of Camel Hills Bore (GSWA 139464)

Locality 7b: Mylonite zone (MGA 474078E 7172884N)

This quartz-vein ridge is a mylonite zone that consists of mylonitized granite and vein quartz. Sheath folds are present (Fig. 13) and are consistent with vertical movements. The foliation in the granite increases in intensity closer to the mylonite zone, grading into it with no evidence of overprinting. The foliation in the granite is regarded as the same age as the mylonite, corresponding to the regional D_{1n} event (Table 2).

Locality 8: Diatexite of the Quartpot Pelite (Camel Hills Metamorphics; MGA 470872E 7180091N)

From Locality 7b return to the station track and then to Erong Road. Turn right and follow the road for 5.7 km. Turn right (at MGA 468206E 7176893N) and follow a station track for 600 m to Randell Well. Find a track to the east following a cleared line through a creek. Follow this indistinct and grass-covered track (initially beside a creek) to the southeast for 4.3 km, to Locality 8. You will be travelling through rocky outcrop for the last 2.3 km.

The Quartpot Pelite of the Camel Hills Metamorphics (Sheppard and Occhipinti, 2000) forms low rocky outcrops throughout the Errabiddy Shear Zone and largely consists of pelitic and psammitic schist or gneiss interlayered with minor quartzite, calc-silicate gneiss, and amphibolite. At this locality the unit is migmatized forming a diatexite (Figs 14a–c). Diatexite migmatites extend east of this locality and into a region northeast of Errabiddy Homestead (Sheppard and Occhipinti, 2000). To the southeast and west-southwest, medium-grade metasedimentary rocks of both the Quartpot Pelite and the Petter Calc-silicate are present, consistent with a decrease in metamorphic grade from northwest to southeast across the Errabiddy Shear Zone.

The mineral assemblage in the migmatized Quartpot Pelite in this part of the Errabiddy Shear Zone consists of biotite, garnet, muscovite, sericite mats (possibly

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Figure 13. Sheath folds in a mylonite zone within the Errabiddy Shear Zone at Locality 7b. Sheath folds are recognized by the change of plunge of fold axes. The long axes of the two pens have been placed subparallel to the plunge of folds in the photograph

after sillimanite), plagioclase (andesine), quartz, and K-feldspar. This mineral assemblage is consistent with in situ partial melting at upper amphibolite-facies conditions or higher. Metamorphism took place during the Glenburgh Orogeny synchronous with the development of a subhorizontal, bedding-parallel foliation (S_{2g}), which is interpreted to have developed during D_{2g} (Table 2). The S_{2g} foliation is largely defined by the alignment of sillimanite and biotite, together with differentiation into sillimanite–biotite and quartz–plagioclase domains. Garnet crystallized synchronously with, or just after D_{2g}, although garnet porphyroblasts may be partially or completely pseudomorphed by chloritoid or chlorite. West-southwest of Locality 7 around Paperbark Well, the Quartpot Pelite consists of medium-grade pelitic and psammitic schist and gneiss, which has not been migmatized, and has been metamorphosed in the amphibolite facies. This indicates that there is a metamorphic isograd between these outcrops of Quartpot Pelite and those to the east. This isograd was folded during the Capricorn Orogeny (D_{1n}).

The migmatites commonly show stromatic, schollen (raft), and nebulitic structures, with medium-grained, heterogeneous siliceous diatexite melt locally forming up to 70% of the rock. Within the melt phase, lenticular rafts of restite consisting of refractory psammite and biotite-rich material are preserved. In places, veins of more homogeneous, externally derived melt cut the in situ migmatite. The stromatic migmatites sometimes grade into nebulitic migmatite, indicating local increased degrees of in situ partial melting.

The S_{2g} foliation was tightly folded during the Capricorn Orogeny (F_{1n}), and a well-developed S_{1n} crenulation cleavage is present locally, largely defined by the alignment of muscovite. At this locality the F_{1n} folds and S_{1n} crenulation cleavage are westerly trending and plunge moderately to steeply to the west, parallel to the regional trend of the Errabiddy Shear Zone (Figs 2, 3, and 11). Elsewhere within the shear zone, D_{1n} structures may be northeasterly or northerly trending (Figs 2, 3, and 11).

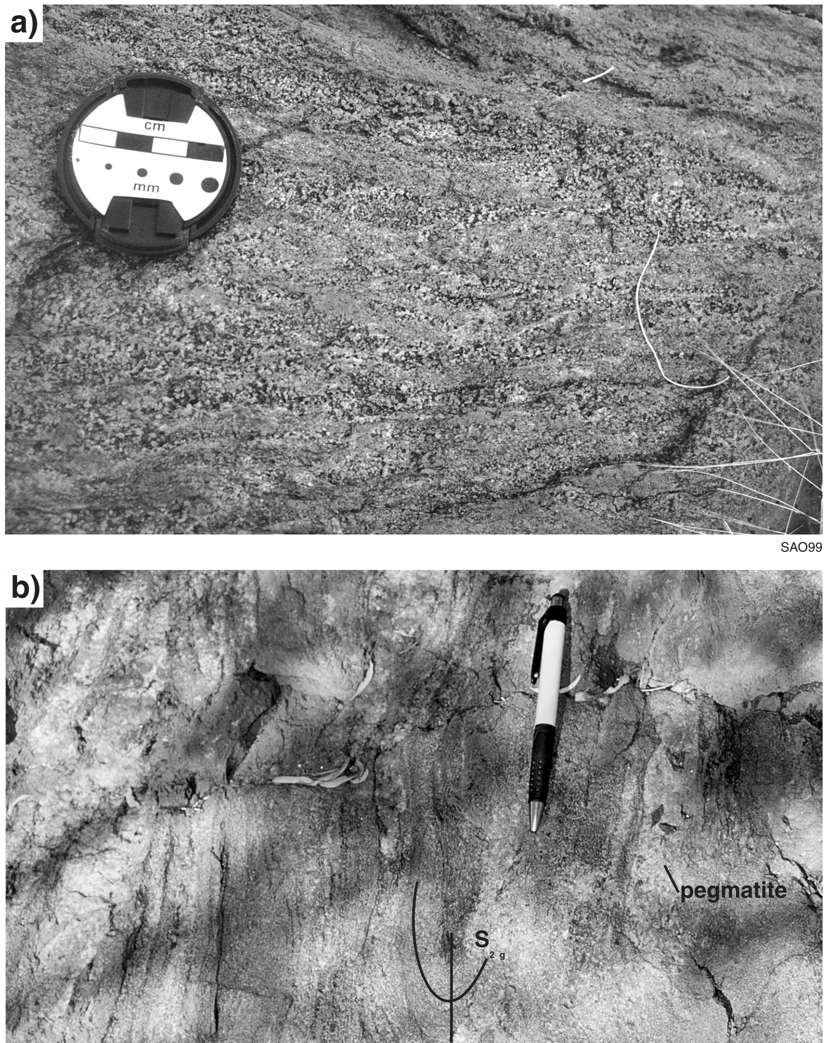


Figure 14. Migmatitic pelitic gneiss of the Quartpot Pelite at Locality 8: a) melt patches in diatexite; b) thin pegmatite veins cutting diatexite subparallel to the fold axial surface of small-scale F_{1n} folds; c) a thin vein of pegmatite cutting the S_{1g} fabric in diatexite but folded by F_{1n}

The Quartpot Pelite has been sampled for SHRIMP U–Pb zircon geochronology at three localities within the Errabiddy Shear Zone. The samples are: GSWA 142905 located 11 km south of Pines Bore (448000E 7210600N; Nelson, 1998), GSWA 142910 located 4 km south of Pannikan Bore (478000E 7284300N; Nelson, 1998), and GSWA 168944 located 2 km east of Paperbark Well (449826E 7171523N; Nelson, in press).

At the first two localities the diatexite melt phase was sampled, which locally comprised over 50% of the outcrop's volume. Pitted zircon grains were present in both samples with ages of 2550–2025 Ma and were interpreted to be of detrital origin (Fig. 15). The youngest zircon populations in GSWA 142905 and 142910 (Figs 15a,b) were obtained from rims and cores of pitted grains with very low Th/U ratios. In both samples the youngest zircon populations consisted of only two zircons dated at 1951 ± 13 and 1966 ± 5 Ma. Based on the pitted nature of the zircon surfaces, Nelson (1998, 1999) suggested that the rims were incorporated into the zircon prior to detrital sedimentary transport, therefore providing a maximum depositional age for the sedimentary protolith. However, this interpretation implies that no new zircon growth took place during migmatization, and is contradicted by the 1970 ± 15 Ma age (Nelson, 1998) of a trondhjemite dyke that cuts the migmatized Quartpot Pelite at the locality south of Pannikin Bore. The pitted zircon surfaces and the low Th–U ratios of the zircon rims are similar to those described by Fraser et al. (1997) and Tyler et al. (1999), which are related to zircon growth during partial melting of metasedimentary rocks. This suggests that the age of high-grade metamorphism and its associated migmatization was c. 1960 Ma, and that the maximum age of deposition of the protolith to the Quartpot Pelite was c. 2025 Ma.

At the Paperbark Well locality, a medium-grade, unmigmatized medium-grained quartz–plagioclase–biotite–muscovite psammite of the Quartpot Pelite was sampled for SHRIMP U–Pb zircon geochronology. Nineteen concordant to slightly discordant analyses of 19 zircons gave a date of 2028 ± 5 Ma. Nelson (in press; GSWA 168944; Fig. 15c) described most of these zircons as containing rounded and pitted surfaces, indicating they have undergone detrital transport, and interpreted the date of 2028 ± 5 Ma as the maximum age for deposition of the sandstone precursor to the psammite. A younger age of 1985 ± 14 Ma was also obtained; however, this is based on a single analysis on an irregular zircon fragment (Nelson, in press). Older populations ranging from 2250 to 2700 Ma are also present (Nelson, in press).

Most of the detrital zircons dated from the Quartpot Pelite were probably sourced from the Glenburgh Terrane of the southern Gascoyne Complex. Only a minority of zircons were sourced from the Yilgarn Craton (Fig. 15), and a few c. 2250 Ma zircons are from an unknown source.

Outcrops of the Quartpot Pelite are locally extensively intruded by dykes and veins of coarse-grained, even-textured, leucocratic biotite(–muscovite–tourmaline) granite or pegmatite. These granites contain the same metamorphic and structural relations as abundant c. 1800 Ma granite throughout the region, and are therefore correlated with them.

Locality 9: Warrigal Gneiss (MGA 470691E 7178872N; time permitting)

From Locality 8, return to the south along the track for 1.3 km to a prominent rocky hill. This is Locality 9.

At this locality (Fig. 11) the Warrigal Gneiss, which is locally pegmatite banded, consists of well-foliated to banded, interleaved mesocratic and leucocratic granite phases. The mesocratic phase has been dated at a site 5 km to the west-southwest (see **Locality 11**) giving a SHRIMP U–Pb zircon age of c. 2700 Ma (Nelson, 2000). The

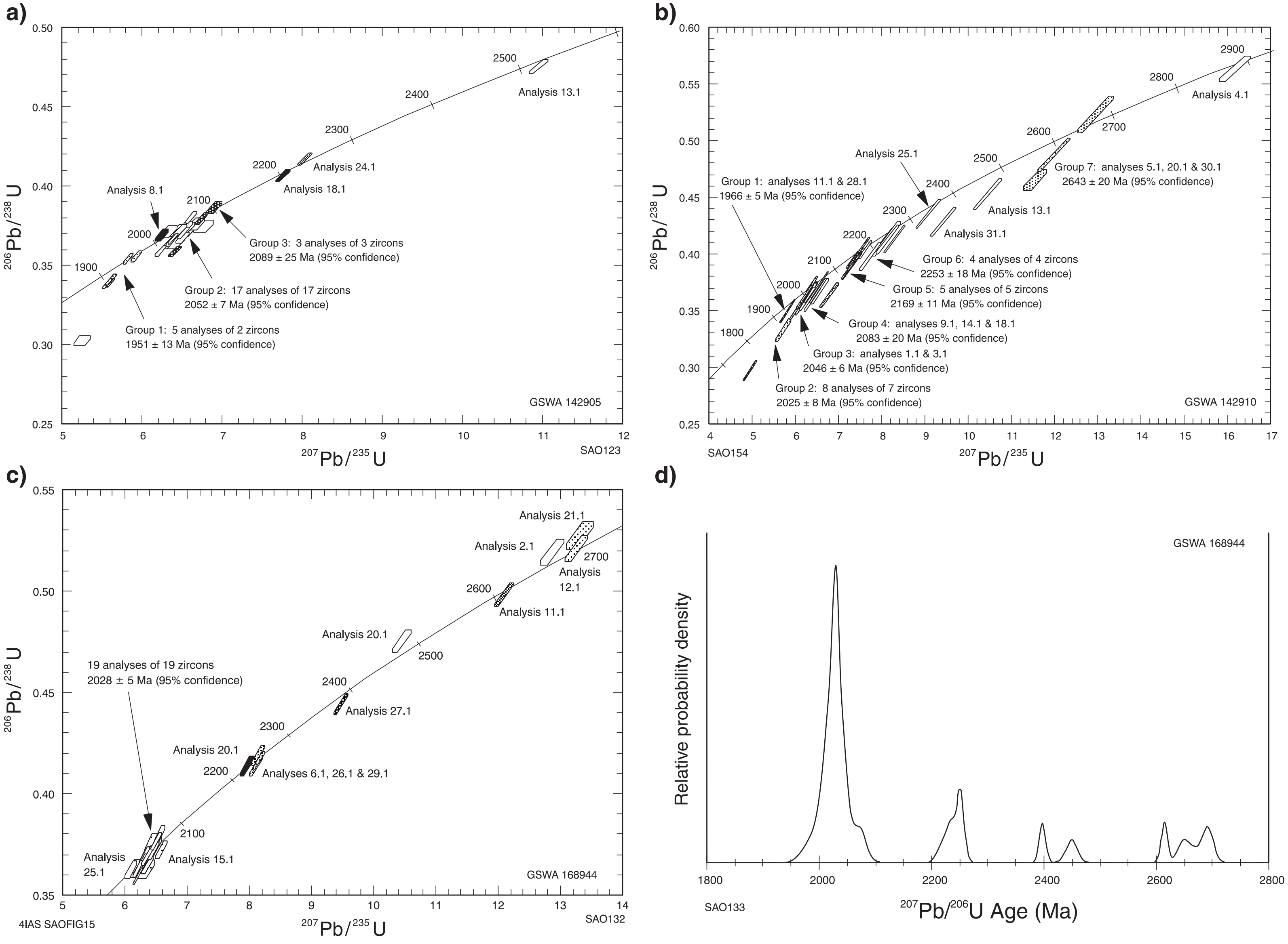


Figure 15. Geochronology results for samples of the Quartpot Pelite (Camel Hills Metamorphics): a) concordia plot for GSWA 142905; b) concordia plot for GSWA 142910; c) concordia plot for GSWA 168944; d) Guassian-summation probability density plot for GSWA 168944 (from Nelson, 1998, in prep.)

late Archaean age of the precursor granites to the Warrigal Gneiss distinguishes it from the Narryer Gneisses (i.e. Meeberrie, Dugel, and Eurada Gneisses).

