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PROTEROZOIC GEOLOGY OF THE CAPRICORN OROGEN WESTERN AUSTRALIA — A FIELD GUIDE

by S. A. Occhipinti, S. Sheppard, I. M. Tyler,
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SGTSG
Kalbarri
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GEOLOGICAL SOCIETY OF AUSTRALIA
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Geological Survey of Western Australia

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Kalbarri, Western Australia, 2003

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GEOLOGICAL SURVEY OF WESTERN AUSTRALIA

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by
S. A. Occhipinti[1], S. Sheppard[2], I. M. Tyler[2], K. N. Sircombe[3], S. Reddy[1], D. Hollingsworth[1], D. McB. Martin[2], and A. M. Thorne[2]

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Specialist Group in Tectonics and Structural Geology
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Proterozoic geology of the Capricorn Orogen, Western Australia — a field guide

by

S. A. Occhipinti[1], S. Sheppard[2], I. M. Tyler[2], K. Sircombe[3], S. Reddy[1], D. Hollingsworth[1], D. Martin[2], and A. Thorne[2]

Introduction

by

S. Sheppard and S. A. Occhipinti

This field guide accompanies an excursion of the Specialist Group in Tectonics and Structural Geology (SGTSG) of the Geological Society of Australia to the northwestern margin of the Yilgarn Craton (Errabiddy Shear Zone), and north into the adjacent Gascoyne Complex, Edmund Basin, Ashburton Basin, and southern Pilbara Craton (Fig. 1).

The excursion will examine:

- the nature of the boundary between the Archaean Yilgarn Craton and the latest Archaean to Palaeoproterozoic Glenburgh Terrane (southern Gascoyne Complex);
- rocks of the 2540–1970 Ma Glenburgh Terrane of the Gascoyne Complex;
- the nature of metamorphism, deformation, and magmatism during the 2000–1950 Ma Glenburgh Orogeny and the 1830–1780 Ma Capricorn Orogeny;
- the nature of sedimentation within the Ashburton Basin, and overlying Edmund Basin;
- deformation during the 1070–750 Ma Edmundian Orogeny, and its effects on earlier developed structures within the basement;
- the relationship between the Hamersley Group on the southern edge of the Pilbara Craton, and deformation and metamorphism to the south within the Capricorn Orogen.

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The field excursion has six main geological components:

- the Errabiddy Shear Zone (southern margin of the Glenburgh Terrane of the Gascoyne Complex; and the northern margin of the Archaean Narryer Terrane of the Yilgarn Craton);
- the Glenburgh Terrane (southern Gascoyne Complex);
- the central part of the Gascoyne Complex;
- the Mesoproterozoic Edmund Basin;
- the Ashburton Basin;
- the Hamersley Basin on the southern margin of the Pilbara Craton.

Much of the excursion illustrates the results of an ongoing Geological Survey of Western Australia (GSWA) program integrating regional mapping and geochronological, structural, metamorphic, and geochemical studies in the latest Archaean to Palaeoproterozoic Gascoyne Complex (Occhipinti et al., in press; Martin and Thorne, in press), and studies by Curtin University of Technology (Occhipinti and Reddy, in press) and the University of Western Australia. Thus parts of the guide on the Glenburgh Terrane and Errabiddy Shear Zone are directly taken from Occhipinti et al. (2001) and Occhipinti and Reddy (in press) respectively. So far, work in the Capricorn Orogen 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;

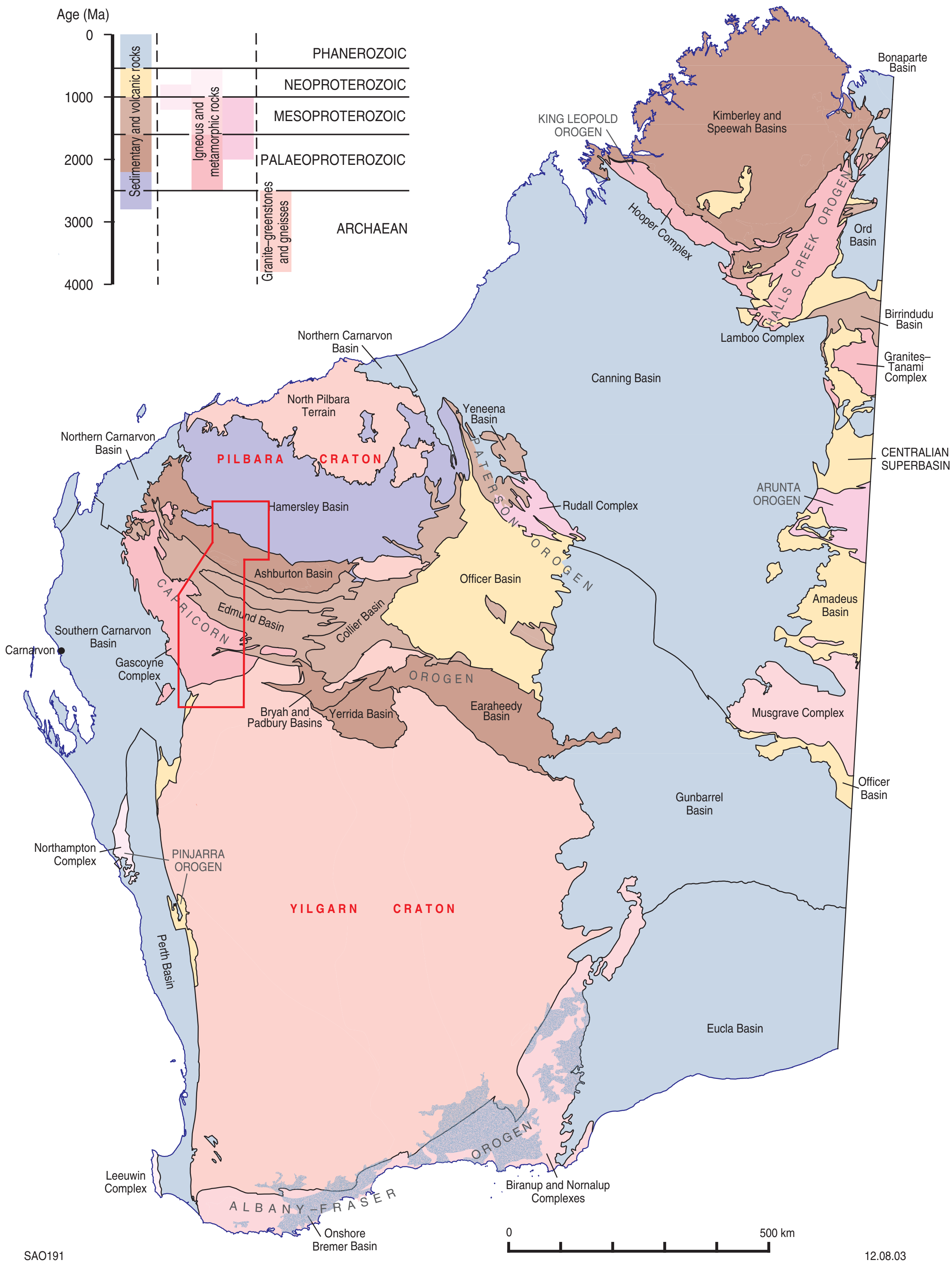

Figure 1. Tectonic units map of Western Australia, after Tyler and Hocking (2001). The red box shows the area covered by this excursion

- possible reactivation accompanied by transpression during the Capricorn Orogeny;
- a major c. 900 Ma deformation episode that accompanied a period of uplift of much of the southern Capricorn Orogen;
- the possibility of a post-Capricorn (1830–1780 Ma), pre-Edmundian (1070–750 Ma) orogenic event involving magmatism, metamorphism, and deformation in the region, and the possible relationship of basement structures to overlying structures in the Edmund Basin;
- the development, deformation, and metamorphism of the Hamersley and Ashburton Basins.

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 (pop. 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 (pop. 1100) to the south and Gascoyne Junction (pop. 35) to the west is provided by the Carnarvon–Mullewa Road, whereas access to Paraburdoo (pop. 2000) to the north is by the Paraburdoo Road.

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 200 mm. Rainfall in the summer months (November–April) results from rain-bearing depressions from the northwest (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 northern part of the area, rain mostly falls from January to July. In the southern part, lesser amounts of rain in winter are provided by cold fronts. All watercourses in the region are ephemeral and the Gascoyne, Murchison, Thomas, and Lyons rivers flow only after heavy rain. However, these river systems are vast and may carry large volumes of water for several months after heavy rain.

Regional geology

Tectonic units

The excursion is a cross section through the Palaeoproterozoic Capricorn Orogen and as such includes the Errabiddy Shear Zone, Hamersley Basin, Gascoyne Complex, Ashburton Basin, and Edmund Basin (Figs 1–3).

The 1830–1780 Ma Capricorn Orogeny has in the past been thought to record the collision of the Archaean Pilbara (to the north) and Yilgarn (to the south) Cratons (Figs 1–3; Tyler and Thorne, 1990a). More recently, however, this theory has been questioned (Occhipinti et al., in press; Sheppard et al., in press). 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, 1990a; Thorne and Seymour, 1991; Martin et al., 1998; Occhipinti et al., 1998, 1999a).

The Errabiddy Shear Zone forms the northwestern boundary between the Archaean Yilgarn Craton, in the south, and the Gascoyne Complex. The shear zone consists of several fault-bound tectonic units, including rocks of the Narryer Terrane, Yarlarweelor Gneiss Complex, Bertibubba Supersuite, Moorarie Supersuite, Camel Hills Metamorphics, Coor-de-wandy Formation, and Mount James Formation.

The Archaean Narryer Terrane (Myers, 1990c) is part of the Yilgarn Craton, and is one of the largest fragments of early Archaean (>3300 Ma) crust on Earth that 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 metasedimentary and mafic meta-igneous rocks (Williams and Myers, 1987; Nutman et al., 1991).

The Narryer Terrane is separated from the Gascoyne Complex to the north by the Errabiddy Shear Zone (Williams et al., 1983b, Figs 2 and 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; Occhipinti and Reddy, in press). Scattered outcrops of the Palaeoproterozoic Coor-de-wandy Formation locally form fault-bound lenses within the Errabiddy Shear Zone. Their low metamorphic grade and simple deformation history suggest that these metasedimentary rocks are significantly younger than the Camel Hills Metamorphics and other rocks that form part of the Errabiddy Shear Zone.

Rocks that lie within the Errabiddy Shear Zone are in faulted contact with those in 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

* Climate data from the Commonwealth Bureau of Meteorology website, 2001, <http://www.bom.gov.au>.

† 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, for example, monzogranite, syenogranite, and so on (Streckeisen, 1976).

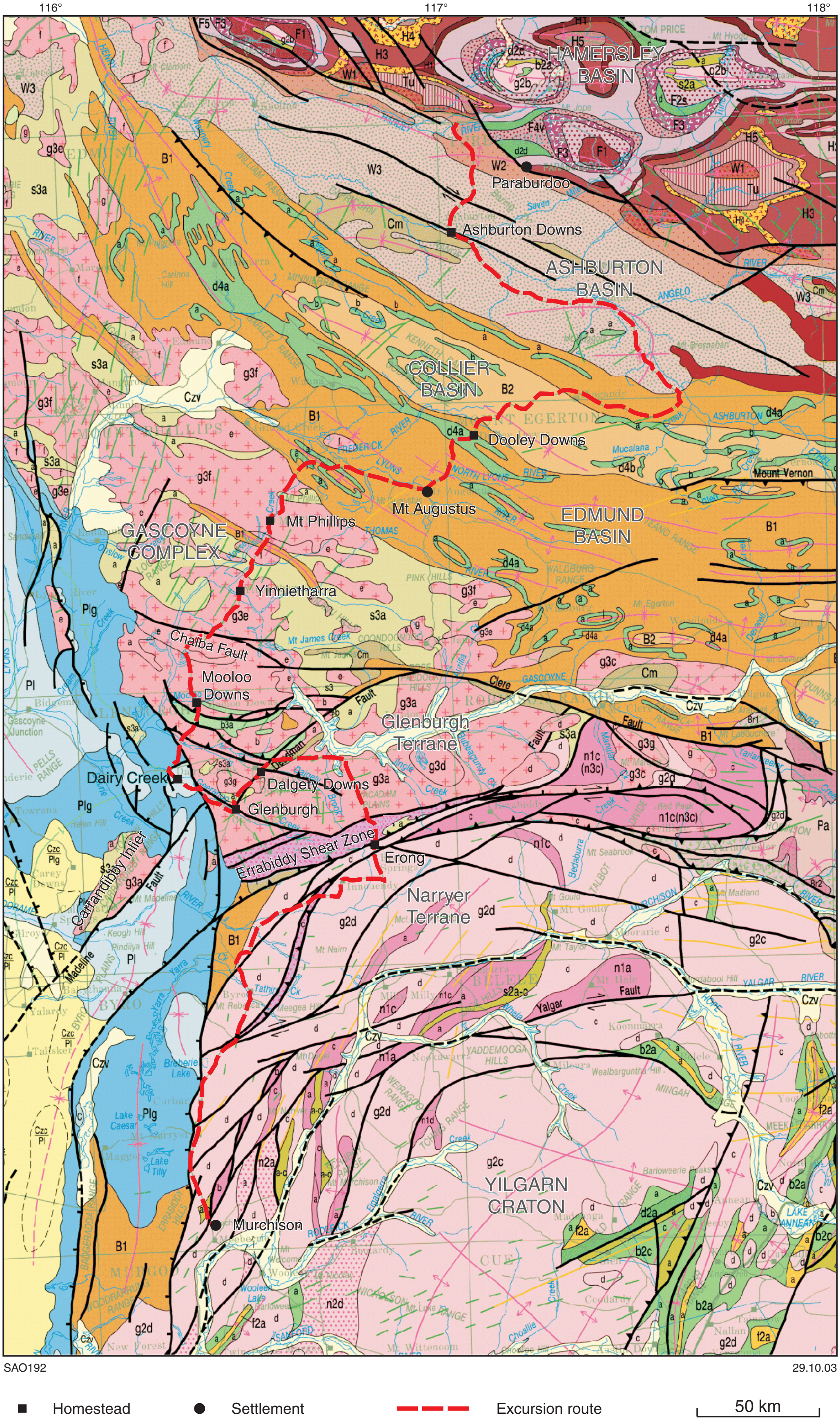

Figure 2. **Portion of the geological map of Western Australia (after Myers and Hocking, 1998) showing major geological units and their geological relationships, and main access roads used during the excursion**

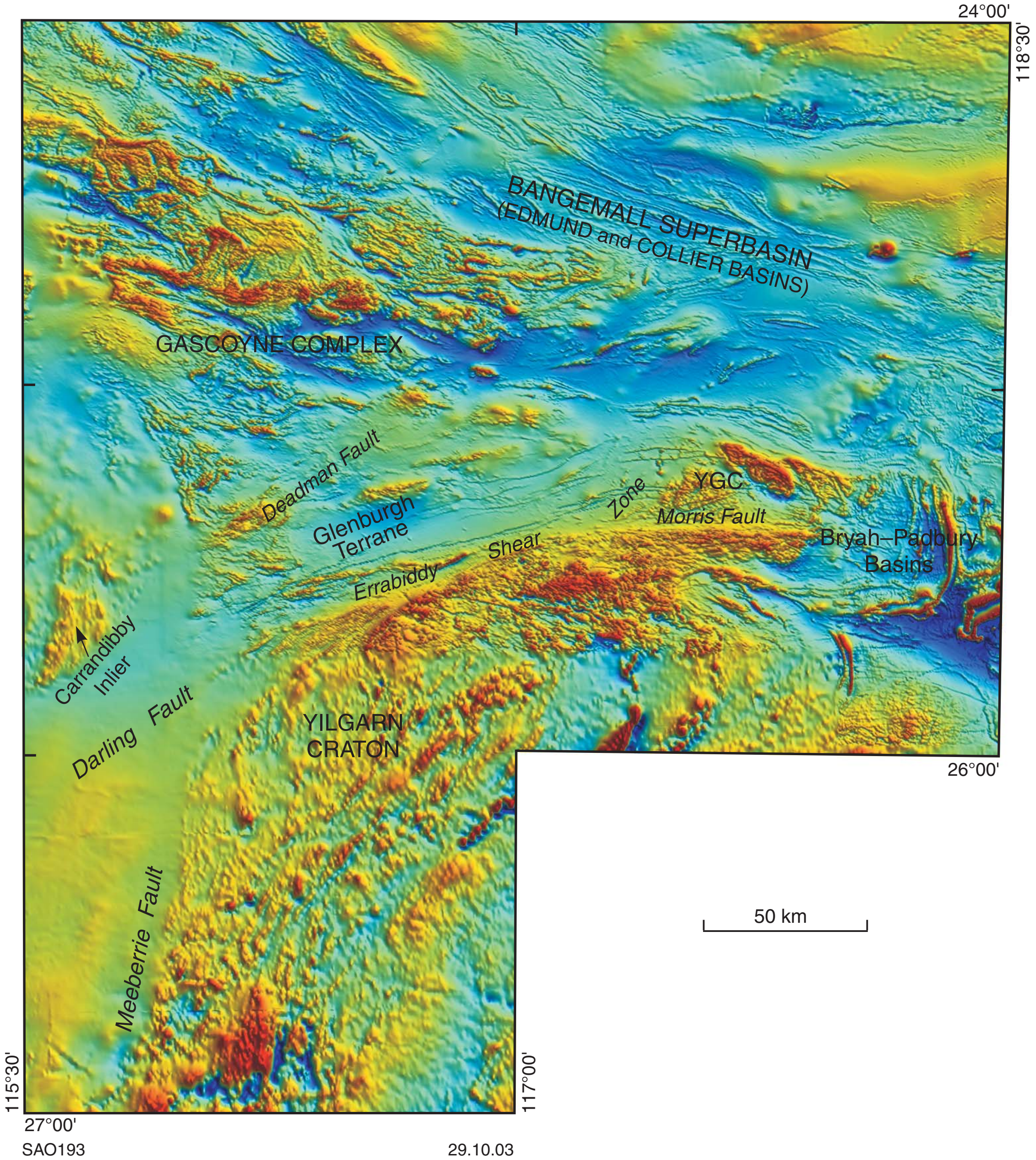

Figure 3. Aeromagnetic image of the northwest Archaean Yilgarn Craton, Errabiddy Shear Zone, Yarlarweelor Gneiss Complex (YGC), Palaeoproterozoic Bryah and Padbury Basins, and Gascoyne Complex (including the Glenburgh Terrane and Carrandibby Inlier), and the latest Palaeoproterozoic to Mesoproterozoic Bangemall Superbasin (Edmund and Collier Basins). After Occhipinti et al. (2001)

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 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 dykes of the 1965–1945 Ma Bertibubba Supersuite and large plutons of potassic and silicic granite that belong to the 1830–1780 Ma Moorarie Supersuite. North of the Glenburgh Terrane the remainder of the Gascoyne Complex commonly consists of granite plutons, sheets, and dykes, granitic gneiss, and metasedimentary rocks. The metasedimentary rocks have been metamorphosed to greenschist to granulite facies, and are locally extensively migmatized (Varvell, 2001; Culver, 2001). The age of the granites in the northern and central Gascoyne Complex is variable, with some granites dated at c. 1780 Ma (Krapez and McNaughton,

1999) that may form part of the Moorarie Supersuite, and others dated at 1670–1620 Ma (Martin and Thorne, 2001; Nelson, 2001).

The Hamersley Basin contains metasedimentary and metavolcanic rocks of the Fortescue, Hamersley, and Turee Creek Groups (Mount Bruce Supergroup collectively; Thorne et al., 1991). Rocks of the Mount Bruce Supergroup were deposited over a period of 340 million years on top of granite–greenstones of the Pilbara Craton between 2770 and 2430 Ma (Thorne and Trendall, 2001). These rocks were deformed into large folds during the Palaeoproterozoic Opthalmia and Capricorn Orogenies.

During the Capricorn Orogeny the mainly turbiditic sedimentary rocks of the Wyloo Group were deposited in the Ashburton Basin in a foreland-basin setting on top of the Mount Bruce Supergroup, and were deformed and metamorphosed into a fold-and-thrust belt (Thorne and Trendall, 2001).

The Gascoyne Complex (including the Glenburgh Terrane) and the northern edge of the Yilgarn Craton are unconformably overlain by scattered outcrops of the Palaeoproterozoic c. 1800 Ma Mount James Formation, and by the Palaeoproterozoic–Mesoproterozoic Edmund Group of the Bangemall Supergroup (Williams et al., 1983b; Drew, 1999a; Martin et al., 1999; Occhipinti and Sheppard, 2001). To the north the Hamersley and Ashburton Basins are unconformably overlain by rocks of the Bangemall Supergroup, and their relationship to the Mount James Formation is unknown.

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), but 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 2000–1950 Ma Glenburgh Orogeny and the 1830–1780 Ma Capricorn Orogeny. During the Glenburgh Orogeny (Occhipinti et al., 1999b), the Glenburgh Terrane, 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 1965–1945 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) intruded the northwestern margin of the Yilgarn Craton, and 1965–1945 Ma granite dykes intruded the southern Glenburgh Terrane and Camel Hills Metamorphics.

The Glenburgh Terrane and Camel Hills Metamorphics were further deformed and metamorphosed at low to medium grade during the 1830–1780 Ma Capricorn Orogeny, 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-bound 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). Palaeoproterozoic to 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. 1070 and c. 750 Ma (Wingate and Giddings, 2000; 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. Late Carboniferous to Early 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 are indicated by the pattern of Rb–Sr ages of biotite reported by Libby et al. (1999; Fig. 4). These authors found that Rb–Sr ages on biotite increased from 817–739 Ma in the southern Gascoyne Complex and Yarlarweelor Gneiss Complex to a minimum age of c. 1544 Ma in the northern part of the Narryer Terrane, closest to the Errabiddy Shear Zone. The Rb–Sr ages progressively increase southwards across the Narryer Terrane. Myers (1990b) suggested that the Rb–Sr age data reflect the extent of substantial southward thrust-stacking of Proterozoic rocks of the southern Capricorn Orogen over the autochthonous Narryer Terrane. He also suggested that the Rb–Sr data supported c. 1700 Ma reflecting a time of cooling, when the Narryer Terrane and its overlying thrust pile were 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 were 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 by 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 1670–1630 Ma (Pearson, 1996; Nelson, 2001) have been reported from granite plutons.

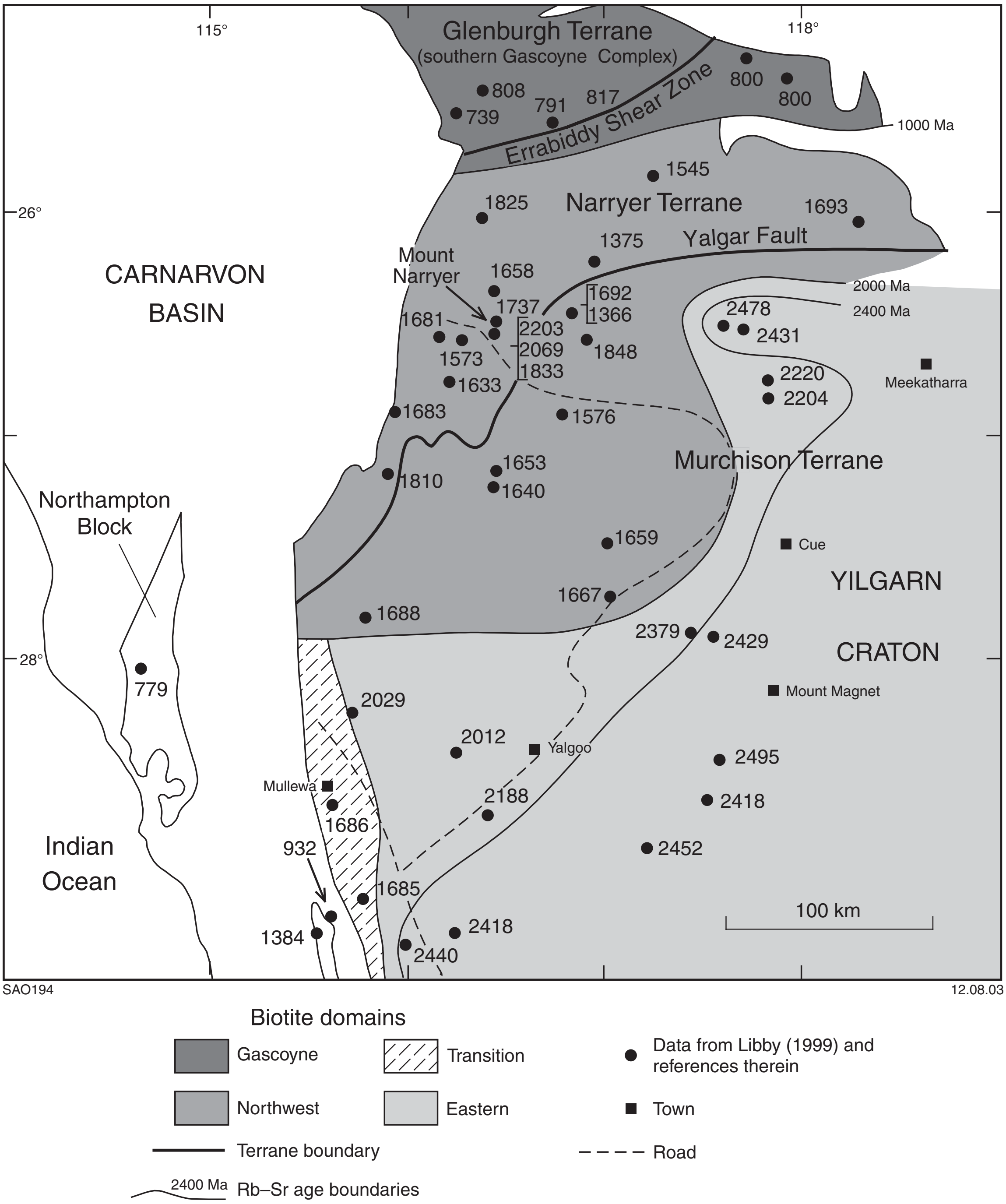

Figure 4. Rubidium–strontium biotite ages and domains in the Yilgarn Craton and the southern Gascoyne Complex. Modified from Libby et al. (1999)

Errabiddy Shear Zone

by

I. M. Tyler, S. A. Occhipinti, S. Sheppard, and S. M. Reddy

Introduction

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 1–3, 5, and 6).

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, 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 metasedimentary Palaeoproterozoic rocks that are confined to the Errabiddy Shear Zone, and include 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, and c. 1950 Ma granite dykes

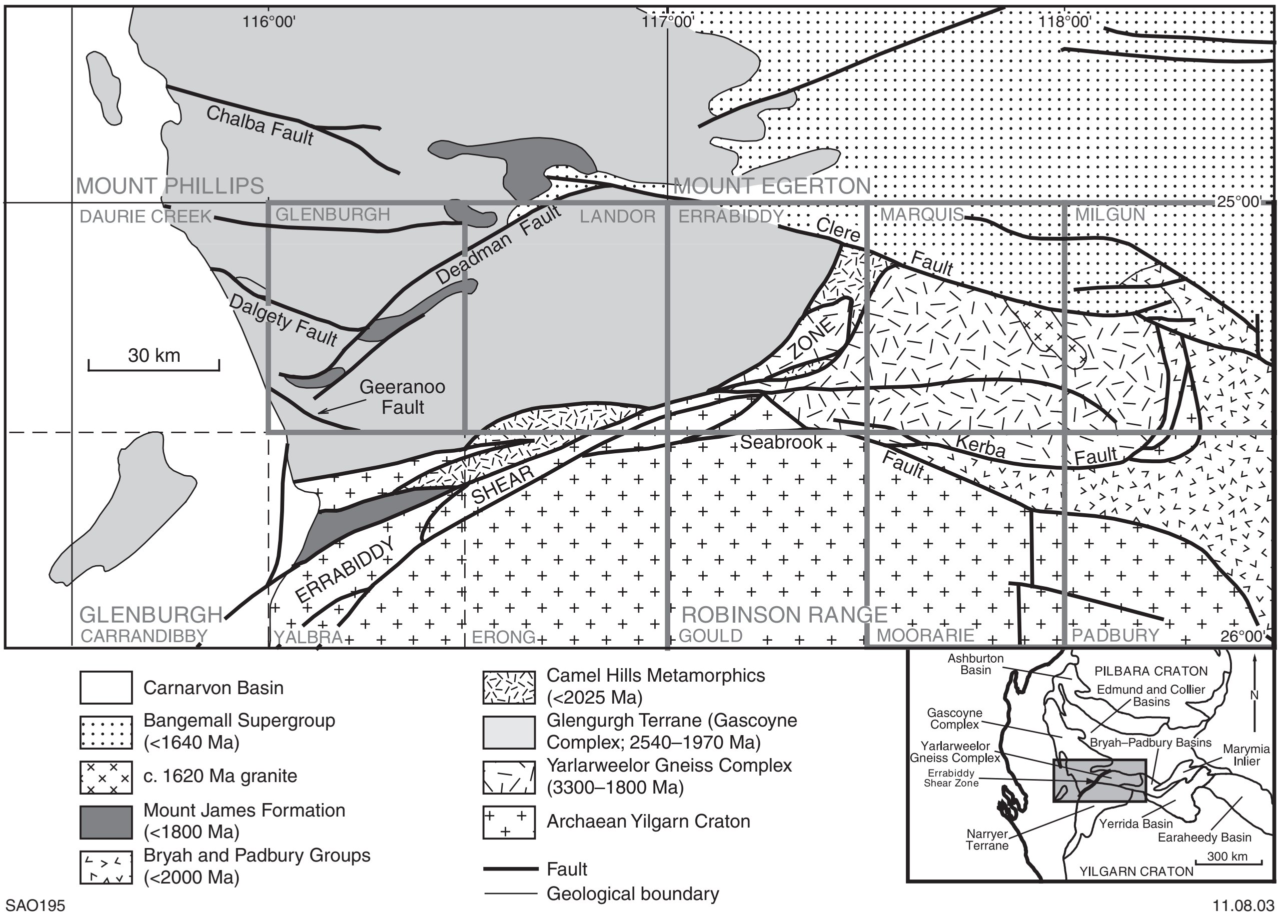

Figure 5. Simplified geological map of the southern part of the Gascoyne Complex (Glenburgh Terrane) and part of the Errabiddy Shear Zone. Bold outlined areas represent 1:100 000-scale maps recently published by GSWA (1996–2003). After Occhipinti et al. (2001)

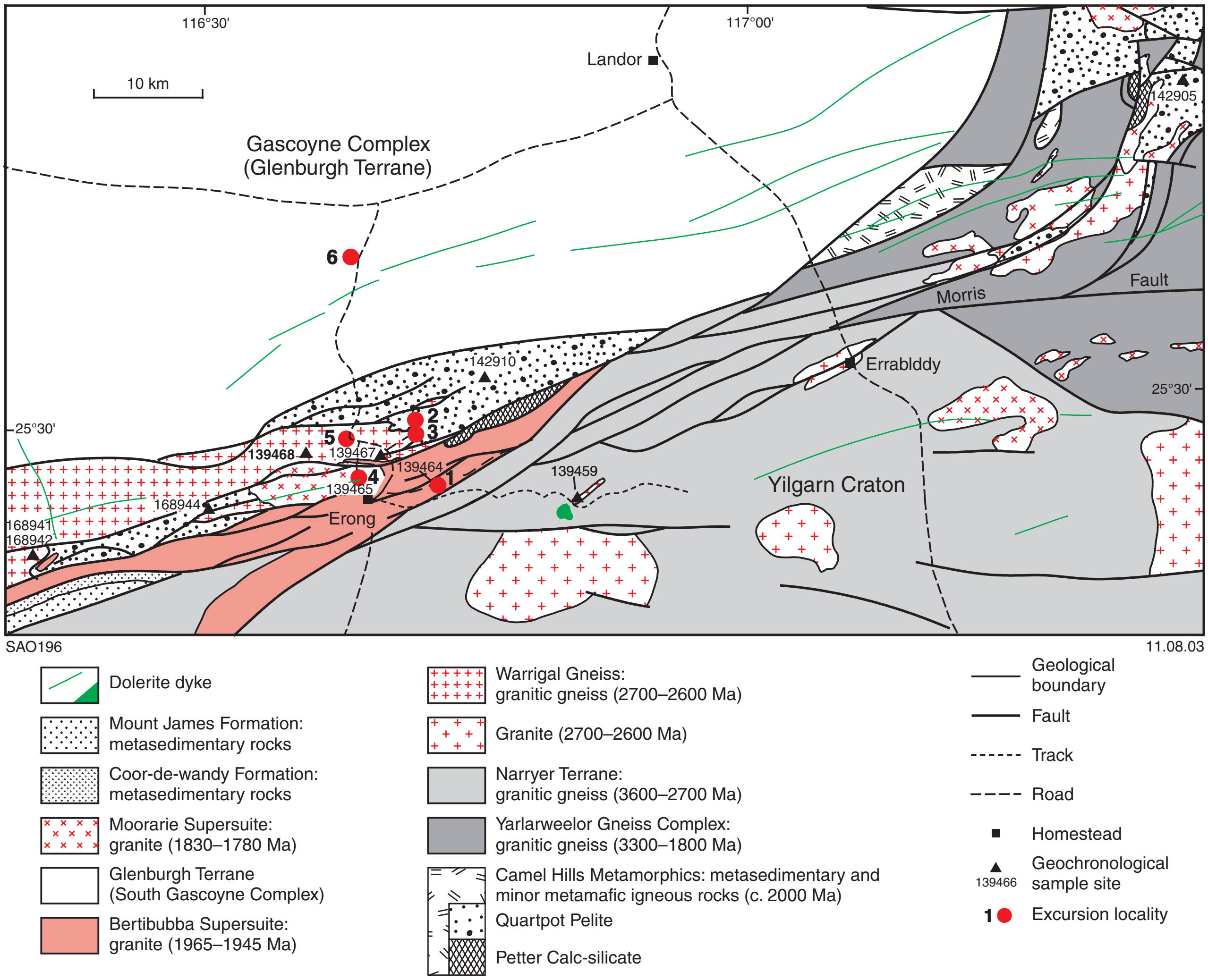

Figure 6. Simplified geological map of the Errabiddy Shear Zone showing main rock units, geochronological sites, and Localities 1–6. Modified from Occhipinti et al. (2001)

that cut the Dalgaringa Supersuite in the southernmost Glenburgh Terrane (Figs 2, 3, and 6);

- Coor-de-wandy Formation — older-than-1800 Ma low-grade micaceous metasedimentary schists;
- ?Mount James Formation — younger-than-1800 Ma low-grade metasiliciclastic rocks, largely composed of quartz sandstone and quartzite.

