Government of **Western Australia**
Department of **Mines and Petroleum**

RECORD 2010/3

EVOLUTION OF ACTIVE PLATE MARGINS: WEST PILBARA SUPERTERRANE, DE GREY SUPERBASIN, AND THE FORTESCUE AND HAMERSLEY BASINS — A FIELD GUIDE

by

AH Hickman, RH Smithies, and IM Tyler

2010 Perth 5ias

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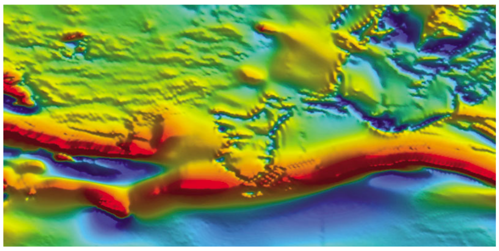


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Perth 2010

Geological Survey of Western Australia

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Preface

This field guide was prepared for a six-day excursion, associated with the Fifth International Archean Symposium, across the northwestern Pilbara Craton and sections of the Fortescue and Hamersley Basins (Figs 1 and 2). The guide has two parts: **Part 1** reviews the geology of the tectonic units; **Part 2** provides descriptions of excursion localities and directions for travel between them.

The Mesoarchean volcanic and sedimentary successions of the west Pilbara provide the best evidence on Earth that plate tectonic processes similar to those of today were operating by at least c. 3200 Ma; the field localities described in this guide provide the evidence for interpreting the geotectonic evolution of the West Pilbara Superterrane as a Mesoarchean active plate margin. The excursion route includes sections through all three terranes of the West Pilbara Superterrane, including fragments of a 3220–3180 Ma passive margin succession, which overlies >3270 Ma continental crust; a 3220–3165 Ma oceanic slab obducted between 3160 and 3070 Ma; and a 3130–3110 Ma volcanic arc. The guide also covers younger successions of the De Grey Superbasin, including a 3010–2990 Ma continental volcanic succession (either a pull-apart basin or a volcanic arc), and 2950 Ma boninite-like rocks and sanukitoids related to late-tectonic orogenic relaxation, slab-breakoff, and remobilization of a trace-element enriched mantle source. Overall, the sequence of tectonic settings in the west Pilbara is very similar to that of a Wilson cycle.

In addition, two days during the excursion provide a traverse across the Neoarchean–Paleoproterozoic Fortescue and Hamersley Basins looking at evidence for their tectonic settings, from rifting at 2780 Ma, through a passive margin setting at 2630–2400 Ma, to collision during the Ophthalmian Orogeny at c. 2210 Ma.

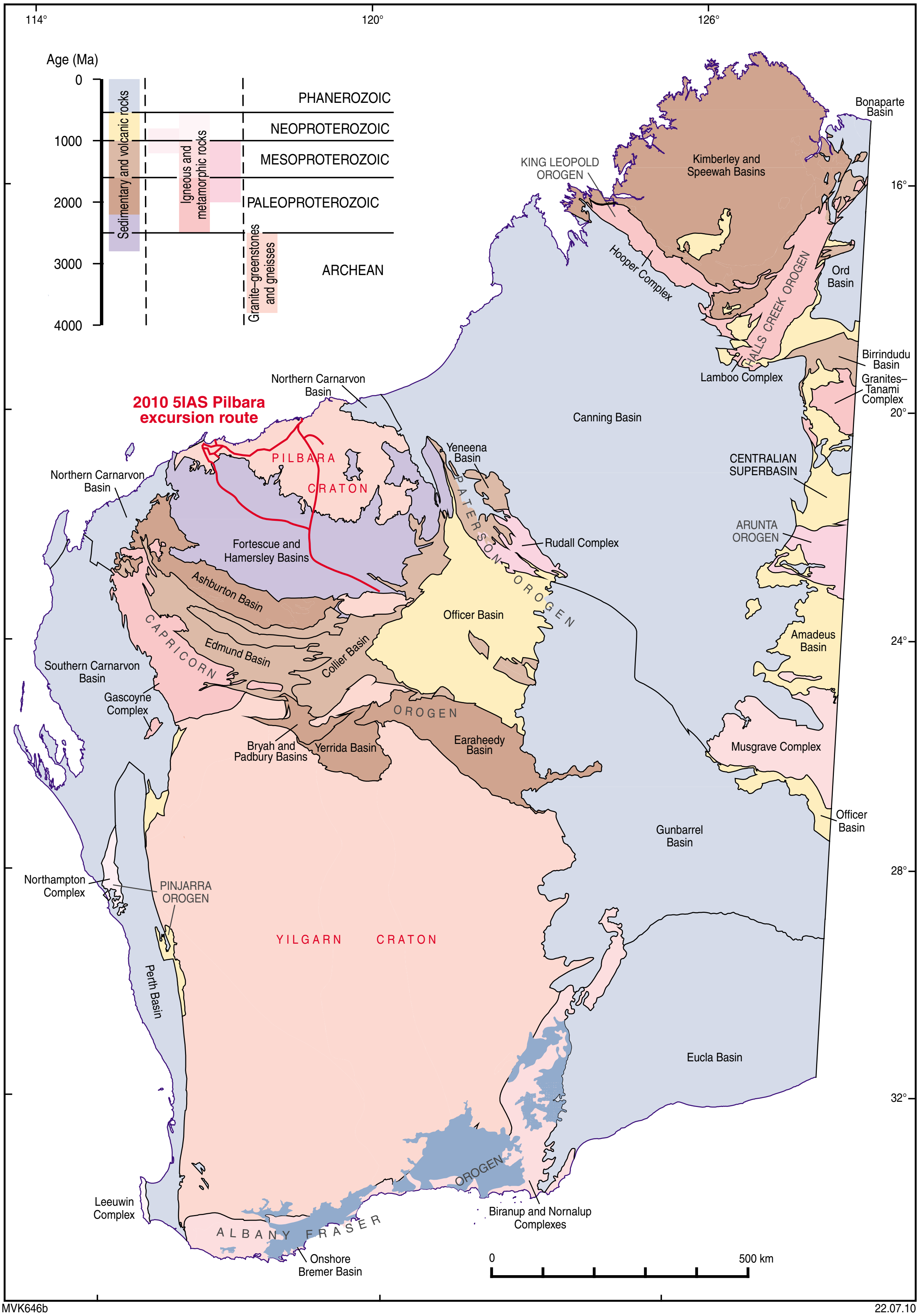


Figure 1. Map of Western Australia, showing the broad tectonic units, and location of the fieldtrip area covered in this guidebook.

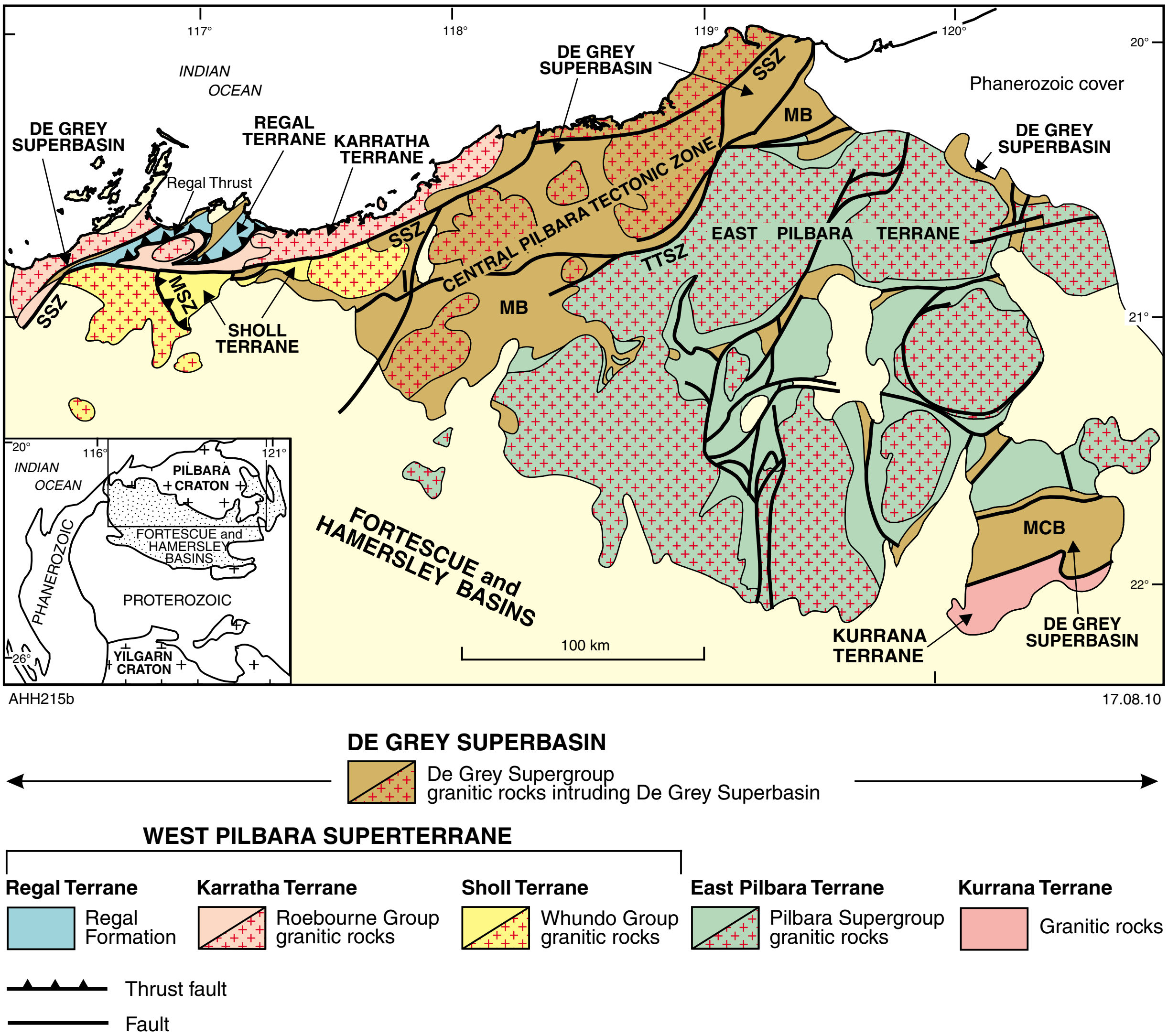


Figure 2. Simplified geological map of the northwestern Pilbara Craton, showing terranes plus the De Grey Superbasin. Abbreviations: MB = Mallina Basin; MCB = Mosquito Creek Basin; SSZ = Sholl Shear Zone; MSZ = Maitland Shear Zone; TTSZ = Tabba Tabba Shear Zone

Part 1. Evolution of the West Pilbara Superterrane, De Grey Superbasin, and the Fortescue and Hamersley Basins

Pilbara Craton

The Pilbara Craton is a 250 000 km^2 ovoid segment of 3800–2830 Ma Archean crust underlying the northwestern part of Western Australia. The southern 70% of the craton is concealed by unconformably overlying 2775–2450 Ma rocks of the Fortescue and Hamersley Basins, but in the north the craton is exposed over 60,000 km^2. The Pilbara Craton was recently redefined (Van Kranendonk et al., 2006) to exclude the overlying Fortescue and Hamersley Basins; although Trendall (1990a) previously included these Neoarchean–Paleoproterozoic successions on the grounds that both they and the underlying granite–greenstones did not attain stability until 2400 Ma. However, recent mapping and geochronology have revealed that stability of the terranes and basins of the Pilbara Craton was essentially attained at c. 2895 Ma. This followed the final stage of terrane accretion, with a major event of orogenic deformation and metamorphism; in essence, this was the 'cratonization event'. Rifting of the Pilbara Craton at c. 2775 Ma, and the first outpouring of Fortescue Group continental flood basalts (Thorne and Trendall, 2001), marked the beginning of a new stage in the evolution of the Pilbara crust. Some of the major structures (especially synforms) in the granite–greenstones of the East Pilbara Terrane were reactivated during deposition of the lower part of the 2775–2630 Ma Fortescue Group, but this deformation was comparatively minor, and occurred at least 140 million years after the c. 2900 Ma cratonization event.

Geological knowledge of the Pilbara Craton is derived almost entirely from the northern Pilbara, where it comprises six main tectonic units, in order of decreasing age:

1. Early crust, 3800–3530 Ma: remnants are only very rarely exposed, and its existence is inferred mainly from geochronological data
2. East Pilbara Terrane, 3525–3220 Ma: a granite–greenstone terrane consisting of the predominantly volcanic succession of the Pilbara Supergroup, and four contemporaneous granitic supersuites
3. West Pilbara Superterrane, 3270–3070 Ma: three granite–greenstone terranes, independently formed between 3270 and 3100 Ma, and accreted by plate convergence at 3070 Ma
4. Kurrana Terrane, 3200–2895 Ma: two granitic supersuites, with greenstones of uncertain age
5. De Grey Superbasin, 3050–2930 Ma: five depositional basins, and three contemporaneous mainly granitic supersuites
6. Split Rock Supersuite, 2890–2830 Ma: late- to post-tectonic granitic intrusions

The interpreted bedrock geology of the Pilbara Craton is shown in Figure 3, and the regional lithostratigraphy of the craton (excluding the Kurrana Terrane) is summarized in Figure 4.

Tectonic models

A variety of tectonic models have been applied to explain the tectonic evolution of Archean granite–greenstones including the Pilbara Craton. Early geological interpretations and tectonic models have previously been reviewed by Anhaeusser et al. (1969), Windley and Bridgwater (1971), and Windley (1977). These early accounts were disadvantaged by very limited geochronology, perhaps explaining why little consideration was apparently given to the possibility that processes dominating tectonic evolution may have changed from the Paleoarchean (e.g. development of early crust in the east Pilbara and southern Africa) to the Neoarchean (e.g. Superior Province and the eastern Yilgarn Craton).

Many workers in the Archean accept the uniformitarian interpretation that tectonic processes during this eon were similar to tectonic processes operating today. Accordingly, many Archean granite–greenstones terrains have been interpreted using Phanerozoic-style plate tectonic paradigms (Trendall, 1995a). Other workers have argued that there is no conclusive evidence for plate-tectonic environments anywhere in the Archean geological record (Hamilton, 1998; McCall, 2003, 2010); and still others have suggested that plate tectonics were absent until the late Archean (e.g. Taylor and McLennan, 2009).

A large amount of accumulated evidence from the Pilbara Craton (Hickman, 2001a, 2004a; Smithies, 2000; Smithies and Champion, 2000; Smithies et al., 2003, 2004, 2005a,b, 2007a; Van Kranendonk et al., 2006, 2007a,b, 2010) indicates that there was a major change in crustal processes and tectonic environments between c. 3220 and c. 3070 Ma. From 3220 Ma to 2900 Ma the northwestern part of the craton evolved through stages characteristic of Phanerozoic-style plate tectonics. These stages, outlined

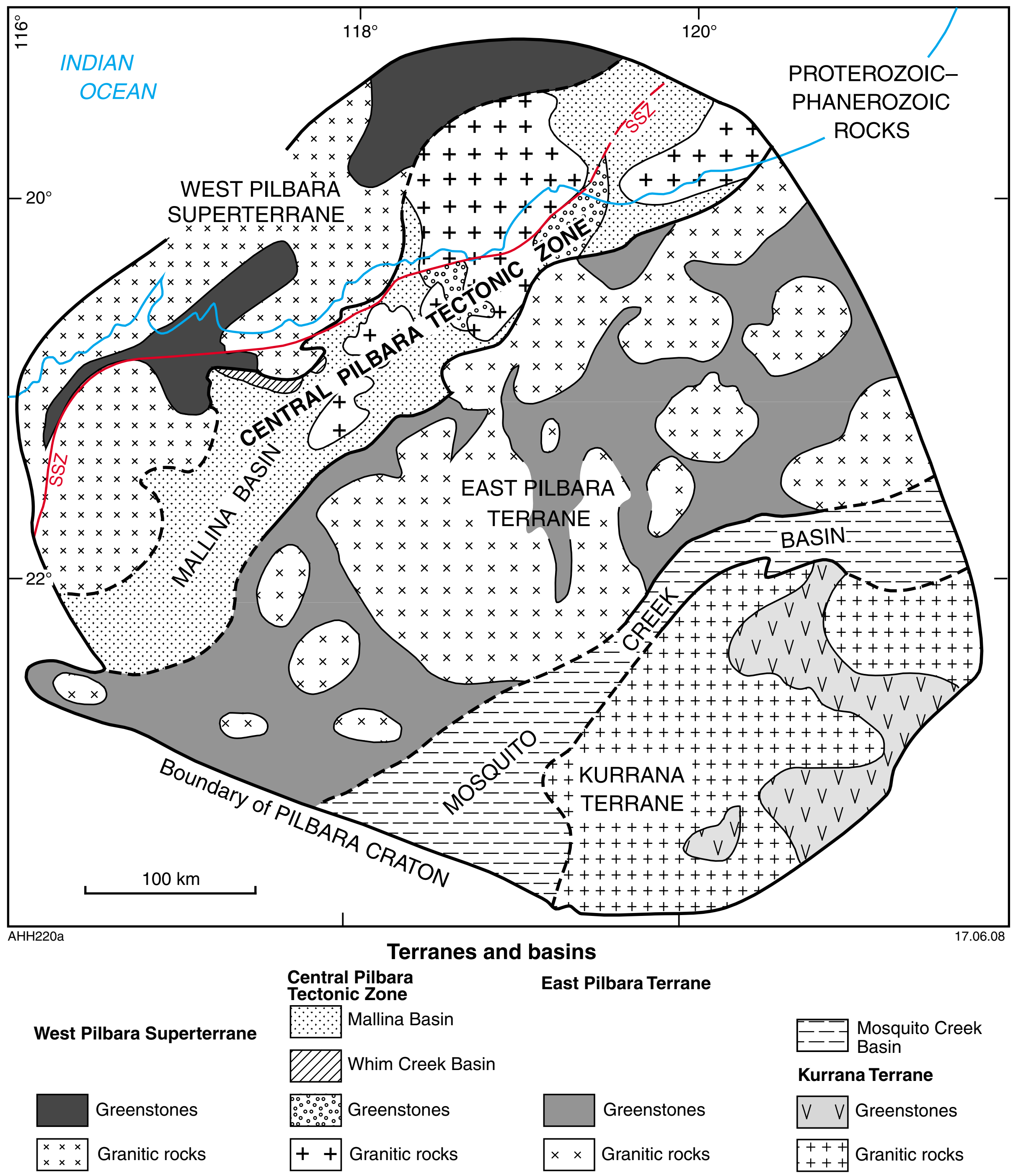


Figure 3. **Interpretation of the granite–greenstone geology of the Pilbara Craton beneath the Fortescue and Hamersley Basins (modified from Hickman, 2004a; based on regional gravity and magnetic data in Blewett et al., 2000)**

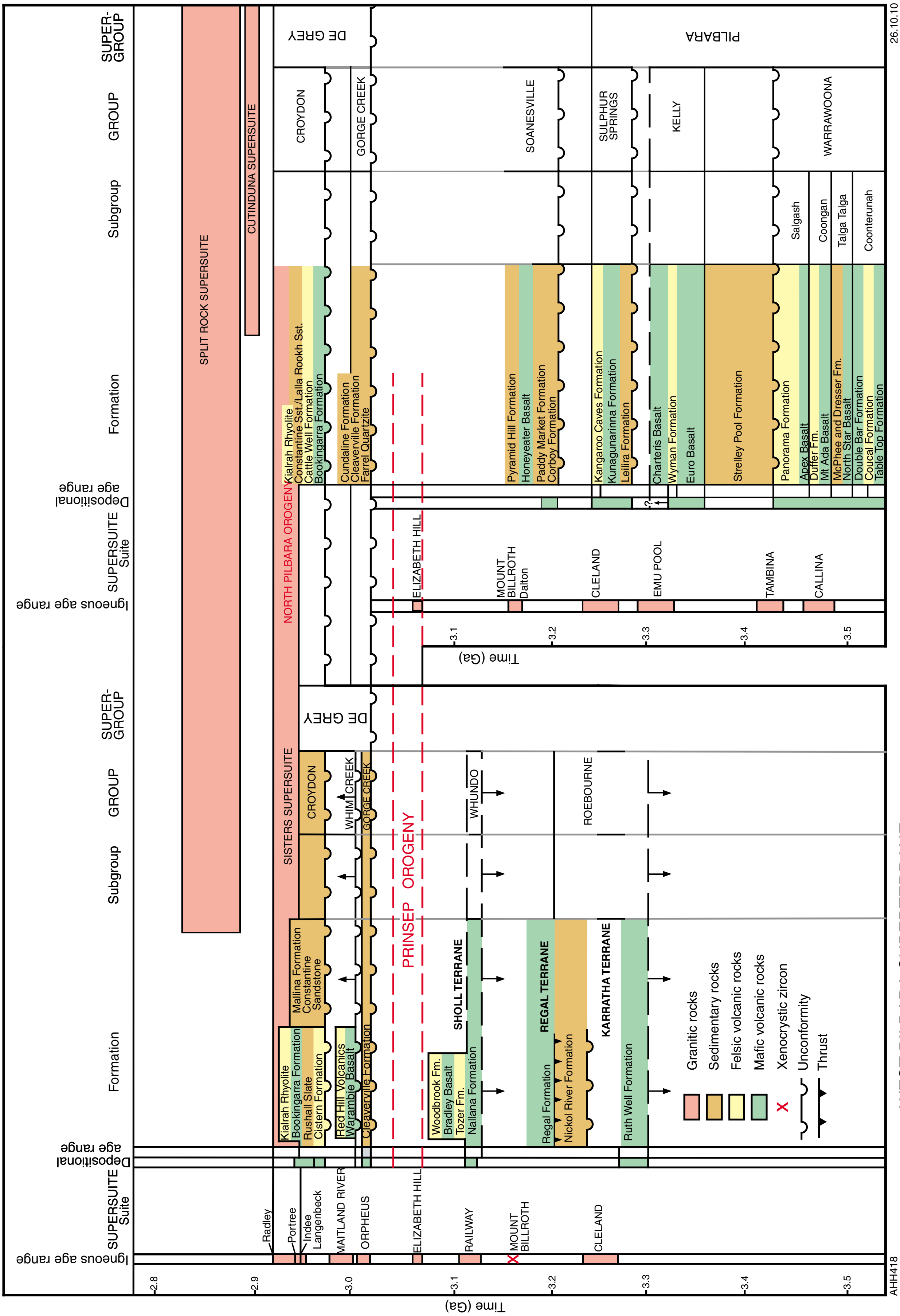


Figure 4. **Comparative lithostratigraphy of the East Pilbara Terrane, West Pilbara Superterrane, and De Grey Superbasin, Pilbara Craton (modified from Van Kranendonk et al., 2006).**

below, have collectively been interpreted to represent a Mesoarchean Wilson cycle (Van Kranendonk et al., 2010).

- rifted continental margins with continental shelf deposition (3220–3165 Ma)
- formation of MORB-like oceanic crust (3200–3150 Ma)
- convergence, subduction of oceanic crust, and formation of an intra-oceanic volcanic arc (3135–3100 Ma)
- accretion and collisional orogeny (3070–3060 Ma)
- peneplanation, post-orogenic extension, and subsidence, with development of a continental basin (3050–3016 Ma)
- subduction northwest of the present craton margin with a southeast-dipping slab, and the establishment of a magmatic arc and back-arc basin parallel to the northwestern margin of the craton (3010–2980 Ma)
- post-subduction removal (break-off) of the fossil slab, and southeastward migration of hot mantle material, with melting of subduction-modified mantle beneath the central part of the western Pilbara, and granitic intrusion of overlying clastic sediments in strike–slip basins (2950–2940 Ma)
- final convergence, closure of all basins, and major orogeny, completing cratonization (2940–2900 Ma)

In various parts of the Paleoarchean East Pilbara Terrane, some researchers have invoked plate tectonic processes to explain their observations (Bickle et al., 1980, 1983, 1985, 1989; Zegers et al., 1996, 2001; van Haaften and White, 1998; Kloppenburg et al., 2001), but there is now considerable regional evidence (combined stratigraphic, structural, metamorphic, geochronological, and geochemical) suggesting that Phanerozoic-style plate tectonic processes did not play a significant role (e.g. Green et al., 2000; Smithies, 2000; Hickman and Van Kranendonk, 2004; Van Kranendonk and Pirajno, 2004; Smithies et al., 2005b, 2007b; Van Kranendonk et al., 2007a,b).

Stratigraphy and evolution

The summary presented here of the lithostratigraphy and tectonic evolution of the northern Pilbara Craton, and the Fortescue and Hamersley Basins, is based on extensive field and analytical data collected by the Geological Survey of Western Australia (GSWA) and others since the 1980s. This guide briefly reviews previous interpretations with comments on their regional application (see '**Previous stratigraphic and tectonic interpretations**'). The field excursion described in Part 2 does not include visits to the East Pilbara and Kurrana Terranes; however, all the major tectonic units in the Pilbara Craton are reviewed. .

Early crust

The Warrawagine Dome in the northeastern Pilbara Craton (Fig. 5) includes enclaves of 3655–3576 Ma biotite tonalite gneiss (GSWA 142870: Nelson, 1999) within post-3500 Ma granodiorite and monzogranite (Williams, 2001a), and xenoliths of 3578 Ma gabbroic anorthosite have been reported (McNaughton et al., 1988) within c. 3430 Ma granitic rocks on the western margin of the Shaw Dome. Extensive SHRIMP U–Pb geochronology across the northern Pilbara Craton has revealed that many Paleoarchean and Mesoarchean siliciclastic formations contain abundant c. 3600 Ma detrital zircons. Even older detrital zircons are present in the De Grey Supergroup (3795 Ma in the Constantine Sandstone — GSWA 142942: Nelson, 2000; 3714 Ma in the Mosquito Creek Formation — GSWA 178010: Nelson, 2005), and the Warrawoona Group (3724 Ma in the Panorama Formation; Thorpe et al., 1992).

The presence of a pre-3500 Ma basement to the East Pilbara Terrane is supported by Nd-isotopic data (Fig. 6), and Jahn et al. (1981) also reported isotopic data indicating that the lower part of the Warrawoona Group was derived from crust older than 3560 Ma. Smithies et al. (2009) showed that most pre- c. 3300 Ma granites within the northeastern Pilbara Craton formed through infracrustal melting of a source older than 3500 Ma (Fig. 7). Smith et al. (1998) reported Sm–Nd data that parts of the Karratha Terrane were derived from crust only slightly younger (T_{DM} 3494–3470 Ma).

East Pilbara Terrane

The 3525–3225 Ma East Pilbara Terrane provides the world's most complete record of Paleoarchean crustal evolution. The major crustal structures of the East Pilbara Terrane (granite–greenstone domes) have no analogues in Phanerozoic crust, and few close analogues in any of the world's numerous Proterozoic terranes. Suggestions that these structures can be explained by either cross-folding (Blewett, 2002), or deformation of metamorphic core complexes (Zegers et al., 1996; Kloppenburg et al., 2001), are considered to be based on inadequate geological evidence, as discussed by Hickman and Van Kranendonk (2004). The exposed part of the East Pilbara Terrane is composed of eight steep-sided granite–greenstone domes, each separated from its neighbours by vertical, ring-shaped boundary faults. Each dome is composed of a central core of granitic rocks surrounded by an envelope of 'greenstones' (supracrustal volcanic and sedimentary rocks metamorphosed to greenschist facies). The greenstone envelopes of adjacent domes meet at the boundary faults, across which the supracrustal succession of the Pilbara Supergroup (see below) shows relative vertical displacements up to 10 km. Deformation fabrics along the boundary faults are dominated by extreme near-vertical stretching, whereas on the dome flanks stretching is radial (Hickman and Van Kranendonk, 2004). The vertical profiles of the domes, some of which extend to depths in excess of 14 km (Wellman, 2000), show crests of low-dipping, relatively undeformed greenstones containing discordant granitic intrusions, whereas the flanks of the domes have steeply inclined, strongly sheared granite–greenstone contacts (Hickman, 1984). Geochronology indicates that the domes formed in five stages over 500 million years, during which time they were progressively inflated and steepened. In the early stages of this process, uplift and inflation of the domes was mainly

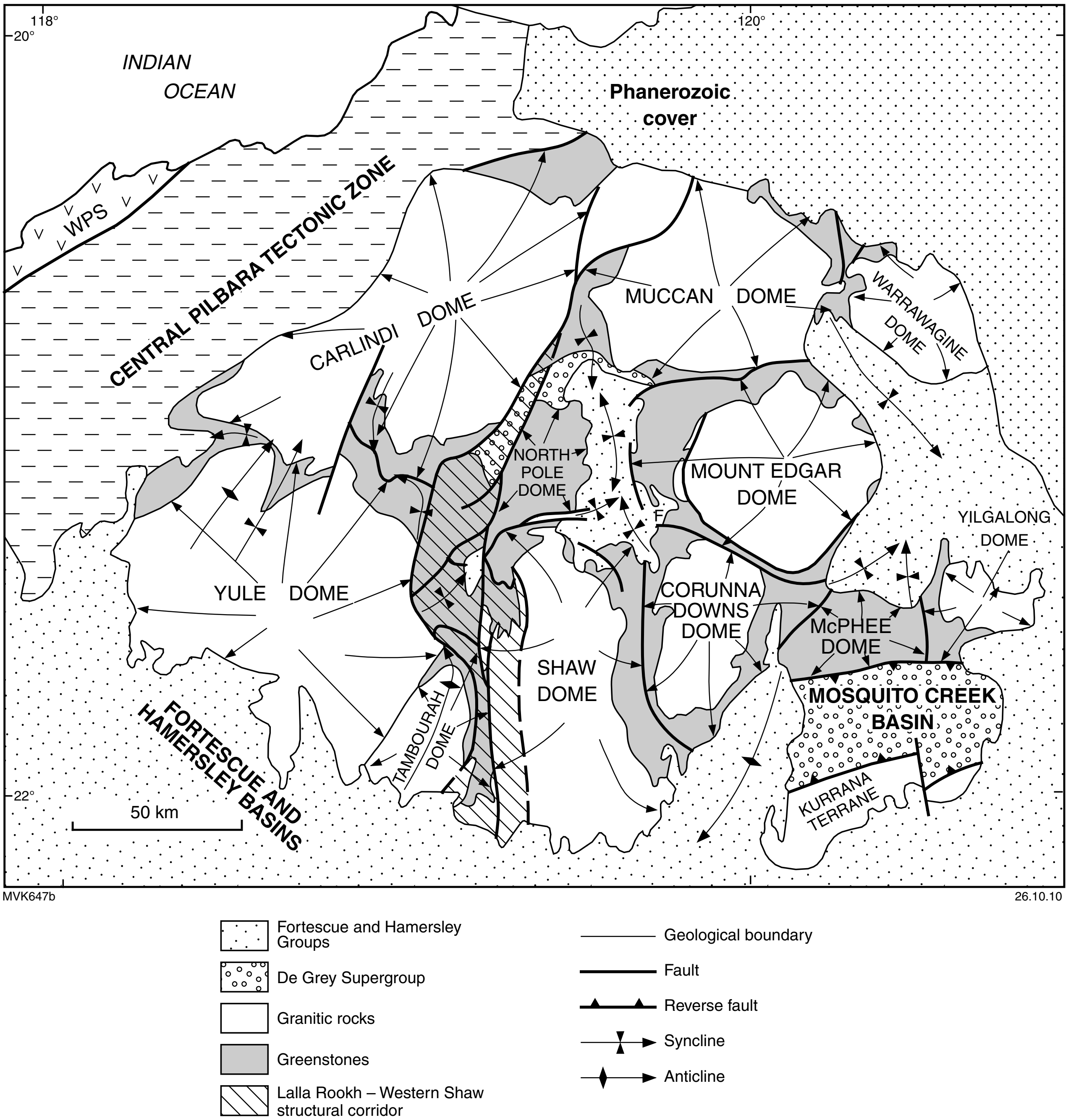


Figure 5. Principal structural elements of the East Pilbara Terrane, including structural domes cored by granitic rocks and flanking greenstones, axial faults in greenstone synclines, and the Lalla Rookh – Western Shaw structural corridor of concentrated c. 2940 Ma deformation. The Central Pilbara Tectonic Zone is another zone of concentrated c. 2940 Ma deformation across the Mallina Basin (modified from Van Kranendonk et al., 2002).

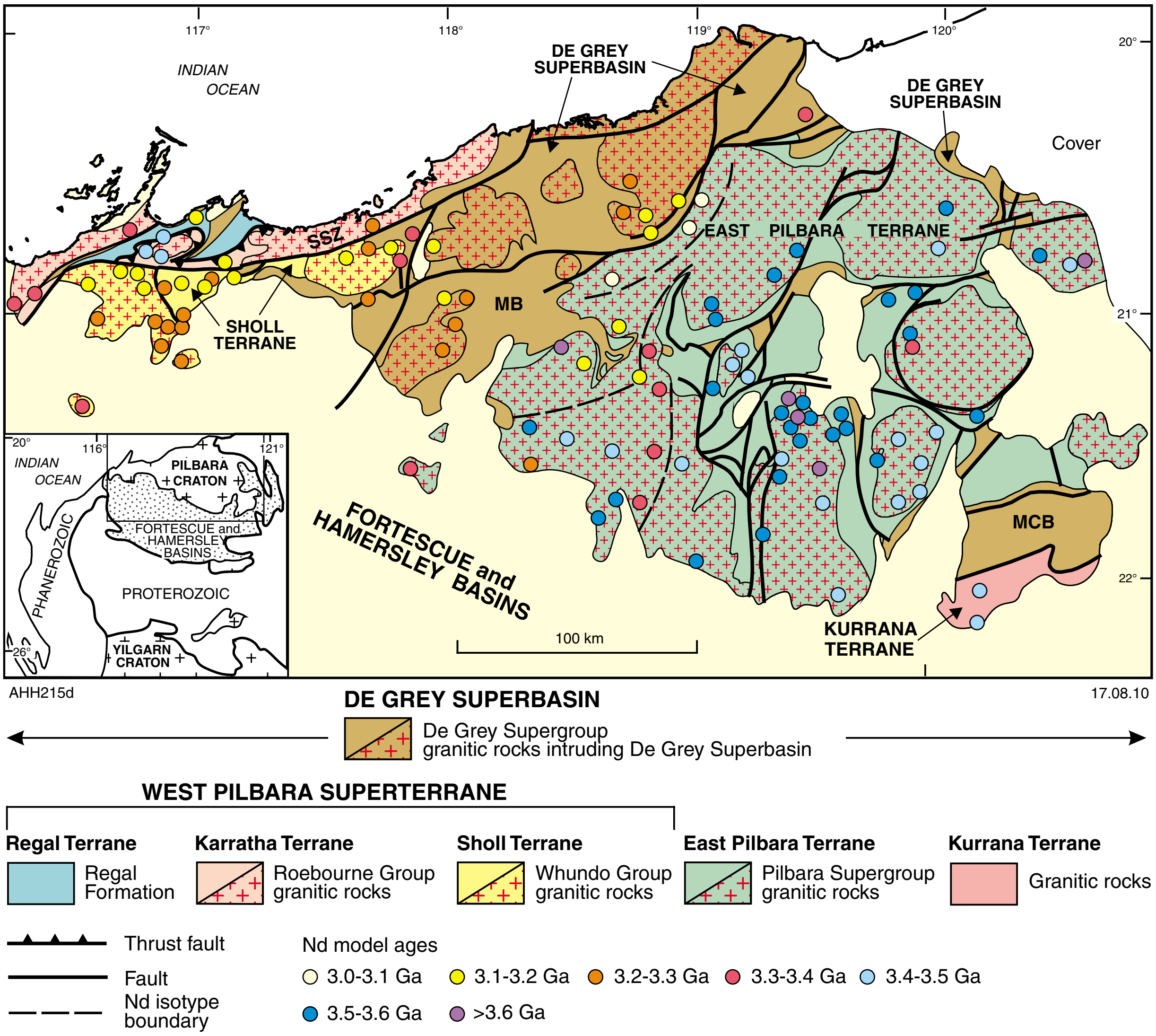


Figure 6. **Nd model age data from the Pilbara Craton, illustrating the relatively young ages (<3250 Ma) from the Sholl Terrane and Mallina Basin. Abbreviations: MB = Mallina Basin; MCB = Mosquito Creek Basin (based on data in Smithies et al., 2007a).**

by granitic intrusion (Hickman and Van Kranendonk, 2004), but, as lateral crustal heterogeneity developed with time, gravity-driven diapirism (Hickman, 1975, 1983) became more important.

The supracrustal succession of the East Pilbara Terrane is assigned to the 3525–3235 Ma Pilbara Supergroup (Table 1). The supergroup records a history of at least eight successive ultramafic–mafic–felsic cyclic volcanism events, which collectively deposited a 15–20 km thick succession. Each volcanic cycle lasted about 10–15 million years, and resulted from heating events in the mantle and lower crust (Van Kranendonk et al., 2002, 2007a,b; Hickman and Van Kranendonk, 2004; Van Kranendonk and Pirajno, 2004; Smithies et al., 2005b, 2009; Pirajno, 2007). The scale and type of volcanism is similar to that of a Large Igneous Province (LIP). In the Pilbara Supergroup, there was a c. 75 million year break in volcanism, from 3425 Ma to 3350 Ma, during which time erosion of the older uplifted and metamorphosed volcanic and granitic rocks was accompanied by deposition of shallow-water sedimentary rocks (Strelley Pool Formation). The Strelley Pool Formation separates the 3525–3427 Ma Warrawoona Group from the 3350–3315 Ma Kelly Group (Hickman, 2008). Deposition of the Kelly Group was followed by a second major break in volcanism, from 3315 to 3270 Ma. Granitic intrusion, diapiric doming, greenschist–amphibolite facies metamorphism, and erosion resulted in a second regional unconformity, overlain by conglomerate, sandstone, and volcaniclastic rocks at the base of the 3270–3235 Ma Sulphur Springs Group. The upper part of this group is composed of komatiite and komatiitic basalt, overlain by basalt, andesite, dacite, rhyolite, and chert, and is interpreted to represent the final mantle plume event of the Pilbara Supergroup.

Felsic volcanic formations of the Pilbara Supergroup are closely contemporaneous with four granitic supersuites (Van Kranendonk et al., 2006). With a few exceptions, magmatic compositions of these felsic volcanic rocks

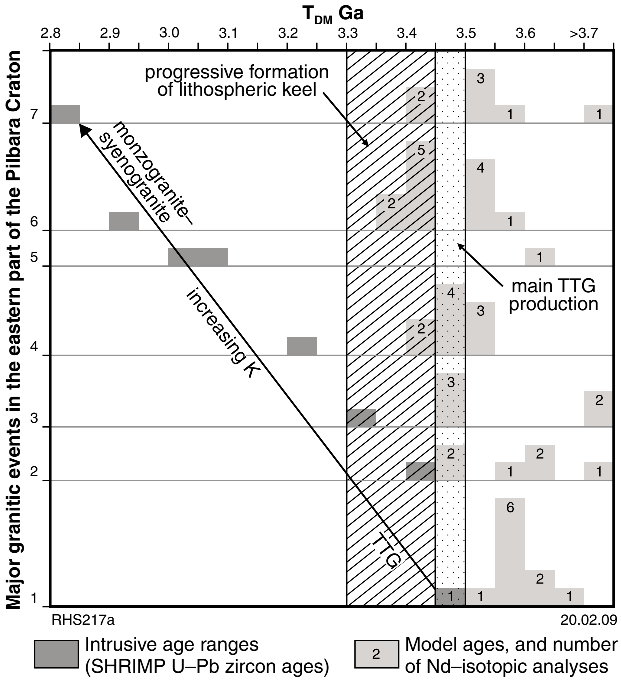


Figure 7. Variation in the depleted mantle model ages (T_{DM}) determined on granites from seven distinct felsic intrusive periods in the eastern part of the Pilbara Craton (Modified from Smithies et al., 2003). Note that event 7 includes sanukitoid plutons representing subduction along the northwestern margin of the East Pilbara Terrane (e.g. Smithies et al., 2003), and these granites, with younger model ages (< 3.45 Ga) that reflect mixed crust-mantle sources, do not belong to the East Pilbara Terrane. Stippled area shows the peak of TTG production, whereas hatch area identifies the period throughout which the thick lithospheric keel to the East Pilbara Terrane developed (Smithies et al., 2005b). After Smithies et al. (2009).

were significantly different from the granitic rocks, indicating different sources (Smithies et al., 2007b, 2009), although the same thermal events were responsible for the both volcanic rocks and granites. The youngest granitic supersuite in the East Pilbara Terrane is the 3275–3225 Ma Cleland Supersuite. Most dated intrusions of this supersuite are 3250–3230 Ma, but geochronological data indicate an earlier, c. 3275–3260 Ma suite, that includes the Karratha Granodiorite of the west Pilbara. In the East Pilbara Terrane, the thermal event responsible for the Kelly Group and the Emu Pool Supersuite ceased at c. 3290 Ma (Van Kranendonk et al., 2006), and was followed by an event of uplift, metamorphism, and erosion, prior to the deposition of the Sulphur Springs Group at c. 3250 Ma. This event probably marked the onset of the last mantle plume event, which was accompanied by intrusion of granitic rocks. By 3225 Ma, the East Pilbara Terrane was a thick and geologically complex body of continental crust.

West Pilbara Superterrane

The West Pilbara Superterrane forms the northwestern 20% of the Pilbara Craton (Fig. 3), and outcrops between Port Hedland and Cape Preston. The superterrane (Figs 2, 3, and 8) is a collage of the 3270–3250 Ma Karratha Terrane (containing the Ruth Well Formation and Cleland Supersuite, and interpreted to be a rifted fragment of the East Pilbara Terrane), the c. 3220 Ma Nickol River Formation, the c. 3200 Ma Regal Terrane (a 2 km thick basalt succession of the Regal Formation), and the 3130–3100 Ma Sholl Terrane (containing the Whundo Group volcanics and Railway Supersuite). These terranes accreted during collision with the East Pilbara Terrane at 3070 Ma. This collision, accompanied by granitic magmatism forming the Elizabeth Hill Supersuite, and various types of deformation, is referred to as the Prinsep Orogeny.

