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EVOLUTION OF THE WEST PILBARA GRANITE–GREENSTONE TERRANE AND MALLINA BASIN WESTERN AUSTRALIA — A FIELD GUIDE

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4th International Archaean Symposium

Geological Survey of Western Australia

GEOLOGICAL SURVEY OF WESTERN AUSTRALIA

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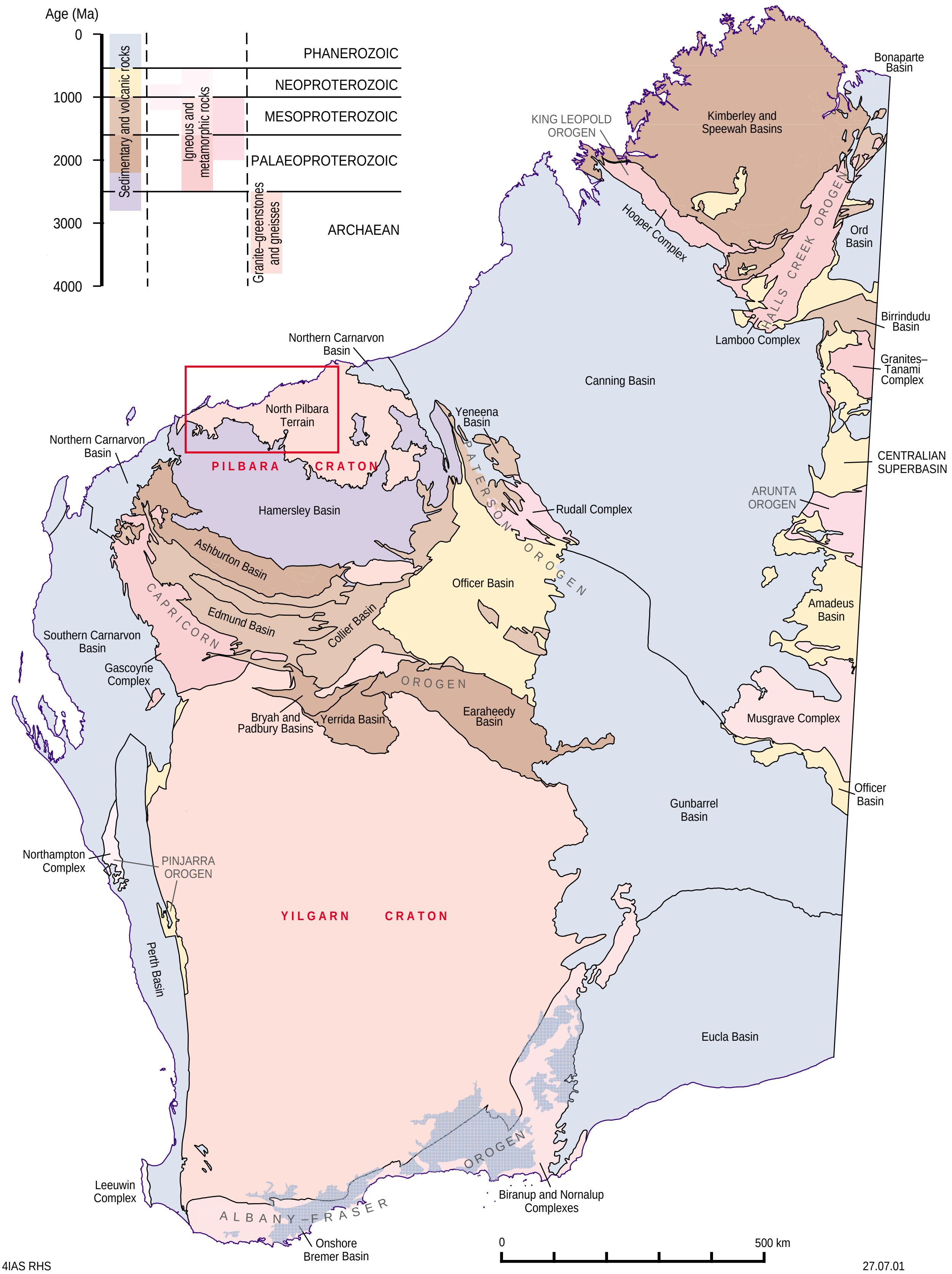

Evolution of the West Pilbara Granite–Greenstone Terrane and Mallina Basin, Western Australia — a field guide

by

A. H. Hickman[1], R. H. Smithies[1], G. Pike[2], T. R. Farrell[1], and K. A. Beintema[3]

Introduction

Since the previous Archaean Symposium excursion to the Pilbara (Hickman, 1990) a large amount of new information has been obtained through geoscientific research and geological remapping. Regional high-quality airborne magnetic and radiometric data obtained in 1995, and precise geochronology using a sensitive high-resolution ion microprobe (SHRIMP), have been extensively used during six years of mapping using 1:25 000 scale coloured aerial photography. The geochronology, in particular, has been of immense value in constraining the timing of magmatism and sedimentary deposition, and in rigorously testing stratigraphic correlations. As a consequence of this work, the review of the granite–greenstone terrain that was presented in 1990 has been found to require major revision. Most importantly, the granite–greenstone terrain, now referred to as the North Pilbara Terrain, has been divided into three granite–greenstone terranes that are separated by two sedimentary basins (Fig. 1). The c. 3270–2920 Ma West Pilbara Granite–Greenstone Terrane (WPGGT) is separated from the c. 3660–2850 Ma East Pilbara Granite–Greenstone Terrane (EPGGT) by the c. 3010–2940 Ma Mallina Basin. On the southeastern side of the EPGGT is the Mosquito Creek Basin (similar in age to the Mallina Basin), which further to the southeast is in faulted contact with the dominantly granitic, and as yet poorly documented, c. 3300 Ma Kurrana Terrane.

Reviewing this subdivision, Van Kranendonk et al. (in prep.) interpreted the WPGGT and Mallina Basin as products of post-3300 Ma rifting on the northwestern margin of the EPGGT. Deposition of the Mallina Basin was controlled by rifting and strike-slip faulting within the east-northeasterly trending zone that separates the WPGGT from the EPGGT (Smithies et al., 1999; Smithies et al., 2001). This zone of deformation and granitoid intrusion is referred to as the Central Pilbara Tectonic Zone. The purpose of this field excursion is to provide participants with a geological transect from the EPGGT, through the Mallina Basin, to the WPGGT.

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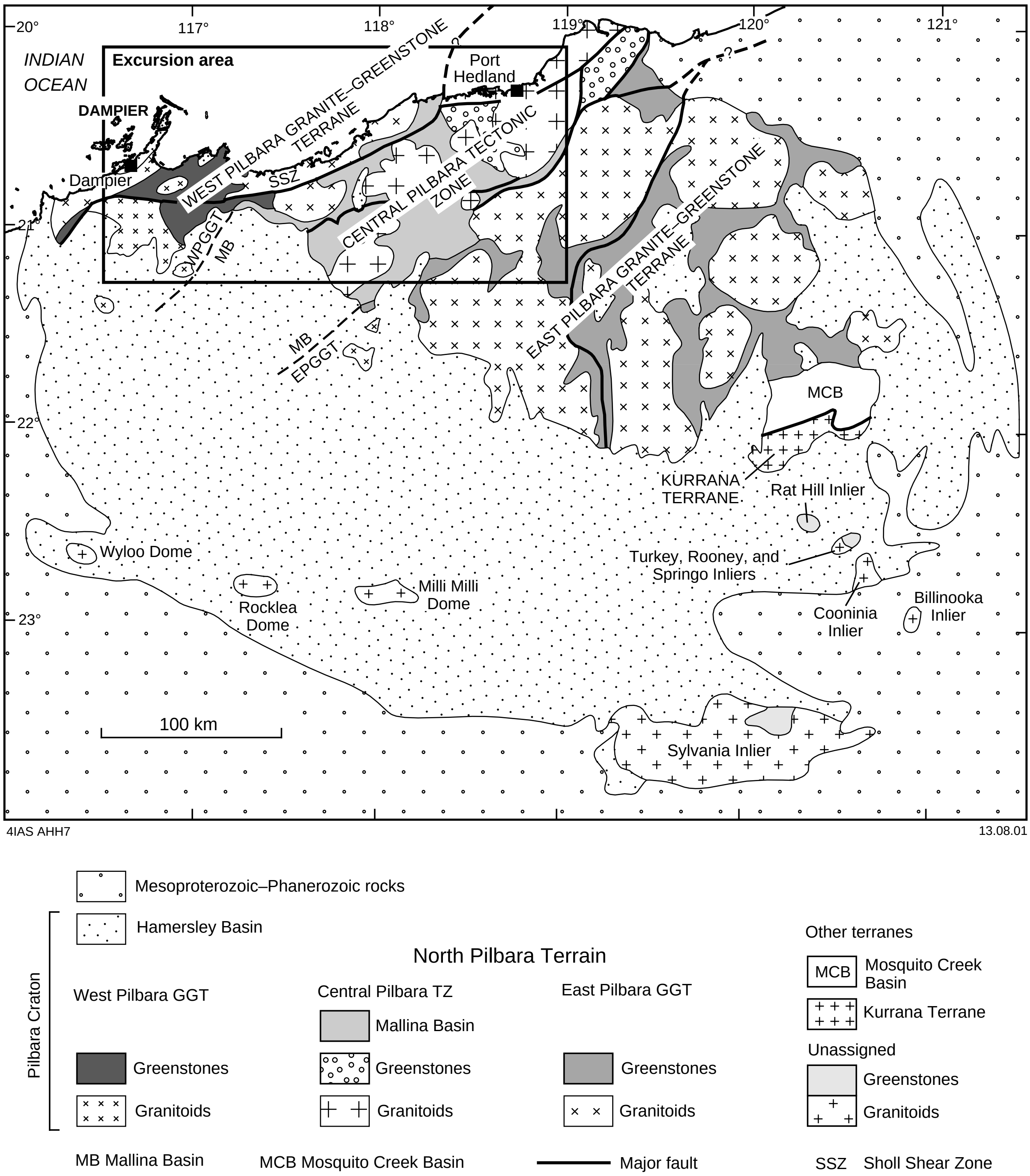


Figure 1. Regional geological setting of the West Pilbara Granite–Greenstone Terrane and the Mallina Basin

Overview of the West Pilbara Granite–Greenstone Terrane

There are major stratigraphic differences between the West Pilbara Granite–Greenstone Terrane and the EPGGT. In particular, the oldest preserved rocks of the WPGGT are 3270 Ma, whereas 3660–3590 Ma gneisses, 3480–3300 Ma granitoids, and 3515–3325 Ma greenstones form parts of the EPGGT. The east Pilbara contains no equivalents of the 3130–3115 Ma Whundo Group, the c. 3010 Ma Whim Creek Group, the c. 2975–2950 Ma Bookingarra Group, or the 3130–3060 Ma granitoids of the west Pilbara, but both terranes contain granitoids spanning 3000–2920 Ma.

The two terranes also exhibit different tectonic styles. The WPGGT is characterized by northeasterly trending granitoid complexes and greenstone belts and by numerous closely spaced easterly and northeasterly striking faults. Granitoids in this area are structurally relatively simple and discordant to the greenstone stratigraphy. In contrast, the EPGGT is dominated by granitoid domes that are separated by faulted, synformal greenstone belts with no preferred structural trend. Granite–greenstone contacts in the EPGGT are generally sheared, and parallel to adjacent greenstone stratigraphy.

Definition of the terrane

The regional setting of the WPGGT is shown in Figure 1, and its major tectono-stratigraphic components are summarized in Figure 2. This definition follows Hickman et al. (2000) and Hickman (2001), and differs from a recent tectonic subdivision proposed by Krapez and Eisenlohr (1998). Whether the WPGGT is a single terrane or actually comprises two terranes depends on the interpretation of the Sholl Shear Zone. Krapez and Eisenlohr (1998) and Smith et al. (1998) argued that the Sholl Shear Zone (Fig. 2) is a terrane boundary because it juxtaposes the 3270–3250 Ma Roebourne Group (interpreted by them as an island-arc succession) with the 3130–3115 Ma Whundo Group (interpreted as a back-arc succession). The distributions of the Roebourne and Whundo Groups are shown in Figure 3. Smith et al. (1998) interpreted the Roebourne Group and granitoids north of the Sholl Shear Zone to be parts of an allochthonous terrane, although they considered this to be indigenous to the North Pilbara Terrain.

Sholl Shear Zone

The Sholl Shear Zone bisects the WPGGT (Fig. 2), and is a major structural break with a long history of displacement and reactivation (Table 1). Over most of its length (at least 250–350 km, depending on geophysical interpretation) it is a nearly vertical zone of mylonite and schist 1000–2000 m wide. Smith et al. (1998) placed the timing of the movement on the shear zone at 2991–2925 Ma, whereas Krapez and Eisenlohr (1998) placed it at 3000–2955 Ma. Both these interpretations advocated sinistral movement, but evidence from geological mapping has shown that post-3020 Ma movement was dextral. Smithies (1998a) has mapped dextral displacement of the c. 3010 Ma Whim Creek Group north of Whim Creek, and Hickman (2001) has mapped dextral displacement of c. 2925 Ma layered intrusions southwest of Roebourne. Dextral strike-slip movement in both these areas can be measured as 30–40 km. Other evidence (Hickman et al., 2000; Hickman, 2001) indicates that dextral movement was late in the history of the Sholl Shear Zone, and minor compared to earlier sinistral movement that may have been 150–200 km.

The present interpretation is that the Sholl Shear Zone developed as one of several major faults within a northeasterly trending belt of rifting and strike-slip faulting on

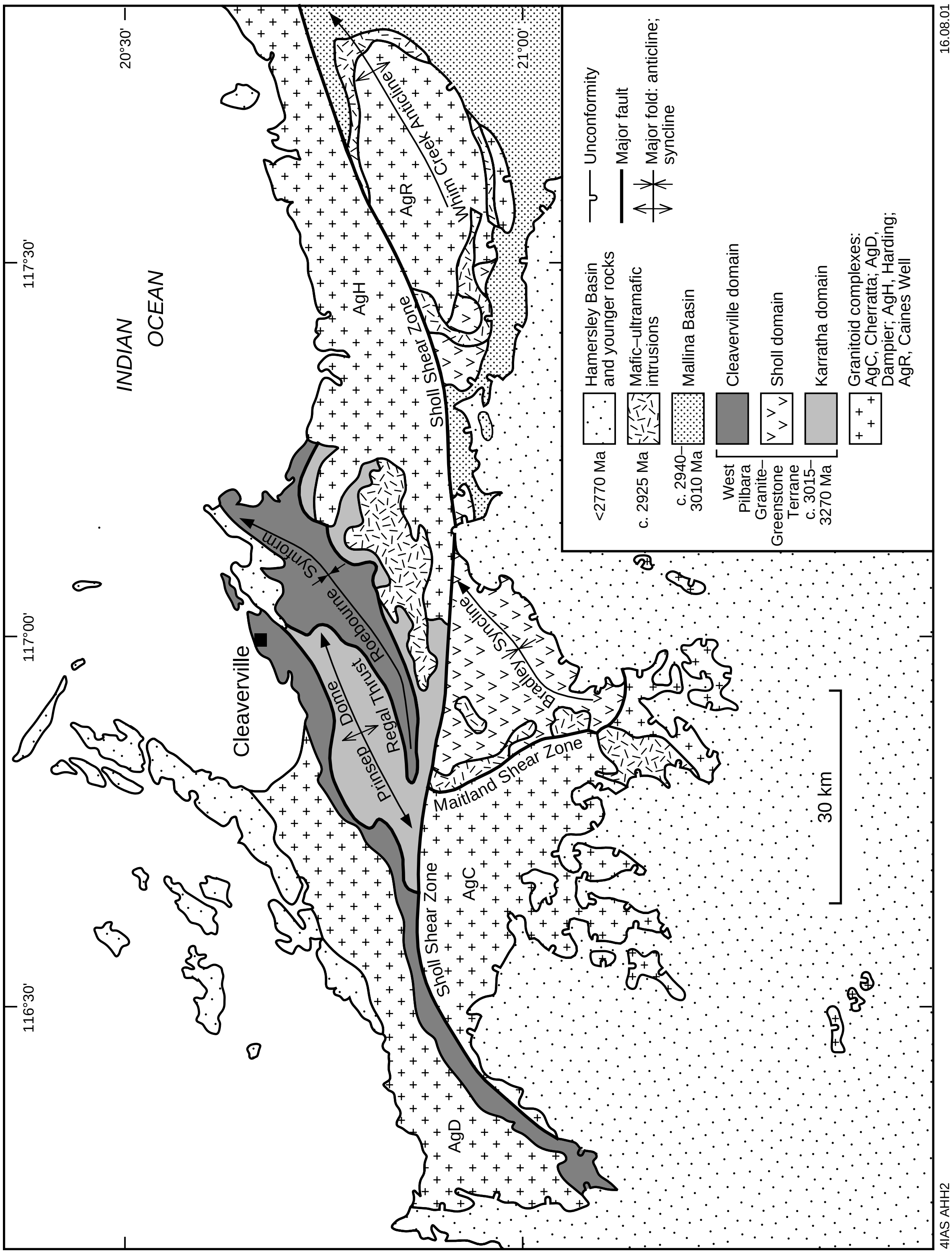


Figure 2. Tectono-stratigraphic domains of the West Pilbara Granite–Greenstone Terrane, showing major structures

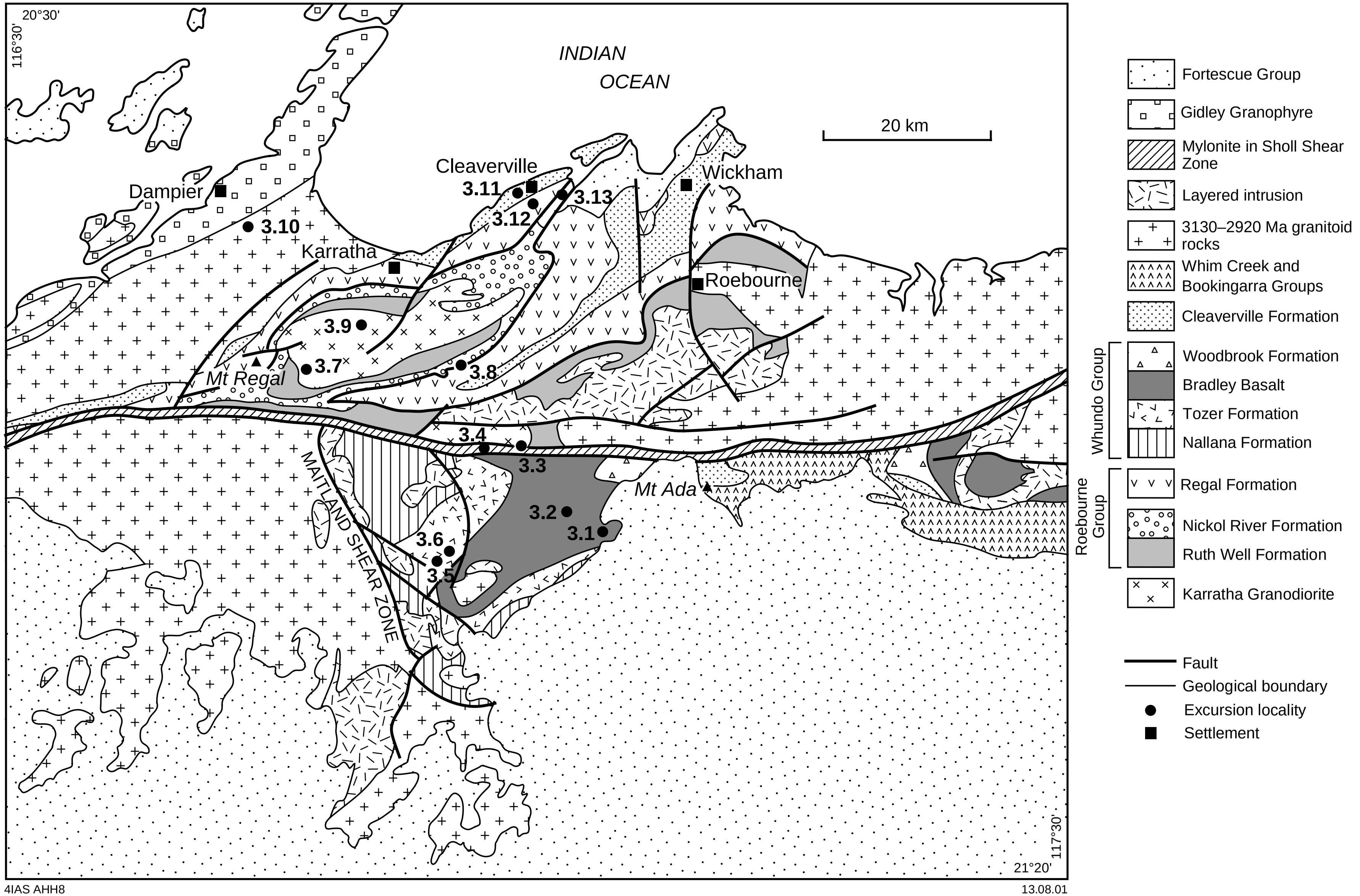


Figure 3. Simplified geological map of the Roebourne–Dampier area. Excursion localities 3.1–3.13 are also indicated

Table 1. Summary of the geological history of the West Pilbara Granite–Greenstone Terrane, the Mallina Basin, and adjacent parts of the Hamersley Basin

Age (Ma)	*Geological event*
3724–3310	Formation of EPGGT sialic crust (not exposed in WPGGT)
3300–3250	Rifting on the northwestern margin of the EPGGT (possibly related to a subduction zone to the northwest); deposition of the Roebourne Group (possibly in an island arc), and intrusion of granitoids
3160–3090	D_1: Thrusting, recumbent folding and c. 3160 Ma granitoid intrusion; deposition of the Whundo Group in a rifted zone southeast of the Roebourne Group (possibly an arc environment); intrusion of c. 3100 Ma granitoids; sinistral movement on the Sholl Shear Zone
3070–3050	Intrusion of tonalite
3070–3020	D_2: Culmination of sinistral strike-slip movement along the Sholl Shear Zone; upright tight to isoclinal transpressional folding and felsic magmatism; erosion
3020–3015	Deposition of the Cleaverville Formation
3015–3010	D_3: Strike-slip movement, felsic magmatism, and transpressional folding of the Whundo Group and Cleaverville Formation; erosion
3010–2990	Deposition of the Whim Creek Group (as redefined by Pike and Cas, in prep.) in the early Mallina Basin, and extensive intrusion of granitoids in the WPGGT
2975–2955	Deposition of De Grey Group and Bookingarra Group in Mallina Basin
2975–2955	D_4: Local thrusting and E–W folding of rocks in the Mallina Basin
c. 2955	D_5: N–S folding of rocks in the Mallina Basin; erosion
2955–2945	Intrusion of Peawah Granodiorite and Portree Granitoid Complex into rocks of the Mallina Basin
2945–2940	Renewed extension across the Mallina Basin and redeposition within the Mallina Basin
2950–2930	$D_{6:}$ Transpressional, northeasterly trending tight to open folding, and commencement of dextral movement along the Sholl Shear Zone; late- to post-tectonic felsic magmatism
c. 2940	D_7: Strike-slip movement on the Maitland Shear Zone, and along faults within the Cherratta Granitoid Complex
c. 2925	Emplacement of layered mafic–ultramafic intrusions, followed by intrusion of granite
c. 2920	D_8: Dextral strike-slip movement along the Sholl Shear Zone, and other east–west and northeasterly striking faults
<2920	D_9: Conjugate faulting produced by north-northwest – south-southeast compression
2920–2770	Erosion
2770–2750	D_{10}: Rifting and deposition of the Mount Roe Basalt and Hardey Formation; intrusion of dolerite dykes
755	Intrusion of northeasterly trending dolerite dykes
545–65	Palaeozoic and Mesozoic erosion
55–present	Uplift and dissection of plateau surface, deposition of Cainozoic units

the northwestern margin of the EPGGT. This fault system evolved into the Central Pilbara Tectonic Zone, and from about 3015 Ma controlled deposition of the Mallina Basin. Less clear at present is the relationship of this zone to volcanism of the c. 3130–3115 Ma Whundo Group, and to intrusion of 3160–3060 Ma granitoids that are restricted to the Central Pilbara Tectonic Zone.