All components have fault-bound 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) is unconformably overlain by the ?Mount James Formation. 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, 5, and 6) initially developed during the 2000–1950 Ma Glenburgh Orogeny (Occhipinti et al., 1999b; Occhipinti et al., in press; Occhipinti and Reddy, in press). 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 1). 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 1; Fig. 7), between 1975 and 1950 Ma. 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 1). Granite of the Bertibubba Supersuite intruded the northwestern margin of the Yilgarn Craton at, or towards the end of, the Glenburgh Orogeny.

During the 1830–1780 Ma Capricorn Orogeny, the Errabiddy Shear Zone, the northwestern Yilgarn Craton

Table 1. Summary of geological history of the Errabiddy Shear Zone and Glenburgh Terrane

Age (Ma)	*Deformation event*	*Domain* — *Errabiddy Shear Zone*	*Domain* — *Glenburgh Terrane*[(a)]
PRE-GLENBURGH OROGENY (>2000 Ma)			
c. 2500			Crystallization of some granite protoliths to the Halfway Gneiss
?			Deposition of protoliths of Moogie Metamorphics
GLENBURGH OROGENY (2000–1950 Ma)			
			Intrusion of Dalgaringa Supersuite
>2000	D_{1g}, S_{1g}; M_{1g}, medium to high grade		**SGT** — Foliation in gneissic and foliated granite of Dalgaringa Supersuite
?c. 2000–1950		Deposition of precursor Camel Hills Metamorphics sedimentary rocks onto ?northern Yilgarn Craton, ?southern 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 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
CAPRICORN OROGENY (1830–1780 Ma)			
c. 1825	D_{1n}, F_{1n}, S_{1n}	Intrusion of plugs, sheets, and veins of Moorarie Supersuite granitic rocks	
c. 1810	M_{1n}; 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_{1n} in the Glenburgh Terrane[(b)], possibly into tight asymmetric folds; deformation of Mount James Formation Deposition of Bangemall Supergroup in intracratonic sag mainly over the Gascoyne Complex (including Glenburgh Terrane)	
		Extension, intrusion of dolerite sills into the Bangemall Superbasin	
EDMUNDIAN OROGENY (1020–750 Ma)			
<c. 1020	D_{1e}, F_{1e}, S_{1e}; subgreenschist facies	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

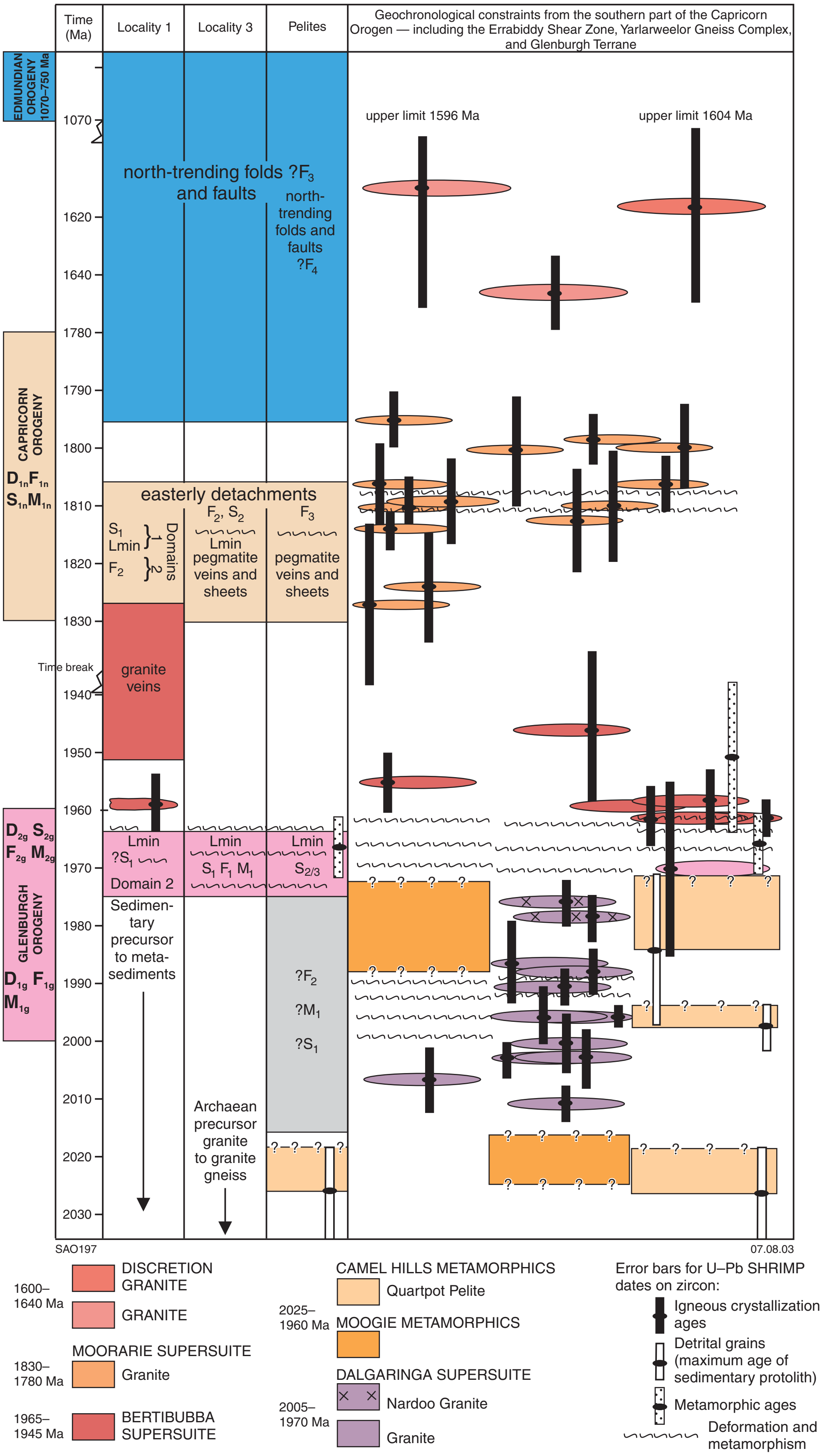

Figure 7. Time–space plot showing different structural elements in the region, and how the structural elements observed on this excursion (Localities 1–3) correlate with them and each other. Geochronological data in the right-hand column were collected as part of a GSWA mapping program between 1996 and 2000

(including the Yarlarweelor Gneiss Complex), and the Glenburgh Terrane were deformed. Rocks that already formed part of the Errabiddy Shear Zone were variably folded, faulted, and foliated during this regional deformation event (Table 1; Fig. 7). These Capricorn Orogeny-aged structures are denoted as D_{1n} structures to distinguish them from structures that developed during the Glenburgh Orogeny (Occhipinti et al., 2001; Occhipinti et al., in press). The F_{1n} folds are commonly tight and upright and contain shallow to steeply plunging fold axes. They were locally deformed by east-southeasterly trending dextral strike-slip movement and north–south reverse and normal faults, which appear to have accompanied a transition from ductile to brittle conditions in the region and are likely to have also occurred during the Capricorn Orogeny. East of the Errabiddy Shear Zone, the Yarlarweelor Gneiss Complex, which largely originated between 1820 and 1800 Ma by pervasive reworking of a tectonic slice of the Narryer Terrane and intrusion of voluminous granite and pegmatite (Occhipinti et al., 1998), was uplifted in a possible restraining bend at this time. Moderately to gently southerly dipping faults, which developed in discrete narrow zones within the northwestern part of the Yilgarn Craton, may have 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 press).

Northerly trending gentle to open folds and kink folds only weakly deformed D_{1n} structures throughout the Errabiddy Shear Zone. These may have formed at any time after the Capricorn Orogeny, but have commonly been correlated with the Mesoproterozoic–Neoproterozoic Edmundian Orogeny (Occhipinti et al., 2001).

Excursion localities — Errabiddy Shear Zone

Day 1

Today, we will drive north into the Errabiddy Shear Zone where we will begin our excursion by visiting a well-exposed mylonite zone within the shear zone (Fig. 6).

Locality 1: Granite of the Bertibubba Supersuite and mylonite zone

Drive to Erong Springs Homestead, north-northeast of Murchison Settlement. Follow the Erong Road east for 1.1 km. 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 away from the track. This is the start of Locality 1.

The main rock types at this locality are weakly to well-foliated granite, which is part of the 1965–1945 Ma Bertibubba Supersuite (Occhipinti et al., in press), well-foliated to mylonitized granite, psammitic metasedimentary rocks, and quartz mylonite. A traverse will be made southwards from an area of relatively low strain in the monzogranite (Locality 1a; part of Domain 1), to a prominent ridge formed by a mylonite zone that comprises mylonitized granite, psammite, and quartz-vein material (collectively called felsic gneiss; Locality 1b; Domain 2; Fig. 8). The contact between the weakly and well-foliated granite (Domain 1) and well-foliated to mylonitized granite, metasedimentary rock, and quartz mylonite (Domain 2) is a high-strain zone.

Locality 1a: Domain 1, biotite monzogranite (start at MGA 473646E 7173731N)

At this locality, variably foliated medium-grained, even-textured and porphyritic biotite monzogranite is veined by granite and pegmatite.

A sensitive high-resolution ion microprobe (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; Fig. 9; Nelson, 2000).

Walk south through the medium- and even-grained, porphyritic biotite monzogranite towards a quartz-vein ridge (Locality 1b; see below).

The porphyritic granite has been heterogeneously deformed, is locally well foliated, and locally contains a strong, subhorizontal to steeply dipping mineral lineation. In places, narrow shear zones are subparallel to the foliation. Rotated feldspar phenocrysts and S–C fabrics indicate both sinistral and dextral movement (Fig. 10b).

Deformation at Locality 1a (Domain 1)

The earliest recorded structure in Domain 1 is a pervasive S_1 foliation that strikes northeast and dips moderately to steeply towards the south or southeast (Fig. 8). This foliation is correlated with the regional D_{1n} deformation event (Fig. 7) and is commonly defined by the alignment of variable amounts of deformed feldspar (variably sericitized), biotite, muscovite, and minor epidote, although epidote is typically randomly oriented and replaces feldspar. Variably recrystallized aggregates of quartz make up domains that follow S_1. Mineral lineations on the S_1 surface are uncommon. However, where present they are defined by aggregates of quartz or muscovite, and elongate feldspar porphyroclasts, now pseudomorphed by sericite. These mineral lineations plunge moderately to steeply to the south, although some are shallowly plunging to the east or west.

Dykes, 1 to 6 m thick, of fine-grained leucogranite intrude the medium-grained granite in Domain 1. These granite dykes commonly strike subparallel to the S_1 foliation and are locally foliated (particularly along their edges), although they are mostly undeformed, indicating strain localization along their edges. Most of the thin leucocratic granite veins that intrude the medium-grained granite in Domain 1 also trend subparallel to the S_1 foliation; however, veins intruding oblique to S_1 are folded into close–tight folds with their fold-axial surfaces parallel to S_1, although they are commonly unfoliated.

Localized shear bands overprint S_1, but are commonly subparallel to slightly oblique to S_1, or to brittle or brittle–

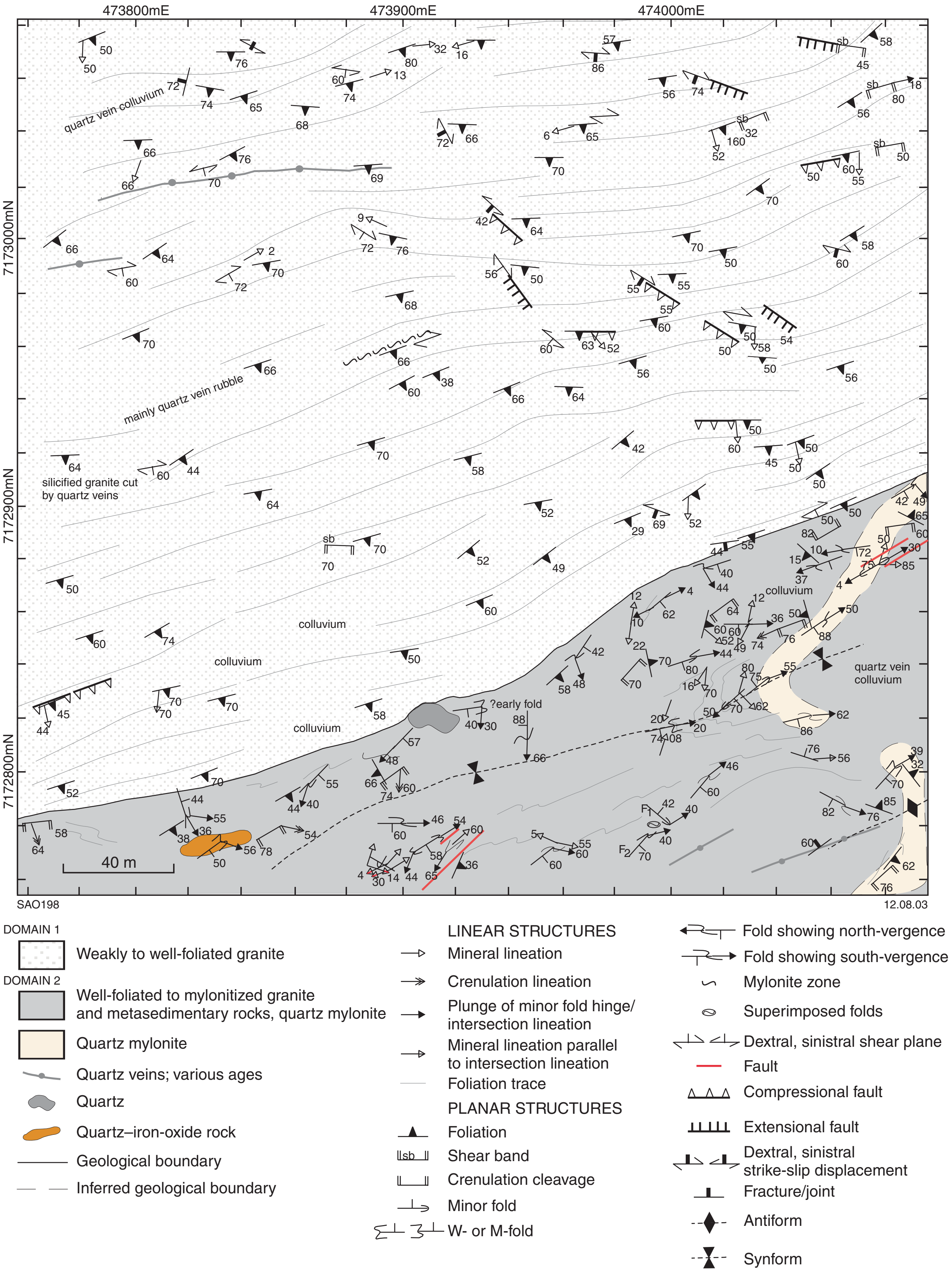

Figure 8. Geological map of Domains 1 and 2 of Locality 1

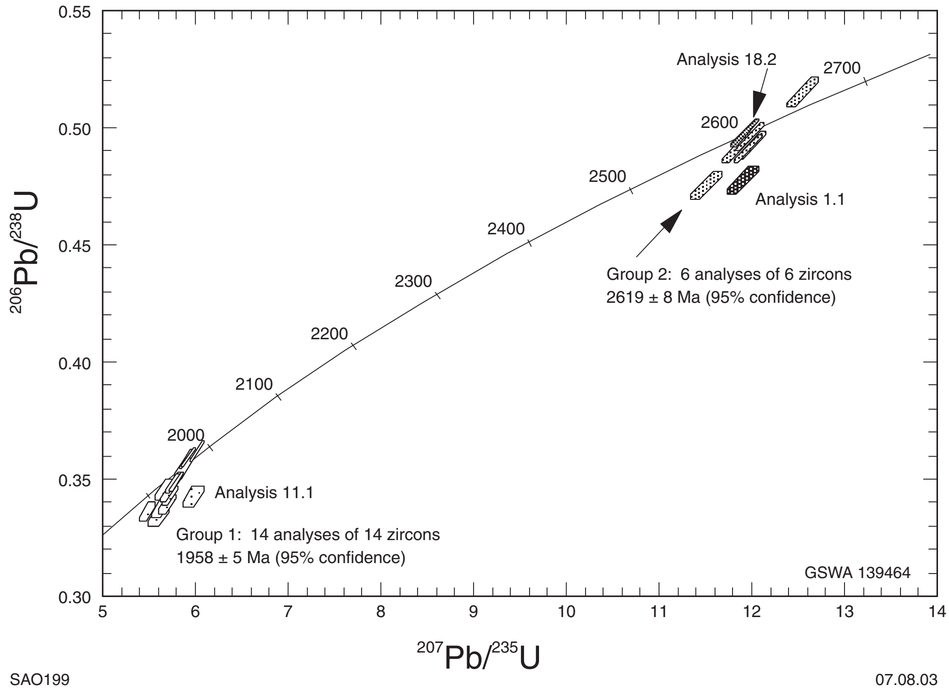

Figure 9. Concordia plot for porphyritic biotite monzogranite of the Bertibubba Supersuite, 3 km southwest of Camel Hills Bore (GSWA 139464). After Occhipinti et al. (2001)

ductile features such as reverse or normal faults in the area (Fig. 10a). Delta and sigma trails of deformed feldspar phenocrysts (Fig. 10b) indicate sinistral and dextral strike-slip movements associated with shearing. Mineral lineations on some of the shear surfaces have identical orientations to those on adjacent S_1 foliation surfaces.

Brittle–ductile faults, which cut the S_1 foliation, include reverse and normal faults, dextral and sinistral strike-slip shears and faults, and fractures (Fig. 10a,b). Locally sinistral or dextral strike-slip movements appear to be associated with reverse faulting, with displacement along the faults being only 1–4 cm. Reverse faults are the most abundant fault type observed in Domain 1 and strike in a southwesterly to southerly or southeasterly direction (Fig. 10a). A few normal faults trend in a similar direction to southwesterly plunging reverse faults.

Easterly to east-northeasterly striking quartz veins cutting the S_1 foliation in Domain 1 (Fig. 8) are typically less than 1 m wide and discontinuous. The S_1 foliation in Domain 1 is locally weakly folded about gentle, steeply plunging, upright, southeasterly striking folds. The relationship of these folds to ductile shear structures and faults in the area is unknown.

Locality 1b: Mylonite zone (MGA 474078E 7172884N)

This quartz-vein ridge is a mylonite zone that consists of mylonitized granite, metasedimentary rocks, and vein quartz. Noncylindrical upright tight to isoclinal folds are prominent. The main foliation in the granite and metasedimentary rocks increases in intensity closer to the mylonite zone, with no evidence of overprinting. The folded foliation in the mylonite zone is regarded as older than the main foliaton in the granite (Domain 1), corresponding to the regional D_{2g} event (Table 1; Fig. 7). However, the upright folds (Fig. 10c,d) in the mylonite zone are most likely to have developed at the same time as the foliation in the granite (Domain 1) and therefore during the regional D_{1n} event.

Deformation at Locality 1b (Domain 2)

In Domain 2 the earliest recognizable structure is a fabric in felsic gneiss (collectively including foliated granite to granitic gneiss and metasedimentary rocks), which may represent an early tectonic foliation, or alternatively bedding (S_0). Locally, shallow to moderately plunging isoclinal, subhorizontal small-scale F_1 folds deform this early fabric (Fig. 10c).

The main regional foliation in Domain 2 (S_1) is schistose to gneissic and subparallel to F_1 fold-axial surfaces (Fig. 10c). This foliation is variably defined by strongly recrystallized granoblastic quartz–feldspar domains and elongate aggregates of biotite or muscovite. Mineral lineations associated with S_1 are defined by elongate quartz and quartz mineral aggregates, and locally by muscovite or biotite aggregates. These mineral lineations are well developed in quartz mylonite and the felsic gneiss and have variable orientations (Fig. 10d).

F_2 folds are tight to isoclinal, upright, east-northeasterly striking, and typically steeply southeasterly plunging (Fig. 10c,d). Mineral lineations that developed on the S_1 surface are deformed by F_2 folds (Fig. 10d). Most of the

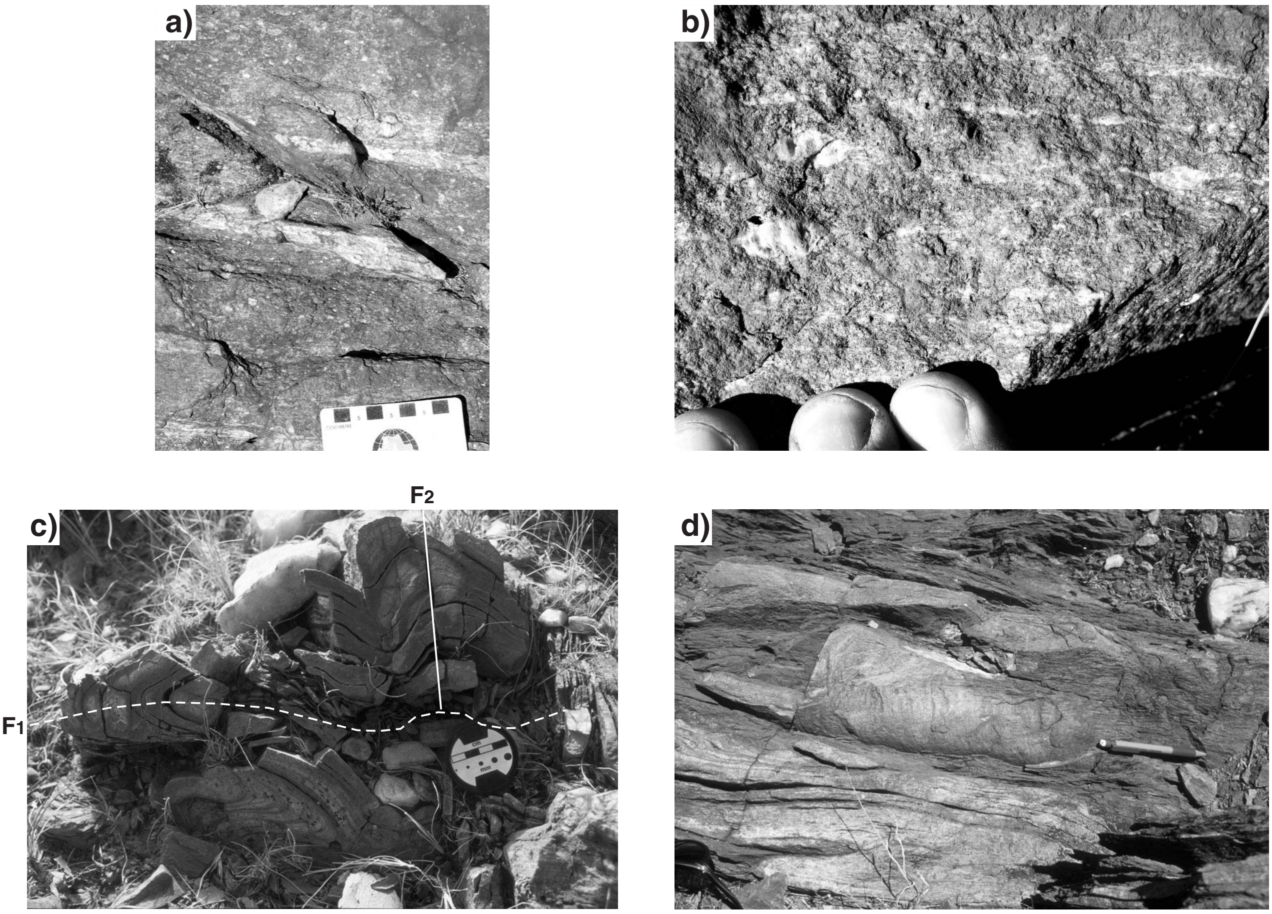

Figure 10. Different structural elements observed in Domains 1 and 2 of Locality 1: a) reverse fault; b) dextral strike-slip kinematics around deformed feldspar phenocrysts; c) F_1 fold refolded by F_2 folds in quartzite; d) folded quartz aggregate mineral lineation around upright shallow-plunging F_2 fold; e) folded quartz aggregate mineral lineation around upright doubly plunging F_2 fold

small-scale F_2 fold-axial surfaces are steeply plunging to the southeast. However, a weak spread in the attitudes of F_2 fold-axial surfaces to northwesterly or northeasterly trending suggests that they may have either been locally refolded, or are curvilinear, or both. Given that F_2 fold-axial surfaces do not show much evidence for possible refolding, whereas their hinge lines do, it is unlikely that any refolding of F_2 in the region was more than just superposition of gentle to open folds. Thus, the spread of F_2 fold hinges is more likely to be due to the noncylindricity of F_2 folds.

Areas of high D_2 strain contain a well-developed crenulation cleavage, and feldspar is typically replaced by sericite, and biotite may be partially replaced by chlorite. These crenulations strike northeast. A large-scale F_2 fold can be traced in quartz mylonite in the eastern part of Domain 2 (Fig. 8). Likewise, large-scale F_2 folds can be traced in well-foliated to mylonitized felsic gneiss to the west. However, near the boundary between Domains 1 and 2, folds in Domain 2 are very tight and may have been deformed at higher strain.

Northeasterly trending quartz veins cut D_2 structures in Domain 2. These veins are typically only a few metres long and less than 1 m wide, but can be up to 30 m long and a few metres wide. One quartz vein, in the southeasternmost corner of Domain 2, contains northeasterly trending joints, but these are not pervasive in the area.

Day 2

Today we will continue looking at different components of the Errabiddy Shear Zone. These will include metasedimentary rocks of the Palaeoproterozoic Camel Hills Metamorphics, granitic gneiss of the Archaean Warrigal Gneiss, and the c. 1800 Ma Erong Granite.

Locality 2: Diatexite of the Quartpot Pelite (Camel Hills Metamorphics; MGA 470872E 7180091N)

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) southeast for 4.3 km, to Locality 2. 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 locally (Fig. 11a–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 meta-sedimentary 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.

Mineral assemblages in the migmatized Quartpot Pelite in this part of the Errabiddy Shear Zone consist of biotite, garnet, muscovite, sericite mats (possibly 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 occurred during the Glenburgh Orogeny, synchronous with the development of a subhorizontal, bedding-parallel foliation, which is interpreted to have developed during D_{2g} (Table 1; Fig. 7). 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 to amphibolite facies. This indicates that there may be a metamorphic isograd between these outcrops of Quartpot Pelite and those to the east. This isograd was folded during the Capricorn Orogeny (D_{1n}).

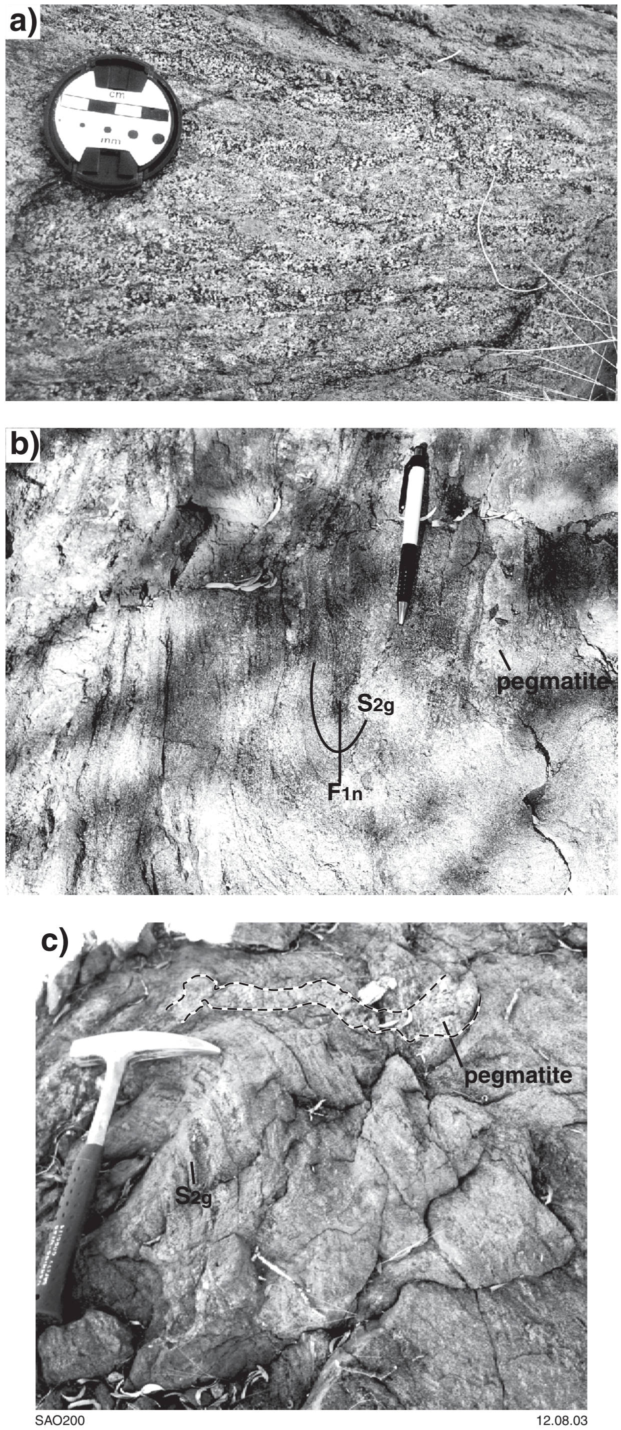

Figure 11. Migmatitic pelitic gneiss of the Quartpot Pelite at Locality 2:
a) melt patches in diatexite;
b) thin pegmatite veins cutting diatexite subparallel to the fold-axial surface of small-scale F_{1n} folds;
c) a thin vein of pegmatite cutting the S_{2g} fabric in diatexite but folded by F_{1n}. After Occhipinti et al. (2001)

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 locally grade into nebulitic migmatite, indicating increased degrees of in situ partial melting in places.