The most distinctive tectonic feature of the West Pilbara Superterrane is a northeasterly structural grain, defined by the elongation of granitic complexes and the trend of greenstone belts, and by numerous closely spaced east- and northeast-striking strike–slip faults. The Karratha and Sholl Terranes are separated by the 1–2 km wide Sholl Shear Zone that bisects the West Pilbara Superterrane over an exposed length of 250 km (Fig. 2). The Sholl Shear Zone is steeply inclined to the northwest, and is composed of mylonite and schist derived from granitic, volcanic, and intrusive rocks. The Karratha Terrane, and the overlying Regal Terrane, outcrop north of the Sholl Shear Zone, and are everywhere separated by the Regal Thrust (Fig. 8). The structural geology of the West Pilbara Superterrane, and the geochemistry of its volcanic and granitic rocks, strongly suggests that this superterrane is a product of Phanerozoic-style plate tectonic processes. This concept was first introduced by Krapež (1993), and was followed by Barley (1997), Smith et al. (1998), and Krapež and Eisenlohr (1998), although subsequent work (Van Kranendonk et al., 2002, 2006; Hickman, 2004a; Smithies et al., 2003, 2004, 2005a, 2007a) has seriously questioned many features of these early interpretations. The Whundo Group (Sholl Terrane) in particular, is exceptional amongst Mesoarchean and Paleoarchean volcanic sequences in that it preserves a range of geological and geochemical features that together provide compelling evidence of modern-style subduction processes (Smithies et al., 2005a). The group includes boninites, interlayered tholeiitic and calc-alkaline volcanics, Nb-enriched basalts, and adakites. Decreasing La/Sm and La/Yb ratios correlate with decreasing LILE, Cr, Ni, and $Mg^{\#}$ values, and with increasing values of Nb, Zr, and Yb, providing evidence for flux melting of an Archean mantle wedge. Although high Ba/La ratios reflect fluid-mediated source metasomatism, a systematic decrease in Ba/La suggests an increasing slab-melt component up the stratigraphic pile of calc-alkaline, culminating in the eruption of adakitic rocks and Nb-enriched basalts. Low Th/La (0.07 – 0.14) and Ce/Yb (<40) ratios, and a lack of evidence for a felsic basement to the Whundo Group, point to an intra-oceanic arc setting.

Karratha Terrane

The main components of the Karratha Terrane are the >3270 Ma Ruth Well Formation, and the 3270–2760 Ma Karratha Granodiorite that intrudes this formation. The Karratha Granodiorite is the same age as the oldest granitic intrusions of the Cleland Supersuite in the East Pilbara Terrane. In the Cherratta Granitic Complex,

Table 1. Generalized lithostratigraphy of the Pilbara Craton

Age (Ma)	*Terrane/Superbasin*	*Basin*	*Group*	*Formation*
c. 2930	De Grey Superbasin (De Grey Supergroup)	Mosquito Creek Basin	Nullagine Group	Mosquito Creek Formation
c. 2940		Mallina Basin	Croydon Group	Mallina Formation
				Bookingarra Formation
				Constantine Sandstone
c. 2950				Rushall Slate
c. 2960				Cistern Formation
c. 2980				Cattle Well Formation
				Coonieena Basalt
c. 3000		Whim Creek Basin	Whim Creek Group	Red Hill Volcanics
				Warambie Basalt
		Gorge Creek Basin	Gorge Creek Group	Cundaline Formation
c. 3020				Cleaverville Formation
				Farrel Quartzite
			Regional Unconformity	
c. 3118	Sholl Terrane	Whundo Basin	Whundo Group	Woodbrook Formation
c. 3120				Bradley Basalt
c. 3120				Tozer Formation
c. 3126				Nallana Formation
			Tectonic Contact —Sholl Shear Zone	
	Regal Terrane		Cleaverville succession†	Port Robinson Basalt
c. 3190				Dixon Island Formation
				Regal Formation
			Tectonic Contact — Regal Thrust	
c. 3180	Rift-related successions	Soanesville Basin	Soanesville Group	Pilbara Well volcanics†
c. 3180				Honeyeater Basalt
				Paddy Market Formation
c. 3190				Corboy Formation
				Cardinal Formation
< 3220		Budjan Creek Basin		Budjan Creek Formation
		Coondamar Basin		Coondamar Formation
			Tectonic Contact	
< 3250	Karratha Terrane		Roebourne Group	Nickol River Formation
> 3270				Ruth Well Formation
			Tectonic Contact	
c. 3235	East Pilbara Terrane (Pilbara Supergroup)		Sulphur Springs Group	Kangaroo Caves Formation
				Kunagunarinna Formation
< 3270				Leilira Formation
			Unconformity	
			Kelly Group	Charteris Basalt
c. 3320				Wyman Formation
c. 3350				Euro Basalt
			Disconformity	
c. 3420				Strelley Pool Formation
			Regional unconformity	
c. 3430			Warrawoona Group	Panorama Formation
				Apex Basalt
c. 3465				Duffer Formation
				Mount Ada Basalt
c. 3480				Dresser/McPhee Formation
> 3490				North Star Basalt
				Double Bar Formation
c. 3515				Coucal Formation
c. 3525				Table Top Formation

NOTE: † Informal stratigraphic name

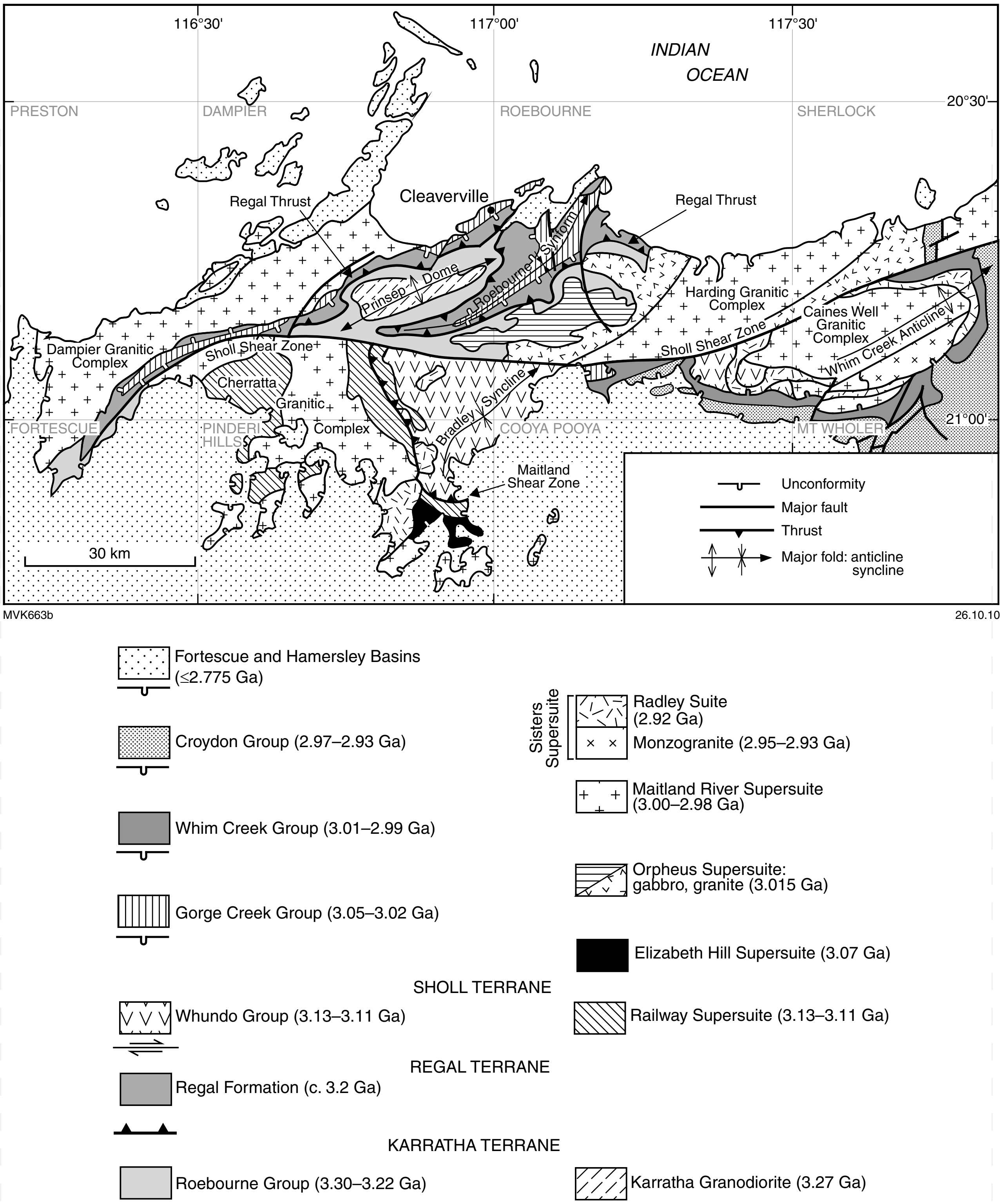


Figure 8. Simplified geological map of the northwestern Pilbara Craton, showing lithostratigraphy, tectonic units, and major structures. Note that the 3160–3070 Ma Regal Thrust is folded by the c. 2950 Ma Prinsep Dome and Roebourne Synform.

south of the Sholl Shear Zone, the 3236 Ma Tarlwa Pool Tonalite is the same age as younger granites of the Cleland Supersuite in the East Pilbara Terrane, and is therefore also included in this supersuite (Van Kranendonk et al., 2006). The Ruth Well Formation, which is almost entirely composed of ultramafic–mafic volcanic rocks, is overlain by the sedimentary Nickol River Formation (Hickman, 1997a). Based on limited geochronology, the Nickol River Formation was previously included, along with the Ruth Well Formation, in the Roebourne Group (Hickman, 1997a), but, as explained below (see '**Nickol River Formation**'), an alternative interpretation is possible.

The Ruth Well Formation is up to 2000 m thick, and consists of metamorphosed ultramafic and mafic volcanic rocks (peridotite to tholeiitic basalt), with a few thin chert intercalations. No stratigraphic base is preserved, lower contacts being either tectonic or intrusive. As it is intruded by the Karratha Granodiorite, the formation must be older than 3270 Ma. Its maximum depositional age is currently very poorly constrained, but likely less than c. 3450 Ma, based on depleted mantle extraction (Nd T_{DM} model) ages (Arndt et al., 2001; Smithies et al., 2007a).

The c. 3270 Ma Karratha Granodiorite ranges from tonalite to granodiorite (GSWA 142433: Nelson, 1998f; Smith et al., 1998). Much older Nd T_{DM} model ages of 3480–3430 Ma, derived from this intrusion (Sun and Hickman, 1998) indicate that magma generation involved older crust or enriched lithospheric mantle, as for the Cleland Supersuite in the East Pilbara Terrane (data in Smithies et al., 2007a).

Nickol River Formation

From its type area around Roebourne and the Nickol River, the Nickol River Formation outcrops almost continuously for 100 km southwestwards, where it extends beneath the Fortescue Group within 25 km of the Fortescue River. The depositional stratigraphy of the formation is dislocated by numerous low-angle thrusts within a regional-scale zone of tectonic transport, referred to as the Regal Thrust (Hickman, 2001b). The formation is well exposed south of Mount Regal, where basal quartzite and chert are overlain by metamorphosed ferruginous shale, banded iron-formation (BIF), and carbonate rocks. The sedimentary rocks are extensively sheared, and mesoscopic recumbent folds are visible in outcrop. Near the Lower Nickol mining area, 20 km east of Mount Regal, Kiyokawa et al. (2002) referred to the same formation as the 'Lydia Mine complex' (informal name), and reached the same conclusion as Hickman (2001b) that it underlies a major thrust. They interpreted the formation to have originated as a continental shelf deposit unconformably overlying the Karratha Granodiorite, and recorded the presence of metamorphosed pebbly sandstone in the formation, in agreement with Hickman and Smithies (2001). Metasandstones of the formation contain rounded clasts of black chert and altered volcanic rocks most likely derived from erosion of the Ruth Well Formation (see **Excursion Locality 14**). Black chert conglomerate mapped on the northwestern coast at Cleaverville (Kiyokawa and Taira, 1998) underlies pillow basalts of the Regal Formation, suggesting that the Nickol River Formation may also be present on the northwest limb of the Cleaverville Synform (Fig. 9).

In the Lower Nickol area, schistose metasandstone of the formation was dated at 3269 ± 2 Ma (GSWA 136819: Nelson, 1998b), a date consistent with the derivation of detrital zircon by erosion of the underlying Karratha Granodiorite. Southeast of Mount Regal, metamorphosed carbonate rocks and ferruginous chert found locally at the base of the formation (in tectonic contact with the underlying Ruth Well Formation) contain metamorphosed porphyritic rhyolite (probably a deformed intrusion). Using the SHRIMP U–Pb zircon method, the crystallization age of this rock was interpreted as 3251 ± 6 Ma (GSWA 118975: Nelson, 1997c), with older zircons, of similar age to the Karratha Granodiorite, interpreted to represent a xenocrystic population. Nelson (1997c) interpreted concordant and slightly discordant zircons younger than the 3251 Ma population as having lost radiogenic Pb, but an alternative interpretation is that the 3251 Ma population is xenocrystic, and the crystallization age is closer to that of the youngest concordant zircon dated at c. 3217 Ma. Deposition of the Ruth Well and the Nickol River Formations was probably separated by at least 20 million years, and possibly more than 50 million years. During that time, the Karratha Granodiorite was intruded into the Ruth Well Formation and then eroded, suggesting that the Ruth Well Formation and Nickol River Formation should not be assigned to the same group (Roebourne Group), although direct dating of the Ruth Well Formation is needed to resolve this question. The Nickol River Formation could be an analogue of the c. 3200 Ma Soanesville Group, a clastic succession that unconformably overlies the northwestern margin of the East Pilbara Terrane (Van Kranendonk et al., 2010) on the opposite side of the 3200 Ma basin of oceanic crust.

Regal Terrane and the Regal Thrust

The Regal Terrane is composed of the Regal Formation, a 2–3 km thick sequence of metamorphosed pillow basalt, local basal komatiitic peridotite, and rare chert. Sills of dolerite and gabbro intrude the formation. In all areas except Cleaverville, the formation has been metamorphosed to amphibolite or upper greenschist facies. Lower grade metamorphism in the succession at Cleaverville is attributed to separation from the main outcrop area by large faults, resulting in only the topmost part of the formation being exposed. Pillow basalt units at Cleaverville, here also assigned to the Regal Formation, were separated as different, informally named formations by Kiyokawa and Taira (1998).

The Regal Terrane is separated from the underlying Karratha Terrane by the Regal Thrust zone, which is mainly composed of tectonically deformed slices of the Nickol River Formation. Nd isotopic data (Smithies et al., 2007a) indicate an approximate depositional age of 3200 Ma for the Regal Formation (Van Kranendonk et al., 2007a), in agreement with a U–Pb zircon age of 3195 Ma obtained from a thin felsic tuff unit in the Dixon Island Formation (Kiyokawa et al., 2002). The Dixon Island Formation has been recognized only on Dixon

Island and in isolated exposures on Cleaverville Beach, where it underlies pillow basalts of the Regal Formation. The minimum age of the Regal Formation is c. 3160 Ma, based on the age of a thermal event that metamorphosed it, the Karratha Granodiotite, and the Ruth Well Formation (Kiyokawa, 1993; Smith et al., 1998; Beintema, 2003).

The lithological composition (pillow basalt and rare thin chert units) and geochemistry (MORB-like) of the Regal Formation are suggestive of oceanic crust (Ohta et al., 1996; Sun and Hickman, 1998; Kiyokawa and Taira, 1998; Beintema, 2003), although the absence of numerous, oriented dykes casts doubt on a mid-ocean ridge origin. Interpretation of the Regal Formation as a slice of oceanic crust (ophiolite) is consistent with its flat REE patterns and ε_{Nd} of ~+3.5, which is close to the depleted mantle value (3.2) at 3200 Ma (Smithies et al., 2007a). The formation's basalts show no geochemical evidence of crustal contamination (Smithies et al., 2007a).

Across 3000 km^2 of the northwest Pilbara, the Regal Formation is sandwiched between the underlying continental crust of the Karratha Terrane, and the overlying Cleaverville Formation. Although this is not the case at Cleaverville, the Cleaverville Formation overlies continental crust elsewhere in the Pilbara Craton, and so, despite being composed of banded iron-formation, chert, and generally fine-grained clastic sedimentary rocks, it cannot be considered a deep-water oceanic deposit. As described below, the Cleaverville Formation is a shallow-water deposit accumulated in an extensional continental basin. The Regal Thrust forms the contact between the Regal Formation and the Karratha Terrane (Hickman, 2001b; Hickman and Smithies, 2001), and this major thrust zone extends over the entire west Pilbara north of the Sholl Shear Zone (Figs 8 and 9). The uppermost thrust plane immediately underlies tectonically stretched and flattened basaltic rocks of the Regal Formation, and includes several mylonite units several metres thick. Below this level, there is a zone of tectonic lensing up to 1 km thick, in which the Nickol River Formation is broken into numerous tectonic slices, bounded by mylonites. These mylonites contain refolded isoclinal folds, the orientation of which led Hickman (2001b, 2004a) to suggest that transport (obduction of oceanic crust) of the Regal Terrane was from the north or northwest. However, Kiyokawa et al. (2002) and Van Kranendonk et al. (2006) instead suggested that the Regal Formation could have been thrust across the Karratha Terrane from the south or southeast; oceanic crust is interpreted to have existed in that direction after 3200 Ma. Establishing the direction of tectonic transport of the Regal Thrust is complicated by the effects of several later phases of deformation (described below), which added shearing and folding to the thrust-related fabrics.

The age of the Regal Thrust must be between 3160 and 3070 Ma. If thrusting occurred at c. 3160–3150 Ma (Hickman, 2001b, 2004a; Hickman et al., 2001; Beintema, 2003), it is unlikely to have been from the south or southeast, because at that time the southeastern area was in an extensional setting. However, if the thrusting occurred during the Prinsep Orogeny at 3070 Ma, it would have been related to the plate collision that formed the West Pilbara Superterrane. Kiyokawa et al. (2002) suggested that thrusting could have occurred as late as 3020 Ma, but that is improbable because it would also have involved the 3130–3110 Ma Whundo Group, which was in tectonic contact with the Karratha Terrane from 3070 Ma.

After c. 3010 Ma, and most probably at c. 2950 Ma, the Regal Thrust was folded by major upright folds (D_6, see Deformation events), so that the thrust and the overlying Regal Formation outcrop on three fold limbs (Figs 8 and 9). Way-up evidence from pillow structures in the Regal Formation confirms the continuity of the structural succession over a strike length of at least 100 km (Fig. 9).

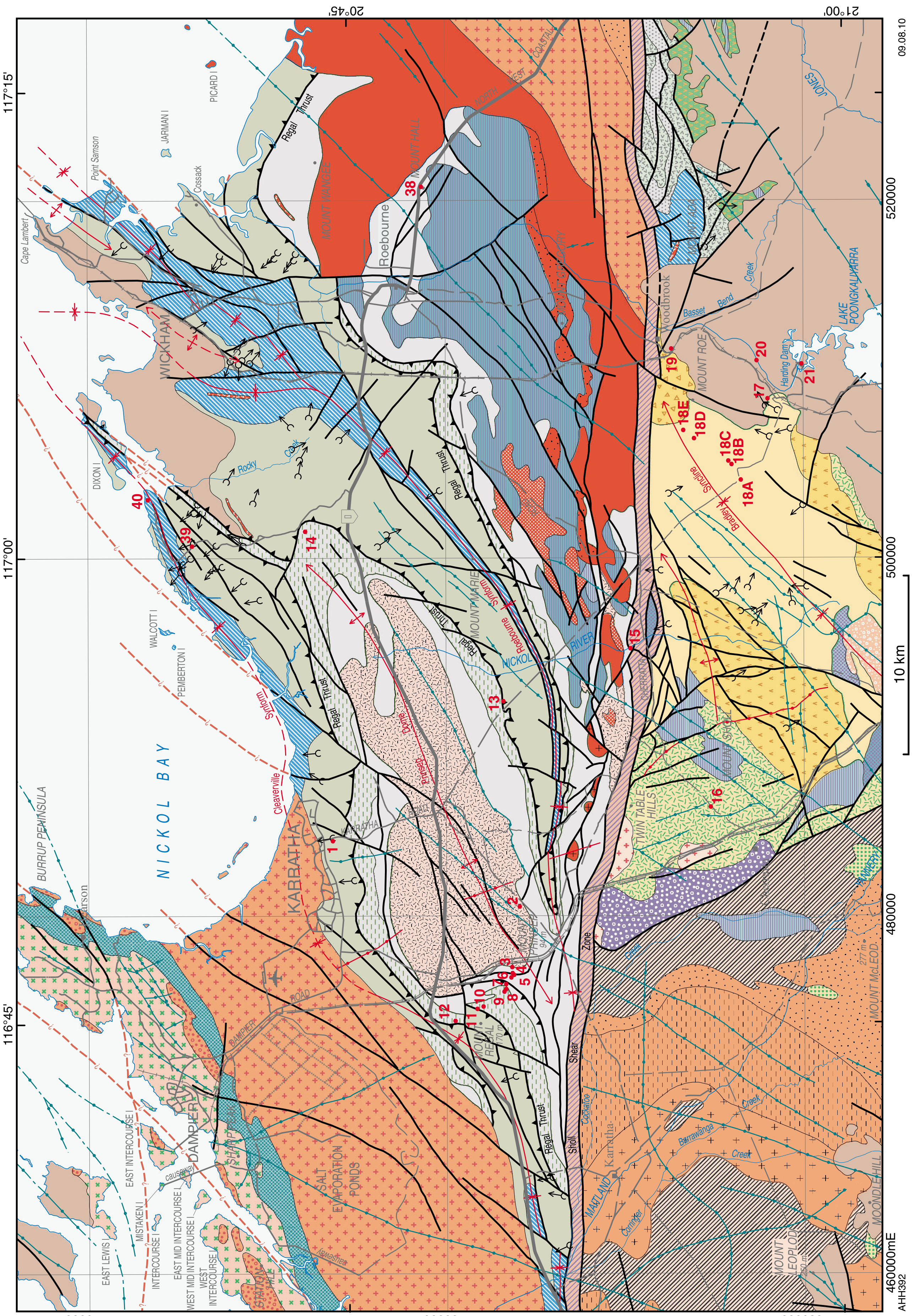
Sholl Terrane

The Sholl Terrane comprises the 3130–3110 Ma Whundo Group and the 3130–3110 Ma Railway Supersuite. Most Railway Supersuite outcrops are within the Cherratta Granitic Complex, but mafic and felsic intrusions of the supersuite also intrude the lower stratigraphic levels of the Whundo Group. The Sholl Terrane was intruded by granitic rocks of the Elizabeth Hill Supersuite during its collision with the Karratha Terrane at 3070 Ma (Prinsep Orogeny).

The Whundo Group is divided into four lithostratigraphic formations (Table 1). No stratigraphic base is exposed, but the group is interpreted to have been deposited above c. 3200 Ma oceanic crust. The exposed basal contact is the Maitland Shear Zone, which separates the Whundo Group from the Railway Supersuite of the Cherratta Granitic Complex, and which was a low-angle thrust prior to folding at c. 2950 Ma. The northern contact of the group is defined by the Sholl Shear Zone (Figs 7 and 9), and the group is unconformably overlain by the Gorge Creek, Whim Creek, and Fortescue Groups.

The lowest formation preserved is the Nallana Formation, which is >2.0 km thick and comprises metabasalt, ultramafic rocks, intermediate pyroclastic rocks, and dolerite sills. The conformably overlying Tozer Formation is 2.5 km thick, and contains basalt, andesite, dacite, rhyolite, and thin metasedimentary units. The conformably overlying Bradley Basalt is more than 4.0 km thick, and consists of metamorphosed, and massive and pillowed basalt, with komatiitic basalt near its base. The Bradley Basalt is conformably overlain by felsic volcanic rocks of the Woodbrook Formation, which comprises metamorphosed rhyolite tuff and agglomerate, and only minor metabasalt. A detailed chemostratigraphic analysis of the Whundo Group (Smithies et al., 2005a) revealed that it developed from a more complex set of igneous processes than can be identified from lithological mapping (Fig. 10). A lower volcanic package (Nallana and lower Tozer Formations) contains calc-alkaline and boninite-like rocks; the middle package (upper Tozer Formation, and lower Bradley Basalt) consists of tholeiitic and minor boninite-like rocks, and rhyolites; and the upper package (upper Bradley Basalt and Woodbrook Formation) contains calc-alkaline rocks, including adakites, Mg-rich basalts, Nb-enriched basalts, and rhyolites.

Geochronological data from all stratigraphic levels of the group provide a well constrained depositional age range between 3126 and 3112 Ma (Hickman et al., 2006).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BURRUP PENINSULA
KARRATHA
Roebourne
WICKHAM
DAMPIER
Cossack
Point Samson
Cape Lambert
JARMAN I
PICARD I
DIXON I
WALCOTT I
PEMBERTON I
EAST LEWIS I
MISTAKEN I
EAST INTERCOURSE I
MOUNT WANGEE
MOUNT HALL
MOUNT MARIE
MOUNT ROE
MOUNT REGAL
MOUNT PRINSEP
MOUNT SHOLL
MOUNT McLEOD
MOUNT LEOPOLD
MOONDEN HILL
TWIN TABLE HILLS
SALT EVAPORATION PONDS
LAKE POONGKALIYARRA
Regal Thrust
Sholl Shear Zone
Roebourne Synform
Cleaverville Synform
Bradley Syncline
Prinsep Dome
NICKOL RIVER
MAITLAND
Karratha
Woodbrook
Harding Dam
Basset Bend Creek
Rocky Creek
Cockatoo Creek
Barrawanga Creek
Cunnigan
NORTH WEST COASTAL
BAMPIER ROAD
causeway
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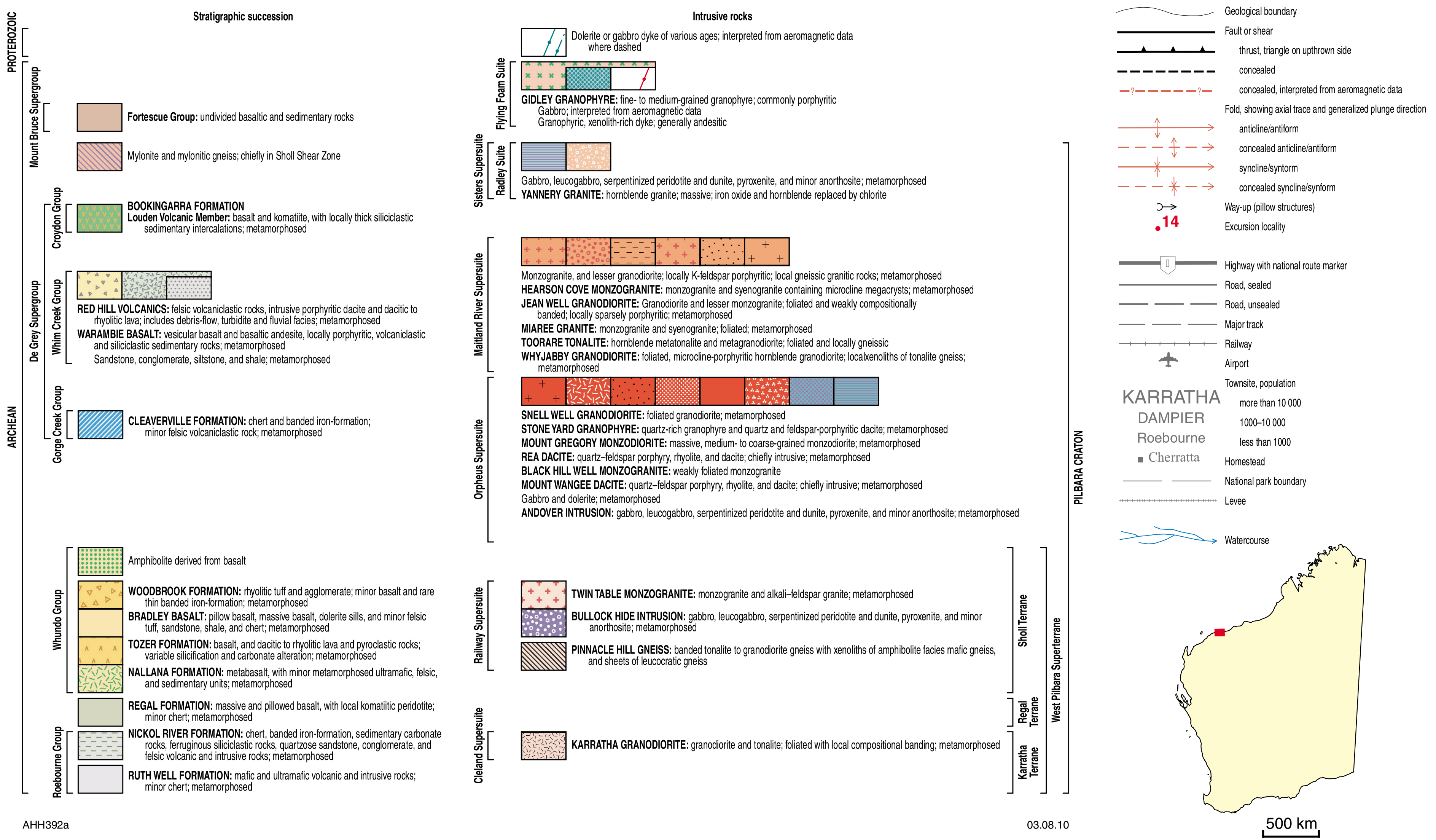

Figure 9. Geological map of the Roebourne–Karratha area, showing major structures and way-up evidence (from pillow structures) in the Regal Formation.

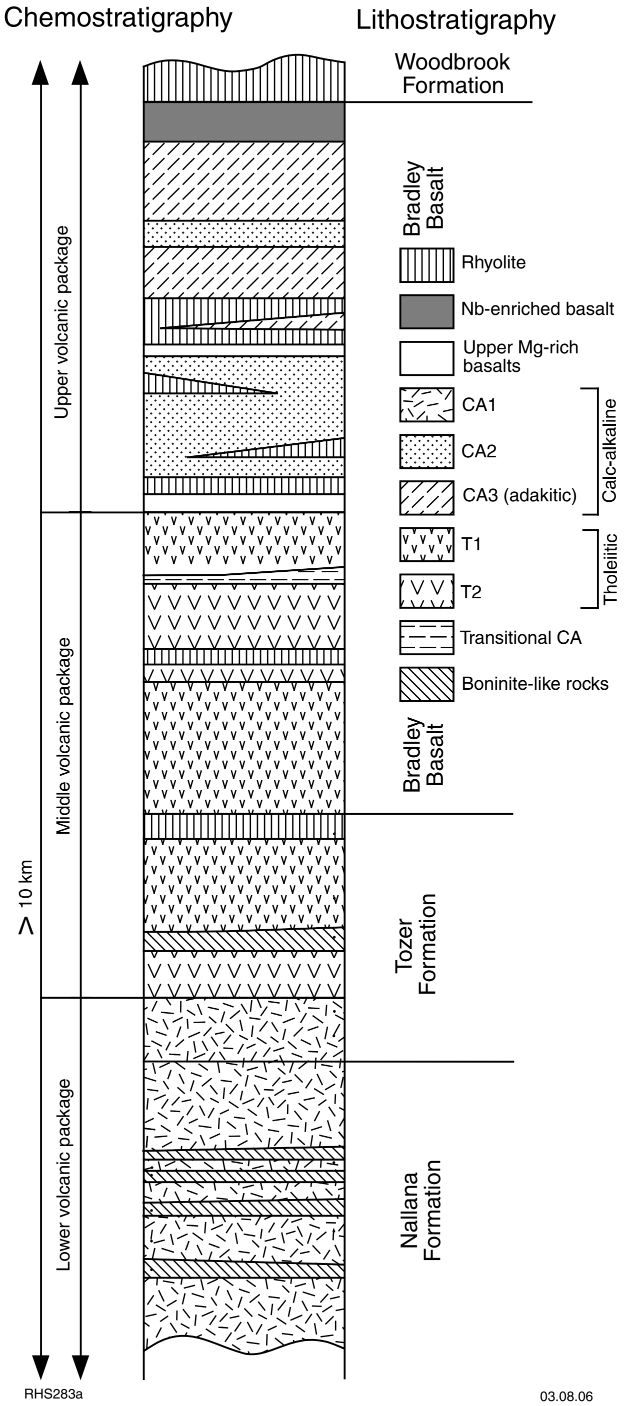

Figure 10. Chemostratigraphic column of the Whundo Group, with lithostratigraphic formation boundaries shown for comparison (after Smithies et al., 2005a).

Nd-isotopic compositions show that the Whundo Group represents a juvenile crustal addition, with values that lie very close to depleted mantle values (ε_{Nd}~+1.5-2.0 at c. 3120 Ma; Sun and Hickman, 1998; Smith et al., 1998; Smithies et al., 2005a).

Smithies et al. (2005a) outlined numerous features of the Whundo Group that can be interpreted in terms of modern-style plate tectonic process. Although many of these features could be individually interpreted in terms of processes not necessarily related to modern-style plate tectonics or to subduction, together they provide a compelling case for a convergent margin setting at 3120 Ma. These features include:

- the fine- and broad-scale intercalation of discrete tholeiitic and calc-alkaline volcanics (Fig. 10)
- Nd isotopic compositions that lie very close to depleted mantle values
- the presence of rocks with a strong boninite affinity near the base of the sequence (Fig. 11)
- the presence of Nb-enriched basalts near the top of the sequence, and the close association of these with lavas of adakitic affinity
- the recognition that assimilation of crust is an unlikely cause of LREE-enrichments in the boninite-like lavas and calc-alkaline lavas
- the identification of a range of source regions, from depleted to undepleted and enriched
- the mixing of source regions and/or primitive magmas derived from undepleted, depleted, calc-alkaline, and tholeiitic (MORB) sources, with distinct periods characterized by enhanced interaction and magmatic diversity
- the recognition that the calc-alkaline lavas are the result of flux melting
- the absence of any intervening sequences containing exotic 'continental' material
- the absence of evidence for continental (felsic) basement material, and the possibility that the sequence was developed over arc-like mid-level crust of similar age

Rocks of the Whundo Group provide the best direct evidence for the operation of modern-style subduction process before ~3000 Ma. Krapež and Eisenlohr (1998), Smith et al. (1998), and Smith (2003) all suggested a back-arc setting for these rocks; however, the lack of evidence for felsic basement, the presence of boninites, and low Th/La, La/Nb, and Ce/Yb ratios is more likely indicative of an intra-oceanic arc setting (Smithies et al., 2005a). Van Kranendonk et al. (2006, 2007a, b) suggested that the arc lay adjacent to a subduction zone within the post-3200 Ma basaltic basin between the East Pilbara Terrane and the Karratha Terrane, revealing a change from spreading to plate convergence at about 3130 Ma.

The other component of the Sholl Terrane, the Railway Supersuite, is composed of metatonalite, metagranodiorite, and metamonzogranite gneiss dated between 3130 and 3114 Ma (Van Kranendonk et al., 2006). The oldest zircons dated from this supersuite are 3149 ± 15 Ma, with Nd T_{DM} model ages of 3230–3210 Ma (Smithies et al., 2007a).

Prinsep Orogeny and Elizabeth Hill Supersuite

Plate convergence, signalled by the eruption of the 3130–3110 Ma Whundo Group and intrusion of the Railway Supersuite, eventually led to accretion of three terranes (Karratha, Regal, and Sholl) to form the West Pilbara

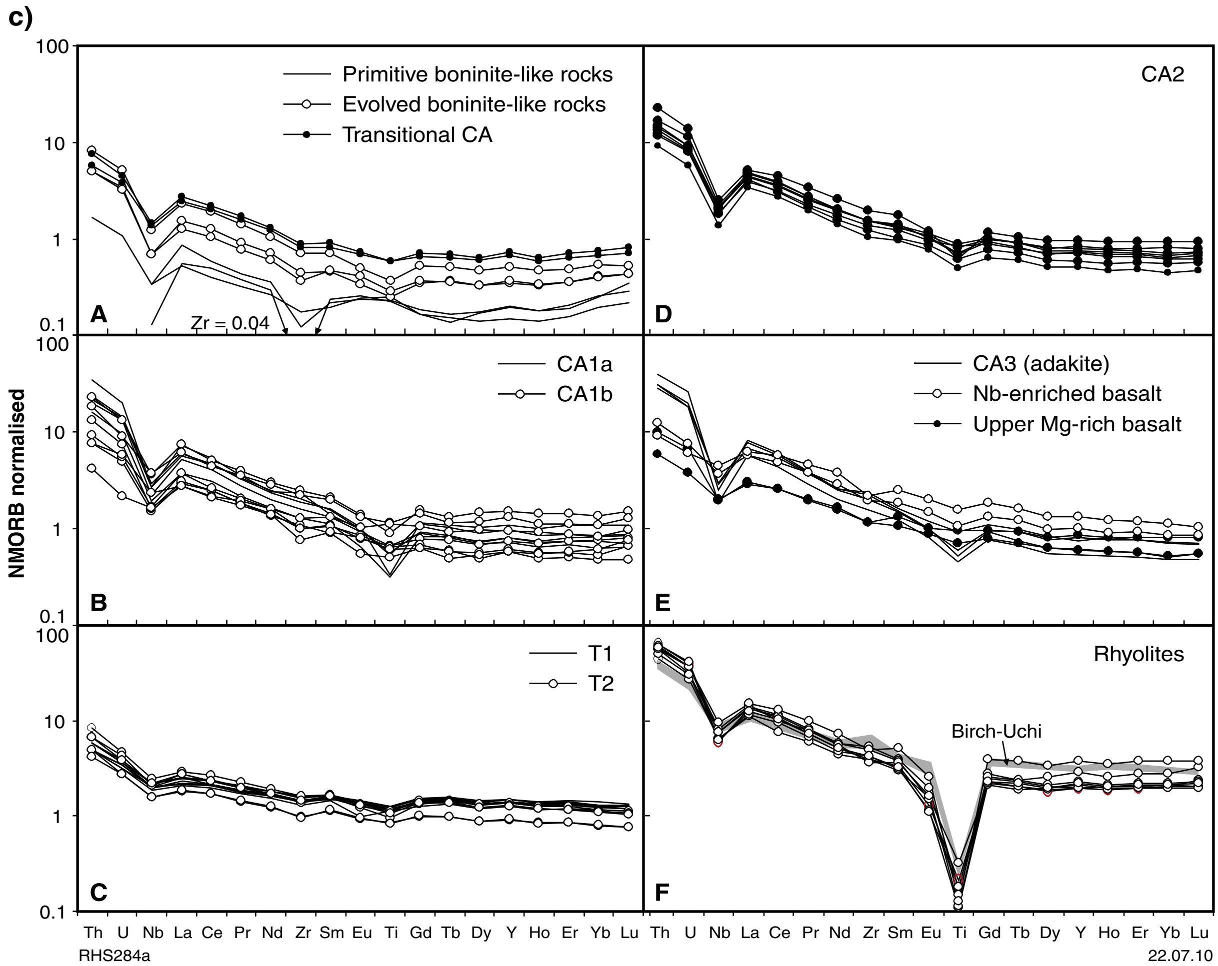

Figure 11. N-MORB normalized trace element diagrams for the various rock types of the Whundo Group

Superterrane at 3070 Ma (Van Kranendonk et al., 2006). At the same time, the West Pilbara Superterrane collided with the East Pilbara Terrane, accompanied by extensive intrusion of 3070 Ma granites (Elizabeth Hill Supersuite) along the northwestern margin of the East Pilbara Terrane. This northwest–southeast collision resulted in the Prinsep Orogeny, which included low-angle thrusting on the Regal Thrust and Maitland Shear Zone, and major sinistral strike–slip movement (at least 200 km) on the Sholl Shear Zone.

The Elizabeth Hill Supersuite intrudes both the West Pilbara Superterrane and the western part of the East Pilbara Terrane, and is not part of either of these major tectonic units. In the west Pilbara, the Elizabeth Hill Supersuite is represented by the Cliff Pool Tonalite, which outcrops over 70 km^2 as part of the Cherratta Granitic Complex east of the Munni Munni Intrusion, and possibly also by gneiss in the Caines Well Granitic Complex. A sample of the Cliff Pool Tonalite (GSWA 142661), collected approximately 6 km south-southeast of the Elizabeth Hill mine, was dated by Nelson (1998g) at 3068 ± 4 Ma. The Cliff Pool Tonalite is a foliated biotite tonalite, which includes xenocrystic zircons of Railway Supersuite age (Nelson, 1998g), supporting the field observation of an intrusive relationship to the c. 3130 Ma Pinnacle Hill Gneiss. Its northern contact with this gneiss is diffuse and interlayered, with sheets of the tonalite intruding rafts of the gneiss. The southern contact of the tonalite is intruded by granitic rocks of the 3000–2980 Ma Maitland River Supersuite.

Kurrana Terrane

Most of the Kurrana Terrane is concealed by the unconformably overlying volcanic and sedimentary formations of the Fortescue and Hamersley Groups. Immediately south of the Mosquito Creek Basin (Figs 1 and 2), the terrane is mainly composed of c. 3180 Ma gneissic metagranodiorite of the Mount Billroth Supersuite, and of virtually unfoliated syenogranite and monzogranite of the 2890–2830 Ma Split Rock Supersuite. This northern exposure of the terrane contains no greenstone belts but the orthogneiss contains numerous amphibolite enclaves. The boundary between the Mosquito Creek Basin and the Kurrana Terrane is the Kurrana Shear Zone, which is intruded by ultramafic and mafic sills assigned to the 3.18 Ga Dalton Suite (Van Kranendonk et al., 2006). However, most deformation on the shear zone is interpreted to be part of the c. 2900 Ma Mosquito Creek

Orogeny, in which the Kurrana Terrane was thrust against the East Pilbara Terrane.