Crustal differences between rocks north and south of the Sholl Shear Zone are indicated by Sm–Nd data (Sun and Hickman, 1998). Additional Nd-isotopic data (Hickman et al., in prep.) support reworking of 3500–3300 Ma crustal material north of the shear zone, but indicate that to the south only crust younger than c. 3300 Ma ever existed in the area of the Central Pilbara Tectonic Zone. This suggests that EPGGT crust, or crust of similar age, underlies the northern area, but must be thin, or even absent, in the Central Pilbara Tectonic Zone. Separation of the pre-Roebourne Group crust north of the Sholl Shear Zone from the main part of the EPGGT may thus have commenced at about 3300 Ma, which significantly coincides with a period of major deformation and felsic magmatism in the EPGGT.

In the above scenario the Sholl Shear Zone, as it currently exists, would coincide with a post-3300 Ma terrane boundary. This conclusion is similar to interpretations by Smith et al. (1998) and Sun and Hickman (1998). However, if the Central Pilbara Tectonic Zone as currently mapped (Fig. 1) is not underlain by crust older than 3300 Ma (current Sm–Nd data) separation of the WPGGT and EPGGT must have been primarily due to rifting. Thus, rifting rather than strike-slip faulting could have juxtaposed the Roebourne and Whundo Groups, and must have been well advanced at 3130 Ma. Abrupt changes in metamorphic grade across the Sholl Shear Zone provide evidence of several kilometres of relative upthrow on its northern side. Therefore, it remains possible that the 3130–3115 Ma Whundo Group (low metamorphic grade) once extended north of the shear zone but has been removed by erosion. This would date the terrane boundary as pre-3130 Ma, and the 3270–3250 Ma Roebourne Group and the Karratha Granodiorite would be the only remnants of the old terranes.

At least 80% of the rocks now separated by the Sholl Shear Zone were formed in their current relative positions. The granitoid complexes on each side of the Sholl Shear Zone were all intruded at about 3015–2940 Ma, and post-3010 Ma relative displacement was no more than 40 km. The WPGGT is here regarded as an essentially coherent geological unit that developed across, and largely concealed, the components of two older terranes.

Tectono-stratigraphic domains

The WPGGT is now divided into four granitoid complexes and three tectono-stratigraphic domains (Fig. 2). Previously, Krapez and Eisenlohr (1998) recognized two tectono-stratigraphic domains (their Domains 5 and 6) separated by the Sholl Shear Zone. Each domain was subdivided into various tectono-stratigraphic units such as granitoid complexes and basins. The interpretation by Krapez and Eisenlohr (1998) was based on previous mapping and a number of local studies such as those of Kiyokawa (1993), in the Cleaverville area, and Smith et al. (1998), on the Sholl Shear Zone. Information from recent mapping, and accompanying geochronology (Nelson, 1996, 1997, 1998), indicates that there are various errors and omissions in the interpretation by Krapez and Eisenlohr (1998), particularly with respect to ‘Domain 5’(Hickman et al., in prep.). The present tectono-stratigraphic subdivision follows that used by Hickman (2001).

The tectono-stratigraphic divisions of the WPGGT are shown in Figure 2. The four granitoid complexes (Dampier, Cherratta, Harding, and Caines Well) are readily identified, but understanding the structure and stratigraphy of greenstone successions has been more difficult. Mapping in the Dampier area (Hickman, 1997; Hickman, 2001) included the recognition of the Regal Thrust, an early (D_1) layer-parallel shear zone that separates the Nickol River Formation from the Regal Formation over an area measuring 70 km east-northeast – west-southwest, and 25 km north-northwest – south-southeast. This thrust separates the Karratha and Cleaverville domains (Fig. 2) and is folded by the Prinsep Dome. Further east the Regal Thrust is folded by the Roebourne Synform. The Cleaverville domain is underlain by the Regal Thrust, and bounded to the south by the Sholl Shear Zone. Stratigraphic components are the Regal Formation and the c. 3020–3015 Ma Cleaverville Formation. The age of the Regal Formation has not been directly established, but is assumed to be similar to that of the Ruth Well and Nickol River Formations. Its contact with the Cleaverville Formation is probably an unconformity, and sandstone or finer grained clastic sedimentary rocks lie at the base of the Cleaverville Formation. Sun and Hickman (1999) found the chemistry of the Regal

Formation to be like mid-ocean ridge basalt (MORB), which is inconsistent with a normal stratigraphic position above clastic sedimentary rocks of the Nickol River Formation. They suggested an explanation of this anomaly might be that the formation was obducted onto the Karratha domain. In this case, the Regal Thrust would be the base of the obducted slab. Very poor exposure obscures the relationship between the Dampier Granitoid Complex and the Regal Formation, but granite veins in the Regal Formation indicate an intrusive relationship.

Near Cleaverville, the Cleaverville domain is structurally more complex, with repetition of the Regal and Cleaverville Formations due to faulting and tight to isoclinal folding. Ohta et al. (1996) interpreted the Cleaverville area as part of an accretionary complex related to subduction of oceanic crust. However, this interpretation conflicts with evidence that the Cleaverville Formation is a shallow-water deposit (Sugitani et al., 1998). Additionally, the Cleaverville Formation is distributed over a wide area of the WPGGT (Hickman, 1997), and has recently been identified 200 km to the east, on the western margin of the EPGGT (Smithies and Farrell, 2000).

At least part of the faulting within the Cleaverville domain post-dates the Fortescue Group, and is probably related to a northeasterly trending belt of reverse faults (D_{11}) that deforms the c. 2725 Ma Gidley Granophyre.

To the south of the Sholl Shear Zone, the Maitland Shear Zone (Hickman, 1997) separates the Cherratta Granitoid Complex from the Sholl domain (Fig. 2). The major layered mafic–ultramafic intrusions of the west Pilbara cross the Maitland Shear Zone without visible deformation, indicating that the zone pre-dates 2925 Ma; additionally, it is truncated by the Sholl Shear Zone. The Sholl domain is mainly composed of the 3125–3115 Ma Whundo Group, but the Cleaverville Formation is also a component in the east.

Lithostratigraphy

Greenstones of the WPGGT were deposited over 250 m.y. in three unconformity-bounded groups. Deposition of the c. 3270–3250 Ma Roebourne Group, composed of two volcanic formations separated by a dominantly sedimentary formation (Table 2), was partly synchronous with intrusion of an underlying sheet of granodiorite and tonalite (Karratha Granodiorite). Rhyolite and dacite in the central part of the Roebourne Group are interpreted to be volcanic rocks related to the granodiorite. The volcanic Whundo Group (Table 2) was deposited at 3130–3115 Ma, but this unit has no preserved stratigraphic contacts with the Roebourne Group. The c. 3020 Ma Cleaverville Formation disconformably or unconformably overlies the Roebourne Group and the Whundo Group, and is assigned to the Gorge Creek Group. This stratigraphic correlation with the lithostratigraphy of the EPGGT is based on apparent continuity of banded iron-formation and chert units between the western margin of the EPGGT and the Gorge Creek Group in areas such as Wodgina, the Ord Ranges, and possibly Goldsworthy. Until such time as the banded iron-formations in the Gorge Creek Group of the EPGGT have been dated this correlation will remain tentative.

Sequence stratigraphy

Horwitz and Krapez (1991) reinterpreted the greenstone successions of the Pilbara Craton using sequence stratigraphy. Krapez (1993) provided a more detailed account of this new stratigraphic scheme, which included a division of the Pilbara granite–greenstone terrane into five tectono-stratigraphic domains. Krapez and Eisenlohr (1998) modified the sequence stratigraphy and domain interpretation of Krapez (1993), and presented a tectono-stratigraphic interpretation of the west Pilbara based on a hypothesis

Table 2. Archaean lithostratigraphy of the Roebourne area

Group	*Formation*	*Thickness (m)*	*Lithology and relationships*
Gorge Creek	Cleaverville Formation	1 500	Banded iron-formation, chert, fine-grained clastic sedimentary rocks. Age c. 3020 Ma
~ *Possible low-angle unconformity* ~			
Whundo	Woodbrook Formation	1 000	Rhyolite tuff and agglomerate; minor basalt and thin banded iron-formation. Age 3117 ± 3 Ma
	Bradley Basalt	>4 000	Pillow basalt, massive basalt, minor units of felsic tuff and chert. Age 3115 ± 5 Ma
	Tozer Formation	2 500	Calc-alkaline volcanics, including felsic pyroclastic units. Minor chert and thin banded iron-formation. Age c. 3120 Ma
	Nallana Formation	2 000	Dominantly basalt, but includes minor ultramafic and felsic units. Felsic tuff dated at 3125 ± 4 Ma. Base of formation intruded by 3130 Ma granitoids and truncated by Maitland Shear Zone
~ *Tectonic contact along Sholl Shear Zone* ~			
Roebourne	Regal Formation	2 000	Basal peridotitic komatiite overlain by pillow basalt and local chert units. Intruded by micro-granite and c. 3015 Ma felsic porphyry
~ *Tectonized contact along Regal Thrust* ~			
	Nickol River Formation	100–500	Banded chert, banded iron-formation, ferruginous clastic sedimentary rocks, quartzite, felsic volcanic rocks, carbonate rocks, and volcanogenic sedimentary rocks and local conglomerate. Schist protolith younger than 3269 ± 2 Ma, and rhyolite dated at 3251 ± 6 Ma
	Ruth Well Formation	1 000–2 000	Basalt and extrusive peridotite with thin chert units. Intruded by granodiorite and tonalite dated at 3270 ± 2 Ma

of global tectonic cycles. In the north Pilbara two ‘Megacycle Sets’, spanning 3500–2775 Ma, are divided into four ‘Megacycles’, each of 190–175 Ma duration. Each ‘Megacycle’ is inferred to contain a ‘Megasequence’ that can be divided into supersequences or basins. The sequence stratigraphy proposed by Krapez and Eisenlohr (1998) has been examined, but is inconsistent with data obtained from mapping and geochronology (Smithies et al., 1999; Hickman et al., in prep.), and therefore has not been applied. Comparisons between the lithostratigraphy and sequence stratigraphy of the two interpretations are shown in Table 3.

Roebourne Group

Table 2 summarizes the lithostratigraphy of the Roebourne Group (Hickman, 1997). Field relations and geochronological data have established that basal mafic and ultramafic metavolcanic rocks of the Ruth Well Formation are the oldest preserved components (Hickman, 1997). These rocks are undated, but must be older than c. 3270–3250 Ma, the age of conformably overlying felsic metavolcanic and metasedimentary rocks of the Nickol River Formation and coeval, intrusive granitoid rocks of the Karratha Granodiorite (Nelson, 1998; Smith et al., 1998; Hickman, 1999). The Karratha Granodiorite crystallized at 3270–3260 Ma, but Nd-isotope depleted-

Table 3. Comparison between lithostratigraphy (Hickman, 1997) and sequence stratigraphy (Krapez and Eisenlohr, 1998) in the Western Pilbara Granite–Greenstone Terrane

Lithostratigraphy		*Sequence stratigraphy*	
Isotopic age (Ma)	*Formation*	*Supracrustal unit*	*Inferred age (Ma)*
3020–3015	Cleaverville Formation		
c. 3115	Woodbrook Formation	Negri Basin	2954–2906
c. 3115 c. 3120 c. 3125	Bradley Basalt Tozer Formation Nallana Formation	Sholl Basin	3136–3089
		Lizard Hills Formation	3227–3180
not determined	Regal Formation	Cleaverville, Regal, and Lydia mine units	3270–3227
3270–3250	Nickol River Formation	Nickol Well unit	c. 3325
c. 3270	Ruth Well Formation	Ruth Well unit	c. 3325

mantle (Nd T_{DM}) model ages are 3480–3430 Ma (Sun and Hickman, 1998), typical of the Warrawoona Group in the EPGGT. The Ruth Well Formation includes serpentinized peridotitic komatiite, talc–chlorite schist, metabasalt, grey-and-white banded chert, and black chert. The overlying Nickol River Formation contains quartzite, ferruginous chert, grey-and-white banded chert, and metamorphosed carbonate sedimentary rocks, fine-grained clastic sedimentary rocks, conglomerate, and felsic volcanic rocks.

Peridotitic komatiite, metabasalt, and chert of the Regal Formation overlie the Nickol River Formation in the central and western parts of the WPGGT, but the contact between the two formations is invariably tectonized. Thus, the Regal Formation may not belong to the Roebourne Group (Hickman, 1997), but stratigraphic redefinition has been deferred pending precise geochronological data on the Regal Formation. As noted above, Sun and Hickman (1999) suggested that the Regal Formation was obducted onto the Nickol River and Ruth Well Formations. Hickman et al. (2000) named the regional tectonic contact the Regal Thrust, and noted that this structure can be traced over a large area of the WPGGT; however, the precise amount of lateral movement involved remains unknown.

Whundo Group

The Whundo Group (Hickman, 1997) is a 10 km-thick succession of mafic and felsic metavolcanic rocks that outcrops south of the Sholl Shear Zone between the Cherratta and Caines Well Granitoid Complexes. Geochronology on felsic volcanic rocks in the central and upper formations of the group indicates an age range of at least 3125–3115 Ma (Horwitz and Pidgeon, 1993; Hickman, 1997; Nelson, 1997, 1998). However, the base of the Nallana Formation is intruded by c. 3130 Ma granodiorite of the Cherratta Granitoid Complex (Nelson, 1999; Hickman et al., in prep.), establishing that the lowermost sections of the Whundo Group pre-date 3130 Ma. The Nallana Formation is 2000 m thick, and comprises metabasalt, ultramafic rocks, intermediate pyroclastic rocks, and sills of dolerite. The Tozer Formation conformably overlies the Nallana Formation, and in most areas is about 2500 m in thickness. Its succession consists of metamorphosed basalt, andesite, dacite, rhyolite, and thin metasedimentary units including chert and banded iron-formation. The most homogeneous formation of the Whundo Group is the Bradley Basalt, which exceeds 4000 m in thickness, and conformably overlies the Tozer Formation. The bulk of the succession consists of massive and pillowed basalt, but spinifex-textured high-Mg basalt is present near its base, and units of felsic tuff are intercalated with basalt and dolerite sills in the upper part of the formation. A felsic tuff 1000 m above the base of the Bradley Basalt was dated at 3115 ± 5 Ma (Nelson, 1996). Units of rhyolite tuff and agglomerate in the uppermost 1000 m of the Whundo Group are assigned to the Woodbrook Formation (Hickman, 1997). Nelson (1998) dated a welded tuff of this formation at 3117 ± 3 Ma, supporting a conformable relationship to the underlying Bradley Basalt. The stratigraphic top of the Whundo Group is concealed by the Fortescue Group.

Gorge Creek Group

The Cleaverville Formation (Ryan and Kriewaldt, 1964) is composed of banded iron-formation, ferruginous chert, grey-white and black chert, shale, siltstone, and minor volcanogenic sedimentary rocks. North of Roebourne, and in the Mount Ada area (15 km south of Roebourne), the formation is intruded by dolerite sills. Clastic sedimentary rocks at the base of the Cleaverville Formation suggest an erosional contact on pillow basalt of the underlying Regal Formation, but an angular unconformity has not been recognized. The depositional age of the Cleaverville Formation is closely constrained by detrital zircons dated at 3018 ± 3 Ma (Nelson, 1998), and by intrusive granophyre dated at 3014 ± 6 Ma (Nelson, 1997). Samples of the Cleaverville Formation also contain detrital zircons with near concordant ages of c. 3287–3236 Ma, and one zircon dated at 3461 ± 8 Ma (Nelson, 1998). The former population was probably derived by erosion of the Karratha Granodiorite, but the 3461 ± 8 Ma age is very significant in that it coincides with the age of the Warrawoona Group in the EPGGT. This suggests that at 3020 Ma EPGGT crust may have been exposed in the Roebourne area.

Granitoids

After intrusion of the Karratha Granodiorite, the greenstones of the WPGGT were intruded by granitoids in two other periods, one spanning 3160–3060 Ma, and a second taking place between 3015 and 2940 Ma. The major granitoid complexes of the WPGGT were probably formed by successive intrusion of sheets and laccoliths beneath the greenstone succession.

Karratha Granodiorite

The Karratha Granodiorite is the oldest identified granitoid unit of the WPGGT. Uranium–lead zircon geochronology (Nelson, 1998; Smith et al., 1998) has established that its components crystallized at 3270–3260 Ma, and Sm–Nd isotopic analyses have given Nd T_{DM} model ages of 3480–3430 Ma (Sun and Hickman, 1998). The c. 200 Ma difference between the emplacement age and the Nd T_{DM} model ages indicates that magma generation involved older crust or enriched lithospheric mantle. One of the samples (JS17) dated by Smith et al. (1998) contained near-concordant zircon cores with $^{207}Pb/^{206}Pb$ ages up to 3311 Ma, supporting involvement of older crust. The Karratha Granodiorite ranges from equigranular tonalite to granodiorite.

Cherratta Granitoid Complex

The Cherratta Granitoid Complex is bounded to the north by the Sholl Shear Zone, and to the southwest it is unconformably overlain by the Fortescue Group. To the east the Maitland Shear Zone separates it from the Whundo Group, although small stocks of monzogranite and syenogranite assigned to the complex locally intrude the greenstones. The oldest granitoid components of the complex (3150–3060 Ma) are restricted to the area east of the Munni Munni Intrusion. However, small inliers of the Cherratta Granitoid Complex in the valley of the Fortescue River expose 3236 Ma granitoids (Nelson, 1999). West of the Maitland River, aeromagnetic lineaments indicate east-northeasterly striking lithological zones, possibly including concealed large greenstone enclaves.

Banded grey granite–tonalite gneiss with xenoliths of amphibolite-facies mafic gneiss, and sheets of leucocratic gneiss, outcrop along the Maitland River, intruded by porphyritic granite and hornblende–biotite monzogranite. Gneiss dated at 2995 ± 11 Ma (Nelson, 1997) gave a Nd T_{DM} model age of 3246 Ma and samples of banded gneiss from other localities provided Nd T_{DM} model ages of 3139 and 3149 Ma (Sun, S.-S., 1997, written comm.). Sun and Hickman (1998) noted that these results were similar to Nd T_{DM} model ages obtained on the 3125–3115 Ma Whundo Group of the Sholl Belt, and in sharp contrast to 3430–3480 Ma Nd T_{DM} model ages that were obtained on the Karratha Granodiorite and greenstones of the Roebourne Group north of the Sholl Shear Zone. Other components of the Cherratta Granitoid Complex range from granite to granodiorite, and include hornblende-rich, and porphyritic granitoids.