The S_{2g} foliation was tightly folded during the Capricorn Orogeny (F_{1n}), and locally a well-developed S_{1n} crenulation cleavage is present, 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 6). Elsewhere within the shear zone D_{1n} structures may be northeasterly or northerly trending (Figs 2, 3, and 6).

The Quartpot Pelite has been sampled for SHRIMP U–Pb zircon geochronology from three localities within the Errabiddy Shear Zone (Nelson, 1998, 1999, 2001; Occhipinti, S. A. and Reddy, S., unpublished data). 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, 2001).

At the first two localities the diatexite melt phase was sampled, which locally comprised more than 50% of the outcrop's volume. Ages of 2550–2025 Ma were obtained from pitted zircon grains in both samples, and were interpreted to be of detrital origin (Fig. 12). The youngest zircon populations in GSWA 142905 and 142910 (Fig. 12a,b) were obtained from rims and cores of pitted grains with very low Th/U values. In both samples Nelson (1998, 1999) found that the youngest zircon populations consisted of only two zircons dated at 1951 ± 13 and 1966 ± 5 Ma for GSWA 142905 and 142910 respectively. However, following cathodo-luminesence (CL) imaging, further analyses of GSWA 142910 gave a concordant to slightly discordant age of 1950 ± 7 Ma from 8 analyses of 7 rims (95% confidences; mean square of weighted deviates — MSWD = 2.0; Occhipinti, S. A. and Reddy, S., unpublished data).

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 occurred 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. Further work on GSWA 142905 and 142910 (Occhipinti, S. A. and Reddy, S., unpublished data) showed that there is more than one generation of metamorphic zircon growth with older metamorphic rims developed prior to the detrital zircon sedimentary transport at 2050–2030 Ma. All generations of metamorphic zircon growth contained high uranium, corresponding to low CL and low thorium contents.

Cathodoluminescence on GSWA 142905, 142910, and 168944 (Occhipinti, S. A. and Reddy, S., unpublished data) illustrated that zircons from migmatized pelitic samples (GSWA 142905 and 142910) and unmigmatized psammitic rock (GSWA 168944) were morphologically different. In the migmatized pelitic rocks, zircons were commonly smooth edged, euhedral to subhedral grains, which were locally rounded and contained thin, low-CL metamorphic rims. In contrast, the unmigmatized psammitic rock (GSWA 168944) commonly contained smaller grains that appeared to be mechanically abraded (irregularly shaped with concentric zones cut off) and low-CL rims were absent or rare.

The data of Nelson (1998, 1999, 2001) and S. A. Occhipinti and S. Reddy (unpublished data) for the Quartpot Pelite indicate that high-grade metamorphism and its localized associated migmatization occurred at c. 1950 Ma, and the maximum age of deposition of the protolith to the Quartpot Pelite was c. 2000 Ma. This implies that most of the detrital zircons dated from the Quartpot Pelite were probably sourced from the Glenburgh Terrane of the southern Gascoyne Complex. Data from Nelson (1998, 1999, 2001) show that a minority of zircons were sourced from the Yilgarn Craton (Fig. 12), 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; however, attempts to date them have been unsuccessful.

Locality 3: Warrigal Gneiss (MGA 470691E 7178872N; time permitting)

From Locality 2, return to the track and follow it south for 1.3 km to a prominent rocky hill. This is Locality 3.

Locality 3 is about 7.5 km northeast of Erong Homestead (Fig. 6). The main rock type at this locality is a granitic gneiss, which is locally pegmatite banded. This gneiss forms part of the Warrigal Gneiss that developed during the 2000–1950 Ma Glenburgh Orogeny. The granitic gneiss is well foliated to banded and consists of interleaved mesocratic and leucocratic granite phases. The granitic protoliths to the gneiss all consist of quartz, feldspar, and mica (variable amounts of biotite and muscovite), and locally contains garnet. Granite protoliths to the gneiss are still recognizable in the field and dominantly consist of medium-grained, even-textured monzogranite, although locally some weakly porphyritic granite is present.

The relationships between different granite phases of the gneiss are complex, but in places dykes of both leucocratic and mesocratic granitic components intrude each other. Four individual granitic components of the Warrigal Gneiss have been dated from three localities

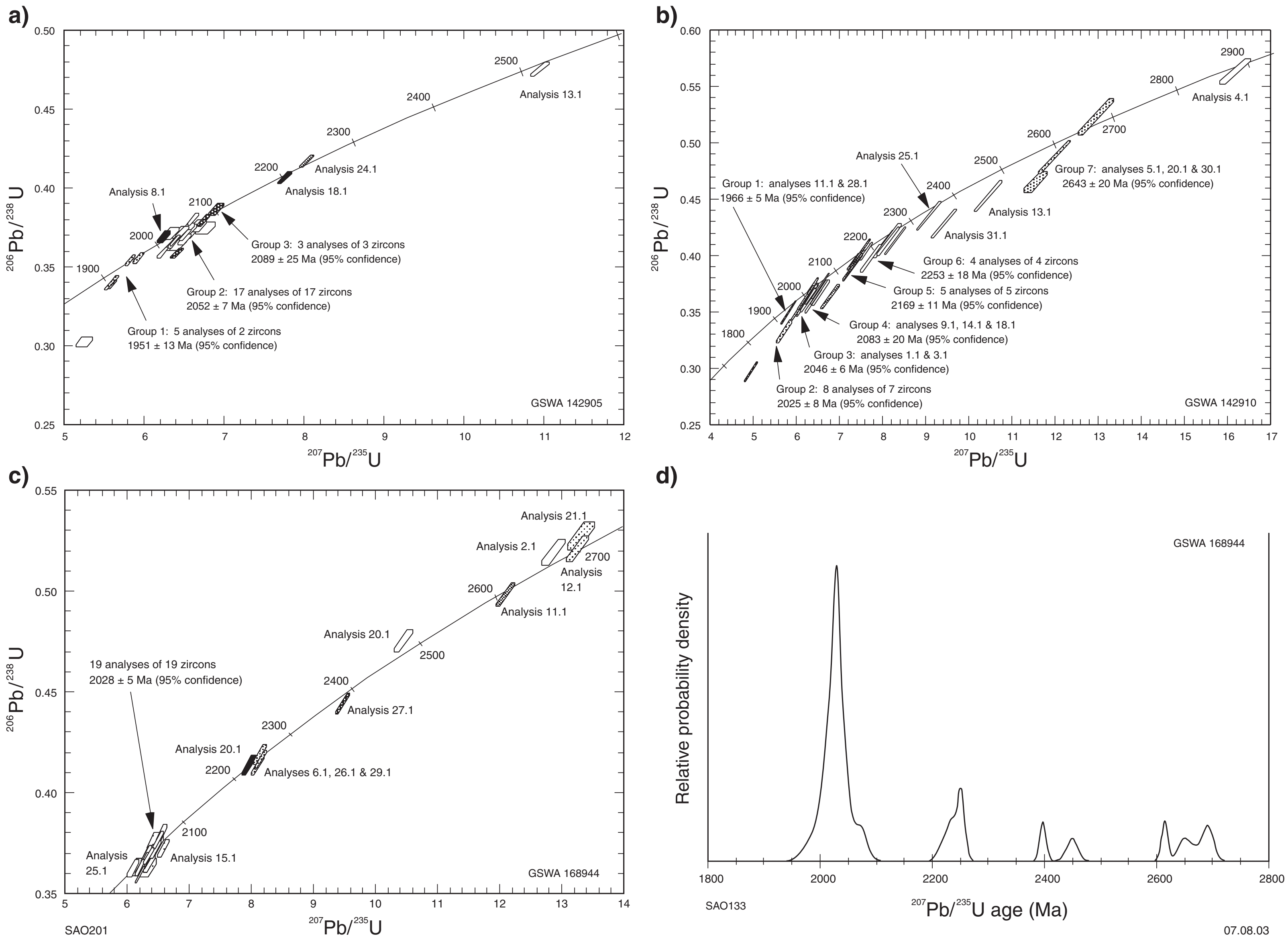

Figure 12. Geochronology results for samples of the Quartpot Pelite (Camel Hills Metamorphics): a) concordia plot for GSWA 142905; b) concordia plot for GSWA 142910; c) concordia plot for GSWA 168944; d) Guassian-summation probability density plot for GSWA 168944 (from Nelson, 1998, 2001). After Occhipinti et al. (2001)

within the Errabiddy Shear zone, giving SHRIMP U–Pb zircon ages that range from c. 2758 to c. 2585 Ma (Nelson, 2000, 2001; Occhipinti et al., 2001).

At Locality 3 the granitic gneiss is interleaved with amphibolite schist and gneiss, calc-silicate gneiss, quartzite, and pegmatite. Gneissic banding within the granitic gneiss is broadly parallel to its contacts with lenses of supracrustal rocks, and to a well-developed foliation. Contacts between the amphibolite, and calc-silicate and granitic gneiss are commonly either not exposed or appear to be tectonic. However, at one locality veins of amphibolite gneiss clearly cut across layering in the granitic gneiss prior to being folded together, indicating that the amphibolites were originally dolerite dykes or veins.

Amphibolite lenses within the granitic gneiss consist of both medium-grained, even-textured gneissic amphibolite and mafic schist. Gneissic amphibolite comprises plagioclase, hornblende, and actinolite–tremolite, whereas the mafic schist comprises varying amounts of actinolite–tremolite and chlorite. The calc-silicate gneiss consists of varying amounts of quartz, clinopyroxene, actinolite–tremolite, and feldspar.

Quartzite outcrops in an elongate zone cutting the granitic gneiss. The quartzite is interleaved on a 0.3 – 1 m scale with the mylonitized granitic gneiss, and both form a narrow (up to 20 m-wide) mylonite zone, which is possibly heterogeneously recrystallized.

Massive pegmatite dykes and veins cut the granitic gneiss, calc-silicate gneiss, amphibolite, and mylonite zone. Exposures of pegmatite range from very good to very poor, and they are mostly weathered. Where pegmatite has intruded as dykes, the edges are commonly parallel to the well-developed foliation in the gneisses. However, along the length of these structures the pegmatite commonly intrudes adjacent granitic gneiss, cutting across foliation planes. The pegmatites range from medium- to coarse-grained, even-textured granites to even-textured pegmatites, or porphyritic pegmatites. They consist of quartz, feldspar, and muscovite.

Deformation at Locality 3

The Petter Bore area of the Errabiddy Shear Zone has been polydeformed and metamorphosed, and has undergone several periods of granite intrusion (Fig. 13). The earliest recorded structure is a well-developed foliation, S_1, in the granitic gneiss, which is correlated with the regional D_{2g} fabric (Fig. 7). This was only observed at four localities, and is generally intensely deformed by D_2 (correlated with regional D_{1n}) to form a composite S_1/S_2 fabric (Fig. 14). However, in low-strain zones S_1 can be observed as having developed subparallel to the axial surface of a tight to isoclinal fold deforming thin (<3 cm-wide) pegmatite veins.

The most prominent structure in the area is the well-developed composite S_1/S_2 fabric (henceforth denoted $S_{1/2}$; Fig. 14). This foliation trends subparallel to contacts between the granitic gneiss, mylonitized granitic gneiss, quartz mylonite, calc-silicate gneiss, and amphibolite (Fig. 13). The foliation is commonly steeply dipping and strikes east-southeast.

The $S_{1/2}$ fabric is parallel to F_2 fold-axial surfaces that are developed throughout the area. The F_2 folds are commonly isoclinal to rootless and are well developed in the granitic gneiss and the amphibolite. F_2 fold hinges are parallel to L_{12} intersection lineations. Locally, tight F_2 folds deform the granitic gneiss in which thin pegmatite veins have been deformed by F_1 isoclinal folds. Here, crenulation cleavages have developed axial planar to the fold-axial surface of F_2. The F_2 folds plunge moderately to steeply towards the west to northwest, although a few intersection lineations steeply plunge to the northeast or east-northeast, and locally deform both fine-grained granitic veins and pegmatite veins. In these areas, shortening of the folded layer perpendicular to the foliation is typically between 30 and 55%.

Mineral lineations on the $S_{1/2}$ planar surfaces are composed of quartz, mica (muscovite or biotite), or amphibole, depending on the rock type in which they are developed. These mineral lineations form a trend similar to, but less well defined than, the L_{12} intersection lineations, commonly plunging shallowly to steeply towards the west or east.

Pegmatite and minor fine- to medium-grained granite sheets and veins of various ages cut the granitic gneiss, quartz mylonite, amphibolite, and calc-silicate gneiss in the area. The oldest pegmatites cut the precursor granite to the granitic gneiss and are folded into small-scale F_1 folds. These are deformed by D_2 and D_3 structures in the area (Fig. 7). Younger pegmatites commonly trend subparallel to the regional $S_{1/2}$ foliation (Fig. 15), and in some cases intrude $S_{1/2}$ crenulation foliation planes. The pervasive foliation that dominates the other components within this area is mostly absent in the pegmatites, although in places small pegmatite veins cutting this foliation have fold-axial surfaces parallel to the foliation.

Minor, gentle 1 m-scale folds (F_3) deform the $S_{1/2}$ foliation and F_2 folds. These folds are north to northwesterly striking and steeply plunging. In places, boudinage of the $S_{1/2}$ foliation in the granitic gneiss around more-competent calc-silicate gneiss has the appearance of larger scale F_3 folding.

East-northeasterly striking vertical detachments and quartz veins cut the $S_{1/2}$ foliation surface, but are locally subparallel to it. Their relationship to the F_3 folds is unknown. Other brittle and brittle–ductile features include dextral strike-slip, sinistral strike-slip, and reverse faults. Kinematic indicators include brittle–ductile displacement of the $S_{1/2}$ foliation, ductile displacement of $S_{1/2}$ foliation, and S–C fabrics. Dextral strike-slip faults with no observed vertical slip strike either in an east-northeasterly, southwesterly, or northwesterly direction, whereas measured reverse and sinistral strike-slip faults are easterly striking. The kinematics of several fractures could not be ascertained. These are steeply dipping and northerly or westerly trending.

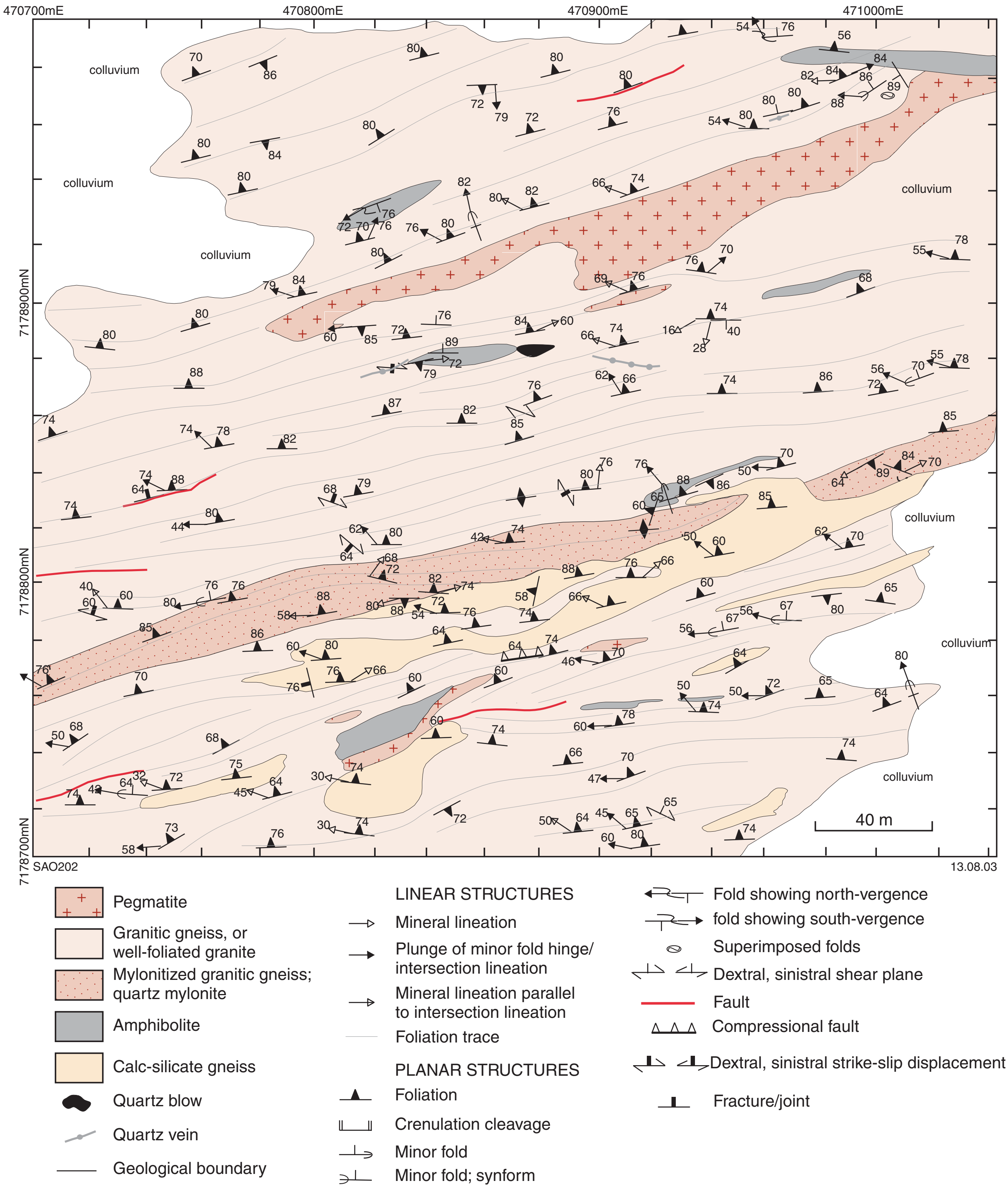

Figure 13. Geological map of Locality 3

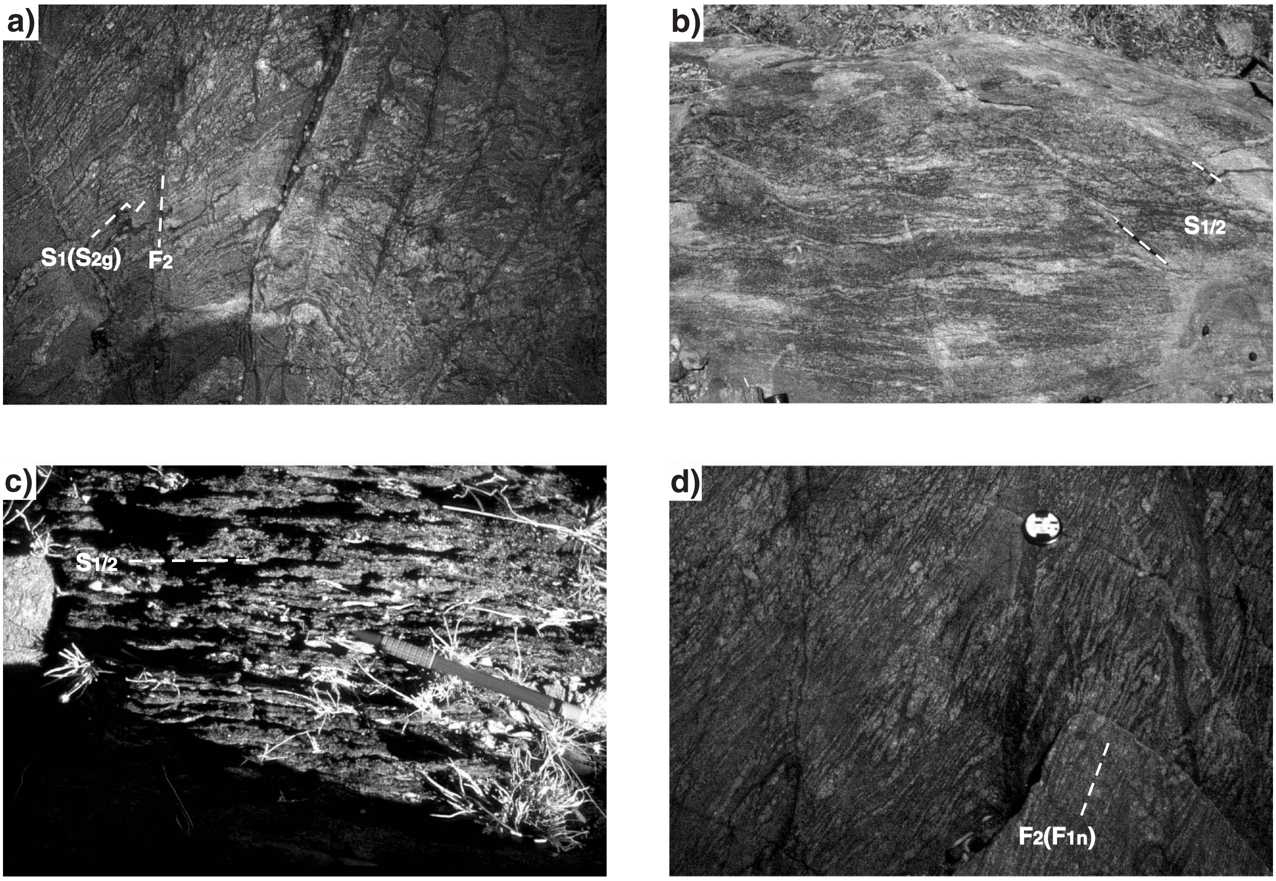

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Figure 14. Rocks and structures at Locality 3: a) S_1 gneissosity in granitic gneiss folded about upright easterly trending, steeply plunging F_2 (correlated with F_{1n} Capricorn-aged folds); b) composite S_1/S_2 foliation in granitic gneiss; c) S_2 crenulations in metamorphosed mafic rock; d) folded pegmatite veins with fold-axial surfaces parallel to S_1 fabric in granitic gneiss

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Figure 15. Pegmatite dyke cutting granitic gneiss and amphibolite

Locality 4: The Petter Calc-silicate of the Camel Hills Metamorphics and the Erong Granite

From Locality 3, 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 south towards the 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 4).

At this locality (Fig. 6) the Petter Calc-silicate of the Camel Hills Metamorphics is intruded by even-textured, medium- to coarse-grained, muscovite–biotite–garnet monzogranite to pegmatite of the Erong Granite, which is thought to be c. 1800 Ma in age.

Locality 4a: 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

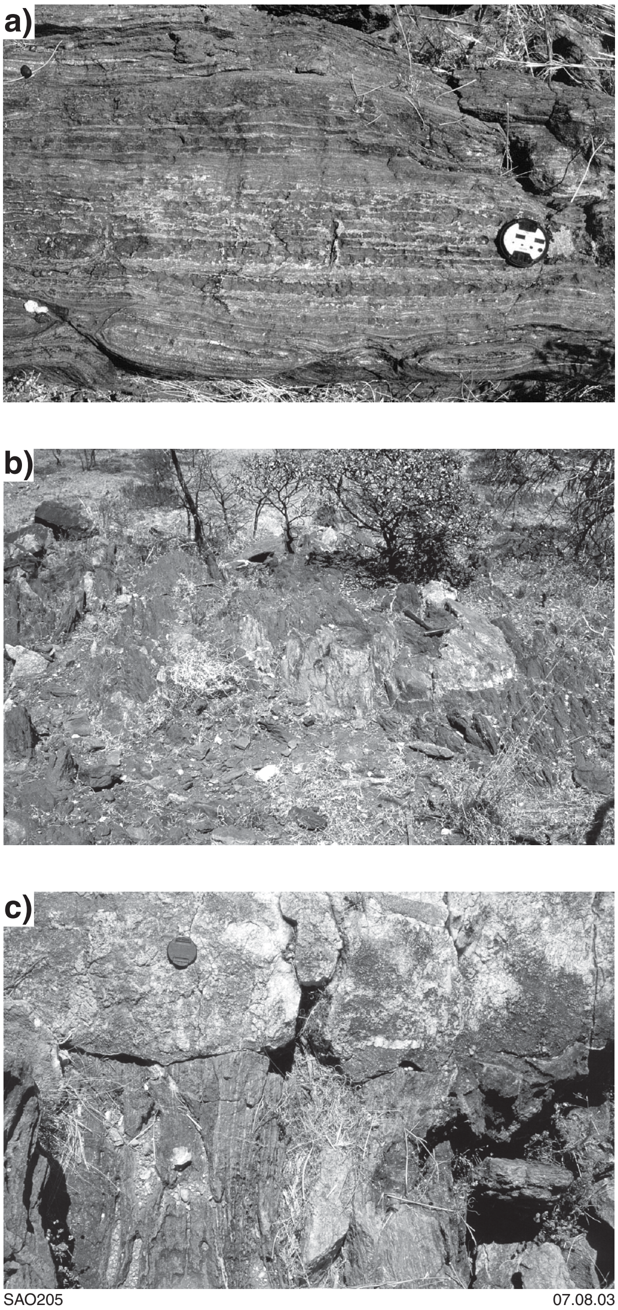

Figure 16. The Petter Calc-silicate at Locality 4a (after Occhipinti et al., 2001):
a) a composite S_0/S_{2g} fabric in the Petter Calc-silicate that is locally boudinaged;
b) a dyke of coarse-grained granite (Erong Granite) cutting the Petter Calc-silicate, but folded with it into a moderately plunging 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

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 contain 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 in the unit, S_1 (correlated with regional S_{2g}), 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-northeast plunging, medium-scale F_2 (correlated with regional F_{1n}) antiform–synform pair (Fig. 7). Locally, a well-developed crenulation cleavage (Fig. 16b) is 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.

The Petter Calc-silicate was sampled 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 that they were probably sourced from the Yilgarn Craton. Three of the zircons have ages older 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 4b: Erong Granite (MGA 465738E 7174598N)

The Erong Granite grades from a medium-grained, even-textured biotite–muscovite–garnet monzogranite into coarse-grained, even-textured granite and pegmatite, and forms numerous granite sheets, dykes, and veins. The granite is typically massive and relatively undeformed, but with a local foliation in narrow low-grade shear zones.

The Erong Granite both cuts and is deformed by c. 1800 Ma D_{1n} structures (Table 1) and is thought to be c. 1800 Ma in age, 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 are mainly Archaean xenocrysts (Fig. 17), perhaps because the low Zr content of the Erong Granite precluded the growth of new zircon.

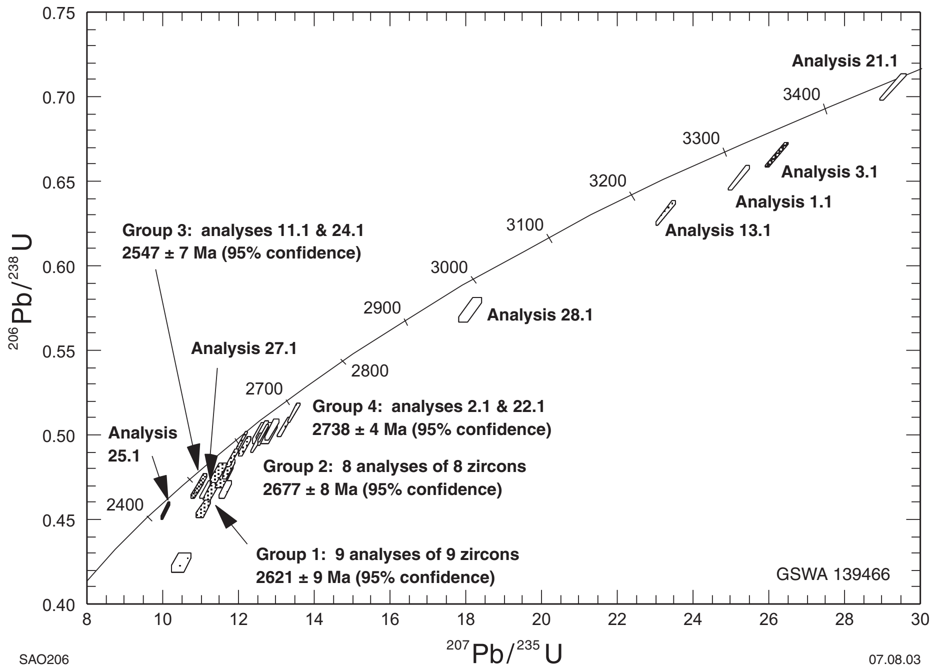

Figure 17. Concordia plot for biotite–muscovite–garnet granodiorite of the Erong Granite from Locality 4b (GSWA 139466; Nelson, 2000). After Occhipinti et al. (2001)

Day 3

Return to Erong Road from Locality 4b. Turn left and follow the road for 800 m. The low rocky outcrop on the left side of the road is Locality 5 (MGA 465011E 7178643N).

Today, we will travel north out of the Errabiddy Shear Zone and into the Glenburgh Terrane of the Gascoyne Complex where we will see variably deformed and metamorphosed granitic components of the Palaeoproterozoic Dalgaringa Supersuite of the Glenburgh Terrane. 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 5: Warrigal Gneiss (MGA 465011E 7178643N; time permitting)

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, 2001; Fig. 18). The gneiss consists of late Archaean granites that intruded the Narryer Terrane of the Yilgarn Craton, and have been deformed and metamorphosed during the Palaeoproterozoic Glenburgh and Capricorn Orogenies.

Structural variation within the Errabiddy Shear Zone

The oldest fabric that can be correlated across the Errabiddy Shear Zone is the main foliation denoted S_1 in Domain 2 of Locality 1, and Localities 2–5. This fabric may be correlated across the Errabiddy Shear Zone by its similar orientation and overprinting relationships, and it having developed at amphibolite facies, and is considered to represent the regional S_{2g} foliation (Occhipinti et al., in press). The main foliation is mostly easterly to east-northeasterly trending, and steeply dipping. However, in areas where it has been least deformed it is northwesterly to northeasterly trending. In Domain 2 of Locality 1, subhorizontal tight to isoclinal F_1 folds have fold-axial surfaces subparallel to the pervasive S_1 foliation in the area.

Mineral lineations on the pervasive foliation surface are most common in Domain 2 of Locality 1, and Locality 3. The orientation of these lineations is highly variable. However, overlap in the orientations of mineral lineations throughout the Errabiddy Shear Zone has been noted (Occhipinti and Reddy, in press). In contrast, there is little orientation overlap between mineral lineations developed in Domain 2 of Locality 1 and the other parts of the Errabiddy Shear Zone.