Old xenocrystic zircons in granitic rocks of the northern Kurrana Terrane range from 3580 to 3460 Ma in age, with Nd T_{DM} model ages from the terrane ranging between 3450 and 3410 Ma (Van Kranendonk et al., 2007a). These dates are consistent with the Kurrana Terrane having originated as a rifted segment of the East Pilbara Terrane. Hickman (2004a) suggested from geophysical data that the Sylvania Inlier, 150 km to the south of the Mosquito Creek Basin, was the southern part of the Kurrana Terrane (Fig. 2), but there are insufficient geochronological data from this inlier to test the interpretation.

De Grey Superbasin

The De Grey Superbasin (Fig. 1) unconformably overlies the East Pilbara Terrane and the West Pilbara Superterrane, and is composed of four successions within four basins: the 3050–3015 Ma Gorge Creek Group, 3010–2990 Ma Whim Creek Group, 2970–2940 Ma Croydon Group, and 2970–2930 Ma Mosquito Creek Formation (Table 1).

Regional unconformity

In the east Pilbara, the base of the Gorge Creek Group, the De Grey Supergroup's oldest group, is an angular unconformity developed above various formations of the Pilbara Supergroup, the Soanesville Group, and also deeply eroded Paleoarchean granitic rocks (Dawes et al., 1995). By contrast, in the west Pilbara, the Gorge Creek Group nonconformably overlies the Regal Formation north of the Sholl Shear Zone, and the Whundo Group south of this zone; the unconformity is not visibly angular. There are two probable explanations for this regional difference. Firstly, the rocks underlying the east Pilbara unconformity are up to 500 million years older than the Gorge Creek Group, with a much longer history of deformation and erosion than the equivalent west Pilbara rocks. Secondly, pre-3050 Ma deformation in the east Pilbara was dominated by doming that produced vertically inclined bedding and tectonic foliation, whereas pre-3050 Ma deformation in the west Pilbara (Prinsep Orogeny) consisted of siubhorizontal thrusting and associated recumbent folding of the crust; at 3050 Ma, the surface geology of the west Pilbara, on which the Gorge Creek Group was deposited, would therefore have consisted of relatively flat-lying metabasalts and metadolerite. As a result, the existence of the unconformity in the west Pilbara is less obvious, but can be established by three lines of evidence: the age difference between the c. 3200 Ma Regal Formation and the c. 3020 Ma Cleaverville Formation (the basal formation of the Gorge Creek Group in the west Pilbara); differences in metamorphic grade between the rocks above and below the unconformity (Hickman and Smithies, 2001); and the very different depositional environments of the rocks above and below the unconformity (deep-water oceanic for most of the Regal Formation, in contrast to shallow-marine and continental for the Cleaverville Formation — see below).

Gorge Creek Basin

The Gorge Creek Basin is composed entirely of the 3050–3020 Ma Gorge Creek Group. In the east Pilbara, this group consists of three formations: a basal unit of metamorphosed sandstone and conglomerate called the Farrel Quartzite (Van Kranendonk et al., 2006); a central unit of chert, banded iron-formation (BIF), carbonaceous shale, and local siltstone and sandstone, named the Cleaverville Formation; and an upper unit of conglomerate, sandstone, and shale, named the Cundaline Formation. In the west Pilbara, only the Cleaverville Formation[1] is recognized.

In the east Pilbara, the Farrel Quartzite is composed predominantly of fluviatile polymictic conglomerate and sandstone, with local siltstone, shale, and chert. The thickness of the formation varies from a few metres above granitic basement, to in excess of 1000 m over greenstones. This difference indicates basement topography variations in the East Pilbara Terrane, most likely due to pre-3050 Ma subsidence of the underlying greenstones relative to the granitic domes, and different rates of erosion for the granites and the greenstones. The Cundaline Formation, which is recognized only in the northeast Pilbara, disconformably overlies the Cleaverville Formation and includes local basal conglomerate and breccia derived from this latter unit (Williams, 1999). In the west Pilbara, the Farrel Quartzite may be represented by lenticular units of sandstone and conglomerate that locally underlie the Cleaverville Formation and unconformably overlie the Regal Formation, such as the informally named 'Sixty-six Hills Member' at Cleaverville (Kiyokawa and Taira, 1998). However, in the west Pilbara, the Gorge Creek Basin overlies basaltic crust (Regal Formation and the Whundo Group) that would have contributed far less quartzofeldspathic sand than rocks of the East Pilbara Terrane. Underlying basaltic crust in the west Pilbara may also account for the low Al_2O_3/TiO_2 values reported from chert units of the Cleaverville Formation at Point Samson (Sugitani et al., 1996). The upper part of the Cleaverville Formation in the west Pilbara includes sandstone and shale of similar facies to the Cundaline Formation.

The Gorge Creek Basin overlies an erosion surface that developed across the Regal, Sholl, and East Pilbara Terranes after the Prinsep Orogeny at 3070 Ma. Deposition was initially shallow-water, and included evaporitic and fluviatile deposits (Sugitani et al., 1996, 1998, 2003). The thickness of the group (up to 2 km) testifies to the amount of basin subsidence, although lateral facies changes in the east Pilbara suggest that the basin was broken by islands, and local erosion of the Cleaverville Formation during deposition of the Cundaline Formation suggests localized areas of uplift. In the east Pilbara, U–Pb dating of detrital zircons from the Farrel Quartzite (GSWA 143995: Nelson, 1998i) indicates detritus derived only from the East Pilbara Terrane. By contrast, in the west Pilbara,

[1] The name Cleaverville Formation is applied to a formally defined stratigraphic unit (Ryan and Kriewaldt, 1964; Hickman, 1983), and is not equivalent to any of the informally named 'Cleaverville succession', 'Cleaverville unit', or 'Cleaverville group' (Kiyokawa, 1993; Kiyokawa and Taira, 1998; Kiyokawa et al., 2002), which include rocks 180 million years older than the c. 3020 Ma Cleaverville Formation.

where the basin was developed on terranes with a younger evolutionary history, the ages of detrital zircon populations in the Cleaverville Formation and immediately underlying sandstones range between 3270 Ma and 3016 Ma, with c. 3016 Ma zircons probably derived from volcanic ash (Hickman et al., 2006).

Whim Creek Basin

Although the ages of the Gorge Creek Group (3050–3020 Ma) and the Whim Creek Group (3010–2990 Ma) are relatively close, the two successions are separated by a high-angle unconformity and a major change in metamorphic grade (Hickman, 2002). The Whim Creek Basin is located above, and immediately to the southeast of, the Sholl Shear Zone, which is the 3070 Ma suture between the Karratha and Sholl Terranes (Smith et al., 1998; Hickman, 2001b, 2004a). Between 3020 and 3010 Ma, the Sholl Shear Zone was reactivated by a north–south convergence, resulting in transpressional, tight to isoclinal folding of the Cleaverville and Regal Formations (D_3; Hickman and Smithies, 2001; Hickman, 2004a), and upper greenschist to lower amphibolite facies metamorphism in adjacent parts of the Whundo Group. Movement along the Sholl Shear Zone was sinistral, combined with relative uplift of the northern block. Evidence of this uplift is provided by local fans of conglomerate and sandstone at the base of the Whim Creek Group immediately south of the shear zone (Hickman and Smithies, 2001).

The 3010–2990 Ma Whim Creek Basin is entirely composed of the Whim Creek Group, which comprises volcanic, intrusive, and volcaniclastic rocks of the Warambie Basalt and Red Hill Volcanics (Fig. 12). The group is unconformably overlain by 2970–2930 Ma sedimentary and volcanic formations of the Croydon Group. Together, the Whim Creek and Croydon Groups form the supracrustal succession that makes up the east-northeast trending, 150 km long Whim Creek greenstone belt, located between Roebourne and Port Hedland. The boundary between this greenstone belt and the Mallina Basin is defined by three faults:,by the western section of the Mallina Shear Zone in the south, by the Loudens Fault in the southeast, and by the eastern section of the Sholl Shear Zone in the northeast (Fig. 13). Some early workers interpreted the depositional setting of the Whim Creek Group (when it was defined as including younger formations later assigned to the Croydon Group) as a pull-apart basin (Barley, 1987; Krapež, 1993). An alternative view, based on a detailed examination of its volcanic and sedimentary facies, was that of an ensialic back-arc basin (Pike and Cas, 2002). Smith (2003) suggested that granites of the 3000–2980 Ma Maitland River Supersuite, which form a linear zone that parallels the Sholl Shear Zone northwest of the Whim Creek greenstone belt, reflect the roots of a continental arc, and Smithies et al. (2004) later suggested that enrichment of the mantle beneath the Mallina Basin possibly occurred at this time.

The geochemistry of the Warambie Basalt, particularly its enrichment in Th and LREE (Smithies et al., 2007a), is consistent with mantle sources similar to those that feed modern subduction-related basaltic magmas, but may also reflect assimilation of crust. Smithies et al. (2007a) noted that compared to otherwise similar subduction-related tholeiites, such as those of the Whundo Group, the Warambie Basalt is enriched in Nb and, in particular, Na_2O (up to 4.5 wt%) — both features of rift-related magmas.

Mallina Basin

The 2970–2930 Ma Mallina Basin (Figs 1, 12, and 13) is 100 km wide, over 250 km long, and filled by a 2–4 km thick succession of conglomerate, sandstone, and shale, deposited in submarine fans (Eriksson, 1981). The

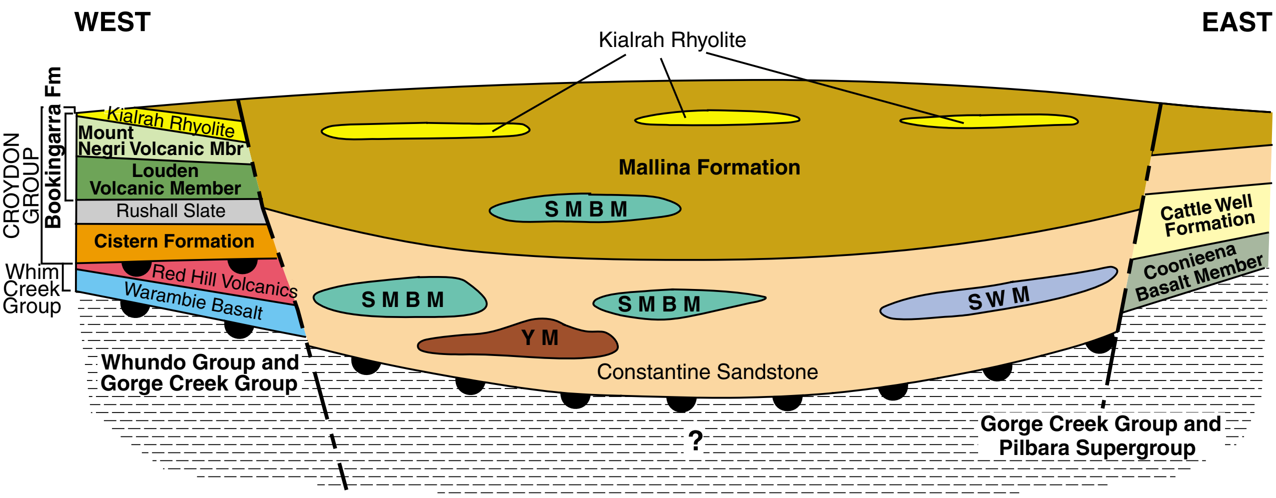

Figure 12. Diagrammatic cross section of the Mallina Basin, showing units of the Whim Creek and Croydon groups. The Louden Volcanic Member, Mount Negri Volcanic Member, South Mallina Basalt Member (SMBM), Salt Well Member (SWM), Yareweeree Member (YM), and Coonieena Basalt Member are all components of the Bookingarra Formation of the Croydon Group (after Van Kranendonk et al., 2006).

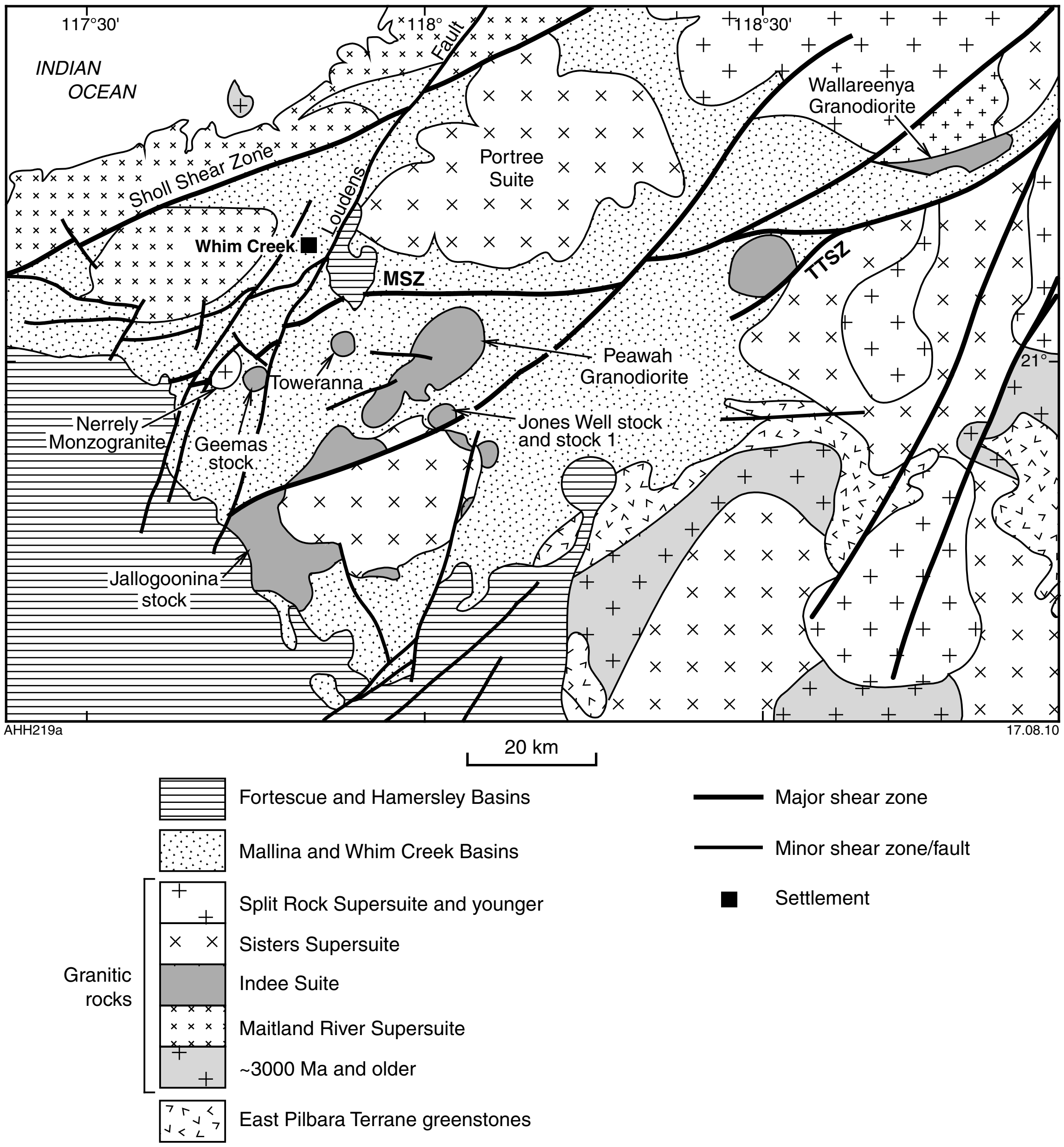

Figure 13. Simplified geological map of the western part of the De Grey Superbasin, showing major faults. MSZ, Mallina Shear Zone; TTSZ, Tabba Tabba Shear Zone.

basin overlies the 3070 Ma collision zone between the East Pilbara Terrane and the West Pilbara Superterrane, and is unconformably underlain by metamorphosed c. 3200 Ma oceanic crust, bimodal volcanics of the 3130–3110 Ma Whundo Group, and the 3050–3015 Ma Gorge Creek Group. Along its northwestern margin, the basin unconformably overlies the Whim Creek Group and Maitland River Supersuite, but these 3010–2980 Ma units lie in a linear belt parallel to the Pilbara coast, and there is no evidence they extend to the southeastern side of the basin. Nd T_{DM} model ages from the Croydon Group and Sisters Supersuite within the main part of the basin (excluding the Whim Creek greenstone belt) are younger than 3280 Ma (Smithies et al., 2007a).

The stratigraphy and structural history of the basin records periods of extension and rift-related deposition (clastic sediments), intrusion (ultramafic–mafic sills), and minor basaltic magmatism, alternating with folding, strike–slip faulting, and thrusting during northwest–southeast compression. The intensity of deformation increases in a southwesterly direction, towards the Tabba Tabba Shear Zone, the basin's tectonic contact with the East Pilbara Terrane (Fig. 1). Minor basaltic lava flows, locally including hyaloclastite breccias, are of siliceous high-Mg basalt (Smithies et al., 2004a). Between 2950 and 2940 Ma, intrusion of the Sisters Supersuite into the basin included alkali granites, high-Mg diorites (sanukitoids) and monzogranites (Smithies and Champion, 2000). The high-

Mg diorites followed emplacement of mafic subvolcanic sills with boninite-like compositions, and collectively provide evidence for the partial melting of subduction-enriched mantle (Smithies et al., 2004a).

Within the Whim Creek greenstone belt, the Croydon Group is represented by one volcanic and two clastic formations — all with direct lithostratigraphic equivalents within the southeastern part of the Mallina Basin. The basal Cistern Formation is an upward-fining succession of subaqueous fanglomerate, breccia, and sandstone, deposited unconformably on uplifted and eroded rocks of the Whim Creek Group (Pike and Cas, 2002). This formation is overlain by shale (Rushall Slate) deposited in isolated half graben sub-basins (Pike et al., 2006). Overlying the Rushall Slate is the Bookingarra Formation, composed of two siliceous high-Mg basalt members — the Mount Negri Volcanic Member and the Louden Volcanic Member.

Mosquito Creek Basin

The Mosquito Creek Basin (Fig. 1) trends between east–west and northeast–southwest, over an estimated strike length of 400 km (Hickman, 2004a). The present width of the basin, about 30 km, is the result of orogenic compression at 2900 Ma. Hickman (1978, 1983) defined the basin as containing a single, predominantly clastic sedimentary unit, the Mosquito Creek Formation, although he noted that the lowermost 1000 m of the succession, which unconformably overlies the Warrawoona Group, is a mixed package of metamorphosed clastic sedimentary rocks, chert, and mafic and felsic volcanic rocks, intruded by ultramafic–mafic sills. Farrell (2006) used more detailed geological mapping and the earliest geochronology obtained from the basin to divide the basin fill into two separate formations. In Farrell's stratigraphic interpretation, the lower succession of mafic–ultramafic rocks and associated sedimentary rocks is assigned to the Coondamar Formation, whereas the bulk of the fill remained within the Mosquito Creek Formation. Current geochronological data indicate that the Mosquito Creek Formation was deposited between 2970 and 2930 Ma (Bagas et al., 2008), whereas the Coondamar Formation is c. 3200 Ma.

Granitic supersuites (3025–2920 Ma)

Three granitic supersuites — the 3025–3010 Ma Orpheus Supersuite, the 3000–2980 Ma Maitland River Supersuite, and the 2955–2920 Ma Sisters Supersuite — intruded the De Grey Superbasin during its deposition, and also intruded both the West Pilbara Superterrane and East Pilbara Terrane.

Orpheus Supersuite

Following the Prinsep Orogeny, the 3025–3010 Ma Orpheus Supersuite was emplaced into rocks on both sides of the Sholl Shear Zone, both above and below the Regal Thrust. Small stocks of granodiorite, and sills of porphyritic dacite, intruded units as young as the Cleaverville Formation. Limited geochronological data suggest that the Andover Intrusion (a layered ultramafic–mafic intrusion south of Roebourne) was also emplaced at this time, although another possibility is that it is a faulted section of the c. 3120 Ma Bullock Hide Intrusion (Railway Supersuite).

Maitland River Supersuite

Large c. 2990 Ma felsic intrusions (Barley et al., 1984) form much of the Red Hill Volcanics, and the group is both underlain and flanked to the northwest by intrusions of tonalite, granodiorite, and monzogranite that form part of the 3000–2980 Ma Maitland River Supersuite (Van Kranendonk et al., 2006). This supersuite makes up much of the extensive Cherratta Granitic Complex in the far west Pilbara, but also forms a northeast-trending zone of granitic intrusion, over 200 km long and 50–100 km wide, immediately northwest of the Whim Creek greenstone belt (Hickman, 2004a). According to Smith (2003), most of the supersuite comprises rocks of the tonalite–trondhjemite–granodiorite (TTG) series. This suggests that the supersuite was derived by partial melting of subducted oceanic crust, and represents generation of new felsic crust as a continental magmatic arc. The age and distribution of the supersuite are consistent with it forming part of the same magmatic system as the Whim Creek Group, which Pike and Cas (2002) placed in an ensialic back-arc basin. As noted previously, Barley (1987), Krapež, (1993), and Van Kranendonk et al. (2007b) all suggested that the Whim Creek Group formed in a pull-apart basin, perhaps as part of the same tensional regime that facilitated intrusion of the slightly older Orpheus Supersuite.

If a 3010–2980 Ma subduction zone existed to the northwest of the Whim Creek greenstone belt, parallel to what is now the northwestern margin of the Pilbara Craton, the southeastwards subduction of mafic crust beneath the Whim Creek greenstone belt and Mallina Basin region may have provided the metasomatized mantle source required to produce many of the magmas of the 2955–2920 Ma Sisters Supersuite, which intruded into both the Mallina Basin and the northwestern part of the East Pilbara Terrane (Smithies et al., 2004).

Sisters Supersuite

The 2955–2920 Ma Sisters Supersuite is mainly composed of leucocratic, high-K monzogranite, derived from remelting of older basement components beneath the Mallina Basin. Widespread monzogranite magmatism developed earliest and was most voluminous in, and immediately adjacent to, the Mallina Basin, and thereafter migrated to the east. However, four other compositionally distinct suites of intrusive rocks are also recognized. Within the Mallina Basin, these include: 2955–2945 Ma sanukitoids of the Indee Suite (Smithies and Champion, 2000; Smithies et al., 2004a; Van Kranendonk et al., 2006); unusual LREE-enriched gabbroic rocks of the Langenbeck Suite (Smithies, 2002); and the Portree Suite of 2950 Ma lower crustal, high-temperature alkaline granites (Fig. 13; Smithies and Champion, 2000). In the West Pilbara Superterrane, the c. 2925 Ma Radley Suite of ultramafic–mafic intrusions was intruded contemporaneously with late dextral movement on the Sholl Shear Zone.

The magmatic history of the Mallina Basin at c. 2950 Ma (Langenbeck and Indee Suites) reflects melting of a mantle source, but also involves a Th-, Zr-, and LREE-enriched crustal component (Smithies and Champion, 2000; Smithies, 2002; Smithies et al., 2004a,b). Although individual members of the Langenbeck Suite significantly differ geochemically, they share relatively constant La/Sm and La/Zr ratios that cannot be accounted for through assimilation of any known Pilbara crustal component (Smithies et al., 2004a). Likewise, sanukitoids of the Indee Suite show trends to sympathetic enrichments in LILE, Cr, Ni, and Mg#, which cannot be tied to crustal assimilation. Rather, the Langenbeck and Indee Suite magmatism is interpreted to derive from a mantle source that was modified (enriched) during an earlier subduction event, possibly that which formed the Whundo Group (see Sholl Terrane).

Nd model age data from high-K monzogranites of the Sisters Supersuite indicate multiple sources. Southeast of the Mallina Basin, these sources include Paleoarchean rocks of the East Pilbara Terrane, whereas within the Mallina Basin the sources were younger than c. 3200 Ma, and likely included the 3200–3165 Ma Mount Billroth Supersuite and the 3130–3110 Whundo Group. T_{DM} data from high-K monzogranites of the Sisters Supersuite also show a linear zone along northwestern margin of the East Pilbara Terrane, where the bulk granite source region was unusually young (3100–3000 Ma) and juvenile; this may include the Elizabeth Hill Supersuite (which intrudes both the East Pilbara Terrane and West Pilbara Superterrane), the Whim Creek Group, or possibly the Maitland River Supersuite. Alternatively, there may have been a direct mantle contribution to the bulk source along this zone.

Cutinduna and Split Rock Supersuites (2910–2830 Ma)

The 2910–2890 Ma Cutinduna Supersuite, composed of monzogranite, is restricted to the southeastern part of the East Pilbara Terrane, and inliers of the Kurrana Terrane within the Fortescue Group. The supersuite is the same age as the Mosquito Creek Orogeny, and occurs on both sides of the Mosquito Creek Basin.

The 2890–2830 Ma Split Rock Supersuite consists of highly fractionated, Sn–Ta–Li bearing, post-tectonic monzogranites that were emplaced in a linear belt trending southeast–northwest across the Kurrana and East Pilbara Terranes. Nd model age data indicate derivation of the Split Rock Supersuite from partial melting of much older granitic crust, commonly with model ages of between 3400 Ma and >3700 Ma (Smithies et al., 2003).

Previous stratigraphic and tectonic interpretations

Based on data from GSWA mapping in the northern part of the craton, the supracrustal successions of the Pilbara Craton (Warrawoona Group and all younger units) were interpreted as deposited on older continental crust (Hickman, 1981). In this interpretation, the Warrawoona Group (a dominantly volcanic succession) was followed by lithologically more diverse successions contemporaneous with successive periods of deformation, metamorphism, and granitic intrusion. Lateral variations in the supracrustal stratigraphy were recognized within, and between, different greenstone belts but, with the exception of the Whim Creek Group, the main stratigraphic groups (Warrawoona and Gorge Greek) were thought to be represented across the northern half of the craton (Hickman, 1981, 1983). The Whim Creek Group and overlying basaltic formations ('Louden Volcanics' and 'Negro Volcanics') were interpreted to be remnants of younger succession that had been deposited unconformably on the Warrawoona and Gorge Creek Groups in only the Whim Creek area of the west Pilbara.

Some research published during the 1980s suggested that the Pilbara Craton was an assemblage of discrete tectonostratigraphic domains, separated by major northeast-trending lineaments (Groves and Batt, 1984; Krapež and Barley, 1987). From 1990 onwards, several researchers working in the Cleaverville area reported evidence for accretion at a convergent margin (Isozaki et al., 1991; Kiyokawa, 1993; Ohta et al., 1996; Kato et al., 1998; Kiyokawa and Taira, 1998); however, these interpretations at Cleaverville were based on a premise that a greenstone assemblage of pillow basalt, chert, shale, and BIF is diagnostic of an oceanic succession. The importance of this premise in framing the resulting tectonic models is illustrated by Isozaki and Maruyama (1992), who stated that 'ancient accretionary complexes are best recognized by the combination of oceanic plate stratigraphy (OPS) and imbricated thrust sheets'. They described OPS as gabbro or basalt 'of MORB affinity', overlain successively by pelagic chert, hemipelagic mudstone, and terrigenous sandstone and mudstone. Isozaki and Maruyama (1992) further stated, 'When combined with the field occurrence of imbricated thrust sheets, OPS can exclusively define an accretionary complex, regardless of age'. There are four major obstacles to applying this conclusion to Archean granite–greenstone terranes:

1. As discussed by Arndt (1999), submarine basalts are very common in Archean greenstone belts worldwide, but there is strong evidence that many of these basalts were deposited on submerged continental platforms. In the Pilbara Supergroup of the East Pilbara Terrane, there are several thick pillow basalt–chert formations that cannot be ophiolites because they unconformably overlie continental crust; examples include the Apex Basalt, Euro Basalt, and Kunagunarrina Formation (Van Kranendonk et al., 2007a).

2. Precambrian BIF – black shale successions do not provide evidence of deep-sea pelagic sedimentation, contrary to the assumption made by Isozaki et al. (1991). Isozaki et al. (1991) used the stratigraphy of a Phanerozoic accretionary orogen in southwest Japan as an analogue for the Cleaverville Formation. However, across the east Pilbara, the Cleaverville Formation overlies granites and greenstones of the East Pilbara Terrane across an erosional unconformity, and this formation was evidently deposited in either an intracontinental basin or on a continental shelf.

3. The existence of tectonic imbrication at Cleaverville has not been satisfactorily demonstrated, and remains highly contentious. Kiyokawa and Taira (1998) rejected the interpretation of Ohta et al. (1996) because they found that the Cleaverville succession is not a single, tectonically duplicated, 'OPS' succession. Instead, they interpreted it as a cyclic succession of three volcanosedimentary units. Faults and shear zones are present in the Cleaverville Formation at Cleaverville, but some are of Fortescue Group age (Hickman and Smithies, 2001), and others are related to a sinistral shear zone interpreted to lie just off the coast (Kiyokawa and Taira, 1998).

4. In the accretionary complex model used at Cleaverville, accretion must have taken place after the deposition of the Cleaverville Formation at 3020 Ma. Post-3020 Ma accretion of an 'oceanic' Cleaverville succession from the northwest is inconsistent with the fact that the Dampier Granitic Complex, which extends at least 80 km northwest of Cleaverville (Fig. 3), includes granitic rocks substantially older than 3020 Ma. Kiyokawa and Taira (1998) dated granite from the complex at 3140 ± 40 Ma, whereas zircon xenocrysts in other granites range in age up to 3255 Ma (GSWA 136844: Nelson 1998e), and Nd T_{DM} model ages on two granites from the Dampier Granitic Complex are 3360–3330 Ma (Smithies et al., 2007a).

Kiyokawa and Taira (1998) argued that the lack of oriented dyke complexes in the Cleaverville succession does not support a mid-oceanic ridge setting, and that the basalt–rhyolite–chert/BIF cycles do not support an oceanic (or back-arc) spreading ridge. They concluded that the Cleaverville succession was deposited in an immature island-arc setting. In this interpretation, subsequent thrusting was northwards, and occurred at sometime between 3190 and 3020 Ma. Kiyokawa and Taira (1998) included not only the Cleaverville and Regal Formations in the interpreted immature island-arc assemblage, but also the Ruth Well Formation, Andover Intrusion, and Harding Granitic Complex, an interpretation inconsistent with the stratigraphy and geochronology of the units involved. A full critical discussion is beyond the scope of this field guide, and will be presented in a future GSWA Report on the West Pilbara Superterrane. However, some problems with Kiyokawa and Taira's (1998) interpretation include:

1. The entire stratigraphy of the Karratha and Regal Terranes, and of the overlying Cleaverville Formation, is folded around the Roebourne Synform, Prinsep Dome, and the Cleaverville Synform (Fig. 9). Way-up evidence from pillow structures shows that the Regal Formation occupies the limbs of all these major fold structures, but Kiyokawa et al. (2002) recognized only the Cleaverville Synform. Without reference to way-up evidence outside the Cleaverville area, Kiyokawa et al. (2002) assigned the Regal Formation to three different stratigraphic units separated by thrusts. Figure 9 of this field guide shows way-up observations in the Regal Formation between Roebourne and Mount Regal, confirming its repetition by the three major folds.

2. The Ruth Well Formation is intruded by the c. 3270 Ma Karratha Granodiorite, and stratigraphically underlies the <3250 Ma Nickol River Formation, both of which were interpreted by Kiyokawa et al. (2002) to be continental crust underlying the thrusts. Consequently, these units cannot be part of the interpreted <3200 Ma immature island-arc assemblage.

3. The c. 3270 Ma Karratha Granodiorite forms a substantial part of the 'thrust slice' that wasinterpreted by Kiyokawa et al. (2002) to occur immediately north of the Sholl Shear Zone. Consequently, their interpretation, that this section of stratigraphy is part of the proposed <3200 Ma immature island-arc assemblage, cannot be correct.

4. SHRIMP U–Pb zircon geochronology on volcanogenic sedimentary rocks of the Cleaverville Formation in the west Pilbara indicates that the maximum depositional age of this formation is between 3022 Ma (GSWA 127330: Nelson, 1998a) and 3018 Ma (GSWA 142830: Nelson, 1998h), whereas its minimum age is indicated by felsic intrusions dated at 3015 Ma (GSWA 136899: Nelson, 1998d) and 3014 Ma (GSWA 127320: Nelson, 1997d). Consequently, the Cleaverville Formation could not have been deformed by thrusting older than 3020 Ma.

Krapež (1993) divided the Pilbara Craton into five tectonostratigraphic domains, separated by four northeasterly trending lineaments. These lineaments were said to define the 'dominant structural fabric' of the craton (Krapež, 1993), and to have had a long history of development and reactivation. Based on this interpretation, Krapež (1993) questioned the validity of the regional lithostratigraphic correlations made by Hickman (1983) in the Pilbara Craton, because these correlations extended across his domain boundaries. Accordingly, he replaced the previous lithostratigraphy with a completely new interpretation based on principles of sequence stratigraphy. This interpretation was later expanded upon in a subsequent publication (Krapež and Eisenlohr, 1998).

In a review of Pilbara Craton geology, Barley (1997) interpreted the craton to be composed of accreted terranes, with accretion taking place through a process of westward growth, and related the supersequences of Krapež (1993) to the opening and closing of ocean basins. Hickman (2004a) argued that this 'westward growth model' could not be correct for several reasons. Firstly, the two most easterly 'domain-boundary lineaments' (Krapež, 1993), within what is now the East Pilbara Terrane, are actually late-stage (post-2950 Ma) fault zones, cross-cutting a continuous older stratigraphy (Van Kranendonk et al., 2002). Secondly, the stratigraphy of the central and west Pilbara does not decrease in age westwards. Thirdly, Barley's (1997) interpretation that the West Pilbara Superterrane was accreted onto the rest of the Pilbara Craton between 3000 and 2900 Ma was incorrect because the c. 3020 Ma Cleaverville Formation extends unconformably across both these tectonic units and any accretion must therefore have been before 3020 Ma.

A detailed interpretation of west Pilbara plate tectonic environments was first published by Krapež and Eisenlohr (1998). This interpretation marked an important step towards understanding the tectonic evolution of the west

Pilbara terranes, although later systematic geological mapping, supported by geochronology and geochemistry, have required many changes to the model. Most importantly, the stratigraphic interpretation by Krapež and Eisenlohr (1998) was largely unconstrained by geochronological data, and was based on a hypothesis of global tectonic cycles. Two 'megacycle sets', spanning 3500–2775 Ma, were divided into four megacycles, each of 190–175 Ma duration, with each megacycle inferred to contain a megasequence that could be divided into supersequences or basins. Subsequent geochronology has not supported this hypothesis.

Beintema (2003) used all previously published data, plus new information derived from mapping and geochronology, to interpret the evolution of the west Pilbara in terms of a mid-Archean, active continental margin. In contrast to most previous workers, she found no evidence for terrane accretion such as island arcs, although the tectonic settings of Mesoarchean units in the west Pilbara were related to low-angle, southeast-directed subduction outside, and to the northwest of, the craton. Two subduction events (3265–3150 Ma and c. 3010 Ma) were recognized, with intervening and later periods of transpression and transtension including formation of the Sholl and Tabba Tabba Shear Zones. Some of Beintema's (2003) conclusions are consistent with available data, but other parts of her tectonic model are not supported, including:

1. Geochemical evidence indicates that the 3270–3235 Ma Sulphur Springs Group and contemporaneous Cleland Supersuite, correlated with the Roebourne Group and Karratha Granodiorite of the west Pilbara, were the result of a mantle plume event (Van Kranendonk et al., 2002, 2006, 2007a, b), not a subduction-related continental arc as interpreted by Beintema (2003).

2. Geochemical evidence indicates that the 3130–3110 Ma Whundo Group was deposited in an oceanic-arc setting (Smithies et al., 2005a, 2007c), not in an intracontinental rift as interpreted by Beintema (2003).

Deformation events

In this field guide, D events are numbered as they occur in the west Pilbara; the existence of several Paleoarchean deformation events in the East Pilbara Terrane requires a different numbering system in that area. D_1 of the West Pilbara Terrane is absent in the east Pilbara, and D_2 of the West Pilbara Terrane is considered to be equivalent to D_5 of the East Pilbara Terrane.

Rifting event, 3220–3165 Ma

Rifting of the East Pilbara Terrane followed deposition of the Sulphur Springs Group (Hickman, 2004a) and intrusion of the Cleland Supersuite, and evolved to establish a northeast–southwest trending basin of mafic oceanic crust between the East Pilbara Terrane and the Karratha Terrane (Van Kranendonk et al., 2007a). In the East Pilbara Terrane, horst and graben faults (Wilhelmij and Dunlop, 1984) accompanied unconformable deposition of the Soanesville Group above the East Pilbara Terrane. The southeastern margin of the basin was locally intruded by c. 3180 Ma mafic intrusions of the Dalton Suite and 3180–3165 Ma granitic intrusions of the Mount Billroth Supersuite; intrusions of this age have not yet been identified along the northwestern margin of the basin in the West Pilbara Superterrane.

D_{1WP} deformation, c. 3160 Ma

D_1 of the West Pilbara Superterrane is restricted to the area northwest of the Sholl Shear Zone, and does not extend into the East Pilbara Terrane. Its age may coincide with a c. 3160 Ma thermotectonic event in the Karratha Granodiorite (Kiyokawa, 1993; Smith et al., 1998). Beintema (2003) recorded a 3144 ± 35 Ma $^{40}Ar/^{39}Ar$ age in the Ruth Well Formation, which she interpreted as the cooling age of amphibolite-facies metamorphism simultaneous with D_1 thrusting. Hickman (2004a) suggested that D_1 was the result of obduction of the Regal Formation across the Karratha Terrane. The major thrust zone (not a single fault plane) at the base of the Regal Formation is the Regal Thrust. Structures within the Regal Thrust zone include recumbent folds, and are likewise assigned to D_1. A bedding-parallel tectonic foliation (S_1), preserved in metasedimentary rocks of the Nickol River Formation (Hickman et al., 2000) and in metabasalt of the Regal Formation, is interpreted to have initially formed parallel to the D_1 thrusts, but was reactivated by parallel shearing during later tectonic events.

The precise age of the Regal Thrust has not been directly established by geochronology, and work by Beintema (2003) suggests that $^{40}Ar/^{39}Ar$ geochronology is unlikely to be successful due to thermal disturbance events in the west Pilbara at 2870, 2700, 1900, 1200, and 900 Ma. Van Kranendonk et al. (2006) suggested that obduction of the Regal Formation was part of an event at 3070 Ma (D_2, below), although evidence of a metamorphic event at 3070 Ma has yet to be obtained in the West Pilbara Superterrane.

D_{2WP} deformation, c. 3070 Ma

Between 3110 Ma (late-stage deposition of the Whundo Group and intrusion of the Railway Supersuite) and c. 3050 Ma (earliest deposition of the Gorge Creek Group), the Sholl Terrane collided with the Karratha and Regal Terranes along the Sholl Shear Zone. Deformation related to this event (D_2) included sinistral movement on the Sholl Shear Zone, and thrusting of the Whundo Group onto the Cherratta Granitic Complex across the Maitland Shear Zone. There is surprisingly little geochronological data on the precise age of D_2, with evidence mainly limited to the age of intrusions in the Elizabeth Hill Supersuite, dated at c. 3070 Ma. Following 3070 Ma, the West Pilbara Superterrane and the East Pilbara Terrane shared major deformation events and some magmatic events, which indicates that collision between these tectonic regions formed part of D_2. For this reason, an orogenic event, the Prinsep Orogeny, was inferred by Van Kranendonk et al. (2006).

D_{3WP} deformation, c. 3020 Ma

D_3 deformation is represented by extensional growth faults in rocks of the 3050–3015 Ma Gorge Creek Group, which were deposited in a post-collision extensional basin. Early minor intrusions of the Orpheus Supersuite were probably controlled by these fractures.

D_{4WP} deformation, 3015–3010 Ma

D_4 structures are most strongly developed along major northeast–southwest trending faults in the Gorge Creek Basin within the west Pilbara. Tight to isoclinal, east-northeasterly trending folds in the Cleaverville Formation between Karratha and Cleaverville are attributed to the D_4 event. The same type of folding occurs in the Cleaverville Formation between Miaree Pool and Mount Regal. In the Regal Formation, S_4 is locally synchronous with intrusive sheets of 3015 Ma quartz–feldspar porphyry and porphyritic microgranite of the Orpheus Supersuite. At Mount Ada (15 km south of Roebourne), easterly trending, upright, tight to isoclinal D_4 folds in the Cleaverville Formation contain a sill of 3014 Ma granophyre (GSWA 127320: Nelson, 1997d), and these fold structures are unconformably overlain by the c. 3010 Ma Warambie Basalt of the Whim Creek Group.

D_{5WP} deformation, 3000–2960 Ma

D_5 structures are largely restricted to the Whim Creek and Croydon Basins. Thrusts in the Warambie Basalt east of Mount Ada (Hickman, 2002) predate dextral movement on the Sholl Shear Zone, and may be equivalent to Phase 3, recognized in the Whim Creek area by Krapež and Eisenlohr (1998). Originally east–west trending 'D_1 folds' on MOUNT WOHLER[1] (Smithies, 1998) and SATIRIST (Smithies and Farrell, 2000) may belong to the same event. These correlations suggest that D_5 occurred between 3000 and 2960 Ma.

D_{6WP} deformation, 2950–2930 Ma

In the west Pilbara, D_6 deformation was most strongly developed in the Mallina Basin (major D_3 folds of Smithies, (1998), and are equivalent to Phase 4 structures of Krapež and Eisenlohr (1998), but the event also formed major northeasterly trending tight to open folds such as the Prinsep Dome and the Roebourne Synform, which folded the Regal Thrust, and the Bradley Syncline which folded the Whundo Group and the underlying layer-parallel Maitland Shear Zone. Geochronology in the Mallina Basin (Smithies, 1998) establishes that the age of these structures must be 2950–2930 Ma. D_6 folds are oblique to the Sholl Shear Zone and to other strike-slip faults of the West Pilbara Superterrane, and are probably transpressional folds within a post-2950 Ma, ENE–WSW belt of dextral strike-slip movement. Minor D_6 structures include a steeply dipping, east-northeasterly striking axial-plane foliation (S_6) in the Prinsep Dome and in an anticline east of Mount Sholl. Minor D_6 folds deform S_1 southeast from Mount Regal.