Dampier Granitoid Complex

The Dampier Granitoid Complex is the most northwesterly granitoid complex of the Pilbara Craton and, on land, extends 100 km southwestwards from outcrops on the Burrup Peninsula. On the Burrup Peninsula the complex is intruded by the c. 2725 Ma Gidley Granophyre and, further northwest in the Dampier Archipelago, it is unconformably overlain by the Fortescue Group. On its southeastern side it intrudes the Regal Formation. Most of the complex consists of porphyritic granite–granodiorite and a more mixed assemblage of even-grained granite–granodiorite containing banded gneissic granitoids, syenogranite, and pegmatite. The composition of the complex is thus similar to that of the Cherratta Granitoid Complex, and limited geochronology (two samples dated) indicates that the age of the Dampier Granitoid Complex is about 2990 Ma (Nelson, 1998, 1999). Because post-3020 Ma strike-slip movement on the intervening Sholl Shear Zone was no more than 30–40 km, the two granitoid complexes were formed at the same time (c. 2990 Ma), and in the same part of the WPGGT. A granitoid dated at 2997 ± 3 Ma (Nelson, 1998) included zircon cores with $^{207}Pb/^{206}Pb$ ages of 3255–3106 Ma. The latter age is similar to the ages of the Karratha Granodiorite and the lower part of the Roebourne Group (3270–3250 Ma; Hickman, 1997), and it is possible that either or both contributed material to the complex. Neodymium T_{DM} model ages obtained on the Dampier Granitoid Complex have a range of 3387–3247 Ma (Hickman et al. in prep.).

Harding Granitoid Complex

The Harding Granitoid Complex is almost entirely concealed by Cainozoic alluvial and marine deposits, but aeromagnetic data indicate that it extends about 130 km in an east-northeasterly direction between Roebourne and Port Hedland. Bouguer anomalies (Blewett et al., 2000) suggest that the complex ranges from 40–80 km in width, including its offshore extent. About 15 km southwest and southeast of Roebourne outcrops of the complex are composed of weakly foliated to compositionally banded monzogranite and granodiorite. At its southern margin the Sholl Shear Zone is a 1 km-wide belt of mylonitized granitoids and tectonic lenses of sheared greenstones that separates the complex from the Whundo Group and the Whim Creek Group. A sample of the less deformed variety of monzogranite was dated at 2970 ± 5 Ma, and contained a population of xenocrystic zircons dated at 3018 ± 19 Ma (Nelson, 1999). The same rock provided a Nd T_{DM} model age of 3309 Ma (Hickman et al., in prep.). North of Whim Creek, and immediately north of the Sholl Shear Zone, a sample of granodiorite gneiss from the Forrestier Bay area was dated at 3014 ± 3 Ma (Nelson, 1997), and subsequently gave a Nd T_{DM} model age of 3276 Ma (Sun and Hickman, 1998). A granitoid sample (JS43) assigned to the complex by Smith et al. (1998) is now interpreted to be part of the Karratha Granodiorite.

Caines Well Granitoid Complex

The Caines Well Granitoid Complex lies in the far northeast of the WPGGT. It is in contact with rocks of the Whim Creek greenstone belt to the east, south, and west, and to the north, is truncated by the Sholl Shear Zone. Contacts with rocks of the Whim Creek greenstone belt are either unconformable or faulted. From geophysical data and limited outcrop Smithies (1998a) identified rocks of the granitoid complex and of the Whim Creek Group north of the Sholl Shear Zone. The indicated displacement along the shear zone is about 40 km in a dextral sense. In the south and east, mafic and ultramafic rocks of the Sherlock Intrusion were emplaced along, or close to, the contact between the complex and the rocks of the Whim Creek greenstone belt. Although the Caines Well Granitoid Complex resembles a dome-like structure, the complex forms a doubly plunging anticline produced by transpression.

Three dates from the complex indicate at least three ages of magmatism. Strongly foliated trondhjemite in the east of the complex has an intrusive age of 3093 ± 4 Ma (Nelson, 1997). This is only marginally younger than its Nd T_{DM} model age of 3118 Ma (Sun and Hickman, 1998) and may indicate derivation via partial melting of rocks of similar age to the Whundo Group. The main phase of the Caines Well Granitoid Complex appears to be a porphyritic monzogranite, dated at 2990 ± 5 Ma (Nelson, 2000). Minor amounts of leucocratic monzogranite adjacent to rocks of the Sherlock Intrusion have been dated at 2925 ± 4 Ma (Nelson, 1997).

Episodes of deformation

The tectonic evolution of the WPGGT is summarized in Table 1.

D_1 (3160–3130 Ma)

Blewett (2000) reported structures earlier than intrusion of the c. 3270 Ma Karratha Granodiorite, but these structures were not observed during mapping of the DAMPIER* 1:100 000 map sheet (Hickman, 2001). The existence of pre-3270 Ma structures requires that the granitoids observed by Blewett (2000) are part of the Karratha Granodiorite, and are not c. 3016–2970 Ma granitoids that also occur in the area. At the time of writing, the age of the structures described by Blewett (2000) remains uncertain.

The earliest tectonic structures in the WPGGT, recognized from mapping of DAMPIER (Hickman, 2001), are low-angle thrusts and recumbent folds, apparently produced by southerly directed thrusting. The largest fault is the Regal Thrust (see **Tectono-stratigraphic domains**), but large-scale thrusts also occur within the Regal Formation. Large-scale isoclinal folds in the Nickol River Formation southeast from Mount Regal were recumbent and southwesterly facing prior to tilting by the D_6 Prinsep Dome. Intrafolial isoclines are relatively common within the Nickol River Formation of the Mount Regal area (Hickman et al., 2000), and are probably minor structures related to

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

the recumbent folds. 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 Regal Thrust is exposed 13 km southeast of Karratha on the southern limb of the Prinsep Dome. Here, a finely laminated silicic mylonite has been isoclinally folded. Part of the outcrop (Locality 3.8) shows that the isoclinal folds have been refolded by later tight to isoclinal folds. Plunges of the early isoclines are generally low (up to 30°) east or west, and the prevailing dip of the mylonite is 60–80° south. This indicates thrusting from either the north or the south.

D_1 structures have not been recognized in the Whundo Group, suggesting that they formed prior to 3130 Ma. A thermotectonic event at 3160–3150 Ma (Smith et al., 1998) is indicated by zircon geochronology (Smith et al., 1998) and K–Ar geochronology (Kiyokawa, 1993), and is here interpreted to coincide with D_1. Granitoids of this age have been identified southeast of Dampier (Nelson, 1999) and close to the southeastern margin of the Central Pilbara Tectonic Zone (Nelson, in press), confirming a tectono-magmatic event at this time.

D_2 (mainly c. 3050–3020 Ma)

Measurable strike-slip movement on the Sholl Shear Zone is dextral, and about 30–40 km. However, the dextral movement was preceded by greater sinistral movement (see above). Sun and Hickman (1998) suggested that the c. 3270–3250 Ma Roebourne Group might be equivalent to the c. 3260–3235 Ma Sulphur Springs Group of the EPGGT.

Direct evidence for sinistral movement is provided by porphyroclasts within fine-scale mylonitic lamination (Hickman, in prep.). These shear-sense indicators consistently point to sinistral movement along the foliation planes of the mylonite. Evidence for major early movement is also provided by the stratigraphic mismatch across the fault (see **Tectono-stratigraphic domains**) that cannot be explained by the measurable 30–40 km of dextral displacement (D_8) that occurred after deposition of the Whim Creek Group.

D_3 (3015–3010 Ma)

Tight to isoclinal east-northeasterly trending folds in the Cleaverville Formation between Karratha and Cleaverville are attributed to the D_3 event. The same type of folding occurs in the Cleaverville Formation between Miaree Pool and Mount Regal. In the basalt, S_3 is locally synchronous with intrusive sheets of 3015 Ma quartz–feldspar porphyry and porphyritic microgranite.

At Mount Ada (15 km south of Roebourne), easterly trending, upright tight to isoclinal D_3 folds in the Cleaverville Formation contain a sill of 3014 Ma granophyre (Nelson, 1997), and these fold structures are unconformably overlain by the c. 3010 Ma Warambie Basalt.

D_4 (c. 2990–2960 Ma)

Thrusts in the Warambie Basalt east of Mount Ada (Hickman, in prep.) pre-date dextral movement along the Sholl Shear Zone, and may be equivalent to Phase 3 that Krapez and Eisenlohr (1998) recognized in the Whim Creek area. Originally easterly trending 'D_1 folds' on MOUNT WOHLER (Smithies, 1998b) and SATIRIST (Smithies and Farrell,

2000) may belong to the same event. These correlations suggest that D_4 occurred between 2990 and 2960 Ma.

D_5 (pre-2950 Ma)

Northerly trending folds recognized within the Central Pilbara Tectonic Zone (Smithies, 1998b; Smithies and Farrell, 2000) appear to have no equivalent structures in the WPGGT.

D_6 (c. 2950–2940 Ma)

The D_6 event formed major northeasterly trending tight to open folds such as the Prinsep Dome, Roebourne Synform, and Bradley Syncline (Fig. 2). These structures are chronologically correlated with major D_3 folds in the Mallina Basin (Smithies, 1998c), and are equivalent to Phase 4 structures described by Krapez and Eisenlohr (1998). However, geochronology in the Mallina Basin (Smithies, 1998b) establishes that the age of these structures must be 2950–2930 Ma, not 2906–2863 Ma as suggested by Krapez and Eisenlohr (1998, fig. 2). D_6 folds are oblique to the Sholl Shear Zone and to other strike-slip faults of the WPGGT and Central Pilbara Tectonic Zone. The folds are probably transpressional folds within a post-2950 Ma, easterly trending 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 and 1 km southwest from Nickol Well. The Mount Regal folds plunge southwest, and the Nickol Well folds plunge northeast.

D_7 (c. 2940 Ma)

The north-northwesterly striking Maitland Shear Zone 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 granitoids of the Cherratta Granitoid Complex. North of Bullock Hide Well all these structures are truncated by the Sholl Shear Zone. Shear zones also occur within the Cherratta Granitoid Complex where gneiss contains late zircon populations dated at 2944 ± 5 Ma and 2925 ± 2 Ma (Nelson, 1997).

D_8 (c. 2920 Ma)

The latest movement on the Sholl Shear Zone was dextral. South of Roebourne a subsidiary dextral strike-slip fault, the Black Hill Shear Zone, displaces the Andover Intrusion by 10 km (Fig. 3; Hickman, in prep.). As noted by Krapez 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 could have coincided with D_8.

Minor D_8 structures in the Sholl Shear Zone include dextral drag folding and isoclinal folding (Locality 3.8) of S_2 mylonite lamination, and associated small-scale faulting and brecciation.

D_9 (<2920 Ma)

The Sholl Shear Zone and earlier structures are deformed by a conjugate system of north-northeasterly striking sinistral faults and west-northwesterly striking dextral faults

(Hickman, 2001). The precise age of D_9 faults is unknown, but well-developed conjugate structures have not been observed in the Fortescue Group.

Metamorphism

Greenstones of the Roebourne Group north of the Sholl Shear Zone have been metamorphosed to amphibolite facies, whereas south of the shear zone the Whundo Group contains rocks at lower greenschist facies. An exception to the above generalization is in the Cleaverville domain between Karratha and Cleaverville. Here, downward movement on steep faults on the northwestern side of the Prinsep Dome has preserved greenschist-facies sections of the Regal Formation. The metamorphic change across the Sholl Shear Zone is attributed to upward movement on its northern side, although the timing of this is uncertain.

The granitoid complexes contain greenstone enclaves that are metamorphosed to amphibolite facies, and the granitoids show evidence of retrogression from amphibolite facies.

Overview of the Mallina Basin

The Mallina Basin is a key tectonic component of the granite–greenstone terrain of the Archaean Pilbara Craton and developed over the boundary between the EPGGT and the WPGGT (Smithies et al., 1999; Van Kranendonk et al., in prep.). The preserved portion of the basin extends in a northeasterly direction for over 150 km (Fig. 1). It includes siliciclastic rocks of the Constantine Sandstone and Mallina Formation, which Hickman (1977) and Eriksson (1982) considered to be of turbiditic origin, and rocks of the Whim Creek greenstone belt, which form the northwestern portion of the basin (Fig. 4). Fitton et al. (1975) placed the bulk of the stratigraphic succession of the Whim Creek greenstone belt into the Whim Creek Group. Pike and Cas (in press) presented evidence that this succussion represents two distinct depositional and intrusive cycles, separated by a disconformity that accounts for up to 40 m.y., and refer to these two sequences as the (redefined) Whim Creek Group and the overlying Bookingarra Group (Table 4). In the discussion below, all references to the Whim Creek Group are to the redefined succession of Pike and Cas (in press).

Controversy exists over the location and nature of the northern boundary of the basin, the age of the Mallina Basin, and, in particular, on how the siliciclastic turbidite succession that forms the Mallina Formation and underlying Constantine Sandstone relate to the volcaniclastic successions that lie in the Whim Creek greenstone belt (Fig. 4).

The Loudens Fault forms the contact between the Whim Creek greenstone belt and rocks of the Mallina Formation (Fig. 4) and has been variously described as either a minor and late feature that separates two facies-equivalent successions (Fitton et al., 1975; Horwitz, 1990; Smithies et al., 1999) or as a domain boundary (Krapez, 1993; Eriksson et al., 1994; Krapez and Eisenlohr, 1998).

According to Barley (1987) and Krapez and Eisenlohr (1998), rocks of the Whim Creek greenstone belt accumulated in a c. 2990 Ma pull-apart basin. The Mallina Basin was thought to contain only the Mallina Formation and Constantine Sandstone, and was inferred to have accumulated between 3090 and 3045 Ma in a hybrid retro-arc and remnant-ocean basin, as a result of eastward subduction of oceanic crust towards the Pilbara Craton. However, Smithies et al., (1999) suggested that the outcrop extent of the Whim Creek Group was bounded by two independent fault sets that do not define a pull-apart structure (Fig. 5) and that none of these faults represented a domain

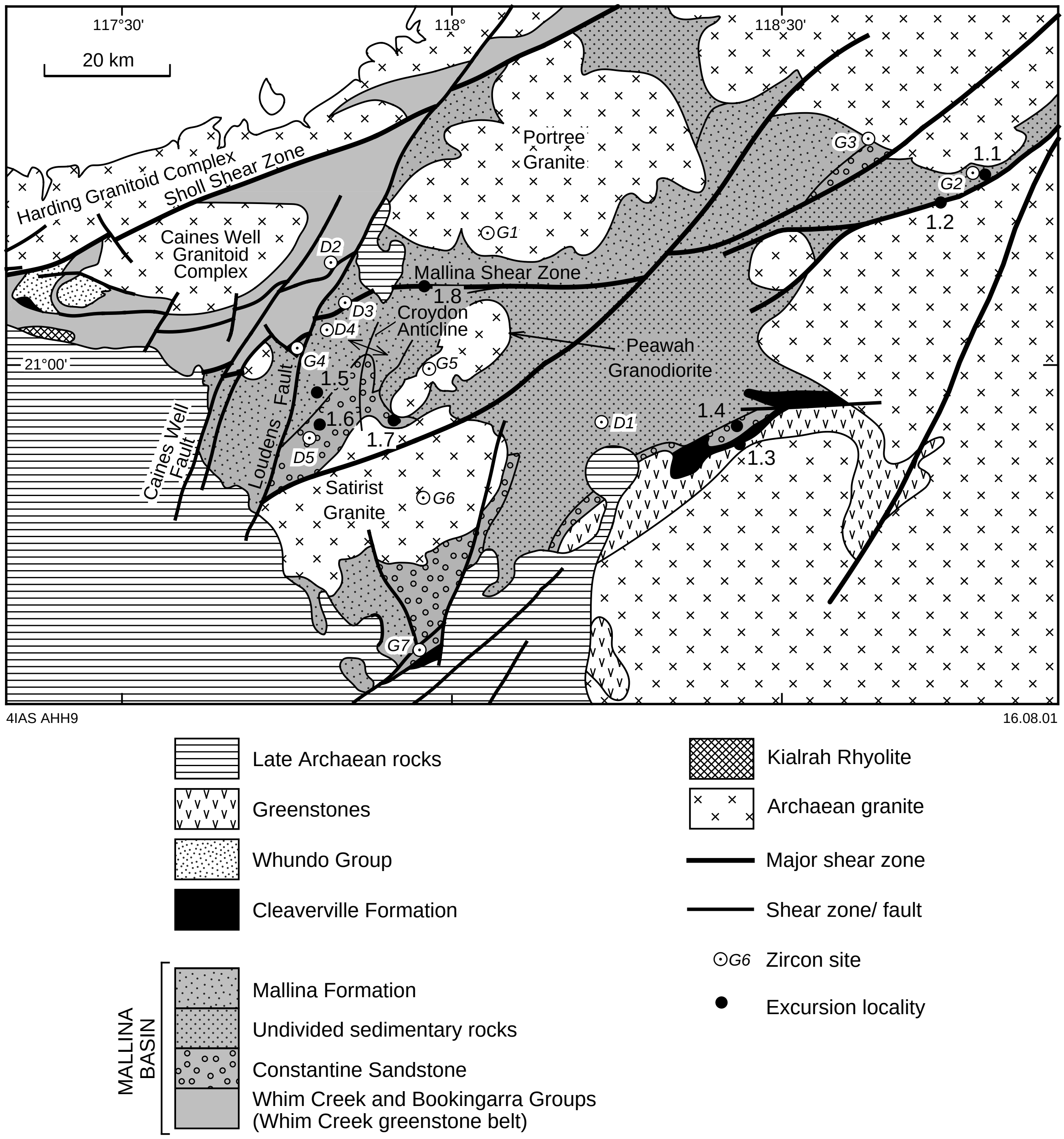


Figure 4. Simplified geological map of the Mallina Basin, showing major lithological subdivisions and excursion localities 1.1–1.8. See Figure 7. Modified from Smithies et al. (2001)

Table 4. Lithostratigraphy of the Whim Creek belt, with maximum preserved thicknesses

Group	*Formation*	*Thickness (m)*	*Description*
Bookingarra Group	Louden Volcanics	2 000	Extensive pillow lava and sheet lava. Commonly exhibits pyroxene-spinifex textures. Interbedded with quartzite and shale
	Negri Volcanics		Extensive sheet lava with some pillow breccia and rare, subaerial basaltic spatter breccia
	Rushall Slate	300	Laminated shale with sandstone interbeds. Derived from a source rich in mafic rocks and granite
	Cistern Formation	800	Conglomerate/breccia to sandstone with upward-decreasing volcaniclastic debris and grain size. Rare pumice
Whim Creek Group	Mons Cupri dacite	160	Dacitic intrusions. Largest intrusion about 250 km^3, typically much smaller (<1 km^3). Rare peperitic and flow-banded margins
	Red Hill volcanics	290	Andesitic sandstone and breccia. Dense volcaniclastic material with monomictic tube-pumice turbidite (3009 ± 4 Ma)
	Warambie Basalt	200	Basaltic lava with autobreccia and hyalo-clastite. Limited epiclastic breccia and granite-derived turbidite sandstone

boundary. Furthermore, SHRIMP U–Pb zircon dates have been obtained on detrital grains from rocks forming the basement to the Mallina Basin, on detrital grains from rocks within the basin, from granitoids that have intruded rocks of the Mallina Basin, and on zircon xenocrysts within those granitoids (Nelson, 1997, 1998, 1999, 2000). According to Smithies et al., (1999, 2001), these data indicate that the Constantine Sandstone and Mallina Formation are younger than c. 2970 Ma, that they are at least partially equivalent in age to the Bookingarra Group, and that there is a systematic variation in the age ranges of detrital zircons, that is not affected by the Loudens Fault. These interpretations support the earlier arguments of Fitton et al. (1975) and Horwitz (1990), that the Mallina Formation and rocks of the Whim Creek greenstone belt are simply lateral facies equivalents within the same Mallina Basin.

Local geology

The southeastern margin of the Mallina Basin is marked by an unconformity on a unit of chert. A maximum depositional age of 3016 ± 13 Ma was obtained on detrital zircons from a sample of volcanolithic sandstone taken from beds located conformably beneath this basement chert (Nelson, 1998; Smithies et al., 1999). This age is consistent with a correlation of the chert with the Cleaverville Formation exposed further to the northwest. The northern margin of the basin ranges from an intrusive contact with younger (c. 2990–2925 Ma) phases of the Caines Well Granitoid Complex (Pike and Cas, in press) to a faulted unconformity between the rocks of the Whim Creek Group and older rocks of the Whundo Group, the Cleaverville Formation, and the oldest (c. 3100 Ma) phase of the Caines Well Granitoid Complex. Figure 6 summarizes the relationship between the stratigraphy of the Mallina Basin and deformation and magmatic events.