The gneissic banding in the Warrigal Gneiss developed during the 2000–1960 Ma Glenburgh Orogeny. The dominant foliation, or gneissic banding, in the gneiss is subparallel to the regional S_{2g} gneissosity in the Camel Hills Metamorphics. A ‘flaser’ texture within it, which consists of leucocratic veinlets up to 2 mm wide, may represent possible in situ minimum melt of the gneiss. This texture appears to have developed during D_{2g} and is consistent with the age of migmatization in the adjacent Quartpot Pelite. The foliation is folded into tight upright F_{1n} folds.

The gneiss is intruded by c. 1800 Ma pegmatite dykes and veins, which either trend subparallel to the F_{1n} upright folds, or cut them at low to high angles. Those subparallel to the F_{1n} folds may be locally sheared or foliated. On the top of the hill is a narrow zone of sheared and mylonitized vein quartz, amphibolite, and ?altered granite.

Locality 10: The Petter Calc-silicate of the Camel Hills Metamorphics and the Erong Granite (start at MGA 465651E 7174737N)

From Locality 9, return past Randell Well to Erong Road and turn right. Follow Erong Road for 4.1 km (to MGA 465313E 7178103N) and turn left onto the track just before the road sign. Travel along this for 1 km to a Y-junction, bear left past a collapsed windmill, and after 100 m go through the old fence and follow an old track to the south towards low rocky hills. After 3 km leave the track, turning left after crossing a creek (at MGA 465268E 7174430N). Drive cross-country to the base of a rocky hill (Locality 10).

At this locality (Fig. 11) the Petter Calc-silicate of the Camel Hills Metamorphics (Locality 10a) is intruded by even-textured, medium to coarse-grained, muscovite–biotite–garnet monzogranite to pegmatite of the Erong Granite (Locality 10b), which is thought to be c. 1800 Ma in age.

Locality 10a: Petter Calc-silicate of the Camel Hills Metamorphics (MGA 465651E 7174737N)

Here, at a small knobbly hill, the Petter Calc-silicate (Fig. 16) forms a large xenolith of country rock within medium- to coarse-grained granite and pegmatite of the Erong Granite. At this locality the Petter Calc-silicate consists of:

- calc-silicate rock and a distinctive para-amphibolite, which largely comprises tremolite, diopside, and talc;
- pelitic and psammitic schist, which largely comprises biotite, quartz, plagioclase, epidote, and garnet.

The calc-silicate and para-amphibolite contains a fine, 1–10 mm-thick compositional layering, which is interpreted as original bedding (S_0; Fig. 16a). The Mg-rich mineralogy may be indicative of an evaporite protolith.

The dominant foliation (S_{2g}) in the unit, along which the metamorphic minerals are aligned, trends subparallel to the boudinaged compositional layering (S_0). The S_0/S_{2g} fabric is deformed by an upright, north-northeasterly plunging, medium-scale F_{1n} antiform–synform pair. Locally, a well-developed crenulation cleavage is present (Fig. 16b) parallel to the axial surface of these folds, particularly in the pelite–psammite unit.

The Petter Calc-silicate is intruded by dykes and veins of leucocratic granite and pegmatite (Fig. 16c), which cut or are folded by the F_{1n} folds (Fig. 16b) depending on their orientation relative to the axial surface. Late easterly trending unmetamorphosed dolerite dykes cut both the Petter Calc-silicate and the Erong Granite.

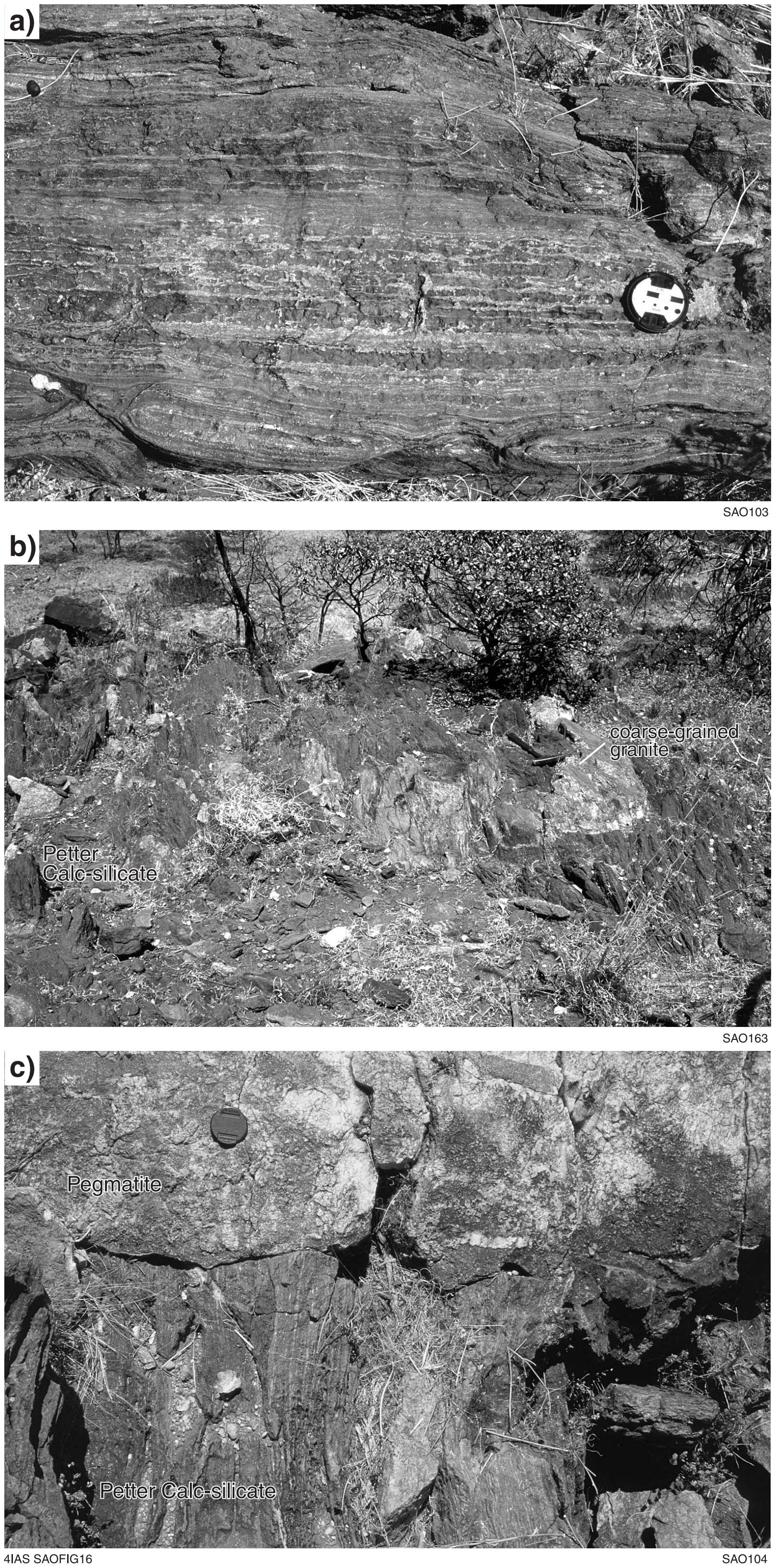


Figure 16. The Petter Calc-silicate at Locality 10a: a) a composite S_0/S_{2g} fabric in the Petter Calc-silicate that is locally boudinaged; b) a dyke of coarse-grained granite (Erong Granite) cutting the Petter Calc-silicate but folded with it into a moderately plunging F_{1n} antiform. The Petter Calc-silicate contains a well-developed S_{1n} foliation; c) a dyke of coarse-grained granite to pegmatite of the Erong Granite cutting the S_0/S_{2g} fabric in the Petter Calc-silicate

The Petter Calc-silicate was sampled to the east of this locality (about 6 km west-southwest of Packsaddle Bore at MGA 481000E 7281300N) for SHRIMP U–Pb zircon geochronology (GSWA 142908; Nelson, 1999). Most of the zircons (22 of 26) have ages between 2700 and 2600 Ma indicating they were probably sourced from the Yilgarn Craton. Three of the zircons have ages greater than 3000 Ma, and one was dated at 1944 ± 5 Ma. Nelson (1999) described the youngest zircon as rounded and pitted and interpreted the date on this grain as providing the maximum age of deposition of the precursor to the calc-silicate gneiss. However, this age is younger than the 1970 ± 15 Ma trondhjemite sheet that cuts the S_{2g} gneissic layering within the Quartpot Pelite (Camel Hills Metamorphics). The provenance of the Petter Calc-silicate may have largely been from the Yilgarn Craton and thus was different to that of the Quartpot Pelite.

Locality 10b: Erong Granite (MGA 465738E 7174598N)

The Erong Granite (Fig. 11) grades from a medium-grained, even-textured biotite–muscovite–garnet monzogranite into coarse-grained, even-textured granite and pegmatite. It has intruded as numerous granite sheets, dykes, and veins. The granite is typically massive and relatively undeformed, but locally a foliation is present in narrow low-grade shear zones.

The Erong Granite both cuts, and is deformed by, c. 1800 Ma D_{1n} structures (Table 2) and is thought to be c. 1800 Ma old, based on correlations with similar granites in the region. These include a plug of similar biotite–muscovite granite from within the Errabiddy Shear Zone to the northeast that gave a SHRIMP U–Pb age of 1802 ± 9 Ma (GSWA 142900; Nelson, 1998). However, zircons dated from the Erong Granite comprise mainly Archaean xenocrysts (Fig. 17). The low zirconium content of the Erong Granite suggests that the rock crystallized from a low-temperature minimum melt.

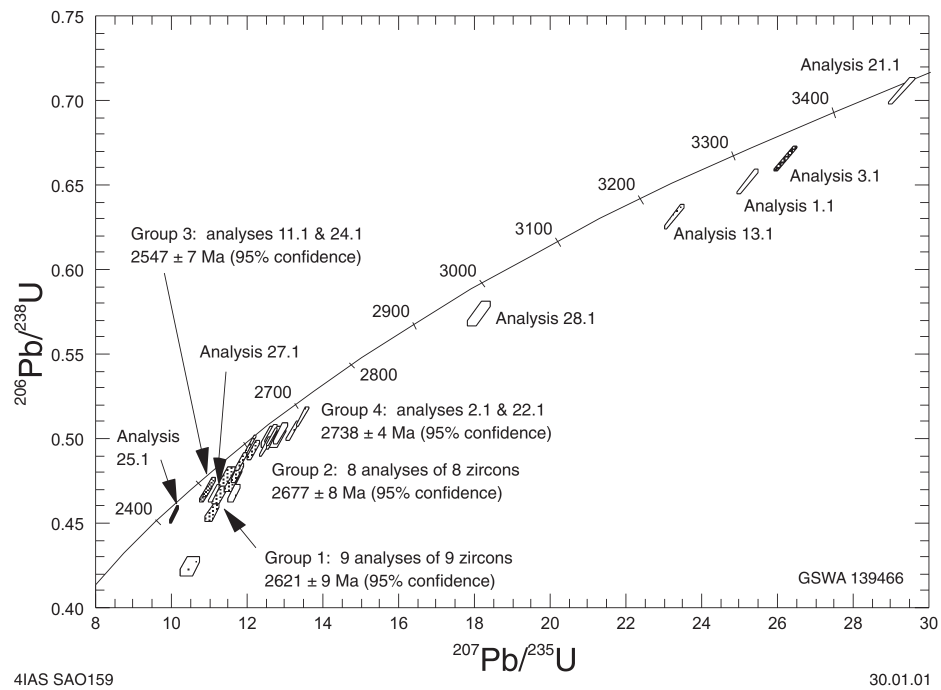


Figure 17. Concordia plot for biotite–muscovite–garnet granodiorite of the Erong Granite from Locality 10b (GSWA 139466; Nelson, 2000)

Locality 11: Warrigal Gneiss (MGA 465011E 7178643N)

From Locality 10b return to Erong Road. Turn left and follow the road for 800 m. The low rocky outcrop on left side of road is Locality 11.

Low outcrops and pavements of the Warrigal Gneiss are present at this locality. The gneiss consists of sparsely porphyritic, medium-grained biotite monzogranite and even-textured, medium-grained biotite monzogranite. The Warrigal Gneiss is folded into upright, gently plunging easterly trending folds, which have been correlated with D_{1n} structures. Locally, the gneiss contains a well-developed, spaced, vertically dipping axial-planar cleavage. Amphibolites, which are interleaved with the granitic gneiss, are also folded into the upright folds. Easterly trending dolerite dykes cut all other rock types at this locality.

The Warrigal Gneiss has been dated at three localities within the Errabiddy Shear Zone, giving SHRIMP U–Pb zircon ages for a number of granitic components that range from c. 2758 to c. 2585 Ma (Nelson, 2000, in press; Fig. 18). The gneiss consists of late Archaean granites that were intruded into the Narryer Terrane of the Yilgarn Craton, and have been deformed and metamorphosed during the Palaeoproterozoic Glenburgh and Capricorn Orogenies.