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

D_{7WP} deformation, c. 2930 Ma

In the west Pilbara, localized deformation is present along the north-northwesterly striking section of the Maitland Shear Zone, where it truncates major, northeasterly trending D_6 folds of the Mount Sholl area. A parallel tectonic foliation (S_7) is developed in the adjacent greenstones of the Whundo Group, and in the rocks of the Cherratta Granitic Complex. Shear zones also occur within the Cherratta Granitic Complex where gneiss, interpreted to include deformed syntectonic granitic veins, contains late zircon populations dated at 2944 ± 5 Ma and 2925 ± 2 Ma (GSWA 136826: Nelson, 1997e).

D_{8WP} deformation, c. 2920 Ma

The latest movement on the Sholl Shear Zone, and probably on most other major faults in the West Pilbara Superterrane and Mallina Basin, was dextral. A subsidiary dextral strike–slip fault, the Black Hill Shear Zone, displaces the Andover Intrusion by 10 km, south of Roebourne. As noted by Krapež and Eisenlohr (1998), zircon geochronology on several rock units close to the Sholl Shear Zone has revealed a metamorphic disturbance event at about 2920 Ma, and this event may have coincided with D_8. Minor D_8 structures in the Sholl Shear Zone include dextral drag folding and isoclinal folding of mylonite lamination, and associated small-scale faulting and brecciation.

Geology of the Fortescue Basin

The Fortescue Basin consists entirely of the Fortescue Group, a ~6 km thick, predominantly volcanic succession that unconformably overlies the Pilbara Craton across 250 000 km^2. There is evidence that this basin originally included the Ventersdorp Supergroup of southern Africa, and that it formed part of the Vaalbara supercontinent (Cheney et al., 1988; de Kock et al., 2009). If this is the case, the depositional thickness of the predominantly basaltic succession was up to 8 km, with a probable areal extent originally in excess of 1 000 000 km^2; in other words, the basin forms the volcanic succession of a Neoarchean LIP.

Deposition of the Fortescue Group spans 150 million years (2780–2630 Ma), with the basaltic rocks that distinguish this group erupted in three main events, each lasting no more than 10 million years. Almost all published accounts describe the volcanics as continental flood basalts, but basaltic andesite is more widespread than basalt (Thorne and Trendall, 2001), and the volcanic rocks in the upper part of the group are mafic–felsic volcanic cycles in the northwest of the basin (Kojan and Hickman, 1998). This volcanic cyclicity has been interpreted as the result of mantle plume events (Arndt et al., 2001). The dominantly volcanic part of the group

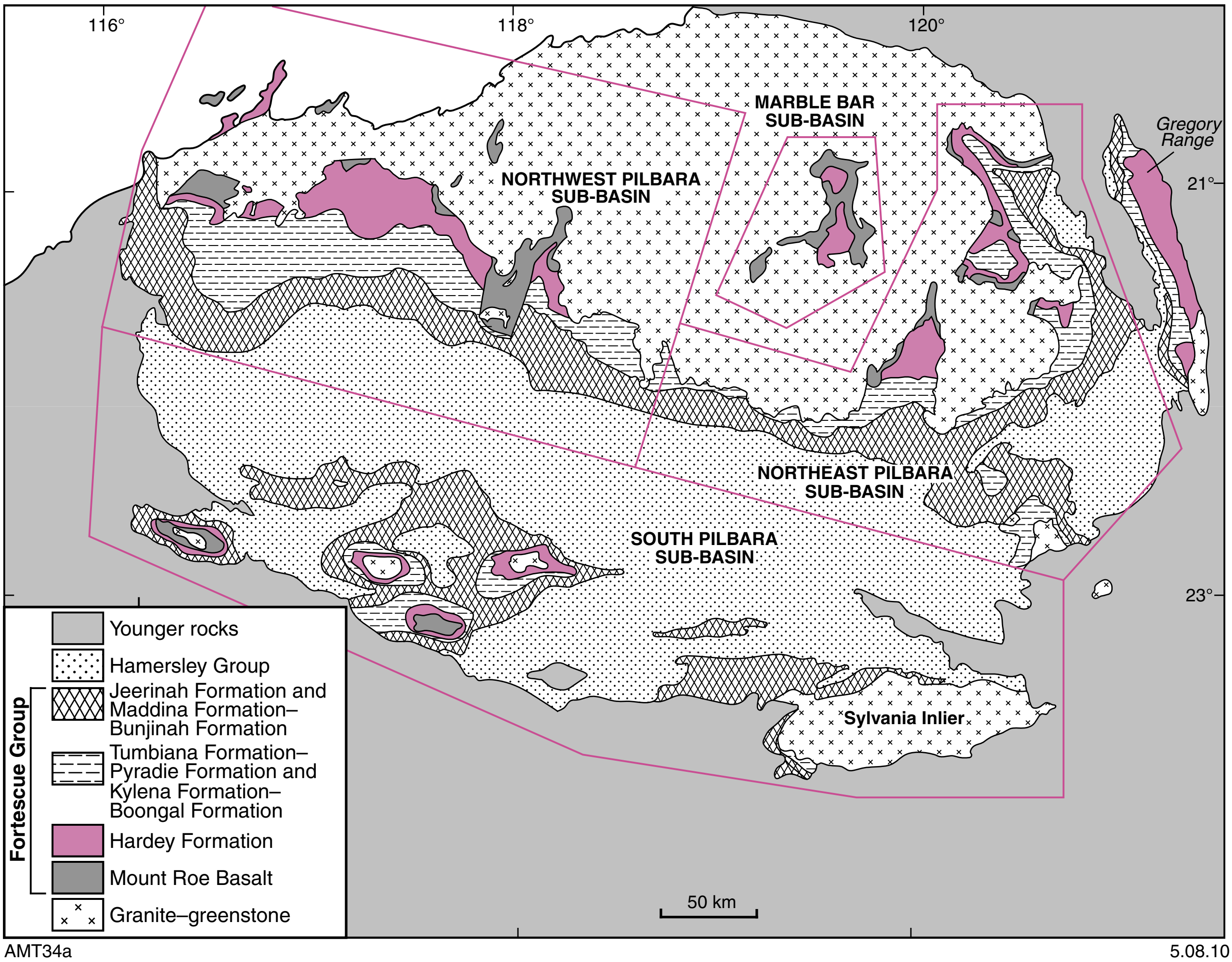

Figure 14. Geological map of the Pilbara Craton, and the Fortescue and Hamersley Basins, showing the location of the sub-basins of the Fortescue Basin (after Blake, 1984; modified from Thorne and Trendall, 2001).

can be divided into three depositional sequences, each of which represents a separate stage in depositional basin evolution (Thorne and Trendall, 2001). The upper part of the Fortescue Group, represented by the 2710–2630 Ma Jeerinah Formation (mainly composed of shale, chert, and BIF), marked a regional change from volcanism and shallow-water sedimentary deposition, to deeper water deposition. This part of the group is assigned to a fourth depositional sequence, which also includes the overlying Hamersley Group. Overall, the depositional sequences of the Fortescue and Hamersley Groups are interpreted as representing early crustal extension (Stages 1 and 2), breakup of the Pilbara Craton (Stage 3), and post-breakup shelf subsidence (Stage 4) at the southern margin of the Pilbara Craton (Tyler and Thorne, 1990; Blake and Barley, 1992; Blake, 1993; Thorne and Trendall, 2001; Cawood and Tyler; 2004).

In the northern Pilbara, Blake (1993) separated the Fortescue Group into three subaerial to shallow-water sub-basins, whereas in the south, a single subsiding sub-basin was recognized (Fig. 14). The lithostratigraphic successions in both the northern and southern sub-basins are summarized in Figure 15.

Stage 1 (2780–2770 Ma) was a period of crustal extension, with rifting and extrusion of the Mount Roe Basalt, fed by a swarm of north-northeasterly trending dolerite dykes (Black Range Dolerite Suite). In the northern Pilbara, extension was to the west-northwest–east-southeast, orthogonal to the southern Pilbara, where extension was towards the south-southwest (Fig. 16a; Thorne and Trendall, 2001). Sequence 1 rocks were deposited on a rugged, subaerial landscape with local relief of more than 500 m (Blake, 1993). In the northwest, northeast, and southwest Pilbara, basal sedimentary units including the Bellary Formation, and much of the pre-Fortescue Group topography, were buried beneath the Mount Roe Basalt, a pile of mostly subaerial basaltic flows, and minor lacustrine pillow lava and hyaloclastite up to 2.5 km thick. Dykes of the Black Range Dolerite Suite intruded both greenstones and granites, although the earliest basalt flows were channelled into valleys, where they appear to have 'ponded' to locally build up piles in excess of 1 km thick. Basalt flows proximal to the vents in granitic areas are not preserved, probably due to uplift and erosion prior to the unconformable deposition of the Stage 2 succession. Within the deeper paleovalleys, the Mount Roe Basalt is separated from the Fortescue Group's basal

NORTHERN FORTESCUE BASIN			***SOUTHERN FORTESCUE BASIN***	***TECTONO-STRATIGRAPHIC SEQUENCES***
MARBLE BAR SUB-BASIN	***NORTHEAST PILBARA SUB-BASIN*** **(Excluding Gregory Range)**	***NORTHWEST PILBARA SUB-BASIN***	***SOUTH PILBARA SUB-BASIN***	
	JEERINAH FORMATION Roy Hill Member Warrie Member Woodiana Member	JEERINAH FORMATION Roy Hill Member Nallanaring Member Warrie Member Woodiana Member	JEERINAH FORMATION	SEQUENCE 4
	MADDINA FORMATION Kuruna Member	MADDINA FORMATION	BUNJINAH FORMATION	SEQUENCE 3
	TUMBIANA FORMATION Meentheena Member Mingah Member	TUMBIANA FORMATION	PYRADIE FORMATION	
	KYLENA FORMATION	KYLENA FORMATION Gidley Granophyre	BOONGAL FORMATION	
HARDEY FORMATION	HARDEY FORMATION Bamboo Creek Member	HARDEY FORMATION Cooya Pooya Dolerite Lyre Creek Member	HARDEY FORMATION	SEQUENCE 2
MOUNT ROE BASALT	MOUNT ROE BASALT	MOUNT ROE BASALT	MOUNT ROE BASALT	SEQUENCE 1
GRANITE–GREENSTONE BASEMENT			BELLARY FORMATION ?	

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Figure 15. Stratigraphic subdivision of the Fortescue Group (modified from Thorne and Trendall, 2001). The locations of the sub-basins are shown in Figure 14.

unconformity by immature clastic sedimentary rocks of the Bellary Formation. Sequence 1 included lacustrine deposition of pillow basalt, plus fluviatile and fan-delta system conglomerate, sandstone and shale, together with local stromatolitic limestone (Hickman, 1983; Thorne and Trendall, 2001; Williams, 2001b).

Stage 2 (2770–2750 Ma) commenced with reactivation of the major Paleoarchean–Mesoarchean domes and faults in the northeast Pilbara, and with renewed movement on Mesoarchean faults and shear zones in the northwest Pilbara. In the northeast, the Mount Roe Basalt was folded into synclines, and formed graben that follow the trend of underlying axial faults of the greenstone belts. Reactivation of 2940 Ma strike–slip faults resulted in locally steep tilting of the Mount Roe Basalt. In a detailed structural study of the Marble Bar Sub-basin, Van Kranendonk (2003) mapped folds within the Mount Roe Basalt that are unconformably overlain by the Hardey Formation (Stage 2). Erosion of areas uplifted by the 2770–2760 Ma deformation led to rapid deposition of poorly sorted fluvial conglomerate and sandstone, and lacustrine shale, in paleovalleys of the northeast Pilbara, and within shallow rift basins of the northwest Pilbara. These epiclastic deposits belong to the 2760–2750 Ma Hardey Formation, and are described in detail by Thorne and Trendall (2001). In both the northeast and northwest Pilbara, the Hardey Formation includes large volumes of felsic volcanic and intrusive rocks, probably related to extensional faulting.

In the South Pilbara Sub-basin, Sequence 2 evolved through time from braided fluvial to deltaic depositional environments (Thorne and Trendall, 2001). As is the case in the north Pilbara, braided fluvial sandstones are compositionally and texturally immature, and were probably derived from basement granitic rocks. Sediment transport directions in fluvial and deltaic distributary channels were principally toward the southwest and west, and indicate the presence of a basement high to the north and northeast of the Milli Milli Dome. The greatly reduced thickness (<100 m) of the Hardey Formation in the SGS–1, WRL–1, and FVG–1 diamond drillholes, and the absence of Hardey Formation from the central north Pilbara, further supports this view. In the Newman area, there is a marked northward thinning of the Hardey Formation across the Sylvania Inlier. This evidence, combined with the absence of thick subaqueous deltaic facies suggests that the basement high north of the Milli Milli Dome may have extended close to the northern margin of the Sylvania Inlier (Fig. 16b).

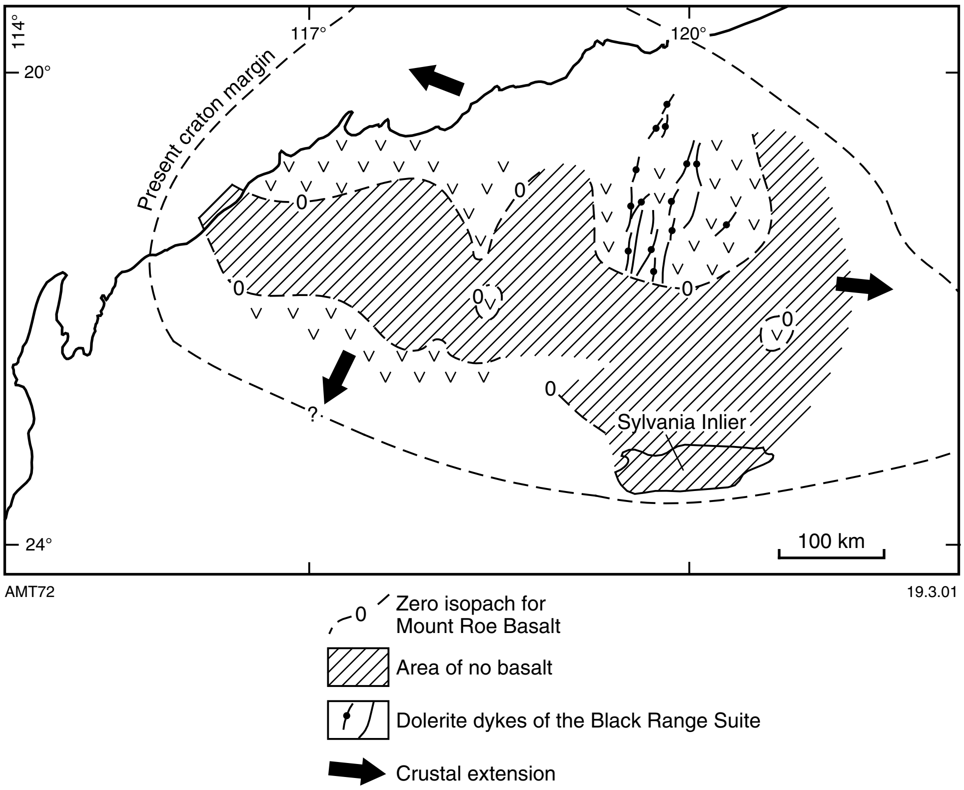
Stage 3 (2750–2710 Ma) volcanism and sedimentation in the northern Pilbara sub-basins differed from that of Stage 1 in several important respects. Firstly, most of the paleovalleys had already been largely filled by deposition of the Hardey Formation, with the result that Stage 3's three main units — the Kylena, Tumbiana, and Maddina Formations — were deposited across most of the northern Pilbara. Secondly, with the exception of the Marble Bar Sub-basin, rifting and folding had almost ceased by Stage 3 and the succession contains no conglomerate or arkosic sandstone units except at unconformable contacts with the Pilbara Craton. Evidence of regional uplift between Stages 2 and 3 is seen in the west Pilbara, where the Kylena Formation is locally absent between the Hardey and Tumbiana Formations (Hickman, 2004b); the Kylena Formation instead shows transgression onto uplifted areas of the Hardey Formation. Thirdly, each of the two volcanic formations are ultramafic–mafic–felsic volcanic cycles that include komatiitic basalt, basalt, andesite, dacite, and local rhyolite (Kojan and Hickman, 1998). These two volcanic cycles are separated by quartz sandstones, fine-grained volcaniclastic rocks, shale, and stromatolitic limestone of the 2727–2715 Ma Tumbiana Formation. In most areas, the lower and upper contacts of the Tumbiana Formation are apparently conformable with the volcanic formations above and below. However, angularly unconformable contacts with the underlying Kylena Formation are present in the west Pilbara (Hickman and Kojan, 2003), and Sakurai et al. (2005) recorded an unconformity with the overlying Maddina Formation in the central section of the Chichester Range. Thorne and Trendall (2001) described the Tumbiana Formation as a marine-shelf deposit, but recent workers have obtained sedimentological and geochemical evidence that it was deposited in a lacustrine environment (Bolhar and Van Kranendonk, 2007; Awramik and Buchheim, 2009). The volcanic cycles of Stage 3 have similarities to those of the Paleoarchean Pilbara Supergroup, and suggest derivation from plume-related melting of the mantle and lower crust (Arndt et al., 2001; Condie, 2001).

In the southern Pilbara, equivalent units to those in the northern Pilbara are the Boongal, Pyradie, and Bunjinah Formations, respectively, which have a combined thickness of about 3 km (Thorne and Trendall, 2001). Thorne and Trendal (2001) summarized the major tectonic and paleogeographic controls on Sequence 3. From north to south across the Pilbara Craton, there was a general change from subaerial and shallow-marine continental volcanism and sedimentation, to 'deeper' submarine volcanism. This change was accompanied by a marked thickening of the sequence toward the south (Horwitz, 1980; Morris and Horwitz, 1983), consistent with the main rift axis being to the south-southwest. The Yule–Sylvania High, which was a dominant feature of Stage 1 and 2 Fortescue Group paleogeography, was gradually onlapped and buried by the end of Tumbiana/Pyradie Formation deposition (Blake, 1993). In the southern Pilbara, lower and upper parts of Sequence 3 (Boongal and Bunjinah Formations) are dominated by 'tholeiitic' basalts, with massive flows, associated with pillow lava and hyaloclastite breccia, abundant in the 'deeper' marine-shelf settings. Strongly amygdaloidal flows, evident in the upper parts of the Bunjinah Formation, are also interpreted as submarine facies. Although reasons for increased vesicularity are unclear, it may be related to regional shallowing, and a consequent decrease in hydrostatic pressure caused by rapid aggradation of the volcanic pile.

The contrast between the northern and southern Pilbara successions is strongest when comparing the middle Sequence 3 units (Tumbiana and Pyradie Formations). Paleocurrent and provenance data suggest that much of the continental detritus in the northwestern stratigraphy was introduced from a low-relief granite–greenstone landmass, which existed to the north of the present day Chichester Range. The spatial and temporal distribution of pyroclastic air-fall tuff in the Tumbiana Formation suggests that there were at least two major hydrovolcanic eruptive centres in the north Pilbara during the deposition of this unit. The first of these eruptive centres was in the northeast Pilbara, and was mostly active during deposition of the lower Tumbiana Formation; the second centre affected the northwest Pilbara during deposition of the upper part of the stratigraphy. Coastal and shallow-marine facies of the Tumbiana Formation extend south to a poorly defined line connecting the northern Sylvania Inlier to the northern limb of the Jeerinah Anticline. South of this line, equivalent rocks of the Pyradie Formation are about 1 km thick, and comprise 'deeper' marine, pyroxene-spinifex textured basalt and hyaloclastite, komatiite, argillite, and chert (Fig. 16c).

Stage 4 (2690–2630 Ma) marked an important change in the style of deposition within the Fortescue Group, with the advent of predominantly marine deposition in a deepening basin (Thorne and Trendall, 2001). Thorne and Trendall (2001) reviewed previous interpretations of the contact between the Maddina and Jeerinah Formations, which represents the transition from Stage 3 to Stage 4. Although some workers had reported an unconformity in this position, and the boundary does mark a significant change from volcanism to predominantly sedimentary deposition, Thorne and Trendall (2001) reported no evidence of an erosional unconformity. Accordingly, they retained the previous stratigraphic interpretation, suggesting that the Jeerinah Formation forms the upper part of the Fortescue Group, rather than the lower part of the Hamersley Group.

a)



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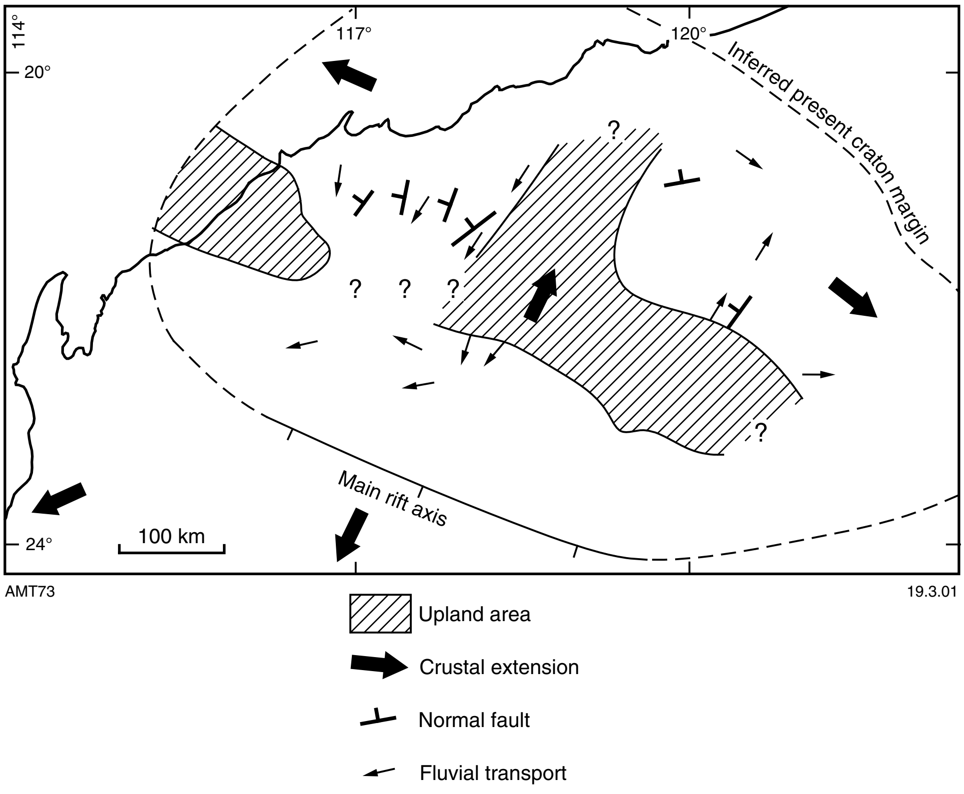

Figure 16. Tectonic and paleogeographic reconstructions for the Pilbara during deposition of Sequence 1 (a) Sequence 2 (b) and Sequence 3 (c) Thorne and Trendall (2001, fig. 14.1).

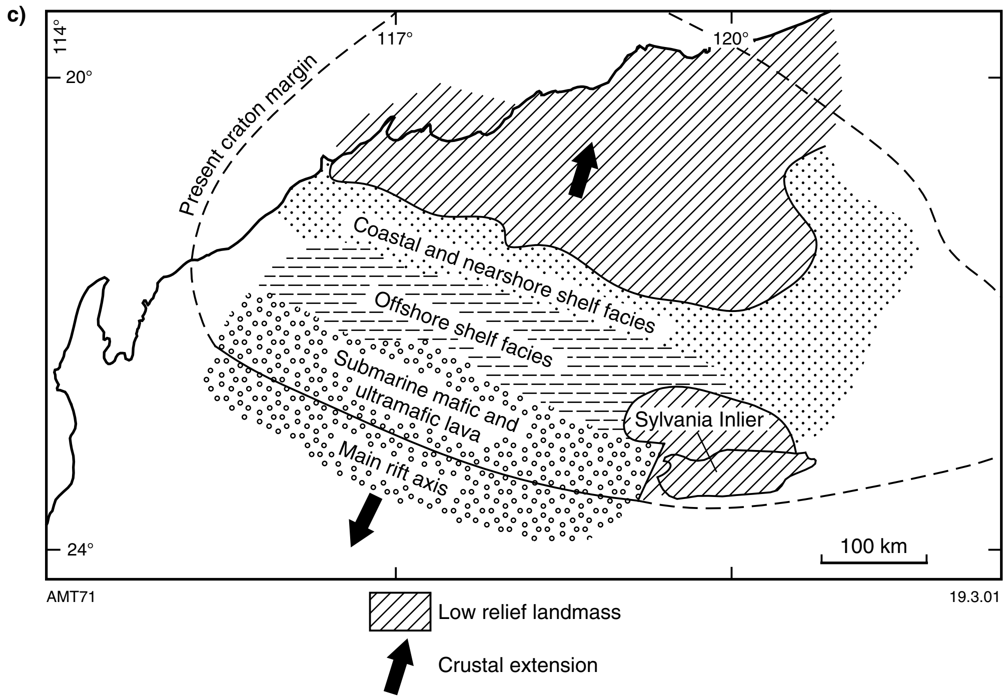

Figure 16 (continued)

Over much of the central and eastern Pilbara, the Jeerinah Formation is about 150 m thick and is subdivided into a lower division of nearshore shelf deposits up to 60 m thick (Woodiana Member; Packer and Walter, 1986; Packer, 1990; Thorne and Trendall, 2001), and an upper section of 'deeper' offshore shelf facies (carbonaceous argillite, carbonate, chert, volcanic sandstone, and BIF). In contrast, the Jeerinah Formation in the southern Pilbara ranges from 1 to 2 km thick, and consists of a laterally variable assemblage of 'deeper' shelf deposits comprising massive basaltic lava, pillow lava, hyaloclastite, argillite, chert, dolomite, and sandstone. Thorne and Trendall (2001) recorded marine basalts in parts of the Jeerinah Formation within the South Pilbara Sub-basin where mafic sills are common, and marine basalts are also present in the far southeast of the Northeast Pilbara Sub-basin.

Mafic sills form a significant part of both the Jeerinah Formation and the Hamersley Group's Weeli Wolli Formation (Thorne and Trendall, 2001). Within the Jeerinah Formation, sills constitute up to 60% of the stratigraphy, and are most abundant south of a line extending from the northwestern Sylvania Inlier to the northern Jeerinah Anticline. In this area, Tyler (1991) noted that the sills are confined to the region west of the Fortescue River Fault. Similarly, in the Milli Milli Dome – Jeerinah Anticline area, there is a relatively sharp boundary between Jeerinah Formation outcrops containing mafic sills and those that do not. This boundary coincides with the line of separation between shallow-shelf and deeper shelf facies in Sequence 4, and may also be fault-controlled (Thorne and Trendall, 2001). Thorne (1990a) suggested that an easterly to east-southeasterly trending, south-block-down growth-fault system, the Jeerinah–Sylvania Fault Zone, may have been active in the southern Pilbara at this time.

Geology of the Hamersley Basin and McGrath Trough

The Hamersley Basin is a latest Neoarchean to early Paleoproterozoic (2630–2450 Ma; Trendall et al., 2004) depositional basin, which overlies most of the southern part of the Pilbara Craton (Fig. 14; Trendall, 1983; Trendall, 1990b; Thorne and Trendall, 2001), and contains the BIF-dominated Hamersley Group (Trendall and Blockley, 1970). The apparently conformably overlying rocks of the McGrath Trough (Fig. 17) are made up predominantly of shale and clastic metasedimentary rocks of the Turee Creek Group, which includes glacigene rocks (Martin et al., 2000). The Turee Creek Group has not been directly dated, but was deposited between c. 2450 Ma and c. 2210 Ma (Trendall et al., 2004; Martin et al., 1998; Rasmussen et al., 2005; Martin and Morris, 2010). The Hamersley Basin initially represents continuation of the post-breakup shelf subsidence (Stage 4) seen at the southern margin of the Pilbara Craton (Tyler and Thorne, 1990; Blake and Barley, 1992; Blake, 1993; Thorne and Trendall, 2001; Cawood and Tyler, 2004). This margin has been interpreted to convert to a continental back-arc setting during deposition of the upper Hamersley Group (Brockman Iron Formation and above; Blake and Barley, 1992). The asymmetric, east-southeasterly trending McGrath Trough, which thins northwards, has been interpreted as a foreland basin in front of the northward-advancing Ophthalmia Fold Belt, and represent an accretionary, or collisional, foreland setting (Horwitz, 1982; Powell et al., 1999; Martin et al., 2000; Martin and Morris, 2010).

Thorne and Trendall (2001) and Trendall et al. (2004) have carried out assessments of the depositional history

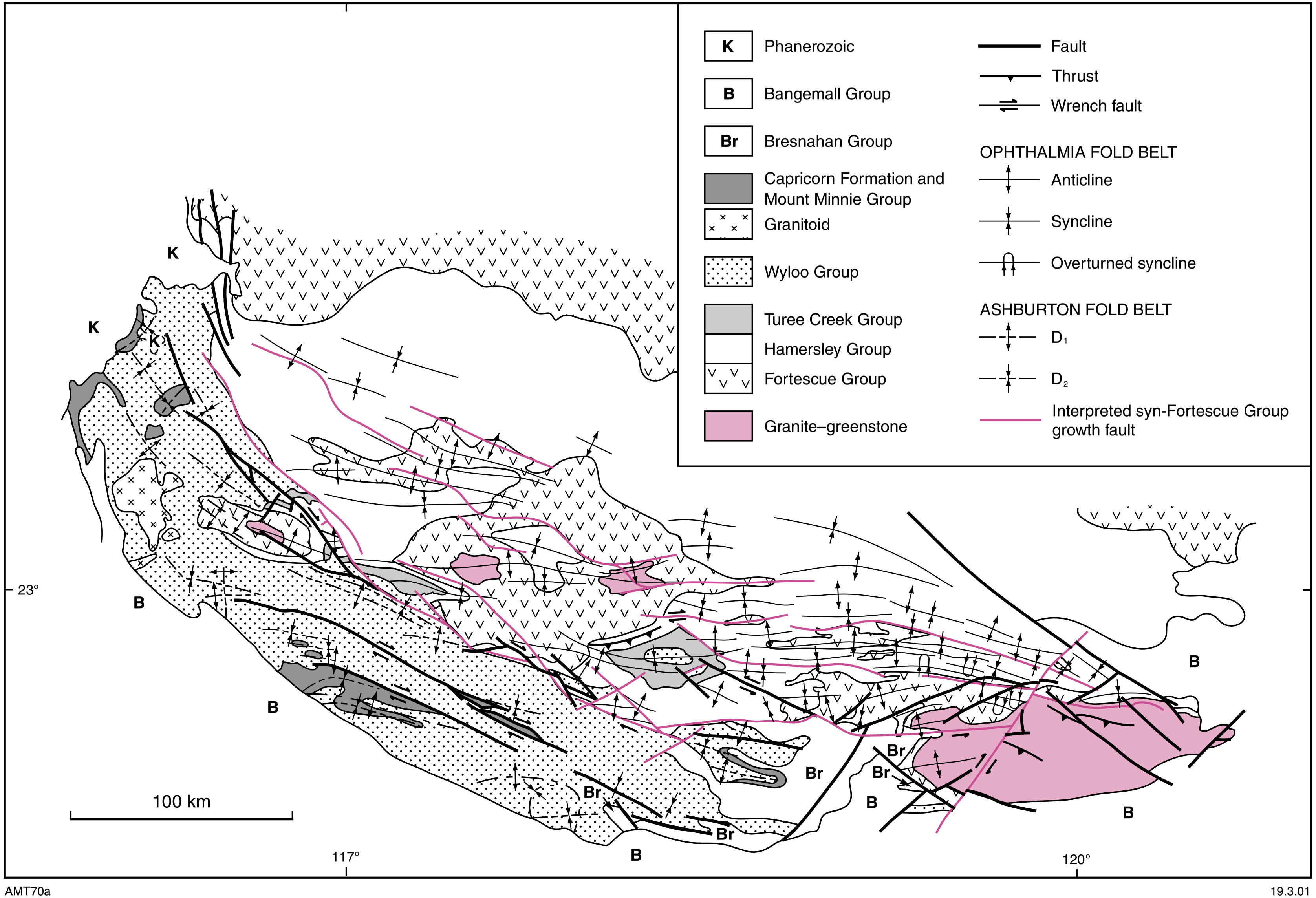

Figure 17. Simplified geological map of the southern Pilbara showing the main Ophthalmian (Ophthalmia Fold Belt) and Capricorn (Ashburton Fold Belt) Orogeny structural features and the probable location and orientation of the major syn-Fortescue Group faults. Thorne and Trendall (2001, figure 14.2, modified after Tyler, 1991).

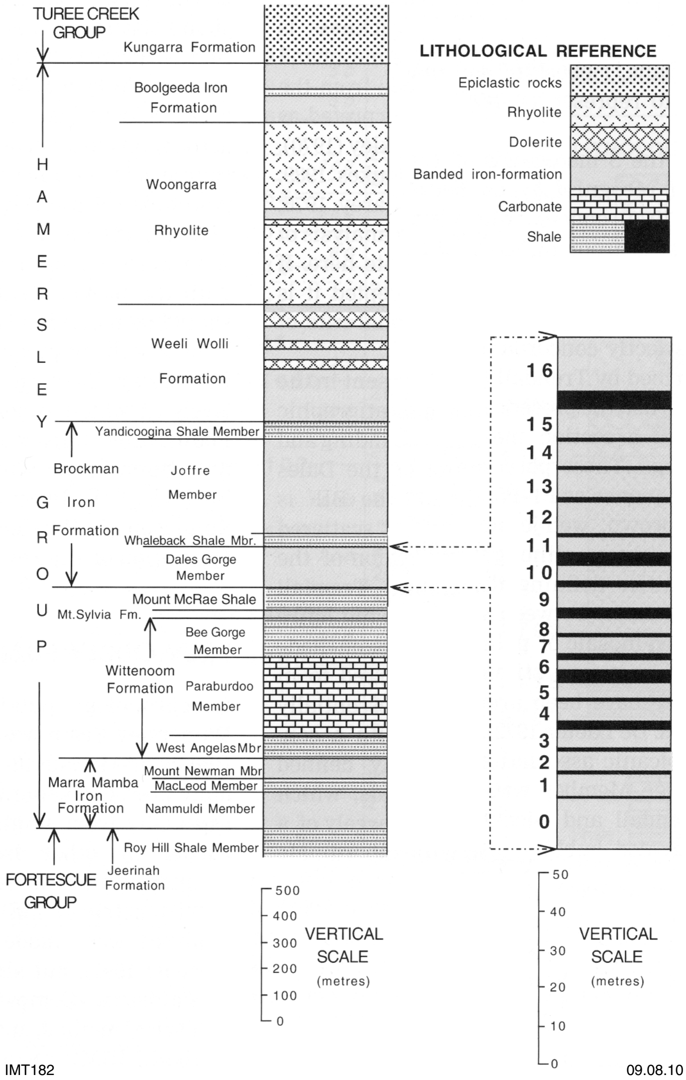

Figure 18. Lithostratigraphy of the Hamersley Group. Modified from Trendall et al (2004)

of the Hamersley Basin and its paleogeography. The Hamersley Group forms the remainder of Sequence 4. It is ~2.5 km thick, and includes BIF-dominated stratigraphic formations (Marra Mamba, Brockman, and Boolgeeda Iron Formations, and the Weeli Wolli Formation) up to 360 m thick (Fig. 18). The major stratigraphic units show relatively little regional variation in thickness or lithology across the basin (Trendall and Blockley, 1970; Trendall, 1983), and reflect an alternation of BIF with rocks other than BIF (Trendall,1990b). The Marra Mamba and Brockman Iron Formations are separated by the carbonate-dominated Wittenoom Formation, plus the Mount Sylvia Formation and Mount McRae Shale. The Woongarra Rhyolite has been interpreted by Trendall (1995b) as an intrusive, sill-like unit, and separates the Weeli Wolli Formation from the Boolgeeda Iron Formation, the uppermost BIF unit. Shallow water sedimentary structures are absent from the Hamersley Group (Simonsen et al. 1993; Thorne and Trendall, 2001) with the exception of the Carawine Dolomite, a stromatolitic shallow water carbonate unit in the northeast Hamersley Basin that is a lateral equivalent of the Wittenoom Formation, and parts of the Wittenoom Formation itself (Kargel et al., 1996). The group consists mostly of 'deeper' marine-shelf deposits (BIF, argillite, carbonate, and fine-grained volcaniclastic rock) deposited in a sediment-starved basin or shelf, or otherwise on a shelf platform, with a regional paleoslope from the northeast to the south and west (Trendall and Blockley, 1970; Morris and Horwitz, 1983; Simonson et al., 1993).

The BIF-dominated units are made up of generally thicker well-banded BIF, composed almost entirely of chert and iron oxides, interbedded with thinner units that are iron-rich but having a more shaly appearance due to their significant sheet-silicate content. Trendall and Blockley (1970) recognized distinct scales of banding — macrobanding, mesobanding, and microbanding — in the Dales Gorge Member of the Brockman Iron Formation. The BIF macrobands are interlayered with stilpnomelane-rich 'shale' macrobands, and consist of alternating silica-rich chert mesobands and iron-rich mesobands. The chert mesobands containing regularly spaced laminae or microbands are interpreted as annual layers (Trendall and Blockley, 1970). The regular sequence of macrobands is not seen in the Joffre Member or overlying Weeli Wolli Formation. The Weeli Wolli Formation is distinguished by the presence of metadolerite sills. The 'shale' macrobands are interpreted as volcaniclastic rocks, probably ash-fall tuffs.

In a detailed geochronological study of the Hamersley Group, Trendall et al. (2004) found that the average depositional rate for the BIF lithologies was 180 m per million years, compared to 12 m per million years for the carbonate lithologies, and 5 m per million years for the shale lithologies. The total body of SHRIMP U–Pb data from the Fortescue and Hamersley Groups was considered consistent with at least 330 million years of continuous basin-fill accumulation.

The Turee Creek Group was deposited in the McGrath Trough, and was derived predominantly from the south, becoming more mature upwards. Martin et al. (2000) identified two detrital sources; a cratonic source similar to the granite–greenstones of the Pilbara Craton, and a lesser source from BIF, and volcanic and sedimentary rocks of the Fortescue and Hamersley Groups. The Woogarra Rhyolite is a common source in the lower Turee Creek Group, and in particular the glacigene rocks, but becomes less common upsection.

The Ophthalmian Orogeny

The Ophthalmian Orogeny, which affected the southern Fortescue and Hamersley Basins (Fig. 17) to form the McGrath Trough and the Ophthalmia Fold Belt, is characterized by west- to northwest-trending, northerly vergent folds and thrusts (Horwitz, 1982; Tyler and Thorne, 1990; Cawood and Tyler, 2004). It forms the northern foreland to the Paleoproterozoic Capricorn Orogen, which formed during a series of suturing events that brought together the Archean Yilgarn and Pilbara Cratons by c. 1960 Ma, following Neoarchean to earliest Paleoproterozoic rifting, breakup, and drifting in the Fortescue and Hamersley Basins. Powell and Horwitz (1994; see also Martin et al., 1998; Martin and Morris, 2010) noted that the Ophthalmia Fold Belt developed during deposition of the upper Turee Creek Group, and corresponded in part with the angular unconformity at the base of the lower Wyloo Group. Deformation is known to have affected the lower Wyloo Group, ending before the intrusion of dolerite dykes at c. 2010 Ma (Muller et al., 2005), and deposition of the c. 1830–1805 Ma upper Wyloo Group in the Ashburton Basin (Cawood and Tyler, 2004). An age of around 2210 Ma for the Cheela Springs Basalt, within the lower Wyloo Group, gives an approximate age for this event (Martin et al., 1998); however, a direct age is provided by SHRIMP U–Pb dates of c. 2215 Ma and c. 2145 Ma, derived from monazite and xenotime crystals that grew during low-grade metamorphism (Rasmussen et al., 2005), and by SHRIMP U–Pb baddelyite and zircon ages of c. 2210 Ma and c. 2030 Ma for the mafic sills and volcaniclastic rocks that bracket the orogeny (Muller et al., 2005). The north-verging structural syle of this phase of orogeny suggests it may have been driven by accretion or collision from the south (Blake and Barley, 1992; Powell et al., 1999; Martin et al., 1998, 2000), with the most likely candidate being the older elements of the Glenburgh Terrane in the Gascoyne Province (Fig. 19; Johnson et al., 2010).

The Ophthalmia Fold Belt

Two groups of folds can be identified within the Ophthalmia Fold Belt (Fig. 17; Tyler and Thorne, 1990; Tyler, 1991). In the southwest, folds form broad-scale, open dome-and-basin structures that have a mainly northwest trend. In the southeast, folds have an easterly trend, are close to tight, and have short wavelengths. Tyler (1991) identified numerous shear zones linked to folding in the Ophthalmia Fold Belt in the Sylvania Inlier, an Archean granite–greenstone inlier in the southeast of the Fortescue and Hamersley basins (Fig. 17).