Constantine Sandstone and Mallina Formation

Historically, the rocks in the south and east of the Mallina Basin have been assigned (Fitton et al., 1975; Hickman, 1990) to two formations; either the Mallina Formation (comprising interbedded, well-graded, fine- to medium-grained wacke and shale), or

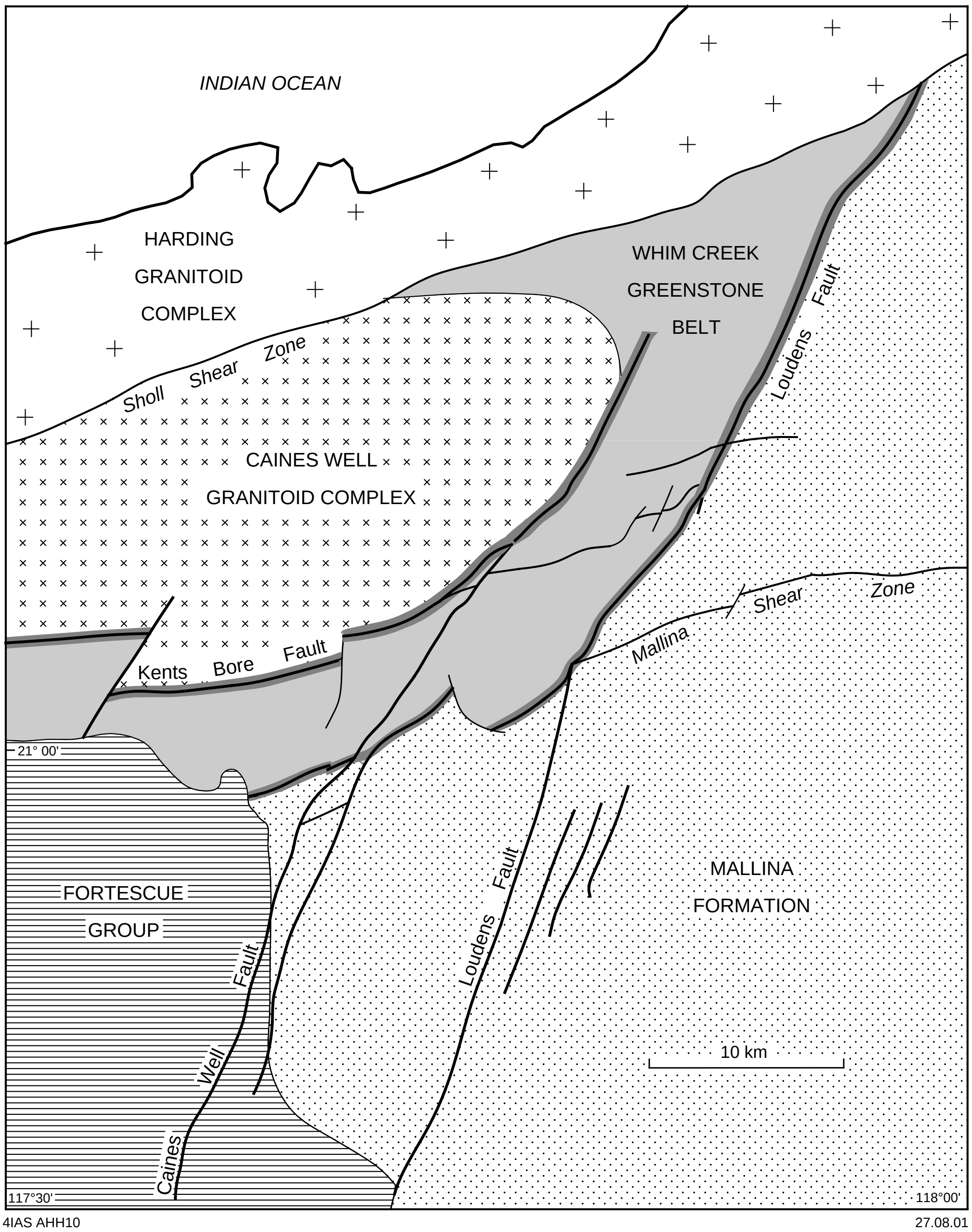


Figure 5. Simplified geological map of the northwestern part of the Mallina Basin, showing independent fault-sets defining the margins of the Whim Creek greenstone belt ('boundary' faults highlighted by thick shade)

the underlying Constantine Sandstone (comprising medium- to coarse-grained, poorly sorted subarkose to wacke, locally with thick conglomerate layers). Fine- and coarse-grained facies are locally interbedded throughout the succession and such units may be difficult to assign to either the Constantine Sandstone or the Mallina Formation. The Constantine Sandstone has an estimated thickness of up to 3.5 km (Smithies, 1998b) whereas estimates of the thickness of the Mallina Formation vary between 2.5 km (Fitton et al., 1975) and 10 km (Miller, 1975). Both units were probably deposited as submarine fans (Hickman, 1977; Eriksson, 1982), with the coarse-grained Constantine Sandstone, locally including chert-cobble conglomerate, representing generally proximal, or upper fan, depositional environments characterized by high rates of

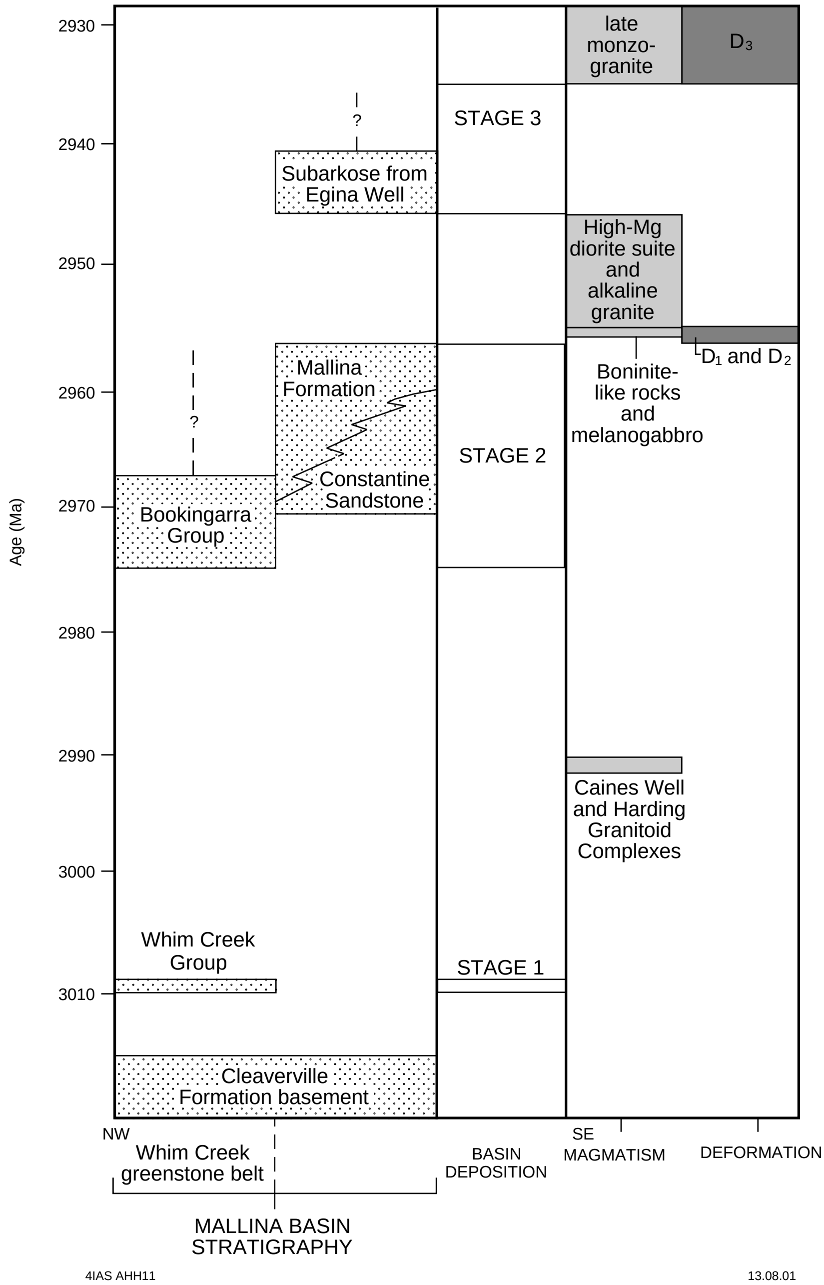


Figure 6. Schematic representation of the stratigraphy of the Mallina Basin and the timing of magmatic and deformational events (modified from Smithies et al., 2001)

sediment deposition. The Constantine Sandstone is not always developed, and in areas distal from the source of sediment influx, wacke and shale that closely resembles the Mallina Formation forms the stratigraphically lowest unit of the basin succession.

Age of the Constantine Sandstone and Mallina Formation

Smithies et al. (1999) presented SHRIMP U–Pb zircon dates on detrital zircon grains extracted from an arkose of the Mallina Formation. Most grains gave ages older than c. 3140 Ma, but three out of 31 concordant or near-concordant analyses gave a combined age of 2997 ± 20 Ma. Detrital zircon grains extracted from a subarkose of the Constantine Sandstone included 15 (out of 31) concordant or near-concordant analyses that gave a combined age of 2994 ± 4 Ma (Nelson, 2000). These ages are interpreted as the ages of source components for the host sedimentary rocks, and thus the maximum depositional ages of the host rocks (Smithies et al., 1999; Nelson, 2000). However, many of the granitoids that have intruded the Mallina Basin (see below) have zircon xenocrysts that are dated at c. 2970 Ma (Fig. 7; Nelson, 1997, 1999, 2000). These grains must be derived from rocks that are the same age as, or older than, the clastic rocks into which the granitoids intruded, and the most reasonable interpretation is that they were derived from clastic rocks in the lower parts of the basin. If this is the case, the c. 2970 Ma zircons provide a maximum age for deposition of the Constantine Sandstone and Mallina Formation (Smithies et al., 2001).

A population of four detrital zircons obtained from a wacke within the southeastern part of the Mallina Basin indicates a maximum depositional age of 2941 ± 9 Ma (Nelson, 1999). The sample was presumed to be from the Mallina Formation. However, the age is interpreted as a maximum depositional age for the host sedimentary rock and so the sample cannot be from the Mallina Formation as this formation was folded during D_2 (see below), prior to the intrusion of granitoids, the oldest of which has been dated at 2954 ± 4 Ma (Nelson, 1999). Therefore, these detrital zircons may suggest the presence of an unrecognized unconformity and a late second depositional cycle. Furthermore, the remaining detrital zircons within this sample do not fall into distinct age groups but, rather, show a wide scatter of ages spanning the ranges shown by detrital zircon populations from the other siliciclastic samples from the Mallina Basin (Fig. 7). This suggests that the wacke was derived via reworking of the Mallina Formation and Constantine Sandstone.

Deformation

Rocks of the Constantine Sandstone and Mallina Formation have been folded during three deformation events (Smithies, 1998b; Smithies et al., 1999). The first event resulted in open east-trending folds that are poorly preserved. The second event produced large north- to northeast-trending folds that have exposed the Constantine Sandstone in their cores. The third deformation event resulted in the dominant east-northeast structural fabric of the region. Numerous east-northeast shear zones, such as the Mallina Shear Zone, are thought to reflect reactivation of early basin-developing faults within the basement, with dominantly south-side-up movement (Smithies, 1998b).

The age of these deformation events is constrained by their relationship to dated episodes of granite magmatism. Granitoids with intrusive ages between c. 2955 and 2945 Ma have truncated structures related to the second phase of deformation, suggesting that this deformation event pre-dates 2954 Ma. The same granitoids are affected by the third phase of deformation, which must post-date c. 2945 Ma. Younger granites, which intruded the sedimentary rocks of the Mallina Basin between c. 2940 and 2930 Ma, are early to syntectonic with respect to the third phase of deformation. Many of these granites show a strong flow alignment of K-feldspar megacrysts that parallels either the pluton margins or the structural trend of the third phase of deformation (Smithies, 1998b). Flow alignment is locally overprinted by a weak schistosity developed during this third phase of deformation (Smithies, 1998b).

Whim Creek and Bookingarra Groups

The Whim Creek greenstone belt forms the northwestern part of the Mallina Basin and includes rocks of the Whim Creek Group and the overlying Bookingarra Group (Fig. 8, Table 4; Pike and Cas, in press). However, all contacts between these rocks and those of the Mallina Formation and Constantine Sandstone are faulted, either along the Mallina Shear Zone or the northeast-trending Loudens Fault.

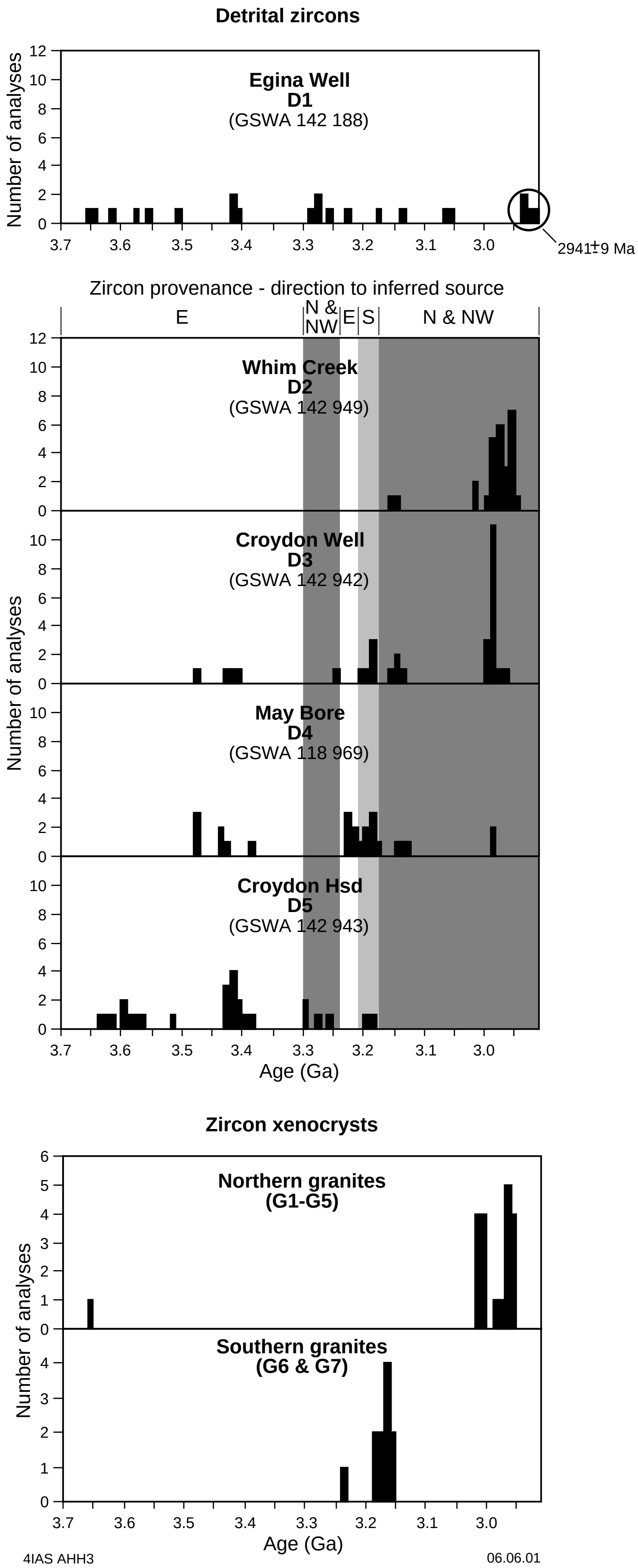


Figure 7. Age distribution histograms for SHRIMP U–Pb analyses of detrital zircons from sedimentary rock samples from the Mallina Basin. Sample localities are shown on Figure 4 (modified from Smithies et al., 2001)

The Whim Creek Group has recently been studied in detail by Pike and Cas (in press) and Pike et al. (in press). It contains a complex association of coeval felsic to mafic volcanic, intrusive, and volcaniclastic rocks. Voluminous, juvenile, felsic volcaniclastic material containing abundant pumice fragments in a glass-shard matrix was deposited by turbidity currents and debris flows, with limited palaeocurrent data suggesting a source to the south. These rocks were transported into the depositional basin and are petrographically and geochemically distinct from associated, bimodal felsic and basaltic volcanic and intrusive rocks (Pike et al., in press). The volcaniclastic rocks are interleaved with subaqueous basaltic lavas and breccia, and the entire succession is intruded by a voluminous, synsedimentary, dacite–rhyodacite sill that includes apophyses into the overlying volcaniclastic rocks. Smaller dacite–rhyodacite intrusions are also noted and typically form elongate bodies up to 2000 m long and 200 m thick (Pike et al., in press). Fitton et al. (1975) first referred to the mafic volcanic

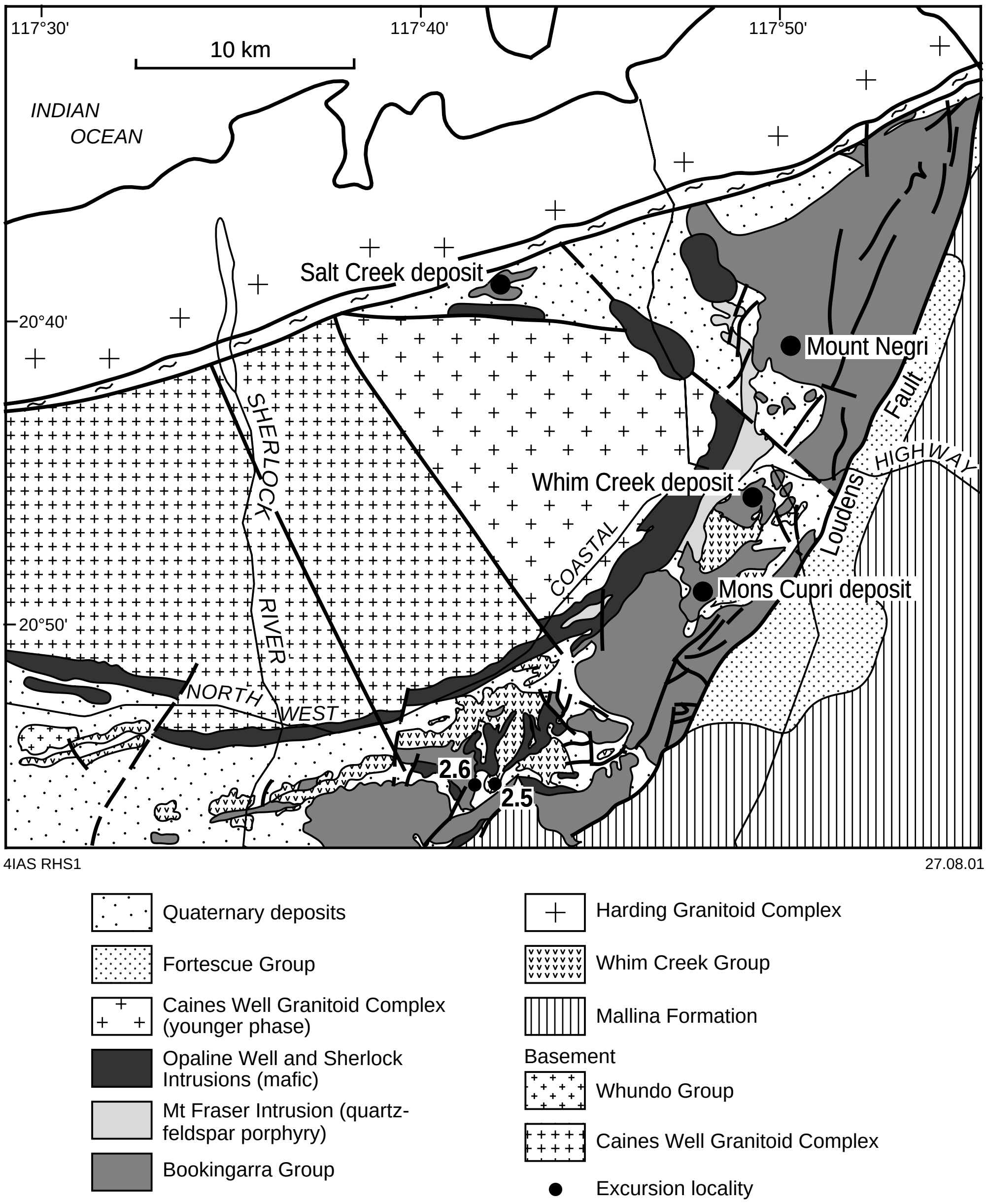


Figure 8. Simplified geological map of the Whim Creek greenstone belt, showing distribution of the Whim Creek and Bookingarra Groups and excursion localities 2.5 and 2.6

and volcaniclastic rocks as the Warambie Basalt and to the felsic rocks as the Mons Cupri Volcanics. Pike and Cas (in press) have subdivided the Mons Cupri Volcanics into the Red Hill volcanics, comprising juvenile, felsic volcaniclastic rocks, and the Mons Cupri dacite, comprising a synsedimentary, dacite–rhyodacite sill and associated smaller intrusions.

A sample of juvenile felsic pumiceous volcaniclastic rocks from the Red Hill volcanics gave a SHRIMP U–Pb zircon age of 3009 ± 4 (Nelson, 1998), which was interpreted as the age of felsic volcanism; that is, synvolcanic redeposition (Nelson, 1998; Smithies et al., 1999; Pike et al., in prep.). The rock contains up to 70% large, angular pumice fragments. Pike et al. (in prep.) also reported that these fragments lie in a glass-shard matrix. It is most likely that the rocks formed by redeposition of unconsolidated material.

The Bookingarra Group contains abundant volcaniclastic material derived from the underlying Whim Creek Group. The Cistern Formation forms the lowest unit and shows an upward increase in non-volcanogenic siliciclastic material (Pike and Cas, in press) and a concomitant decrease in grain size, culminating in the deposition of the Rushall Slate. The Rushall Slate is overlain by locally spinifex-textured and variolitic siliceous high-Mg basalts, which Hickman (1977) subdivided into the Louden Volcanics and the Mount Negri Volcanics. Contacts between the basalts and the volcaniclastic and siliciclastic rocks are generally conformable (Smithies, 1998a) or form a low-angle unconformity (Hickman, 1997). Interbedded sandstone, shale, and basaltic conglomerate near Mount Negri suggest volcanic quiescence during deposition of the Cistern Formation and Rushall Slate. Pike and Cas (in press) recognized peperite-like contacts, where basalts have locally intruded unlithified sedimentary material, indicating that basaltic volcanism and clastic deposition in the Bookingarra Group overlapped in time. For this reason, the Louden Volcanics and the Mount Negri Volcanics are included within the Bookingarra Group.

In the southwestern part of the Whim Creek greenstone belt, the Kialrah Rhyolite overlies the Louden Volcanics and has a maximum depositional age of 2975 ± 4 Ma (Nelson, 1998; Smithies et al., in press). Detrital zircons from a dominantly siliciclastic unit within the Bookingarra Group (Cistern Formation) also provide a maximum depositional age of c. 2975 Ma. Sulfide mineralization within the upper part of the Whim Creek Group has a Pb-isotopic model age of 2948–2942 Ma, which provides a minimum depositional age for that part of the group, irrespective of whether the mineralization is interpreted as syngenetic or epigenetic (Huston et al., 2000). This depositional age range is identical to that for the Constantine Sandstone and Mallina Formation, in the southern part of the Mallina Basin.