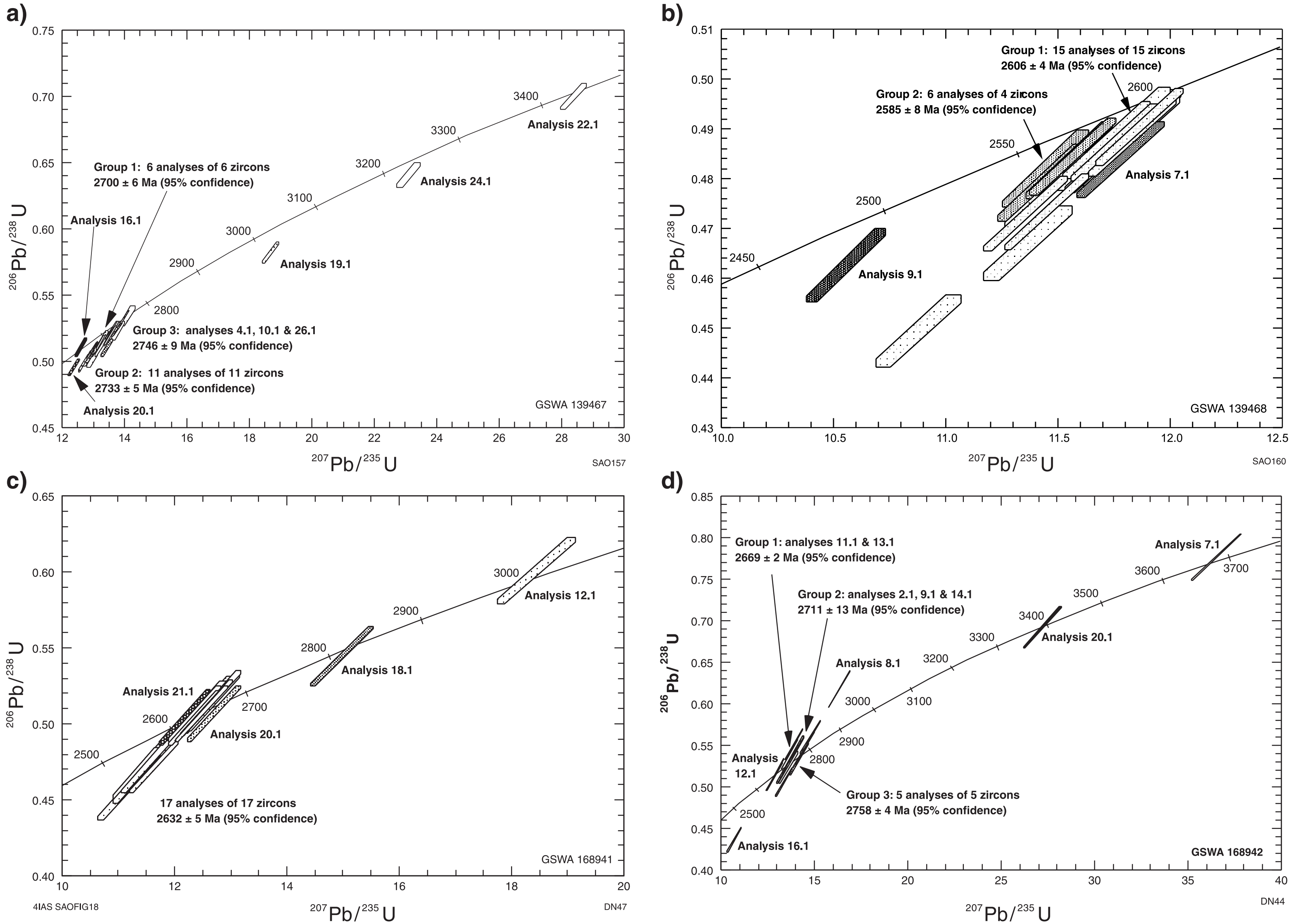


Figure 18. Concordia plots for samples of the Warrigal Gneiss: a) well-foliated porphyritic biotite monzogranite (from MGA 466400E 7176400N; GSWA 139467; Nelson, 2000); b) strongly foliated porphyritic biotite monzogranite (from MGA 460100E 7176700N; GSWA 139468; Nelson, 2000); c) metamorphosed biotite–tonalite granite (from MGA 433700E 7167600N; GSWA 168941; Nelson, in press); d) porphyritic biotite–muscovite monzogranite (from MGA 433700E 7167600N; GSWA 168942; Nelson, in press)

Glenburgh Terrane

by

S. A. Occhipinti, S. Sheppard, and D. R. Nelson

Introduction

The southern Gascoyne Complex can be divided into three main units: the Glenburgh Terrane, the Camel Hills Metamorphics, and the Moorarie Supersuite. Of these the Camel Hills Metamorphics outcrops exclusively in the Errabiddy Shear Zone, and the Moorarie Supersuite outcrops throughout the southern Gascoyne Complex. It has been suggested that the southern Gascoyne Complex comprises reworked Archaean gneisses of the Yilgarn Craton (Williams, 1986), or is the para-autochthonous Yilgarn Craton interleaved with Proterozoic rocks (Myers, 1990a). However, these interpretations were contradicted by reconnaissance SHRIMP U–Pb dating that failed to identify Archaean crust older than 2610 Ma (Nutman and Kinny, 1994).

Recent geological mapping (Sheppard and Occhipinti, 2000; Occhipinti and Sheppard, 2001) combined with SHRIMP U–Pb zircon geochronology (Nelson, 1998, 1999, 2000, in press) indicates that the southern Gascoyne Complex (Fig. 2) comprises mainly Palaeoproterozoic meta-igneous and metasedimentary rocks. However, some Archaean c. 2540 Ma foliated to gneissic granites that are present in the northern part of the Glenburgh Terrane (Occhipinti et al., 1999a,b; Nelson, 2000; Occhipinti and Sheppard, 2001) appear to be tectonically interleaved with, and juxtaposed against, the Palaeoproterozoic rocks (Figs 19–21). In the Carrandibby Inlier (the southwesternmost exposed part of the Glenburgh Terrane), Nutman and Kinny (1994) also reported a c. 2500 Ma age for granitic gneiss. Although the absolute age date on one sample of the granitic gneiss components is enigmatic (see **Locality 16**), these latest Archaean foliated to gneissic granites are younger than any dated rocks from the northwestern part of the Yilgarn Craton, which are all older than c. 2600 Ma (Wiedenbeck and Watkins, 1993; Myers, 1995; Schiøtte and Campbell, 1996; Pidgeon and Hallberg, 2000). In addition, the Pilbara Craton does not contain granite dated at younger than c. 2750 Ma (Nelson et al., 1999). Therefore, the Glenburgh Terrane may be part of a terrane separate from both the Archaean Yilgarn and Pilbara Cratons.

The Glenburgh Terrane consists of latest Archaean – Palaeoproterozoic granitic gneiss of the Halfway Gneiss interleaved with metasedimentary rocks of the Moogie Metamorphics (Fig. 22), and the 2005–1970 Ma granitic gneiss and granite of the Dalgaringa Supersuite. The Moogie Metamorphics consists of medium- to low-grade metasedimentary and mafic meta-igneous rocks that form large outcrops within the granitic gneiss, or discontinuous lenses within granite or granitic gneiss. Palaeoproterozoic medium- to high-grade metasedimentary rocks of the Camel Hills Metamorphics are confined to the Errabiddy Shear Zone; their relationship to the Moogie Metamorphics is unknown.

The Halfway Gneiss, Dalgaringa Supersuite, Moogie Metamorphics, and Camel Hills Metamorphics were deformed and metamorphosed during the 2000–1960 Ma Glenburgh Orogeny (Occhipinti et al., 1999a,b; Sheppard et al., 1999a,b). Subsequently, during the Capricorn Orogeny at 1830–1780 Ma, the rocks were deformed and metamorphosed at low to medium grade and intruded by voluminous granite and pegmatite dykes, plugs, and sheets of the Moorarie Supersuite (Fig. 22; Sheppard et al., 1999a).

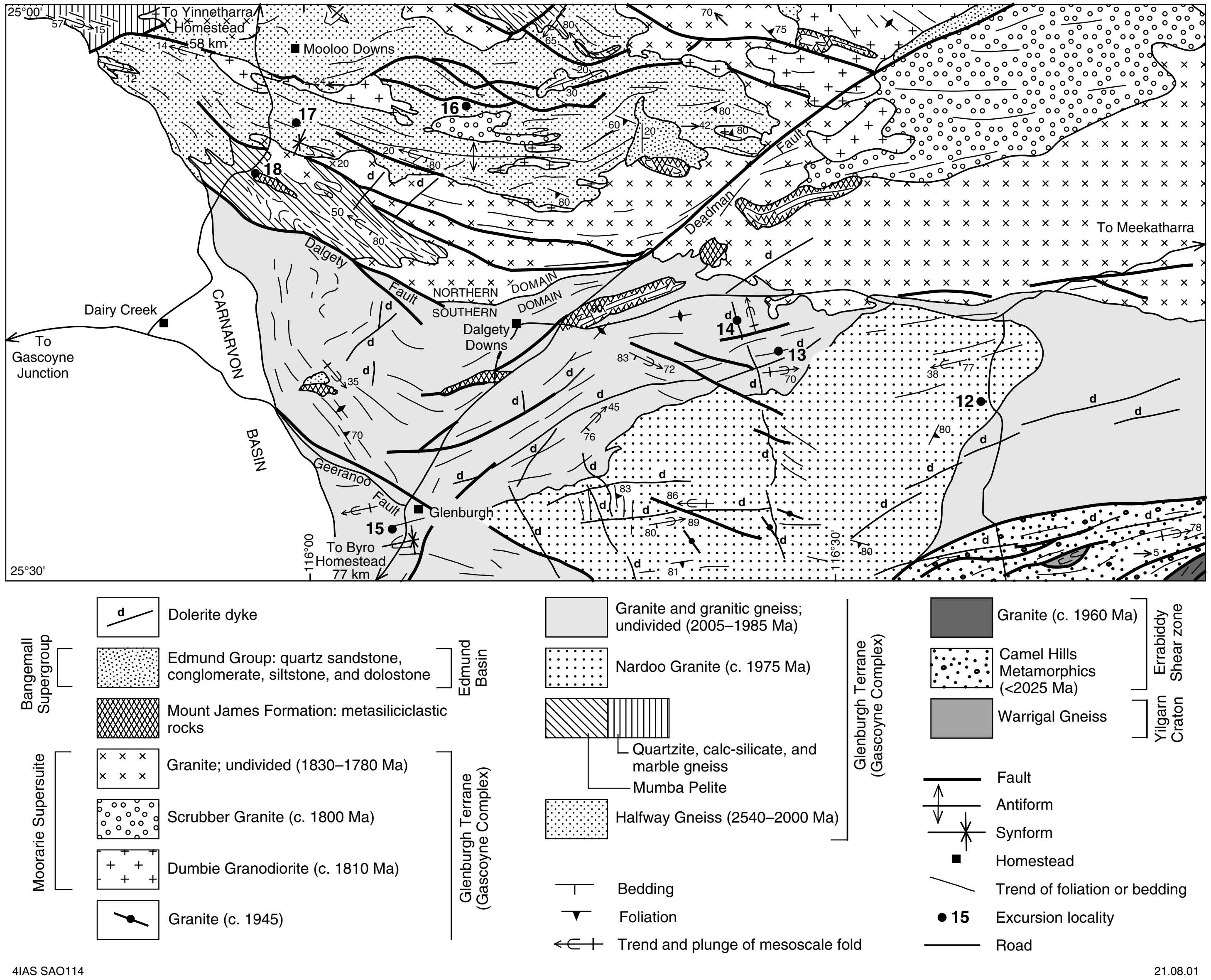
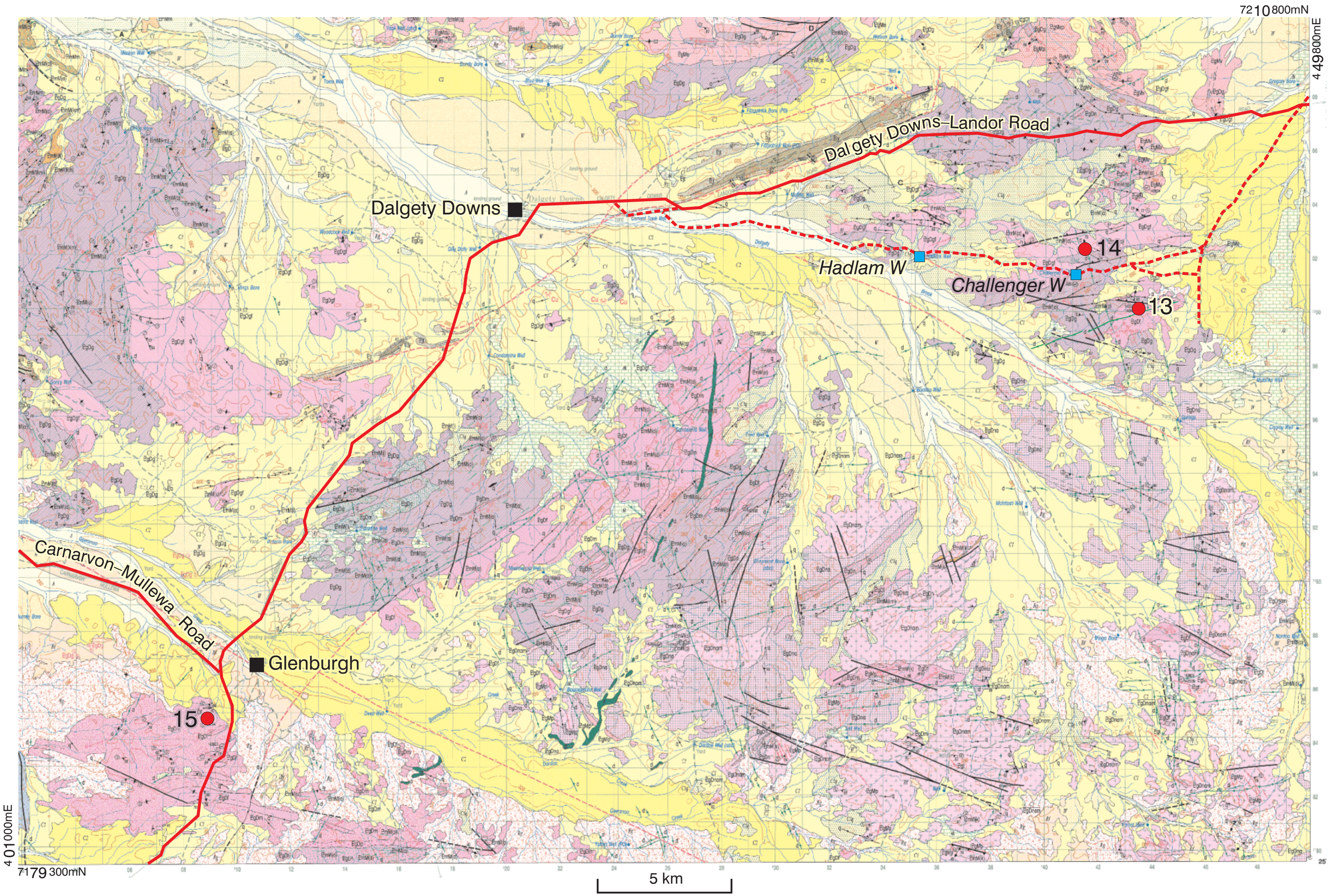


Figure 19. Simplified geological map of the Glenburgh Terrane, showing the northern and southern parts (which correlate with the structural domains of Occhipinti and Sheppard, 2001), the main rock units, and the positions of Localities 12–18



Excursion locality Road Track Well/bore Homestead

Figure 20. Portion of the GLENBURGH 1:100 000-scale geological map (Occhipinti and Sheppard, 2001), showing the positions of Localities 13–15 and main roads and tracks used on the excursion