Two phases of deformation have been recognized, with early, small-scale, layer-parallel folds at particular stratigraphic horizons across the basins (including the

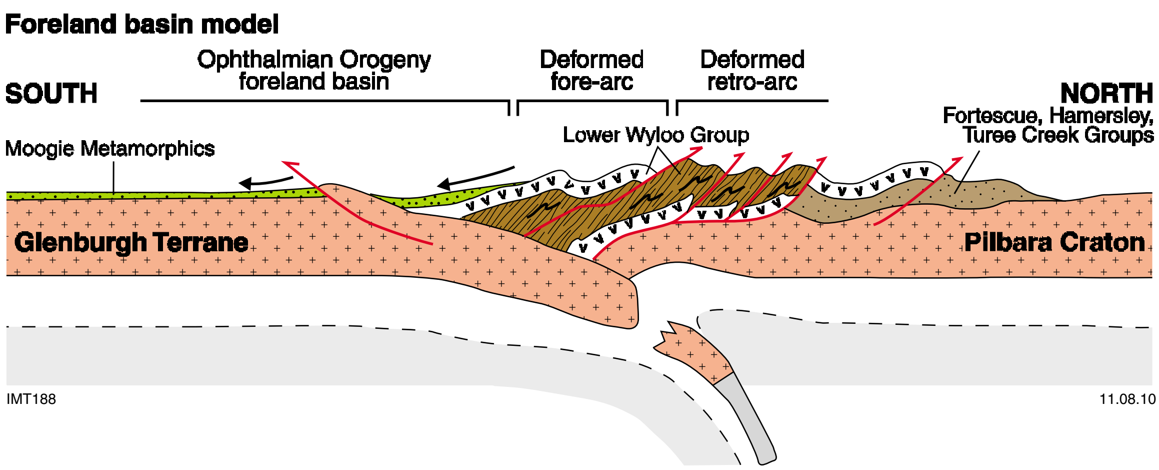

Figure 19. Schematic cross section, showing possible tectonic setting for the Ophthalmian Orogeny. Modified from Johnson et al. (2010)

upper Jeerinah Formation and the basal Boolgeeda Iron Formation). Tyler (1991) interpreted these deformation phases as being due to gravitational collapse in front of an advancing thrust system, but as movement directions have not been identified, they could also represent extensional features formed on the original basin paleoslope.

The main deformation event in the southeastern part of the fold belt takes the form of buckle-type folds with rounded hinges that are non-cylindrical and impersitant, dying out both laterally and vertically along their axial surfaces at all scales. Mesoscopic and macroscopic fold profiles are parallel to flattened parallel, and folds vary from open to isoclinal, often with a conjugate box-like form. Plunges are usually gentle to moderate, either to the east or west.

Folding can be separated based on its intensity into two main areas, with an area of high strain immediately to the north of the Sylvania Inlier, and an area of lower strain farther to the northwest and west. Folding passes through asymmetrical overturned north-facing folds, into reverse faulting, and steeply inclined upright folding (Fig. 17). An axial plane cleavage is well developed north of the Sylvania Inlier, but in the lower strain areas this cleavage is less prominent or absent.

In general, deformation dies out in the area recognized in the paleogeography of the Fortescue Basin as the former Yule–Sylvania High (Fig. 16). The distribution of faulting and folding reflects a strong basement control, suggesting that the compressional structures represent reactivated extensional structures formed during syn-Fortescue Group rifting.

Reactivation of the Nanjilgardy Fault System in the western Ophthalmia Fold Belt at this time also reflects control by original syn-Fortescue Group basement faults, producing complex and much-debated structural and stratigraphic relationships (a discussion beyond the scope of this field guide) between the Hamersley, and Ashburton Basins, plus the development of the McGrath Trough (Horwitz, 1982; Tyler and Thorne, 1990; Powell et al., 1999; Martin et al., 1998, 2000; Muller et al., 2005; Martin and Morris, 2010).

Post Ophthalmia Fold Belt faults

Major south- and southeast-dipping normal faults, including the Mount Whaleback Fault and the Homestead Fault, occur at the southeast margins of the Hamersley and Fortescue Group (Fig. 17). This extensional fault system, probably formed during the formation of the Bresnahan Basin along the margin of the Pilbara Craton during the Mangaroon Orogeny at c. 1690 Ma (Fig. 17; Cawood and Tyler, 2004).The world-class Mount Whaleback hematite iron ore deposit occurs in the hanging wall of the Mount Whaleback Fault.

Part 2.
Excursion localities

This six-day excursion for the Fifth International Archean Symposium starts and finishes at Karratha, and visits a series of geological sites that collectively indicate that plate tectonic processes operated from at least 3200 Ma. Areas visited include the >3270–3070 Ma West Pilbara Superterrane, where, from 3220 Ma, the sequence of tectonic environments closely resembles that seen in a Wilson cycle. The excursion also visits successions of the De Grey Superbasin, which include a 3010–2990 Ma continental volcanic arc, and 2950 Ma boninite-like rocks and sanukitoids related to late-tectonic orogenic relaxation and slab breakoff. Two days of the excursion take participants on a traverse across the Neoarchean–Paleoproterozoic Fortescue and Hamersley Basins to examine evidence that their tectonic settings changed from rifting at 2775 Ma, through a passive margin setting at 2630–2400 Ma, to collision during the Ophthalmian Orogeny at c. 2200 Ma.

Although the guide provides sufficient location detail to permit its later use by anyone wishing to independently follow the excursion route, it should be noted that access to some of the sites requires prior approval from landowners, and for many localities the nature of the terrain makes a four-wheel drive vehicle necessary. A general view of the excursion route and the localities is shown on Figure 20, and Figures 9 and 13 provide more local geological detail.

DAY 1.
Regal and Karratha Terranes: thrusting of c. 3200 Ma oceanic crust across at least 3000 km^2 of Paleoarchean continental crust

The first day of the excursion visits localities providing structural evidence for regional-scale thrusting of the 3200 Ma Regal Formation across continental crust comprising the c. 3300 Ma Ruth Well Formation, the 3270 Ma Karratha Granodiorite, and the <3250 Ma Nickol River Formation. Lunch will be taken at Locality 3.

Locality 1: Oceanic crust of the Regal Terrane above continental crust of the Karratha Terrane (Dampier, MGA 484200E 7706350N)

Locality 1 is on the ridge overlooking Karratha, close to the water tanks near the Karratha Tourist Office (Dampier MGA 484200E 7706350N).

The strongest structural evidence for Mesoarchean plate tectonic processes in the West Pilbara Superterrane lies in the fact that at least 3000 km^2 of the Paleoarchean continental crust of the Karratha Terrane is overlain by c. 3200 Ma oceanic crust of the Regal Terrane (see Part 1). The contact between these two terranes, the Regal Thrust, is a 1 km thick bedding-parallel tectonic complex that includes the upper Karratha, and basal Regal Terranes. These components are juxtaposed in tectonic slices separated by mylonites. Although metasedimentary rocks of the Nickol River Formation make up most of the thrust zone, some thrusts extend into the underlying Ruth Well Formation. The magnitude of deformation in this thrust zone is consistent with the Regal Terrane being highly allochthonous.

Locality 1 is a convenient point to commence the excursion because it serves as a viewing point for the area around Karratha, and also because it provides outcrops of the amphibolite-facies metabasalts of the Regal Formation (see Part 1). These metabasalts are typically strongly foliated parallel to the underlying thrust zone, and pillow structures are difficult to recognize in most outcrops along the ridge. However, in areas where the formation is less deformed, the Regal Formation can be seen to be composed mainly of pillow basalt containing thin units of chert, banded iron-formation, and shale, and is commonly intruded by sills of dolerite and gabbro. One locality where pillow structures are preserved is conveniently located at the footbridge leading to the Karratha Tourist Office, only 400 m to the east of Locality 1.

The prominent dolerite dykes that intrude the Regal Formation along the ridge remain undated, although their low metamorphic grade, lack of any metamorphic foliation, and northerly trend suggest that they represent feeder dykes for Neoarchean basalts of the unconformably overlying Fortescue Group.

Locality 2: Karratha Granodiorite, Karratha Terrane (Dampier, MGA 480520E 7695920N)

From the Karratha Tourist Office, drive 7 km southwards to Highway 1, then turn right and drive 10.5 km westwards to the Tom Price Road. Turn left and drive 7 km south to the crossing on the Hamersley Iron Railway. From this point, access to Locality 2 requires prior approval from the manager of Karratha Station. East from the railway crossing, enter Karratha Station through a gate, and drive approximately 3 km northeastwards along a good gravel track to Locality 2 (Dampier MGA 480520E 7695920N).

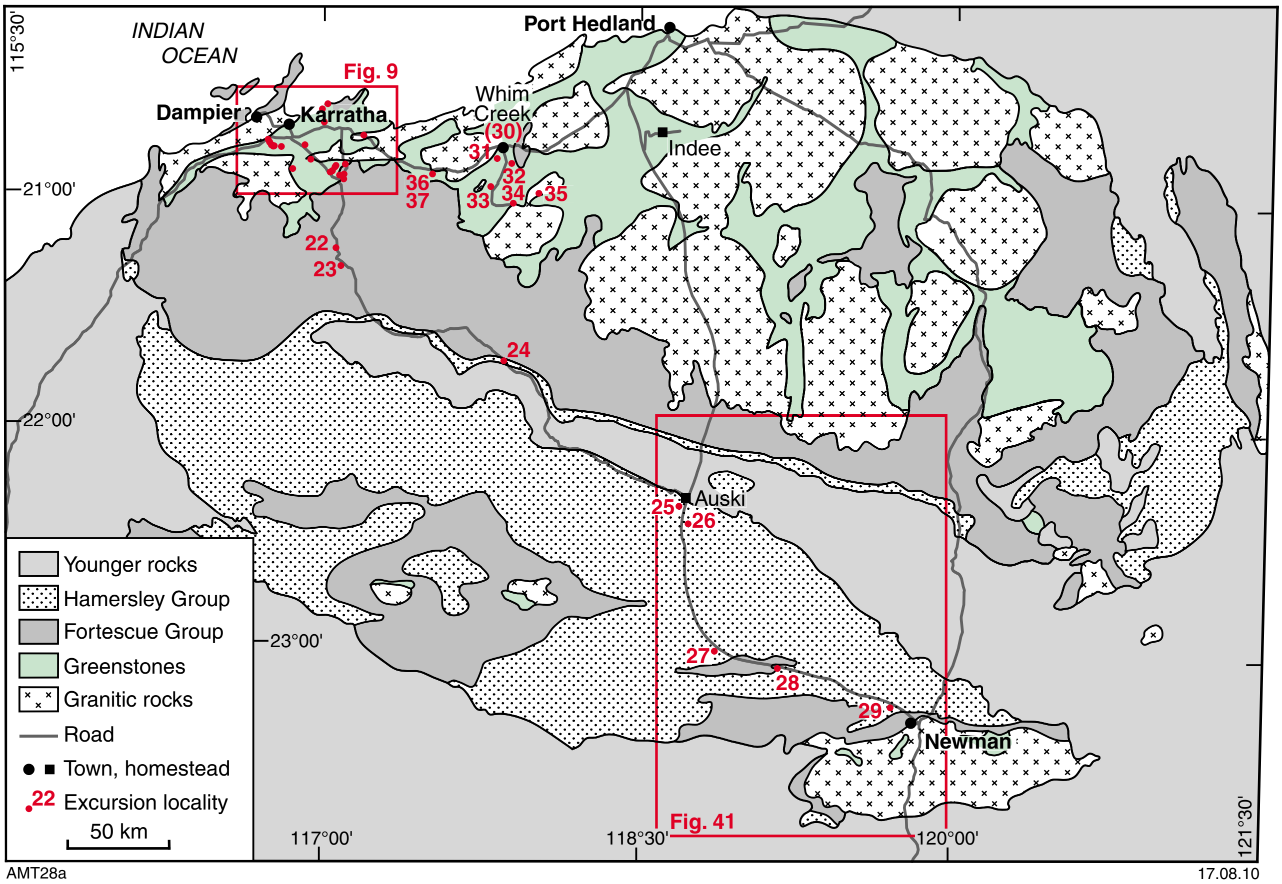

Figure 20. Simplified geological map of the Pilbara Craton, and the Fortescue and Hamersley Basins, showing excursion localities.

Figure 21. Outcrop of the Karratha Granodiorite at Locality 2. The c. 3270 Ma granodiorite contains scattered xenoliths of older granitic rocks (undated), as seen at the top of this boulder.

Figure 22. Geochronology site (GSWA 142433, collected 1996) in the Karratha Granodiorite at Locality 3. The crystallization age (by the SHRIMP U–Pb zircon method) of the sample is 3270 Ma, and the Nd T_{DM} model age is 3480 Ma (data in Smithies et al., 2007a).

Locality 2 provides one of the few large, fresh exposures of the c. 3270 Ma Karratha Granodiorite in the core of the Prinsep Dome. No geochronology has been undertaken at this locality but three other exposures within the dome (including Locality 3) have provided ages between 3270 Ma and 3261 Ma (GSWA 142433: Nelson, 1998f; Smith et al., 1998; Smith, 2003). Additionally, two samples of the same granodiorite exposed on the southern limb of the Roebourne Synform have crystallization ages of 3268 Ma and 3265 Ma (Smith et al., 1998; Smith, 2003). Nd T_{DM} model ages of 3480–3430 Ma (Sun and Hickman, 1998) and 3494–3479 Ma (Smith et al., 1998) indicate that magma generation involved crust of the same age as that which forms the East Pilbara Terrane. As noted in Part 1, the present interpretation is that the Karratha Granodiorite and >3270 Ma Ruth Well Formation originated in the East Pilbara Terrane before the c. 3220 Ma rifting event led to the separate development of the West Pilbara Superterrane.

The rock at Locality 2 is biotite granodiorite containing rare inclusions of mafic rock and black chert, similar to rocks in the Ruth Well Formation (Localities 4 and 6). The granodiorite also contains undated xenoliths of granitic rock of unknown origin (Fig. 21).

Locality 3: Geochronology site in the Karratha Granodiorite (DAMPIER, MGA 477140E 7696350N)

Return to the Tom Price Road, cross the railway line, and turn right on the gravel access road along the western side of the railway. Note that a permit (obtainable from the Karratha Tourist Office) is required to drive on this road. Drive 3 km north, to a point where the fence on the west side of the road ends, and the water pipeline is locally buried (MGA 477400E 7696100N). Turn west along the northern side of a poorly defined track running along an east–west fence. Drive about 250 m west to a 5 m long outcrop of the Karratha Granodiorite (MGA 477140E 7696350N), about 75 m north of the fence.

This locality provides one of the few fresh outcrops of tonalite belonging to the Karratha Granodiorite (Fig. 22). A sample blasted from the exposure in 1996 (GSWA 142433) was dated at 3270 ± 2 Ma using the SHRIMP U–Pb zircon method (GSWA 142433: Nelson, 1998f). This result is consistent with U–Pb data (sample JS17; Smith et al., 1998) from a tonalite 8 km to the east, and indicates that the crystallization age of the Karratha Granodiorite was 3270–3260 Ma. However, sample JS17 also contains xenocrystic zircons dated at about 3300 Ma (Smith et al., 1998), suggesting contamination by older crust. The tonalite locally intrudes metabasalt of the Ruth Well Formation to the southwest. The tonalite is mainly composed of plagioclase and quartz, with minor K-feldspar, biotite, hornblende, epidote, and sericite, and is very weakly foliated (biotite and hornblende). The rock is slightly porphyritic, and contains mafic inclusions, presumably small xenoliths from the adjacent greenstones. These inclusions are aligned to define a weak planar fabric dipping 40° towards 300°, which is parallel to bedding in the Ruth Well Formation.

Locality 4: Metabasalt in the Ruth Well Formation (DAMPIER, MGA 476801E 7696325N)

Continue 400 m west along the track, and then 100 m north to low hills of metabasalt (MGA 476801E 7696325N).

Localities 4 to 11 provide a traverse from the Karratha Granodiorite, through the Ruth Well Formation and the overlying Nickol River Formation, to the base of the Regal Formation. Key features of this traverse are the wide range of lithologies encountered, the pervasive bedding-parallel tectonic foliation S_1 (D_{1WP}, see Part 1), and the presence of several mylonite units along thrust planes. The thrusts dislocate the original stratigraphy of the Nickol River Formation, with the result that few members are continuous laterally for more than 1 km. Such deformation of the Nickol River Formation is not confined to this area, but is present beneath the Regal Thrust throughout the west Pilbara region.

Locality 4 is a brief stop to examine schistose metabasalt in the lower part of the Ruth Well Formation. About 500 m south of Locality 4, the metabasalt is intruded by veins of the Karratha Granodiorite, establishing that its depositional age is greater than 3270 Ma. The maximum depositional age of the Ruth Well Formation is very poorly constrained due to an apparent absence of felsic volcanic rock suitable for U–Pb zircon dating. Nd T_{DM} model ages (Arndt et al., 2001; Smithies et al., 2007a) for three basaltic samples in the area are each approximately 3450 Ma, suggesting sources of similar age to that of the Karratha Granodiorite.

Locality 5: Serpentinite in the Ruth Well Formation (DAMPIER, MGA 476700E 7696240N)

Drive 100 m southwest, across metabasalt and a dolerite dyke, to a small ridge of serpentinite north of the fence (MGA 476700E 7696224N).

The schistose metabasalt of the Ruth Well Formation at Locality 4 is overlain by a 1 m thick grey-white banded chert unit, above which is a 50 m thick serpentinite. This serpentinite is probably a metamorphosed extrusive peridotite, similar to much thicker ultramafic units that outcrop 12 km to the east-southeast at Ruth Well. The Ruth Well succession includes spinifex-textured komatiitic basalt and komatiitic peridotite (Tomich, 1974; Nisbet and Chinner, 1981; Arndt et al., 2001). At Locality 5, textures in the foliated serpentinite are poorly preserved, although some exposures show cumulate olivine pseudomorphs composed of carbonate and chlorite. The serpentinite is overlain by schistose metabasalt, and the underlying chert dips 20° northwest.

Locality 6: Carbonaceous chert in the Ruth Well Formation (Dampier, MGA 476560E 7696400N)

Drive 200 m to the northwest, crossing a fault zone marked by quartz veining, to outcrops of dark grey chert (MGA 476560E 7696400N).

Locality 6 provides a good outcrop of interbedded grey and white banded chert in the Ruth Well Formation (Fig. 23). The chert dips 30° west-northwest, and bedding planes show stretching lineations plunging 25° northwest. Black and dark grey chert units are relatively common in the Ruth Well Formation around the Prinsep Dome, and are interpreted to represent silicified carbonaceous shale units within the mafic and ultramafic volcanic rocks. The chert is overlain by metadolerite and approximately 150 m of metamorphosed tholeiitic basalt, interlayered with units of talc–chlorite schist probably derived from komatiitic basalt.

Locality 7: Mafic mylonite in the upper part of the Ruth Well Formation (Dampier, MGA 476190E 7696690N)

Drive 500 m to the northwest, first across the metabasalt succession followed by poorly exposed talc–chlorite and talc–carbonate schist. Locality 7 is located on outcrops of mafic mylonite that form a low north-northeasterly trending ridge (MGA 476190E 7696690N).

The ridge at Locality 7 is underlain by strongly sheared ultramafic and mafic rocks that include a 3–5 m thick unit of mylonite (Fig. 24a). In some outcrops, the mylonite appears relatively felsic, but this appearance is probably due to silicification. It is evident that the mylonite marks a major fault because of the abrupt change of strike from the underlying succession (strike north-northwest) to the overlying succession (strike north-northeast). Furthermore, geological mapping (Hickman, 1997b) shows that the underlying succession between Localities 6 and 7 is progressively tectonically removed northwards along the mylonite unit. The mylonite dips 60° west-northwest, and marks a thrust in the upper part of the Ruth Well Formation. Stretching lineations on the mylonite foliation plunge 45° northwest (Fig. 24b).

Figure 23. Grey and white banded chert in the Ruth Well Formation at Locality 6. Dark grey chert is interpreted to represent silicified carbonaceous shale, and the white chert includes veins of microcrystalline quartz.

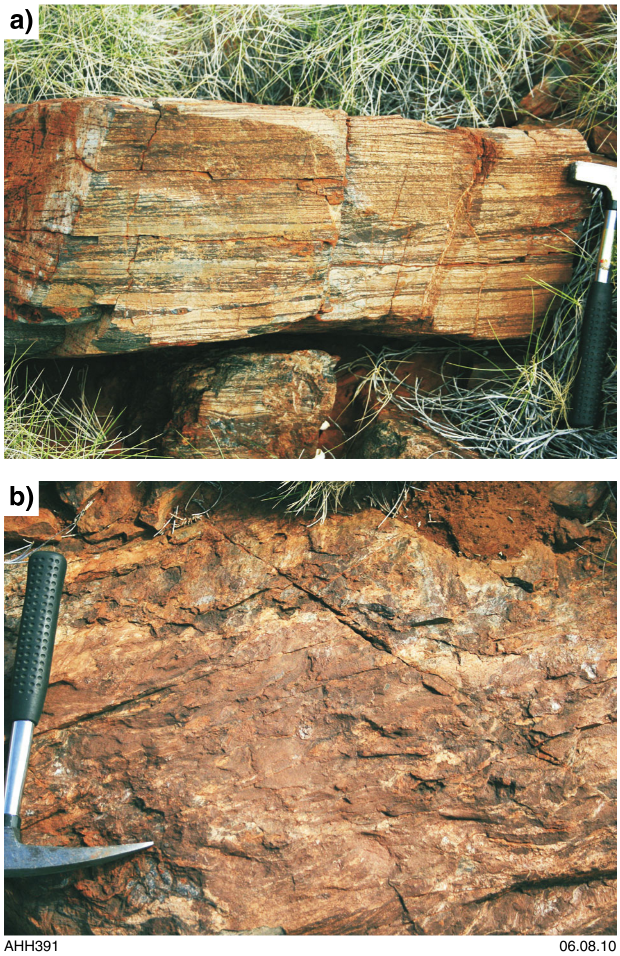

Figure 24. Mafic mylonite in the thrust at Locality 7: a) alternating mafic and felsic layers in the mylonite represent variably silicified sheared metabasalt, and possibly equally deformed more felsic intrusive rock. The foliation in the mylonite is parallel to the schistose foliation (S_1) in overlying metabasalt of the Ruth Well Formation, but strongly discordant to the stratigraphy of underlying metabasalts in the same formation; b) lineations on S_1 plunge 45° northwest.

Locality 8: Fuchsitic quartzite at the base of the Nickol River Formation (Dampier, MGA 475904E 7696693N)

Drive 300 m to the west across a valley underlain by metabasalt, then up a gradual slope to outcrops of quartzite near the top of a low ridge (MGA 475904E 7696693N).

Locality 8 provides outcrops of the basal section of the Nickol River Formation. The main rock types here are flaggy (sheared) quartzite and siliceous schist, with units of green (possibly fuchsitic) quartzite, and thin layers of black chert. Isolated small oval aggregates of chert in the quartzite may represent deformed pebbles. North of Locality 8, the quartzite is separated from the underlying metabasalt of the Ruth Well Formation by flaggy black chert and highly siliceous green mylonite, revealing another tectonic contact. About 1.5 km southeast of Mount Regal, this tectonic contact is folded by tight to isoclinal folds visible at 1:100 000 scale (Hickman, 1997b). These folds were probably produced during a late phase of the D_1 deformation, and their geometry at 1:100 000 scale suggests overthrusting from the north or northwest.

Locality 9: Mylonitic quartzite and black chert in the Nickol River Formation (Dampier, MGA 475760E 7696785N)

Drive 200 m west across the ridge, through a small valley, and up an east-facing slope to outcrops of quartzite at the top of a second ridge (MGA 475760E 7696785N).

The mylonitic quartzite at Locality 9 provides the largest outcrops of the lower part of the Nickol River Formation near Mount Regal. The tectonic foliation dips 30° northwest and stretching lineations on the foliation plunge 30° northwest. About 80 m to the west-southwest, the quartzite is overlain by black chert, which 100 m farther west is itself overlain by ferruginous chert and schistose carbonate rocks. The ferruginous chert is interpreted to represent a silicified succession of pyritic shale interbedded with BIF. The frequency of carbonate units increases westwards up the succession, and some are being quarried for aggregate 700 m west-northwest of Locality 9.

Locality 10: Recumbently folded carbonate schists in the Nickol River Formation, east of Mount Regal (Dampier, MGA 474930E 7697960N)

Return to the Hamersley Iron Railway access road, drive north to Highway 1, turn left, drive west 500 m, and then turn left onto a good quality gravel road (MGA 475525E 7700750N). Drive south 3.2 km to a gate at the entrance of a mining area (MGA 474840E 769800N). Walk across the creek to the east, and climb approximately 25 m up the side of the hill to outcrops of carbonate rocks at Locality 10 (MGA 474930E 7697960N).

The upper part of the Nickol River Formation at Mount Regal is composed of a strongly deformed carbonate-shale unit that is partly tectonically interleaved with mafic and ultramafic schist. At Locality 10, approximately 150 m vertically beneath the overlying Regal Formation, the rocks consist of carbonate-rich schist showing numerous intrafolial recumbent folds (Fig. 25). The present lithologies in these outcrops indicate a metamorphosed

Figure 25. Folded carbonate-rich metasedimentary rocks in the Nickol River Formation at Locality 10. Recumbent folds visible in this outcrop close about east–west to northeast–southwest trending axes. Other lithologies include ferruginous chert and metamorphosed shale.

and strongly deformed primary association of carbonate sediments, shales, and ferruginous chert. The succession was extensively sheared during overthrusting of the Regal Formation, and the resulting tectonic foliation was subsequently deformed by recumbent folding. Closures of the recumbent folds are only clearly visible on the eastern and southeastern sides of the outcrops, indicating east–west trending axes and north–south compression.

Locality 11: Terrane boundary: mylonite beneath the Regal Formation southeast of Mount Regal (Dampier, MGA 474800E 7698290N)

Return to the road, and walk or drive approximately 300 m north-northwest to Locality 11 (MGA 474800E 7698290N), located just beneath the cliffs of metabasalt on the southeastern slopes of Mount Regal.

The first outcrops to be visited at Locality 11 are situated approximately 50 m from the base of the Regal Formation, and consist of finely laminated mylonite (Fig. 26a,b). This mylonite is located within the upper part of the Nickol River Formation, which is locally composed of ferruginous chert and metamorphosed carbonate rocks. However, at this locality the upper Nickol River Formation is essentially a tectonic assemblage that also contains slices of mafic and ultramafic schist derived from the overlying Regal Formation, and accordingly, the composition of the mylonite varies along strike. The mylonite is immediately overlain by ferruginous chert and schistose metasedimentary rocks, which are in turn overlain by sheared ultramafic rocks that locally contain minor chrysotile. Outcrops at the foot of the cliffs are composed of sheared metabasalt (Regal Formation) in the Regal Terrane.

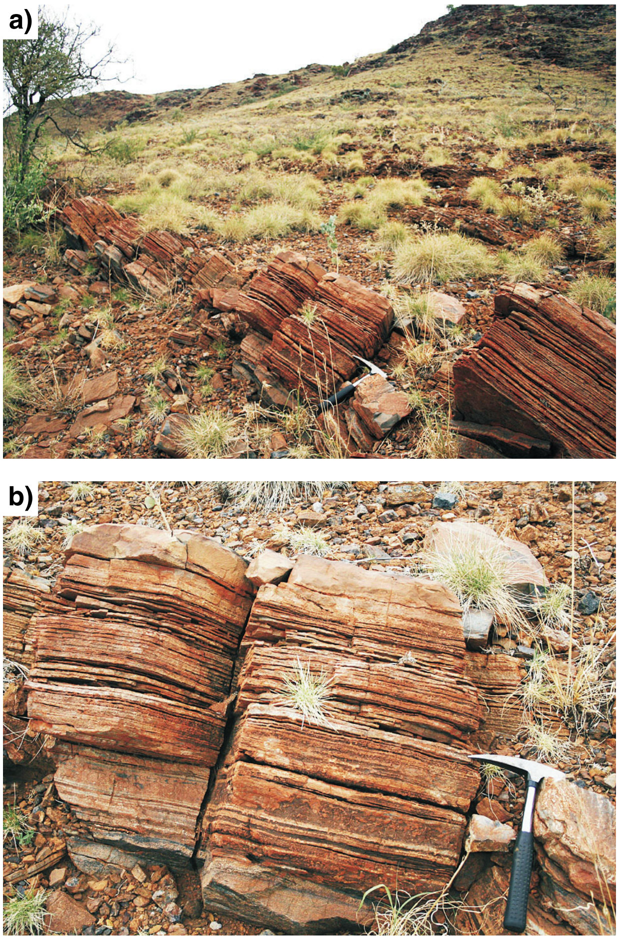

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Figure 26. Outcrop of mylonite in the Regal Thrust at Locality 11: a) mylonite dips northwestwards beneath the metabasalts of the Regal Formation (top right), although there are intervening metasedimentary rocks of the Nickol River Formation, including banded iron formation; b) close-up view of the mylonite showing its strongly planar tectonic foliation.

Locality 12: Siliceous mylonite in the Nickol River Formation, north of Mount Regal (Dampier, MGA 474130E 7699525N)

Return to the road, and drive approximately 1.2 km north to a well-used, westward-leading track (MGA 475120E 7699200N). Drive 1.1 km west on this track, avoiding a track to the south that leads up the northern side of Mount Regal. Close to the end of the track (MGA 473900E 7699005N) is a less distinct track leading north-northeast. Follow this track for 600 m to Locality 12 (MGA 474130E 7699525N).

Locality 12 provides good exposures of a siliceous mylonite (Fig. 27) about 100 below the top of the Nickol River Formation, and provides additional evidence of the tectonic complexity of the Nickol River Formation in the Regal Thrust zone. The protolith of the mylonite could have been chert, a quartzite, or a felsic intrusive rock. Parts of the mylonite are pale green, which is a feature of mylonites developed close to ultramafic rocks seen at various localities around the Prinsep Dome. Away from the mylonite, the Nickol River Formation is locally composed of schistose carbonate rocks, ferruginous chert, and BIF. There are also tectonic slices of mafic and ultramafic rock and veins of monzogranite, the latter probably originating from c. 2990 Ma intrusions of the Dampier Granitic Complex to the north.

DAY 2.
The Sholl Terrane: 3130–3115 Ma intraoceanic arc volcanics (Whundo Group) accreted against the Karratha and Regal Terranes along the Sholl Shear Zone

On the morning of the second day, the excursion visits key outcrops in the upper part of the Nickol River Formation on the southern and northeastern sides of the Prinsep Dome. Following an examination of the Sholl Shear Zone at Nickol River, the rest of the day focuses on the 3130–3115 Ma Whundo Group of the Sholl Terrane, and the geochemical evidence this group provides for early Mesoarchean subduction of oceanic crust.

Lunch will be taken at Locality 15.

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Figure 27. Siliceous mylonite beneath the main plane of the Regal Thrust at Locality 12. The protolith of the mylonite is unknown, but its composition is consistent with either chert, vein quartz, or quartz-rich sandstone. Parts of the mylonite are pale green, suggesting minor disseminated chromium muscovite (fuchsite). Green chert is commonly located adjacent to ultramafic rocks in Pilbara greenstone successions.

Locality 13: Mylonite at the Regal Thrust, southeast of the Karratha Granodiorite (Dampier, MGA 492000E 7696800N)

From Karratha, drive to Highway 1. Turn right for about 50 m, before turning sharp left onto a road that follows a water pipeline southeastwards. Follow this road for 7 km (MGA 491200E 7696800N). Leave the track about 100 m southeast of the creek, and drive east about 600 m to Locality 13.

Locality 13 provides excellent exposures of the Regal Thrust on the southern limb of the Prinsep Dome. Close examination of the chert-like rock forming the low cliffs at Locality 13 reveals a finely laminated siliceous mylonite with ubiquitous internal isoclinal folding (Fig. 28a). In addition to intrafolial isoclines, the mylonite contains refolded isoclines (Fig. 28b) and sheath folds (Fig. 28c), and it is strongly lineated in all outcrops (plunge of lineations 0–30°, generally to the west). The prevailing dip of the mylonite is 60–80° south.

Underlying the mylonite to the north is a mixed lithological assemblage of strongly sheared quartzofeldspathic sandstone, chert, quartzite, carbonate rocks, and ultramafic schist belonging to the Nickol River Formation. About 10 km northeast of Locality 13, strongly sheared quartzofeldspathic sandstone of the Nickol River Formation was dated by the SHRIMP zircon U–Pb method (GSWA 136819: Nelson, 1998b) at 3269 ±2 Ma, a date consistent with the derivation of detritus from the underlying 3270 Ma Karratha Granodiorite. Above the mylonite, the Regal Formation is composed of strongly foliated amphibole–chlorite schist.

The mylonite at Locality 13 represents the upper part of the tectonic zone referred to as the Regal Thrust. The Regal Thrust is folded around the 2950 Ma Prinsep Dome and Roebourne Synform, and as a result, its total outcrop length in the Karratha–Roebourne area exceeds 100 km (Figs 2 and 9).

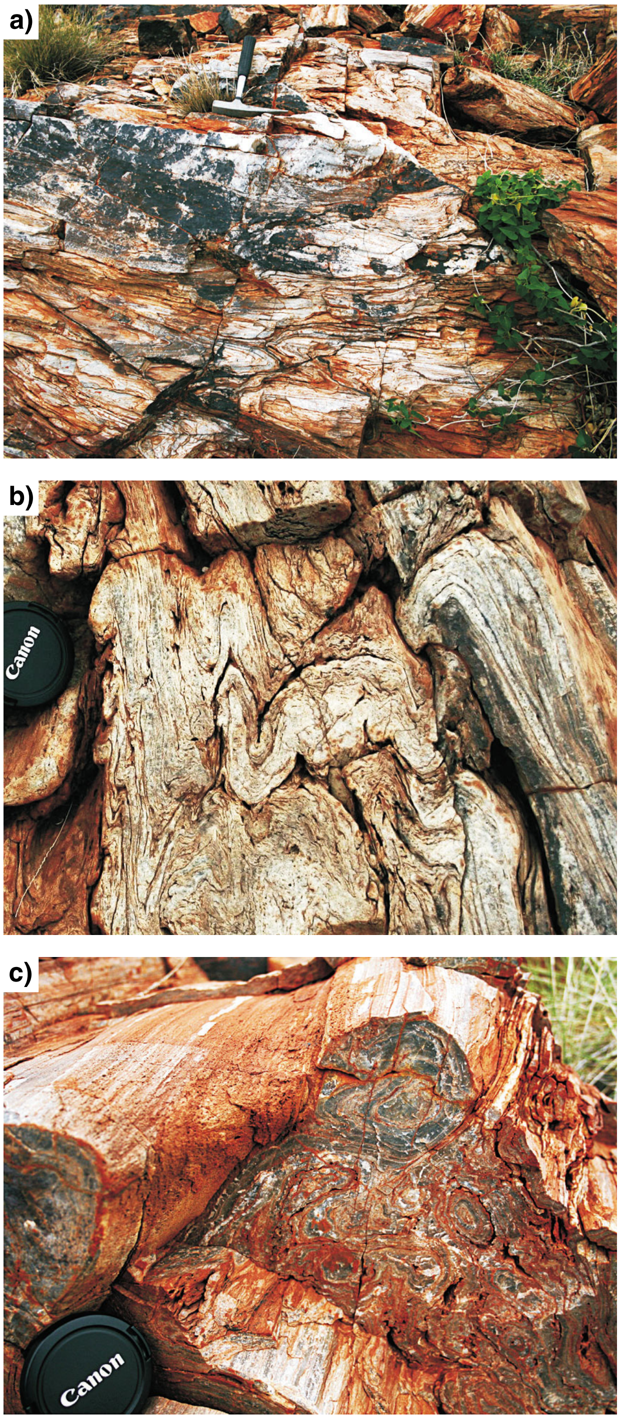

Figure 28. Structures in mylonite in the Regal Thrust at Locality 13: a) finely laminated silicic mylonite with ubiquitous internal isoclinal folding; b) close-up of a refolded isocline; c) sheath folds with a parallel mineral lineation.

Locality 14: Stretched pebbles in conglomerate of the Nickol River Formation (Roebourne, MGA 501505E 7707865)

Return to Highway 1, turn right, and drive 20 km east to the turn-off signposted to Cleaverville Beach (MGA 503700E 7705300). Drive 3.8 km north to a track that leads to the west (MGA 501790E 7707865N). Follow this track west for 750 m, then turn left onto a minor track (MGA 501240E 7707815N). Follow this track south for 150 m (MGA 501340E 7707720N). Leave the track and drive or walk 200 m east-northeast up a gradual slope to outcrops at Locality 14 (MGA 501505E 7707865N).

At Locality 14, a metamorphosed and tectonically attenuated pebbly sandstone of the Nickol River Formation dips 30º east-northeast on the northeastern margin of the Prinsep Dome. Numerous clasts of chert in the sandstone are acutely stretched in a northeasterly direction (Fig. 29a). A few larger chert pebbles up to 10 cm in diameter are less deformed (Fig. 29b). The sandstone is situated about 100 m beneath the uppermost fault plane of the Regal Thrust, which in this area is folded around the locally northeasterly plunging Prinsep Dome. The Prinsep Dome also folds the tectonic foliation of the metasandstone, and a parallel tectonic foliation in the metabasalt of the overlying Regal Formation. Further evidence constraining the age of the foliation lies in that within the Regal Formation this foliation is intruded by c. 3015 Ma dacite sills of the Orpheus Supersuite. Therefore, the foliation was a product of the thrusting event D_1 (see Part 1). To the southwest, towards the core of the Prinsep Dome, the sandstone is underlain by sheared clastic metasediments that overlie ultramafic rocks and black chert in the Ruth Well Formation.

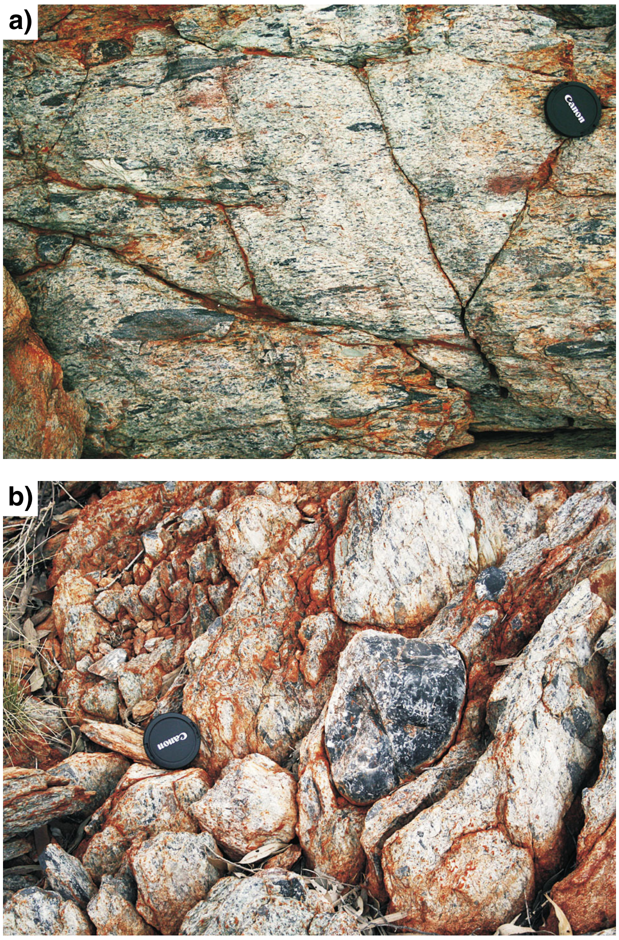

Figure 29. Sheared conglomerate in the Nickol River Formation at Locality 14: a) intensely stretched clasts of grey chert and politic schist in a mylonitized coarse lithic sandstone matrix; b) a small cobble of grey chert, largely undeformed within a strongly sheared and lineated metasandstone matrix.

Locality 15: Sholl Shear Zone between the Karratha and Sholl Terranes, Nickol River (Dampier, MGA 494950E 7689700N)

Return to Highway 1, drive westwards to the Karratha turnoff, and take the water pipeline track southeast for 11 km to the crossroad intersection with the Roebourne–Cherratta road (MGA 499700E 7689200N). Turn right, and follow the Roebourne–Cherratta road west for about 4 km to a four-wheel drive track (MGA 495820E 7688993N) about 200 m before the crossing over Nickol River. Take the track to the north, drive north 1 km (MGA 495250E 7689860N), then turn west onto an indistinct track leading to rock pavements in the bed of Nickol River. Park at the end of the track (MGA 495060E 7689760N).

Rock pavements in the Nickol River provide excellent exposures (Fig. 30a) of the northern section of the Sholl Shear Zone. The mylonite is dominantly felsic, and represents extremely sheared granitic rocks, but there are also layers of amphibolite. The mylonitic lamination is folded by isoclinal folds, and tight, west-plunging Z-folds that may be related to late dextral movement (Fig. 30b). All of these structures are displaced by late brittle fractures, which are locally filled by pseudotachylite. These fractures are probably related to a post- lower Fortescue Group (post-2725 Ma) north–south compressional event, which produced a conjugate fault system in the Dampier–Roebourne area (Hickman, 2001b).

Smith et al. (1998) reported a zircon age of 3024 ± 4 Ma from the mylonite, which is interpreted as the age of the local granitic precursor. Shear-sense indictors elsewhere in the area (Hickman, 2001b) indicate that the dominant early shearing was sinistral, and may have been partly contemporaneous with the Orpheus Supersuite that intrudes the Sholl Shear Zone and outcrops sporadically for several kilometres to the north. Late movement (the post-3.01 Ga Whim Creek Group) was clearly dextral because the Whim Creek Group and Caines Well Granitoid Complex are displaced about 30–40 km across the shear zone. Metamorphic grades immediately north of the shear zone range from upper greenschist to amphibolite facies, whereas grades to the south are of greenschist facies. However, the timing of the implied north-side-up movement, and the amount of early sinistral movement are both unclear. This is largely due to the fact that in the Dampier–Roebourne area, the Maitland River Supersuite intruded large areas of greenstones both north and south of the shear zone, and these granitic intrusions appear to have totally obscured pre-3010 Ma displacements of stratigraphy.