The variably trending mineral lineations on the pervasive foliation in Domain 2 of Locality 1 are the effect

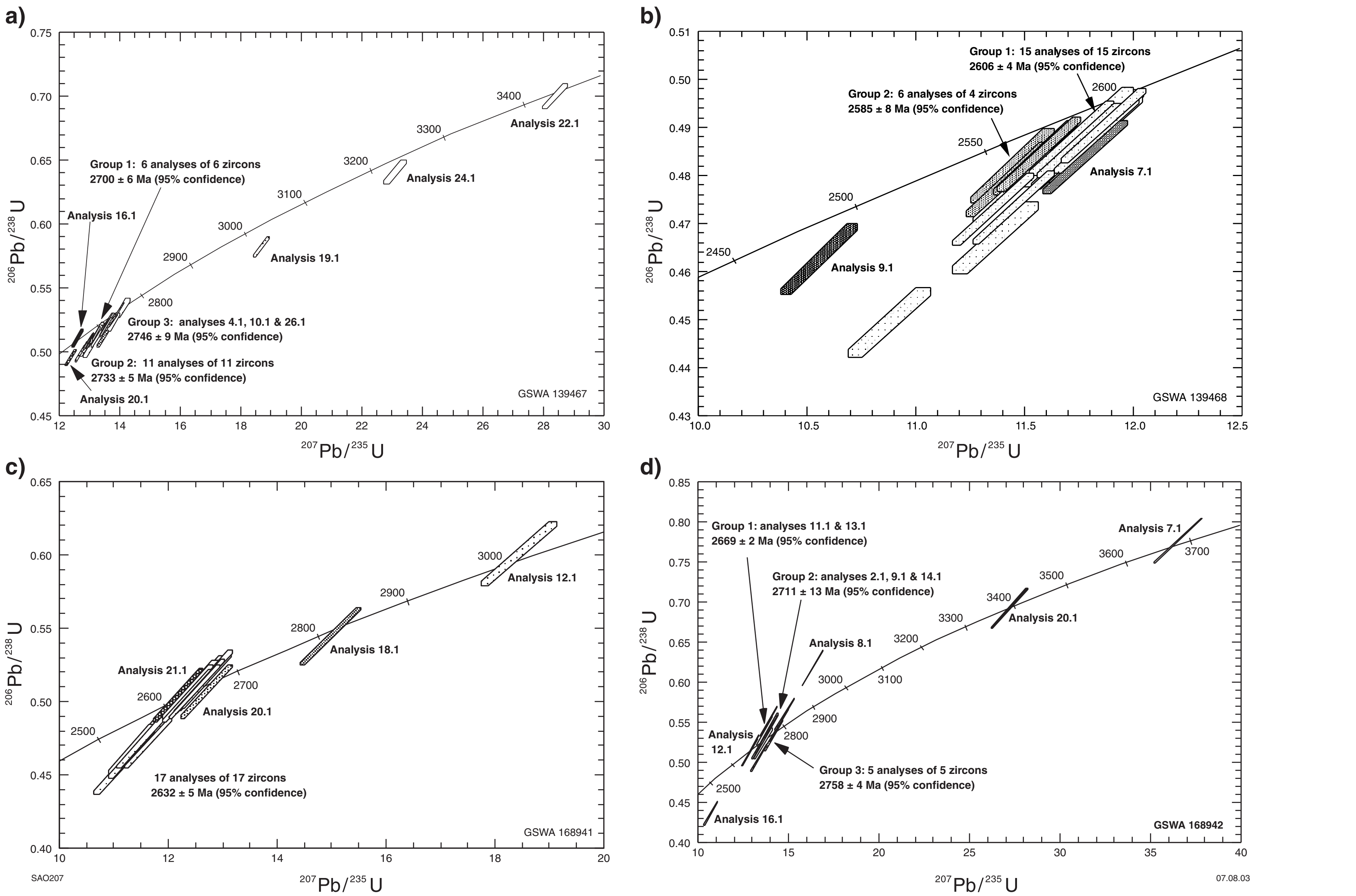

Figure 18. Concordia plots for samples of the Warrigal Gneiss: a) well-foliated porphyritic biotite monzogranite (MGA 466400E 7176400N; GSWA 139467; Nelson, 2000); b) strongly foliated porphyritic biotite monzogranite (MGA 460100E 7176700N; GSWA 139468; Nelson, 2000); c) metamorphosed biotite–tonalite granite (MGA 433700E 7167600N; GSWA 168941; Nelson, 2001); d) porphyritic biotite–muscovite monzogranite (MGA 433700E 7167600N; GSWA 168942; Nelson, 2001). After Occhipinti et al. (2001)

of re-orientation of originally shallowly plunging, possibly northeasterly trending mineral lineations during a subsequent deformation about upright tight to isoclinal curviplanar folds. Elsewhere, mineral lineations also vary in their orientation around younger D_{1n} folds (Occhipinti and Reddy, in press), where the mineral lineations in low-D_{1n} strain zones are oblique to the major fold axes, and those in high-D_{1n} strain zones are subparallel to the major fold axes. This suggests that in higher strain deformation zones, earlier developed mineral lineations may be recrystallized into a new orientation, whereas outside these high-strain zones they remain less affected.

Upright tight to isoclinal folds and crenulations (locally denoted F_2) that deform the main foliation in Domain 2 of Locality 1 and at Localities 2–5 are geometrically similar and developed at greenschist facies during the Capricorn Orogeny (D_{1n}). These folds are generally vertical to subvertical, easterly to east-northeasterly striking, subvertically to shallowly plunging structures with variable fold-axis orientations. A wide range in fold-axis orientations is prevalent in Domain 2 of Locality 1, and locally in other parts of the Errabiddy Shear Zone (Occhipinti and Reddy, in press), although for the most part the fold hinge lines are all fairly similar and generally steeply plunging to the west. In Domain 2 of Locality 1, the variability of the orientation of the fold axes is due to the curvilinear nature of these folds, which is probably the effect of refolding of different parts of a large-scale subhorizontal fold, although they could also be in part constrictional folds with a component of vertical stretch.

The intensity of the upright, tight to isoclinal, easterly to east-northeasterly trending F_{1n} folds between and within the five localities and throughout the Errabiddy Shear Zone is variable, suggesting heterogeneous deformation.

Upright easterly to east-northeasterly trending folds did not develop in Domain 1 of Locality 1; however, the S_1 foliation, which is the dominant structure throughout this domain, is subparallel to the fold-axial surfaces of the tight to isoclinal folds in Domain 2, and also formed at greenschist facies, suggesting that these structures formed coevally.

Mineral lineations related to the formation of the S_1 foliation in Domain 1 of Locality 1 are on both the S_1 and shear band surfaces, slightly oblique to S_1. These mineral lineations have a range in orientations and are shallow to steeply plunging. However, there is no evidence in the field that this range is related to folding of the mineral lineation, suggesting that the movement direction within Domain 1 may not be consistent across the zone.

Detachments that cut the upright tight to isoclinal fold-axial surfaces within the region, but trend subparallel to them, are throughout the Errabiddy Shear Zone and probably also formed during the same deformation event.

Northerly trending open kinks and folds deformed the main easterly to east-northeasterly trending folds throughout the area, with variable intensities.

Kinematic comparison

Kinematic indicators associated with mineral lineations that developed on the main foliation surfaces were not observed in the field, and further microstructural work is being undertaken to determine if there are any kinematics associated with these lineations and to further classify these mineral lineations.

No kinematic indicators have been found to be associated with the upright tight to isoclinal easterly trending folds within the Errabiddy Shear Zone; however, kinematic indicators such as deformed feldspar porphyroclasts and shear bands in Domain 1 of Locality 1 are commonly associated with mineral lineations on the S_1 surface, and are also present in areas where mineral lineations were not recognized (Occhipinti and Reddy, in press). Both dextral and sinistral east-northeasterly trending (along map strike; Fig. 2) strike-slip movements are present throughout the zone; however, dextral strike-slip kinematics are most prevalent. Brittle–ductile normal and reverse faults are locally associated with sinistral or dextral strike-slip movements. These opposing kinematic-sense structures are commonly only a few metres apart and therefore suggest a bulk coaxial flow. The reverse and normal faults cut the along-strike kinematic indicators, indicating that they may have developed slightly later. However, both shear bands and deformed feldspar phenocrysts, and in places a mineral lineation, which all developed at greenschist facies, are associated with the indicators, suggesting that they could have all formed during the same progressive deformation event.

The brittle detachments that trend subparallel to fold-axial surfaces of the easterly trending upright folds commonly show dextral strike-slip movements around them (although they are locally associated with sinistral faulting). The main foliation or shear surfaces associated with these detachments do not commonly have mineral lineations, but displacement of the easterly trending main foliation and folds suggests that the detachments have mostly dextral displacement.

Implications of structural variation within the Errabiddy Shear Zone

The Errabiddy Shear Zone developed as a probably north-northeasterly or northerly trending shear zone (Sheppard et al., in press; Occhipinti et al., 2001, in press) between 2000 and 1950 Ma. During this time D_{1g} and D_{2g} structures formed within the Glenburgh Terrane; however, only D_{2g} structures formed in the Errabiddy Shear Zone. It is known that in the southern part of the Capricorn Orogen, deformation, metamorphism, and magmatism occurred due to oblique approximately north–south shortening caused by either intercratonic or intracratonic processes (Occhipinti et al., in press) between the Archaean Pilbara and Yilgarn Cratons.

The Errabiddy Shear Zone was probably reoriented from its originally north-northeasterly or northerly trend into its approximate present position during the Capricorn Orogeny. The F_{1n} easterly to east-northeasterly trending folds, which were the dominant structures developed during

D_{1n} in the Errabiddy Shear Zone, are not associated with any kinematic movement indicators, and any mineral lineations around these folds appear to have developed prior to their formation. Thus pure shear and shortening probably operated during the early stages of D_{1n} in parts of the Errabiddy Shear Zone. Shortening was approximately north–south directed and 'shear zone boundary-normal' in the areas studied. The foliation in Domain 1 of Locality 1 also developed due to northwest–southeast shortening, but dextral strike-slip also appears to have occurred at the same time in this area.

The variable orientation of F_{1n} fold-axial surfaces that mimic the curvature of the Errabiddy Shear Zone (Fig. 2) is suggestive of either refolding after D_{1n} or variation of the strain geometry throughout the Errabiddy Shear Zone during D_{1n} and therefore during the Capricorn Orogeny. This suggests that a component of simple shear acted in the Errabiddy Shear Zone, with a 'boundary-parallel' component, at the same time as pure shear 'boundary-normal' processes. This scenario is similar to that presented recently by Lin and Jiang (2001), who reported that pure shear dominated in a transpressional shear zone. Dextrally displacing vertical detachment faults that cut the F_{1n} folds at Localities 2–4 suggest that boundary-parallel simple shear also occurred after pure shear and shortening. These observations are consistent with dextral transpression operating in the Errabiddy Shear Zone during D_{1n}.

Northerly trending upright folds that developed after the Capricorn Orogeny only weakly deformed D_{1n} structures and were not responsible for the reorientation of the Errabiddy Shear Zone from approximately easterly striking in the west to northerly striking in the east, producing the bend observed in the shear zone east of the excursion area (Figs 1 and 2). This implies that reactivation did not influence the orientation of D_{1n} structures around this bend in the shear zone, but that a variation of the strain geometry during the Capricorn Orogeny operated throughout the shear zone, thus further supporting transpression occurring during D_{1n}.

Glenburgh Terrane

by

S. A. Occhipinti and S. Sheppard

Introduction

The southern Gascoyne Complex can be divided into three main units: the Glenburgh Terrane, Camel Hills Metamorphics, and 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 (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, 2001), indicate 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 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 component is enigmatic, 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 to Palaeoproterozoic granitic gneiss of the Halfway Gneiss interleaved with metasedimentary rocks of the Moogie Metamorphics (Fig. 22), and the 2000–1975 Ma granitic gneiss and granite of the Dalgaringa Supersuite. The Moogie Metamorphics contain 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 (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).

Siliciclastic sedimentary rocks of the Mount James Formation were deposited in a series of small fault-bound basins on top of the Glenburgh Terrane, Camel Hills Metamorphics, 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 latest Palaeoproterozoic to Mesoproterozoic, sedimentary rocks of the Edmund Group (Bangemall Supergroup) were deposited on the Gascoyne Complex, Yilgarn Craton, Mount James Formation, and in the Bryah and Padbury Basins. The Edmund Group of the Bangemall Supergroup was intruded by latest Mesoproterozoic dolerite sills at 1465 and 1070 Ma (Wingate and Giddings, 2000; Nelson, 2001), and then deformed during the Edmundian Orogeny between c. 1070 and c. 750 Ma (Sheppard and Occhipinti, 2000; Wingate and Giddings, 2000). Late Carboniferous to Early 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 19–20). The supersuite comprises two episodes of magmatism that are separated by a deformation and high-grade regional metamorphic event (Table 1 and Fig. 7). 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 younger foliated to gneissic granites.

The Dalgaringa Supersuite is dominated by diorite, tonalite, and granodiorite, in contrast to most Palaeo-

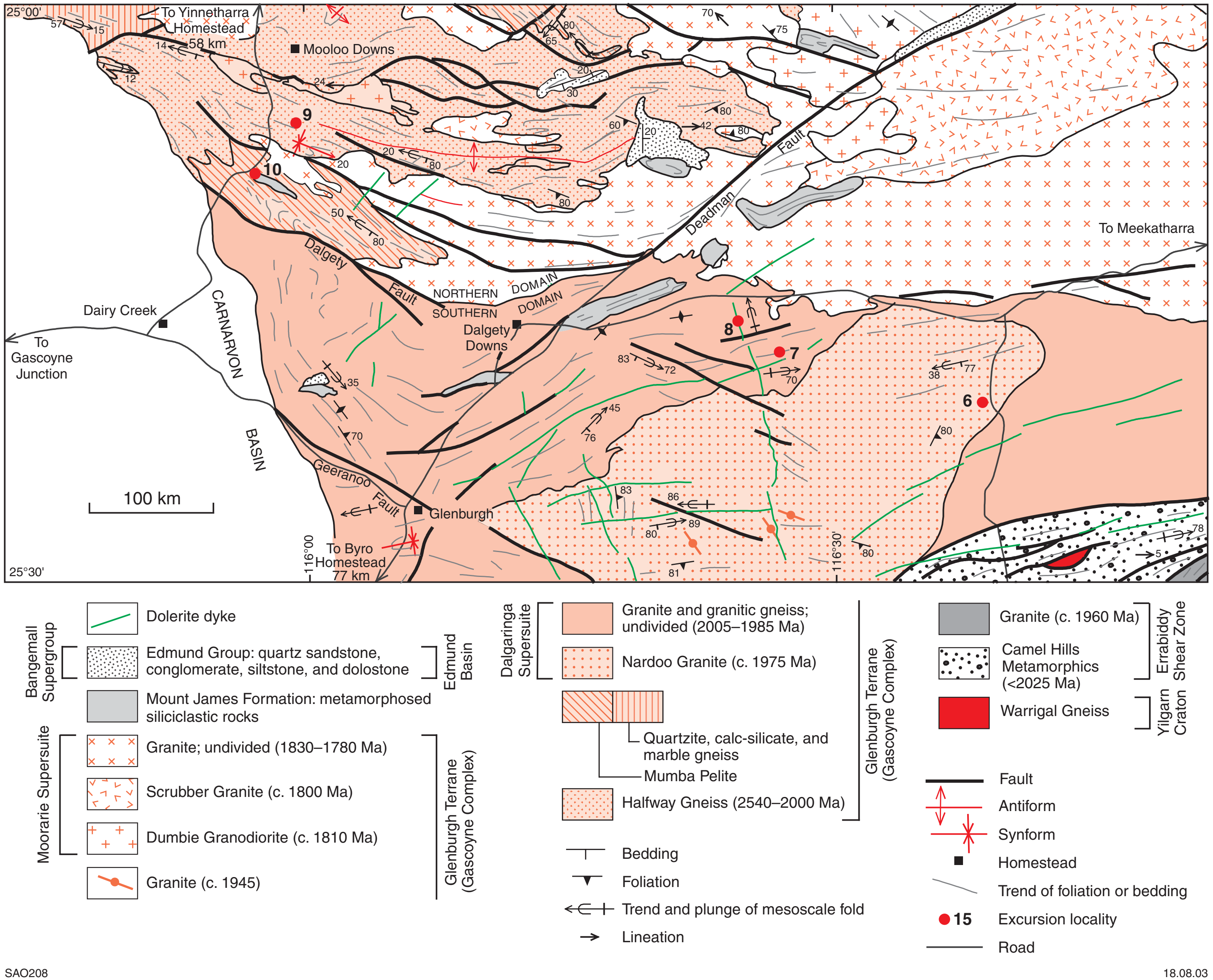

Figure 19. Simplified geological map of the Glenburgh Terrane, showing the northern and southern parts (which correlate with the structural domains of Occhipinti and Sheppard, 2001), the main rock units, and Localities 6–10. Modified from Occhipinti et al. (2001)

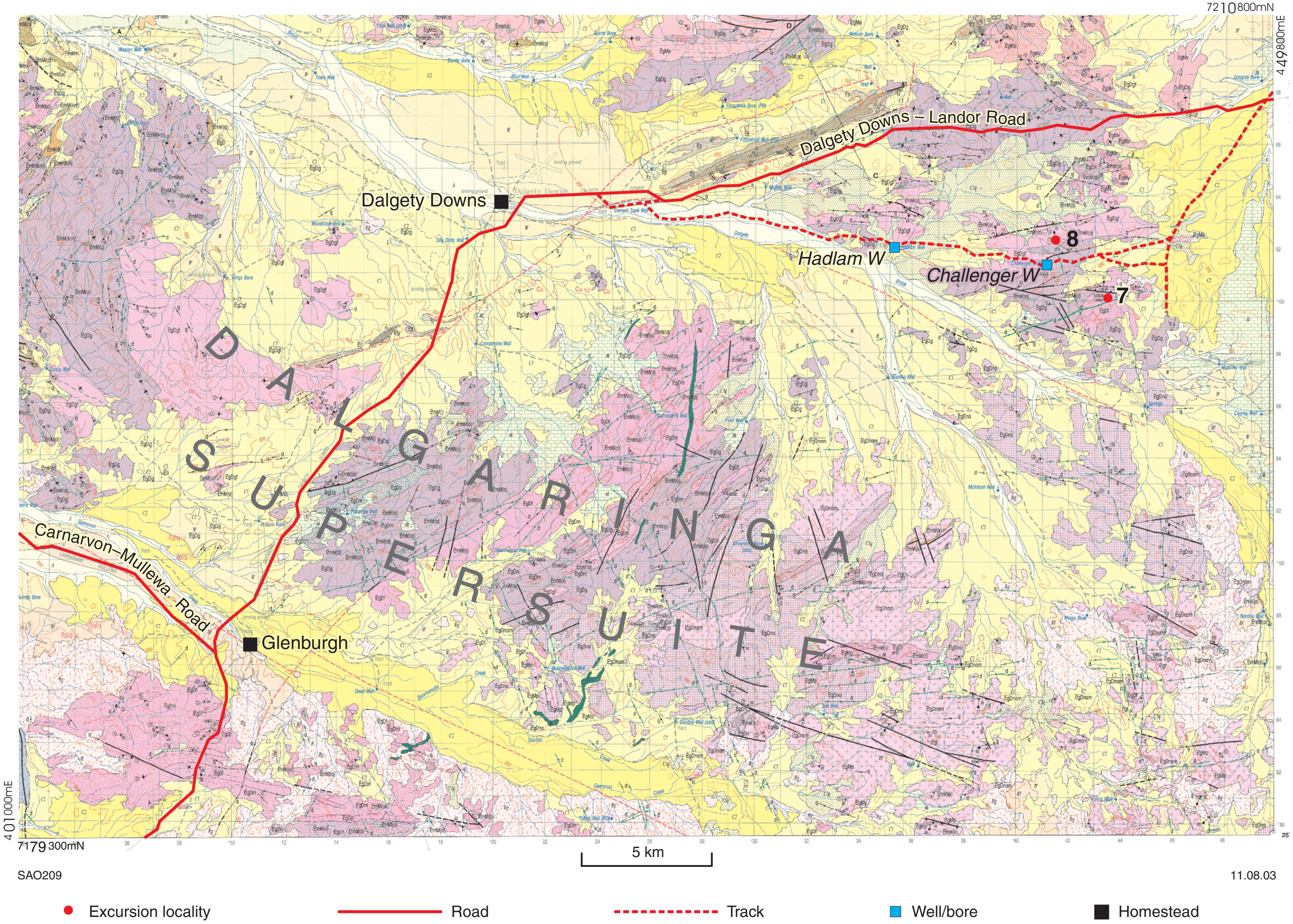

Figure 20. Portion of the Glenburgh 1:100 000-scale geological map (Occhipinti and Sheppard, 2001), showing Localities 7 and 8 and main roads and tracks used on the excursion. After Occhipinti et al. (2001)

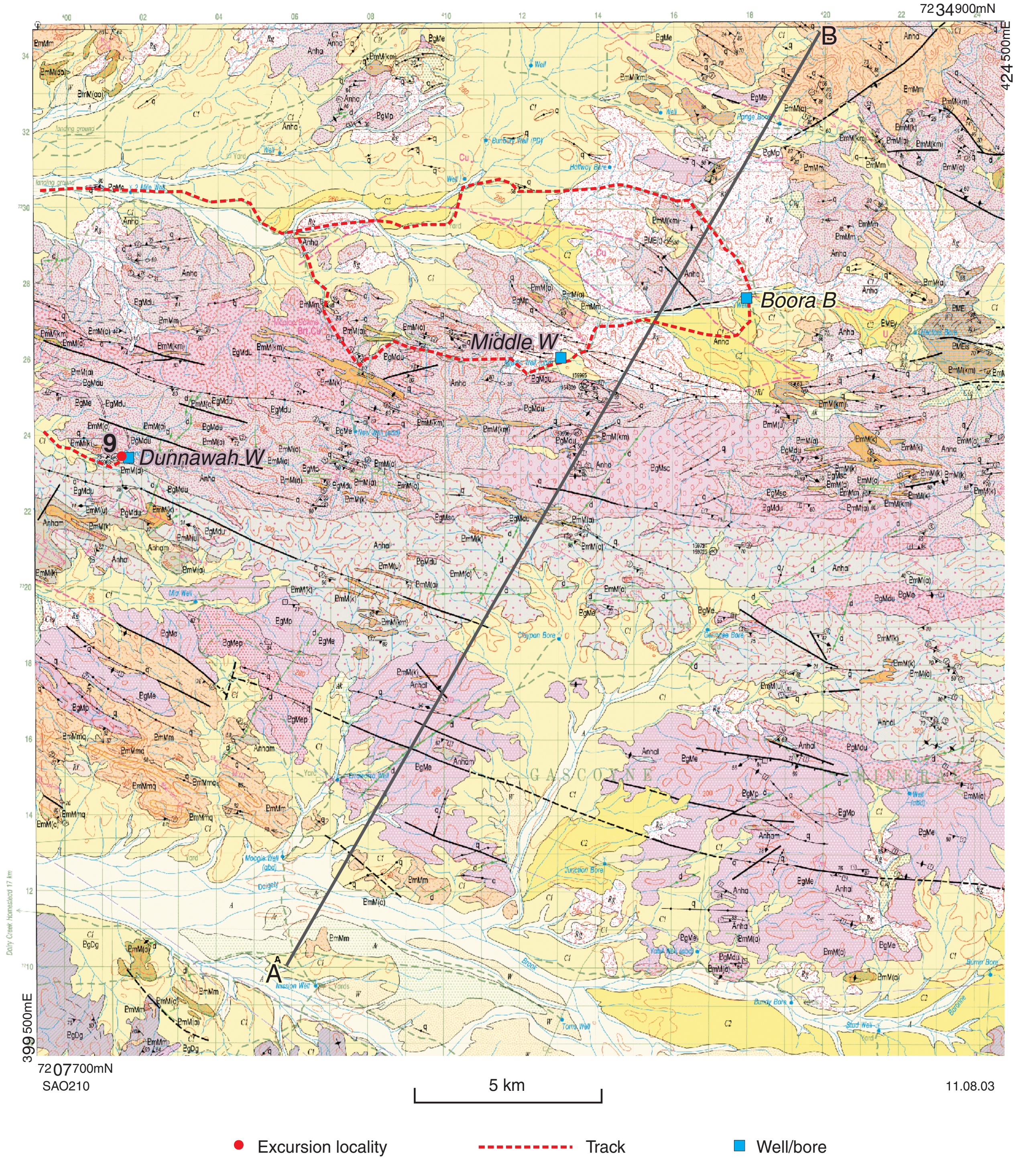

Figure 21. Portion of the GLENBURGH 1:100 000-scale geological map (Occhipinti and Sheppard, 2001), showing Locality 9 and main roads and tracks used on the excursion (after Occhipinti et al., 2001). Cross section A–B is shown in Figure 22

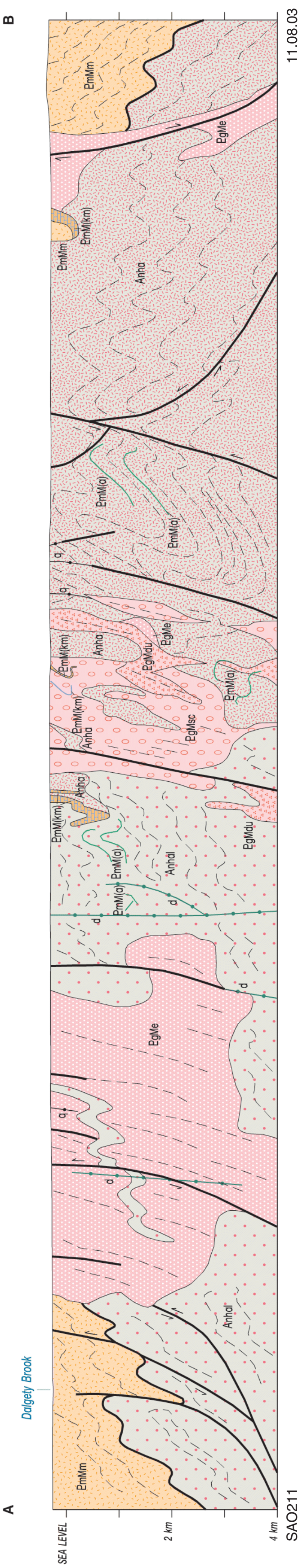

Figure 22. Cross section (see Fig. 21 for location) from the GLENBURGH 1:100 000-scale geological map (Occhipinti and Sheppard, 2001), showing the faulted boundaries between the Halfway Gneiss and metasedimentary rocks of the Moogie Metamorphics. The Halfway Gneiss is interpreted to be folded into an antiform, with granites of the Moorarie Supersuite typically intruded into this structure. EgMsc, EgMdu = Moorarie Supersuite; EmMa, EmMm, EmMkm = Moogie Metamorphics metasedimentary rocks; Anhal = Halfway Gneiss. After Occhipinti et al. (2001)

proterozoic 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 a late Archaean to Palaeoproterozoic continent or microcontinent, which subsequently collided with the passive margin of the Yilgarn Craton.

Locality 6: Nardoo Granite (MGA 465582E 7200762N; time permitting)

From Locality 5 follow Erong Road for 27 km to Locality 6.

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, but 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 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 exposed 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 7: Foliated and gneissic granites of the Dalgaringa Supersuite (MGA 443424E 7200086N)

From Locality 6 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 west-northwest for 1.1 km, and then turn left onto indistinct wheel marks (at MGA 444880E 7200459N). Follow the wheel tracks onto a west-southwesterly trending dolerite dyke for 1.6 km. Locality 7 is in the rocky outcrops 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

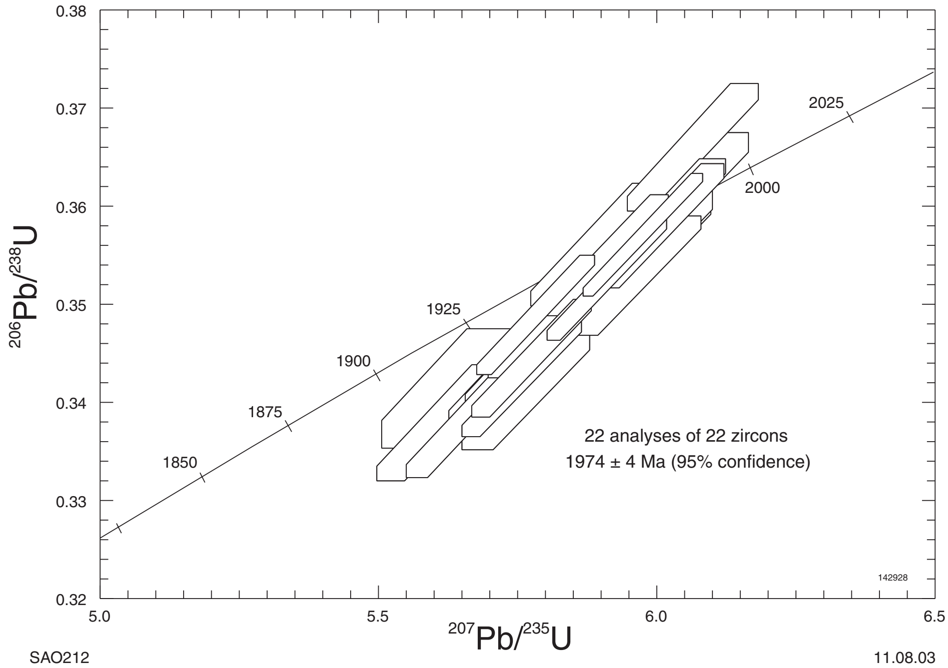

Figure 23. Concordia plot for porphyritic biotite tonalite of the Nardoo Granite (MGA 445000E 7186400N; GSWA 142928; Nelson, 1999). After Occhipinti et al. (2001)

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 in 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 the 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 locally 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 8: Foliated granites of the Dalgaringa Supersuite (MGA 441548E 7202202N)

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 8.