Major shear zones

Mallina Shear Zone

The Mallina Shear Zone runs east-northeast along the axis of the Mallina Basin. The shear zone consists of very low grade, intensely folded and locally sheared metasedimentary rocks of the Mallina Basin. The observed structure is interpreted to be the upper crustal expression of a reactivated structure in the basement and can be observed on Landsat Thematic Mapper (TM) and aeromagnetic images.

The subvertical shear zone records south-side-up movement. The structures are best developed in the wackes. In shaly horizons the structures are flat lying, in the more competent wackes they form ramps and display south-side-up displacement. This phase of deformation has also produced the main generation of open east-northeasterly trending folds in the Mallina Basin which are poorly preserved.

The steeply south-dipping structures are overprinted by several phases of strike-slip activity of the Mallina Shear Zone. This deformation phase locally produced dextrally verging folds and dextral shears within the shear zone. The folds are at millimetre to metre scale and their fold axes plunge steeply to the southwest. Locally a sinistral shear sense has been observed, indicating a complex history of reactivation of the Mallina Shear Zone.

Tabba Tabba Shear Zone

The Tabba Tabba Shear Zone forms the eastern margin of the Mallina Basin. The rocks within the shear are generally well exposed between the Marble Bar road in the northeast and the Great Northern Highway in the southwest. Aeromagnetic images show the northeastern extension of the shear zone, but the structure cannot clearly be traced to the southwest beneath the Mallina Basin.

The structures in the Tabba Tabba Shear Zone record three deformation phases: an early dextral phase, followed by a major oblique sinistral phase, and a minor late dextral phase. A regional foliation that rotates into the shear zone with a dextral geometry at kilometre scale is taken as evidence for the early dextral phase. This foliation is only observed on the southeast side of the Tabba Tabba Shear Zone.

During the main oblique sinistral phase the northwest block moved down relative to the southeast block. Lenses of sedimentary rocks belonging to the Mallina Basin have been caught up in the Tabba Tabba Shear Zone and metamorphic grade increases from greenschist facies on the northwest side of the shear zone to amphibolite facies on the southeast side. Deformation resulted in the formation of moderately northwest-plunging mineral and stretching lineations in the southwestern section of the shear zone and subvertical lineations in the northern section. This geometry suggests the steep, generally northwest-dipping Tabba Tabba Shear Zone had a normal sense of displacement during this event. A minor dextral fabric overprints the main foliation in the Tabba Tabba Shear Zone.

During the second phase of deformation the Tabba Tabba Shear Zone was intruded by early gabbros and a later suite of more evolved gabbros, and by dioritic to granodioritic rocks of the Pilbara high-Mg diorite suite (see **Intrusive and extrusive rocks**). All of these intrusions are of mantle origin and, at least in the case of the high-Mg diorite suite, were emplaced into extensional jogs created during shearing. Dating of the high-Mg diorite suite constrains the age of the main phase of shearing to c. 2950 Ma, consistent with a Nd T_{DM} model age of c. 2970 Ma from a mylonitized granite to the southeast of the shear zone (Champion, D. C., 2000, written comm.). There are a number of lenses of breccia within the shear zone. The breccia contains clasts with olivine-cumulate texture together with clasts of mafic schist, and is rich in graphite. Such features suggest a diatreme-like origin for the rock, and the presence of olivine and graphite indicate the Tabba Tabba Shear Zone reached at least as deep as the upper mantle.

Zircon provenance

In the absence of sufficient palaeocurrent data, Smithies et al. (2001) speculated about the provenance of the zircon populations based upon current knowledge of age distributions within the Pilbara Craton (Fig. 7). Detritus interpreted to have been derived from the south, north, northwest, and east is present in all samples. Interestingly, there is a progressive northerly increase in the contribution of young (<3020 Ma) detritus from an inferred north and northwesterly source, and the detrital population sampled from the Bookingarra Group forms an end-member of that progression (Fig. 7). This is consistent with the Bookingarra Group forming part of the Mallina Basin succession.

Intrusive and extrusive rocks

The Whim Creek greenstone belt was intruded by high-level felsic magmas of the Mons Cupri dacite at c. 3010 Ma and by quartz–feldspar porphyry at c. 2970 Ma or younger. Compositions of these rocks range from dacite to rhyolite, with features more typical of calk-alkaline (medium- to high-K) than Archaean tonalite–trondhjemite–granodiorite (TTG) magmatism (Fig. 9). The geochemistry of the Mons Cupri dacite is characterized by enrichment in Th and Ce coupled with depletion in Sr, Ta, Nb, TiO_2, P_2O_5 and shows strong enrichment of LILE[1] over HFSE[2] (Fig. 9). Normalized REE[3] data also suggest calc-alkaline compositions with fractionated LREE, flat HREE and negative Eu anomalies (Fig. 9).

Within the Cistern Formation, rare sand- to granule-sized bubble-pumice clasts record a felsic volcanic event at c. 2975 Ma or younger. This suggests that the Bookingarra Group formed in a regional setting in which volcanic processes were taking place. However, it is not clear if this volcanic activity was related to the development of the Mallina Basin.

Mafic volcanism within the Whim Creek greenstone belt includes the c. 3010 Ma Warambie Basalt at the base of the Whim Creek Group, and the Louden and Mount Negri Volcanics, which form the upper part of the Bookingarra Group. The Warambie Basalt is tholeiitic, whereas the Louden and Mount Negri Volcanics are siliceous high-Mg basalts (Sun et al., 1989). All of these rocks show strong enrichments in incompatible trace elements (Sun et al., 1989). Siliceous high-Mg basalts (SHMB) combine basaltic to andesitic compositions and high MgO contents (up to 16 wt%) with enriched LILE and LREE patterns (Fig. 10). Their origin has been linked to crustal contamination of komatiitic magmas, with which they are commonly associated (Arndt and Jenner 1986; Sun et al., 1989). However, rocks with >18 wt% MgO are rare in the Louden Volcanics, and many of these are cumulate rocks. Sun et al. (1989) also noted that the Louden Volcanics has anomalously low Ti (<0.5 wt%), Sc (<30 ppm), and V (<170 ppm) contents for Archaean SHMB, and new data (Smithies, R. H., unpublished data) shows them to have low normalized Gd/Lu ratios (~1). On the basis of such compositional features, Sun et al. (1989) suggested that partial melting of a subduction-enriched, refractory harzburgitic source was possibly a more reasonable model for the Louden Volcanics, although they preferred contamination of komatiitic magmas as a general model for SHMB.

Two compositionally distinct basaltic units are present within the southern and central parts of the Mallina Basin. The stratigraphically lowest unit is interbedded with rocks of the Constantine Sandstone, and is inferred to lie within the stratigraphically lowest portions of the basin succession. It is not clear if these rocks are of extrusive or high-level intrusive origin. They are tholeiitic to high-Mg basalts, characterized by strongly fractionated LREE and HREE patterns (Fig. 10). These patterns require derivation from a refractory mantle source, but also require LREE (and LILE) enrichment, either of that source or at a later stage through crustal contamination. These rocks closely resemble modern boninites in composition, but occur in an intracontinental setting, temporally and spatially unrelated to subduction (Smithies, in prep.). The stratigraphically higher unit of basalt and high-Mg basalt are flows that represent the only clearly volcanic rocks so far identified within the southern and central parts of the Mallina Basin. They occur within rocks that appear transitional between those of the Constantine Sandstone and those of the Mallina Formation. Significantly, these volcanic rocks show many compositional similarities to the siliceous high-Mg

1 LILE = large ion lithophile element
2 HFSE = high field strength elements
3 REE = rare earth elements, including light REE (LREE) and heavy REE (HREE)

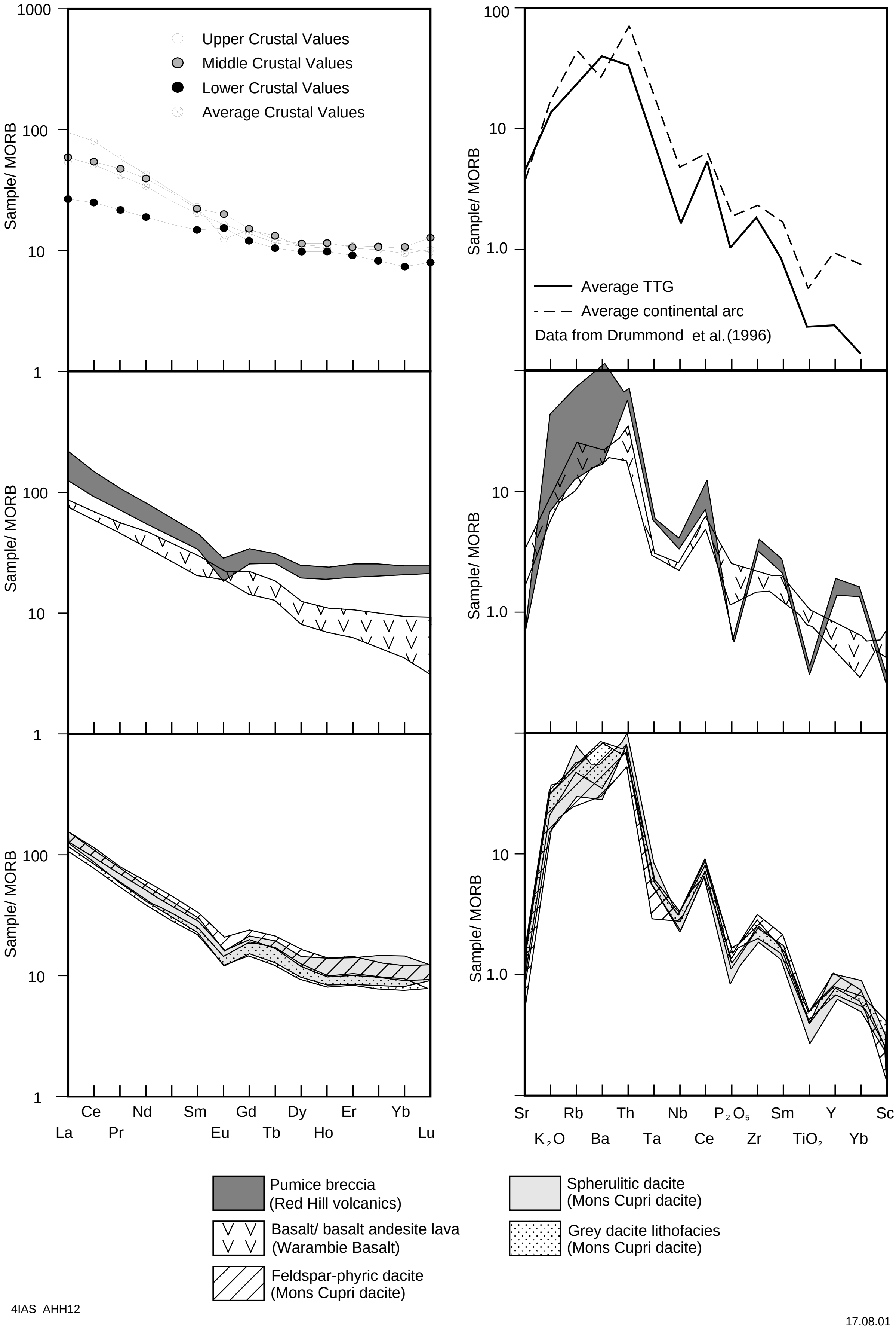


Figure 9. Geochemical plots of felsic volcanic and volcaniclastic rocks from the Whim Creek greenstone belt

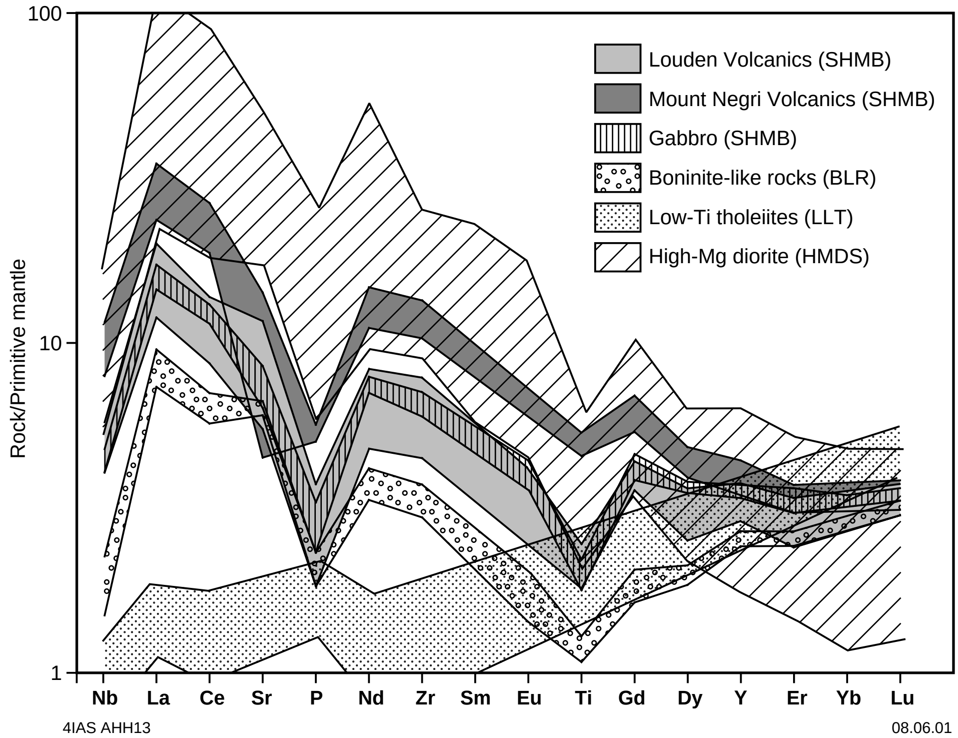


Figure 10. Primitive mantle-normalized trace-element plot of mafic rocks from the Mallina Basin

basalts (Sun et al., 1989) that comprise the Louden Volcanics, within the Bookingarra Group, including strong enrichments in incompatible trace elements such as Th, La, Ce, Zr, and moderate to low concentrations of Ti (Fig. 10; Smithies, R. H., unpublished data).

Three suites of layered mafic-ultramafic rocks have intruded the Mallina Basin. The layered Sherlock Intrusion ranges in composition from leucogabbro to pyroxenite, and has intruded the contact between the Caines Well Granite Complex and the Whim Creek greenstone belt and marginal sections of the granite complex, close to that contact. The layered intrusion contains extensive reserves of V and Ti.

The Opaline Well Intrusion of layered gabbro and olivine gabbro intrudes the Whim Creek greenstone belt. This intrusion was interpreted by Smithies (1998a) to be the subvolcanic equivalent of the Louden or Mount Negri Volcanics (or both). Similar gabbro sills intrude the Constantine Sandstone and Mallina Formation in the southern part of the basin, and have trace-element compositions that are virtually indistinguishable from the Louden Volcanics (Fig. 10).

The Millindinna Intrusion is a regionally extensive, but generally thin (<200 m but up to 400 m), sill that intrudes the Constantine Sandstone and Mallina Formation in the southern part of the Mallina Basin. This sill varies from lherzolite at the base to melanogabbro at the top. This layering persists over a minimum continuous strike distance of about 20 km. Outcrops of ultramafic rock in the central and eastern parts of the Mallina Basin may be tectonically dismembered components of this sill; however, they show no clear evidence for layering, and some show evidence for an extrusive origin. Both the Millindinna Intrusion and the gabbros that intrude the siliciclastic turbidites were emplaced between c. 2970 and 2950 Ma and, thus, may be similar in age to the Louden Volcanics.

A number of granitoids have intruded the Mallina Basin and have been dated at between 2954 ± 4 and 2931 ± 5 Ma (Fig. 6). These can be broadly divided into two groups based on intrusive age and on composition (Fig. 11). Those emplaced between c. 2955 and 2945 Ma belong to either the Pilbara high-Mg diorite suite or the Portree Granitoid Complex, are either in the compositional range of diorite–tonalite and granodiorite, or are alkaline granites, and are late- to post-tectonic with respect to the second phase of deformation to affect the rocks of the Mallina Basin. Younger granites, which intruded between c. 2940 and 2930 Ma, are K-feldspar porphyritic monzogranites or syenogranites, and are early to syntectonic with respect to the third phase of deformation (Smithies, 1998b). A small stock of K-feldspar porphyritic monzogranite, the Opaline Well Granite, lies adjacent to the intersection of the Mallina Shear Zone and the Loudens Fault. This monzogranite locally contains quartz–tourmaline orbicules, up to 10 cm in diameter. However, with an age of 2765 ± 5 Ma (Nelson, 1997), the Opaline Well Granite is significantly younger than the other monzogranites and is coeval with magmatism that formed the lower part of the Fortescue Group.

The 2946 ± 6 Ma (Nelson, 1999) Portree Granitoid Complex (Fig. 11) is a large nested plutonic complex of alkaline granite, which is locally hypersolvus and contains sodic pyroxene. Intrusion of the high-Mg diorite, or sanukitoid, suite was coeval with intrusion of the alkaline granite, with ages ranging between 2954 ± 4 and 2945 ± 6 Ma (Nelson, 2000). Intrusions of the high-Mg diorite suite are of hornblende–biotite (–clinopyroxene–orthopyroxene)-bearing rocks and are virtually confined to the Mallina Basin. They comprise a belt of intrusions that parallel the axis of the Mallina Basin, indicating a strong structural control (Fig. 11). Distinct aeromagnetic anomalies suggest the presence of a number of these intrusions within and adjacent to the major east-northeast trending Mallina Shear Zone. Low-pressure metamorphic aureoles, and geobarometric calculations constrain the depths of emplacement to less than 7 km (Smithies and Champion, 2000). Many of the intrusions are partially surrounded by earlier intrusions of gabbro. Most intrusions also contain abundant rounded enclaves, up to 30 cm in diameter. Most of these are cognate inclusions of diorite and gabbro. Some intrusions preserve a chilled margin of fine-grained melanodiorite, which is also present in 1–2 m-thick dykes and sills in country rock. The high-Mg diorite suite is analogous to the sanukitoid suite described by Shirey and Hanson (1984) and Stern et al. (1989) from the Superior Province of North America. The rocks are characterized by high Mg-number*, Cr, Ni, and LILE contents. At about 60 wt% SiO_2, high-Mg diorite has a Mg-number of about 60, about 200 ppm Cr, and about 100 ppm Ni, requiring a source significantly more mafic than typical Archaean basaltic crust; that is, a mantle source. High LILE concentrations cannot be explained through crustal contamination of a mafic magma, and indicate that the mantle source was LILE enriched, probably through an earlier episode of subduction (Shirey and Hanson, 1984; Stern et al., 1989; Smithies and Champion, 2000).

The c. 2940–2930 Ma K-feldspar porphyritic monzogranites and syenogranites that intruded the sedimentary rocks of the Mallina Basin represent remelting of earlier crust (Champion and Smithies, 1999). Intrusion of these granites was concentrated along the southern margin of the Mallina Basin, and systematically decreased in both age and abundance further to the southeast (Smithies and Champion, 2000).

Geological evolution and tectonic setting

The simplest interpretation of all the available data is that the Whim Creek Group, Bookingarra Group, Constantine Sandstone, and Mallina Formation are part of a single

* Mg-number = $Mg^{2+}/(Mg^{2+} + Fe_{Total}) \times 100$, with Fe_{Total} as Fe^{2+}

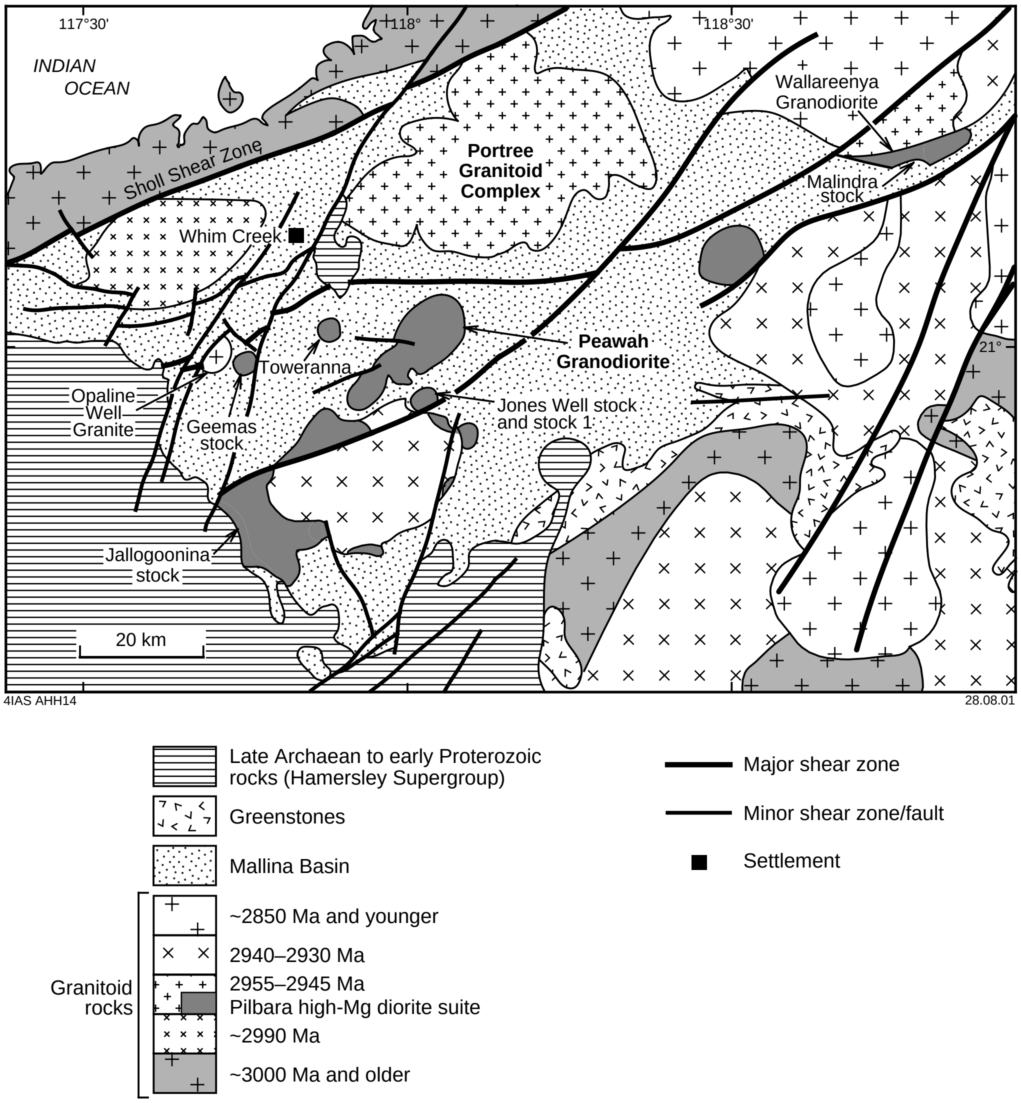


Figure 11. Simplified geological map of the Mallina Basin showing the distribution of granitoid rocks of various age groups

depositional basin — the Mallina Basin — that may include a series of stacked, sequential basins. The following section outlines, in point form, the evolution and tectonic setting of this basin.