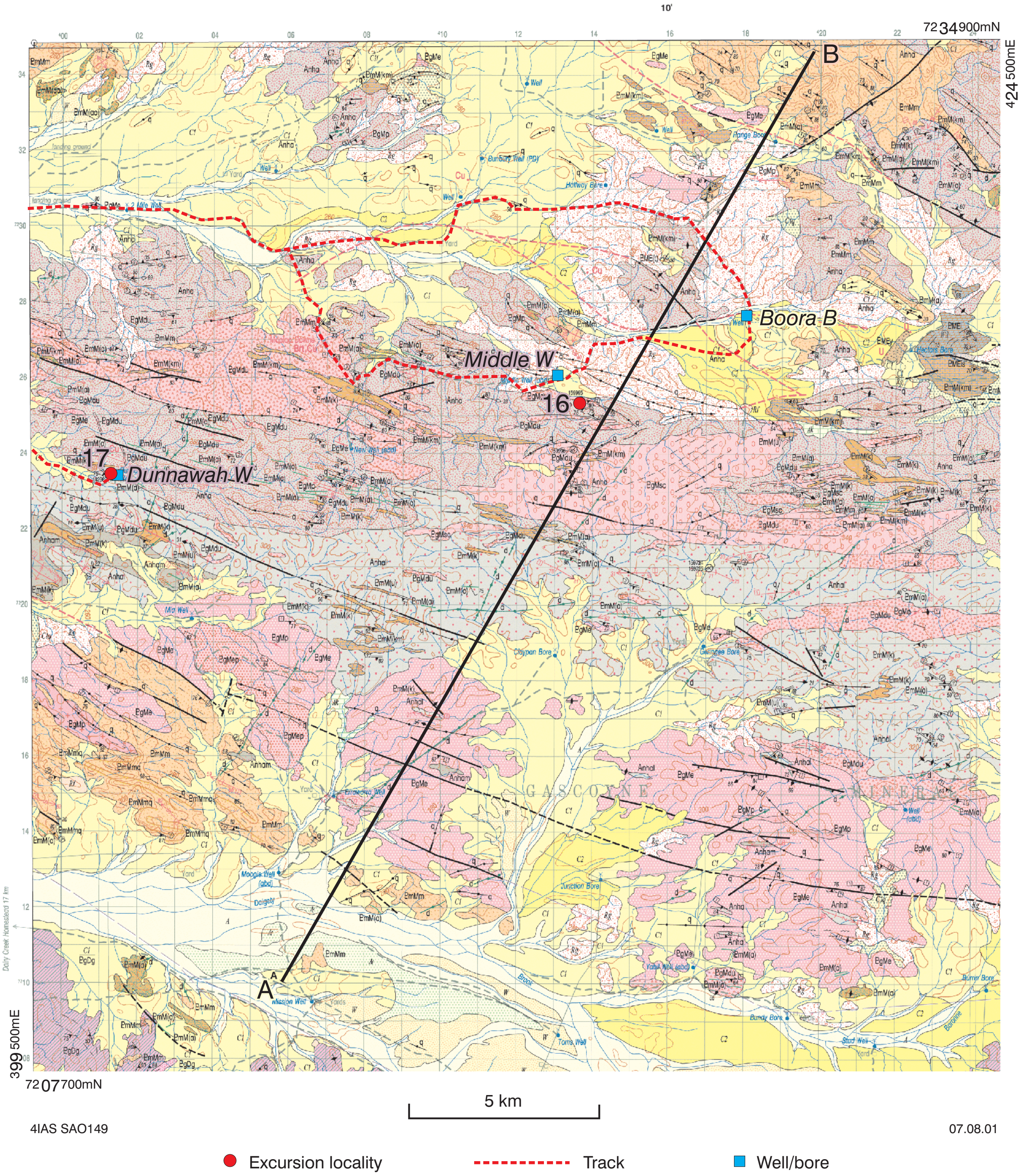


Figure 21. Portion of the GLENBURGH 1:100 000-scale geological map (Occhipinti and Sheppard, 2001), showing the positions of Localities 16 and 17 and main roads and tracks used on excursion. Cross section A–B is shown in Figure 22

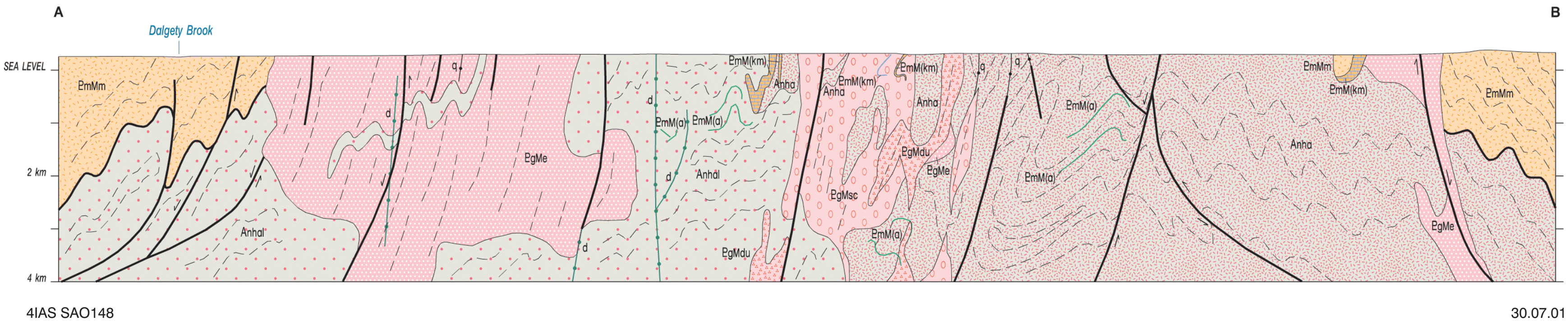


Figure 22. Diagrammatic cross section (see Fig. 21 for location) from the GLENBURGH 1:100 000-scale geological map (Occhipinti and Sheppard, 2001), showing the faulted boundaries between the Halfway Gneiss and metasedimentary rocks of the Moogie Metamorphics. The Halfway Gneiss is interpreted to be folded into an antiform, with granites of the Moorarie Supersuite typically intruded into this structure. Ᵽgmsc, Ᵽgmdu, Ᵽgme — granite, Moorarie Supersuite; Ᵽmma, Ᵽmmm, Ᵽmmkm — metasedimentary rocks and minor amphibolite of the Moogie Metamorphics; Anhal — Palaeoproterozoic to Archaean Halfway Gneiss

Siliciclastic sedimentary rocks of the Mount James Formation were deposited in a series of small fault-bounded basins on top of the Glenburgh Terrane, the Camel Hills Metamorphics, the granites of the Moorarie Supersuite, and the northwestern edge of the Yilgarn Craton at c. 1800 Ma. These sedimentary rocks were probably deposited during the latter stages of the Capricorn Orogeny (Occhipinti et al., 1999a). During the Mesoproterozoic, sedimentary rocks of the Edmund Group (Bangemall Supergroup) were deposited on the Gascoyne Complex, Yilgarn Craton, Mount James Formation, and the Bryah and Padbury Basins. The Edmund Group was intruded by dolerite sills at 1465 Ma and 1070 Ma (Wingate, M., 1999, 2000, written comm.; Nelson, in press), and then deformed during the Edmundian Orogeny between c. 1070 and c. 750 Ma (Sheppard and Occhipinti, 2000; Wingate and Giddings, 2000). Upper Carboniferous to Lower Permian glacigene rocks of the Carnarvon Basin were deposited on top of all other tectonic units and locally folded and faulted into northerly trending structures.

Excursion localities — Dalgaringa Supersuite

The Dalgaringa Supersuite consists of massive, foliated, and gneissic granites dated at 2005–1970 Ma (Sheppard et al., 1999b), and forms extensive outcrops throughout the southernmost part of the Glenburgh Terrane, south of the Dalgety Fault (Figs 19 and 20). The supersuite comprises two episodes of magmatism that are separated by a deformation and high-grade regional metamorphic event (Table 2). The two magmatic episodes consist of 2005–1985 Ma foliated to gneissic quartz diorite, tonalite, granodiorite, and monzogranite, and c. 1975 Ma tonalite and granodiorite of the Nardoo Granite. After the deformation and regional metamorphism event, sheets of foliated leucocratic monzogranite intruded the early foliated to gneissic granites.

The Dalgaringa Supersuite is dominated by diorite, tonalite, and granodiorite, in contrast to most Palaeoproterozoic batholiths of northern Australia, which largely consist of monzogranite and granodiorite (Wyborn et al., 1992). Sheppard et al. (1999b) noted that the Nardoo Granite has a composition similar to Phanerozoic subduction-related granites, and suggested that the supersuite may have formed in an Andean-type setting along the margin of an Upper Archaean to Palaeoproterozoic continent or microcontinent, which subsequently collided with the passive margin of the Yilgarn Craton.

Day 5

Variably deformed and metamorphosed granitic components of the Palaeoproterozoic Dalgaringa Supersuite of the Glenburgh Terrane will be seen today. In particular, we will visit some geochronological sampling sites where the heterogeneous nature of the supersuite is well illustrated. In the afternoon we will drive into the northern part of the Glenburgh Terrane.

Locality 12: Nardoo Granite (MGA 465582E 7200762N)

From Locality 11 follow Erong Road for 27 km to Locality 12.

At this locality (Fig. 19), representative rock types of the Nardoo Granite are exposed as tors and low whalebacks on both sides of Erong Road. The Nardoo Granite consists of two intrusive phases: medium-grained, mesocratic, even-textured or porphyritic biotite tonalite, and a lighter coloured, medium-grained, weakly porphyritic biotite tonalite and granodiorite. Contacts between the two phases are commonly sharp, with the lighter coloured phase consistently intruding the mesocratic tonalite; however, the two also locally grade into each other. SHRIMP U–Pb zircon dating of both phases indicates that they are essentially coeval. A sample of mesocratic tonalite (GSWA

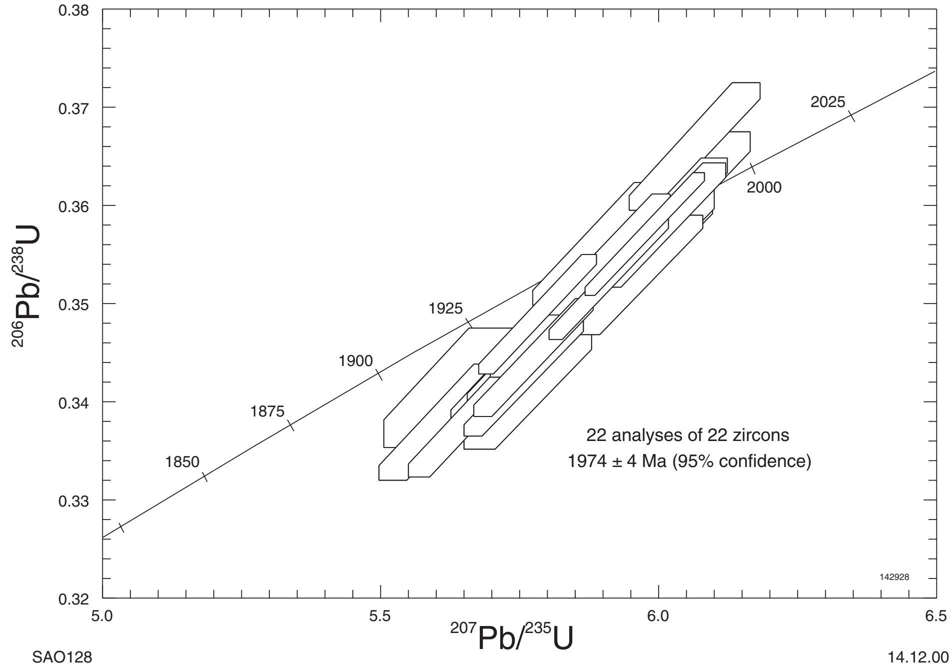


Figure 23. Concordia plot for porphyritic biotite tonalite of the Nardoo Granite (MGA 445000E 7186400N; GSWA 142928; Nelson, 1999)

142932) has an igneous crystallization age of 1977 ± 4 Ma, whereas a sample of lighter coloured granodiorite (GSWA 142928) was dated at 1974 ± 4 Ma (Nelson, 1999; Fig. 23).

Mesocratic porphyritic biotite tonalite and leucocratic biotite granodiorite to tonalite are both present at this locality. Both phases contain inclusions of fine-grained biotite tonalite, psammite, and round, green mafic clots, 5–15 mm in diameter, composed of biotite and chlorite. Inclusions of tonalite and the mafic clots are much more abundant in the mescocratic phase. Locally, the granite contains a weak igneous flow banding that may be attenuated by deformation; however, more commonly, an easterly trending tectonic foliation is oblique to the igneous banding. The foliation is cut by narrow, northerly trending and vertically dipping shear zones.

Locality 13: Foliated and gneissic granites of the Dalgaringa Supersuite (MGA 443424E 7200086N)

From Locality 12 travel north for 5.7 km along Erong Road. Turn left at the T-junction onto the Dalgety Downs – Landor road. Travel for 18.2 km to a southerly trending station track (at MGA 449309E 7207604N) about 20 m after a grid. Turn left and travel south along the track. About 6.2 km down the track is a Y-junction; take the left fork. From here travel another 2.8 km. Turn right, off the track (at MGA 445710E 7199896N). Follow an old station track to the west-northwest for 1.1 km, and then turn left onto indistinct wheel tracks (at MGA 444880E 7200459N). Follow the wheel tracks onto a west–southwesterly trending dolerite dyke for 1.6 km. Locality 13 is in the rocky outcrops to the north of the flat grassy area.

This locality (Figs 19 and 20) is a low-strain zone in which igneous contact relationships in granites of the Dalgaringa Supersuite are preserved. Two granite samples from this locality were dated using SHRIMP U–Pb zircon geochronology.

Foliated and gneissic granites, 2005–1985 Ma in age, outcrop over a wide area on the southern part of GLENBURGH*. They form the older component of the Dalgaringa Supersuite and are intruded by the c. 1975 Ma Nardoo Granite. The rocks range from strongly deformed and completely recrystallized foliated and gneissic granite in zones of high strain to statically recrystallized granites with intrusive relationships in areas of low strain. All the rocks have been metamorphosed at medium to high grade. In zones of moderate to high strain, the rocks are pegmatite banded and strongly resemble Archaean mesocratic granitic gneiss of the Narryer Terrane (Occhipinti et al., 1998; Sheppard and Swager, 1999). However, SHRIMP U–Pb zircon dating here and elsewhere in the Glenburgh Terrane demonstrates that these rocks are Palaeoproterozoic in age and therefore do not represent reworked Narryer Terrane (Nutman and Kinny, 1994; Sheppard et al., 1999b; Sheppard and Occhipinti, 2000; Occhipinti and Sheppard, 2001; Fig. 24).

The high-strain zones at this locality are related to the Glenburgh Orogeny. Pinch and swell structures parallel to the main foliation are well developed, and small box-folds deform the foliation (Fig. 25a). The rocks are also cut by narrow, easterly to east-northeasterly trending shear zones, which are probably related to the Capricorn Orogeny.

At this locality net-veining between several different types of granite of the Dalgaringa Supersuite implies coeval intrusion of intermediate and acid magmas (Fig. 25b). A medium-grained, variably porphyritic biotite monzogranite, net-veins fine-grained tonalite, monzogranite, and mesocratic granodiorite. The latter three rock types form lobate inclusions or ‘pillows’ enclosed by veins of the medium-grained monzogranite. Locally, the fine-grained biotite monzogranite also veins the fine-grained tonalite. Inclusions of the tonalite are sometimes present in the mafic granodiorite. A sample of fine-grained tonalite (GSWA 142926; Fig. 24a) from this locality has an igneous crystallization age of 2002 ± 2 Ma (Nelson, 1999). A sample of fine-grained biotite monzogranite (GSWA 142927; Fig. 24b), also from this locality, gave a SHRIMP U–Pb zircon age of 1999 ± 5 Ma, which is interpreted as the age of igneous crystallization (Nelson, 1999).

Locality 14: Foliated granites of the Dalgaringa Supersuite (MGA 441548E 7202202N)

From Locality 13 retrace your tracks for 300 m then turn left off the dolerite dyke (at MGA 443789E 7200047N) and travel in a northerly direction, keeping the outcrop to the left. Head towards the east side of a low quartz-vein hill, and then continue a few hundred metres to an easterly trending station track. Turn left and drive for 2.2 km. At this point (MGA 441585E 7201676N) turn off the track to the right (north) and head towards the base of prominent rocky outcrops (at MGA 441453E 7202135N). Park and walk up the hill to Locality 14.