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 7, 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

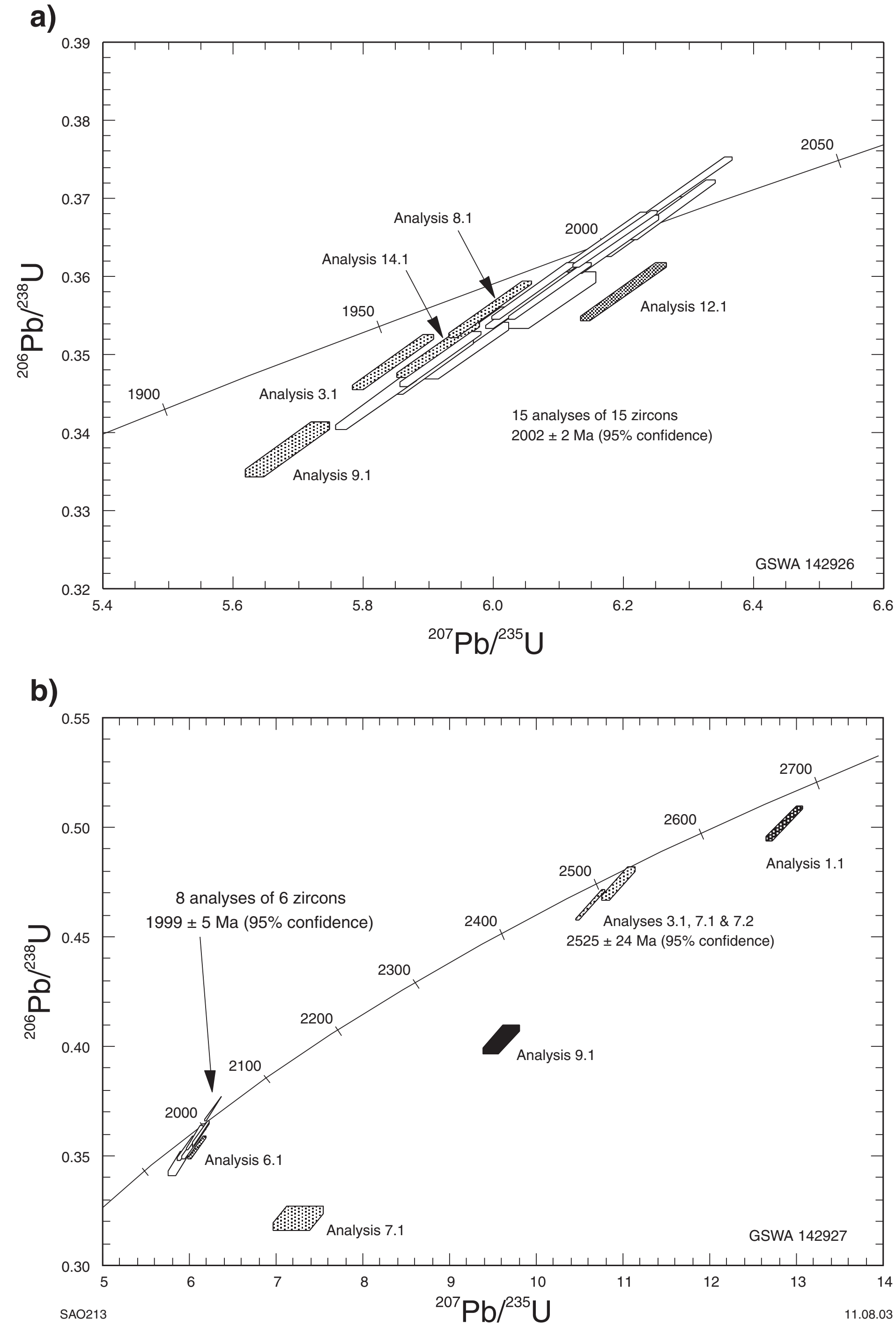

Figure 24. Concordia plots for c. 2000 Ma granite from the Dalgaringa Supersuite: a) foliated biotite tonalite (MGA 443424E 7200086N; GSWA 142926); b) foliated fine-grained biotite–oligoclase granodiorite (GSWA 142927) from 100 m east of GSWA 142926 sample site. After Occhipinti et al. (2001)

granodiorite. The porphyritic mafic granodiorite 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 inclusions are 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.

An L-tectonite fabric 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:

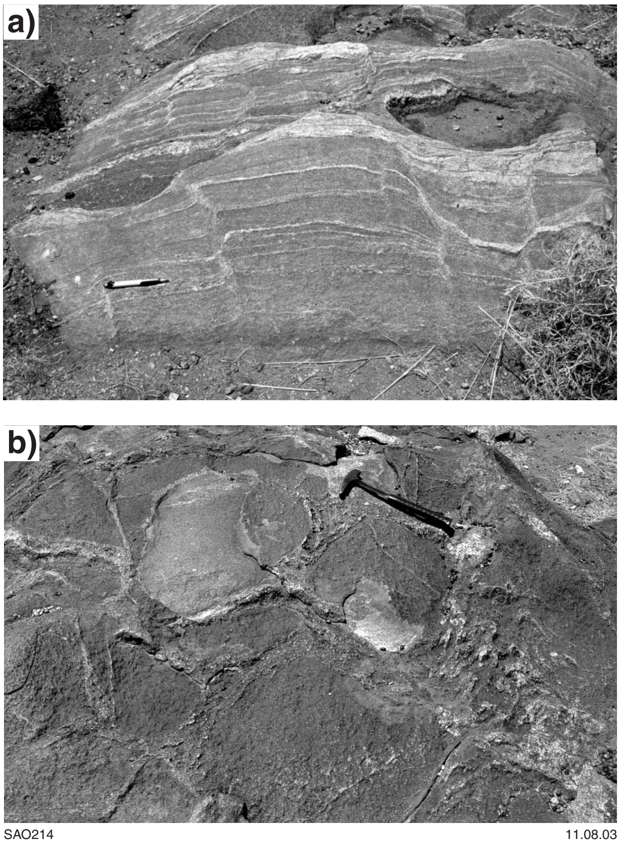

Figure 25. Outcrops of tonalite, granodiorite, and monzogranite from the Dalgaringa Supersuite at Locality 7: a) in the lower half of the photograph a banded granitic gneiss is deformed into box-folds. In the upper part, pinch and swell structures are present (i.e. the weathered-out pillow-like depressions 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. After Occhipinti et al. (2001)

Figure 26. Thin inclusions of tonalite in a mafic granodiorite at Locality 8. This texture has developed because the porphyritic mafic granodiorite intruded rigid, solidified tonalite along a pre-existing tectonic fabric. After Occhipinti et al. (2001)

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

Excursion localities — Halfway Gneiss

Day 4

Today we will look at the latest Archaean to Palaeoproterozoic granitic gneiss of the Halfway Gneiss, the metasedimentary Moogie Metamorphics, younger granite that intrudes them, and the Mount James Formation, which unconformably overlies them. We will also drive north through the central Gascoyne Complex and into the overlying Mesoproterozoic Edmund Basin.

The Halfway Gneiss (*Anhal*) outcrops in an easterly trending 6–10 km-wide belt in the northern part of the Glenburgh Terrane (Figs 19, 21, 22, and 27), 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, which is locally pegmatite banded (Fig. 27). The different rock types are interleaved on both mesoscopic and megascopic scales and contacts between them are typically tectonic, although in 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, 2001; GSWA 168947, 164309, 142988, and 168950; Fig. 28) has demonstrated that the gneiss has both latest Archaean and Palaeoproterozoic granitic components. Samples from Locality 9

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Figure 27. Examples of different components in the Halfway Gneiss from the northern domain of the Glenburgh Terrane (Gascoyne Complex; after Occhipinti et al., 2001):
a) pavement of augen gneiss, or well-foliated porphyritic granite intruded by probable c. 1800 Ma leucocratic pegmatite (at MGA 413630E 7225305N). A sample for geochronology (GSWA 164309) was taken about 3 m south of this site;
b) pegmatite-banded tonalite gneiss (mesocratic granitic gneiss) from Dunnawah Well, Locality 9. The geological 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

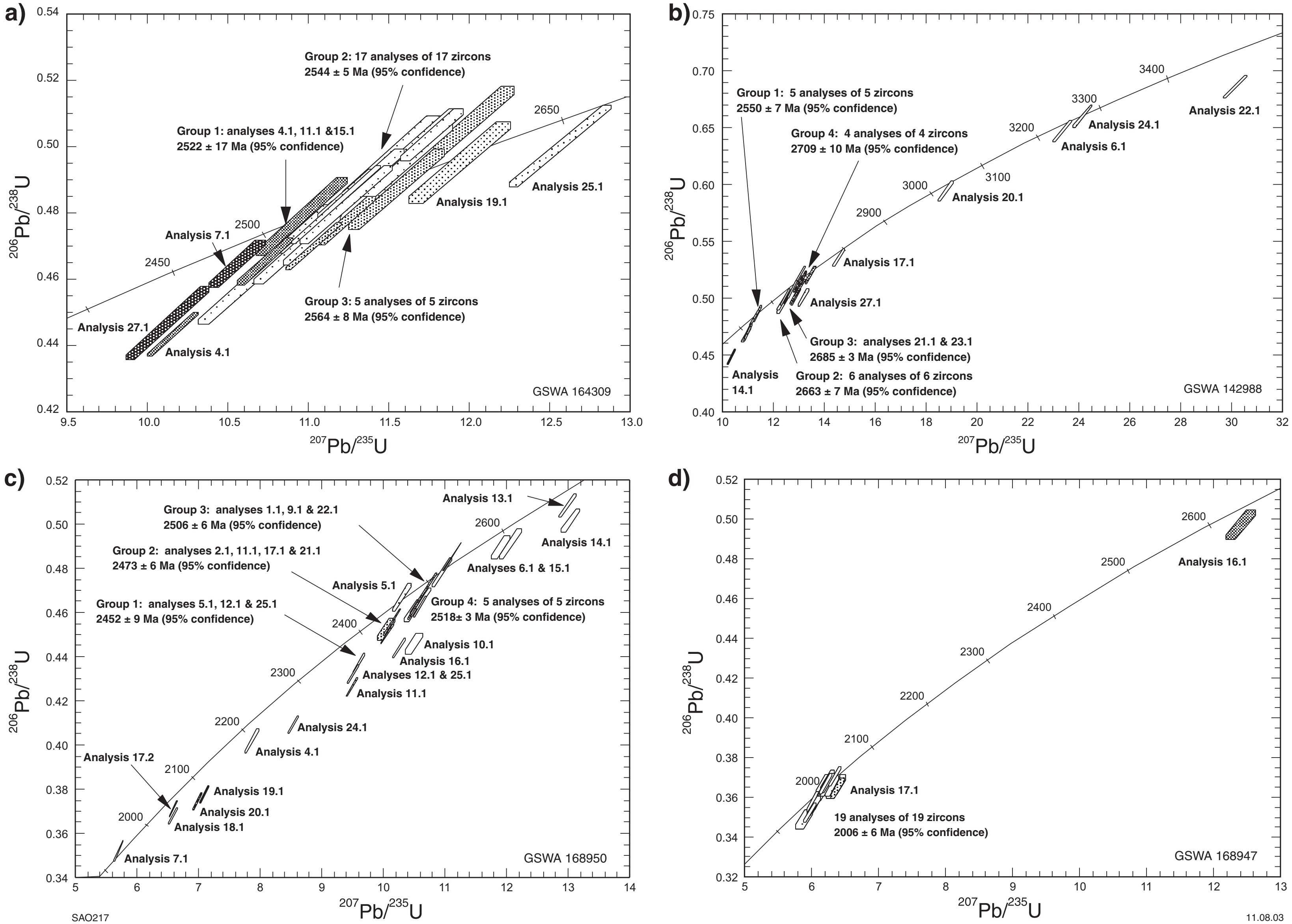

Figure 28. Concordia plots for components of the Halfway Gneiss: a) biotite augen gneiss (GSWA 164309; MGA 413616E 7225303N); b) mesocratic, banded granitic gneiss from Locality 9 (GSWA 142988; MGA 401309E 7223430N); c) pegmatite-banded tonalite gneiss (GSWA 168950; MGA 362200E 7165700N) from the Carrandibby Inlier; d) leucocratic granitic gneiss from 2 km west of Weedarra Homestead (GSWA 168947; MGA 381950E 7230840N). After Occhipinti et al. (2001)

have igneous crystallization ages, as determined by SHRIMP U–Pb zircon analyses, of 2544 ± 5 and 2550 ± 7 Ma (Fig. 28a,b). Nelson (2001) analysed a pegmatite-banded tonalite gneiss (GSWA 168950; Fig. 28c) in the Carandibby Inlier (Fig. 2), which Nutman and Kinny (1994) had previously dated at c. 2500 Ma. These analyses by Nelson (2001) contained four main zircon populations at 2452 ± 9, 2473 ± 6, 2506 ± 6, and 2519 ± 3 Ma (Fig. 28c). 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.

In addition to latest Archaean granite protoliths, Nelson (2001) also identified Palaeoproterozoic protoliths in a sample of leucocratic granitic gneiss (GSWA 168947) from 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. 28d; Nelson, 2001). 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, but 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 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 for 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, the widespread micrographic and myrmekitic textures, and replacement of magnetite and ilmenite by epidote and titanite, respectively, in association with plagioclase alteration.

Locality 9: Pegmatite-banded gneiss of the Halfway Gneiss (MGA 401309E 7223430N)

From Locality 8 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 right and travel 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 to an indistinct station track to the east (at MGA 401409E 7223424N), and travel east for 5.5 km to Dunnawah Well. Park, and walk west along the northern side of the creek for 100 m to a low pavement (Locality 9).

The 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 (Fig. 27b,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 constitute 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. 27b). The sample contains several concordant zircon populations (Fig. 28b). 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 constitute less than 5% of the sample, the youngest population at 2550 ± 7 Ma probably represents the igneous crystallization age of the tonalite precursor (Figs 27b and 28b). 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, 2001), 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 the gneissic layering. Some small, vertically plunging folds are possibly associated with the shear zones.

Excursion localities — Moogie Metamorphics and Mount James Formation

The Moogie Metamorphics (Occhipinti and Sheppard, 2001) include pelitic and psammitic schist (Mumba Pelite), quartzite, calc-silicate gneiss, marble, amphibolite, ultramafic schist, and metamorphosed banded iron-formation (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., 1983a; 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., 1983b). 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. It is also probable that not all the metasedimentary rocks, or the metamorphosed mafic and ultramafic igneous rocks in the Gascoyne Complex, are of 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, forming 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 of which are commonly iron rich), it also locally contains minor quartzite and metamorphosed granular iron-formation. The unit is locally tectonically interleaved with amphibolite and ultramafic schist, but 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 1), 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 or a quartzite outcropping at the base (Sheppard and Occhipinti, 2000). The formation consists of quartzite, meta-arkosic sandstone to quartz–sericite phyllite, meta-quartz 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 are common in the quartzite.

The nature and origin of the Mount James Formation (Drew, 1999a,b) is problematic. Hunter (1990) suggested that a correlation 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, could be made 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 because the Padbury Group is more complexly deformed. It is also probable that the younger-than-2000 Ma Padbury Group is older than the Mount James Formation (see below). Additionally, the Padbury Group outcrops in the Padbury Basin, which is 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 (2001) found that SHRIMP U–Pb ages on detrital zircons in a sample of quartzite (GSWA 168937) from the Mount James Formation had two populations of zircons: c. 1960 and 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 c. 1640 Ma 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 lower greenschist-facies conditions, and therefore its sedimentary protolith is easily discernible.

Locality 10: The Mumba Pelite of the Moogie Metamorphics, and the Mount James Formation

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 two creeks (at MGA 393585E 7218251N and MGA 393617E 7218213N). Park about 50 m north (uphill) of the creek. From this point walk in a northeasterly direction over the hill towards Locality 10a.

At Locality 10 (Figs 19 and 29) 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.

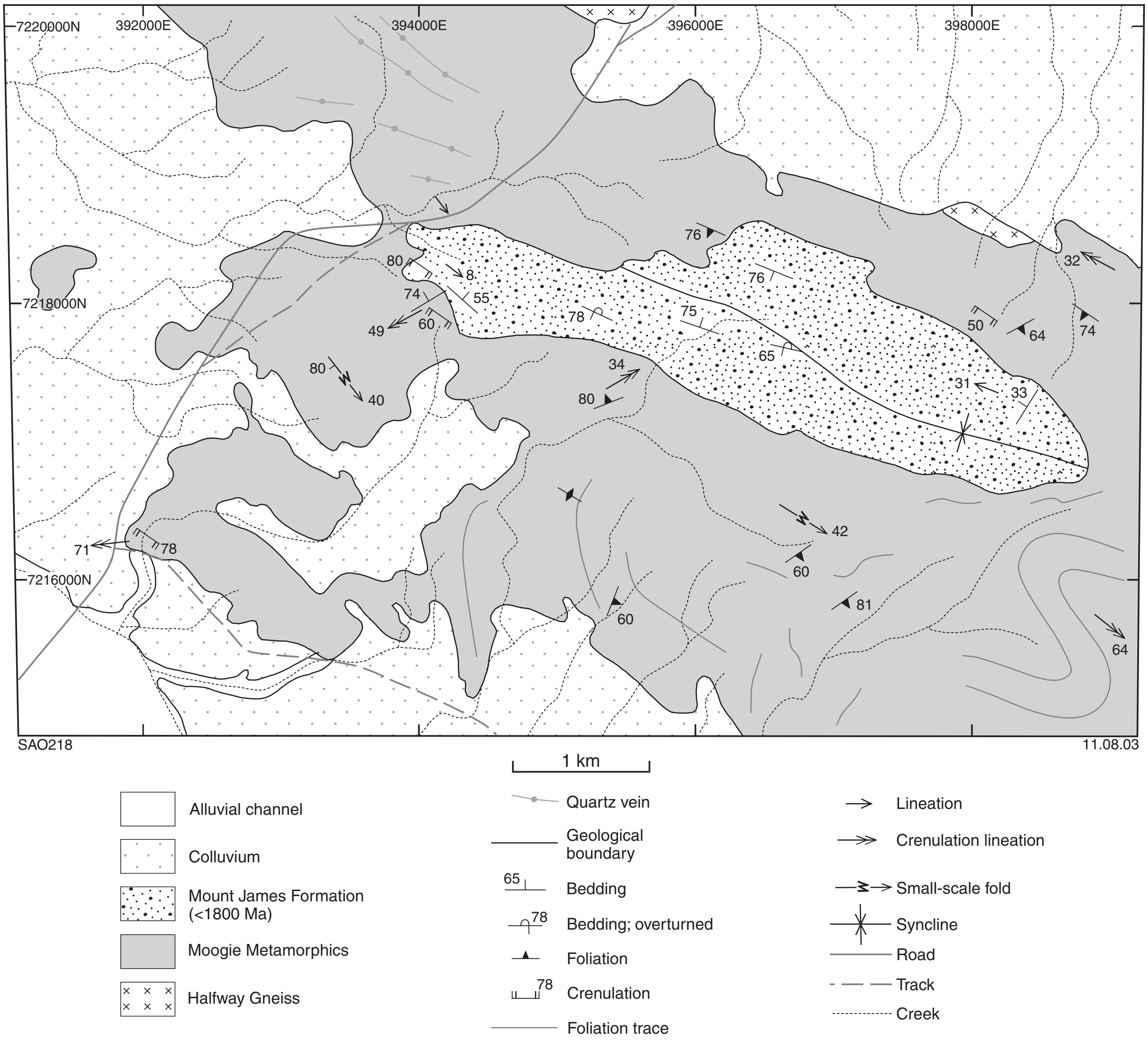

Figure 29. Simplified geological map of the area around Locality 10, showing the distribution of the Mount James Formation and the Moogie Metamorphics. Modified from Occhipinti et al. (2001)

Locality 10a: 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. 30). 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 metamorphosed quartz sandstone that contains clots of opaque minerals (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 1). 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 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 10, where 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 less than 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 10b: 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 at greenschist facies, unconformably overlies the Mumba Pelite of the Moogie Metamorphics. Although sedimentary structures such as bedding (Fig. 31) and fining-up sequences are present, the metaconglomerate is mostly massive.

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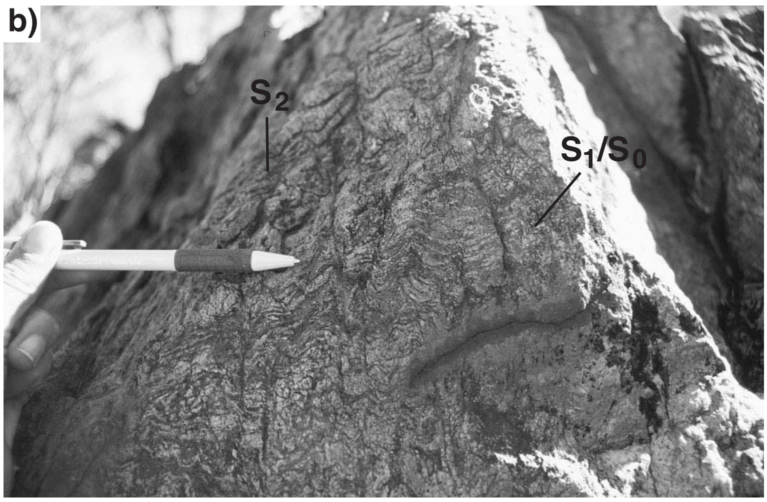

Figure 30. Outcrops of the Mumba Pelite of the Moogie Metamorphics: a) compositional layering defining bedding at Locality 10a (MGA 394192E 7217883N). Dark elongate blotches (well developed on left-hand side) are magnetite–quartz aggregates that probably defined an original bedding, S_0, but have now 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 10a

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, but 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.

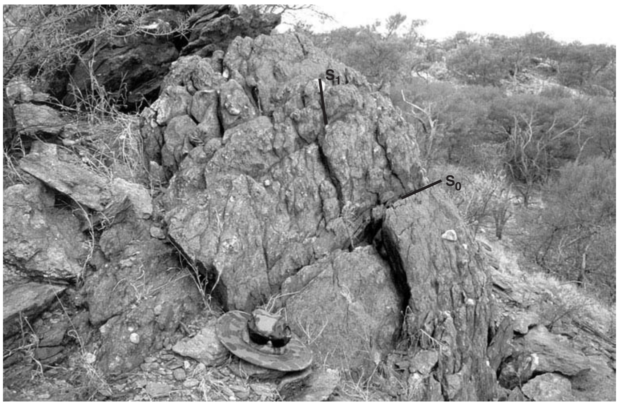

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Figure 31. Bedding in moderately well foliated Mount James Formation metaconglomerate and pebbly sandstone at Locality 10b. After Occhipinti et al. (2001)

Central Gascoyne Complex

by

S. A. Occhipinti

Introduction

The Glenburgh Terrane of the Gascoyne Complex is separated from the central Gascoyne Complex by the west-southwesterly trending Chalba Fault. Much of the central Gascoyne Complex consists of granite sheets and plutons, which are both interleaved with and intrude metasedimentary rocks and minor metamorphosed mafic and ultramafic rocks.

Granitic rocks of the central Gascoyne Complex consist largely of monzogranite, syenogranite, and granodiorite of the 1830–1780 Ma Moorarie Supersuite and the 1680–1620 Ma Durlacher Supersuite (Sheppard et al., 2000). Pelitic and semipelitic metasedimentary rocks of the Morrissey Metamorphic Suite outcrop over the central Gascoyne Complex and their sedimentary protolith has a maximum age of c. 1840 Ma (Varvall, 2001). These rocks are interleaved with granite sheets and contain a pervasive south-dipping foliation, which transposes bedding and is axial planar to small-scale folds (Varvall, 2001). The Morrissey Metamorphic Suite has been variably metamorphosed in the central part of the Gascoyne Complex, with metamorphic facies ranging from greenschist through to middle amphibolite (garnet–staurolite–biotite, staurolite–biotite–kyanite, and biotite–kyanite–chlorite assemblages in pelitic schist, which formed at 7–10 kbar and 600–700°C; Varvall, 2001) from north to south.

Changing metamorphic mineral assemblages in pelitic rocks of the Morrissey Metamorphic Suite from chlorite–muscovite–quartz in the north, just south and adjacent to low-grade rocks of the overlying Mesoproterozoic Edmund Group, to garnet–biotite–chlorite–muscovite–quartz over a distance of about 10 km to the south suggests a steady increase in metamorphic grade with no apparent structural break in between (Culver, 2001). Prograde metamorphism to greenschist facies in the north and amphibolite facies to the south was concomitant with the development of the dominant S_2 foliation within the metasedimentary rocks (Varvall, 2001; Culver, 2001).

After the formation of the dominant S_2 foliation, slightly elongate granite plutons intruded an apparent east-southeasterly lineament that separates the low-grade pelitic rocks in the north and middle amphibolite-facies pelitic rocks in the south (Culver, 2001). This suggests that the lineament might represent a fault or shear zone (Culver, 2001). One of these elongate granite plutons was dated at c. 1650 Ma, by U–Pb SHRIMP analyses of zircon (Culver, 2001), which is within the age range for the regionally significant 1680–1620 Ma Durlacher Supersuite (Sheppard et al., 2000). This suggests that regional deformation and prograde metamorphism occurred in the central Gascoyne Complex prior to c. 1650 Ma and possibly during the Capricorn Orogeny between 1840–1780 Ma.

Rocks of the Mesoproterozoic Edmund Group are both in faulted contact with and unconformably overlie parts of the central Gascoyne Complex and are discussed in **Edmund and Collier Groups**.

Excursion locality — central Gascoyne Complex

Locality 11: Porphyritic granodiorite, Yinnetharra Homestead

From Locality 10 return to the Cobra – Dairy Creek Road. Drive north for about 69 km, past Mooloo Downs Homestead towards Yinnetharra Homestead. Stop at a creek crossing just south of the homestead (you can see the homestead from the road).

A porphyritic biotite granodiorite pluton is well exposed in the creek crossing just south of Yinnetharra Homestead. The granite is not foliated, but contains a well-developed L-tectonite fabric. The age of the granite is unknown. It could be part of the 1680–1620 Ma Durlacher Supersuite or the 1830–1780 Ma Moorarie Supersuite. The granite looks similar to both the c. 1810 Ma Dumbie Granodiorite, which outcrops as abundant sheets and elongate east-southeasterly trending plutons to the south in the Glenburgh Terrane around Mooloo Downs Homestead, and to parts of the c. 1670 Ma Pimbyana Granite, which outcrops to the north around the Gifford Creek and Wanna Homesteads. Both the Dumbie Granodiorite and the much younger Pimbyana Granite are heterogeneously deformed, although an L-tectonite fabric is common in the Dumbie Granodiorite, but has not been observed in the Pimbyana Granite, which appears to more commonly contain an igneous flow foliation.

Edmund and Collier Groups

by

D. McB. Martin and A. M. Thorne

Introduction

The Mesoproterozoic Bangemall Supergroup (Fig. 32), comprising 4–10 km of mostly fine-grained siliciclastic and carbonate sedimentary rocks, is the youngest depositional element in the Capricorn Orogen. The Bangemall Supergroup is divided into the c. 1620–1465 Ma Edmund Group and the c. 1400–1070 Ma Collier Group (Figs 2 and 32). Dolerite sills intruded the Edmund Group at c. 1465 Ma, and both the Edmund and Collier Groups at c. 1070 Ma.

Deposition of the Bangemall Supergroup was strongly controlled by syndepositional faults parallel to the trend of major structures in the underlying Ashburton Fold Belt and Gascoyne Complex. These structures include the Talga Fault, the faulted southwestern limb of the Wanna Syncline, the Lyons River Fault, and the northwest-trending fault system that bounds the Mangaroon and Ti Tree Synclines.

The Edmund Group can be divided into four depositional packages consisting of carbonate and siliciclastic shelf to basinal facies that were strongly to moderately influenced by these syndepositional faults. Source areas for the siliciclastic rocks were to the northwest, northeast, and southeast.

The Collier Group, on the other hand, comprises two depositional packages consisting of deltaic to deep-marine siliciclastic facies derived from the northeast and southeast, which are only mildly influenced by these structures. The depositional packages are bound by basal unconformities or major marine flooding surfaces.

Reactivation of basement structures during the Edmundian Orogeny resulted in open to tight upright F_1 folding of the Bangemall Supergroup. These folds trend east to southeast and are tightest above basement faults. Intrusion of the 755 Ma Mundine Well dyke swarm post-dates F_1, but was followed by localized tightening of F_1 under dextral transpression in the Mangaroon Syncline. A less prominent set of upright open, northeast-trending F_2 folds has refolded F_1 into open, doubly plunging structures.

This excursion will visit two localities in the Edmund Group, between Cobra Homestead and Mount Augustus. The first is in the tightly folded Cobra Syncline, in which the lower to middle Edmund Group (Fig. 32) is preserved. The second is at the unconformity between the Mount Augustus Sandstone and the Gascoyne Complex, at the western end of Mount Augustus.

Excursion localities — Edmund and Collier Groups

Locality 12: Cobra Syncline (MGA 444083E 7323493N)

From Locality 11 continue driving north along the Cobra – Dairy Creek Road for about 70 km. Stop (at MGA 444083E 7323493N) where the Discovery Formation of the Edmund Group is well exposed on both sides of the road.

Tight, upright-folded laminated siltstone of the Discovery Formation (middle Edmund Group) is well exposed on both sides of the Dairy Creek – Mount Augustus Road, about 2.5 km west of Cobra Homestead. This locality is in the core of the Cobra Syncline, on the southern limb of the Bangemall Anticline. Stratigraphic evidence indicates that bedding is overturned here (S_0=136/86°N) and youngs to the south. Cleavage is generally subparallel to bedding, and forms a well-developed intersection lineation (098/81) on S_0, but also truncates upright intrafolial folds (Fig. 33). Strain partitioning into the Discovery Formation, particularly in tightly folded areas, is common in the Bangemall Supergroup.

Locality 13: Mount Augustus (MGA 479000E 7312800N)

From Locality 12 continue driving along the road west past Cobra Homestead. Travel for about 28 km towards Mount Augustus to a south-trending fenceline. Turn right onto the fence track and travel towards a well. At the well turn onto a southeasterly trending track. Travel along this track for about 4 km, then east for about 2 km to Locality 13.

Here, the contact between the Gascoyne Complex and pebbly sandstone of the Mount Augustus Sandstone is exposed. The Gascoyne Complex consists of granite, and phyllite of the Mount James Formation. Correlation of the Mount Augustus Sandstone is unclear at this stage. Facies and palaeocurrent data suggest correlation with the Bresnahan Group, but stratigraphic evidence from the southern limb of the Cobra Syncline supports correlation with the Yilgatherra Formation at the base of the Bangemall Supergroup.

Mount Augustus is an asymmetrical anticline with steep dips on the northern limb and shallow dips on the southern limb. An anastomosing S_1 fracture cleavage striking 099/71°S is locally preserved in the sandstone and is parallel to prominent aerial photograph lineaments at the eastern end of Mount Augustus.

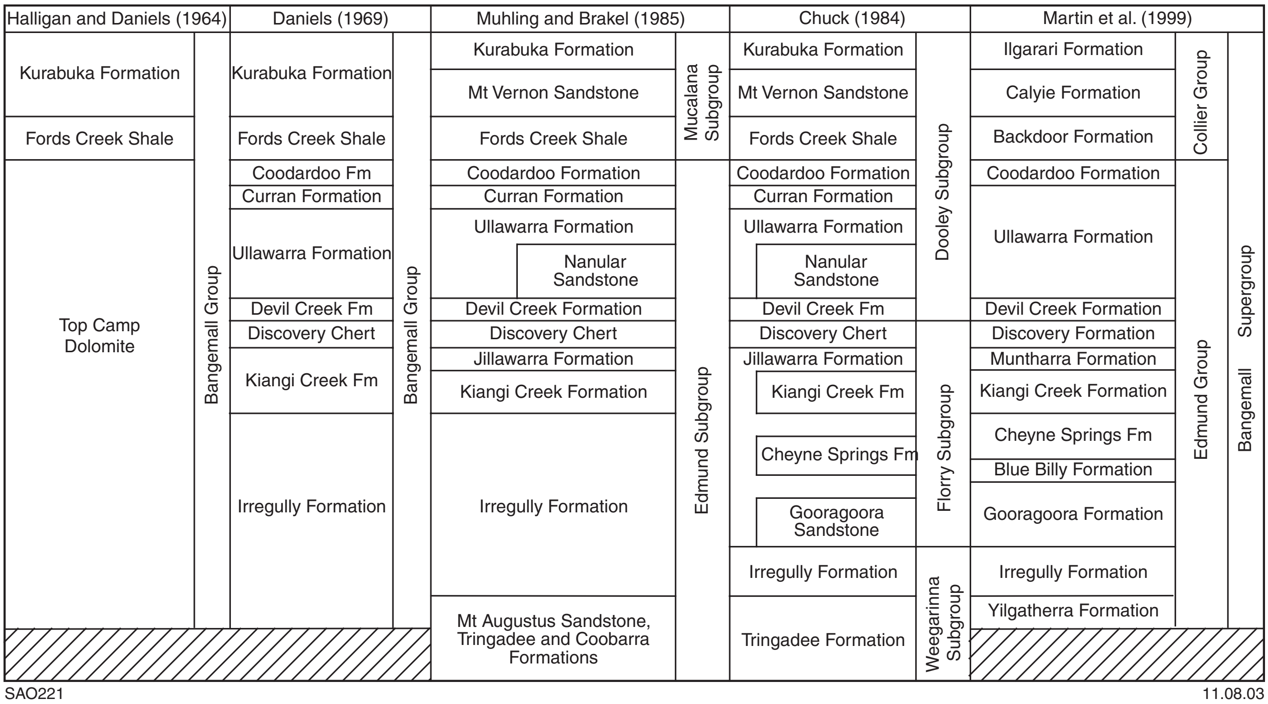

Figure 32. Evolution of stratigraphic nomenclature for the Bangemall Supergroup. After Martin and Thorne (2002)

Figure 33. Upright intrafolial fold within cleaved Discovery Formation on the southern limb of the Bangemall Anticline. Bedding is overturned, and younging is to the left

Ashburton Province

by

K. N. Sircombe

Introduction

The Proterozoic Ashburton Province lies in the northern part of the Capricorn Orogen. It defines an arcuate structure spanning about 450 km from near Pannawonica in the northwest to Turee Creek Homestead in the east (Fig. 34). The term 'Ashburton Province' is used here for the purpose of broad geographical description of all stratigraphy between the Hamersley and Bangemall Basins, including the lower and upper Wyloo Groups, Capricorn Group, Mount Minnie Group, and Bresnahan Group. Stratigraphy and age constraints are outlined in Figure 35. Lower units of the Ashburton Province overlie the southern margin of the Hamersley Basin in the Pilbara Craton. A series of smaller successor basins (Mount Minnie, Blair, and Bresnahan Basins) in the central and southern sections of the Ashburton Province have a variety of structural relationships. The contact between these basins is commonly a significant angular unconformity and typically forms prominent escarpments. The southern exposure of the Ashburton Province is also marked by a prominent unconformity and escarpment with the regionally extensive and generally subhorizontal Bangemall Superbasin.