- The early regional setting of the Mallina Basin was discussed previously (see **Overview of the WPGGT**). The basin developed between the EPGGT and the WPGGT in a zone that is believed to have been characterized by repeated rifting, compression, and magmatism, throughout the period from 3160 to 2920 Ma.
- The Whim Creek Group was deposited in the northwestern part of the Mallina Basin at c. 3010 Ma, on rocks that are c. 3020–3015 Ma and older. It is possible that c. 3015 Ma rocks that can be correlated with the Cleaverville Formation underlie most of the basin.
- There is no evidence from outcrop or from the ages of zircon xenocrysts in granitoids, that the rocks of the Whim Creek Group were deposited in the southern part of the Mallina Basin.
- The Whim Creek Group reflects bimodal, subaqueous volcanism with calc-alkaline characteristics. However, regional geological interpretations of the northwestern and

central parts of the Pilbara Craton (Krapez and Eisenlohr, 1998; Van Kranendonk et al., in prep.) have not invoked subduction at c. 3010 Ma. Consequently, the Whim Creek Group either reflects the only known preserved relic of c. 3010 Ma subduction, or it formed in an alternative setting such as an early rift (Pike and Cas, in press).

- A period of up to 40 m.y. separated deposition of the Whim Creek Group from deposition of the next oldest preserved sequences. The Bookingarra Group was deposited in the northern part of the Mallina Basin at the same time, or slightly before, deposition of the Constantine Sandstone. Deposition of the Cistern Formation and Rushall Slate is limited to kilometre-scale sub-basins (Fig. 8). These are overlain by the Mount Negri Volcanics and more widespread Louden Volcanics. The distribution of these rocks suggests that the Bookingarra Group records half-graben subsidence and sedimentary in-fill, followed by (fault) physical linkage of sub-basins and synchronous sedimentation and basaltic volcanism. Compared to the Bookingarra Group, the much greater thickness of the Constantine Sandstone and Mallina Formation (>10 km combined) suggests that the bulk of subsidence took place in the southern portion of the Mallina Basin.
- Strongly depleted basalts and high-Mg basalts in the Constantine Sandstone reflect melting of a strongly refractory mantle source, and require a large degree of lithospheric extension. These rocks have boninite-like compositions, although their tectonic setting differs from that of modern boninites.
- The Constantine Sandstone and Mallina Formation derived much of their detritus (including zircon grains) from a significantly older elevated land mass to the south and east. However, the proportion of this ‘old’ detritus contributing to rocks of Constantine Sandstone and Mallina Formation decreases systematically from the southern parts of the basin to the northwest. Rocks of the Bookingarra Group, in the far northwest of the Mallina Basin, derived most of their detritus from the underlying c. 3010 Ma Whim Creek Group and from c. 2990 Ma granites exposed to the northwest.
- There is no suggestion from regional geological interpretations (Krapez and Eisenlohr, 1998; Smithies and Champion, 2000; Van Kranendonk, et al., in prep.) that the Mallina Basin evolved at or near an active plate margin. Accordingly, we consider that the basin evolved in an intracontinental rift setting (Smithies et al., 1999; Smithies and Champion, 2000).
- Rocks of the Constantine Sandstone, Mallina Formation, and Bookingarra Group were deposited by c. 2955 Ma, at which stage they were deformed, possibly as a distal result of collision along the margins of the Pilbara Craton.
- High-temperature, lower crustal-derived (Portree Granitoid Complex) and mantle-derived (high-Mg diorite suite) felsic magmas intruded between c. 2955 and 2945 Ma. According to Smithies and Champion (2000), the high-Mg diorite suite of granitoids intruded zones of active dilation along major basin-parallel faults, suggesting that emplacement was controlled by extensional reactivation of structures related to the earlier evolution of the Mallina Basin.
- A zircon population with an age as young as 2941 ± 9 Ma suggests that some sedimentary rocks in the southeastern part of the Mallina Basin were deposited during or following emplacement of the c. 2955–2945 Ma granites, but before the third phase of deformation, which is dated by intrusion of early to syntectonic granites at between c. 2940 and 2930 Ma.
- Within less than 10 m.y. of the high temperature mantle-derived magmatism, large volumes of monzogranite and syenogranite of crustal origin (Champion and Smithies, 1999) swamped the central Pilbara region. These are most voluminous adjacent to the Mallina Basin, and from the available geochronology show a

progressive decrease in age away from the basin towards the southeast. This c. 2940–2930 Ma magmatism overlapped with south-southeast to north-northwest compression, representing the third phase of deformation.

Smithies and Champion (2000) speculated that this c. 2955–2930 Ma cycle of extension (rifting) accompanied by high-temperature magmatism, with subsequent sedimentation that was followed by voluminous crustally derived magmatism, was related either to local lithospheric detachment-delamination or to a small mantle plume beneath the Mallina Basin.

Hamersley Basin

Fortescue Group

The age of the dominantly volcanic Fortescue Group is c. 2770–2680 Ma (Arndt et al., 1991; Nelson et al., 1992; Wingate, 1999). All rocks in the Fortescue Group are weakly metamorphosed to prehnite–pumpellyite facies (Smith et al., 1982). Blake (1993) and Thorne and Trendall (2001) discussed the regional stratigraphy and tectonic evolution of the Fortescue Group.

In the west Pilbara the basal formation of the Fortescue Group is the Mount Roe Basalt, and this unconformably overlies the Cherratta Granitoid Complex. In the Dampier Archipelago, unassigned basalt and andesite of the group unconformably overlie the Dampier Granitoid Complex, and are intruded by the c. 2725 Ma Gidley Granophyre (Wingate, M. T. D., 1997, written comm.). Hickman (2001) suggested that these volcanic rocks probably belong to the Kylena Formation, but no distinguishing stratigraphic markers are present. South of Roebourne, the Mount Roe Basalt unconformably overlies the Whundo Group, whereas north of Roebourne it unconformably overlies the Cleaverville Formation. In contrast to the EPGGT, there is no coincidence of synclinal structures in the Fortescue Group with synclines in the WPGGT. This provides additional evidence that diapirism was not a feature of the evolution of the WPGGT.

Excursion localities

Localities visited are marked on Figures 3, 4, 8, and 12. Please note that access to many localities visited during this excursion involves crossing pastoral leases and mining tenements, and the use of private tracks. Accordingly, anyone intending to visit these localities at a future date should first obtain approval from the landowners and tenement holders involved.

Part 1: Southern part of the Mallina Basin, including basement and major shear zones

Locality 1.1: High-Mg diorite (sanukitoid) suite, Wallareenya Granodiorite (MGA 0688800E 7704500N)

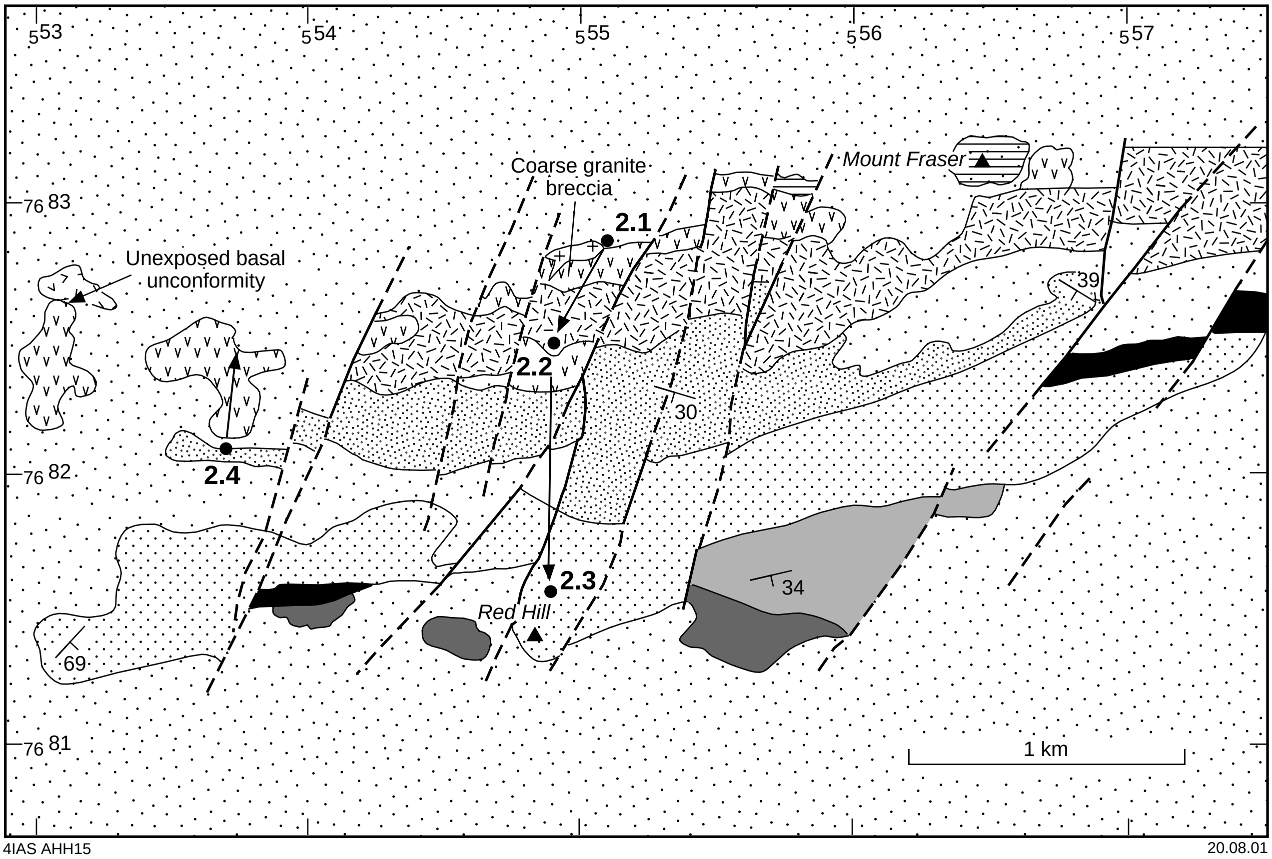
Leave Port Hedland and drive about 55 km south along the Port Hedland – Mount Newman railway service road, to Cookes Hill. Turn left and drive a further 2 km and turn right onto the road to Wallareenya Station. Follow this road eastwards for about 14 km to Wallareenya Homestead. Rocks of the high-Mg diorite suite outcrop at, and to the north of, the homestead.

The granitoid rocks exposed at, and to the north of, Wallareenya Homestead belong to the Wallareenya Granodiorite, and form part of the Pilbara high-Mg diorite suite. The rocks at this locality are mesocratic hornblende–biotite(–clinopyroxene) granodiorites, which form the most common component of the suite. In these rocks, 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}. Euhedral hornblende is intergrown with the early crystallizing plagioclase framework. Subhedral to euhedral hornblende forms an intergranular phase or occurs in allotriomorphic-textured mafic clots. In many of the granodiorites, hornblende contains cores of diopside ($Wo_{45–47}En_{40–41}Fs_{13–14}$) 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, 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.

A ubiquitous feature of rocks of the high-Mg diorite suite, irrespective of the silica content of the rocks, is the presence of abundant rounded to subangular xenoliths, some reaching up to metre scale. The majority of xenoliths are of dioritic to granodioritic composition. They are interpreted to be cognate in origin, based on mineralogical similarity between the xenoliths and rocks from compositionally zoned high-Mg diorite plutons.

Locality 1.2: Tabba Tabba Shear Zone (MGA 0678200E 7700200N)

From Wallareenya Station, return to Cookes Hill, on the Port Hedland – Newman railway service road. Follow this road to the south for about 2 km (MGA 0674600E 7701100N), then turn east. After about 200 m turn right onto the old Port Hedland – Wittenoom road. Follow this road for about 1.5 km to the south (MGA 0675100E 7699300N) and then turn left towards a prominent ridge. Follow this track for 1 km and cross a wide sandy creek. Turn right just after the crossing and before the ridge.



POST-WHIM CREEK GROUP INSTRUSIONS

		Quaternary deposits
LOUDEN VOLCANICS		Basalt lava intrusions
MONS CUPRI DACITE		High-level dacite intrusion
RED HILL VOLCANICS		Rhyodacite pumice breccia, flattened pumice breccia facies
		Tabular bedded sandstone and siltstone facies
WARAMBIE BASALT		Basalt lava, hyaloclastite, autobreccia
WHUNDO GROUP		Amphibolite-facies metavolcanic rocks

Granitoid

Quartz–feldspar porphyry

Dolerite dyke

Fault, (concealed where dashed)

Traverse direction

Excursion locality

Figure 12. Simplified geological map of the Red Hill area in the southwest of the Whim Creek greenstone belt and excursion localities 2.1–2.4

Follow the ridge around on its southern side and follow the track towards the East Turner River. Locality 1.2 is situated on the southern side and easterly tip of the chert ridge just west of Poonthuna Pool.

Near the East Turner River the Tabba Tabba Shear Zone runs east–west. A prominent chert ridge shows a moderately westerly plunging ribbing lineation and sheath folds. Immediately south of the chert, altered ultramafic rocks are in the form of actinolite–chlorite–talc schists. These schists show evidence for three phases of deformation. Intensely folded and strongly weathered metasedimentary rocks lie to the north of the chert but are very poorly exposed.

The main foliation, southwest-plunging lineations, and sinistral shear bands represent the main deformation phase with a major sinistral component. These structures locally overprint and refold an earlier foliation that is parallel to the chert ridge. To the south, this early foliation gradually rotates towards a more northeasterly direction over a distance of several hundred metres. This geometry indicates that there must have been an early major dextral phase of movement on the Tabba Tabba Shear Zone pre-dating the oblique sinistral-normal phase.

All of these structures are affected by a late phase of deformation, with a dextral component represented by crenulations and decimetre- to metre-scale folds with a northeast-striking vertical axial plane and subvertical to southwest-plunging fold axes. This late phase is best developed close to the chert ridge.

The rocks to the south of the chert ridge consistently show moderately to steeply westerly plunging stretching and mineral lineations. Throughout the area, this mineral lineation is overgrown by random hornblende, indicating a thermal overprint post-dating the oblique sinistral-normal phase of deformation with which the mineral lineations are associated.

Locality 1.3: Pilbara Well greenstone belt — basement to the Mallina Basin (MGA 0643900E 7657500N)

Return to the Port Hedland – Newman railway service road, and cross the service road at the marked railway crossing (MGA 0674600E 7701100N). Travel to the west, about 16 km, through Indee Station and onto the Great Northern Highway. Turn south and drive about 40 km (MGA 0677000E 7663600N). Turn west onto a graded gravel road and follow this road for 42 km, to Yandeearra Station. From Yandeearra, follow the Yandeearra–Mallina road to the northwest for 5 km to a bore and cattle yard. Turn right onto a northerly trending track along a fence line. Follow this track for about 7 km then turn right onto a faint track. Follow this track, passing over a northwesterly trending cross-track after about 300 m and continue on to the northeast for about 4 km. The faint track fades away after 1–2 km and the last part of the drive is across country. Locality 1.3 is on the south side of a low hill, to the south of a prominent scarp.

Greenstones in the Pilbara Well greenstone belt form part of the basement to the Mallina Basin, and are assigned to the EPGGT. The greenstone belt contains components of the Gorge Creek Group, and an unassigned mafic–ultramafic unit (‘Teichmans Group’ of Fitton et al., 1975), similar to the Warrawoona Group in the east Pilbara. A thin felsic unit, tentatively assigned to the Gorge Creek Group, contains a range of felsic volcaniclastic rocks, and minor rhyolite and trachyte. A chert–BIF unit at the top of the succession, correlated with the Cleaverville Formation, consists of layered chert, silicate-facies BIF, and subordinate, poorly exposed clastic sedimentary rocks, and foliated basalt.

Rocks in the Pilbara Well greenstone belt are multiply deformed and metamorphosed, and up to three phases of ductile deformation have been recognized (Smithies and Farrell, 2000). The dominant structures are the regional, east to northeasterly trending foliation, and the John Bull synform, a large tight, upright, northeasterly plunging fold that affects rocks of the Pilbara Well greenstone belt and the Mallina Basin.

The mafic–ultramafic sequence is dominated by high-Mg basalt, but also contains ultramafic rocks, clastic sedimentary rocks, layers of chert, and conformable zones of silicification. Most of the succession appears to consist of repeated cycles, comprising high-Mg basalt flows (with or without ultramafic rock, and thin gabbro sills) overlain by silicified mafic-fragmental and poorly sorted sedimentary rocks, and an upper conformable zone of silicification with one or more layers of chert intercalated with sedimentary rock (with or without basalt). Thin bands of dark-grey chert that cut across the underlying stratigraphy are also present in some areas. These chert horizons and zones of silicification may possibly represent quiescent periods during which there was limited volcanic activity and the upper parts of the volcanic pile underwent hydrothermal alteration and resedimentation.

The felsic unit at the base of the Gorge Creek Group contains a variety of felsic volcaniclastic rocks, and minor rhyolite and trachyte. The most abundant rock type in the succession is a medium to thick-bedded, feldspathic sandstone and siltstone. Local coarse-grained zones contain volcaniclastic conglomerate with a variety of felsic rock clasts up to 300 mm in diameter. Silicification of the sequence is widespread and locally intense, thus making the rocks difficult to identify in the field. In addition, silica-cemented hydrothermal breccias are common in some areas. The appearance of felsic volcaniclastic rocks in the stratigraphy is thought to mark a significant shift in the tectonic regime. The limited thickness and areal distribution of the felsic rocks, coupled with the predominance of volcaniclastic material, suggests that the volcanic activity was limited in time and space.

The upper unit of chert and BIF consists of layered chert and silicate-facies BIF, with minor, poorly exposed clastic sedimentary rocks, tremolite–chlorite schist, and foliated basalt. The chert shows considerable variation in texture, from relatively massive chert with diffuse patches and swirls of colour, through to layered and finely laminated chert. The Fe-oxide content is highly variable, and is typically reflected in the colour of the chert, which can vary from cream or pale grey, through to dark blue-grey or dark brown. Banded iron-formation is developed locally, particularly near Kangan Gap, where dark-grey and red layered BIF accompanies tremolite–chlorite schist (?after high-Mg basalt) and strongly foliated ferruginous siltstone.

The age of the mafic–ultramafic succession is equivocal, as there are no geochronological data, and correlation with formations elsewhere in the EPGGT is uncertain. Moreover, the chert horizons may mark significant time breaks in the succession and, therefore, the lower part of the Pilbara Well greenstone belt has not been formally assigned to any specific stratigraphic unit. A SHRIMP U–Pb zircon age of 3421 ± 2 Ma (Nelson, 1999) for the Yallingarrintha Tonalite, adjacent to the southern margin of the Pilbara Well belt, suggests that some components of the Pilbara Well succession may be as old as c. 3.4 Ga. The felsic unit in the Gorge Creek Group has not been directly dated due to the poor exposure. If it is a correlative of the felsic rocks at Nunyerry Gap (Smithies, 1998b,c), which have a maximum age of c. 3015 Ma (Smithies, 1998b), then it seems likely that the felsic rocks are unconformable on the mafic–ultramafic succession.

Traverse through the main components of the Pilbara Well greenstone belt (MGA 0643900E 7657500N to MGA 0643900E 7658200N)

At this locality, the mafic–ultramafic succession consists of metamorphosed and variously deformed basalt, fine-grained gabbro, and basaltic tuff, with thin interspersed units of metamorphosed ultramafic rock. The ultramafic rocks include serpentinized peridotite, tremolite–chlorite schist, and carbonate-altered ultramafic rock. Towards the contact with the overlying felsic unit, the metabasalt has abundant amygdales and local varioles. The amygdales are filled with quartz, plagioclase, epidote, actinolite, carbonate, and opaque minerals. Thin dykes of quartz–feldspar porphyry are common in the upper part of the mafic–ultramafic succession close to the contact with the felsic unit. It is not clear whether these dykes are genetically related to the felsic unit, or whether they are part of a younger c. 2940 Ma suite of porphyry dykes (Smithies and Farrell, 2000).