At this locality (Figs 19 and 20) igneous contact relationships in the foliated and gneissic granites of the Dalgaringa Supersuite are preserved on pavements. In contrast to Locality 13, the magmas here have intruded granite that may not have been much older, but had already solidified.

This locality contains many of the granite types in the Dalgaringa Supersuite. A dark-grey, fine-grained biotite tonalite is intruded by a variably porphyritic mafic granodiorite. The porphyritic mafic granodiorite has intruded along a pre-existing tectonic fabric in the tonalite, and peeled off pieces of the tonalite (Fig. 26). The tonalite forms thin or slabby inclusions in the mafic granodiorite. The tonalite inclusions are

* Capitalized names refer to standard 1:100 000 map sheets, unless otherwise indicated.

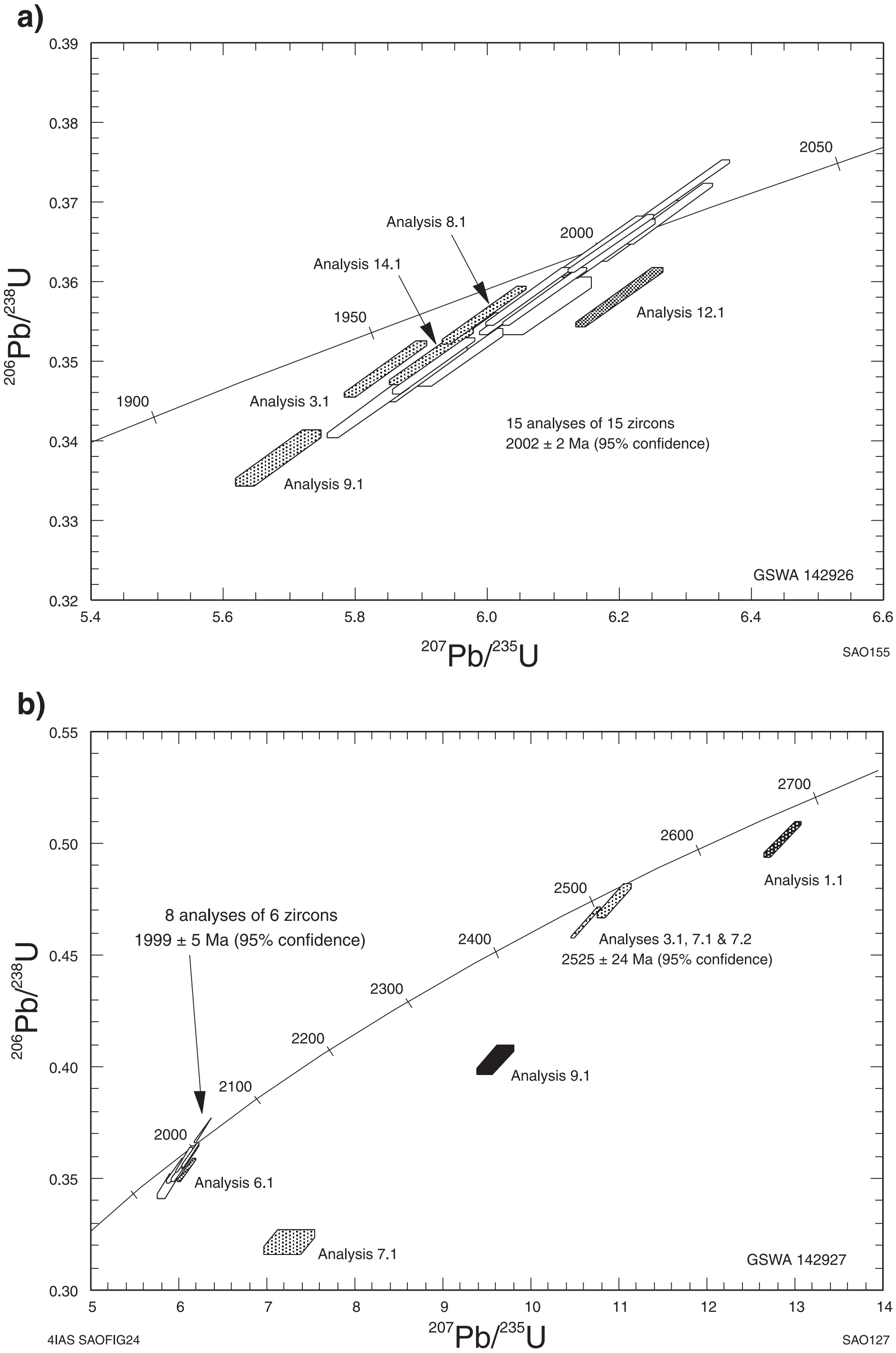
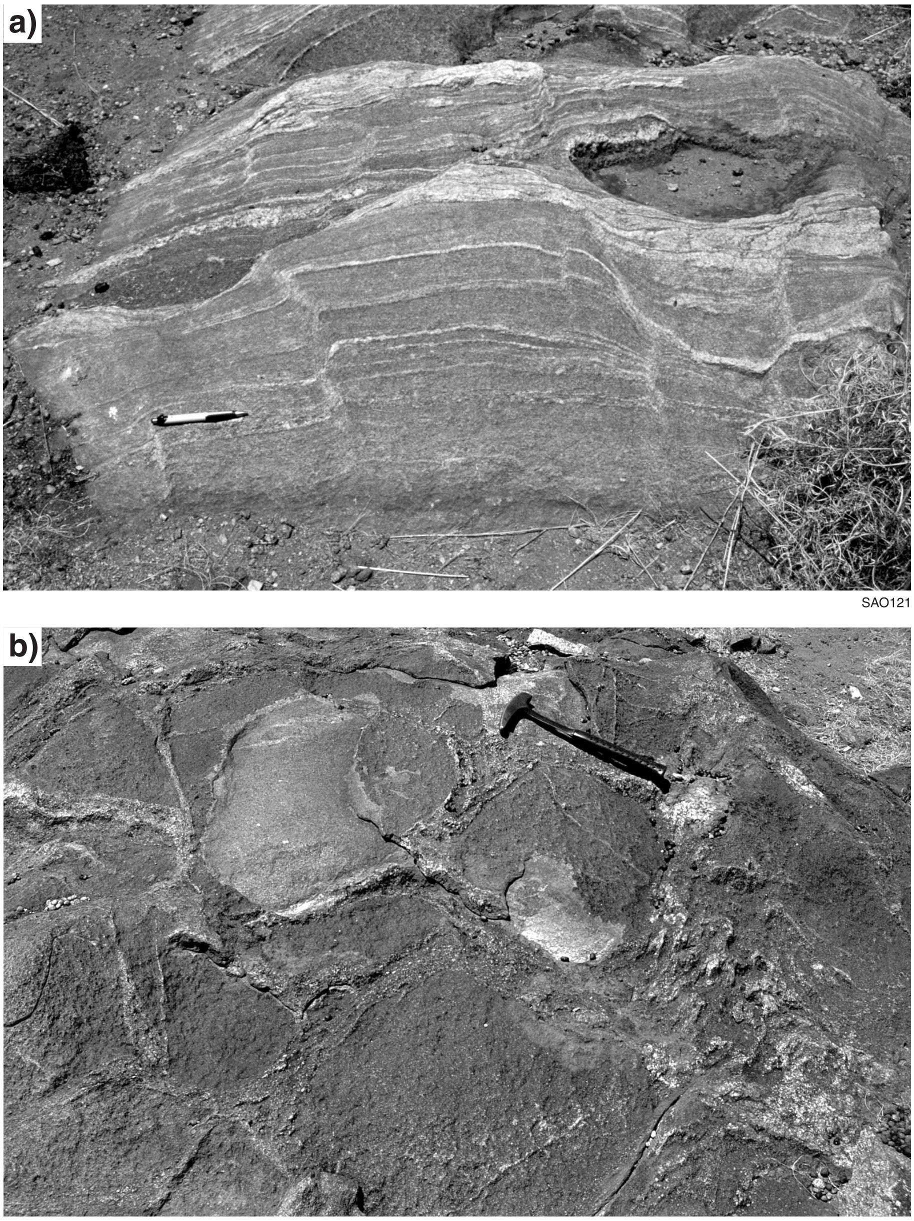


Figure 24. Concordia plots for c. 2000 Ma granite from the Dalgaringa Supersuite: a) foliated biotite tonalite (from MGA 443424E 7200086N; GSWA 142926) b) foliated fine-grained biotite–oligoclase granodiorite (GSWA 142927) from 100 m east of GSWA 142926 sample site

homogeneous and typically angular suggesting that the granodiorite intruded rigid, solidified tonalite (Fig. 26). Locally, a well-foliated, sparsely porphyritic biotite monzogranite is also cut by the mafic granodiorite. At this locality the mafic granodiorite both grades into, and is veined by, a variably porphyritic felsic granodiorite.



SAO121

4IAS SAOFIG25 SAO115

Figure 25. Outcrops of tonalite, granodiorite, and monzogranite from the Dalgaringa Supersuite at Locality 13: a) in the lower half of the photograph a banded granitic gneiss is deformed into box-folds. In the upper part pinch-and-swell structures are present, i.e. the weathered-out pillow-like depressions in the photo are wrapped by the foliation, which pinches in between them; b) pillows and liquid–liquid contacts between tonalite and monzogranite (net-veining). Around this area the pillows are locally flattened

An L-tectonite fabric, which formed during the Glenburgh Orogeny, is well developed in the granodiorite at this locality. However, on a regional scale this fabric is typically a foliation rather than an L-tectonite. Younger deformation features related to the Capricorn Orogeny include:

- 100–110°-striking, narrow quartz–epidote-filled brittle–ductile fractures;
- small-scale 150°-trending, narrow ductile shear zones indicating strike-slip movement.

Figure 26. Thin inclusions of tonalite in a mafic granodiorite at Locality 14. This texture has developed because the porphyritic mafic granodiorite intruded rigid, solidified tonalite along a pre-existing tectonic fabric

Locality 15: Leucocratic monzogranite of the Dalgaringa Supersuite (MGA 409408E 7183770N)

From Locality 14 go back to the easterly trending station track. Turn right and continue past Challenger Well for 6.6 km. Go through a gate in a northerly trending fenceline. Travel for 5.1 km along the track to a Y-junction. Take the right-hand fork for 1.3 km to Mollies Well. Go through a gate in the easterly trending fenceline and continue for 700 m to the Dalgety Downs – Landor road. Turn left and travel for 31.2 km to the T-junction with the Carnarvon–Mullewa road. Turn left at the junction towards Mullewa and travel for 3 km. Park on the right hand side of road and walk about 200 m west to Locality 15.

In the southwestern part of the Glenburgh Terrane and west of the Deadman Fault is a granite pluton of at least 35 km^2. It comprises leucocratic, medium-grained, weakly porphyritic biotite monzogranite, locally with abundant biotite clots after garnet. The

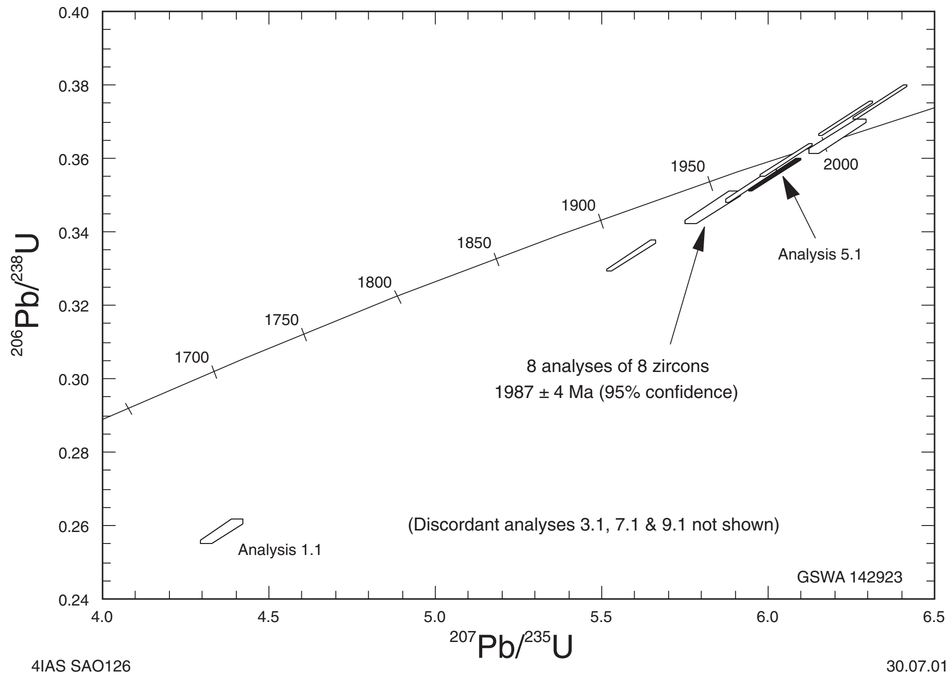


Figure 27. Concordia plot for leucocratic monzogranite from the Dalgaringa Supersuite (at MGA $_4$09408E $_{71}$83770N; GSWA 142923; Nelson, 1999)

pluton consists of a series of granite sheets 10–100 m thick, which contain abundant screens and rafts of interlayered fine-grained mesocratic tonalite and fine-grained granite, up to 2 km long. At and around Locality 15 (Figs 19 and 20), the leucocratic monzogranite is dated at 1987 ± 4 Ma (GSWA 142923; Nutman and Kinny, 1994; Nelson, 1999; Fig. 27), and forms easterly trending sheets into c. 2000 Ma tonalite and mafic granodiorite. Locally, the fine-grained tonalite is folded into easterly trending folds; the monzogranite intrudes subparallel to the fold axes in the tonalites but is not folded, indicating that the maximum age of the F_{2g} folds is c. 1987 Ma at this locality (Table 2). The monzogranite locally contains inclusions of mafic granodiorite and fine-grained biotite tonalite.

Other 1989–1987 Ma granites in the southern part of the Glenburgh Terrane exhibit similar relationships to the D_{2g} folds. For example, a quartz diorite dated at c. 1989 ± 3 Ma (Nelson, 1999) is interpreted to have intruded into ambient granulite-facies conditions (Occhipinti and Sheppard, 2001).

Excursion localities — Halfway Gneiss

Day 6

Today we will look at the latest Archaean to Palaeoproterozoic granitic gneiss of the Halfway Gneiss and younger granite that intrudes it.