Sm–Nd data from the Sholl Terrane and the Cherratta Granitic Complex to the south (Sun and Hickman, 1998), and the Karratha Terrane to the north provides evidence that early movement along the Sholl Shear Zone was much greater than the 30–40 km late dextral movement. Sm–Nd data south of the Sholl Shear Zone show no evidence of underlying c. 3480 Ma crust, whereas Sm–Nd data north of the Sholl Shear Zone do. This difference between the ages of underlying crust on either side of the Sholl Shear Zone supports evidence from stratigraphy, and the width of the mylonite zone, that early sinistral movement was probably at least 200 km.

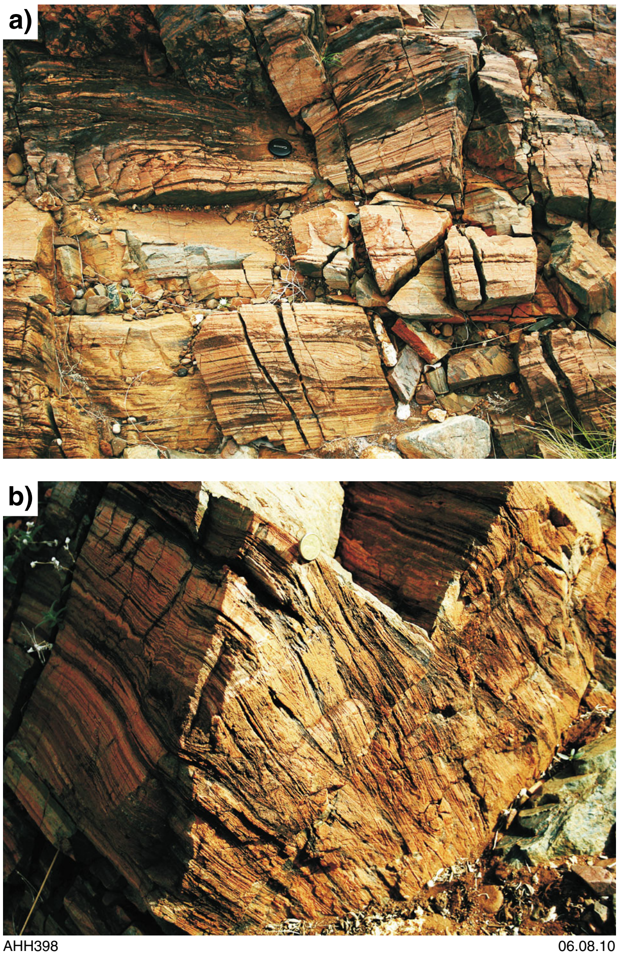

Figure 30. Mylonite in the Sholl Shear Zone at Nickol River, Locality 15: a) felsic mylonite containing darker mafic material, and locally tightly to isoclinally folded (left and right of camera lens cap); b) isoclinal folding of mylonite fabric (coin 2 cm diameter).

Localities 16–23: Whundo Group

The remainder of day two is devoted to the Whundo Group. As noted previously, this group can be subdivided geochemically into three packages, which are broadly similar to the previously defined lithological map divisions (formations). The aim of the following excursion stops is to provide a traverse through as much of this volcanic package as reasonable access conditions and time permit. The route traces previous geochemical sampling traverses (Smithies et al., 2005a) through the least deformed and least altered parts of the Whundo Group.

Locality 16: Lower Volcanic Package — first calc-alkaline basalt and boninites (Dampier, MGA 486086E 7685229N (basalt) and 485258E 7685226N (boninite))

From Locality 15, return to the Roebourne–Cherratta road and turn right. Drive 11 km west to Locality 16 (MGA 486086E 7685229N).

The lower volcanic package is dominated by calc-alkaline basaltic to andesitic lavas (CA1), but also includes ~1–10 m thick flow units with boninitic compositions (Fig. 31), which are interbedded throughout; these bininites form up to ~15% of the overall package. This package forms the most deformed and metamorphosed part of the Whundo Group. Large-scale depositional features such as hyaloclastite layers, pillows, and flow-top breccias are only locally preserved. These locally schistose rocks comprise a greenschist-facies assemblage of quartz–actinolite–chlorite–epidote–plagioclase. Little of the primary mineralogy is preserved, although primary textures are commonly well preserved; later alteration resulted in locally extensive development of a quartz–carbonate–sericite assemblage. Locally preserved original igneous textures show the boninitic lavas to have been vesicular, extremely fine grained, glass-rich rocks containing phenocrysts of subhedral olivine (now chlorite and serpentine) and acicular pyroxene (now actinolite).

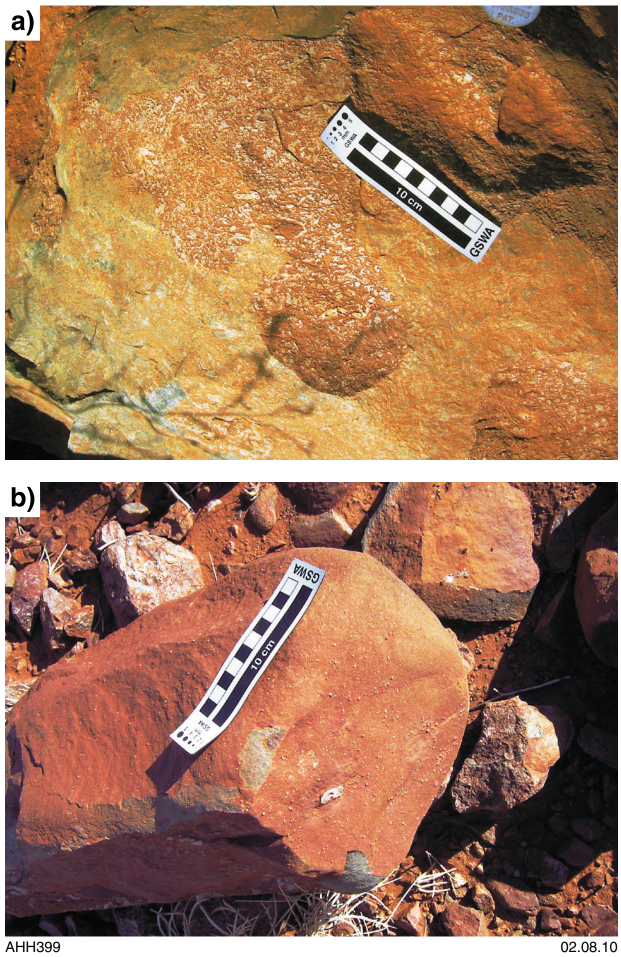

Figure 31. Mafic units in the Lower Volcanic Package of the Whundo Group, showing: a) hyaloclastite breccia in boninite, and b) weakly amygdaloidal boninite flows.

Locality 17: Middle Volcanic Package — lower tholeiitic pillow basalts, Harding Dam (ROEBOURNE, MGA 508900E 7682000N)

Drive approximately 15 km east on the Roebourne–Cherratta road to the water-pipeline road, turn right, and continue southeast for 14.5 km, to where the road bends sharply to the left. A few hundred metres past the bend, a disused track leads into a gravel pit (MGA 508900E 7681850N). Locality 17 is situated on the western side of the small hill 150 m to the north of the gravel pit.

This locality provides excellent three-dimensional exposures of pillow basalt (Fig. 32a,b) in the middle volcanic package (Bradley Basalt) of the Whundo Group. Pillow structures show that the lava flows, which dip 55° towards the northeast, are the right way up.

Tholeiitic lavas (T1 of Smithies et al., 2005a) dominate the middle volcanic package (Fig. 10), although rare flows of boninite-like lavas persist into the lower parts of this package, and rocks compositionally transitional between these and overlying calc-alkaline lavas are found in the upper parts of the section. Large-scale features (pillow structures, hyaloclastite, vesicle-rich layers, and flow top breccias) are common, as are fine-scale igneous textures; however, the original mineralogy shows a moderate to high level of replacement, and fine calcite veining is a locally prominent feature. The tholeiitic rocks are plagioclase-phyric, although the euhedral laths (up to 2 mm long) are typically replaced by calcite, sericite, and epidote. Groundmass plagioclase is better preserved, and locally has only a moderate dusting of alteration minerals. Original mafic minerals — dominantly clinopyroxene, both as phenocrysts and a groundmass phase — have been largely replaced by chlorite and lesser actinolite. The tholeiites contain rare clots of serpentine, possibly replacing olivine, and abundant leucoxene replacing magnetite.

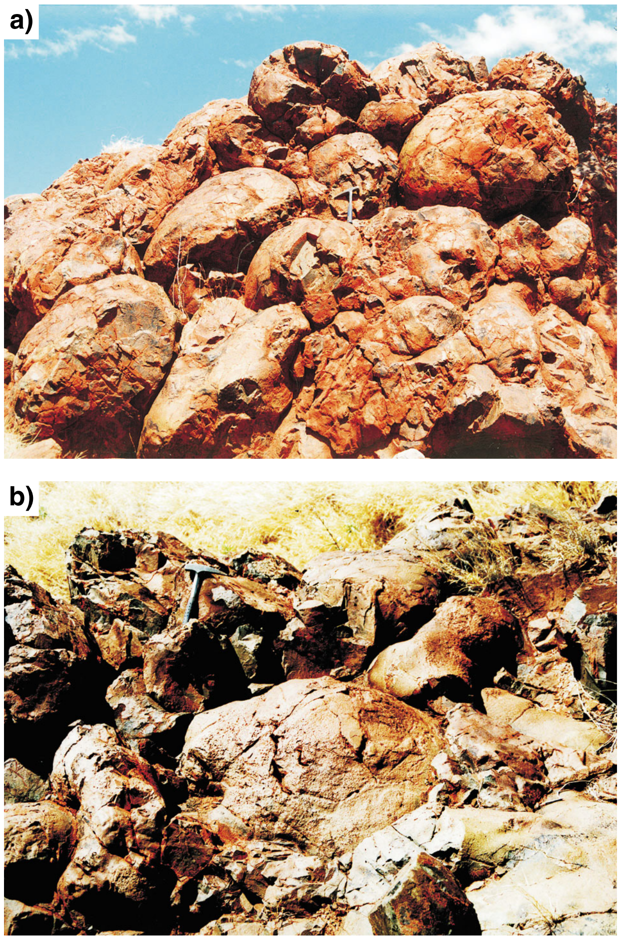

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Figure 32. Pillow lava in the Bradley Basalt 2.5 km northwest of Harding Dam: a) pillow structures, 1.5 to 2 m in diameter, are exceptionally well exposed on the side of the small hill at Locality 17; b) pillow structures viewed from above, showing irregular flow morphology and vesicular tops.

Locality 18: Traverse through the Upper Volcanic Package (begins at ROEBOURNE, MGA 504368E 7683505N)

Return to the water-pipeline road, turn right, and continue west for 6 km to Locality 18, located at the base of a hill 100 m south of the road (MGA 504368E 7683505N). This is the starting point of a 4.3 km traverse that includes five outcrops.

The upper volcanic package contains abundant calc-alkaline basaltic to dacitic flows of vesicular pillowed units, intercalated with hyaloclastite, flow-top breccias, and reworked volcaniclastic deposits. Large-scale features are extremely well preserved, as are fine-scale igneous textures, and primary mineralogy is also locally well preserved. Subhedral to euhedral clinopyroxene phenocrysts, up to 2 mm in size, are partially altered to chlorite and epidote. Plagioclase, both as phenocrysts and as a groundmass phase, is weakly to moderately replaced by calcite, sericite, and epidote. Quartz is a rare interstitial phase. The area from which these calc-alkaline rocks were sampled is one of the few large outcrops where deformation and overprinting alteration is not significant.

Felsic rocks clearly of a volcanic origin first appear within the stratigraphically lower levels of the middle volcanic package, and include laminated ash deposits, pumiceous breccia, and porphyritic quartz–plagioclase flows. The proportion of felsic volcanic and volcaniclastic rocks increases upwards within the upper volcanic package, and they dominate the poorly preserved uppermost part of that package. Near the top of the upper volcanic package, felsic volcanic and volcaniclastic rocks are interbedded with minor flows of Nb-rich basalt and LREE-rich dacite.

A) MGA 504368E 7683505N: at this locality we see an example of reasonably fresh basalt belonging to the CA2 calc-alkaline basalt unit of Smithies et al. (2005a).

Return to the water-pipeline road and drive 650 m east to a bend where a powerline crosses the road. Turn north on an old track (four-wheel drive) that follows the powerline, and continue northeast for 450 m to outcrop B.

B) MGA 505236E 7684022N: here we see one of the lower felsic volcanic and volcaniclastic layers within the Upper Volcanic Package. These include laminated ash deposits (Fig. 33a) and fragmental coarse-grained volcaniclastic sandstone (Fig. 33b,c), and show possible rheomorphic structures and sharp contacts with vesicular CA2 basalts (Fig. 33d).

Continue northeast on the same track for 250 m to outcrop C.

c) MGA 505426E 7684197N: here we see another felsic volcaniclastic unit showing cyclic grain size grading (Fig. 34a,b), well developed flame structures and slump folding (Fig. 34c), and fine-scale cross bedding (Fig. 34d).

Continue northeast on the track for 2.7 km to outcrop D.

D) MGA 506684E 7686113N: here we can observe an outcrop of the upper (adakitic) calc-alkaline unit (CA3).

Drive northeast on the track for 850 m to outcrop E.

E) MGA 507144E 7686713N: here we find an exposure of Nb-enriched basalt. The outcrops at D and E show that, within the context of the Whundo Group in general, the adakitic and Nb-enriched rocks are particularly indistinct in outcrop (Fig. 35a, CA3; Fig. 35b, Nb-enriched); they can only be recognized on the basis of geochemistry.

Return to Karratha by driving back to the water pipeline road and following that northwest to Highway 1.

DAY 3. Fortescue Basin

Days 3 and 4 of the excursion provide a traverse from the basal unconformity of the Neoarchean Fortescue Basin, to the eastern part of the Neoarchean–Paleoproterozoic Hamersley Basin. Localities along this 500 km traverse reveal evidence for changing tectonic settings, from a stable craton between 2900 and 2800 Ma, to rifting at 2780 Ma, through passive-margin and increasingly deep water deposition between 2630 and 2400 Ma, to continental collision and the Ophthalmian Orogeny at c. 2200 Ma.

Lunch will be taken at Locality 22.

Note that the availability of fuel and water is very limited between Roebourne and the Auski Roadhouse at Munjina, a distance of approximately 310 km.

Figure 33. Felsic volcanic and volcaniclastic units in the Upper Volcanic Package of the Whundo Group: a) Laminated ash deposits; b) coarse-grained volcaniclastic sandstone; c), fragmental volcaniclastic rocks; d), sharp contacts with vesicular CA2 lavas

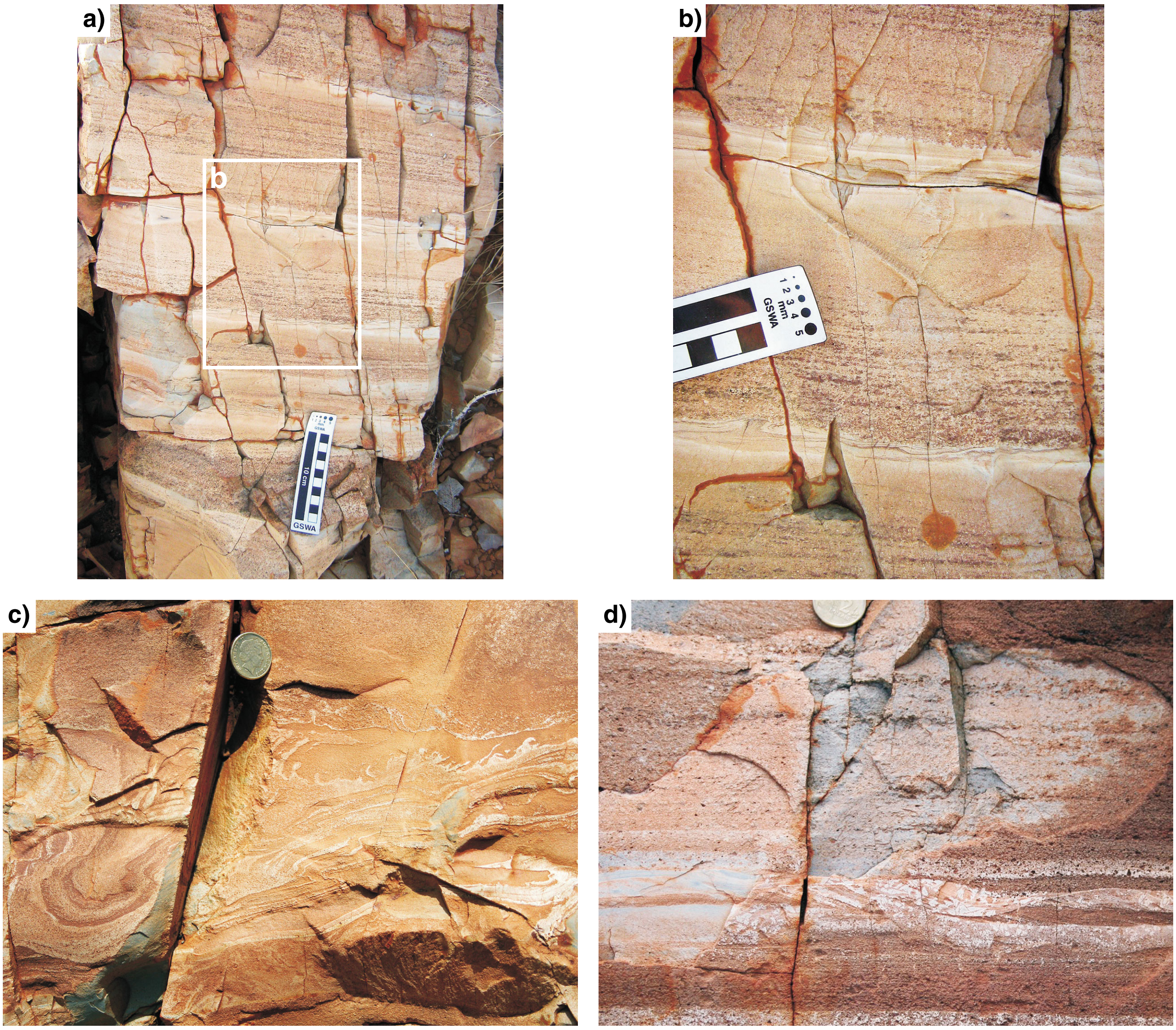

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Figure 34. **Sedimentary structures in felsic volcaniclastic unit of the Bradley Basalt at Locality 18C: a, b) cyclic grainsize garding; c) well developed flame structures (centre and right), and syndepositional slump folding (left), coin 2 cm diameter; d) fine-scale cross bedding in reworked felsic tuffaceous sediment, coin 2 cm diameter.**

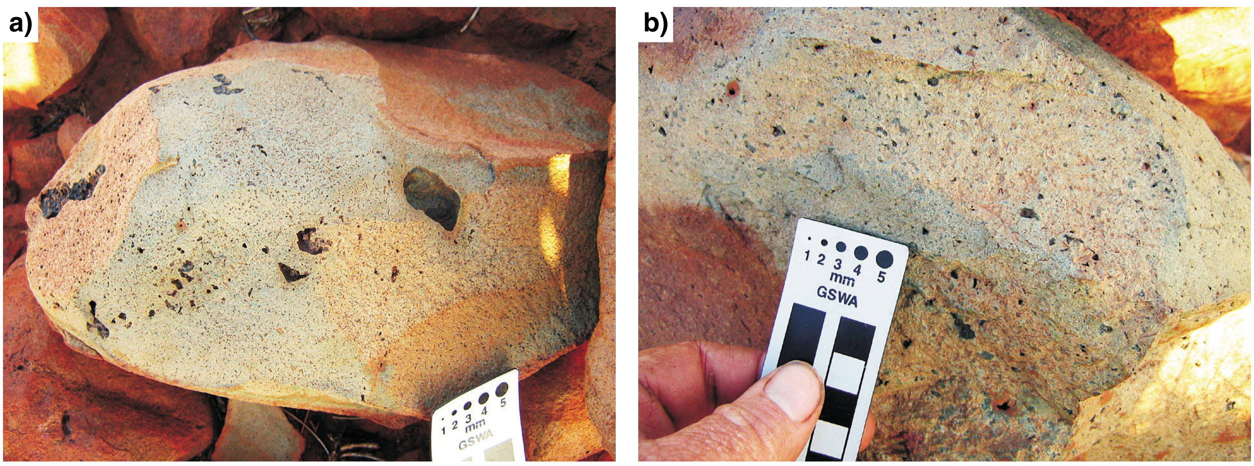

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Figure 35. CA3 lavas (a) and Nb-enriched basalts (b) in the Upper Volcanic Package of the Whundo Group

Locality 19: Mount Roe Basalt and the basal unconformity of the Fortescue Group near Mount Roe (ROEBOURNE, MGA 511705E 7687370N)

From Karratha, drive 36 km to Roebourne, and at the southern end of the town take the road signed for Harding Dam. Approximately 17 km south of Roebourne, the Harding Dam road meets a service road along the Robe River Railway (iron ore). Approximately 150 m north of the road junction, a minor track (MGA 511500E 7687420N) leads to the east. Follow this track for 220 m to the water pipe, and walk 50 m south along an old track to the base of a small hill.

Locality 19 provides exposures of highly amygdaloidal basalt flows (Fig. 36) a few metres above the basal unconformity of the Fortescue Group. The amygdales are up to 10 cm across, and in outcrop are partly filled by calcite, quartz, and chlorite. At this locality, the amygdales are elongated east–west, indicating the local flow direction.

From the top of the hill, the basal unconformity of the Mount Roe Basalt is clearly visible to the west-southwest (Fig. 37); on the hill tops the flows form dip slopes, with the formation inclined approximately 20° to the southeast. In this area, the basal Fortescue Group was unconformably deposited on an irregular paleosurface composed of Whundo Group, Cleaverville Formation, mylonite of the Sholl Shear Zone, and the Whim Creek Group. Kriewaldt (1964) recorded that the Mount Roe Basalt is 2.4 km thick at Mount Roe, yet the formation is less than 50 m thick only 5 km to the southeast and 6 km to the south-southwest, and in places the stratigraphically overlying Hardey Formation rests unconformably on the Whundo and Whim Creek Groups. Although some degree of thickness variation can be attributed to the distribution of paleoridges and -valleys on the c. 2780 Ma land surface, the extremely abrupt thickness changes in some areas must coincide with horst and graben structures formed by the rifting that accompanied volcanism (Blake, 1993). In the west Pilbara, the Fortescue Group was deposited in the Northwest Pilbara Sub-basin (Blake, 1984), where crustal extension was approximately east–west (Thorne and Trendall, 2001). Thorne and Trendall (2001) attributed the intracratonic rifting and volcanism of the lower Fortescue Group to lithospheric stretching of the Pilbara Craton, which subsequently led to crustal separation and ocean basin formation.

Locality 20: Conglomerate of the Hardey Formation near Harding Dam (ROEBOURNE, MGA 511040E 7682600N)

Return to the Harding Dam road and drive 7 km south to MGA 510760E 7682850N. Locality 20 is located on the west bank of the Harding River, 380 m to the southeast of the railway and at the southern end of Pinanular Pool (MGA 511040E 7682600N).

Locality 20 is an excellent exposure of polymictic conglomerate within the lower part of the Hardey Formation (Fig. 38). The pebbles and boulders in the conglomerate beds are dominantly composed of vesicular and non-vesicular basalt derived from the underlying Mount Roe Basalt. However, clasts of granite, gneissic granite, chert, and vein quartz are also present, establishing that the source of the beds was not entirely local. The matrix is composed of coarse sandstone, with many pebbles being matrix-supported. Stratigraphically above the conglomerate, on the eastern bank of the river, there are larger exposures of upward-fining lithic sandstone, probably deposited in an alluvial fan setting.

The significance of this conglomerate is that it establishes that the Mount Roe Basalt was eroded before, or during,

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Figure 36. The Mount Roe Basalt near Mount Roe, Locality 19. Vesicular lava flows near the base of the formation.

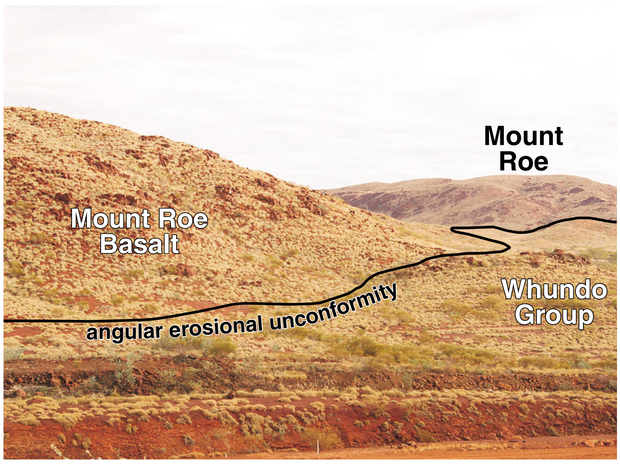

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Figure 37. View of the basal unconformity of the Fortescue Group near Mount Roe. The Mount Roe Basalt overlies the Woodbrook Formation (Whundo Group) above a high-angle erosional unconformity (shown by dashed line).

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Figure 38. Basal conglomerate of the Hardey Formation near Harding Dam, Locality 20. Polymictic conglomerate unconformably overlies the Mount Roe Basalt. Most of the boulders and pebbles in the conglomerate are vesicular or porphyritic basalt typical of lava flows in the Mount Roe Basalt; other rock types, including granite, are also present. Scale, hammer (top centre).

deposition of the Hardey Formation. The Hardey Formation unconformably overlies the Mount Roe Basalt in many areas of the Fortescue Basin, and belongs to the second depositional sequence of the Fortescue Group (see Part 1). Like the Mount Roe Basalt, the thickness of the Hardey Formation is extremely variable in the west Pilbara, and deposition was concentrated in local rift basins.

Locality 21: Cooya Pooya Dolerite at Harding Dam (Roebourne, MGA 510850E 7680095N)

Return to the Harding Dam road, and drive 3.6 km south to the road signposted to Harding Dam. Follow the road to the parking area at the top of the dam wall (MGA 510850E 7680095N).

The Cooya Pooya Dolerite, extends across 3500 km² of the west Pilbara, and is a regionally lenticular intrusive complex, consisting of dolerite laccoliths, sills, and dykes. It intrudes the Hardey Formation beneath the Lyre Creek Member (a felsic to intermediate volcaniclastic member), and also intrudes the contact between the Hardey Formation and the Kylena Formation. Hickman (2004b) explained the Cooya Pooya Dolerite as composed of subvolcanic intrusions of the Kylena Formation. The lower level of the Cooya Pooya Dolerite is up to 150 m thick, and was probably formed by the trapping of basaltic magma beneath massive beds of the Lyre Creek Member. The upper sill, generally no more than 50 m thick, accumulated under basalts at the base of the Kylena Formation, and locally intruded the lower part of the Kylena Formation (Hickman, 2004b). The composition of the Cooya Pooya Dolerite is very similar to that of the lower Kylena Formation, and the Kylena Formation is typically anomalously thin where it is underlain by the Cooya Pooya Dolerite (Hickman, 2004b).

Locality 21 shows typical exposures of the Cooya Pooya Dolerite. It invariably outcrops as black hills, ridges, or large areas of dissected plateau fringed by cliffs (Fig. 39). These outcrops are fringed by steep, bouldery scree slopes without vegetation. The dolerite is massive, and generally jointed at 0.5 to 3 m intervals. In thin section, the rock is a fine- to medium-grained plagioclase-rich and biotite-bearing dolerite, with local olivine–orthopyroxene–clinopyroxene–plagioclase cumulate. The lower levels of the Cooya Pooya Dolerite, as exposed at Harding Dam, contain numerous inclusions of quartzite, representing contact metamorphosed xenoliths derived from the Hardey Formation.

Locality 22: Kylena Formation (Cooya Pooya, MGA 507520E 7647190N)

Return to the Harding Dam road, turn left onto the Robe River Railway access road (permit required), and drive 26 km south to the junction with the sealed Dampier–Tom Price Road. Turn left on this road and drive approximately 11 km south to where there is a railway bridge immediately to the west of the road (MGA 507460E 7647130N). Park north of the bridge and walk to rocky outcrops in the bed of the river (MGA 507520E 7647190N).

Locality 22 provides good exposures of basalt flows belonging to the Kylena Formation in the upper reaches of the Harding River. The formation is approximately 250 m thick in this area, and consists of subaerial flows of massive to amygdaloidal basalt and basaltic andesite. The individual flows are generally between 10 and 25 m thick, and consist of sparsely amygdaloidal lower and central sections, overlain by very amygdaloidal flow tops. As in the Mount Roe Basalt, amygdales are filled by quartz, carbonate, and chlorite. The petrography and geochemistry of the formation were described by Thorne and Trendall (2001). Locality 22 is situated very close to

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Figure 39. Harding Dam and Lake Poongkaliyarra, showing typical outcrops of the Cooya Pooya Dolerite.

the base of the formation, and the underlying unconformity has an irregular surface, due to the uplift and erosion of the Hardey Formation prior to the local deposition of the Kylena Formation. Geochronological data from the northern Pilbara (GSWA database) indicate no overlap in the depositional age ranges of the Hardey and Kylena Formations, although regionally, the time interval between the two formations is only a few million years (2752 Ma to 2749 Ma). Volcanism of the Kylena Formation marked the start of Stage 3 in the evolution of the Fortescue Group (see Part 1).

Locality 23: Stromatolitic limestone of the Tumbiana Formation (COOYA POOYA, MGA 509350E 7640400N)

Drive 9 km south to a minor gravel road (MGA 509940E 7640540N). Take this road, turn left after a few metres, and proceed 500 m to the railway crossing (MGA 510270E 7639820N). On the south side of the railway is an old disused railway access road. Follow this old road (a 'no through road') to the northwest for 1 km to Locality 23 (MGA 509350E 7640400N).

The 2727–2715 Ma Tumbiana Formation of the Northwest Pilbara Sub-basin is predominantly composed of quartz sandstone, fine-grained volcaniclastic sandstone, siltstone, shale, and minor stromatolitic limestone and dolostone. Near Locality 23, the base of the formation is composed of cross-bedded calcareous sandstone and siltstone, approximately 50 m thick, which is probably a braided fluvial deposit. This is overlain by a 20 m thick succession of lithic sandstone and siltstone containing a few thin beds (1–3 m) of stromatolitic carbonate, one of which is well exposed at Locality 23. The stromatolites include conical and domal forms generally no more than 10 cm high, and smooth to pustular mats are visible on most bedding planes. Bedding planes also exhibit ripple marks, and vertical exposures show ripple-drift cross lamination. These stromatolitic carbonates are lithologically and paleontologically very similar to thicker stromatolitic carbonate units in the Northeast Pilbara Sub-basin. Recent geochemical and sedimentological evidence from that latter sub-basin supports a lacustrine depositional setting for these units (Bolhar and Van Kranendonk, 2007; Awramik and Buchheim, 2009).

Locality 24: Contact between the Maddina and Jeerinah Formations, and a view of the Hamersley Range escarpment, (MOUNT BILLROTH, MGA 585400E 7592800N)

Return to the Tom Price Road and turn right. Drive approximately 60 km to the second of two railway crossings (MILLSTREAM, MGA 547340E 7603200N). Take the Munjina road, and continue 45 km to the Coolawanyah access road (MOUNT BILLROTH, MGA 584838E 7594500N). Drive 1.8 km along this road until it enters a small gorge (MOUNT BILLROTH, MGA 585400E 7592800N).

Both sides of the gorge are composed of altered basalt from the top of the Maddina Formation. Overlying the basalt, and forming a low line of cliffs on the ridge west of the gorge, is a 10 m thick quartzite, the Woodiana Member of the 2690–2630 Ma Jeerinah Formation. The contact between the basalt and the quartzite is also the contact between Stages 3 and 4 of the Fortescue Group (discussed in Part 1). Ripple marks are exposed on some bedding planes of the quartzite, and Thorne and Trendall (2001) interpreted the sandstone protolith of this unit as a nearshore facies of the Jeerinah Formation; the overlying succession indicates progressively deepening marine facies.

From the ridge, there is an excellent view of the Hamersley Range escarpment, 20 km to the south. The base of the cliff line exposes the top of the c. 2560 Ma Wittenoom Formation, which is overlain by the Mount Sylvia Formation (30 m succession of shale, dolomitic shale, and three thin BIF members), the Mount McRae Shale (70 m of alternating shale, dolomitic shale, chert, and minor BIF), and the lower part of the overlying Brockman Iron Formation.

From Mount Florence drive 53 km to the intersection with the Nanutarra–Munjina Road. Turn left and drive 68 km to the Auski Roadhouse at Munjina. Vehicles without long-range fuel tanks will need to refuel at Munjina.

DAY 4. Hamersley Basin

The localities to be visited on Day 4 are shown on Figure 40, which combines parts of the ROY HILL and NEWMAN 1:250 000 map sheets to give an overview of the structure of the southeast Ophthalmia Fold Belt.

Locality 25: Conjugate box fold in gently dipping Bee Gorge Member, Wittenoom Formation (MOUNT GEORGE, MGA 673513E 7520596N)

From Munjina Roadhouse travel south along the Great Northern Highway for 3.4 km. Park on the east side of the road in a layby north of a road cutting (MOUNT GEORGE, MGA 673740E 7520865S). Walk into the cutting to view the west wall. Take care of passing traffic.

To the south of the Fortescue Valley, the Great Northern Highway cuts up through the Hamersley Range escarpment through Munjina East Gorge. The approach to the gorge is through the upper Wittenoom Formation, Mount Sylvia Formation, and the Mount McRae Shale, which represent post-breakup shelf subsidence. The cutting at this locality (Fig. 40) passes through a BIF unit towards the top of the 150 to 250 m thick Bee Gorge Member of the Wittenoom Formation (Thorne and Tyler, 1997). A SHRIMP U–Pb zircon age of c. 2565 Ma has been obtained from the Bee Gorge Member (Trendall et al., 2004).

The BIF dips very gently to the southwest. In the western cutting wall a north-facing, open to tight asymmetric fold, plunging to the southeast and with locally developed thrusting along its axial plane, cuts across the wall (Fig. 41a). Beneath this fold, at the base of the cutting, is an open to gentle fold with a conjugate axial surface dipping to the northeast (Fig. 41b). This medium-scale conjugate fold is typical of very low-grade, brittle deformation in what is the very frontal part of the Ophthalmia Fold Belt.

At the north end of the east wall, several vertical veins are present filling fractures. These may relate to fluid flow during very low-grade metamorphism, dated in the orogenic front at c. 2145 Ma (Rasmussen et al., 2005) using authigenic monazite and xenotime.

Locality 26: Munjina East Gorge Lookout and Albert Tognolini Rest Area (Mount George, MGA 678917E 7511815N)

From the cutting, continue south along the Great Northern Highway through Munjuna East Gorge. After 11 km, at the top of the gorge, is a turn off to the Albert Tognolini Rest area and Munjina East Gorge Lookout. Park at the lookout.

The approach to the gorge is through the Mount Sylvia Formation and the Mount McRae Shale. Mount Sylvia Formation is characterized by its triplet of BIF macrobands, the uppermost of which is known as 'Brunos Band' and is a prominent stratigraphic marker traceable across the whole basin. The gorge itself is through the Dales Gorge Member of the Brockman Iron Formation, with the stepped nature of the cliffs reflecting the alternating BIF and shale macrobands.

The Munjina East Gorge Lookout overlooks Munjina East Gorge and the south face of Mount Lockier to the north. The locality provides an excellent overview of the stratigraphy of the Brockman Iron Formation (Fig. 42a and 42b; see also Fig. 18) with the Dales Gorge Member overlain first by the Whaleback Shale, forming a recessive unit, and then by the Joffre Member forming the top of the hills.

SHRIMP U–Pb dating of the Brockman Iron Formation has given ages ranging from c. 2495 Ma near the base of the Dales Gorge Member, through c. 2463 Ma from the Whaleback Shale Member, to c. 2454 Ma near the top of the Joffre Member (Trendall et al., 2004), together representing deposition of a 610 m section over a period of 40 million years. The depositional environment has been interpreted as converting to a continental back-arc setting (Blake and Barley, 1992) at this time. The deposition of BIF is discussed further by Trendall et al. (2004) and papers referred to therein.

Locality 27: Mount Robinson — zone of reverse faulting and open folding, view of the Governor (Governor, MGA 689644E 7450751N)

Return to the Great Northern Highway and continue south for 72 km. The road passes through scenic high plateau country to the east of the Karijini National Park. Railway crossings on the highway are evidence of the scale of iron ore mining in this region. Turn off the Great Northern Highway towards Mount Robinson Rest Area and park (Governor, MGA 689477E 7450565N). Walk along a track that leads into a small gorge below Mount Robinson.

The plateau country through which the Great Northern Highway initially passes from Locality 26 is characterized by broad, open, subhorizontal to gently plunging anticlines and synclines of Brockman Iron Formation and overlying Weeli Wolli Formation, before descending into country characterized by tighter folding and reverse faulting. East–west trending ridges of Brockman Iron Formation form regional synclines, whereas the Marra Mamba Iron Formation is typically surrounded by broad colluvium and alluvium filled valleys, underlain by Wittenoom Formation and the Jeerinah Formation at the top of the Fortescue Group, and crops out in the cores of regional anticlines (Fig. 40).

Mount Robinson is formed by Brockman Iron Formation in the core of the east–west trending, open to tight, upright, regional-scale Mount Robinson Syncline (Fig. 40; Tyler, 1991; Tyler et al., 1991). Locality 27 is on the southern limb of the syncline and folding occurs on all scales. Tight, upright, small- and medium-scale, gently to moderately plunging folds can be seen in BIF of the Dales Gorge Member above the track (Fig. 43a) and in the walls of the gorge. Outcrop beside the track shows the development of a weak, spaced fracture cleavage, axial planar to the folding (Fig. 43b).

Following an upper track back out of the gorge onto the shoulder of the ridge gives excellent views of the stratigraphy of the upper Wittenoom Formation, Mount Sylvia Formation, Mount McRae Shale, and Dales Gorge Member of Brockman Iron Formation, together forming The Governor (Fig. 43c), and of the regional structure looking to the west into the core of the Governor Syncline and Alligator Anticline (Fig. 43d).

Locality 28: Wonmunna Pool — metadolerite sill in Jeerinah Formation. Aboriginal rock art (Ophthalmia, MGA 717938E 7442747N)

Return to the Great Northern Highway and head southeast for 32.4 km. Turn off the highway (Ophthalmia, MGA 718535E 7441857N), and follow a track for 1.0 km. Bear left at a track junction and follow the track for 500 m to a parking and turning area above a low gorge. Walk down the track to a pool in in the gorge. Please take care not to damage the extensive gallery of aboriginal rock art present along the walls of the gorge at this locality (NO HAMMERS!).

This locality is within a metadolerite sill, intruded into the 2 km thick Jeerinah Formation outcropping in the core of the Wonmunna Anticline (Fig. 40; Tyler, 1991; Tyler et al., 1991). The Jeerinah Formation is made up predominantly of phyllite, together with interbedded metamorphosed chert units, felsic volcaniclastic rocks, and thin dolomitic units. SHRIMP U–Pb dating on zircon has given ages of between c. 2690 Ma and c. 2629 Ma (Trendall et al.,

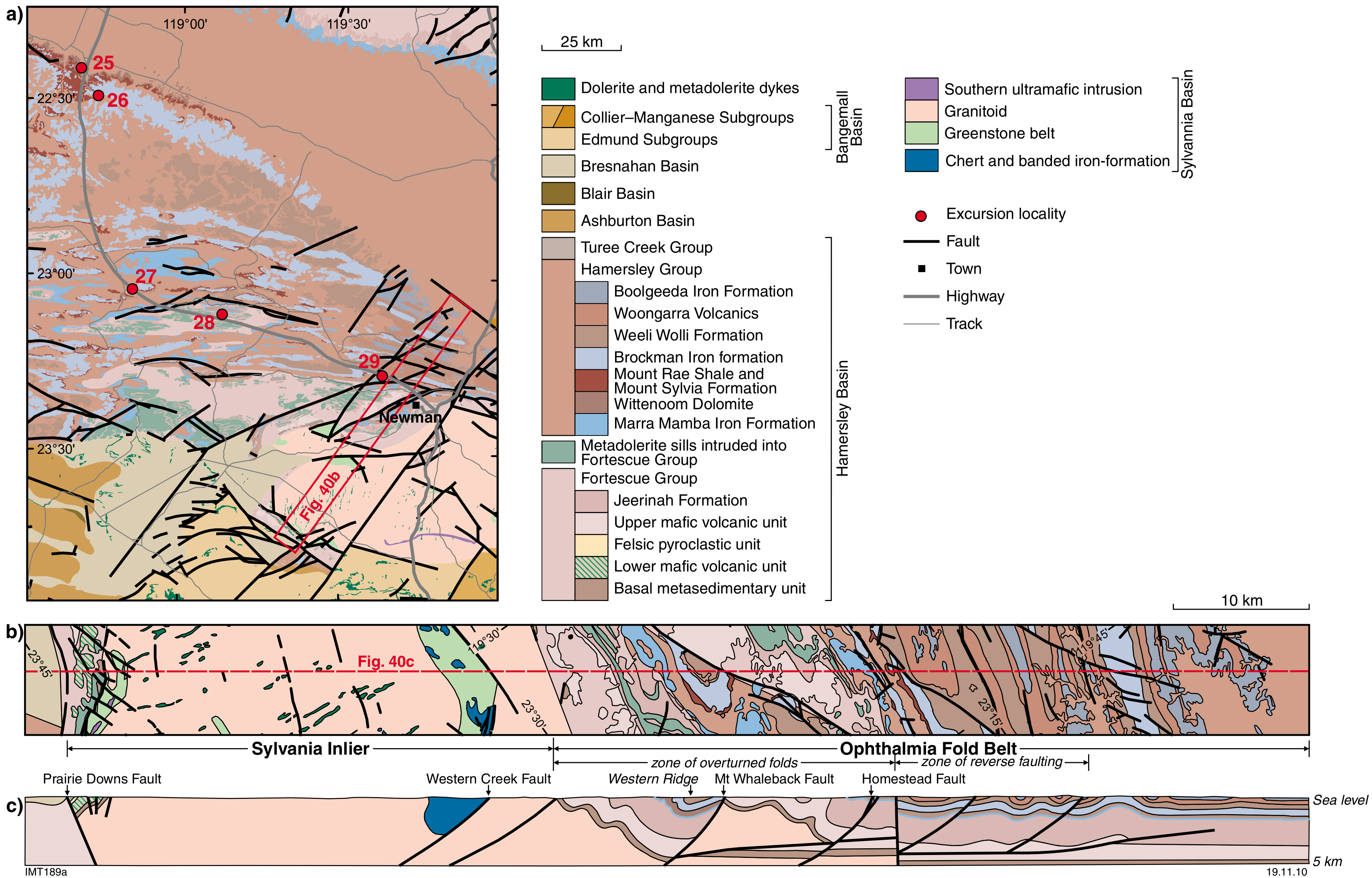

Figure 40. Location of stops 25 to 29, marked on a mosaic of the Roy Hill and Newman 1:250 000 geological map sheet (Thorne and Tyler, 1997; Tyler et al., 1991). Section line for parts b) and c) shown in red.