In the lower part of the felsic unit, the rocks are pale-grey to creamy white, cherty, non-bedded and relatively structureless. Upsequence, the rocks show a poorly developed layering. Farther to the north (stratigraphic top), the unit displays graded beds ranging from coarse-grained sandstone at the base, through fine-grained sandstone with cherty interbeds, to thin, laminated cherty tops. In thin section, the sandstone comprises subangular clasts of plagioclase, K-feldspar, and lithic fragments. Clasts with cuspate shapes may be former vitric fragments.

Chert of the Cleaverville Formation forms a steep ridge at the northern end of the traverse. At this point, the chert is grey with a variably developed layering that appears to be parallel to the regional foliation.

Locality 1.4: Basal part of the Mallina Basin — Constantine Sandstone (MGA 0635400E 7659500N)

From the previous locality, backtrack for 4 km to the northwesterly cross-track. Turn right (north-northwest) and follow an old track for about 10 km, past Black Gin Well, to a track intersection near Satirist Outcamp Well. Turn left at the intersection, and proceed for 1.5 km to a Y-junction with the Yandeearra–Mallina road. Turn left and follow the road south for 2 km. Locality 1.4 is about 1 km off the road to the east, at the base of a hill.

Conglomerate and pebbly sandstone belonging to the Constantine Sandstone outcrop along the northern edge of the Pilbara Well greenstone belt in the area north of Black Gin Well. The original thickness of the conglomerate is difficult to estimate due to the high strain and multiple deformation, but was probably less than about 800 m. The main rock types in this area are poorly sorted coarse-grained lithic sandstone, pebbly sandstone, and conglomerate. The rocks contain clasts of chert (>80 vol.%), unidentified silicified rock, and rare vein quartz. Some of the silicified rock fragments contain carbonaceous material. A variety of chert types are present, including both layered and non-layered varieties. The clasts show a wide range in angularity and sphericity, although most are subangular to subrounded. Most clasts are less than 150 mm in diameter, but there are rare boulders of chert up to 1.5 m in diameter. The matrix is strongly silicified and recrystallized. At the eastern end of the unit, the rocks contain very angular chert fragments and are difficult to distinguish from fractured chert of the Cleaverville Formation.

The contact between the Constantine Sandstone and the greenstone basement is not directly exposed on SATIRIST, but it is broadly concordant with bedding on both sides of the contact. The contact is inferred to be an unconformity, based on the observation that in some areas the Gorge Creek units are missing, and the Constantine Sandstone is juxtaposed against the mafic–ultramafic succession. It is possible that felsic and chert–BIF units could have locally been removed by faulting, as the rocks adjacent to the contact are strongly deformed. However, the locally northwesterly trend of the contact, north of Black Gin Well, is at a high angle to the foliation trends and supports the interpretation that the contact is a locally tectonized unconformity. Moreover, the observation that the contact is refolded into Type 2 and 3 interference patterns suggests that any contact-parallel fault must post-date the last major deformation event in the area. In addition, the conglomerate and poorly sorted breccia in the Constantine Sandstone are similar to rocks in the same formation at Teichmans and at Nunyerry Gap, 20 and 60 km, repectively, to the southwest, where an unconformity is exposed (Horwitz, 1979; Hickman et al., 2000).

Traverse through the Constantine Sandstone to basement (MGA 0635400E 7659500N to MGA 0635600E 7659400N)

The traverse starts in strongly deformed coarse-grained sandstone, pebbly sandstone, and pebble conglomerate. The rocks dominantly contain clasts of chert and extremely silicified rock. In thin section, the finer grained rocks contain clasts of chert, quartz, and various silicified rocks, with diffuse relict textures preserved by variations in abundance of ultrafine opaque material. Scattered boulders and cobbles of layered chert are present locally. The foliation at this location is parallel to the contact with the underlying mafic–ultramafic rocks of the Pilbara Well greenstone belt. To the south, there is no obvious change in grain size approaching the basal contact.

The basement succession is represented by patchy exposures of weakly deformed high-Mg basalt with a relict pyroxene-spinifex texture and small varioles up to 1 mm in diameter. To the south is a relatively uniform, weakly deformed succession of high-Mg basalt.

Locality 1.5: Mallina Formation (MGA 0584000E 7673800N)

From the Pilbara Well area the route to the next stops in the Mallina–Croydon area will depend on road conditions. The area can be accessed directly along the disused Mallina–Yandeearra road. The alternative route is via Yandeearra and onto the Great Northern, and the North West Coastal highways, until the turnoff to Mallina Station, about 20 km west of Whim Creek. From Mallina Station, drive about 29 km to the southwest along the Mallina–Croydon road. From there, leave the road and drive a further 1 km to the west to a low hill comprising outcrop of the Mallina Formation.

This locality shows deformed turbidites of the Mallina Formation on the western limb of the Croydon Anticline. The available geochronology constrains the age of these rocks to between c. 2970 and 2950 Ma. Rocks of the Mallina Formation typically outcrop very 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. Evidence for felsic volcanic or volcaniclastic units is extremely rare. Occasional lithic fragments and pebbles in the turbidites are either shale intraclasts or chert. Two main phases of deformation can be seen, the earlier phase related to formation of the Croydon Anticline. Graded units indicate that the succession faces to the west.

Locality 1.6: Croydon Anticline; Constantine Sandstone (MGA 0583700E 7664400N)

Rejoin the Mallina–Croydon road and drive a further 10 km southwards to Croydon Homestead. Turn east and drive about 500 m.

This locality shows medium- to coarse-grained quartzites of the Constantine Sandstone, which underlies the Mallina Formation. These rocks locally include chert-cobble conglomerate, and have been interpreted to represent typically proximal, or upper fan, depositional environments (Hickman, 1977; Eriksson, 1982) characterized by high rates of sediment deposition.

The age range of detrital zircon populations from the Constantine Sandstone varies widely. A sample taken from this locality contained detrital zircon populations with ages greater than 3.5 Ga, c. 3.4 Ga, c. 3.28 Ga and c. 3.2 Ga (Fig. 7). The ages less than 3.0 Ga that characterize samples of the Constantine Sandstone and Mallina Formation collected to the north, are not found here, and this is interpreted to reflect a systematic variation in the influence with source areas of strongly contrasting age ranges.

Locality 1.7: Boninite-like rocks in the Mallina Basin (MGA 0591700E 7666800N)

Road conditions permitting, drive eastwards from Locality 1.6 for about 11 km.

At this locality, strongly deformed high-Mg basaltic rocks are exposed, and are locally interleaved with clastic rocks of the Constantine Sandstone. The rocks show a well-developed spinifex texture; however, it is not clear if development of this texture is a primary feature or a result of recrystallization in proximity to the Peawah Granodiorite and the Satirist Granite. To the southeast of the Satirist Granite, less deformed exposures of the high-Mg basaltic unit forms irregular, lensoidal outcrops within the Constantine Sandstone possibly indicating synsedimentary eruption of lavas. However, no definitive evidence for an extrusive origin is found, and it is possible that these rocks form a subvolcanic sill complex.

These high-Mg basalts have a geochemical signature that closely matches that of Phanerozoic boninite. However, they form a component of the Mallina Basin, which is thought to have formed in an intracontinental setting rather than at a convergent margin (Smithies et al., 1999, 2001; Van Kranendonk et al., in prep.; see **Overview of the Mallina Basin**). Conflicting factors complicate petrogenetic interpretations of these boninite-like high-Mg basalts. The fractionated HREE signatures of the rocks (Fig. 10) indicate that a refractory mantle source existed beneath the Mallina Basin, and the intrusion of the c. 2950 Ma high-Mg diorite (sanukitoid) suite into the Mallina Basin shows that the sub-Mallina mantle locally also had the appropriate subduction-modified, LILE-enriched compositions for production of boninites. Alternatively, the trace-element composition of these boninite-like rocks can be accurately modelled by contaminating a melt derived from a refractory mantle source with 10–15% of material similar in composition to c. 3170 Ma granitoids that lie to the south of the Mallina Basin.

Locality 1.8: Mallina Shear Zone (MGA 0602600E 7687800N)

Rejoin the Mallina–Croydon road, heading northeast to Mallina Station. About 100 m north of Mallina Station (MGA 0607000E 7690600N) turn west onto a track towards a creek. After crossing the creek the track turns to the southwest. After about 4 km this track crosses the Mallina Shear Zone.

This locality shows the typical nature of the Mallina Shear Zone. The lithology is variable and comprises alternating shale and wacke. The extensively calcretized metasedimentary rocks are strongly folded and locally sheared. The centimetre- to decimetre-scale shears mostly display a dextral geometry. The decimetre- to metre-scale folds also show a dextral asymmetry with fold axes plunging steeply to the south-southwest. Locally, a weak subvertical cleavage is present that is axial planar to these folds. These structures indicate that dextral strike-slip movement represented the last stage of deformation in the Mallina Shear Zone. Locally a sinistral sense of shear can be observed in small shear zones. The early south-side-up event observed elsewhere in the Mallina Shear Zone and associated structures cannot be observed at this locality.

Part 2: Northern part of the Mallina Basin — Whim Creek greenstone belt

Locality 2.1: Whim Creek Group — Warambie Basalt and Mons Cupri dacite (MGA 0555100E 7682850N)

Starting from the Whim Creek Hotel, drive west along the North West Coastal Highway for about 34 km, crossing the Little Sherlock River. Turn left, taking the track through a gate, and continuing south for 1.1 km, to where the track swings west. Continue west along the track for about 3 km, turning south at a fence and continuing for 1 km. Localities 2.1–2.4 are all within walking distance of this point (Fig. 12).

The coarse basal breccia, which is restricted to this locality, represents the lowest part of the Warambie Basalt and is composed of angular boulders of basalt and granite with a poorly sorted coarse sandstone–granulestone matrix. The breccia is interpreted to be derived from basement uplift and erosion and may be proximal to a syndepositional fault. Transport was likely to have been by clast rolling and sliding. Note that the Caines Well Granitoid Complex near this locality is a leucogranitic phase that intrudes the breccia, rather than forming basement to the Whim Creek Group.

Traverse across Warambie Basalt and Mons Cupri dacite (from MGA 0555100E 7682850N to MGA 0554950E 7682450N)

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 (Figs 13 and 14). Vesicle layers are found within the basaltic lavas, which also include aphanitic as well as many plagioclase-phyric types.

The first dacitic rocks encountered along the transect (at MGA 0555500E 7682700N) are dark grey. The contact relationships here are unclear. 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 2.2: Whim Creek Group — basalt–dacite contact relationships (MGA 0554950E 7682450N)

Directions are the same as for Locality 2.1 (see Fig. 12).

At this locality, 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. The preferred interpretation, however, 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, whispy 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.

4IAS RHS7 12.7.01

Figure 13. Hyaloclastite from the top of the Warambie Basalt

Traverse across the Red Hill volcanics (from MGA 0554950E 7682450N to MGA 0554900E 7681600N)

The flat ground at the start of this traverse is underlain by centimetre- to decimetre-scale tabular beds of shard-rich volcaniclastic rocks (Fig. 15). 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 the traverse from MGA 0554900E 7681600N onwards. The first outcrops contain both dense and pumice breccia in massive beds. The volcaniclastic material is rhyodacitic in composition. The angular volcaniclastic breccia represents the onset of felsic volcanism and probably resulted from debris-flows derived from adjacent volcanic areas (Fig. 16).

Locality 2.3: Whim Creek Group — pumice breccia, Red Hill volcanics (MGA 0554900E 7681600N)

Directions are the same as for Locality 2.1 (see Fig. 12).

The hill to the south of this locality is Red Hill and is composed entirely of tube-pumice breccia. Around the hill (to the west), the pumice clasts become progressively flattened. In extreme cases of flattening, clasts are several centimetres in length but only a few millimetres thick. The matrix to these pumice breccias is glass-shard sandstone. The pumice breccias lack bedding and have a monomictic composition. They represent synvolcanic sedimentation. Elsewhere, upward-fining packages (about 20 m thick) indicate deposition from turbidity currents. The large preserved volume of pumiceous material (about 2.5 km^3) points to an ignimbrite origin, as does the presence of relic shard textures. These breccias gave a U–Pb SHRIMP zircon age of 3009 ± 4 Ma (Nelson, 1998).

Locality 2.4: (Optional) Whim Creek Group — complex basaltic lava and breccia, Warambie Basalt (traverse from MGA 0553700E 7682150N to MGA 0553700E 7682450N)

This stop is about 1.4 km to the northwest of Locality 2.3 (see Fig. 12) and can be accessed by foot.

This locality shows the complexity of basaltic lava and related products. The basaltic lavas are aphanitic or plagioclase-phyric and contain many different plagioclase phenocryst types, including simple laths, interpenetrant ‘crossed’ laths, etched blocks, and etched megacrysts to 5 mm in length. The basaltic lavas have been intruded by metre- to several hundred metre-scale ovoid bodies of dacitic rock. Some basaltic rocks contain flow-banded margins, indicating that they are coherent entities. The ovoid intrusions of dacite may be coeval with basaltic volcanism or intruded prior to lithification (Fig. 12).

Locality 2.5: Whim Creek Group — Mons Cupri dacite and clastic rocks of the Red Hill volcanics (MGA 0573500E 7681700N)

Return to the North West Coastal Highway, then drive east for about 13.5 km. Turn right onto a track, proceed through the gate, and drive south a further 4 km to Locality 2.5.

4IAS RHS2 12.7.01

Figure 14. Autobreccia in the Warambie Basalt

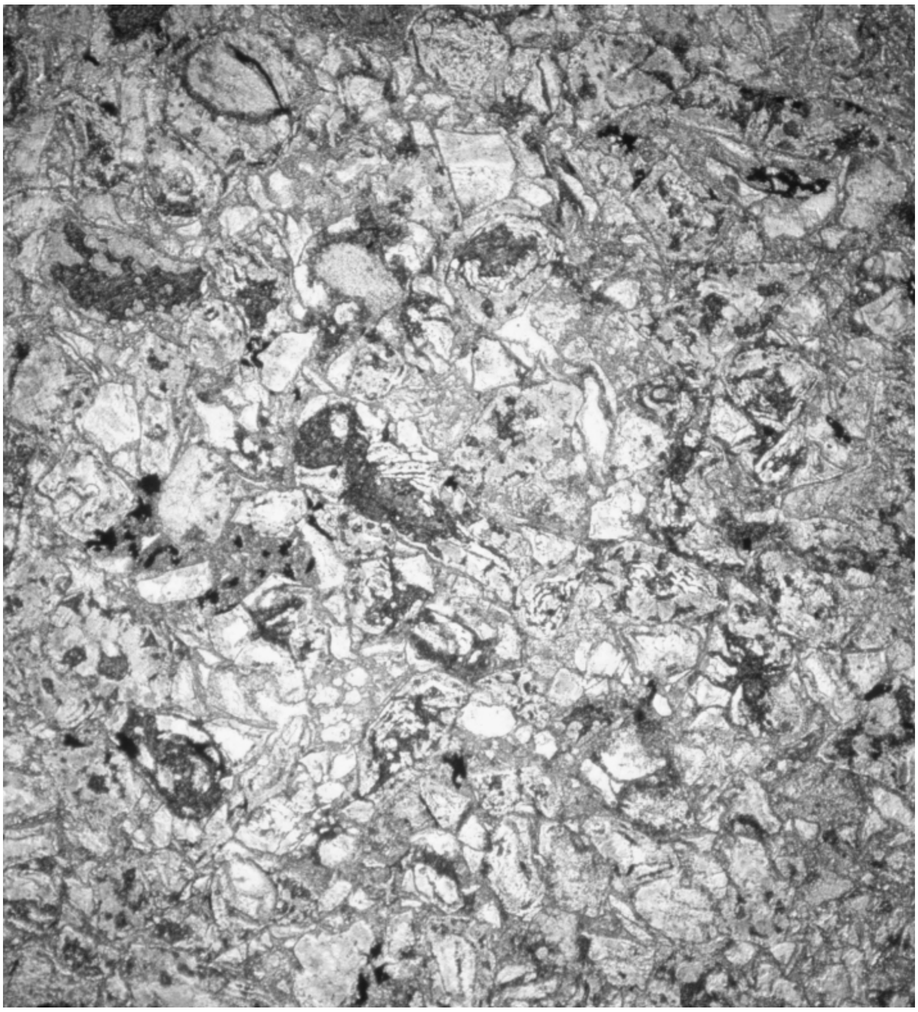

4IAS RHS3 12.7.01

Figure 15. Photomicrograph of a shard-rich sandstone from the Red Hill volcanics

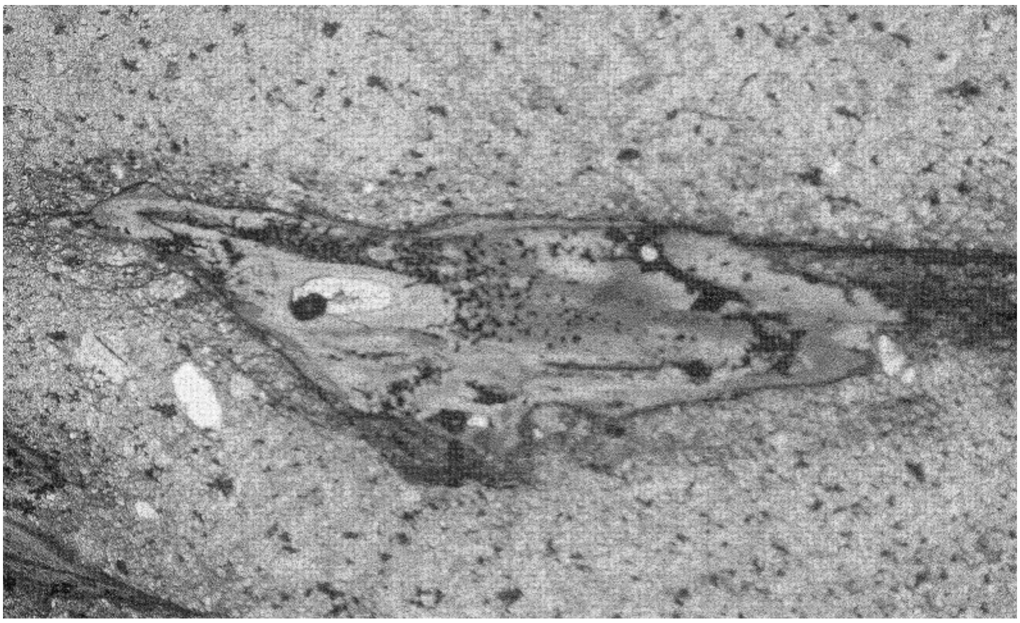

4IAS RHS4 12.7.01

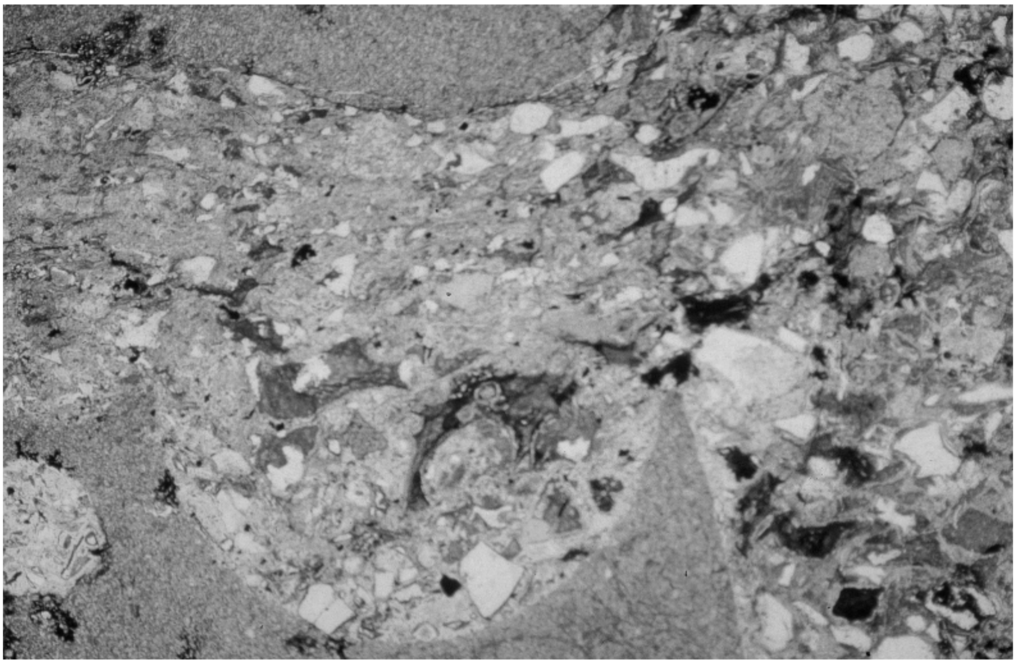
Figure 16. Pumiceous breccia from the Red Hill volcanics

Traverse across a complete dacite intrusion and overlying breccia–sandstone (from MGA 0573500E 7681700N to MGA 0573200E 7681550N)

This traverse covers a number of lithologies. The basal rocks are poorly exposed, charcoal-grey basalt and dolerite. These are overlain by around 12 m of crudely upward-fining polymictic breccia and sandstone. The clastic rocks are massive to diffusely planar bedded and contain dacite clasts (similar to the Mons Cupri dacite) and light- to medium-grey (?andesite) volcanic clasts. The clastic rocks pass into massive dacite across an irregular boundary. Clasts of dacite (probably quenched) are in the clastic rocks. Apophyses of sandstone are also found within the dacite.

Near the top of the dacite unit there is abundant enclave material within flow bands or as discrete, boulder-size inclusions around which flow bands have deformed. The dacite is locally strongly flow banded and brecciated. Clasts of dacite in sedimentary rock and apophyses of sedimentary material within the dacite can be found at both the upper and lower contacts of the dacite layer and reflect soft-sediment intrusive contacts (Fig. 17). Consequently, the dacite probably intruded the sedimentary rocks prior to lithification.