The Halfway Gneiss outcrops in a 6–10 km-wide easterly trending belt in the northern part of the Glenburgh Terrane (Figs 19, 21, 22, and 28), and as rafts within granite of the Carrandibby Inlier (Figs 2 and 3). The gneiss consists of augen gneiss, banded granitic gneiss, and well-foliated leucocratic or mesocratic granite that is locally pegmatite banded (Fig. 28). The different rock types are interleaved on both mesoscopic and megascopic scales and contacts between them are typically tectonic, although, in

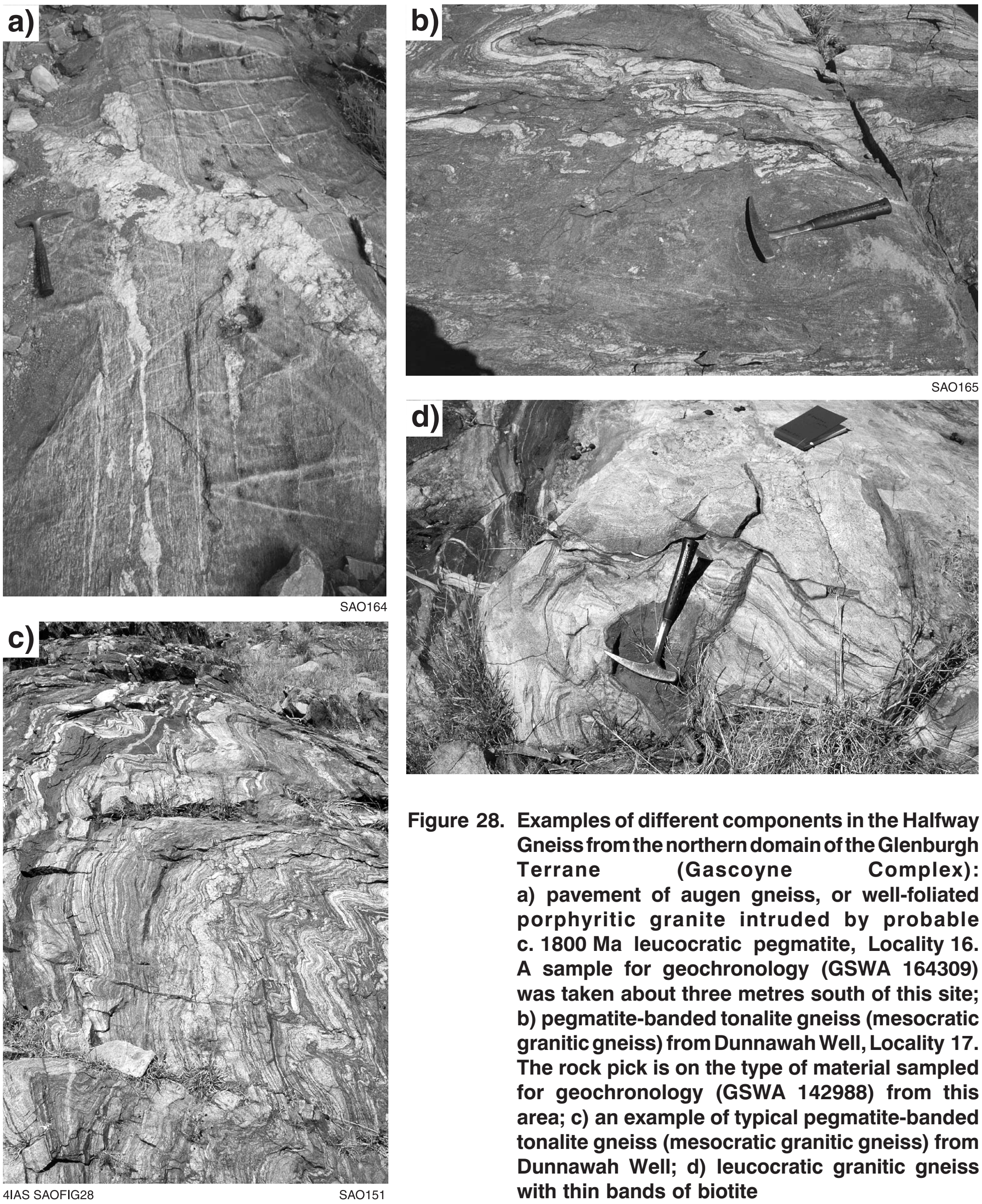


Figure 28. Examples of different components in the Halfway Gneiss from the northern domain of the Glenburgh Terrane (Gascoyne Complex): a) pavement of augen gneiss, or well-foliated porphyritic granite intruded by probable c. 1800 Ma leucocratic pegmatite, Locality 16. A sample for geochronology (GSWA 164309) was taken about three metres south of this site; b) pegmatite-banded tonalite gneiss (mesocratic granitic gneiss) from Dunnawah Well, Locality 17. The rock pick is on the type of material sampled for geochronology (GSWA 142988) from this area; c) an example of typical pegmatite-banded tonalite gneiss (mesocratic granitic gneiss) from Dunnawah Well; d) leucocratic granitic gneiss with thin bands of biotite

places, igneous intrusive relationships are preserved. The Halfway Gneiss has been heterogeneously deformed and metamorphosed to at least amphibolite facies. Despite the metamorphism, the original igneous components can be recognized in areas of low strain.

Geochronology by Nelson (2000, in press; GSWA 168947; GSWA 164309; GSWA 142988; GSWA 168950; Fig. 29) has demonstrated that the gneiss has both latest Archaean and Palaeoproterozoic granitic components. Samples from Localities 16a and 17 have igneous crystallization ages, determined by SHRIMP U–Pb zircon analyses, of 2544 ± 5 and 2550 ± 7 Ma respectively (Figs 29a,c). Nelson (in press) analysed a pegmatite-banded tonalite gneiss (GSWA 168950; Fig. 29c) in the Carandibby Inlier (Fig. 2), which Nutman and Kinny (1994) had previously dated at c. 2500 Ma, and found four main zircon populations at 2452 ± 9, 2473 ± 6, 2506 ± 6, and 2519 ± 3 Ma (Fig. 29c). These ages are interpreted to be from zones (commonly cores, but not exclusively) within zircon grains that formed at the time of igneous crystallization of the granite protolith phases identified within the gneiss.

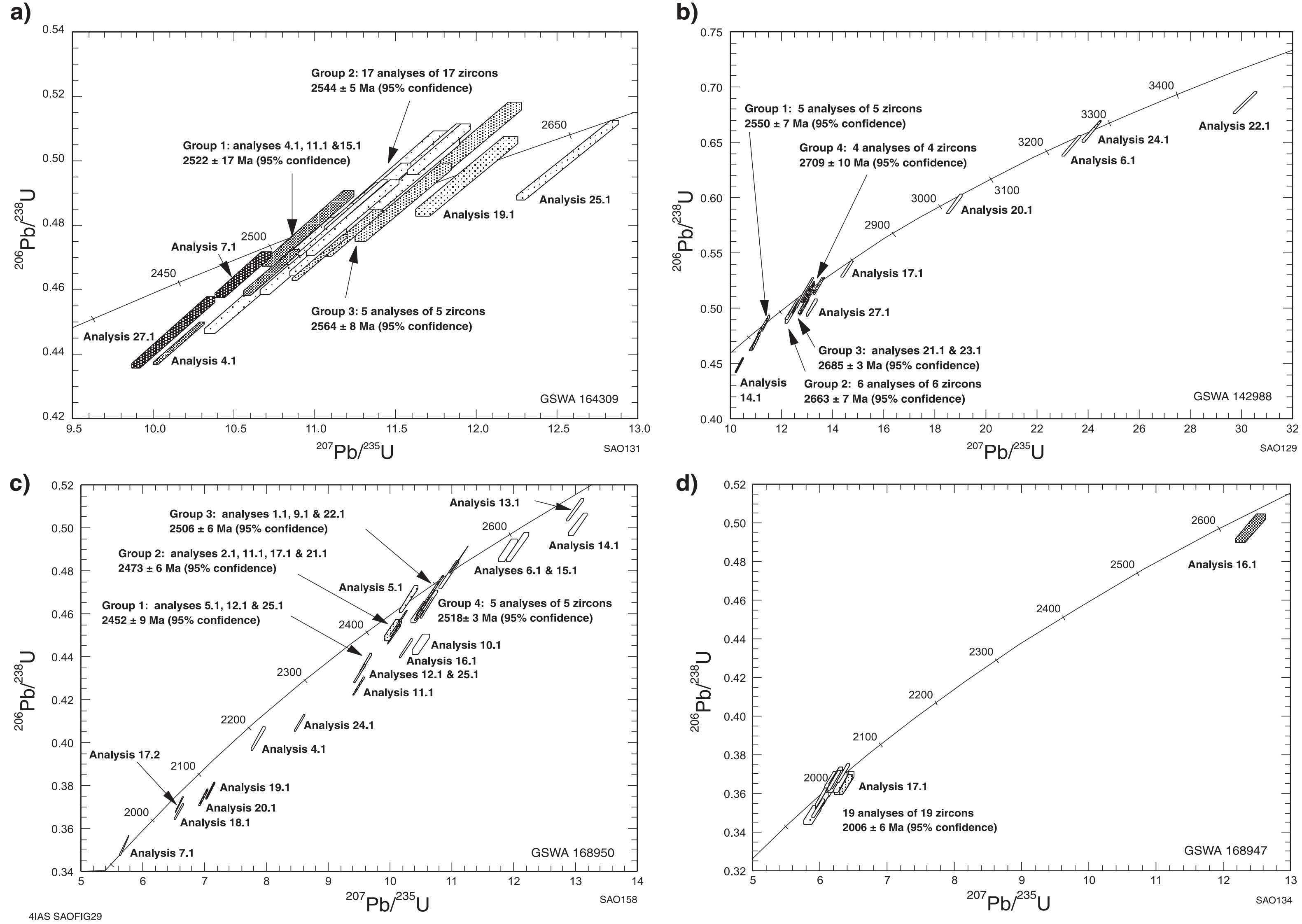


Figure 29. Concordia plots for components of the Halfway Gneiss: a) biotite augen gneiss from Locality 16a (GSWA 164309; MGA 413616E 7225303N); b) mesocratic, banded granitic gneiss from Locality 17 (GSWA 142988; MGA 401309E 7223430N); c) pegmatite-banded tonalite gneiss (GSWA 168950; MGA 362200E 7165700N) from the Carrandibby Inlier; d) leucocratic granitic gneiss from 2 km west of Weedarra Homestead (GSWA 168947; MGA 381950E 7230840N)

In addition to latest Archaean granite protoliths, Nelson (in press) also identified Palaeoproterozoic protoliths in a sample of leucocratic granitic gneiss (GSWA 168947) about 2 km west of Weedarra Homestead (MGA 381950E 7230840N). A single population of 19 concordant to slightly discordant analyses on 19 zircons defines an igneous crystallization age of 2006 ± 6 Ma (Fig. 29d; Nelson, in press). This age is within error of igneous crystallization ages for gneissic and foliated granites of the Dalgaringa Supersuite, and suggests that some granite of the Dalgaringa Supersuite has been included in the Halfway Gneiss. The original relationship of the c. 2006 Ma component to the c. 2500 Ma components of the Halfway Gneiss is unknown; however, they now appear to be tectonically interleaved.

The Halfway Gneiss is also tectonically interleaved with calc-silicate gneiss, amphibolite, actinolite schist, tremolite schist, and pelitic schist of the Moogie Metamorphics. The original relationship between these supracrustal rocks and the Halfway Gneiss is unknown, although regional structural observations suggest that at least some parts of the Halfway Gneiss are older.

The Halfway Gneiss is extensively intruded by sheets and dykes of the Dumbie Granodiorite, by plutons, dykes, and veins of the Scrubber Granite, and by medium-grained biotite(–muscovite) granite, all of which are part of the 1830 to 1780 Ma Moorarie Supersuite.

All of the constituent rock types of the Halfway Gneiss display a variety of textures reflecting different strain states and overprinting by lower grade metamorphic events. In thin section, most samples show evidence of extensive static recrystallization at low metamorphic grade. Evidence includes recrystallization of quartz to fine polygonal aggregates, replacement of plagioclase by albite–oligoclase, sericite, and epidote, widespread micrographic and myrmekitic textures, and replacement of magnetite and ilmenite by epidote and titanite respectively, in association with plagioclase alteration.

Locality 16: Augen gneiss (Halfway Gneiss) and the Dumbie Granodiorite

From Locality 15 travel back towards the T-junction and continue to the west for 30.8 km along the Carnarvon–Mullewa road to the intersection of the Dairy Creek – Cobra road. Turn right (to the north) and travel along the Dairy Creek – Cobra road for 34.9 km to the turnoff to Mooloo Downs Homestead. Turn right and travel along the track towards the homestead. At the airstrip, about 100 m before the homestead, turn right and travel east along a station track for 22.9 km. At this point there is an old fence and a well to the right. Turn right and drive past the eastern side of the well to an old station track. About 2 km south of the well, turn right, ignoring the track to the south. At 5.1 km from the well, go through a steep creek and at the following Y-junction take the left track. Travel for a further 1.1 km to another Y-junction and take the left track. About 700 m further on turn left off the track (at MGA 413417E 7226113N), and travel south to an easterly trending creek. Go through the creek and park. Walk south past Locality 16b and on to Locality 16a (at MGA 413630E 7225305N).

At this locality (Figs 19 and 21), the Halfway Gneiss (Locality 16a) and a sheet of the Dumbie Granodiorite of the Moorarie Supersuite (Locality 16b) have been dated by SHRIMP U–Pb zircon analysis (Fig. 21).

Locality 16a: c. 2540 Ma augen gneiss of the Halfway Gneiss (MGA 413630E 7225305N)

Augen gneiss, or well-foliated porphyritic biotite granite of the Halfway Gneiss, outcrops in, and just south of, a creek on the northern side of a low rubbly hill. This augen gneiss appears to be the simplest, most homogeneous component of the Halfway

Gneiss mapped to date. A SHRIMP U–Pb zircon date of 2544 ± 5 Ma (Fig. 29a) obtained for the gneiss at this locality is interpreted as the igneous crystallization age of the biotite granodiorite precursor (Nelson, 2000). Five zircons defining an older population at 2563 ± 8 Ma probably represent xenocrysts. The augen gneiss is folded into tight, upright, gently plunging folds and is intruded by both pegmatite veins (Fig. 28a) and the Dumbie Granodiorite (Locality 16b). The pegmatite veins are interpreted to be associated with c. 1800 Ma granite in the region, either being intruded subparallel to F_{1n} folds or oblique to them. Thus, the veins are locally boudinaged or folded.

The c. 2540 Ma age of the Halfway Gneiss is the first clear indication that there are Archaean rocks present in the southern Gascoyne Complex. However, it also distinguished the Glenburgh Terrane from the Yilgarn Craton, as rocks of this Archaean age have not been dated in the northern part of the Yilgarn Craton.

Walk back to Locality 16b.

Locality 16b: Dumbie Granodiorite (MGA 413649E 7225430N)

In the northeastern part of the Glenburgh Terrane, the Dumbie Granodiorite forms a large pluton apparently comprising numerous easterly trending sheets (Figs 19, 20, and 30). The actual extent of this pluton is unknown as it has yet to be mapped out to the north. The Dumbie Granodiorite outcrops as boulders, tors, and scattered whalebacks. At this locality, the unit consists of porphyritic fine- to medium-grained granodiorite (Fig. 30a) with tabular phenocrysts of sanidine mostly up to 1 cm long, whereas elsewhere the unit may consist of tabular or rounded microcline phenocrysts up to 3 cm long in a medium-grained groundmass of plagioclase, quartz, and biotite.