The Ashburton Province commonly consists of low-grade metasedimentary and metavolcanic rocks. Metamorphic grade tends to increase towards the west where possible equivalents in the Gascoyne Complex are termed the Morrissey Metamorphic Suite. Contact metamorphism with the Boolaloo Granodiorite further complicates the geology in the western margin of the basin.

Two general phases of deformation, referred to as the Ashburton Fold Belt, are recognized in the broader Ashburton Province, although lower units in the succession on the northern margin display structures related to the Opthalmia Fold Belt. The Mount Minnie and Blair Basins only record one phase of deformation, but their distribution is strongly controlled by associated northwest-trending faults. The Bresnahan Basin records neither phase of deformation and is tectonically controlled by a later phase of extensional deformation.

Known mineral resources in the Ashburton Province are relatively limited, with minor, but numerous, gold, copper, lead and silver mineralization commonly associated with D_{2a} quartz veins. These small deposits were commercially exploited mainly in the late 19th and early 20th Centuries, with a few remaining low-key operations at present. The recent discovery and development of the Mount Olympus gold deposit southeast of Paraburdoo has renewed interest in the mineralization potential of the region.

From a geological history perspective, the Ashburton Province provides a record of the evolution of the southern margin of the Pilbara Craton, and its interaction with the Gascoyne Complex during amalgamation with the Yilgarn Craton during the Capricorn Orogeny. Numerous models have been proposed and debate continues.

Stratigraphy and age

Pilbara Craton and Hamersley Basin

The oldest rocks in the region are Archaean granite and greenstone units in the Pilbara Craton with several inliers along the southern margin of the craton (e.g. Wyloo and Rocklea Domes, and Sylvania Inlier to the east). The Hamersley Basin unconformably overlies the Archaean rocks and consists of the Mount Bruce Supergroup subdivided as the Fortescue, Hamersley, and Turee Creek Groups (Trendall, 1990). Volcanic units within the Hamersley Group range in age from 2597 to 2449 Ma (Trendall et al., 1998; Barley et al., 1997). The Boolgeeda Iron Formation is the uppermost BIF of the Hamersley Group and is gradationally overlain by turbidites, shallow-marine carbonate rocks, and fluvial and marine siliciclastic rocks of the Turee Creek Group (Martin et al., 2000).

Wyloo Group

The base of the Wyloo Group is defined by an unconformity below the shallow-marine Beasley River Quartzite (typically 220–360 m thick; this and following thicknesses from Thorne and Seymour, 1991), which in turn is overlain by continental tholeiites of the Cheela Springs Basalt (maximum thickness 2000 m). Zircons from a volcaniclastic breccia in the Cheela Springs Basalt yielded a youngest component at 2209 ± 15 Ma (Martin et al., 1998). The Cheela Springs Basalt extrusion environment evolves from terrestrial at the base to shallow marine, and is partially overlain by dolomitic siltstone and mudstone of the Wooly Dolomite (maximum thickness 325 m).

A significant unconformity at the base of the Mount McGrath Formation is taken as the division between the lower and upper Wyloo Groups (Thorne and Seymour, 1991). The Mount McGrath Formation consists of deltaic and shallow-marine cycles of ferruginous conglomerate and siltstone with a maximum thickness of up to 1200 m. The Duck Creek Dolomite conformably overlies the Mount McGrath Formation, overlaps older units to the west of Wyloo Dome, and reaches a thickness of more than 1000 m.

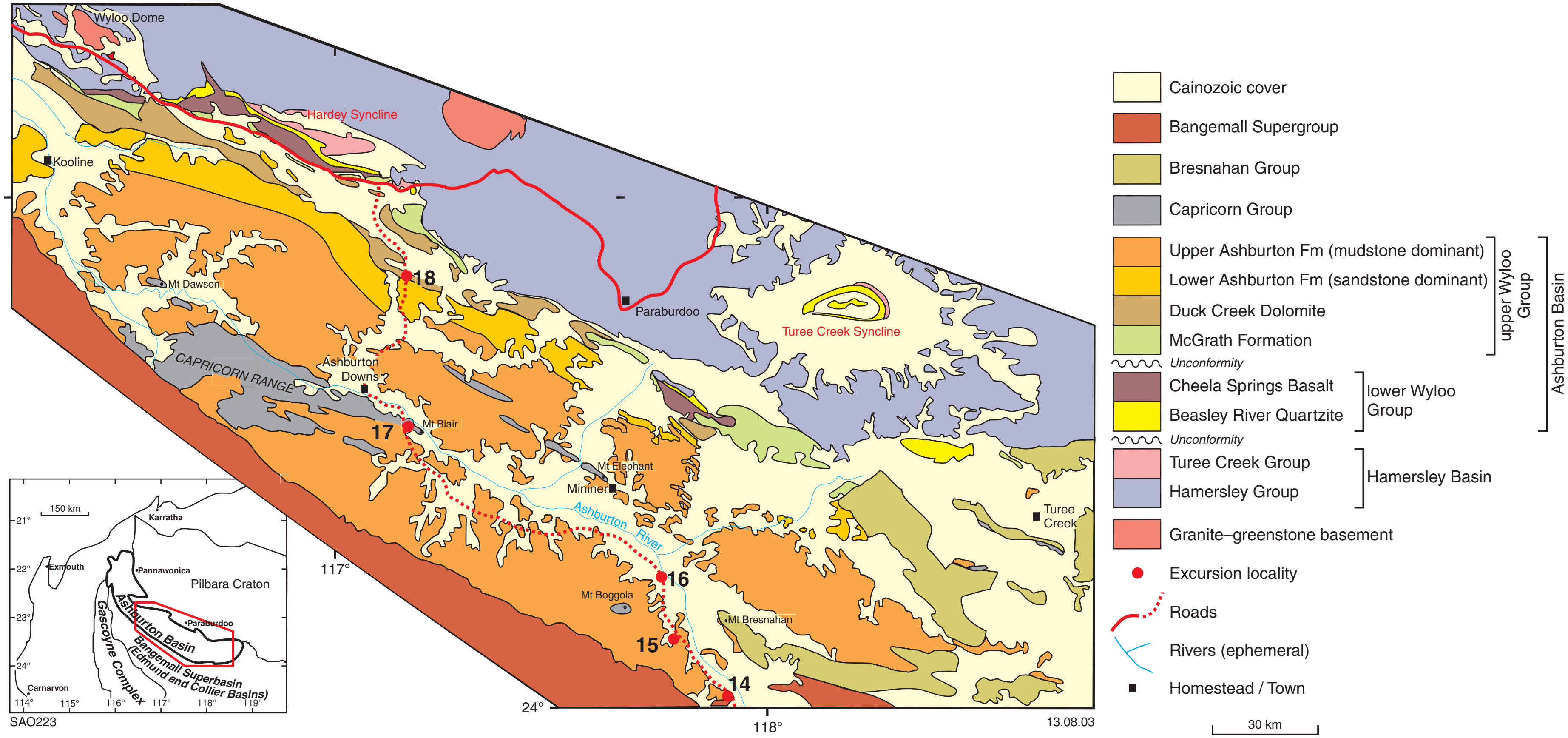

Figure 34. Geological map of central and eastern sections of the Ashburton Province illustrating excursion localities and route

Group	Formation	Age	Interpretation
Bangemall	Irregully		**Bangemall Basin** intracontinental
	Coobarra	1638 ± 14[1]	
Bresnahan		?	**Bresnahan Basin** intracontinental, east–west extension
Mount Minnie		?	**Mount Minnie Basin** ?intracontinental/continental margin, ?late D_{2a}
Capricorn	(Capricorn formerly)	1804 ± 7[2]	**Blair Basin** intracontinental, fluvial
Upper Wyloo W	Ashburton – Cane River (?)		**Ashburton Trough** continental margin with converging orogenic belt from southwest; palaeodirections from north (lower) and east (upper), diachronous deposition — younging westward
	June Hill Volcanics	1799 ± 8[3]	
	Ashburton – Wandarry (?)	c. 1819[4]	
	Mount Boggola	1828 ± 5[5]	
E	Ashburton – Mininer (?)	c. 1860[4]	
	Duck Creek Dolomite		
	Mount McGrath		
Lower Wyloo	Wooly Dolomite		**McGrath Trough** foreland basin, palaeodirections from south — uplifted Pilbara margin?
	Cheela Springs Basalt	2209 ± 15[6]	
	Beasley River Quartzite		
Turee Creek	Kazput		
	Koolbye		
	Kungarra		
Hamersley	Boolgeeda		
	Woongarra Rhyolite	2449 ± 3[7]	**Hamersley Platform**

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Angular unconformity — Conformable contact

Figure 35. Stratigraphic subdivisions and tectonic interpretations of the Ashburton Province region. Geochronology: [1] Nelson (1995); [2] Hall et al. (2001); [3] Evans et al. (in prep.); Sircombe, K. N., unpublished data ([4] detrital zircon minimum age components, [5] felsic tuff above mafic volcanic rocks); [6] Martin et al. (1998); and [7] Barley et al. (1993)

Ashburton Formation

The Ashburton Formation conformably overlies the Duck Creek Dolomite and consists of mudstone and sandstone interbedded with minor amounts of conglomerate, BIF and chert, dolomite, and mafic to felsic volcanic rocks. Features associated with turbidite deposition are common. This unit forms the bulk of the Ashburton Province, but the lack of marker horizons and structural complexity make assessing true thickness difficult. Thorne and Seymour (1991) estimated a thickness of between 5 and 12 km, and suggested a broad stratigraphy with three facies subdivisions as follows, from lower to upper levels:
- mudstone, thin-bedded sandstone, BIF–chert (20–500 m thick);
- massive sandstone, thin- to medium-bedded sandstone, mudstone, minor conglomerate (2.4 – 5 km thick);
- mudstone, thin-bedded sandstone, BIF–chert, numerous conglomerate and sandstone units (2.5 – 6 km thick).

The latter two associations are the thickest and are shown on the geological map (Fig. 34). The stratigraphy can be broadly summarized as the lower Ashburton Formation being more sandstone rich and the thicker, upper succession being more mudstone rich. Both are interpreted as being deposited in an elongate deep-marine basin draining from the east, with lower successions deposited in a braided distributary channel system that evolved into a few major channels with a decrease in supply of coarse clastic rocks (Thorne and Seymour, 1991). Areas of the Ashburton Formation around the Capricorn Range and Mount Dawson display palaeocurrent indicators directed from the southwest; these are interpreted as evidence for submarine fans prograding northeastward into the basin due to uplifting south of the basin.

Establishing a geochronological framework of the Ashburton Formation has been difficult, given its compositional homogeneity and structural complexity. Dating the Ashburton Formation, and indeed the whole of the Wyloo Group, previously relied on a multigrain U–Pb zircon age of 1843 ± 2 Ma for the June Hill Volcanics immediately underlying the Ashburton Formation northwest of the Wyloo Dome (Pidgeon and Horwitz, 1991). This age had been coupled with a 1804 ± 7 Ma age from a tuff in the overlying Capricorn Group (Hall et al., 2001) to provide a reasonable approximately 40 million-year framework for the deposition and deformation of the considerably thick Ashburton Formation.

However, recent work (Evans et al., in prep.) indicated that the June Hill Volcanics age of Pidgeon and Horwitz (1991) is actually an amalgam of crystallization and inherited ages, and an alternative single-grain SHRIMP U–Pb age of 1799 ± 8 Ma is provided — remarkably similar to the Capricorn Group age. Furthermore, the June Hill Volcanics were correlated with similar volcanic rocks at Mount Boggola in the southeastern section of the basin, but a recent SHRIMP age of 1828 ± 5 Ma from a felsic unit above mafic volcanic rocks has contradicted this (Sircombe, in prep.).

These recent dates have significant implications for the geological evolution of the Ashburton Formation. Evans et al. (in prep.) proposed that deposition of the Ashburton Formation was diachronous from east to west across the Ashburton Province, and what had been previously mapped as a single homogeneous unit may be divided into at least two successions. Such a stratigraphic amendment is broadly similar to the 'Cane River supersequence' proposed by Krapez (1999) on the basis of eustatic models, but instead this succession in the northwest may be the correlative to the Capricorn Group rather than other parts of the Ashburton Formation. In turn, this age also implies that the Mount Minnie Group in the northwest is younger than both the 'Cane River succession' and the Capricorn Group. The significance of any stratigraphic division created by the Mount Boggola volcanic rocks (loosely using the Mininer and Wandarry nomenclature in Fig. 35) remains uncertain.

Detrital zircon geochronology of the Ashburton Formation also confirms the two-stage stratigraphy of the succession (Sircombe, K. N., unpublished data). Detrital zircon ages are commonly concentrated in the Palaeoproterozoic, with only minor Archaean contributions. Overall, results can be divided into two groups. The first, derived from lower in the Ashburton Formation succession, displays a variety of ages between c. 1800 and 2500 Ma. Substantial individual modes are difficult to establish, but a c. 1860 Ma mode can be discerned in two samples. The second group, derived from higher in the Ashburton Formation succession, tends to be strongly unimodal with prominent modes at 1819 and 1840 Ma.

Capricorn Group

The Capricorn Group overlies the Ashburton Formation with a prominent angular unconformity. The composition of fluvial to shallow-marine coarse sandstones, siltstones, and minor conglomerate and dolomite also differs markedly from the Ashburton Formation. Overall, the succession is 800 m thick (Thorne and Seymour, 1991) and has been recently elevated to group status and subdivided into the Bywash and Mooline Formations (Thorne et al., 2002). The outcrop is strongly controlled by northwest-trending faults and folds associated with D_{2a} deformation, with several exposures infolded into D_{2a} structures well to the east of the central Capricorn Range, even as far as Turee Creek Homestead. The Capricorn Range section of the formation also contains a thick felsic lapilli tuff layer (Koonong Member), which has been dated by Hall et al. (2001) as 1804 ± 7 Ma.

The depositional setting of the Capricorn Group is interpreted as a southeast-trending basin with quartz-rich sand draining from the west, southwest, and east into a central shallow-marine (or lacustrine) environment (Thorne and Seymour, 1991). The succession appears to have been influenced by a relative rise in sea or lake level before a renewed influx of detritus. Volcanic activity was also a locally prominent contributor.

Mount Minnie Group

The Mount Minnie Group forms a series of outliers in the northwestern section of the Ashburton Province, unconformably overlying the Ashburton Formation with a maximum thickness of about 2600 m (Seymour et al., 1988). The group is conformably subdivided into the Brodagee Sandstone (quartz sandstone, conglomerate), Wabco Shale (mudstone with thin interbeds of quartz sandstone), and Warramboo Sandstone (quartz sandstone, minor siltstone, and mudstone). The absolute age of the Mount Minnie Group is uncertain, although as discussed in **Ashburton Formation**, the one implication of a younger June Hill Volcanics age is that the Mount Minnie Group is also considerably younger than c. 1799 Ma.

The depositional setting of the Mount Minnie Group is interpreted as a basin, lying west of the uplifted Wyloo Group, being filled by a series of alluvial fans. The Wabco Shale and Warramboo Sandstone are interpreted as a marine transgression into this basin, followed by progradation back into the northern part of the basin (Thorne and Seymour, 1991). Whether this basin was intracontinental or faced an open ocean margin is uncertain.

Bresnahan Group

The Bresnahan Group unconformably overlies the Wyloo and Capricorn Groups and has a maximum measured thickness of 4000 m (Thorne and Seymour, 1991). The group consists of conglomerate, sandstone, and siltstone–mudstone. Three facies associations are recognized, with basal localized accumulations of conglomerate infilling underlying topography. This facies grades laterally and vertically into very coarse grained sandstone and pebbly sandstone with medium- to large-scale trough cross-stratification. A strong unimodal easterly directed palaeoflow is indicated. A lacustrine facies of mudstone–siltstone with interlayered sandstone is recognized, particularly in the eastern section of the basin.

The absolute age of the Bresnahan Group is uncertain, but structurally it is younger than D_{2a} deformation and appears to have been tectonically controlled by post-D_{2a} southeasterly directed extension along the southern Pilbara Craton margin that included development of the Mount Whaleback Fault system (Tyler et al., 1991). The overlying Bangemall Supergroup provides a stratigraphic constraint on the age.

The depositional setting of the Bresnahan Group is interpreted as a series of basins bounded by normal-faulted western margins and uplifted Wyloo Group and Capricorn Group rocks. Close to faulted margins, debris-flow conglomerates filled topography on the pre-Bresnahan Group surface and alluvial-fan complexes drained the uplifted western highlands (Hunter, 1990).

Bangemall Supergroup

The Bangemall Supergroup (as recently revised by Martin and Thorne, 2002) unconformably overlies the Wyloo, Capricorn, Mount Minnie, and Bresnahan Groups. The contact with this supergroup delineates the southern margin of the Ashburton Province and is commonly marked by a prominent escarpment formed in the subhorizontal Bangemall Supergroup successions.

In the southern part of the Ashburton Province concerning this guide, the lowermost member of the Bangemall Supergroup is the Irregully Formation, which is about 500 m thick and largely consists of dolarenite and dololutite with interbeds of pebble conglomerate, quartz sandstone, siltstone–mudstone, and chert (Thorne et al., 1991). Domical stromatolites are common and the depositional setting is interpreted as a coastal lagoon system (Muhling and Brakel, 1985).

The age of the Bangemall Supergroup has recently been constrained by the dating of underlying granitic rocks in the central part of the Bangemall Superbasin, with an age of 1638 ± 14 Ma providing a maximum age of deposition (Nelson, 1995; Thorne and Martin, 2002).

Structure

The lower succession on the northern margin is marked by an earlier series of deformations associated with the Opthalmia Fold Belt (Tyler and Thorne, 1990a). In the central and southern sections of the Ashburton Province, two phases of deformation are recognized (Tyler and Thorne, 1990a): D_{1a} structures post-date deposition of the Ashburton Formation and pre-date deposition of the Capricorn Group, Mount Minnie Group, and intrusion of the Boolaloo Batholith; D_{2a} structures post-date the Capricorn Group and are the dominant structures. The Bresnahan Group is associated with a series of northeast-trending extensional fault systems post-dating D_{2a}. Large-scale, west-northwesterly trending open folding in this group also pre-dates the Bangemall Superbasin (Tyler et al., 1991). Evidence for a subsequent phase of post-Bangemall Supergroup deformation is seen in the western extreme of the Ashburton Province (Thorne and Seymour, 1991).

In the area covered by this guide, the notable features associated with the two phases of deformation are:

- An S_1 penetrative cleavage seen within Ashburton Formation mudstones and sandstones, but not in the overlying successions. S_1 cleavage is commonly crenulated by an S_2 cleavage.
- F_2 structures spanning the Ashburton and successor basins (except the Bresnahan Basin). These are commonly large, and trend west-northwesterly with northwesterly or southeasterly plunges of 10°–40°. Numerous wrench faults parallel the fold trends, although the lack of marker horizons makes estimating displacement difficult. Parts of the overlying basins, particularly outliers of the Blair Basin, are notably fault bound. The orientation of F_1 structures is difficult to determine, although relationships at Mount Blair suggest tight folds with steep southward-plunging axial surfaces (Thorne and Seymour, 1991).

Tectonic evolution

A number of strongly contrasting models have been proposed for the tectonic setting of the Ashburton Province. Tyler and Thorne (1990a) and Thorne and Seymour (1991) proposed a post-rift convergent, foreland-basin setting for the entire Wyloo Group, deposited on the lower plate of the Yilgarn–Pilbara Craton collision, with detritus derived from the north and east. Upper successions (upper Ashburton Formation, and Capricorn, Mount Minnie, and Bresnahan Groups) represent the encroachment of the Yilgarn Craton, development of the Gascoyne Complex, and associated uplift and volcanism with detritus sourced from the west and south. In particular, the upper Ashburton Formation marks a transition, with more detritus sourced from the southwest rather than from the uplifted Sylvania Inlier in the east.

Blake and Barley (1992) considered that continental collision had culminated earlier and that the Beasley River Quartzite marked the transition to extension and passive-margin subsidence with detritus sourced from the north. The identity of the southern continent in this model is unclear. Krapez (1999) further developed this model and suggested that the southern conjugate continent was the Gawler Craton and that the lower Wyloo Group and lower successions of the Ashburton Formation were deposited in a failed-rift setting. In this model, the June Hill Volcanics and Mount Boggola volcanic rocks are interpreted as the onset of a compressional regime with the oblique collision of the Yilgarn Craton. The upper successions of the Ashburton Formation were deposited in a remnant deep-marine basin behind, and sourced from, an encroaching fore-arc that was to become the Gascoyne Complex. The Capricorn, Mount Minnie, and Bresnahan Groups represent the final stages of the collision, with the remnant basins overwhelmed by detritus from the southwestern orogenic belt.

In contrast, Horwitz (1982) recognized a post-Hamersley Group pre-Ashburton Formation basin termed the McGrath Trough that included the Turee Creek and lower Wyloo Groups (Fig. 35). Subsequent development of this model by Powell and Horwitz (1994) and Martin et al. (2000) proposed a foreland-basin setting for the McGrath Trough, with detritus derived from an uplifted margin of the Hamersley Basin to the south. The Mount McGrath Formation marks a significant change in tectonic setting, with palaeodirections reversing towards the south and an extensional rift-drift environment being interpreted as the beginning of the Ashburton Trough (Fig. 35).

Recent SHRIMP age data further refines or refutes these models, although a new coherent model incorporating proven features of each proposal is still emerging. The palaeodirectional and compositional provenance data of Martin et al. (2000) appears comprehensive in establishing a foreland basin in the McGrath Trough, reaching up to the Mount McGrath Formation. Detrital zircon geochronology on these formations would be a further test of this model. The theoretical timing of the Krapez (1999) model is seriously challenged by recent data, particularly the younger June Hill Volcanics age of 1799 ± 8 Ma (Evans et al., in prep.) and the significant difference with the Mount Boggola volcanic rocks at 1825 ± 5 Ma (Sircombe, K. N., unpublished data). However, the younger June Hill Volcanics age does support the concept of a younger 'Cane River supersequence' of the Ashburton Formation in the northwest of the Ashburton Province, and a new model of diachroneity is emerging (Evans et al., in prep.).

Elsewhere in the Ashburton Formation, detrital zircon geochronology has confirmed the two-stage evolution of this unit as postulated by Thorne and Seymour (1991), with a transition from polymodal to unimodal age distributions (Sircombe, K. N., unpublished data). The youngest of these unimodal components from higher in the Ashburton Formation is dated at c. 1819 Ma and emphasizes that the Ashburton Trough was already in transition before the rapid deformation and uplift represented by the overlying Capricorn Group. However, the lack of Archaean-aged grains does challenge the suggestion of predominant derivation from the uplifted Sylvania Inlier. Indeed, the youngest components in the lower successions of the Ashburton Formation are c. 1860 Ma, suggesting that: a) the Ashburton Trough was already deriving material from an encroaching Gascoyne Complex; and b) a substantial hiatus of about 200 million years must be accommodated between the Ashburton Formation and Cheela Springs Basalt.

Further zircon dating should also aid the understanding of the age and provenance of the Capricorn, Mount Minnie, and Bresnahan Groups and their role in the last stages of formation of the Ashburton Province.

Excursion localities — Ashburton Province

Day 5

Today we will look at rocks of the Ashburton Province, including the mafic volcanic rocks in the Ashburton Formation, and metasedimentary rocks of the Capricorn Group. We will also look at the relationship of the Ashburton Formation and the overlying Bangemall Supergroup.

Locality 14: 'Tchintaby Ridge' (MGA 593741E 7347031N)

From Locality 13, travel west towards Mount Augustus tourist park. Turn left into the tourist park and then travel north through it and onto the Pingandy Road. Continue travelling north towards Pingandy Homestead. From Pingandy Homestead drive west. At the intersection of the Pingandy Road and Ashburton Downs – Meekatharra Road turn left (northward). Travel about 2 km. The road follows a gentle descent through an open gorge with a prominent plateau to the west and a series of knolls in the east. Turn off the road and park in the open area at the base of a north-facing escarpment. Please note that this is a steep cliff section and a high degree of fitness, agility, and care is required.

North of the escarpment (Fig. 34) is a knoll (1 on Fig. 36; MGA 593676E 7347210N) providing a typical Ashburton Formation shale outcrop: orange-red weathering, abundant quartz veining (10–30 cm thick), and a pencil cleavage lineation illustrating two prominent cleavage planes.

From the top of the knoll, patches of grey-coloured outcrop in the Ashburton Formation can also be seen nearby; these are probably zones of kaolinization associated with potential mineralization. Many such patches in the Ashburton Province bear signs of older excavation or more-recent geochemical sampling, or both. On the horizon to the northwest (bearing 310°), a range of low hills marks Mount Boggola (681 m), where a series of mafic volcanic rocks in the Ashburton Formation is capped by the Capricorn Group. To the north the distinctly rounded range is Mount Bresnahan (683 m), which comprises Bresnahan Group conglomerates.

Looking southward provides an overview of the escarpment section (Fig. 36). 'Tchintaby Ridge' is a good illustration of the structural relationships between the Ashburton Formation and the overlying Bangemall Supergroup, with the latter providing the bulk of the vertical outcrop in the cliffs. This section also includes a presumed outcrop of Capricorn Group between the Ashburton Formation and Bangemall Supergroup.

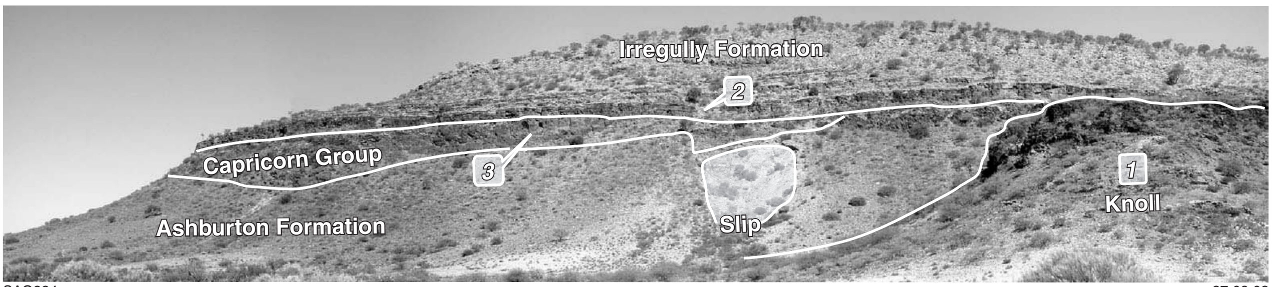

Figure 36. Overview of 'Tchintaby Ridge' (Locality 14) with the three formations exposed. Numbers indicate locations discussed in text

From the knoll walk towards the cliff section. A suggested route up the escarpment is immediately to the east of a prominent slip exposing grey Ashburton Formation. The vertical exposure is less at this point (2 on Fig. 36) thus allowing relatively simple access to higher parts of the section if required.

At the top of the section the Bangemall Supergroup is represented at this location by its lowermost member, the Irregully Formation (Thorne et al., 1991), which is a stromatolitic dolarenite–dololutite with interbedded thin sheets of quartzitic sandstone. The outcrop weathers to a distinct pale-orange colour. Chert nodules are common and the unit is overlain by massive chert near the top of the ridge. Beds dip about 10°–15° towards the southeast and south-southeast. Stromatolitic features are subdued with commonly only small domes and planar shapes and local tee-pee structures.

Walk about 120 m towards the eastern end of the escarpment, staying below the cliff. A prominent gully (3 on Fig. 36; MGA 593741E 7347031N) provides an excellent exposure of the unconformity.

The gully at this location exposes the angular unconformity between the Irregully Formation and Capricorn Group (Fig. 37). The lower unit consists of massive medium-grained arenites interbedded with granular sandstones and conglomerates. Bedding dips 35°–45° towards the south-southeast. Unlike the underlying Ashburton Formation, there is only one prominent cleavage plane, with a steep dip (76°) towards the south-southwest. This cleavage and associated quartz veins are truncated at the unconformity. These stratigraphic, lithological, and structural features correlate with the broader features of the Capricorn Group farther north; therefore, this outcrop is interpreted as the southernmost exposure of this unit in the Ashburton region.

Although there is little evidence for weathering at the contact, dolomite penetrates the formation up to 10 m below the unconformity. Conglomeratic beds contain clasts of shale, fine sandstone, chert, and dolomite. Although both the Ashburton Formation and Capricorn Group contain dolomite elsewhere, the prominence at this location suggests a relatively local and substantial source.

The contact between the Capricorn Group and Ashburton Formation is not well exposed at this locality. A 'triple junction' between the Capricorn Group and the Ashburton and Irregully Formations is immediately above and to the west of the prominent slip (Fig. 36).

Locality 15: Little Pingandy Creek crossing (MGA 580757E 7360000N)

From Locality 14, return to the Ashburton Downs – Meekatharra Road and continue north for about 19 km. The road crosses a well-defined creek bed at a concrete ford.

The Little Pingandy Creek crossing (Locality 15; Fig. 34) provides a convenient and relatively unweathered exposure of the Ashburton Formation. The outcrop consists of graded siltstones and fine-grained sandstones with decimetre-scale bedding, and some flame structures (Fig. 38). The bedding youngs northward at this locality, but is slightly overturned. A weak cleavage is also developed in the finer grained parts of the section.