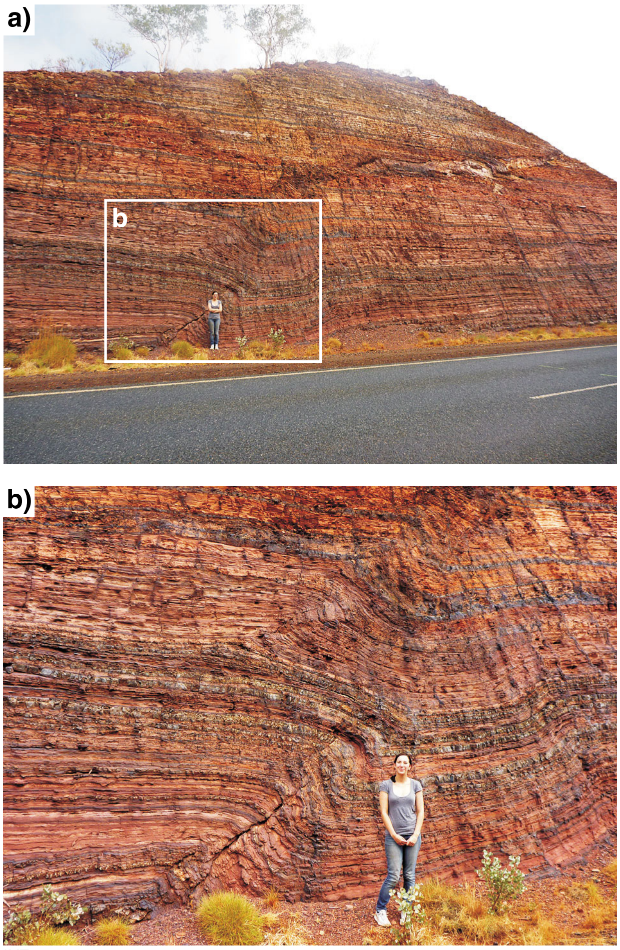

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Figure 41. West wall of road cutting: a) showing open to tight asymmetric fold cutting across the wall. Beneath this fold, at the base of the cutting, is open to gentle fold with a conjugate axial surface; b) close up of the conjugate box fold at the base of the cutting wall

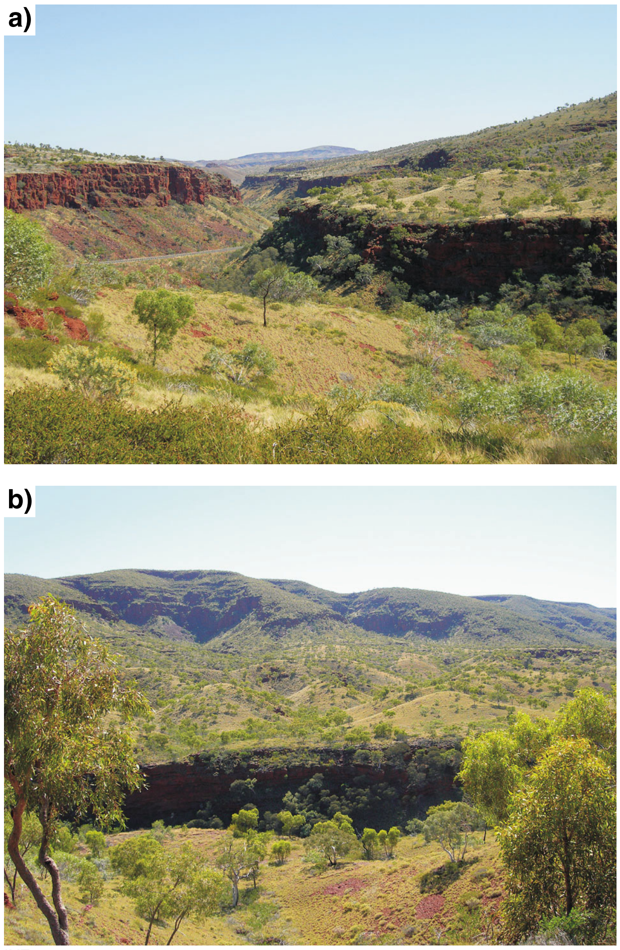

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Figure 42. From the lookout: a) looking down Munjana East Gorge. Dales Gorge Member of the Brockman Iron Formation forms the walls of the gorge; b) Mount Lockier, with the Whaleback Shale Member forming the recessive unit through the centre of the face and the Joffre Member of the Brockman Iron Formation forming the upper slopes.

2004). The Marra Mamba Iron Formation at the base of the overlying Hamersley Group has given a SHRIMP U–Pb zircon age of c. 2597 Ma (Trendall et al., 2004). These units represent the early stages of post-breakup shelf subsidence.

Metadolerite sills make up to 850 m of the Jeerinah Formation section in this area (Fig. 44). The metadolerite has been described by Tyler et al. (1991) as fine to coarse grained, with a preserved ophitic to poikilitic igneous texture. Clinopyroxene is preserved as plates up to 7 mm across that show alteration to chlorite and/or actinolite. Plagioclase laths are invariably albitized; some are replaced by intergrowths of sericite and epidote–clinozoizite. Iron oxides are leucoxenized. Pumpellyite is present in some samples, and the rocks are affected by prehnite–pumpellyite facies metamorphism (Tyler, 1991). Large-scale primary igneous layering can occur with the development of metapyroxenite at the base of some sills.

Thorne and Trendall (2001) pointed out that the occurrence of extensive metadolerite sills in the Jeerinah Formation was restricted to the south-southwest side of the Jeerinah–Sylvania Fault Zone, and used this, together with a deepening depositional environment, to imply the presence of syn-depositional growth faults in the Fortescue Basin. The zone of reverse faulting and upright folding identified by Tyler and Thorne (1990) and Tyler (1991) is interpreted as being controlled by inversion of this growth fault system at the southern edge of the Yule–Sylvania High.

Locality 29: View of Mount Newman, zone of overturned folds. Section through vertically dipping Brockman Iron Formation (Newman, MGA 768426E 7423277N)

Return to the Great Northern Highway and continue southeast for 65 km, a drive that passes to the north of the Ophthalmia Range, skirting around Pamelia Hill. At the east end of the Ophthalmia Range, the highway descends to pass through Cathedral Gorge. Park at a road junction

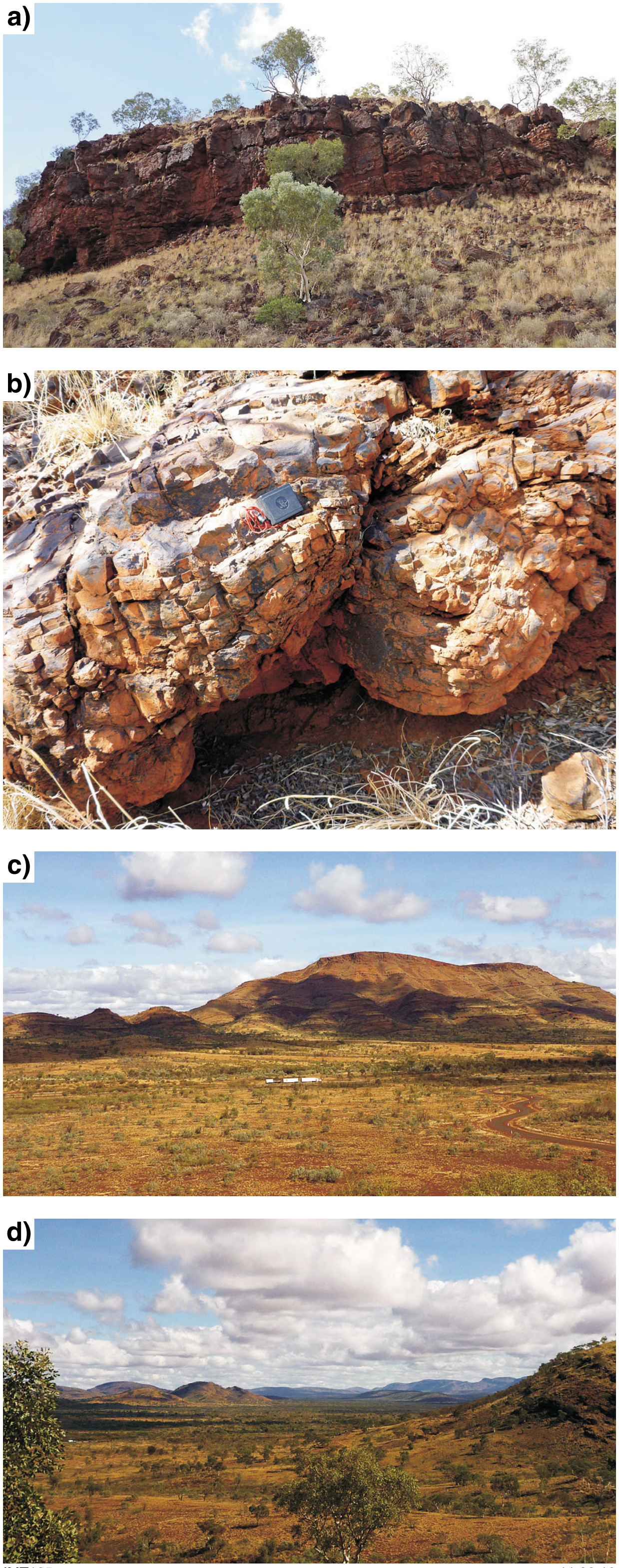

Figure 43. At the Mount Robinson Rest Area: a) tight, upright small- and medium-scale, gently to moderately plunging folds in BIF of the Dales Gorge Member; b) a weak spaced fracture cleavage axial planar to the folds; c) Upper Wittenoom Formation to Dales Gorge Member stratigraphy seen in The Governor; d) looking to the west into the core of the Governor Syncline and the Alligator Anticline

at the south end of the gorge.

Continuing along the Great Northern Highway 16 km to the east would bring you to the iron ore mining town of Newman, which is dominated by the Mount Whaleback Mine, developed in the 1960s on a world-class 1600 million ton hematite iron orebody.

From the road junction the view to the west is of the east face of Mount Newman (1057 m). The top of the face is formed by gently west-dipping Brockman Iron Formation, and the stratigraphy is the same as that at Munjina East Gorge and at The Governor. The 'Brunos Band', seen in the cliff face, is a marker at the top of the Mount Sylvia Formation. To the north, the bedding is folded to near vertical and overturned, forming an asymmetrical, north-facing regional-scale anticline (Fig. 45a). Cathedral Gorge is on the north limb of the anticline. To the south, the Marra Mamba Iron Formation forms the core of the anticline. The next fold to the south is the Mount Newman Syncline. Further to the southeast is the Sylvania Inlier, which represents Archean basement thrust up from the edge of the Pilbara Craton during the c. 2210 Ma Ophthalmian Orogeny, driven from the south by collision with the Glenburgh Terrane of the Gascoyne Province (Johnson et al., 2010).

The anticline is cut off by the Homestead Fault, part of a later, east- to northeast-trending normal fault system that includes the Mount Whaleback Fault, which forms the footwall to the Mount Whaleback orebody. This extensional fault system (see Fig. 40) probably formed during the formation of the Bresnahan Basin along the margin of the Pilbara Craton during the Mangaroon Orogeny at c. 1690 Ma (Fig. 17; Cawood and Tyler, 2004).

Walk into the gorge taking care of passing traffic.

The vertical to overturned bedding at this locality gives an opportunity to look at features in the BIF in detail. These features include the BIF and Shale macrobands (see Fig. 18), the mesobanding within the BIF macrobands, and the microbanding within the chert mesobands.

Figure 44. Metadolerite at Wonmunna Pool

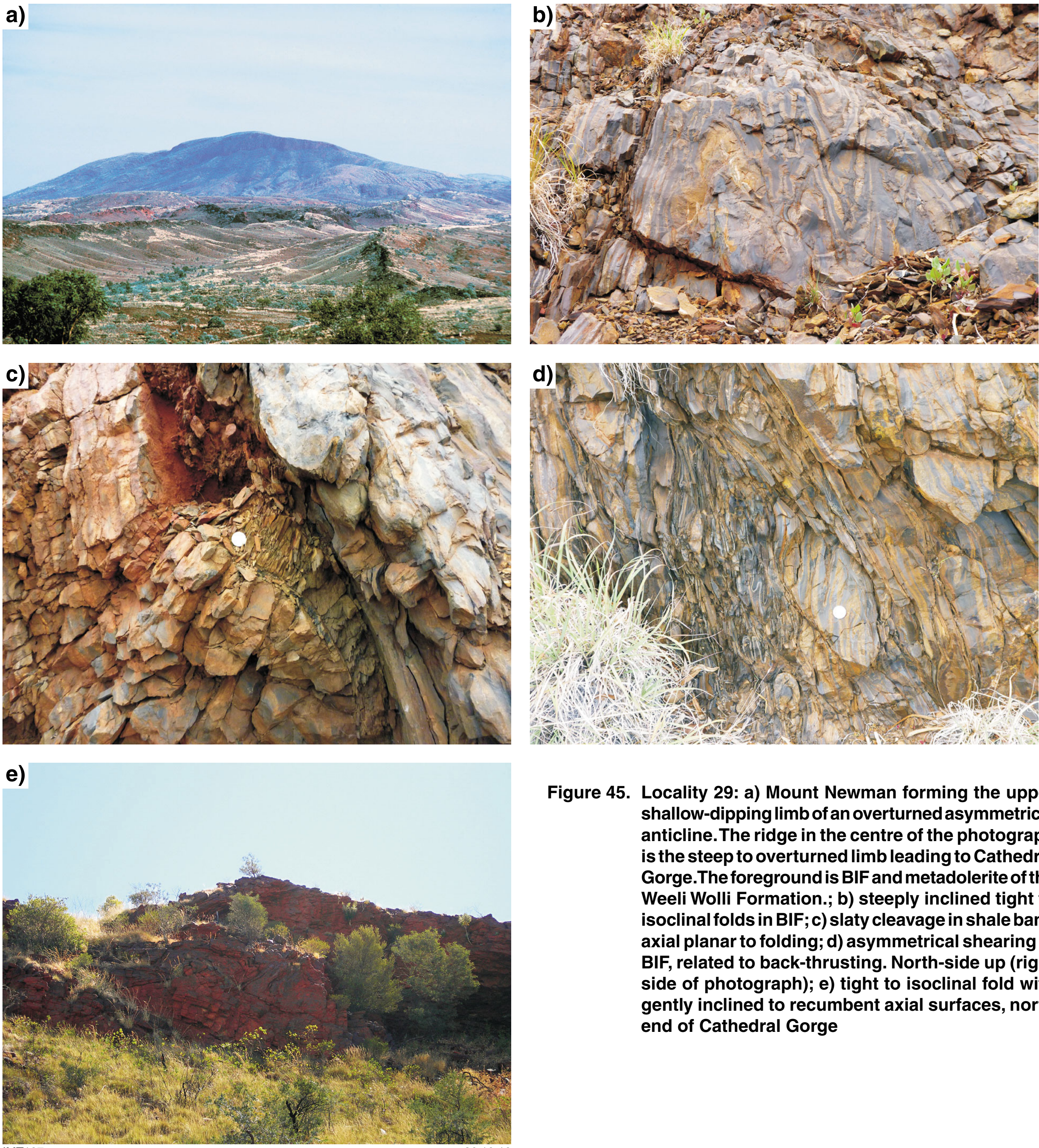

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Figure 45. Locality 29: a) Mount Newman forming the upper shallow-dipping limb of an overturned asymmetrical anticline. The ridge in the centre of the photograph is the steep to overturned limb leading to Cathedral Gorge. The foreground is BIF and metadolerite of the Weeli Wolli Formation.; b) steeply inclined tight to isoclinal folds in BIF; c) slaty cleavage in shale band axial planar to folding; d) asymmetrical shearing in BIF, related to back-thrusting. North-side up (right side of photograph); e) tight to isoclinal fold with gently inclined to recumbent axial surfaces, north end of Cathedral Gorge

Open to isoclinal folding is present on various scales, with axial surfaces generally steeply to moderately inclined or recumbent (Fig. 45b). A pervasive axial plane slaty cleavage is well developed in the recessive volcanogenic shale macrobands (Fig. 45c), and in the Whaleback Shale Member, exposed along the walls of the gorge. Shearing is developed (Fig. 45d) related to local back-thrusting. In the BIF macrobands, and in the BIF of the Joffre Member, a spaced axial planar cleavage is developed, defined by seams of hematite (Tyler and Thorne, 1990; Tyler, 1991).

The gorge leads into the overlying interbedded BIF and metadolerite of the Weeli Wolli Formation. Figure 45e shows tight to isoclinal folding with moderately to gently inclined or recumbent axial surfaces in BIF. The core of the syncline to the north is occupied by the Woongarra Rhyolite, which has given a SHRIMP U–Pb zircon age of c. 2450 Ma (Trendall et al., 2004). Local reverse faults and thrusts occur along the northern limb of this syncline. The bimodal magmatism represented by the dolerite sills and rhyolite has been interpreted as consistent with a continental back-arc setting (Blake and Barley, 1992.

Return 165 km to Munjina Roadhouse, and from there drive northwards towards Port Hedland. Overnight accommodation on the excursion will be at Indee Station, which is signposted approximately 170 km north of

Munjina. Indee homestead is approximately 7 km east of the highway.

DAY 5. Mallina Basin and Sisters Supersuite

On day five of the excursion we look at the Croydon Group — the sedimentary and volcanic rocks that form the fill of the Mallina Basin. We also examine several of the geochemically specialized intrusive suites (Sisters Supersuite) that have intruded that basin.

The first two stops look at units within the northwestern portion of the basin, where it forms part of the Whim Creek greenstone belt. In this area, the basal unit — coarse clastic rocks of the Cistern Formation — is not readily accessible due to recent mining activity. Rocks of the two overlying units — the Rushall Slate and the Bookingarra Formation (Louden Volcanic Member) — remain accessible. Following an examination of these two units, we will visit correlative units within the southeastern part of the Mallina Basin.

The Cistern Formation has been dated at 2964 ± 6 Ma (Huston et al., 2002), whereas rhyolite overlying the Bookingarra Formation (Kialrah Rhyolite) gives a maximum depositional age of 2943 ± 7 Ma (GSWA 144261). In the Whim Creek greenstone belt, VMS mineralization in rocks of both the Whim Creek and Croydon groups is dated at between 2950 and 2920 Ma (Huston et al., 2002). To the southeast of the Whim Creek greenstone belt, detrital zircons from the fine- to coarse-grained siliciclastic rocks of the Croydon Group (Mallina Formation and Constantine Sandstone) yield a maximum depositional age of 2994 ± 4 Ma (GSWA 142942), although a study of xenocrystic zircons from granites that intrude the basin indicates a more realistic maximum depositional age of c. 2970 Ma (Smithies et al., 2001). The main sedimentary fill was folded prior to intrusion of granites, beginning at c. 2955 Ma. However, detrital zircons from a shale in the eastern part of the basin indicate a second depositional cycle (a later basin) at 2941 ± 9 Ma (GSWA 142188). This is corroborated by a depositional age of 2948 ± 3 Ma (GSWA 169025) from a rhyolitic unit in the same region.

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Figure 46. Coarse pyroxene-spinifex in rocks of the Louden Volcanic Member

Locality 30: Rushall Slate — lower part of the Croydon Group in the Whim Creek greenstone belt (Sherlock, MGA 587683E 7695513N)

From Indee, return to the Great Northern Highway and drive 25 km north to Highway 1. At the T-junction, turn left, and drive 74 km west to a gate on the southern side of the highway, 500m west of the bridge over Whim Creek (MGA 587700E 7695800N), Go through the gate, turn right, and follow the track south for approximately 1 km.

In the Whim Creek greenstone belt, the lower part of the Croydon Group comprises the Cistern Formation and the overlying Rushall Slate. The 2964 ± 6 Ma date from the Cistern Formation (Huston et al., 2002) indicates a significant time-break separating deposition of the Croydon Group and the unconformably underlying Whim Creek Group. Pike et al. (2002) noted that, in general, facies of the Cistern Formation fine upwards. Basal polymictic conglomerate includes abundant clasts derived from the Whim Creek Group and passes upwards into meter- to decimetre-bedded litharenite turbidite, and centimetre-bedded, finely rippled litharenite sandstone and lithic wacke. These sandstones then pass upwards into the laminated shale with minor sandstone laminae of the Rushall Slate. Thus together, the Cistern Formation and Rushall Slate represent a continuously upwards-fining siliciclastic sequence. Throughout this single fining cycle, the abundance of epiclastic volcanic debris from the Whim Creek Group decreases upwards, and amounts of quartz and clay increase upwards (Pike et al., 2002). According to Pike (2001) and Pike and Cas (2002), initial deposition of the Croydon Group in the Whim Creek greenstone belt was unconformable on uplifted and eroded rocks of the Whim Creek Group (Cistern Formation), and later progressed to deep-water distal turbidite deposition in isolated half graben sub-basins (Rushall Slate), with sediment thickness up to 300m.

This stop provides a brief examination of a typical outcrop of the Rushall Slate. The unit is dominated by mudstone with rare laminae of fine sandstone. Pike (2001) noted that the mudstone contains a high proportion of mafic lithic grains. It is also important to note that although the commonly pillowed mafic lavas of the Bookingarra Formation primarily overlie the Rushall Slate, thin laminated mudstone intervals are locally interlayered with the lavas, and contacts are locally peperitic (Pike et al., 2002). Thus, deposition of mudstones and extrusion of mafic lavas overlapped in time.

Figure 47. Outcrop photograph of a rare ultramafic flow unit of Louden Volcanic Member

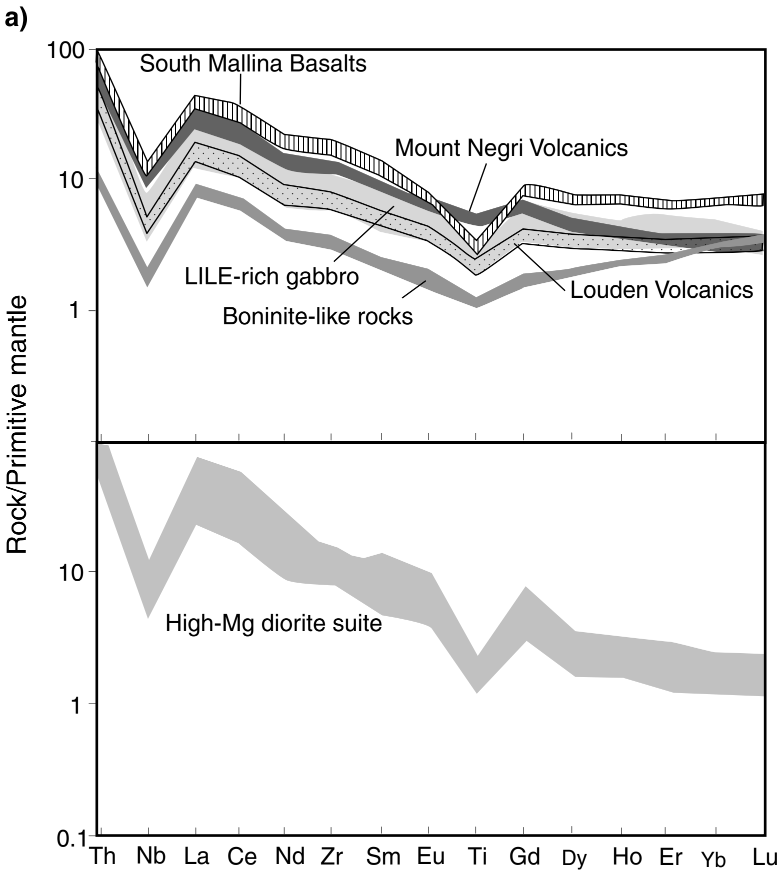

Figure 48. Primitive mantle normalized trace-element diagrams for: (top) the mafic lavas and intrusions of the Croydon Group; (bottom) high-Mg diorite (sanukitoids) of the Sisters Supersuite (Smithies et al., 2004a).

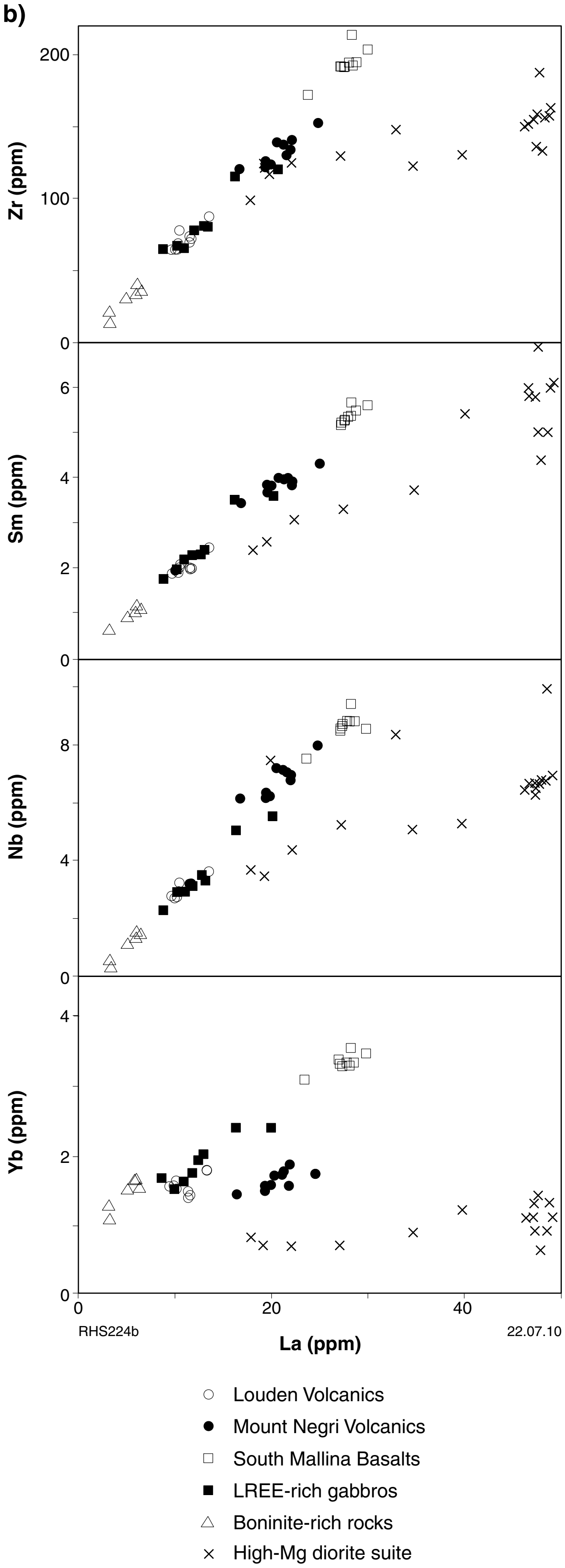

Figure 49. La vs. Zr, Sm, Nb, and Yb diagrams for the igneous rocks of the Mallina Basin (Smithies et al., 2004a)

Locality 31: Louden Volcanic Member (Bookingarra Formation) — upper part of the Croydon Group in the Whim Creek greenstone belt. Two closely spaced stops showing two variants of spinifex-texture (a) SHERLOCK, MGA 585857E 7688054N, and b) SHERLOCK, MGA 585446E 7688065N)

Return north to the gate and join the main track to the south. Drive south approximately 8 km to an intersection (MGA 589320E 7691196N). Turn to the west and follow track for 4 km to Localities 31a and 31b.

In the Whim Creek greenstone belt, the upper part of the Croydon Group is dominated by siliceous high-Mg basalt flows of the Bookingarra Formation. Included within this unit are thin and discontinuous layers of laminated mudstone, and medium- to coarse-grained quartz sandstone. Two volcanic members are recognized — a lower Mount Negri Volcanic Member, and the upper Louden Volcanic Member. They show slight, but persistent, geochemical differences and are typically easily distinguished in the field — the Mount Negri Volcanic Member is variolitic and characterized by only fine spinifex-texture, whereas the Louden Volcanic Member is never variolitic and forms units that include coarsely spinifex-textured varieties. In both cases, the spinifex texture is formed by acicular clinopyroxene, often with cores of either orthopyroxene or olivine. Deposition of both volcanic members was almost entirely subaqueous, and although pillow structures are common, non-pillowed rocks are equally abundant, reflecting rapid rates of extrusion.

The two stops at this locality provide an typical example of a pyroxene spinifex-textured flow (Stop a); Fig. 46) and a rare example of an ultramafic flow unit (Stop b); Fig. 47). Although the two volcanic members do show slight, but persistent, geochemical differences (Fig. 48), together they define a remarkably narrow range in various incompatible trace element ratios (e.g. La/Nb, La/Sm, and La/Zr; Fig. 49). Other basaltic flows, interleaved with siliciclastic rocks in the southeastern part of the Mallina Basin (i.e. in the Mallina Formation and Constantine Sandstone), also share this distinct geochemical feature (further discussed below).

Locality 32: Mallina Formation — Croydon Group to the southeast of the Whim Creek greenstone belt (SHERLOCK, MGA 589320E 7691196N)

Return east to the track intersection (MGA 589320E 7691196N). Locality 32 is on the south side of the track.

To the southeast of the Whim Creek greenstone belt, the Croydon Group has been traditionally subdivided into the quartzites of the lower Constantine Sandstone, and the finer-grained siliciclastic rocks (mudstones to medium-grained sandstones) of the upper Mallina Formation.


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Figure 50. Rarely preserved volcanic structures in siliceous high-Mg basalt flows of the South Mallina Basalt Member

This location is only marginally southeast of the southern faulted margin of the Whim Creek greenstone belt (the Loudens Fault), but provides a typical example of outcrop of the Mallina Formation. These folded turbidites generally outcrop quite poorly and discontinuously throughout the Mallina Basin. Little variation is seen in terms of sedimentary structures, bed thickness, and lithological range. The feldspathic component of these rocks is overwhelmingly dominated by plagioclase, reflecting a tonalitic or trondhjemitic source. Apart from a thin rhyolitic unit in the upper stratigraphic parts of the sequence in the far east of the Mallina Basin, there is no evidence for felsic volcanic or volcaniclastic interleaves within the siliciclastic pile. The chloritic composition of the rocks, however, likely points to a source that included a high proportion of mafic material and basaltic flow, and hyaloclastite units are observed at several stratigraphic levels. In this respect, the Mallina Formation shares close similarities with the Rushall Slate.

Tight to isoclinal folding of the Croydon Group relates to two main folding events: the first proceeding intrusion of the c. 2955 Sisters Supersuite (i.e. between c. 2970 and 2955 Ma), and the second accompanying intrusion of the c. 2935 Ma Satirist Granite of the Sisters Supersuite.

Locality 33: Siliceous high-Mg basalt flows (South Mallina Basalt Member) in the Mallina Formation — Croydon Group (Sherlock, MGA 586063E 7680635N)

From Locality 32, continue south along main track for approximately 7.5 km (MGA 587300E 7680600N). Leave the track and drive west approximately 1.3 km to Locality 33.

Siliceous high-Mg basalt, interlayed with siliciclastic rocks of the Mallina Formation, forms a minor but important part of the Mallina Basin stratigraphy southeast of the Whim Creek greenstone belt. These basalts range from massive to vesicular (Fig. 50), and locally show autobrecciation, both as flow-tops, and more commonly as hyaloclastite. Although these rocks can be distinguished geochemically from the volcanic member of the Bookingarra Formation (Fig. 48), they share several important geochemical similarities in the Whim Creek greenstone belt. In particular, all of these lavas have a common, narrow range in various incompatible trace element ratios (e.g. La/Nb, La/Sm, La/Zr: Fig. 49).

Locality 34: Boninite-like rocks and low-Ti tholeiites in the Constantine Sandstone — Croydon Group (Mount Wholer, MGA 588220E 7667395N)

From Locality 33, return east to the main track and drive south a further 17.5 km to Croydon Station (abandoned). Turn east immediately before the entrance gate to the station, and drive along a track northeast for approximately 6.5 km to Location 34.

This locality lies within the core of the Croydon Anticline, where the thickly developed Constantine sandstone of the Croydon Croup is tightly folded about a north-trending axis. In this area, two geochemically specialized mafic magma types — low-Ti tholeiites, and rocks with boninite-like compositions — are present within the contact areole of the c. 2935 Ma Satirist Granite and the c. 2950 Ma Peawah Granodiorite. The rocks at this locality are typically strongly deformed and the nature of most contacts is not clear. Nevertheless, the low-Ti tholeiites are typically medium-grained and appear to have intruded both the quartzite and the boninite-like rocks.

The boninite-like rocks outcrop over a strike length of more than 50 km, but their original extent has been obscured by intrusion of the Satirist Granite. They are typically a schistose assemblage of actinolite, chlorite, and serpentine, with accessory epidote, talc, plagioclase, and quartz. At other localities (e.g. on the southeastern edge of the Satirist Granite) that replicate this structural and stratigraphic setting, rarely preserved igneous textures are either fine-grained quench textures, or, more commonly, medium-grained olivine-cumulate textures. The former include possible hopper-textured actinolite (after clinopyroxene) in a weakly metamorphosed plagioclase-bearing rock, and oriented pyroxene-spinifex textures. Despite evidence for quenching, no definitive evidence for an extrusive origin was found. The extensive continuity of this thin unit within coarse-grained sedimentary rocks that were deposited in a high-energy environment favours an intrusive (subvolcanic) origin.

According to Smithies (2002), low-Ti tholeiites and boninites form an association typical of arc settings, the geological evidence suggests that the Mallina Basin formed in an intracontinental setting. Smithies (2002) and Smithies et al. (2004b) suggest that mantle source regions for these rocks were locally to regionally modified during a previous subduction event (perhaps related to formation of the Whundo Group, or subduction of Whundo-type material during formation of the Whim Creek Group),

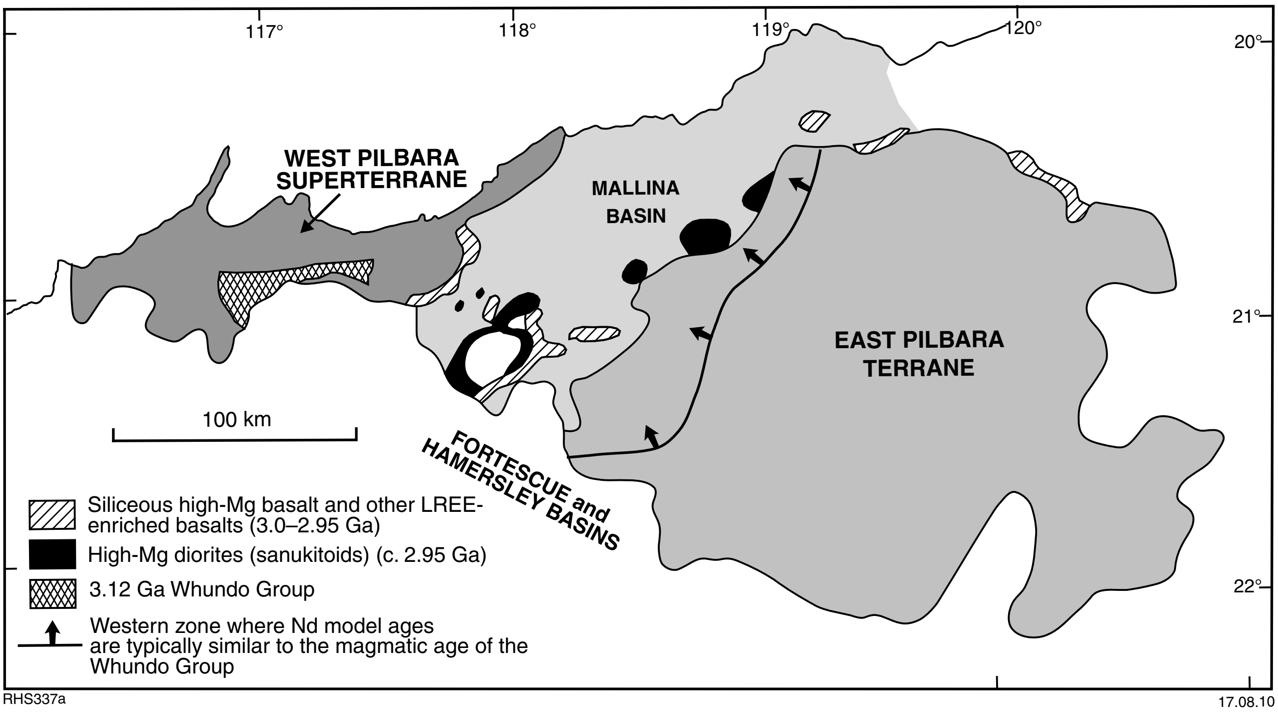

Figure 51. Map showing the distribution of various subduction-enriched magmatic units of the West Pilbara Superterrane and the Mallina Basin. Also shown is the approximate eastern limit of young (~3.12 Ga or younger) Nd depleted-mantle model ages that extend into the western margin of the East Pilbara Terrane (modified from Smithies et al., 2007c).

and remelted during a subsequent event, perhaps related to removal of a fossil slab at c. 2955 Ma (e.g. Van Kranendonk et al., 2007b). The low-Ti tholeiite relates to a mantle source already depleted during a previous mantle melting event, whereas the source for the boninite-like rocks suggests that this depleted source was modified by addition of an enriched, subduction-related component prior to mantle melting. Trace element data show that the component that enriched the mantle source of the boninite-like rocks was the same component identified in the mantle source for the lavas of the Bookingarra Group and for the South Mallina Basalt Member (Smithies et al., 2004b). This same enriched component can also be identified in the source of similarly aged siliceous high-Mg basalts and dolerites as far northwest as the Cleaverville area (Smithies et al., 2004b), and to the east and northeast across the northern margin of the East Pilbara terrane (Salt Well Member), and on the northeastern margin of the East Pilbara Terrane (Coonieena Basalt Member: Fig. 51; Smithies et al., 2007a; Van Kranendonk et al., 2007b; Pease et al., 2008).

Smithies et al. (2004b) suggested that c. 3.12 Ga Whundo-like mafic crust and homogeneous sediment derived from old (>3.3 Ga) Archean terrains was subducted to the southeast, and that partial melts derived from the subducted sediments infiltrated the mantle wedge. Resulting metasomatism was homogeneous in regards to ratios involving LREE and Zr, at least over an area large enough to provide the source that later produced the boninite-like rocks and the siliceous high-Mg basalts.

Locality 35: High-Mg diorites (sanukitoids) of the Sisters Supersuite (Mount Wohler, MGA 592867E 7667496N)

From Locality 34, rejoin the track and drive east a further 4.5 km. Leave the track and drive approximately north 400m to Locality 35.

At this locality, sanukitoids of the Peawah Granodiorite have intruded the Constantine Sandstone, plus the low-Ti tholeiites and the boninite-like rocks, prior to the folding and formation of the Croydon Anticline and the intrusion of the Satirist Granite.