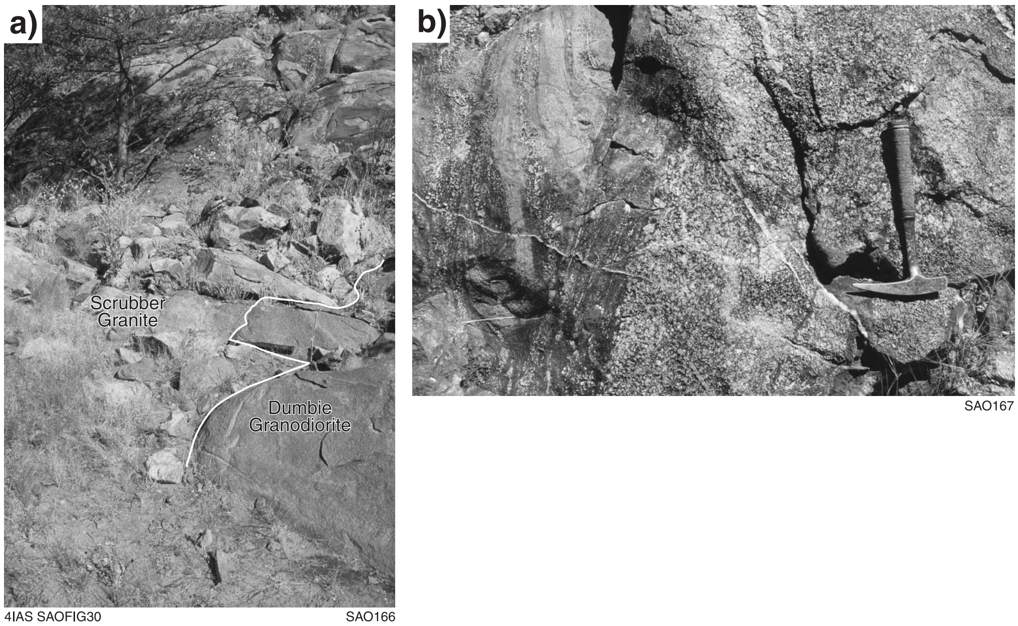


Figure 30. Different phases of the Dumbie Granodiorite: a) contact between fine-grained porphyritic biotite granodiorite (Dumbie Granodiorite) and leucocratic, medium to even-grained biotite monzogranite (Scrubber Granite) at Locality 16b on the GLENBURGH 1:100 000-scale geological map (MGA 413649E 7225430N). In the foreground on the left hand-side is a dyke of Scrubber Granite. The background mostly compises Scrubber Granite; b) medium-grained porphyritic biotite granodiorite cutting well-foliated and folded fine-grained biotite tonalite gneiss (possibly Halfway Gneiss) on the MOUNT PHILLIPS 1:250 000-scale geological map (MGA 453200E 7235300N)

The Dumbie Granodiorite intrudes both the Halfway Gneiss (Fig. 30b) and the Moogie Metamorphics, mostly as sheets and veins that largely trend subparallel to the fold-axial surfaces of F_{1n} folds. It is intruded by veins and dykes of medium-grained, even-textured biotite(–muscovite–tourmaline) granite (including the c. 1796 Ma Scrubber Granite; Occhipinti and Sheppard, 2001), coarse-grained granite, and pegmatite, and aplite. The Dumbie Granodiorite is either foliated, contains an L-tectonite fabric, or is locally massive.

SHRIMP U–Pb zircon ages have been obtained from the Dumbie Granodiorite at two localities (Figs 31a,b). A sample from the Dumbie Granodiorite at Locality 16b (GSWA 159995) gave an igneous crystallization age of 1811 ± 6 Ma (Nelson, 2000; Figs 30a and 31a). To the northeast on the MOUNT PHILLIPS 1:250 000 map sheet (MGA 453200E 7235300N; Nelson, 2000), GSWA 159987, which is a medium-grained porphyritic biotite granodiorite, contains three zircon populations. A date of 1810 ± 9 Ma on seven analyses of seven zircons was interpreted as the igneous crystallization age of the granodiorite (Fig. 31b). Two populations consisting of six analyses of six zircons dated at 2430 ± 11 Ma, and nine analyses of nine zircons dated at 2113 ± 8 Ma, are interpreted as xenocrysts.

The most abundant rock type in the Dumbie Granodiorite is a grey, fine- to medium-grained granodiorite (or less commonly monzogranite) with up to 30% feldspar phenocrysts. Some of the phenocrysts show minor development of microperthite. The groundmass consists of anhedral, granular, fine-grained plagioclase, quartz, green-brown biotite, and minor microcline. Accessory minerals comprise magnetite, ilmenite, allanite, zircon, and apatite. Locally, the Dumbie Granodiorite contains small clusters of fine magnetite crystals. Brown prismatic crystals of hydrated allanite, up to 1.5 mm long, are a prominent accessory mineral.

The rocks show evidence of static or dynamic recrystallization in the greenschist facies, namely: plagioclase is extensively replaced by albite–oligoclase, epidote, and sericite; quartz consists of fine-grained polygonal aggregates; and ilmenite and magnetite are rimmed or replaced by titanite and epidote, respectively.

Locality 17: Pegmatite-banded gneiss of the Halfway Gneiss (MGA 401309E 7223430N)

From Locality 16 drive back past Mooloo Downs Homestead to the Cobra – Dairy Creek road. Turn left at the intersection and drive south for 6.1 km to a left-hand bend before a creek crossing. Turn off the road (at MGA 396552E 7224551N) and travel east along an initially indistinct station track for 5.5 km to Dunnawah Well. Park, and walk to the west along the southern side of creek for 100 m to a low pavement (Locality 17).

The creek pavement at this locality (Figs 19 and 21) illustrates some of the complexity in the Halfway Gneiss. The main component of this gneiss has been dated as latest Archaean in age.

The gneiss comprises two main phases, both of which contain numerous thin pegmatite bands. The gneiss, including the thin pegmatite bands, is tightly folded (Figs 28b,c), but there may also be small refolds of the pegmatite parallel to the gneissic layering. The main component to the gneiss is a dark-grey, variably porphyritic, fine-grained biotite tonalite. Rounded plagioclase phenocrysts comprise up to 10% of the rock. The other phase consists of layers of pale-grey, fine- to medium-grained biotite granodiorite and quartz-rich tonalite.

Nelson (2000) dated the dark-grey tonalitic component of the gneiss (GSWA 142988) at this locality (Fig. 28b). The sample contains several concordant zircon populations

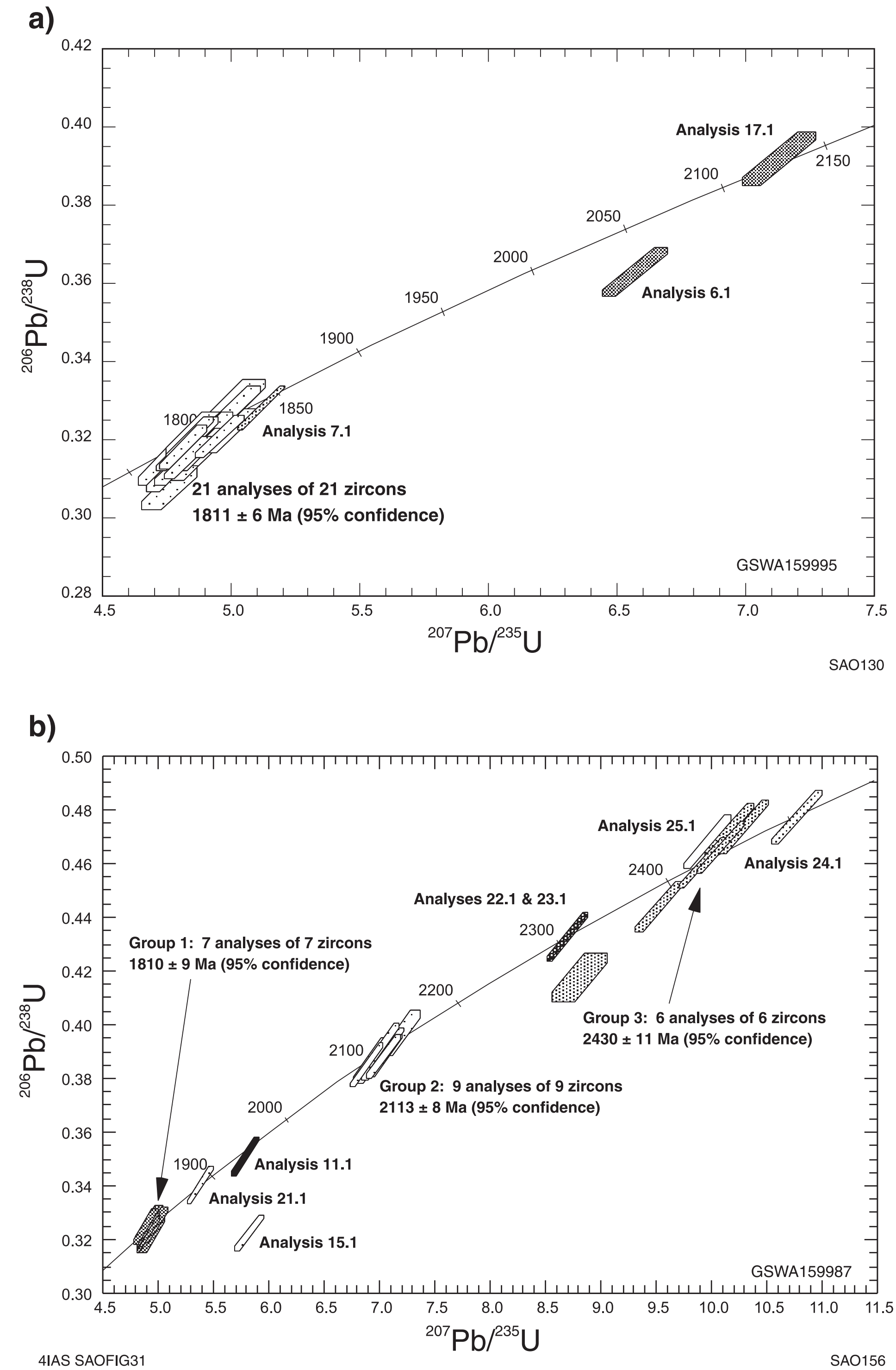


Figure 31. Concordia plots for the Dumbie Granodiorite: a) fine-grained porphyritic biotite granodiorite (Locality 16b; GSWA 159995; MGA 453200E 7235300N); b) well-foliated medium-grained porphyritic biotite granodiorite (GSWA 159987; MGA 453200E 7235300N)

(Fig. 29d). The youngest population consists of five analyses of five zircons with a pooled age of 2550 ± 7 Ma. Six analyses of six zircons define a date of 2663 ± 7 Ma, four analyses of a further four zircons define a date of 2709 ± 10 Ma, whereas the remaining zircons are c. 3300 Ma or older. Nelson (2000) suggested that the date of 2663 ± 7 Ma is the igneous crystallization age of the tonalite precursor, and that the 2550 ± 7 Ma date corresponds to the age of thin pegmatite veins in the sample. However, Occhipinti and Sheppard (2001) suggested that since pegmatite veins comprise less than 5% of the sample, the youngest population at 2550 ± 7 Ma probably represents the igneous crystallization age of the tonalite precursor (Figs 28b and 29b). This would suggest that the older zircons are xenocrystic populations.

Both the dark- and pale-grey phases of the granitic gneiss contain a grain-flattening fabric defined by quartz and feldspar, and some amoeboid textures, implying metamorphism at medium to high grade. This metamorphism and deformation also affects components of the Halfway Gneiss dated at 2006 ± 6 Ma (Nelson, in press), and therefore is related to the Glenburgh Orogeny. The gneiss is intruded by veins and dykes of foliated, leucocratic biotite pegmatite, which itself is cut by a medium-grained aplitic granite; the pegmatite and granite crosscut the gneissic layering and folds.

Narrow (<10 cm wide) shear zones marked by chlorite, epidote, and strongly flattened quartz crystals cut the gneiss at this locality. These shear zones also cut the granite and pegmatite veins that intrude across the gneissic layering. Some small, vertically plunging folds are possibly associated with the shear zones.

Moogie Metamorphics and Mount James Formation

The Moogie Metamorphics (Occhipinti and Sheppard, 2001) includes pelitic and psammitic schist (Mumba Pelite), quartzite, calc-silicate gneiss, marble, amphibolite, ultramafic schist, and metamorphosed BIF. Like the Camel Hills Metamorphics, these metasedimentary and metamorphosed mafic and ultramafic igneous rocks were previously included within the Morrissey Metamorphic Suite (Williams et al., 1983b; Williams, 1986). The Morrissey Metamorphic Suite was defined on the MOUNT PHILLIPS 1:250 000 sheet, and represented a group of metamorphosed and deformed Palaeoproterozoic sedimentary rocks thought to outcrop throughout the Gascoyne Complex (Williams et al., 1983c). These rocks were considered to be the metamorphosed and deformed equivalents of sedimentary rocks of the Wyloo Group to the north and the ‘Glengarry Group’, now the Yerrida and Bryah Groups (Pirajno et al., 1998), to the east and southeast.

Many of the components included in the Morrissey Metamorphic Suite by Williams (1986) are separated by large areas of granite and by major faults. In addition, it is probable that not all the metasedimentary rocks, or the metamorphosed mafic and ultramafic igneous rocks in the Gascoyne Complex, are the same age or were metamorphosed at the same time. Therefore, the metasedimentary and meta-igneous rocks in the Glenburgh Terrane have been grouped into the Moogie Metamorphics.

In the northwest Glenburgh Terrane, the Moogie Metamorphics are dominated by abundant quartzite derived from quartz sandstone, which forms large strike ridges. The quartzite is interlayered with calc-silicate gneiss and marble, and together they probably formed a coherent sedimentary package, with bedding traces preserved both internally and between the three rock types. The Mumba Pelite is also interlayered with calc-silicate gneiss and marble, suggesting that it was part of the same sedimentary package. All units have been multiply deformed and variably metamorphosed under prograde

?amphibolite and retrograde greenschist-facies conditions.

The Mumba Pelite forms a major component of the Moogie Metamorphics. In addition to comprising pelitic and psammitic schist (both which are commonly iron rich), it also locally contains minor quartzite and metamorphosed granular iron-formation. It is locally tectonically interleaved with amphibolite and ultramafic schist. The original relationship with these rocks is unknown.