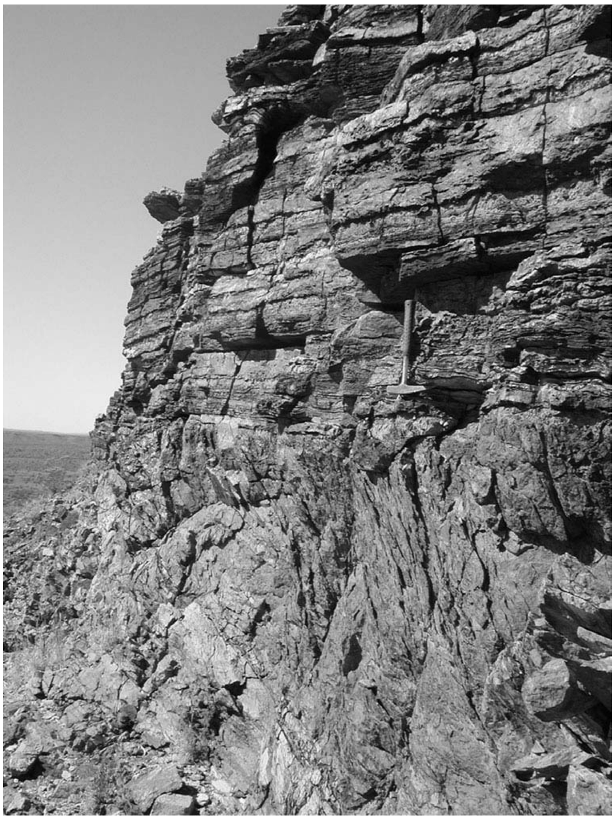

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Figure 37. Angular unconformity between the Capricorn Group (below) and Irregully Formation (above) at 'Tchintaby Ridge', Locality 14

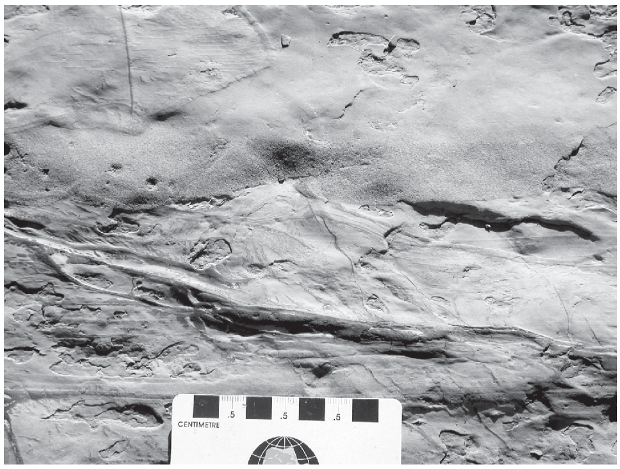

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Figure 38. Flame structure in the Ashburton Formation at Little Pingandy Creek crossing, Locality 15

The next section of road provides the best views of both Mount Bresnahan to the east and Mount Boggola to the west.

Locality 16: Banded iron-formation outcrop (MGA 577084E 7372741N)

From Locality 15, return to the Ashburton Downs – Meekatharra Road and continue north for about 14 km. The road rises between two knolls, with the one on the west topped by a survey mark.

Volcanic rocks and associated units constitute only a small proportion (1%; Thorne and Seymour, 1991) of the Ashburton Formation, thus compounding the difficulty in dating and understanding the geological evolution of this extensive sedimentary succession. The major locality for such volcanic rocks is in the low hills north of Mount Boggola, where an extensive sequence of vesicular pillow lava is overlain by a series of BIF, chert, ferruginous pelite, mafic volcaniclastic rocks, and rare felsic tuffs. The outcrop at Locality 16 (Fig. 34) represents a well-weathered eastern extension of the BIF – ferruginous pelite sequence and provides a good view of Mount Boggola.

Looking west-southwest from Locality 16, Mount Boggola forms a prominent peak on the horizon, with the Capricorn Group in a shallow syncline capping underlying the Ashburton Formation. Farther north, a series of low hills marks the outcrop of the mafic volcanic rocks and overlying successions. Recent work on a felsic unit overlying the mafic volcanic rocks yielded a SHRIMP $^{207}Pb/^{206}Pb$ weighted mean age of 1828 ± 5 Ma (2 s.d., 1.1 MSWD; Sircombe, in prep.). A number of the zircons in this unit were notably acicular, suggesting that it may represent either a tuff or at least a felsic volcaniclastic unit with minimal transportation.

From here the Ashburton Downs – Meekatharra Road skirts along the southern edge of the Ashburton River floodplain and is entirely underlain by the Ashburton Formation. Good outcrops are sparse, although a number of prominent calcrete-capped mesas rise about 30 m above the current valley floor. These are interpreted as groundwater deposits of a former valley floor of indeterminate, but possibly Tertiary, age (Mann and Horwitz, 1979). North of the Ashburton River in the distance, Mount Elephant (491 m) consists of an outlier of Capricorn Group bound by D_2 northwest-trending structures.

Locality 17: Mount Blair unconformity (MGA 513427E 7407433N)

From Locality 16, return to the Ashburton Downs – Meekatharra Road and continue north for about 81 km. Park just before the road narrows and enters a small gorge (coincidentally, this gorge also marks the Tropic of Capricorn). The locality is east of the road in a gully on the southern side of Mount Blair. Walk about 550 m from the road to the section. Note that gullies are a favourite haunt of much wildlife in this arid environment. In particular, when amongst the low trees and scrub, keep your head up to avoid encounters with orb spiders and their webs. Although large they are generally harmless, but excited attempts to extricate them from hair, hats, and other belongings have been known to result in injury.

The angular unconformity between the Ashburton Formation and the overlying Capricorn Group is well exposed in this section (Fig. 34). Ashburton Formation mudstone and sandstone beds display dips of up to 40° to the south within open folds (Fig. 39). Two cleavage planes are particularly seen in the finer grained material. The Capricorn Group dips northward at about 55° and the basal sequence consists of a thin stromatolitic dolomite grading upward into a succession of mudstones and cross-stratified sandstones (Fig. 40).

The Mount Blair ridge effectively forms a northwest-trending asymmetrical anticline — the contact on the northern side of the ridge dips southward at about 30°. Mount Blair itself is the eastern extension of the Capricorn Range (Fig. 34), where a broadly similar synclinal structure continues for about 50 km to the northwest before swinging southwest.

Felsic tuffs within the Capricorn Group provide some age constraints on the geological history of the Ashburton Province. These tuffs are particularly prominent in the central part of the Capricorn Range and reach up to 15 m in thickness (Thorne and Seymour, 1991). Hall et al. (2001) dated a lapilli tuff about 24 km west-northwest of Locality 17, which yielded a $^{207}Pb/^{206}Pb$ SHRIMP age of 1804 ± 7 Ma. The youngest detrital-zircon age components within the underlying Ashburton Formation in this area are c. 1819 Ma (Sircombe, K. N., unpublished data). This marked similarity in age across such a prominent unconformity has a number of implications. Firstly, the central Capricorn Range area was relatively proximal to a volcanic source and that source was already active in the closing stages of Ashburton Formation deposition. Secondly, the deformation associated with creating the angular unconformity was notably rapid.

Figure 39. Open folds in the Ashburton Formation immediately below the Mount Blair unconformity on the western side of a gully at Locality 17

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Figure 40. Cross-stratification in the Capricorn Group above the Mount Blair unconformity at Locality 17

After passing through the gorge, the road follows the northern margin of the Capricorn Ranges — and the northern limb of the anticline in the Capricorn Group — before heading north across the Ashburton River at Ashburton Downs Homestead. The surface is entirely underlain by the Ashburton Formation and the weathering patterns remain typical of this unit. However, subtle changes occur as the road passes lower into the Ashburton Formation stratigraphy and into more-sandstone-dominated sequences.

Locality 18: Duck Creek Dolomite (MGA 514454E 7440266N)

From Locality 17, return to the Ashburton Downs – Meekatharra Road and continue north for about 45 km (pass Ashburton Downs Homestead and cross the Ashburton River after 14 km). After passing across a generally flat landscape, the road will rise over a prominent ridge. Park near the crest of the ridge.

This ridge marks the base of the Ashburton Formation and the transition into the Duck Creek Dolomite (Fig. 34). At this location the dolomite is locally silicified into a white chert, although a pavement with more-typical dolomite outcrop and weathering patterns can be seen on the northern side of the ridge (e.g. MGA 514619E 7440313N). North of this locality the surface weathering is a distinctly paler colour, reflecting the underlying dolomite.

The Hardey Syncline area, Hamersley Province

by

S. M. Reddy and D. Hollingsworth

Introduction

The Hamersley Province of the southern Pilbara Craton is probably best known for the number of large iron ore deposits that are associated with geographically extensive BIFs (Trendall and Blockley, 1970). However, the Hamersley Province also records a complex depositional and structural history that provides a unique picture of sedimentological and tectonic processes from the Archaean through to the Palaeoproterozoic. One of the key areas for understanding the geological evolution of the Hamersley Province is around the Hardey Syncline, where the relationships between basin formation, sedimentation, and fold–thrust belt development can be inferred.

Regional background and stratigraphy

The Pilbara Craton in the northern part of Western Australia comprises two different lithotectonic sequences (Fig. 41). The oldest, ranging in age from c. 3.55 to 2.85 Ga, is dominated by granite–greenstone basement. To the south a series of younger Archaean to Palaeoproterozoic (2765–2470 Ma; Thorne and Tyler, 1996) volcano-sedimentary rocks, referred to as the Mount Bruce Supergroup, unconformably overlie this granite–greenstone basement complex. The relationship between basement and cover is preserved in a series of structural domes within the Mount Bruce Supergroup that expose the underlying Pilbara Craton basement and enable it to be traced to close to the southern edge of the craton.

The Hamersley Province is the regional term commonly used to define the large (~100 000 km^2) area over which the Mount Bruce Supergroup outcrops. This supergroup covers a vast geographic area and can be subdivided into three stratigraphic groups: the Fortescue, Hamersley, and Turee Creek Groups (Figs 42 and 43). The stratigraphy of the Mount Bruce Supergroup and the overlying unconformable Wyloo Group has been well known for some time (e.g. de la Hunty, 1965). However, over recent years there have been some new interpretations of the tectonic setting of basin formation and deformation history that have allowed this stratigraphy to be placed within a geodynamic framework. The aim of this day of the excursion is to outline both structural and sedimentological aspects of Mount Bruce Supergroup evolution and outline the evidence for this geodynamic setting.

The stratigraphically lowest group of the Mount Bruce Supergroup — the Fortescue Group — has an age range from 2765 to 2687 Ma (Arndt et al., 1991), and unconformably overlies the Pilbara Craton basement. The group contains seven formations, two of which are dominated by terrigenous clastic rocks, and the other five comprising mafic volcanic rocks including aerial mafic flows, lapilli tuffs, subaqueous pillow lavas, and hyaloclastites. Much of the Fortescue Group, particularly the uppermost formations, are extensively intruded by mafic sills that can account for up to 60% of the total thickness of the formation.

The Hamersley Group (MacLeod et al., 1963) overlies the Fortescue Group and includes the BIFs that contain the iron ore deposits for which the Hamersley Province is best known. Volcanic-derived shale, dolomite, chert, and rhyolite are also important components of the stratigraphy of the Hamersley Group, but clastic sedimentation is rare. The Hamersley Group is divided into a number of different formations (Fig. 43), with the Brockman Iron Formation containing the main ore-grade mineralization in the Hamersley Province. The stratigraphy of the Hamersley Group shows a cyclic depositional sequence at many scales. Banded iron-formation, shale, and chert layers repeat in the stratigraphy (Fig. 43) with only minor variations regionally. It is evident that deposition occurred at the same time throughout the Hamersley Province, and correlation of bedding at centimetre scale is possible over hundreds of kilometres.

The Turee Creek Group is the youngest group of the Mount Bruce Supergroup and lies conformably on the uppermost iron formation of the underlying Hamersley Group (Fig. 43). Rocks of the Turee Creek Group indicate a change from chemical to clastic sedimentation in the Hamersley Province. Originally, the Turee Creek Group was considered part of a single conformable sedimentary sequence. However, recent work has documented the presence of several unconformities within the upper part of the Turee Creek Group (Martin et al., 2000). These provide important chronological constraints between basin formation, associated sedimentation, and north–south shortening.

Unconformable on the Turee Creek Group, the Wyloo Group overlies the Mount Bruce Supergroup and is itself divided into subgroups (lower and upper Wyloo Groups) based on significant variations in depositional environment, established after official naming of the group. Granite–greenstones and Hamersley Province rocks at the southern margin of the Pilbara Craton are also unconformably overlain by the Wyloo Group, which was deposited sometime between 2200 and 1700 Ma.

Depositional history

In the past, the sedimentary history of the region has been considered in terms of two basins: deposition of the Mount

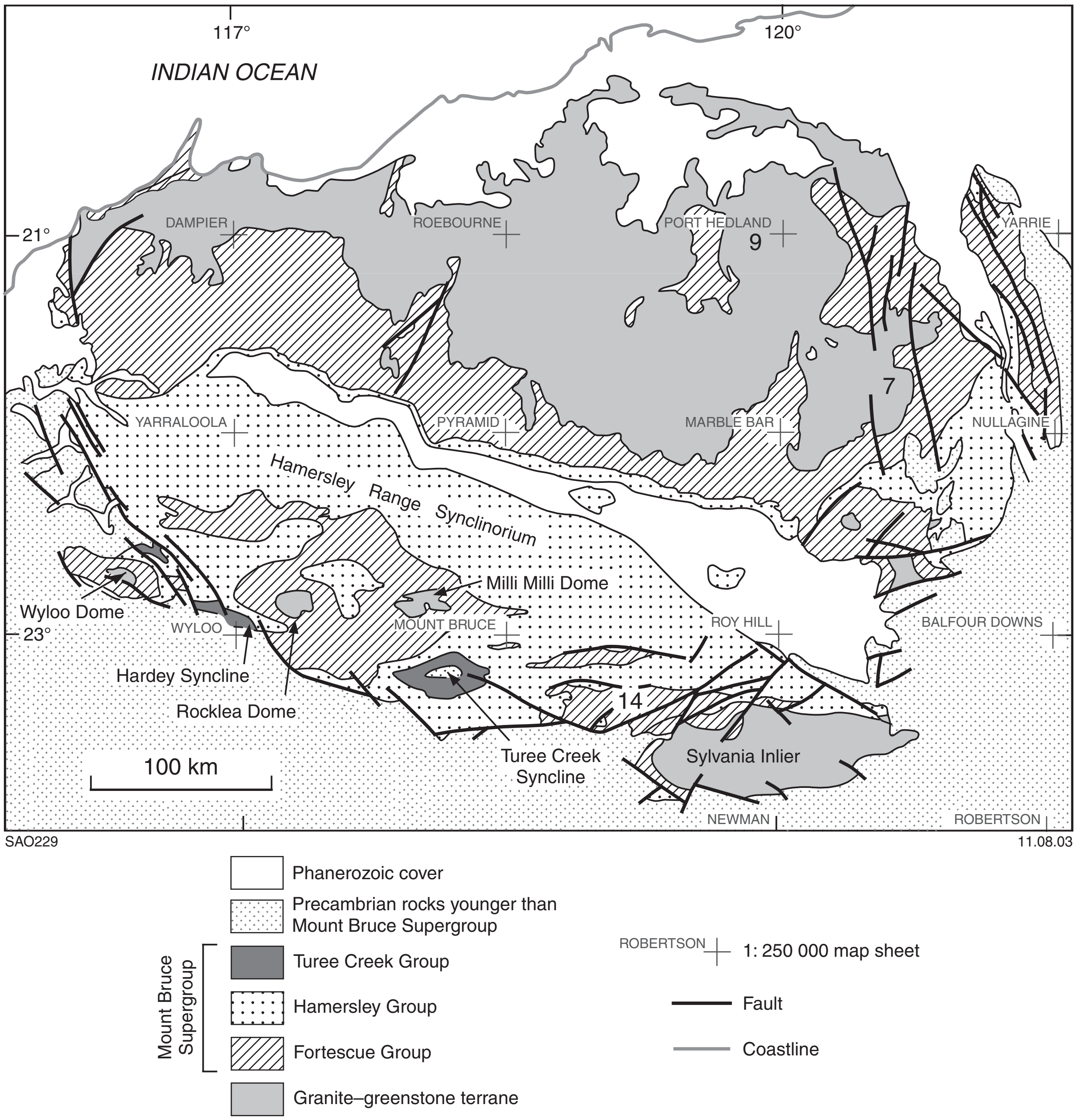

Figure 41. Geological setting of the Pilbara Craton showing the granite–greenstone basement and overlying Mount Bruce Supergroup and younger rocks. Also shown are some of the dome and syncline structures (including the Hardey Syncline) that characterize the southwestern Hamersley Province. After Geological Survey of Western Australia (1990, p. 211)

Bruce Supergroup in the Hamersley Basin, and deposition of the Wyloo Group in the younger Ashburton Basin. However, recently it has been argued that three distinct depocentres existed, with much of the Turee Creek Group and the lower Wyloo Group being deposited in the McGrath Trough (a foreland basin), the older Hamersley Basin, and the younger Ashburton Basin (Martin et al., 2000).

Deposition or accumulation of the Mount Bruce Supergroup began during the late Archaean with deposition of the lower Fortescue Group at around 2765 Ma (Arndt et al., 1991). Accumulation of the lower Fortescue Group began in the western part of the Hamersley Province, with deposition of the Bellary Formation, Mount Roe Basalt, and Hardey Formation in a continental rift environment associated with extensive volcanism (Blake and Groves, 1987). Accumulation of the middle Fortescue Group reflects a change from rifting to a passive-margin environment that, during upper Fortescue Group times, developed into a shelf environment with deep-water facies evident in the south-southwest of the

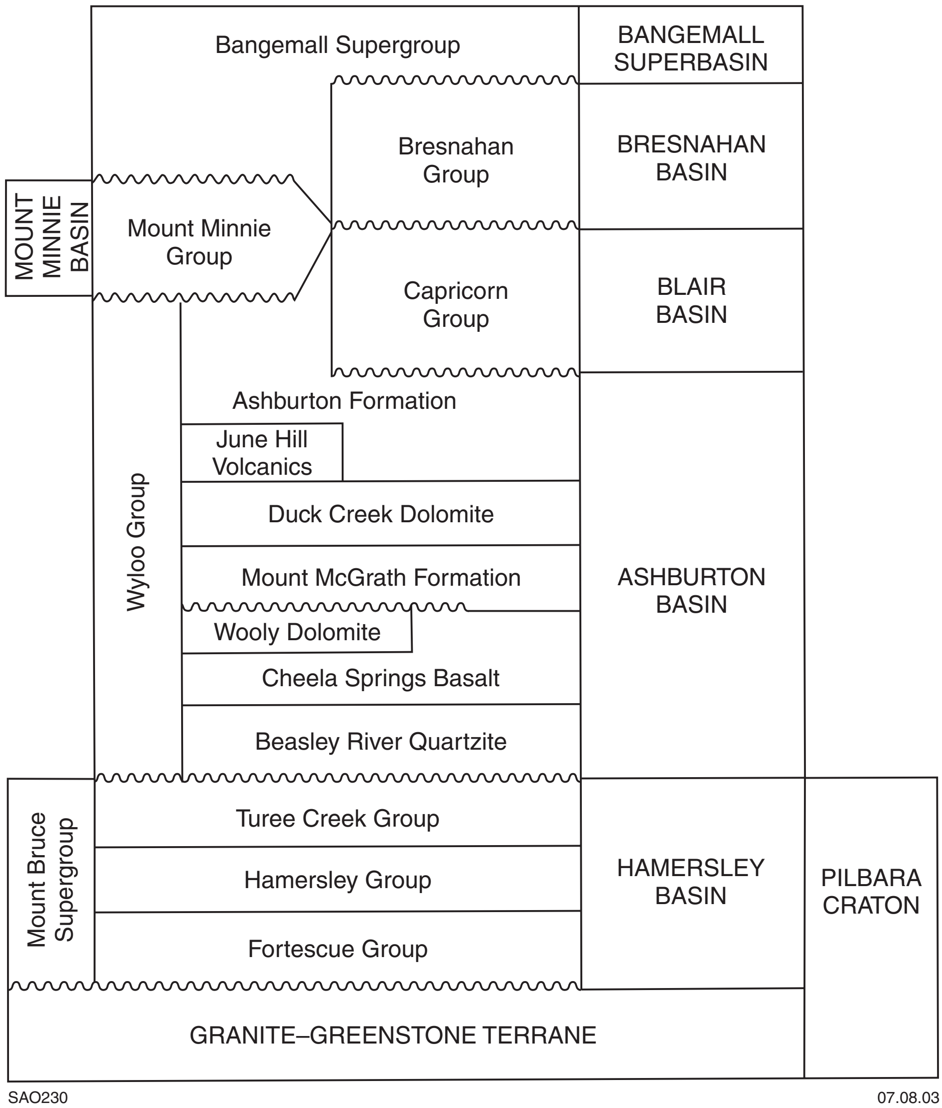

Figure 42. Stratigraphy of the Hamersley Province and nearby younger basins. After Geological Survey of Western Australia (1990, p. 212)

present-day Hamersley Province (Morris and Horwitz, 1983). The Hamersley Group and the lower part of the Turee Creek Group were deposited on the shelf, and represent an essentially stable environment from about 2680 Ma through to around 2450 Ma, at which time tectonic activity began to disrupt deposition of the Turee Creek Group.

The onset of deformation in the southwest of the Pilbara Craton is evident in the formation of the McGrath Trough, which is a foreland basin associated with northward advance of the Opthalmia Fold Belt. Deposition within the McGrath Trough began around 2450 Ma and continued throughout lower Wyloo Group times. Hiati and unconformities are common throughout the depositional units of the McGrath Trough and deformation was ongoing throughout its evolution. A major unconformity exists at the base of the lower Wyloo Group over much of the southern Hamersley Province. The Beasley River Quartzite unconformably overlies rocks as young as the Turee Creek Group and as old as the Fortescue Group. Mapping within the Hardey Syncline has shown that folding about an axis trending about 110° occurred before deposition of the Beasley River Quartzite, and continued during deposition of the lower Wyloo Group (Powell and Li, 1991). Folds within the Turee Creek Group are truncated by the unconformity, whereas the axial-planar cleavage below and above the unconformity remains consistent, as does the orientation of folds. Relationships preserved in the Hardey Syncline are a good indication of the ongoing deformation during deposition of the lower Wyloo Group.

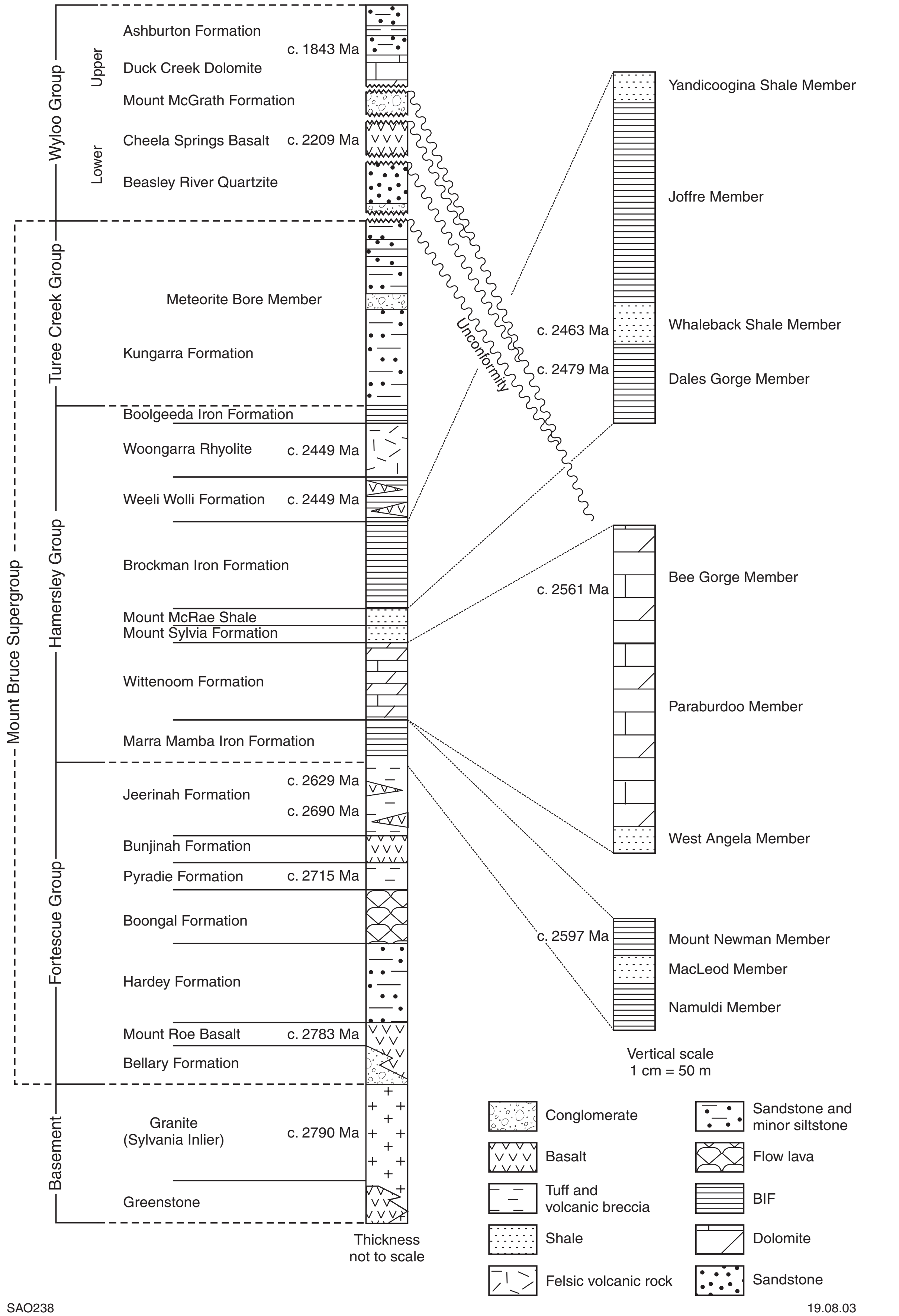

Figure 43. Stratigraphic relationships of the Hamersley Province, with representative geochronological data from Arndt et al. (1991); Pidgeon and Horwitz (1991); Barley et al. (1997); Trendall et al. (1998); Martin et al. (1998); Wingate and Giddings (2000); Trendall, A. F., 2003, (unpublished data); Figure after Cawood and Hollingsworth (2002)

Deformation history

Two regional folding events (F_2 and F_3) have been distinguished at the southern margin of the Hamersley Province. An older set of F_1 layer-parallel recumbent folds is locally within the Fortescue and Hamersley Groups. Easterly trending, northward-verging F_2 structures form overturned to recumbent folds in the southeastern Hamersley Province, but form open, upright to steeply inclined dome and basin structures in the southwest. These structures, commonly referred to as the Opthalmia Fold Belt, reflect south to north transport of the Hamersley Province over the granite–greenstone basement. These structures die out towards the northern Hamersley Province, where the Mount Bruce Supergroup is undeformed. The F_2 structures are thought to have developed synchronously with the McGrath Trough foreland basin and the deposition of the Turee Creek Group, and perhaps the lower Wyloo Group within this basin. A younger generation of northwesterly to easterly trending F_3 structures deform the upper and lower Wyloo Group and the Mount Bruce Supergroup, and are commonly referred to as the Ashburton Fold Belt. Some workers consider both F_2 and F_3 structures to be manifestations of the Capricorn Orogeny, which is thought to reflect continental collision between the Pilbara and Yilgarn Cratons (Trendall and Blockley, 1970; Tyler and Thorne, 1990b; Tyler, 1991). Other workers prefer to treat earlier F_2 structures as belonging to a distinct 2400 to 2200 Ma Opthalmian Orogeny that is unrelated to the Capricorn Orogeny (Powell and Horwitz, 1994; Martin et al., 1998).

Deformation that has affected the rocks of the Mount Bruce Supergroup can be summarized as a number of extensional events before, between, and after two major compressional events. Early rifting during deposition of the lower Fortescue Group is the only event for which a corresponding earlier compression is not evident. Northwest–southeast oriented extension occurred in the granite–greenstone basement in the northern Pilbara region during sedimentation of the lower Fortescue Group. Recent work near Rocklea Dome suggests that during deposition of the Bellary Formation and the Mount Roe Basalt, extension was perpendicular to that in the northern Pilbara region (Hall, 2002). Palaeocurrent data and detailed mapping of the stratigraphy have revealed easterly trending clastic wedges sitting in half-grabens. The easterly trend of the clastic wedges suggests a north–south oriented extension, and the distribution of the stratigraphic units indicates the presence of a major south-dipping detachment zone within the basement (Hall, 2002).

The major compressional events to affect the Mount Bruce Supergroup occurred from c. 2450 to 2200 Ma (Opthalmian Orogeny) and c. 1800 to 1650 Ma (Capricorn Orogeny; Powell, 1999). The Ophthalmian Orogeny was characterized by north–south compression and the northward propagation of a fold-and-thrust belt (F_2). Uplift and erosion occurred along the southern margin of the Hamersley Province at the same time as deposition in a foreland basin just to the north (McGrath Trough). Northward-propagating deformation during deposition resulted in intraformational unconformities and conglomerates containing clasts of underlying units that are conformable elsewhere in the province. Extension associated with orogenic collapse followed soon after the compressional event and resulted in low-angle normal faults oriented east–west and a series of dolerite dykes oriented southeast–northwest.

Following the compression of the Ophthalmian Orogeny, the uplifted southern margin of the Pilbara Craton was rifted and deep-water sediments of the Ashburton Formation were deposited in the region where the terrestrial source for sediments of the McGrath Trough had once been. The second major compression event occurred between c. 1800 and 1650 Ma and is associated with amalgamation of the Pilbara and Yilgarn Cratons. This compression event resulted in the southeast-trending Ashburton Fold Belt (F_3). In the southwest of the Hamersley Province, folding associated with the Ashburton Fold Belt overlaps with that of the Ophthalmia fold-and-thrust belt.

The Hardey Syncline

The Hardey Syncline is at the southern margin of the Hamersley Province and is on the southwestern corner of the MOUNT BRUCE 1:250 000 map sheet (Figs 41 and 44). The structure is a large F_2 structure that exposes an almost complete stratigraphy of the southern Hamersley Province without the need to travel long distances. Excellent exposure of the Fortescue Group, Hamersley Group, and most of the Turee Creek and Wyloo Groups can be seen in the area. The core of the syncline allows the relationship between unconformities and deformation to be examined, and this provides the critical evidence in establishing the timing of deformation and the tectonic evolution of the whole Hamersley Province. The area around the structure therefore provides an excellent opportunity to address some of the controversy regarding the timing of regional-scale F_2 and F_3 fold structures and whether they developed during Opthalmian or Capricorn orogenesis.

Excursion localities — the Hardey Syncline area

Day 6

Today we will visit part of the Hamersley Province that provides a unique and picturesque look at both sedimentological and tectonic processes that occurred from the Archaean through to the Palaeoproterozoic.