Siliciclastic and mafic intrusive and extrusive rocks of the Mallina Basin were intruded by high-Mg diorites (sanukitoids) of the Sisters Supersuite between 2.955 and 2.945 Ga (Smithies and Champion, 2000). The sanukitoids form a northeast-trending chain of high-level intrusions extending for over 150 km along the axis of the Mallina Basin (Figs 51 and 52). In the east of the basin, the sanukitoids were emplaced into zones of active dilation related to extension along crustal-scale basin-parallel faults; most of the intrusions are stocks of <~20km² in areal extent. The Peawah Granodiorite is the largest intrusion, covering an area of ~180km², and is a composite body that can be geochemically subdivided into the Peawah Granodiorite (east) and the Peawah Granodiorite (west). The Peawah Granodiorite (west) has been dated at 2948±5 Ma (GSWA 118967: Nelson, 1997b). Many of the intrusions are partially surrounded by earlier intrusions of gabbro. Most intrusions also contain abundant rounded

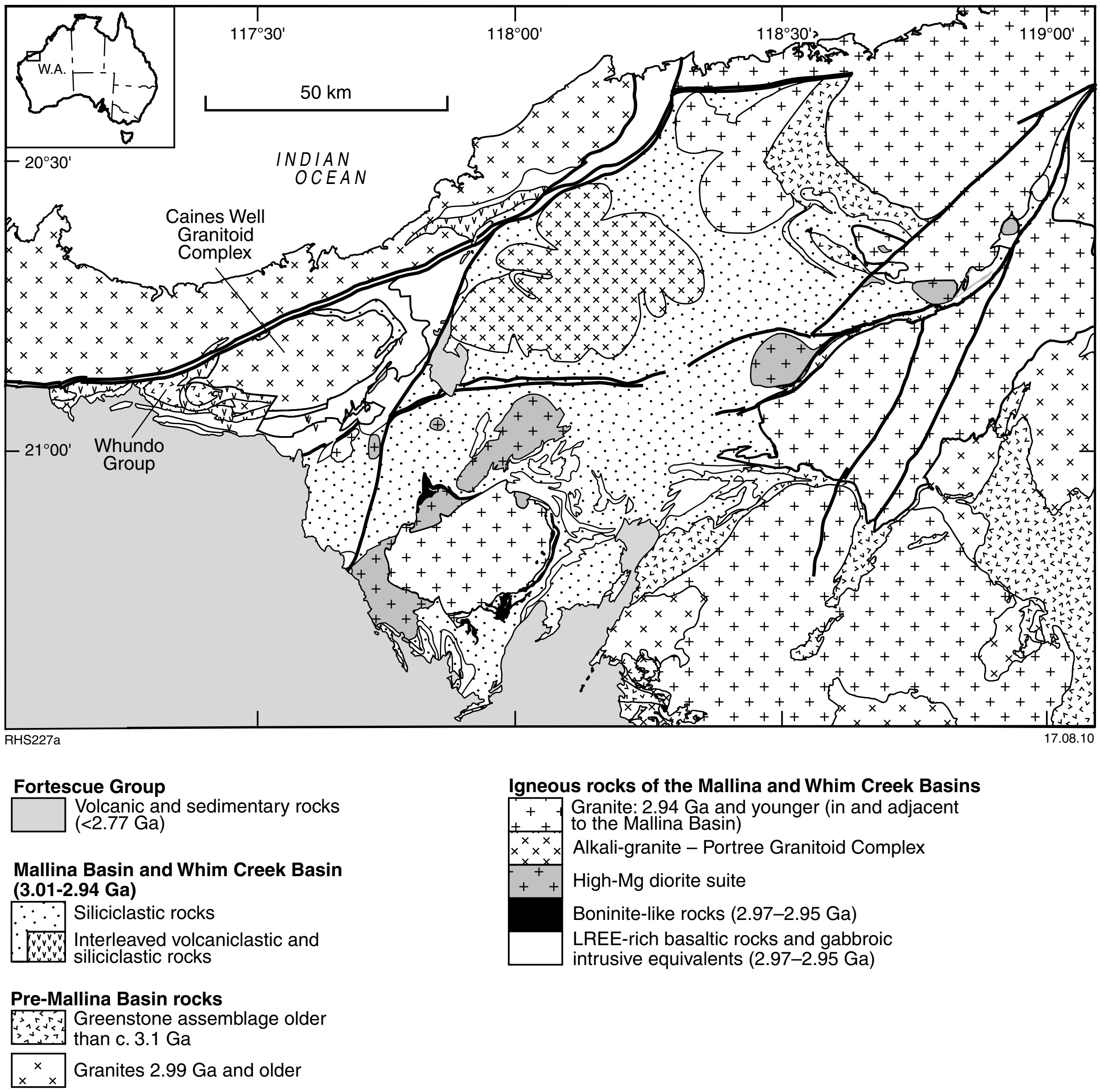

Figure 52. Geology of the Mallina Basin, highlighting main igneous rock types; modified after Smithies (2002)

enclaves up to 30 cm in diameter, mostly cognate inclusions of diorite and gabbro. Some intrusions preserve a chilled margin of fine-grained melanodiorite, which also occurs in 1–2 m thick dykes and sills in country-rock.

The sanukitoids range from diorite and monzodiorite, to tonalite and granodiorite. Mesocratic granodiorite is the most common rock type and ranges in texture from equigranular to seriate to porphyritic, with plagioclase phenocrysts up to 1cm long. Plagioclase forms a connected framework of euhedral crystals, many of which show well-developed compositional zoning from inner zones of An_{35} to sodic rims of An_{18}. Small sericite- and calcite-altered cores suggest compositions more calcic than An_{35}. Hornblende is the dominant mafic mineral, and in most granodiorites contains cores of diopside ($Wo_{45-47}En_{40-41}Fs_{13-14}$ – $Mg^{\#}$ ~ 75), variably altered to actinolite. The abundance of these cores indicates the rocks were initially clinopyroxene-rich. Biotite mantles hornblende, or forms an anhedral intergranular phase. Quartz and minor microcline are intergranular phases. Accessory minerals include magnetite, sphene, apatite, and zircon, all of which are concentrated in hornblende and biotite. Mafic clots, up to 1 cm in diameter, are locally abundant and contain hornblende with lesser diopside, biotite, plagioclase, and magnetite.

Fine-grained melanodiorite forms a chilled margin to some intrusions. It contains 2–4 mm long phenocrysts of euhedral plagioclase (An_{43} cores to rims of An_{21}), diopside ($Wo_{41}En_{42}Fs_{17}$ – $Mg^{\#}$ ~ 72) and hornblende ($Mg^{\#}$ = 65) ± orthopyroxene, within a flow-aligned

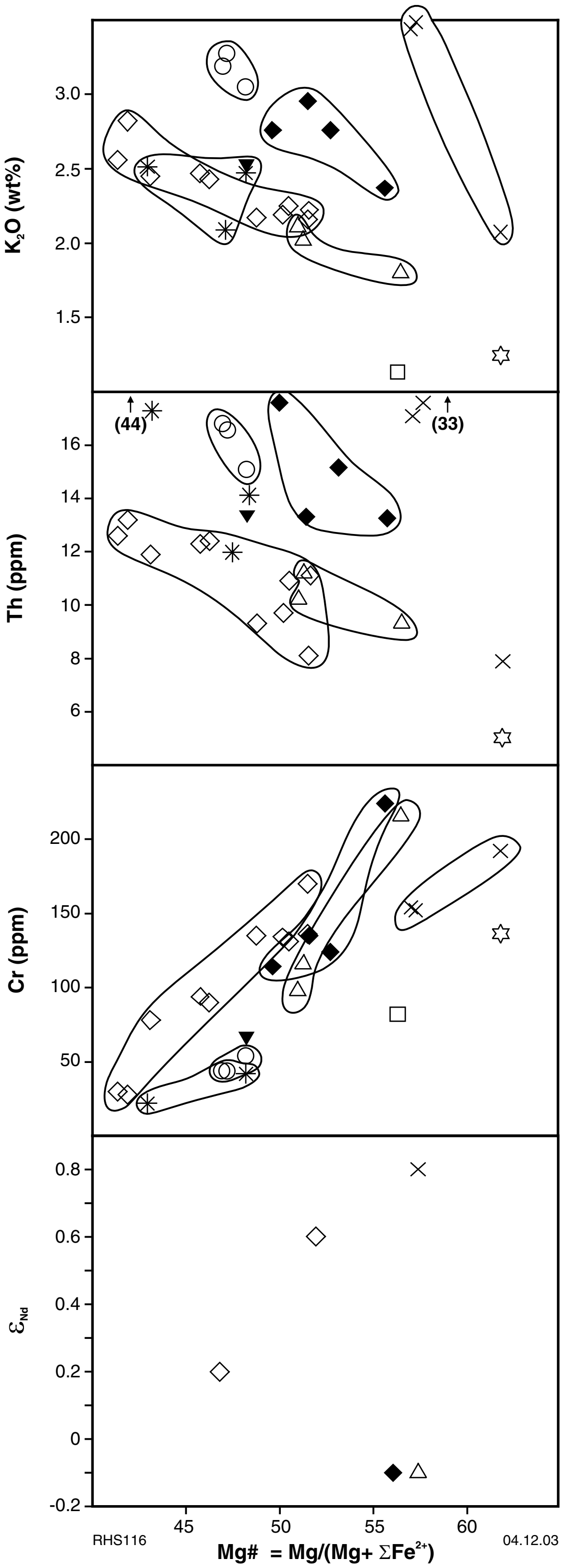

Figure 53. Compositional variation diagrams plotting Mg# against K_2O, Th, Cr, and $\varepsilon_{Nd(2.95\ Ga)}$ for high-Mg diorite (sanukitoids) of the Sisiters Supersuite. Symbols define specific plutons as described by Smithies and Champion (2000): circles = Jallagoonina; open diamond = Peawah West; closed diamond = Peawah East; double cross = Mallindra; triangle = Jones Well; cross = Wallareenya; star = Stock 1 (Smithies et al., 2004a).

groundmass of plagioclase and hornblende, with lesser biotite ($Mg^{\#} = 52$), quartz, and magnetite. Orthopyroxene occurs as discrete subhedral crystals, as granoblastic clots containing minor diopside and plagioclase, or as anhedral cores in diopside. Analysis of coexisting clinopyroxene and orthopyroxene give equilibration temperatures of ~1020°C.

The term 'Archean Sanukitoid suite' was first used by Shirey and Hanson (1984) to describe a suite of dioritic to granodioritic rocks, the more mafic of which (60% SiO_2 or less) having MgO > 6 wt %, $Mg^{\#}$ >60, and Ni and Cr >100 ppm, coupled with high LILE and Th, low HFSE (Fig. 53), and high Na_2O, K_2O, LREE, and LREE/HREE [e.g., $(La/Yb)_{PM}$ ~15–55]. Such compositions required a mantle source region to achieve the high $Mg^{\#}$, Cr, and Ni, but also a significantly enriched component to explain the high LILE, Th, and LREE. Archean sanukitoids are now recognized as a widespread, but minor (probably <<5% of Archean granitic rocks), component of many Archean terranes younger than ~2.8 Ga (Shirey and Hanson, 1984; Stern and Hanson, 1991). The only older documented examples are sanukitoids of the Sisters Supersuite, which intrude the Mallina Basin (Smithies and Champion, 2000). In the case of these sanukitoids, Smithies and Champion (2000) showed that the high LILE content could not be a result of varying degrees of either magma mixing or assimilation of felsic crust. The enriched source component was most likely an addition to the mantle source region.

However, compared to the source for the boninite-like rocks and for the siliceous high-Mg basalts of the Bookingarra Group and Croydon Group, contamination of the source for the sanukitoids was by a component with a considerably lower ratio of old Pilbara-like crust to basaltic Whundo-like crust. The significantly more radiogenic Nd-isotopic compositions of the sanukitoids form one line of evidence for this. If this is the case, then

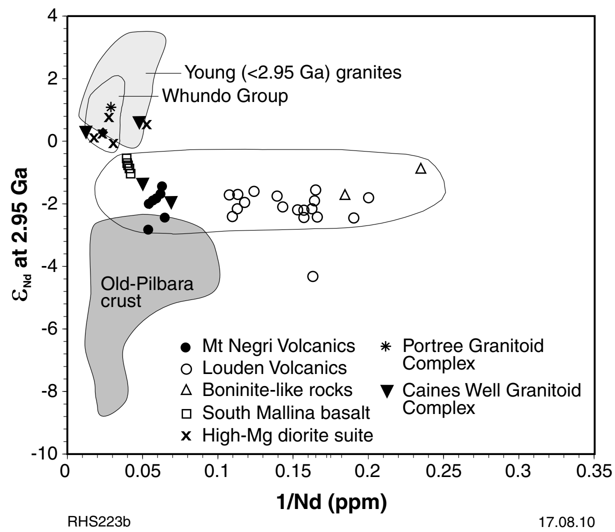

Figure 54. $\varepsilon_{Nd(2.95\ Ga)}$ vs. 1/Nd diagram, comparing the isotopic compositions of the Mallina igneous rocks with regional sources of potential contamination. Errors are typically ±0.5 epsilon units (Smithies et al., 2004a)

by c. 2950 Ma two distinct enriched mantle sources had developed beneath the central part of the Pilbara Craton. The combined geographical extent of the boninite-like rocks, siliceous high-Mg basalts, and sanukitoids shows complete geographical overlap. The entire magmatic period probably occurred within the narrow interval, between c. 2.955 Ga and c. 2.945 Ga. Consequently, the respective mantle source regions almost certainly coexisted and were not widely separated spatially.

Smithies et al. (2004b) suggested that c. 3120 Ma Whundo-like mafic crust and homogeneous sediment derived from old (>3.3 Ga) Archean terrains was subducted to the southeast, and that partial melts derived from the subducted sediments infiltrated the mantle wedge. This most likely occurred at c. 3010 Ma, the depositional age of the Whim Creek Group, but might be as old as c. 3120 Ma. Mantle in that region was refractory harzburgite, having previously yielded mafic and ultramafic magmas, including those that formed the Whundo Group.

Resulting metasomatism was homogeneous in regards to ratios involving LREE and Zr, at least over an area large enough to provide the source that later produced the boninite-like rocks and siliceous high-Mg basalts. Because relative Nb-depletions in the boninite-like rocks and the siliceous high-Mg basalts are not matched by relative depletions in other fluid immobile elements like Th and Zr, the metasomatic medium was most likely a partial melt rather than a fluid.

Compared to the source for the boninite-like rocks and the siliceous high-Mg basalts, contamination of the source for the sanukitoids was by a component with a considerably lower ratio of old Pilbara-like crust to basaltic Whundo-like crust. Smithies and Champion (2000) suggested that the mantle source for the high-Mg diorite suite was metasomatized by the addition of up to 40% of slab-melt prior to magma genesis. This would explain the more radiogenic Nd-isotopic compositions and the higher LREE concentrations and LREE/HREE ratios of the sanukitoids, particularly if melting of the slab occurred at pressures high enough to stabilize garnet. Selected trace element variations also show clear compositional differences between the respective source regions of the sanukitoids, and the boninite-like rocks and siliceous high-Mg basalts (Fig. 54). It is suggested that at greater depths, within the stability field of garnet, mafic Whundo-like crust itself partially melted to produce small volumes of adakitic magmas. These continuously interacted with mantle material as they ascended, but because of low melt/wall-rock ratios, froze close to the base of the lithosphere, where they provided the source for subsequent sanukitoid magmatism.

The variably metasomatized sub-Mallina mantle remained essentially inert for between ~50 to ~160 m.y. until, at c. 2955 Ma, it partially melted in a non-subduction setting. Smithies and Champion (2000) have attributed the late tectonothermal anomaly that caused this magmatism to either a plume or to active rifting of the Mallina Basin, whereas Smithies et al. (2004b) and Van Kranendonk et al. (2007b) suggest it more likely relates to break-off of a fossil slab. The observation that slightly later (c. 2940–2930 Ma) monzogranitic magmatism of the Sisters Supersuite becomes less voluminous and younger away from the Mallina Basin indicates that this tectonothermal anomaly was centred on the Mallina Basin.

DAY 6. Whim and Gorge Creek Groups

Day 6 of the excursion visits important outcrops of the Whim Creek and Gorge Creek Groups within pre-3000 Ma volcanic and sedimentary units of the De Grey Supergroup.

If time pernits, lunch will be taken at the historic Pilbara port of Cossack, between Localities 38 and 39.

Locality 36: 3.01 Ga Warambie Basalt and Red Hill Volcanics, Whim Creek Basin, near Red Hill (Sherlock, MGA 555230E 7682950N)

From Indee, return to the Great Northern Highway and drive 25 km north to Highway 1. At the T-junction, turn left, and drive 110 km west to a gate on the southern side of the highway, 1 km west of the bridge over Little Sherlock River (MGA 557400E 7684750N), Go through the gate and follow the track south for 1.1 km, almost to Mount Fraser, where the track swings west. Drive west for approximately 1.8 km, then turn south to a low ridge of basalt approximately 400 m south of the track. Localities 36 and 37 are within walking distance of this point.

Localities 36 and 37 are located in the Whim Creek greenstone belt, which contains all exposures of the Whim Creek Group, and the most northwesterly exposures of the Croydon Group. The geology of the Red Hill area and Localities 36 and 37 are shown on Figure 55. Coarse basal breccia at this locality represents the lowest part of the Warambie Basalt, and is composed of angular boulders of basalt and granite within a poorly sorted coarse sandstone/granulestone matrix (Fig. 56). The granite boulders were derived from the adjacent Caines Well Granitic Complex, which elsewhere includes rocks as old as 3090 Ma (GSWA 118965: Nelson, 1997a). Approximately 1.5 km to the west, the Warambie Basalt unconformably overlies amphibolite-facies metabasalt of the 3130–3110 Ma Whundo Group. The basal breccia is interpreted to be derived from basement uplift and erosion, and may be proximal to a syn-depositional fault related to 3010 Ma extension. Transport was likely to have been by clast rolling and sliding.

The Warambie Basalt contains a range of sheet lava types, including thick (~24 m) tabular flows and thin (0.5 – 2 m) irregular flows. At this level, the Warambie Basalt is dominated by coherent lava with lesser autobreccia and rare hyaloclastite. Vesicle layers occur within the basaltic lavas, which also include aphanitic and many plagioclase-glomeroporphyritic types.

Dacitic rocks of the Mons Cupri Dacite Member of the Red Hill Volcanics are encountered about 200 m southwest

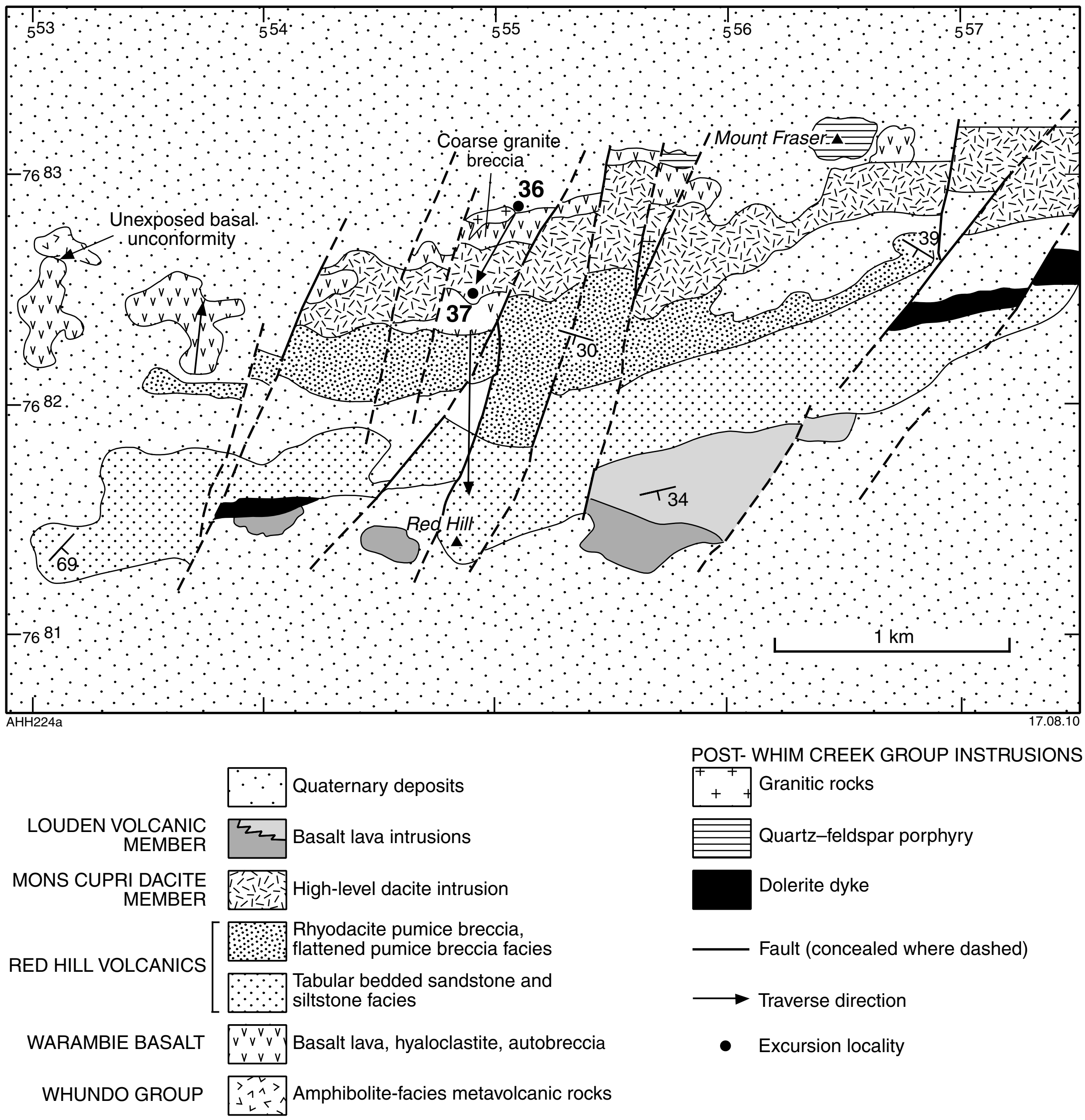

Figure 55. Simplified geological map of the Red Hill area in the Whim Creek greenstone belt, showing excursion localities 36 and 37.

Figure 56. Coarse basal breccia of the Warambie Basalt at Locality 36. Angular boulders of basalt and granite are set in a poorly sorted sandstone/granulestone matrix.

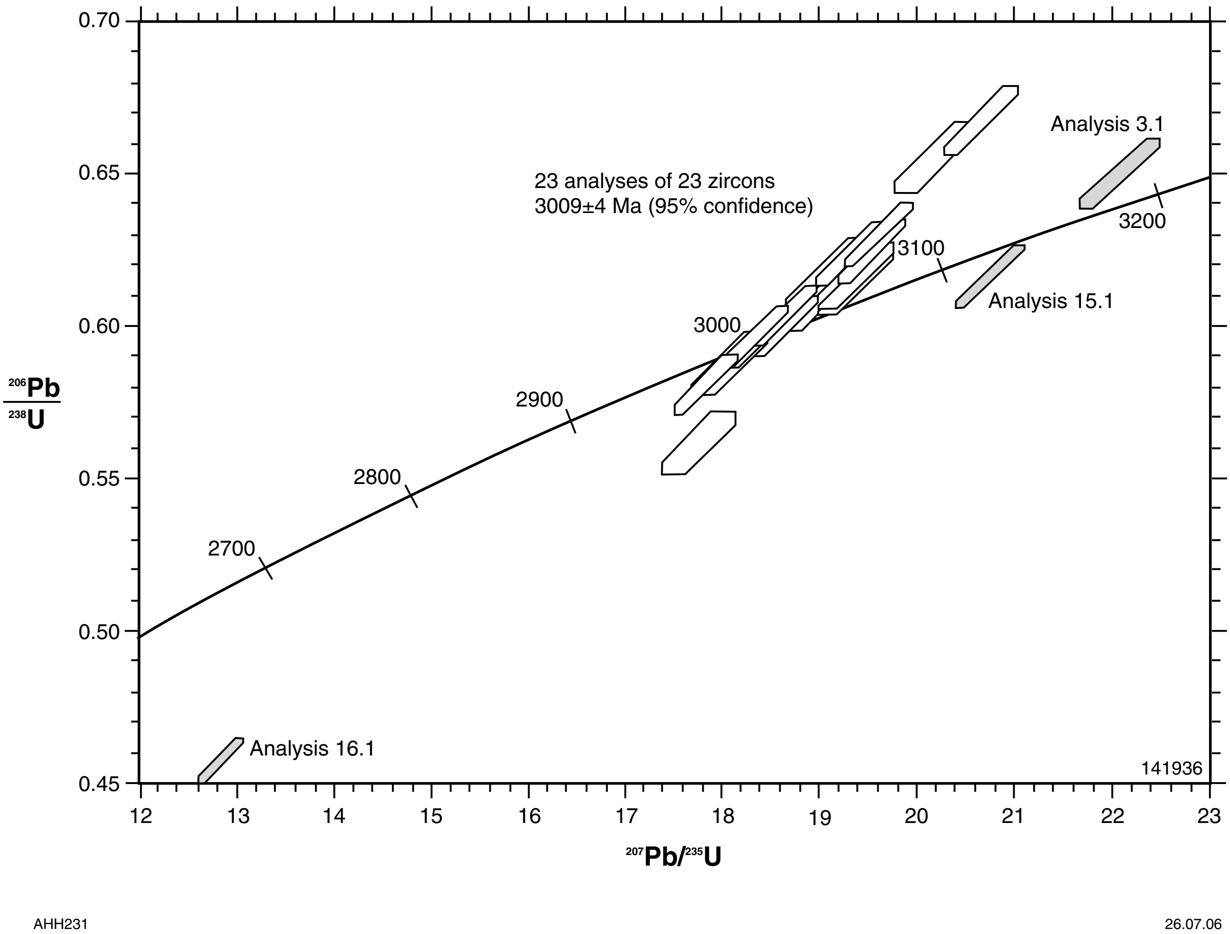

Figure 57. Concordia plot for GSWA 141936, Red Hill Volcanics, Red Hill (after Nelson, 1998c)

of the basal Warambie Basalt (at MGA 555150 7682800), and are dark grey in colour. The dacite is wholly coherent and contains a spherulitic matrix indicative of high-temperature devitrification of volcanic glass. The dacite also contains plagioclase phenocrysts and abundant, cherty fragments that may be early-formed crystals.

Locality 37: Traverse across the Red Hill Volcanics at Red Hill (Sherlock, MGA 554950E 7682450N)

Locality 37 is situated about 400 m south-southwest of Locality 36, and provides a good starting point for a north–south traverse east of a fence line, across the upper Warambie Basalt and the lower and central part of the Red Hill Volcanics

At the beginning of the traverse, dark grey basaltic breccia has an irregular contact with the upper surface of the dacite observed at the last locality. Tracing out the contact reveals a lobate upper margin to the dacite that may be original topography. However, the preferred interpretation is that this highly irregular contact is the result of injection of dacite into the basaltic breccia. If this interpretation is correct, then the presence of spherulitic matrix in the dacite suggests a relatively high-level of emplacement (subvolcanic sill?).

The basaltic breccia is jigsaw-fit and is composed of dark grey aphanitic clasts with distinct, margin-parallel banding. There are rare, wispy apophyses of the same material that are clearly not clasts. The breccia is an excellent example of hyaloclastite in which basaltic magma has undergone quenching in the presence of water. The clasts have rapidly contracted and shattered in situ. Clast-rotated breccias may indicate limited movement. The basaltic apophyses represent ‘fingers’ of basaltic magma that have remained whole, perhaps protected by the breccia carapace.

The flat ground at the start of the 500 m traverse from Locality 37 to a position (MGA 554950E 7681950N) 300 m northwest of Red Hill is underlain by centimetre- to decimetre-scale tabular beds of shard-rich volcaniclastic rocks. These rocks outcrop only in deep creek beds, and are easily overlooked. Interpretation is difficult due to limited outcrop, but they appear to be the medial to distal products of turbidity currents. Pumice-rich breccia underlies much of the traverse. The first outcrops contain both dense and pumiceous breccia in massive beds. This volcaniclastic material is rhyodacitic in composition. This angular volcaniclastic breccia represents the onset of felsic volcanism, and probably resulted from debris-flows derived from adjacent volcanic areas.

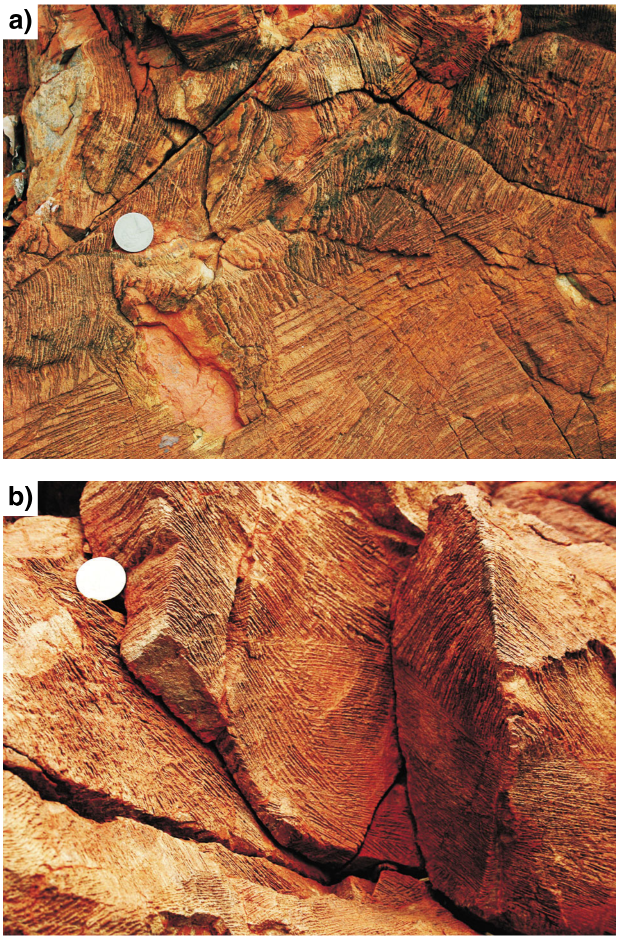

Figure 58. Bladed olivine-spinifex texture in the Ruth Well Formation at Mount Hall, Locality 38: a) individual platy olivine crystals, pseudomorphed by serpentine, tremolite, and chlorite, are up to 50 cm long; b) close-up of interlocking olivine crystals, scale in both pictures, 3 cm diameter coin.

To the south, Red Hill is composed entirely of angular tube pumice breccia. The matrix to the pumice breccia is glass-shard sandstone. The pumice breccia lacks bedding and has a monomictic composition, and represents syn-volcanic sedimentation. The large amount of pumiceous material points to an ignimbrite origin, as does the presence of relic shard textures. A sample of the breccia, collected 50 m west of the fence line and 300 m northwest of Red Hill, gave a U–Pb SHRIMP zircon an age of 3009 ± 4 Ma (GSWA 141936: Nelson, 1998c; Fig. 57). Individual concordant or slightly discordant zircon grains gave $^{207}Pb/^{206}Pb$ ages between 3150 Ma to 2995 Ma, confirming a significant detrital component in the rock, with possible sources including the Whundo Group and Railway Supersuite.

Locality 38: Komatiite flows in the Ruth Well Formation, Roebourne Group, at Mount Hall (Roebourne, MGA 520770E 7701337N)

Return to Highway 1, turn left and drive 43 km west to Locality 38, which is located on a bend (MGA 520810E 7701370N), 6 km east of Roebourne.

Roadside exposures at Mount Hall provide one of the few accessible exposures of well preserved sheaf and random spinifex-textured ultramafic komatiite flows in the Ruth Well Formation. Olivine plates (pseudomorphed by serpentine and tremolite) are up to 50 cm long (Fig. 58a,b). Microscopic examination of the sheaf spinifex texture reveals a highly magnesian rock comprised mainly of tremolite, serpentine, chlorite, and a pale yellow, birefringent phyllosilicate (probably vermiculite). Serpentine, and tremolite or vermiculite has replaced olivine blades. Tremolite and serpentine also form very fine intergrowths, and chlorite forms randomly oriented flakes. Opaque minerals are disseminated through the rock as anhedral grains and granular aggregates, and as discontinuous linings along narrow fractures or veinlets. Some of the opaques have a translucent, reddish-brown character around their outer margins, suggesting they are iron oxides such as hematite. Layers of different spinifex texture indicate that the ultramafic flows are about 2 m thick.

The komatiitic flows are overlain by a thin unit of quartzite and ferruginous chert, the bedding in which shows that the succession dips 30 to 50° southwards. Further south, this is overlain by amphibolite (metabasalt), with gabbro, pyroxenite, and serpentinized peridotite of the Andover Intrusion. At Mount Wangee, 7 km to the north, similar spinifex-textured ultramafic flows are overlain by massive and pillowed basalts of the Regal Formation, but the succession in that area dips northwestward. There are no Archean exposures between these localities, but aeromagnetic data indicate that the Harding Granitic Complex underlies Cenozoic deposits of the coastal plain. The opposing inclinations of the Ruth Well Formation at Mount Hall and Mount Wangee indicate that the Harding Granitic Complex occupies the core of an anticline in the Ruth Well Formation.

Figure 59. Pillow basalt in the Regal Formation at Cleaverville, Locality 39. The morphology of the pillows (convex tops and 'tails' at bases) consistently shows way-up to the northwest in this area.

Locality 39: Pillow basalt of the Regal Formation underlying the Cleaverville Formation (Roebourne, MGA 500694E 7714188N)

From Locality 38, drive 6 km west to Roebourne. Continue west on Highway 1 for 12 km to the turnoff signposted to Cleaverville Beach (MGA 503700E 7705300). Follow the Cleaverville Beach road approximately 9 km north to the second of two causeways across a tidal lagoon (MGA 500694E 7714188). Locality 39 is situated 50 m east of the northern end of the causeway.

At Locality 39, marine erosion has produced a wave-cut platform of weathered pillow basalt (Fig. 59). Here, and within a 1 km radius of this locality, pillows consistently show way-up to the northwest, towards the BIF of the Cleaverville Formation on Cleaverville Peninsula. Available geochronology indicates that the depositional age of the pillow basalt at Cleaverville is c. 3200 Ma (see Part 1). However, previous workers have reached different opinions on the age and tectonic setting of the overlying Cleaverville Formation (see Locality 40). The contact between the pillow basalt and the Cleaverville Formation is locally a sheared pebbly sandstone, and Nelson (1998a: GSWA 127330) used the SHRIMP U–Pb zircon method to interpret the maximum depositional age of this sandstone as 3058 ± 7 Ma. This depositional age is consistent with the interpretation (Part 1) that the Cleaverville Formation at Cleaverville belongs to the Gorge Creek Group (De Grey Superbasin), whereas the underlying pillow basalts belong to the Regal Formation. Nelson (1998a: GSWA 127330) dated two concordant zircon grains in the sandstone at c. 3287 and 3461 Ma respectively, indicating either Paleoarchean sources for part of the detritus, or erosion of younger rocks that had Paleoarchean sources.

Locality 40: Cleaverville Formation at Cleaverville Beach (Roebourne, MGA 503290E 7716644N)

Continue northwards along the road for 1.4 km, and turn right at the T-junction. Follow the graded gravel road east for 3.5 km to the most easterly beach (MGA 503290E 7716644N).

Significance of the Cleaverville area

Geological investigations of the Cleaverville area have promoted the interpretation that plate tectonic processes were responsible for the Mesoarchean tectonic evolution of northwest Pilbara Craton. Prior to 1991, the Cleaverville Formation (BIF and chert interbedded with shale, mudstone, and tuff) was interpreted as part of a Pilbara-wide sedimentary succession (Gorge Creek Group) overlying the Warrawoona Group (Hickman, 1983, 1990). In the east Pilbara, Eriksson (1981) interpreted the Gorge Creek Group to have been deposited in submarine mid-fan, outer-fan, and basin-plain settings due to continental to marine transition following continental rifting. Making comparisons to the east Pilbara, Hickman (1983) suggested that the pillow basalts at Cleaverville might

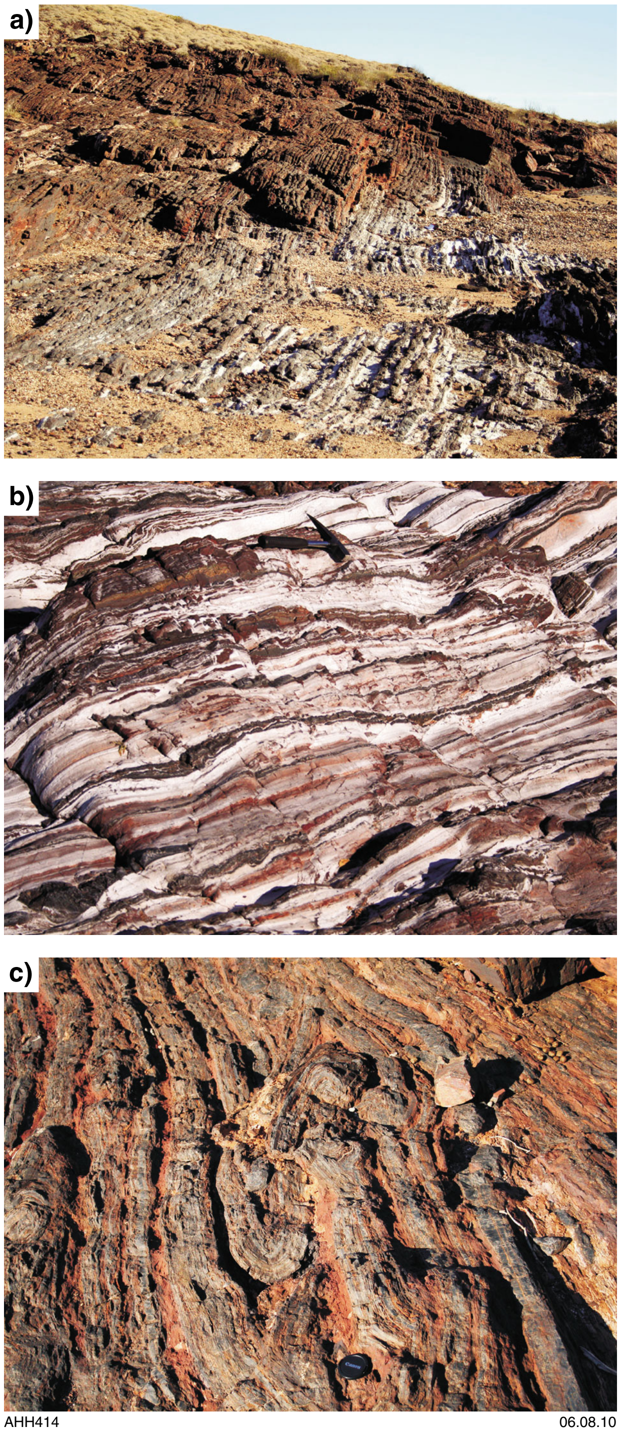

Figure 60. Outcrops of the Cleaverville Formation at Cleaverville Beach, Locality 40: a) the weathered aspect of the Cleaverville Formation differs markedly between the ridge and the wave-cut platform; b) outcrop in the tidal zone, where bleached fine-grained clastic rocks predominate over chert and iron formation; c) outcrop on the ridge, where the main rock type appears to be banded iron-formation, here deformed by tight folding and faulting.

belong to the upper part of the Warrawoona Group, but no geochronology was then available to test the correlation. Subsequent geochronology has established that the 1983 correlation is incorrect, and these basalts are now assigned to the Regal Formation (Hickman, 1997a).

Isozaki et al. (1991), Ohta et al. (1996), and Kato et al. (1998) interpreted the Cleaverville succession to be Mesoarchean mid-ocean ridge crust accreted onto the northwest margin of the Pilbara Craton in a complex series of tectonic slices. Kiyokawa and Taira (1998) and Kiyokawa et al. (2002) interpreted the same assemblage to be an immature island-arc succession thrust across the Karratha Granodiorite and the Nickol River Formation from the south. Krapež and Eisenlohr (1998) used information from Kiyokawa (1993) and Kiyokawa and Taira (1998) to propose that the Cleaverville succession formed on a platform within an intra-arc basin, but emphasized that relationships to successions farther southeast indicated that it was not part of an exotic terrane. These interpretations are briefly critically reviewed in Part 1.

SHRIMP U–Pb zircon geochronology has indicated that all outcrops correlated with the Cleaverville Formation in the west Pilbara, and outcrops of similar BIF in the extreme northwestern part of the east Pilbara, are the same age (3020–3015 Ma). Several attempts to precisely date the major iron formations of the Gorge Creek Group in the east Pilbara have been unsuccessful. In the west Pilbara, the Cleaverville Formation is up to 1500 m thick, and is composed of BIF, ferruginous chert, grey-white and black chert, shale, siltstone, and minor volcanogenic sedimentary rocks. As it includes magnetite-bearing iron formation, the Cleaverville Formation is readily identifiable on regional aeromagnetic images, and iron ore deposits exist southeast of Cleaverville. In the Roebourne–Dampier area, these images confirm that the Cleaverville Formation is folded around the Prinsep Dome (Fig. 9). Outcrops at Cleaverville are located on the northwestern limb of this fold, but a reversal of way-up in pillow basalts to the northwest of the Cleaverville Formation suggest that the main BIF ridge of the Cleaverville Peninsula forms the core of an isoclinal synform. This narrow synform (the 'Cleaverville Syncline', Beintema, 2003) probably extends southwest along the line of the coast to Karratha, then inland to the area immediately north of Mount Regal (Fig. 9).

Beach exposures, and geoscientific drilling

Wave-cut platforms at Cleaverville Beach provide excellent, but weathered, exposures of steeply inclined BIF and interbedded fine-grained clastic sedimentary rocks (shale and mudstone; Fig. 60a). Exposures on the ridge are silicified, and resemble normal BIF, with cycles of chert–hematite/goethite and shale at relatively constant 10–20 cm intervals. Aeromagnetic data indicate that the formation is a chert–magnetite BIF at depth, and mineral exploration drilling 5 km to the east of Cleaverville has confirmed that the formation is partly composed of chert–magnetite BIF.

On wave-cut platforms, along strike from the ridge outcrops, the formation appears to contain more mudstone and shale (Fig. 60b). Deformation of the succession includes tight to isoclinal folding (Fig. 60c) and minor shear zones.

In August 2007, Japanese researchers, mainly from Kyushu University, drilled three diamond drill holes, two through the Cleaverville Formation and one through the Dixon Island Formations. The plan was to investigate two units previously mapped by Kiyokawa (1993) and Kiyokawa and Taira (1998); the main objectives were to establish the detailed stratigraphy of the sections drilled, and to search for Archean microfossils similar to possible microfossil structures discovered by surface investigations (Kiyokawa et al., 2006). Near-surface alteration at Cleaverville was found to extend to vertical depths of between 30 and 50 m, making the upper sections of each drill hole unsuitable for detailed study. Below the alteration zone, diamond drill core was collected for geochemical investigation (major, trace, and rare earth element analysis, and organic carbon, nitrogen, sulfur isotope studies), petrological examination, and paleomagnetic measurements.

In the holes through the Cleaverville Formation, the dominant lithology was black shale, with frequent zones of reddish black shale, possibly in oxidation zones along shears. No BIF was intersected, possibly because the c. 100 m holes did not extend far enough below the ridge of BIF outcrop (Fig. 60a). The hole aimed at intersecting the Dixon Island Formation was commenced in altered pillow basalt of the Regal Formation and, inclined northwards, intersected a black shale succession containing thin pyrite layers; on the surface these shales mainly outcrop as chert. As in the two holes through the Cleaverville Formation, there were several zones of fragmentation and shearing.

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