The age of the Moogie Metamorphics is poorly constrained. In the southern part of the Glenburgh Terrane metasedimentary rocks are intruded by granites of the Dalgaringa Supersuite, and so must be older than c. 2005 Ma. However, the relationship between these metasedimentary rocks and the remainder of the Moogie Metamorphics (including the Mumba Pelite) is unknown. Rocks of the Moogie Metamorphics, including the Mumba Pelite are faulted against the Halfway Gneiss, so the relative age of the two units is uncertain. Early, originally subhorizontal layer-parallel folds deforming bedding within the Mumba Pelite and the quartzite and calc-silicate gneiss of the Moogie Metamorphics (in the northern part of the Glenburgh Terrane) also deform a well-developed gneissic layering in the Halfway Gneiss. Thus, if the two units initially developed in the same terrane, the Moogie Metamorphics must be younger than the Halfway Gneiss. However, if the two units initially developed in separate terranes and were juxtaposed by layer-parallel deformation (D_{2g} of the Glenburgh Orogeny; Table 2), then it is not possible to constrain their relative ages. Rocks of the Mumba Pelite are intruded by coarse-grained granite and pegmatite correlated with c. 1800 Ma granite of the Moorarie Supersuite. They are also unconformably overlain by the c. 1800 Ma Mount James Formation, which is not intruded by granite. Thus, they must be older than c. 1800 Ma.

The Mount James Formation comprises deformed and metamorphosed siliciclastic sedimentary rocks, which outcrop as several strips throughout the Gascoyne Complex. It typically outcrops along steeply dipping faults or high-strain zones. Different successions may be present in these strips, with either a meta-arkosic conglomerate (see **Locality 17b**) or a quartzite outcropping at the base (Sheppard and Occhipinti, 2000). The formation consists of quartzite, meta-arkosic sandstone to quartz–sericite phyllite, metaquartz pebble- to cobble-conglomerate, metapolymictic conglomerate, and quartz–chlorite–sericite phyllite. Locally, sedimentary structures such as cross-bedding or fining-up sequences are discernible. Original rounded to subrounded sand-sized quartz grains within quartzite and meta-arkosic sandstone of the Mount James Formation have locally been recrystallized to form almost polygonal quartz grains with seriate grain boundaries and undulose extinction. In many cases quartz grains are slightly elongate or ‘flattened’, defining a foliation in the rock. Minor detrital sphene or tourmaline is often present in the quartzite.

The nature and origin of the Mount James Formation (Drew, 1999a,b) is problematic. Hunter (1990) suggested that a correlation can be made between the Mount James Formation and parts of the Padbury Group in the Padbury Basin, and the Mount Minnie Group and Capricorn Formation in the Ashburton and Blair Basins, on the basis of lithological and structural relationships. Occhipinti et al. (1996) and Sheppard and Occhipinti (2000) suggested that the Mount James Formation should not be correlated with the Padbury Group, as the Padbury Group is more complexly deformed. It is also probable that the <2000 Ma Padbury Group is older than the Mount James Formation (see below). Additionally, the Padbury Group outcrops in the Padbury Basin, a distinctive tectonic unit that overlies the Bryah Basin, and is confined to the east of the Errabiddy Shear Zone (Occhipinti and Sheppard, 2001). The relationship of the Mount James Formation to the Bryah Basin is unknown. The adjacent fault zones are thought to have controlled initial sedimentation of the Mount James Formation (Hunter, 1990).

Nelson (in press) found that SHRIMP U–Pb ages on detrital zircons in one sample of quartzite (GSWA 168937) from the Mount James Formation contained two populations of zircons: one dated at c. 1960 Ma and the other dated at c. 1800 Ma. Thus, c. 1800 Ma provides a maximum age for deposition of the formation. The Mount James Formation is unconformably overlain by the lower part of the Mesoproterozoic Bangemall Supergroup (Nelson, 1995), providing a minimum age for the deposition of the Mount James Formation.

The Mount James Formation is tightly folded into easterly to northeasterly trending, gently inclined, moderately to steeply plunging folds. Some folds are discontinuous and asymmetric, possibly due to syn- to post-folding faulting. The Mount James Formation has been metamorphosed at sub- to low-greenschist facies conditions, thus its sedimentary protolith is easily discernible.

Excursion localities — Mumba Pelite and Mount James Formation

Day 7

This morning, the metasedimentary Moogie Metamorphics of the Glenburgh Terrane and the overlying metasedimentary rocks of the Mount James Formation will be our only excursion stop.

Locality 18: The Mumba Pelite of the Moogie Metamorphics, and the Mount James Formation

From Locality 17 return to the Cobra – Dairy Creek road. Turn left and drive south for 8 km. Turn left off the road (at MGA 393508E 7218424N). Travel east, crossing a creek (at MGA 393585E 7218251N), and a second creek (at 393617E 7218213N). Park about 50 m north (up hill) from the creek. From this point walk in a southeasterly direction over the hill towards Locality 18a.

At Locality 18 (Figs 19 and 32) the Palaeoproterozoic Mumba Pelite (Moogie Metamorphics) is folded into an easterly trending tight antiform and is unconformably overlain by metaconglomerate of the Mount James Formation, which is folded into an easterly trending tight syncline.

Locality 18a: Mumba Pelite of the Moogie Metamorphics (MGA 394192E 7217883N)

Here, the Mumba Pelite of the Moogie Metamorphics displays obvious compositional layering, defining original bedding (S_0; Fig. 33). A foliation, S_1, is well developed slightly oblique to S_0. Both these fabrics are folded into an easterly trending upright antiform that plunges moderately to steeply towards the west**.**

The outcrop largely comprises an iron-rich metaquartz sandstone that contains clots of opaques (magnetite and hematite) and quartz. Finer grained rocks around this locality largely comprise chloritoid–sericite–quartz schist. Chloritoid commonly forms unaligned sprays on the S_{2g} cleavage planes.

The first regional foliation in the Mumba Pelite is a subhorizontal or gently dipping foliation, S_{2g}, which formed subparallel to bedding (Table 2). This foliation is subparallel to the early faults in the Halfway Gneiss, and to the contact between the Halfway Gneiss and Moogie Metamorphics. Here, S_{2g} is locally subparallel or oblique

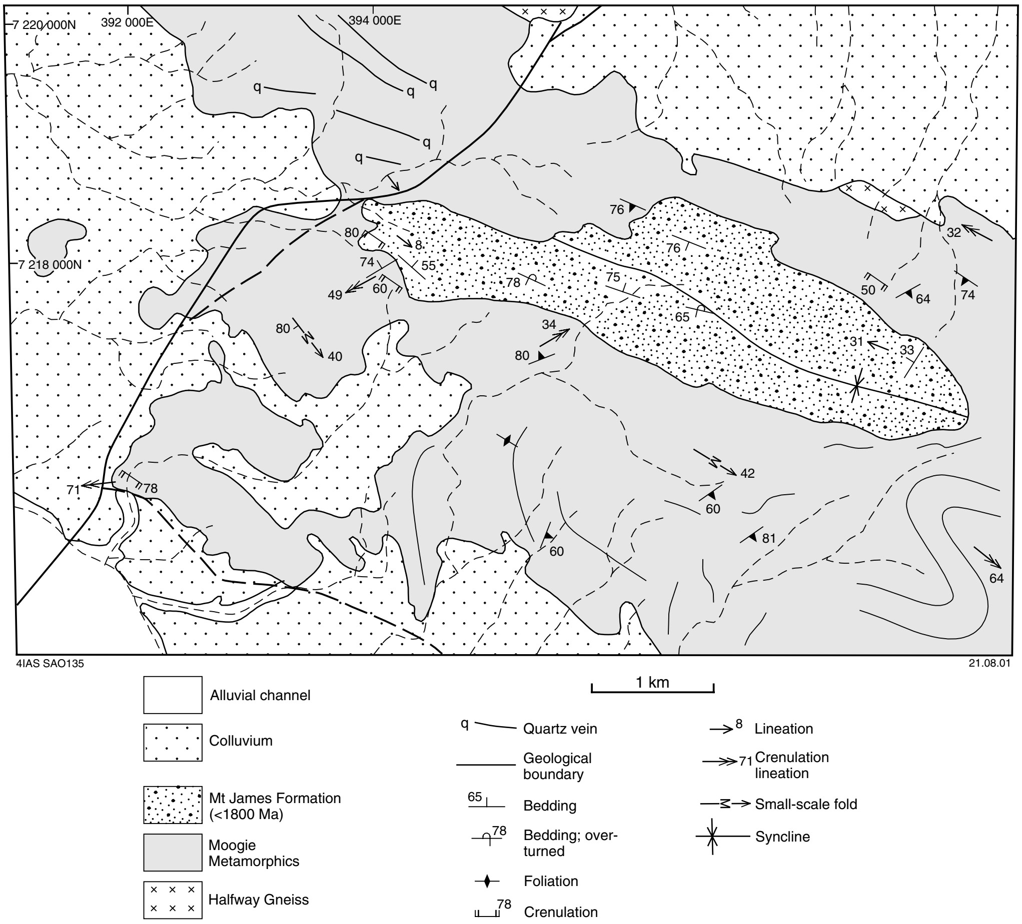
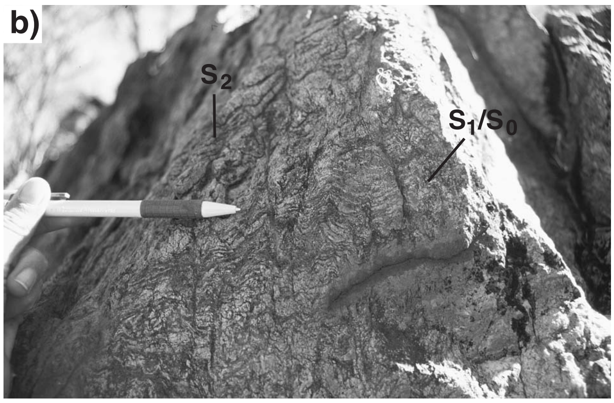


Figure 32. Simplified geological map of the area around Locality 18, showing the distribution of the Mount James Formation and the Moogie Metamorphics

4IAS SAOFIG33 SAO168



SAO120

Figure 33. Outcrops of the Mumba Pelite of the Moogie Metamorphics: a) compositional layering defining bedding at Locality 18a (MGA 394192E 7217883N). Dark elongate blotches (well-developed on the left-hand side) are magnetite–quartz aggregates that probably defined an original bedding, S_0, but have now been attenuated subparallel to S_0 into S_1; b) typical composite S_1/S_0 fabric folded into upright, gently plunging, easterly trending S_2 crenulations, east of Locality 18a

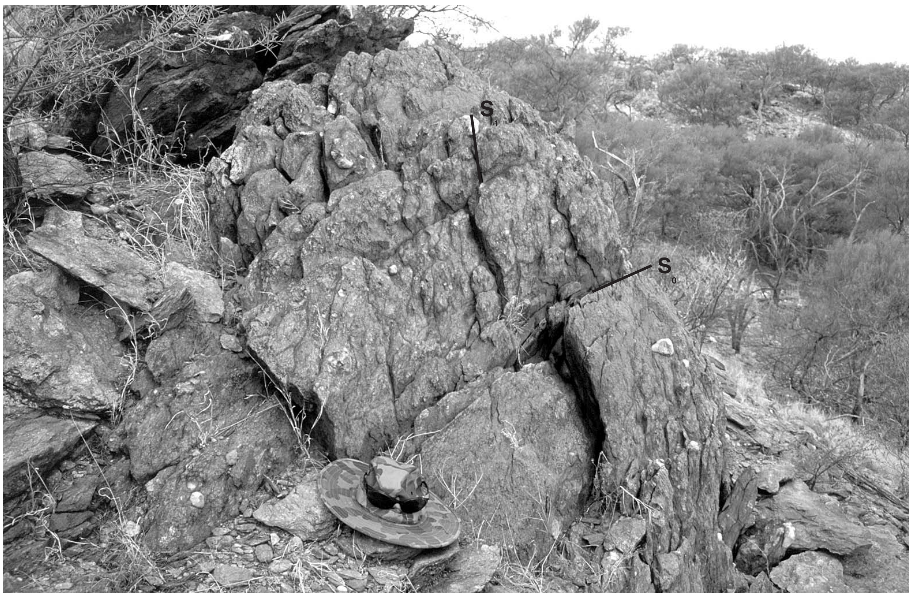


Figure 34. Bedding in moderately well foliated Mount James Formation metaconglomerate and pebbly sandstone at Locality 18b

to S_0, and is best developed in the finer grained, more pelitic layers. The S_{2g} fabric is parallel to the axial surface of subhorizontal folds of bedding locally observed in the Moogie Metamorphics northwest of Locality 17. There they deform quartzite, metapsammite, and calc-silicate, and locally a well-developed S_{1g} foliation in the Halfway Gneiss.

Dark elongate clots, which consist of quartz and magnetite aggregates, probably originally formed a bed that has been broken up (sheared and transposed) during subsequent deformation (D_{1n}). In addition, small, commonly <0.5 cm subrounded quartz–magnetite aggregates are interpreted as iron-rich sedimentary ‘concretions’. These ‘concretions’ now contain seriate grain boundaries due to deformation and are flattened parallel to the F_{1n} fold axis and metamorphic mineral lineation.

The original mineral assemblages developed in M_{2g} in the Mumba Pelite are typically completely overprinted by lower grade mineral assemblages. Locally, garnet, which developed during M_{2g}, forms an ‘annealed’ texture with quartz. In more pelitic components of the Mumba Pelite, mats of sericite may represent completely pseudomorphed sillimanite or ?staurolite, and chloritoid and chlorite have locally partially or completely pseudomorphed garnet. Chlorite may also have pseudomorphed biotite. This suggests that the Mumba Pelite was probably metamorphosed at medium grade (?amphibolite facies) during M_{2g}.

Regionally, the Mumba Pelite is intruded by coarse-grained biotite–muscovite granite and pegmatite, correlated with c. 1800 Ma granite, and unconformably overlain by the Mount James Formation.

Locality 18b: Conglomerate — Mount James Formation (MGA 394252E 7217936N)

At this locality a matrix-supported pebble to boulder conglomerate at the base of the Mount James Formation, which has been metamorphosed in the greenschist facies, unconformably overlies the Mumba Pelite of the Moogie Metamorphics. Although sedimentary structures such as bedding (Fig. 34) and fining-up sequences are present, the metaconglomerate is mostly massive.

The metaconglomerate consists of pebble- to boulder-sized, subrounded to subangular clasts of vein quartz and quartzite (90–95%), granite (including pegmatite), amphibolite, and metasedimentary rock (5–10%) in a matrix of arkosic sandstone. The sandstone consists of quartz, sericite, chlorite, and acicular opaque minerals comprising ?ilmenite or ?rutile. Quartz is the dominant component in the rock and typically forms sand-sized grains that have been largely recrystallized and contain seriate grain boundaries. Mats of sericite, in which relict feldspar may be preserved, make up about 20–30% of the rock. Chlorite, with lesser sericite, defines a well-developed foliation. The clasts have undergone plane strain or have been flattened with the long axes of the pebbles aligned subparallel to the foliation; however, in some places possible S–C fabrics and sheared quartz-pebble clasts indicate a dextral shear sense. There is typically a lack of asymmetry around most of the quartz pebbles, indicating that they have not undergone rotational strain.

The well-developed foliation in the metaconglomerate is parallel to the axial surface of a regional-scale, gently plunging, easterly trending tight syncline. The fold-axial surface of the syncline trends subparallel to the fold-axial surface of an older, coplanar, regional, steeply plunging tight antiform in the underlying Mumba Pelite.

Return to the Cobra – Dairy Creek road. Drive south to the T-junction with the Carnarvon–Mullewa road. Turn left onto the Carnarvon–Mullewa road and drive towards Perth.

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