Locality 19: Folding and cleavage development in the Hardey Syncline

Return to the Ashburton Downs Road and drive north to a T-junction with the Nannutarra–Paraburdoo Road. At this T-junction turn left and drive along the Nannutarra–Paraburdoo Road towards the east and take a track off to the north at an old roadworks area. The track continues a short distance and enters a creek bed. We will examine a number of localities around this creek bed and on the adjacent hills. Leave the vehicle and continue along the

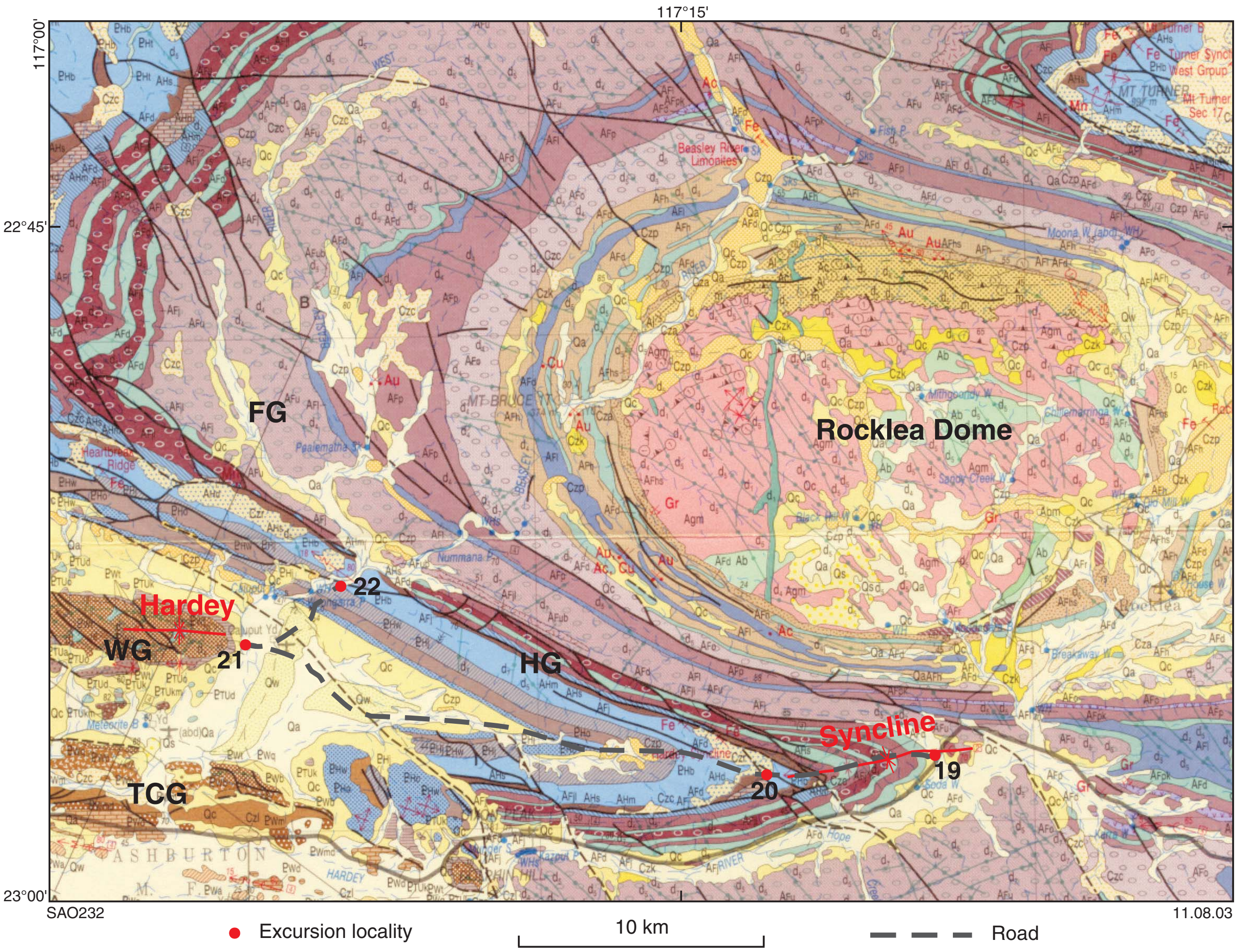

Figure 44. The location of the Hardey Syncline and the Rocklea Dome and Localities 19–22 shown on part of the MOUNT BRUCE 1:250 000 map sheet (Thorne et al., 1996). FG = Fortescue Group, HG = Hamersley Group, TCG = Turee Creek Group, WG = Wyloo Group

creek bed for about 300 m. The outcrop is best viewed from the top of a dolerite mound on the eastern bank of the creek.

At this locality the Marra Mamba Iron Formation is strongly folded in the core of the Hardey Syncline, with large upright ‘M-folds’ evident in outcrop (Fig. 45). The outcrop belongs to the Nammuldi Member, which is the lowermost member of the Marra Mamba Iron Formation. Extensive faulting is evident in the hinge of the syncline. The most obvious faults are steep and northwesterly striking. However, a number of more-subtle faults with ramp-flat geometries, which probably formed broadly synchronously with folding, give rise to stratigraphic omission (e.g. the Wittenoom Formation and parts of the Fortescue Group are locally missing in this area).

Locality 20: Folding and cleavage development in Bruno’s Band

Continue driving westward along the track or creek bed. An old drill site marks the top of a steep small rise. Leave the vehicle here and walk east-southeast to the top of a small ridge about 1 km away.

This locality allows spectacular structures to be seen both close up (in Bruno’s Band) and at a larger scale (in the overlying Dales Gorge Member). Bruno’s Band is a 5–10 m-wide BIF marker that outcrops beneath the Mount Silvia and Mount McCrae shales and can be traced throughout the Hamersley Basin. Outcrops show the noncylindrical nature of small-scale folding and numerous mesoscopic-scale relationships between folding and cleavage development (Fig. 46).

If time and track conditions allow, we will continue westward along the track to the top of the large hill. The top of this hill gives dramatic views and a regional overview of the Hardey Syncline (Fig. 47).

Locality 21: Unconformity at the nose of the Hardey Syncline

Return to the main road and continue heading west. About 14 km farther west, just past an abandoned airstrip on the left side of the road, a track heads off to the right. About 1.5 km along the track is a gate. Continue through the

SAO233 11.08.03

Figure 45. Folding in the Nammuldi Member of the Marra Mamba Iron Formation, Mount Bruce Supergroup, close to the nose of the large-scale Hardey Syncline

SAO234 27.06.03

Figure 46. Complex fold and foliation relationships in F_2 structures developed in Bruno's Band, Hardey Syncline

gate, leaving it as you found it! Follow the track in a northwesterly direction for about 13.5 km where a track leads off to the left. Turn onto the track and follow it for 1.5 km at which point it should curve around to head south at an airstrip. Leave the vehicle and walk 1.5 km west to the large hills.

The rocks on the lower slopes of the hill are siltstone and shale of the lower Turee Creek Group. The beds dip uniformly at 55° towards 210°. A near-vertical, disjunctive, reticulate to slaty cleavage dips towards 190° and is parallel to the axial-plane cleavage in the overlying uppermost Turee Creek Group. Ribs of fine-grained indurated quartz arenite interbedded with cleaved olive-green mudrock can be traced up the hill where they are erosionally truncated by the overlying flat-lying units. The rocks above the contact are siltstone with some fine-grained cross-bedded sandstone. A few scattered pebbles and cobbles up to 10 cm in diameter are in the fine-grained quartzite. Decimetre-scale cross-beds in the sandstone are herringbone style with bipolar current directions oriented north-northwesterly. About 90 m above the basal contact there is a thin (3 m) discontinuous horizon of coarse-grained, poorly sorted lithic conglomerate containing angular fragments of Weeli Wolli Formation; this correlates with the Three Corner Conglomerate Member, which farther west lies at the base of the Beasley River Quartzite (Trendall, 1979). The conglomerate grades upward to a 3.5 m-thick medium-grained quartzose sandstone. The overall aspect of the 6.5-m coset is of braided-stream or sheetflood-sandstone facies (cf. Thorne and Seymour, 1991), in contrast to the inferred shallow-marine or deltaic facies of rocks below. Cross-beds measured 3–10 km to the west in this jasper-bearing unit give a mean flow direction towards the northeast. Immediately overlying the discontinuous conglomerate are cosets of very fine to fine-grained well-sorted quartz arenite with abundant ripple marks and quadrapolar cross-bedding, indicating a shallow-marine environment with a north-northwesterly trending strandline. Disjunctive cleavage both above and below the unconformity contact is axial-planar to the 110°-trending F_2 folds to which the Hardey Syncline belongs.

Locality 22: The Hamersley Group section

Return east along the track to the T-junction and turn left. Continue on to Woongarra Pool and check the water level before crossing. Continue along the track as it heads into Woongarra Gorge. Follow the track as far as possible up the gorge, allowing for room to turn around. If the vehicle

Figure 47. The shallowly westerly plunging Hardey Syncline looking east from the Joffre Member of the Hamersley Group. The synformal closure is in the middle of the picture. The dark prominent ridge defining the fold is the Marra Mamba Iron Formation. Between the photographer and the ridge are well-exposed units of the Brockman Iron Formation. The low irregular ground beyond the ridge are volcanic units of the Fortescue Group

is unable to continue along the gorge we will have to walk about 2 km. Bring bathers for a dip in the beautiful Woongarra Pool. The swimming here is very pleasant.

The entire Hamersley Group is exposed in a 2-km section along the Beasley River. The upstream part of the section where the river first enters the gorge is the Brockman Iron Formation, with some Marra Mamba Iron Formation outcropping in the north-facing hill slopes to the west of the river bed. Minor folds in the Brockman Iron Formation indicate that the regional structure is an upright syncline plunging at about 20° to the west-northwest. Along the gorge, the rocks of the Brockman Iron Formation are well exposed and dip moderately to the south.

The section through the Woongarra Rhyolite is well exposed on the eastern side of the river and is best examined in afternoon light. The unit contains several flows of different thickness. The basal part of the lowermost thick unit contains dark chloritic fragments and cobble- to boulder-sized breccia clasts with some feldspar phenocrysts. This is overlain by a rather massive rock with sparse feldspar phenocrysts in a dark-blue–grey matrix and sporadic angular cobble-sized clasts, some of which are flow banded. Large cooling columns at a high angle to bedding (which dips ~60° to the south-southwest) can be seen in the middle third of the section, in the upper part of which there is an increase in dark wispy lenticular inclusions possibly representing fiamme (pumice fragments). This lenticle-tuff unit has been dated here at 2439 ± 10 Ma (Pidgeon and Horwitz, 1991). Some intrusive phases have been observed in the general area. The overall aspect of the section is remarkably similar to Devonian ash-flow tuffs of the eastern Lachlan Fold Belt, interpreted to have been produced by resurgent cauldron activity in a setting similar to the modern Taupo Zone of North Island, New Zealand. A different interpretation was given by Trendall (1995) who considered the Woongarra Rhyolite to be a very large synsedimentary sill with ‘pepperite’ margins. The top of the Woongara Rhyolite is marked by a 5-m band of intercalated graded beds (turbidites) and some iron formation. This passes upward into monotonous planar bedded shale with no internal bedding structure, which, in turn, passes into the main part of the Boolgeeda Iron Formation.

References

ARNDT, N. T., NELSON, D. R., COMPSTON, W., TRENDALL, A. F., and THORNE, A. M., 1991, The age of the Fortescue Group, Hamersley Basin, Western Australia, from ion microprobe zircon U–Pb results: Australian Journal of Earth Sciences, v. 38, p. 261–281.

BARLEY, M. E., PICKARD, A. L., and SYLVESTER, P. J., 1997, Emplacement of a large igneous province as a possible cause of banded iron formation 2.45 billion years ago: Nature, v. 385, p. 55–58.

BLAKE, T. S., and GROVES, D. I., 1987, Continental rifting and the Archaean–Proterozoic transition: Geology, v. 15, p. 225–232.

BLAKE, T. S., and BARLEY, M. E., 1992, Tectonic evolution of the Late Archaean to Early Proterozoic Mount Bruce Megasequence Set, Western Australia: Tectonics, v. 11, p. 1415–1425.

CAWOOD, P. A., and HOLLINGSWORTH, D. A., 2002, Resolution of the subsurface structure of the Hamersley Province by multichannel seismic reflection: Western Australian Minerals and Energy Research Institute of W.A. (MERIWA), Report 228.

CULVER, K. E., 2001, Structure, metamorphism and geochronology of the northern margin of the Gurin Gutta Granite, central Gascoyne Complex, Western Australia: Perth, Western Australia, Curtin University of Technology, PhD thesis (unpublished).

de la HUNTY, L. E., 1965, Mount Bruce, W.A. (1st edition): Western Australia Geological Survey, 1:250 000 Geological Series Explanatory Notes, 28p.

DREW, B., 1999a, The geology of the Mount James Formation in the Coor-de-wandy Hill area, Western Australia: University of Western Australia, Geology Department, preliminary mapping report (unpublished).

DREW, B., 1999b, The geology of the Mount James Formation in the Coor-de-wandy Hill study area, Western Australia: University of Western Australia, Geology Department, laboratory report (unpublished).

EVANS, D. A. D., SIRCOMBE, K. N., WINGATE, M. T. D., DOYLE, M., McCARTHY, M., PIDGEON, R. T., and VAN NIEKERK, H. S., in prep., Foreland-basin volcanism and internide plutonism in the western Capricorn Orogen at 1805–1785 Ma: diachroneity of the Pilbara–Yilgarn collision: Australian Journal of Earth Sciences.

GEOLOGICAL SURVEY OF WESTERN AUSTRALIA, 1990, Geology and mineral resources of Western Australia, Memoir 3, 827p.

HALL, C. E., 2002, Sedimentology, geochemistry and palaeogeography of the Lower Forescue Group, Hamersley Province, Western Australia: University of Western Australia, PhD thesis (unpublished).

HALL, C. E., POWELL, C. McA., and BRYANT, J., 2001, Basin setting and age of the Late Palaeoproterozoic Capricorn Formation, Western Australia: Australian Journal of Earth Sciences, v. 48, p. 731–744.

HALLIGAN, R., and DANIELS, J. L., 1964, The Precambrian geology of the Ashburton Valley region, northwest division: Western Australia Geological Survey, Annual Report 1963, p. 38–46.

HORWITZ, R. C., 1982, Geological history of the Early Proterozoic Paraburdoo Hinge Zone, Western Australia: Precambrian Research, v. 19, p. 191–200.

HUNTER, W. M., 1990, Mount James Formation, *in* Geology and mineral resources of Western Australia: Western Australia Geological Survey, Memoir 3, p. 221–223.

KRAPEZ, B., 1999, Stratigraphic record of an Atlantic-type global tectonic cycle in the Palaeoproterozoic Ashburton Province of Western Australia: Australian Journal of Earth Sciences, v. 46, p. 71–87.

KRAPEZ, B., and McNAUGHTON, N. J., 1999, SHRIMP zircon U–Pb age and tectonic significance of the Palaeoproterozoic Boolaloo Granodiorite in the Ashburton Province, Western Australia: Australian Journal of Earth Sciences, v. 46, p. 283–287.

LIBBY, W. G., de LAETER, J. R., and ARMSTRONG, R. A., 1999, Proterozoic biotite Rb–Sr dates in the northwestern part of the Yilgarn Craton, Western Australia: Australian Journal of Earth Sciences, v. 46, p. 851–860.

LIBBY, W. G., de LAETER, J. R., and MYERS, J. S., 1986, Geochronology of the Gascoyne Province: Western Australia Geological Survey, Report 20, 31p.

LIN, S., and JIANG, D., 2001, Using along-strike variation in strain and kinematics to define the movement direction of curved transpressional shear zones: an example from northwestern Superior Province, Manitoba: Geology, v. 29 (9), p. 767–770.

MacLEOD, W. N., de la HUNTY, L. E., JONES, W. R., and HALLIGAN, R., 1963, A preliminary report on the Hamersley Iron Province, northwest division: Western Australia Geological Survey, Annual Report 1962, p. 90–100.

MANN, A. W., and HORWITZ, R. C., 1979, Groundwater calcrete deposits in Australia: Some observations from Western Australia: Journal of the Geological Society of Australia, v. 26, p. 293–303.

MARTIN, D. M., LI, Z. X., NEMCHIN, A. A., and POWELL, C. M., 1998, A pre-2.2 Ga age for giant hematite ores of the Hamersley Province, Australia?: Economic Geology, v. 93, p. 1084–1090.

MARTIN, D. McB., POWELL, C. McA., and GEORGE, A. D., 2000, Stratigraphic architecture and evolution of the early Paleoproterozoic McGrath Trough, Western Australia: Precambrian Research, v. 99, p. 33–64.

MARTIN, D. McB., and THORNE, A. M., in press, Tectonic setting and basin evolution of the Bangemall Supergroup in the northwestern Capricorn Orogen: Precambrian Research.

MARTIN, D. McB., and THORNE, A. M., 2001, New insights into the Bangemall Supergroup: Western Australia Geological Survey, Record 2001/5, p. 1–2.

MARTIN, D. McB., and THORNE, A. M., 2002, Revised lithostratigraphy of the Mesoproterozoic Bangemall Supergroup on the Edmund and Turee Creek 1:250 000 map sheets, Western Australia: Western Australia Geological Survey, Record 2002/15, 27p.

MARTIN, D. McB., THORNE, A. M., and COPP, I. A., 1999, A provisional revised stratigraphy for the Bangemall Group on the Edmund 1:250 000 sheet: Western Australia Geological Survey, Annual Review 1998–99, p. 51–55.

MORRIS, R. C., and HORWITZ, R. C., 1983, The origin of the iron-formation-rich Hamersley Group of Western Australia — deposition on a platform: Precambrian Research, v. 21, p. 273–297.

MUHLING, P. C., and BRAKEL, A. T., 1985, Geology of the Bangemall Group — the evolution of an intracratonic Proterozoic basin: Western Australia Geological Survey, Bulletin 128, 266p.

MYERS, J. S., 1990a, Gascoyne Complex, *in* Geology and mineral resources of Western Australia: Western Australia Geological Survey, Memoir 3, p. 198–202.

MYERS, J. S., 1990b, Part 1 — Summary of the Narryer Gneiss Complex, *in* Third International Archaean Symposium, Perth, 1990, Excursion Guidebook *edited by* S. E. HO, J. S. GLOVER, J. S. MYERS, and J. R. MUHLING: University of Western Australia, Geology Department and University Extension, Publication no. 21, p. 62–71.

MYERS, J. S., 1990c, Precambrian tectonic evolution of part of Gondwana, southwestern Australia: Geology, v. 18, p. 537–540.

MYERS, J. S., 1995, The generation and assembly of an Archaean supercontinent — evidence from the Yilgarn Craton, Western Australia, *in* Early Precambrian processes *edited by* M. P. COWARD and A. C. RIES: The Geological Society of London, Special Publication no. 95, p. 143–154.

MYERS, J. S., and HOCKING, R. M., 1998, Geological map of Western Australia, 1:2 500 000 (13th edition): Western Australia Geological Survey.

NELSON, D. R., 1995, Compilation of SHRIMP U–Pb zircon geochronology data, 1994: Western Australia Geological Survey, Record 1995/3, 243p.

NELSON, D. R., 1997, Compilation of SHRIMP U–Pb zircon geochronology data, 1996: Western Australia Geological Survey, Record 1997/2, 189p.

NELSON, D. R., 1998, Compilation of SHRIMP U–Pb zircon geochronology data, 1997: Western Australia Geological Survey, Record 1998/2, 242p.

NELSON, D. R., 1999, Compilation of geochronology data, 1998: Western Australia Geological Survey, Record 1999/2, 222p.

NELSON, D. R., 2000, Compilation of geochronology data, 1999: Western Australia Geological Survey, Record 2000/2, 251p.

NELSON, D. R., 2001, Compilation of geochronology data, 2000: Western Australia Geological Survey, Record 2001/2, 205p.

NELSON, D. R., TRENDALL, A. F., and ALTERMANN, W., 1999, Chronological correlations between the Pilbara and Kaapvaal Cratons: Precambrian Research, v. 97, p. 165–189.

NUTMAN, A. P., and KINNY, P. D., 1994, SHRIMP zircon geochronology of the southern Gascoyne Province and the northwestern margin of the Yilgarn Craton, W.A., *in* Geoscience Australia — 1994 and beyond 12th Australian Geological Convention; Perth, W.A., 1994: Geological Society of Australia, Abstracts, no. 37, p. 320–321.

NUTMAN, A. P., KINNY, P. D., COMPSTON, W., and WILLIAMS, I. S., 1991, SHRIMP U–Pb zircon geochronology of the Narryer Gneiss Complex, Western Australia: Precambrian Research, v. 52, p. 275–300.

OCCHIPINTI, S. A., and MYERS, J. S., 1999, Geology of the Moorarie 1:100 000 sheet: Western Australia Geological Survey, 1:100 000 Geological Series Explanatory Notes, 20p.

OCCHIPINTI, S. A., MYERS, J. S., SHEPPARD, S., TYLER, I. M., and SWAGER, C. P., in prep., Robinson Range, W.A. (2nd edition): Western Australia Geological Survey, 1:250 000 Geological Series Explanatory Notes.

OCCHIPINTI, S. A., and REDDY, S. M., in press, Deformation in a complex crustal-scale shear zone: Errabiddy Shear Zone, Western Australia: Geological Society of London.

OCCHIPINTI, S. A., and SHEPPARD, S., 2001, Geology of the Glenburgh 1:100 000 sheet: Western Australia Geological Survey, 1:100 000 Geological Series Explanatory Notes, 37p.

OCCHIPINTI, S. A., SHEPPARD, S., MYERS, J. S., TYLER, I. M., and NELSON, D. R., 2001, Archaean and Palaeoproterozoic geology of the Narryer Terrane (Yilgarn Craton) and the southern Gascoyne Complex (Capricorn Orogen), Western Australia — a field guide: Western Australia Geological Survey, Record 2001/8, 70p.

OCCHIPINTI, S. A., SHEPPARD, S., PASSCHIER, C., TYLER, I. M., and NELSON, D. R., in press, Palaeoproterozoic crustal accretion and collision in the southern Gascoyne Complex: The Glenburgh Orogeny: Precambrian Research.

OCCHIPINTI, S. A., SHEPPARD, S., TYLER, I. M., and NELSON, D. R., 1999b, Deformation and metamorphism during the c. 2000 Ma Glenburgh Orogeny and c. 1800 Ma Capricorn Orogeny, *in* Two billion years of tectonics and mineralisation *edited by* G. R. WATT and D. A. D. EVANS: Geological Society of Australia, Abstracts, no. 56, p. 26–29.

OCCHIPINTI, S. A., SHEPPARD, S., NELSON, D. R., MYERS, J. S., and TYLER, I. M., 1998, Syntectonic granite in the southern margin of the Palaeoproterozoic Capricorn Orogen, Western Australia: Australian Journal of Earth Sciences, v. 45, p. 509–512.

OCCHIPINTI, S. A., SHEPPARD, S., and TYLER, I. M., 1999a, The Palaeoproterozoic tectonic evolution of the southern margin of the Capricorn Orogen, Western Australia: Geological Society of Australia, Abstracts, no. 53, p. 173–174.

OCCHIPINTI, S. A., SWAGER, C. P., and PIRAJNO, F., 1996, Structural and stratigraphic relationships of the Padbury Group, Western Australia — implications for tectonic history: Western Australia Geological Survey, Annual Review 1995–96, p. 88–95.

PEARSON, J. M., 1996, Alkaline rocks of the Gifford Creek Complex, Gascoyne Province, Western Australia — their petrogenetic and tectonic significance: University of Western Australia, PhD thesis (unpublished).

PIDGEON, R. T., and HALLBERG, J. A., 2000, Age relationships in supracrustal sequences of the northern part of the Murchison Terrane, Archaean Yilgarn Craton, Western Australia: a combined field and zircon U–Pb study: Australian Journal of Earth Sciences, v. 47, p. 153–165.

PIDGEON, R. T., and HORWITZ, R. C., 1991, The origin of olistoliths in Proterozoic rocks of the Ashburton Trough, Western Australia, using zircon U–Pb isotopic characteristics: Australian Journal of Earth Sciences, v. 38, p. 55–63.

PIRAJNO, F., OCCHIPINTI, S. A., and SWAGER, C. P., 1998, Geology and tectonic evolution of the Palaeoproterozoic Bryah, Padbury, and Yerrida Basins (formerly Glengarry Basin), Western Australia — implications for the history of the south-central Capricorn Orogen: Precambrian Research, v. 90, p. 119–140.

POWELL, C. McA., 1999, Structural and tectonic evolution of the Palaeoproterozoic Ophthalmia Fold-and-Thrust Belt: Geological Society of Australia, Abstracts, no. 53, p. 202–203.

POWELL, C. McA., and HORWITZ, R.C., 1994, Late Archaean and Early Proterozoic tectonics and basin formation of the Hamersley Ranges: Geological Society of Australia (W.A. Division), 12th Australian Geological Convention, Perth, W.A., 1994, Excursion Guidebook no. 4, 53p.

POWELL, C. McA., and LI, Z. X., 1991, New evidence for the age of deformation along the southern margin of the Hamersley Province: relevance to the palaeogeographic evolution and time of iron-ore formation: Geological Society of Australia, Abstracts, no. 25, p. 52–53.

SCHIØTTE, L., and CAMPBELL, I. H., 1996, Chronology of the Mount Magnet granite–greenstone terrain, Yilgarn Craton, Western Australia: implications for field based predictions of the relative timing of granitoid emplacement: Precambrian Research, v. 78, p. 237–260.

SEYMOUR, D. B., THORNE, A. M., and BLIGHT, D. F., 1988, Wyloo, W.A.: Western Australia Geological Survey, 1:250 000 Geological Series Explanatory Notes, 36p.

SHEPPARD, S., MARTIN, D. M., and THORNE, A. M., 2000, Mangaroon, W.A. Sheet 2050 (preliminary): Western Australia Geological Survey, 1:100 000 Geological Series.

SHEPPARD, S., and OCCHIPINTI, S. A., 2000, Geology of the Errabiddy and Landor 1:100 000 sheets: Western Australia Geological Survey, 1:100 000 Geological Series Explanatory Notes, 37p.

SHEPPARD, S., OCCHIPINTI, S. A., NELSON, D. R., and TYLER, I. M., 1999a, Granites of the southern Capricorn Orogen, Western Australia, *in* Two billion years of tectonics and mineralisation, *edited by* G. R. WATT and D. A. D. EVANS: Geological Society of Australia, Abstracts, no. 56, p. 44–46.

SHEPPARD, S., OCCHIPINTI, S. A., NELSON, D. R., and TYLER, I. M., 1999b, The significance of c. 2.0 Ga crust along the southern margin of the Gascoyne Complex: Western Australia Geological Survey, Annual Review 1998–99, p. 56–61.

SHEPPARD, S., OCCHIPINTI, S. A., and TYLER, I. M., 2002. The relationship between the tectonic setting and composition of granitoid magmas, Yarlarweelor Gneiss Complex, Western Australia: Lithos, v. 66, p. 133–154.

SHEPPARD, S., OCCHIPINTI, S. A., and TYLER, I. M., in press, Palaeoproterozoic (2005–1970 Ma) granites related to convergence and collision of the Gascoyne Complex and Yilgarn Craton, Western Australia: Precambrian Research.

SHEPPARD, S., and SWAGER, C. P., 1999, Geology of the Marquis 1:100 000 sheet: Western Australia Geological Survey, 1:100 000 Geological Series Explanatory Notes, 21p.

SIRCOMBE, K. N., in prep., Age of the Mount Boggola Volcanic succession and further geological constraint on the Ashburton Basin, Western Australia: Australian Journal of Earth Sciences.

STRECKEISEN, A., 1976, To each plutonic rock its proper name: Earth Science Reviews, v. 12, p. 1–33.

THORNE, A. M., and MARTIN, D. McB., 2002, Capricorn, W. A. Sheet 2251: Western Australia Geological Survey, 1:100 000 Geological Series.

THORNE, A. M., MARTIN, D. McB., and COPP, I. A., 2002, Capricorn, W.A. Sheet 2251: Western Australia Geological Survey, 1:100 000 Geological Series.

THORNE, A. M., and SEYMOUR, D. B., 1991, The Geology of the Ashburton Basin: Western Australia Geological Survey, Bulletin 139, 141p.

THORNE, A. M., and TRENDALL, A. F., 2001, Geology of the Fortescue Group, Pilbara Craton, Western Australia: Western Australia Geological Survey, Bulletin 144, 249p.

THORNE, A., M., and TYLER, I. M., 1996, Geology of the Rocklea 1: 100 000 sheet: Western Australia Geological Survey, 1:100 000 Geological Series Explanatory Notes, 15p.

THORNE, A. M., TYLER, I. M., BLOCKLEY, J. G., and BLIGHT, D. F., 1996, Mount Bruce, W.A. Sheet SF 50-11 (2nd edition): Western Australia Geological Survey, 1:250 000 Geological Series.

THORNE, A. M., TYLER, I. M., and HUNTER, W. M., 1991, Turee Creek, W.A.: Western Australia Geological Survey, 1:250 000 Geological Series Explanatory Notes, 29p.

TRENDALL, A. F., 1979, A revision of the Mount Bruce Supergroup, *in* Western Australia Geological Survey, Annual Report 1978, p. 63–71.

TRENDALL, A. F., 1990, The Hamersley Basin, *in* Geology and mineral resources of Western Australia: Western Australia Geological Survey, Memoir 3, p. 163–191.

TRENDALL, A. F., 1995, The Woongarra Rhyolite — a giant lava-like felsic sheet in the Hamersley Basin of Western Australia: Western Australia Geological Survey, Report 42, 70p.

TRENDALL, A. F., and BLOCKLEY, J. G., 1970, The iron formations of the Precambrian Hamersley Group, Western Australia: Western Australia Geological Survey, Bulletin 119, 366p.

TRENDALL, A. F., NELSON, D. R., de LAETER, J. R., and HASSLER, S., 1998, Precise zircon U–Pb ages from the Marra Mamba Iron Formation and Wittenoom Formation, Hamersley Group, Western Australia: Australian Journal of Earth Sciences, v. 45, p. 137–142.

TYLER, I. M., 1991, The geology of the Sylvania Inlier and the southeast Hamersley Province: Western Australian Geological Survey, Bulletin 138, 108p.

TYLER, I. M., and HOCKING, R. M., 2001, Tectonic units of Western Australia (scale 1:2 500 000): Western Australia Geological Survey.

TYLER, I. M., and THORNE, A. M., 1990a, The northern margin of the Capricorn Orogen, Western Australia — an example of an early Proterozoic collision zone: Journal of Structural Geology, v. 12, p. 685–701.

TYLER, I. M. and THORNE, A. M., 1990b, Structural evolution of the northern margin (Capricorn Orogen), *in* Geology and mineral resources of Western Australia: Western Australia Geological Survey, Memoir 3, p. 223–232.

TYLER, I. M., HUNTER, W. M., and WILLIAMS, I. R., 1991, Newman W.A.: Western Australia Geological Survey, 1:250 000 Geological Series Explanatory Notes, 36p.

VARVELL, C., 2001, Age, structure and metamorphism of a section of the Morrissey Metamorphic Suite, central Gascoyne Complex, Western Australia: Perth, Western Australia, Curtin University of Technology, PhD thesis (unpublished).

WIEDENBECK, M., and WATKINS, K. P., 1993, A time scale for granitoid emplacement in the Archaean Murchison Province, Western Australia, by single zircon geochronology: Precambrian Research, v. 61, p. 1–26.

WILLIAMS, I. R., and MYERS, J. S., 1987, Archaean geology of the Mount Narryer region, Western Australia: Western Australia Geological Survey, Report 22, 32p.

WILLIAMS, S. J., 1986, Geology of the Gascoyne Province, Western Australia: Western Australia Geological Survey, Report 15, 85p.

WILLIAMS, S. J., WILLIAMS, I. R., CHIN, R. J., MUHLING, P. C., and HOCKING, R. M., 1983a, Mount Phillips W.A.: Western Australia Geological Survey, 1:250 000 Geological Series Explanatory Notes, 29p.

WILLIAMS, S. J., WILLIAMS, I. R., and HOCKING, R. M., 1983b, Glenburgh, W.A.: Western Australia Geological Survey, 1:250 000 Geological Series Explanatory Notes, 25p.

WINGATE, M. T. D., and GIDDINGS, J. W., 2000, Age and palaeomagnetism of the Mundine Well dyke swarm, Western Australia: implications for an Australia–Laurentia connection at 755 Ma: Precambrian Research, v. 100, p. 335–357.

WYBORN, L. A. I., WYBORN, D., WARREN, R. G., and DRUMMOND, B. J., 1992, Proterozoic granite types in Australia: implications for lower crust composition, structure and evolution: Transactions of the Royal Society of Edinburgh: Earth Sciences, v. 83, p. 201–209.

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