The sedimentary succession overlying the dacite is dominated by tabular beds of breccia, coarse-grained sandstone and rare fine-grained sandstone. The breccia is typically massive and matrix supported and is associated with lesser conglomerate. The coarse sandstone is also tabular and may be capped by thin (centimetre-scale) fine-grained sandstone interbeds. Asymmetric ripples, flame structures, and micro-sandstone dykes are also noted. Coarse clasts are dominated by aphanitic and vesicular andesite with lesser dacite and rare granitoid (Fig. 18). These sedimentary rocks are tabular mass-flow deposits derived from a dominantly andesitic source.



4IAS RHS5 12.7.01

Figure 17. Photomicrograph of peperite from the Red Hill volcanics

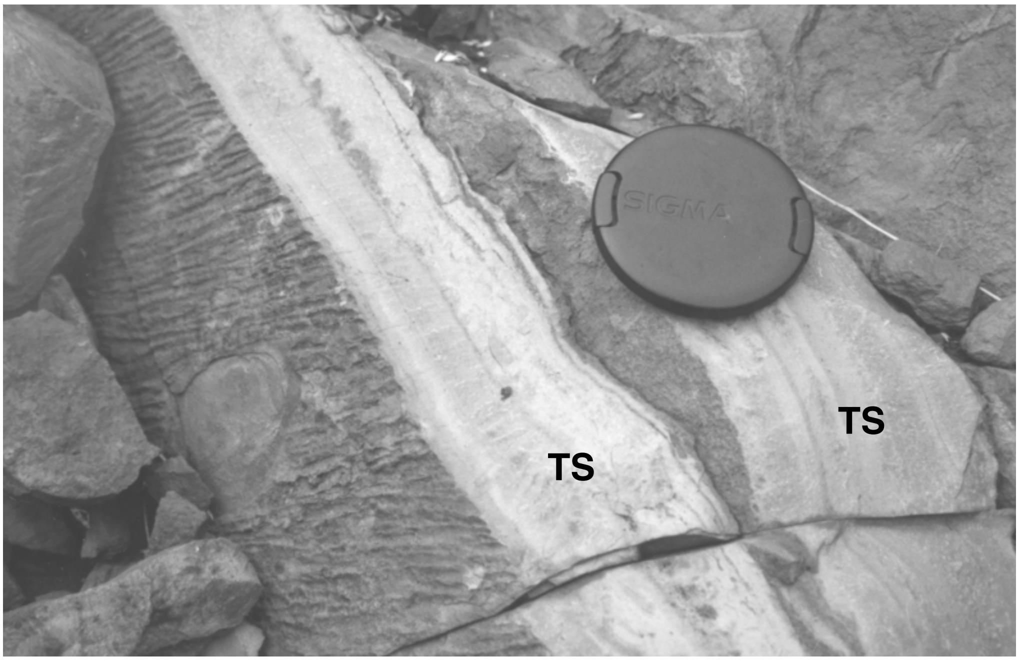


Figure 18. Asymmetric ripples on the surface of tabular volcaniclastic sandstone (TS) form the Red Hill volcanics

Locality 2.6: Bookingarra Group — Cistern Formation (MGA 05724000E 7681150N)

Locality 2.6 is about 1.3 km to the southwest of Locality 2.5. Much of that distance can be travelled by vehicle, the remaining distance being covered on foot.

Traverse across upward-fining breccia, sandstone, and mudstone (from MGA 0572400E 7681150N to MGA 0572100E 7680800N; Fig. 8)

This section looks at an upward-fining succession dominated by mixed volcaniclastic and siliciclastic breccia at the base, passing upward into thick sandstone and a poorly exposed mudstone cap. The polymictic breccia contains clasts of granitoid, dacite, andesite, and vein quartz. The matrix contains a higher proportion of quartz than in similar rocks from the Whim Creek Group and clasts are typically more rounded. Black chert fragments are found only in the Bookingarra Group and are locally imbricate. These, and imbricate rip-up clasts of mudstone, give approximate flow directions to the south or south-southwest. Overlying sandstone and granule- to pebble-breccia forms 5–10 m thick, (diffusely) planar-bedded packages. There is some petrographic evidence that the bases of individual packages contain more volcaniclastic clasts and less quartz than the tops. Rare basalt scoria is also observed towards the top of the succession.

The entire succession is upward fining, interpreted as the result of basin subsidence and deepening. Individual sediment flows were highly erosive upstream (abundant mudstone rip-up clasts) and were probably high-concentration turbidity currents. The volcaniclastic matrix was probably eroded from Whim Creek Group basement whereas plutonic quartz, vein quartz, granitic lithics, and black chert were derived from an unknown continental source. Scoria clasts reflect the onset of Negri Volcanics basaltic volcanism (the thick dolerite sill to the south is interpreted as a subvolcanic sill). In the Whim Creek area, mudstone forms a succession around 300 m thick, whereas mudstone at this locality is typically about 10 m thick.

Part 3: West Pilbara Granite–Greenstone Terrane

Locality 3.1: Bradley Basalt, Whundo Group (MGA 0508900E 7682000N)

Return to the North West Coastal Highway, and drive west 45 km to the bridge 1 km south of Roebourne. Beyond the bridge, do not turn right towards Roebourne, but continue 200 m, and take the left turn to Harding Dam. A second left turn is required shortly after the first. About 25 km along this road, and 3.5 km before reaching Harding Dam, turn sharply right after crossing the southern branch of Miller Creek (MGA 0508900E 7681800N). Follow the water-pipeline road for 100 m and turn right along an old track into a gravel pit. Locality 3.1 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. 19) in the Bradley Basalt of the Whundo Group. Pillow structures show that the lava flows, which dip 55° towards the northeast, are right-way-up. The basalt at this locality is at a stratigraphic level 1000 m above the base of the formation and, from geochronology along strike and higher in the succession, is 3120–3115 Ma (Table 2). The depositional environment of the basalt may have been a back-arc basin (Krapez and Eisenlohr, 1998), and Nd T_{DM} model ages (3250–3150 Ma) from the Whundo Group (Sun and Hickman, 1998) suggested that it was not generated by melting of source rocks significantly older than its depositional age. This contrasts with volcanic rocks of the c. 3270–3250 Ma Roebourne Group that have Nd T_{DM} model ages of 3480–3430 Ma (Sun and Hickman, 1998).

The top of the hill at Locality 3.1 provides views of formations in the Fortescue Group to the north, east, and south. The basal Mount Roe Basalt (c. 2770 Ma) unconformably overlies the Whundo Group, and dips southeastwards at about 20°. To the east the Mount Roe Basalt is overlain by the Hardey Formation, which locally comprises sandstone, tuff, ultramafic lava, and dolerite.

4IAS AHH FIG19 10.08.01

Figure 19. Pillow lava in the Bradley Basalt 2.5 km northwest of Harding Dam, Locality 3.1

Locality 3.2: Felsic pyroclastic rocks, Bradley Basalt (MGA 0505500E 7684300N)

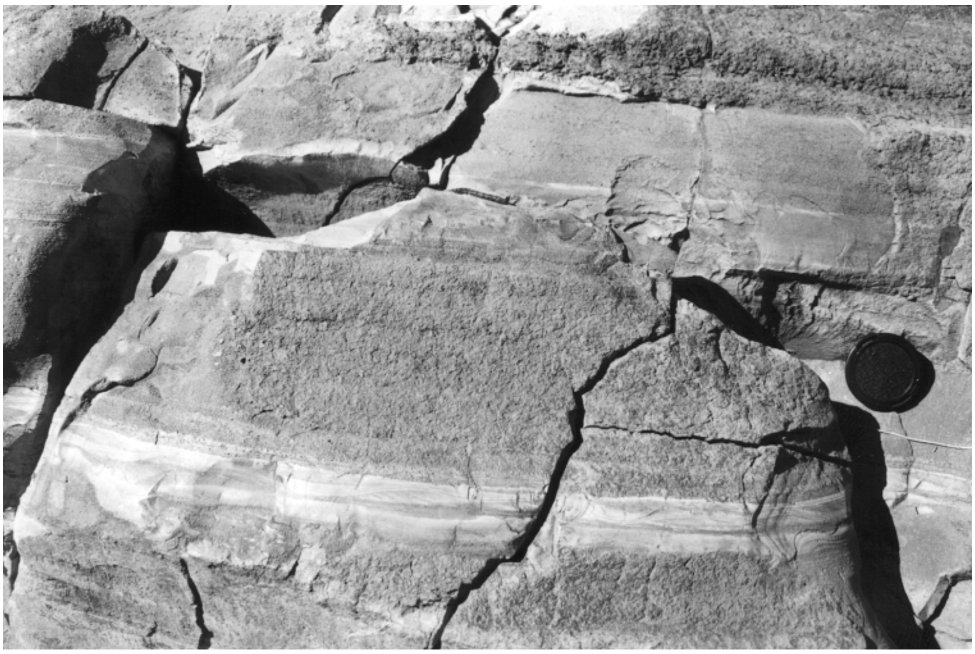
Return to the water-pipeline road, and follow this west for about 6 km to a sharp bend where a powerline crosses the road (MGA 0505000E 7683650N). At the bend, turn right on a four-wheel drive track that follows the powerline to the northeast. Follow this track for 1 km to Locality 3.2. The main exposures are located 50 m west of the track and 50 m south of the nearest pylon.

The central part of the Bradley Basalt contains several units of felsic tuff (including reworked tuff) and agglomerate, with lenticular sills of dolerite. Exposures in a small valley west of the track show beds of felsic tuff and agglomerate dipping 65° towards 010°. The felsic unit is about 100 m thick, underlain by silicified basalt and intruded to the north by dolerite. Graded bedding (upward fining) and fine-scale cross-bedding show the succession to be right-way-up. Fine-grained reworked tuff displays intraformational convolutions and load structures (Figs 20 and 21). Reworking of the tuff by currents suggests a relatively shallow-water depositional environment.

About 2000 m higher in the Whundo Group, the Bradley Basalt is overlain by the felsic pyroclastic Woodbrook Formation, and about 3000 m lower in the group similar felsic pyroclastic rocks form parts of the Tozer Formation. Local sandstone at the top of the Tozer Formation may also indicate shallow-water deposition. Given the approximate 10 km total thickness of the Whundo Group, its depositional basin must have been progressively subsiding.

Locality 3.3: Northern margin of Sholl Shear Zone (MGA 0499600E 7689100N)

Return to the water-pipeline road, and follow this northwestwards for about 8 km to the crossroads intersection with the Roebourne–Cherratta road. Locality 3.3 is the ridge 250 m north of the crossroads.



4IAS AHH FIG20 10.08.01

Figure 20. Graded bedding in reworked rhyolite tuff in the Bradley Basalt, Locality 3.2

Figure 21. Reworked rhyolite tuff in the Bradley Basalt, showing deformed load structures and fine-scale cross-bedding, Locality 3.2

Whether the WPGGT is a single terrane or a composite of two terranes depends on the interpretation of the tectonic significance of the Sholl Shear Zone. Krapez and Eisenlohr (1998) and Smith et al. (1998) argued that the Sholl Shear Zone is a terrane boundary because it juxtaposes the 3270–3250 Ma Roebourne Group (interpreted by them as an island-arc succession) with the 3130–3115 Ma Whundo Group (interpreted as a back-arc succession). Van Kranendonk et al. (in prep.) concluded that the Sholl Shear Zone approximately coincides with a post-3300 Ma terrane boundary.

Stratigraphic differences between rocks north and south of the Sholl Shear Zone are accompanied by Sm–Nd evidence that the Roebourne and Whundo Groups were formed in different tectonic environments (Sun and Hickman, 1998). The Sholl Shear Zone was probably active before and during deposition of the Whundo Group, and certainly prior to deposition of the Cleaverville Formation at c. 3020 Ma. However, because the Cleaverville Formation occurs on both sides of the shear zone, and its stratigraphy is similar in these northern and southern outcrops, most movement on the zone must have occurred before about 3020 Ma. Van Kranendonk et al. (in prep.) pointed out that at least 80% of the rocks (including granitoids) now separated by the Sholl Shear Zone were formed in approximately their current relative positions, and consequently post-date the pre-3020 Ma boundary. Thus, the WPGGT is an essentially coherent geological unit that developed across this boundary, and largely conceals the components of two older terranes.

Over most of its length (at least 250–350 km, depending on geophysical interpretation) the Sholl Shear Zone is a near-vertical zone of mylonite and schist 1000–2000 m wide. Dextral displacement (30–40 km) of the c. 3010 Ma Whim Creek Group is seen north of Whim Creek (Smithies, 1998a), and (Hickman, in prep.) mapped dextral displacement of c. 2925 Ma layered intrusions southwest of Roebourne. Localities 3.3 and 3.4 provide opportunities to examine two of the relatively rare outcrops of this major structure.

At Locality 3.3 a foot traverse from the crossroads to the top of the ridge provides a section through poorly exposed silicic mylonite, strongly sheared amphibolite, and chert. The mylonite section, which is far better exposed at Locality 3.4, represents highly tectonized granitoids and greenstone xenoliths. The amphibolite, on the upper slopes of the ridge, is a highly deformed part of the Ruth Well Formation (Roebourne Group). The chert, which forms the top of the ridge, is also part of the Ruth Well Formation, and outcrops sporadically along the northern side of the Sholl Shear Zone for about 30 km. The chert includes grey-and-white banded chert and ferruginous chert, and although sheared, is interpreted to be a deformed stratigraphic unit. An attempt to date zircons in the chert gave ambiguous data due to disturbance effects.

To the north of the ridge, irregular sheets of monzogranite have intruded the amphibolite of the Ruth Well Formation. The age of the monzogranite is likely to be either 3015–2970 Ma (if part of the Harding Granitoid Complex to the east) or c. 3260 Ma (if equivalent to an intrusive granitoid 3 km to the northwest, dated by Smith et al., 1998).

Locality 3.4: Sholl Shear Zone, Nickol River (MGA 0494800E 7689600N)

Drive west for about 4 km to a four-wheel drive track about 200 m before the crossing over Nickol River. Take the track to the right, drive north about 1 km, then turn left on an indistinct track to rock pavements in the bed of Nickol River.

Rock pavements in the Nickol River provide excellent exposures (Fig. 22) of the northern section of this major shear zone. The mylonite is dominantly silicic, and represents extremely sheared granitoids (Fig. 23), but there are also layers of amphibolite. Mylonite lamination is folded by tight, west-plunging Z folds (Fig. 24), which may be related to late dextral movement, and isoclinal folds (Fig. 25). All these structures are displaced by late brittle fractures (Fig. 26), which are locally filled by pseudotachylite. These are probably related to a post-Fortescue Group north–south compressional event that produced a conjugate fault system in the Dampier–Roebourne area (Hickman, 2001).

Smith et al. (1998) reported a zircon age of 3024 ± 4 Ma for the mylonite, which is interpreted to be the age of the local granitoid precursor. Shear-sense indictors elsewhere (Hickman, 2001) indicate that the dominant early shear was sinistral, and may have been partly contemporaneous with a suite of c. 3020–3015 Ma felsic intrusions along the zone, and outcropping sporadically for several kilometres to the north. Late movement (post-3010 Ma Whim Creek Group) was clearly dextral because the Whim Creek Group and the 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 each unclear. This is largely due to the fact that in the Dampier–Roebourne area, post-3000 Ma granitoids intruded large areas of greenstones north and south of the shear zone. These granitoids appear to have totally obscured pre-3015 Ma displacements of the stratigraphy.

Using Sm–Nd data (Sun and Hickman, 1998) from the Whundo Group and the Cherratta Granitoid Complex south of the shear zone, there is no indication of underlying c. 3480 Ma material. This is in contrast to isotopic data from the Roebourne Group and Karratha Granodiorite north of the shear zone. This implies that the early movement along the Sholl Shear Zone was much greater than the 30–40 km late dextral movement.

4IAS AHH FIG22 10.08.01

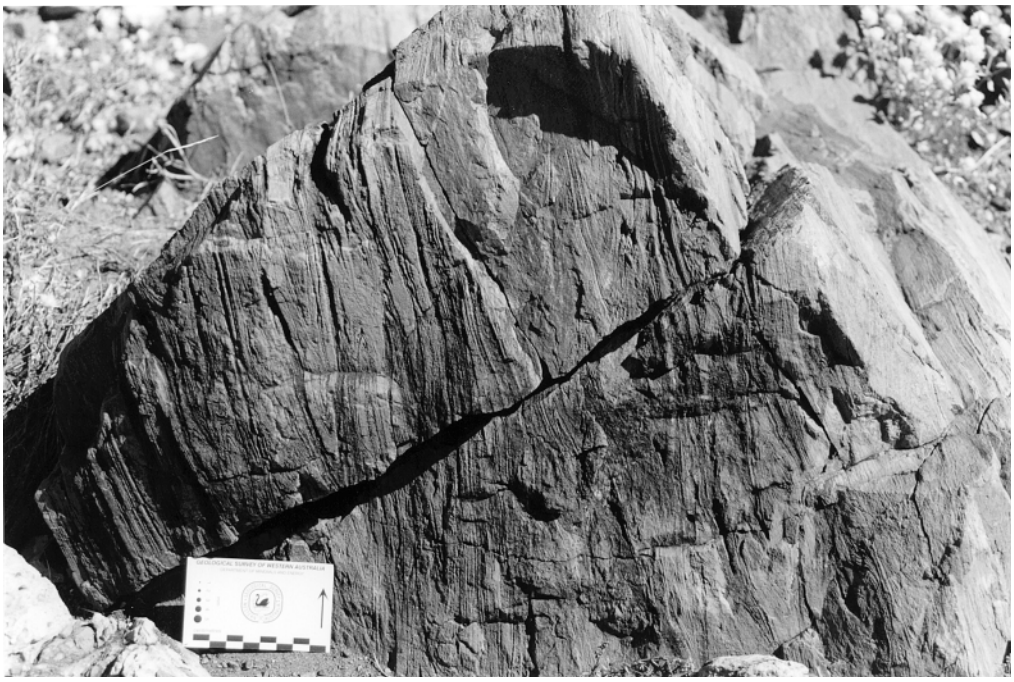
Figure 22. Mylonite of the Sholl Shear Zone in the Nickol River at Locality 3.4

4IAS AHH FIG23 10.08.01

Figure 23. Closely laminated siliceous mylonite at Locality 3.4

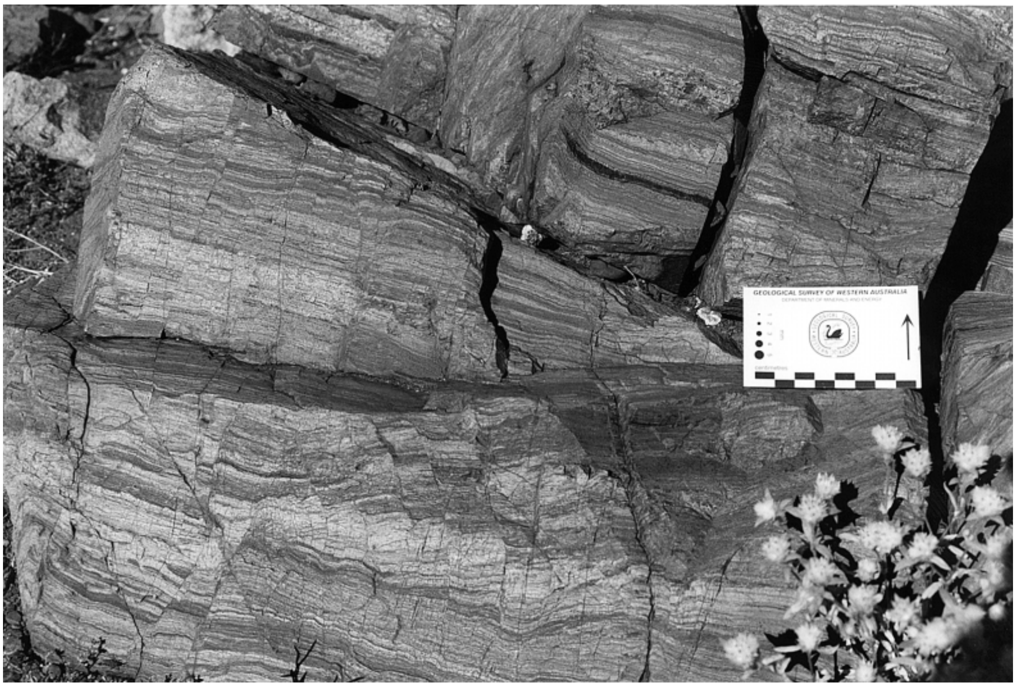
4IAS AHH FIG24 10.08.01

Figure 24. Mylonite foliation at Locality 3.4



4IAS AHH FIG25 10.08.01

Figure 25. Isoclinal folding of mylonite foliation at Locality 3.4



4IAS AHH FIG26 10.08.01

Figure 26. Close-up view of compositional banding in the mylonite at Locality 3.4

Locality 3.5: Tozer Formation felsic tuff (MGA 0492500E 7679200N)

Return to the Cherratta road and drive westwards 15 km to a railway crossing. Cross the railway, turn left, and drive 14 km to another railway crossing (near MGA 0491100E 7674700N). Cross the railway and follow the four-wheel drive track to the northeast for about 7 km to a creek crossing (MGA 0493300E 7679200N). Turn left off the track, and drive (no track) about 800 m westwards to Locality 3.5. This is situated about 200 m north of the west-southwesterly trending creek, and at the northern end of a northwest-trending ridge.

Localities 3.5 and 3.6 are situated in remote country, but provide some of the best outcrops of felsic volcanic rocks in the c. 3120 Ma Tozer Formation. The ridge at Locality 3.5 is composed of a northwesterly striking, xenolith-rich granophyric dyke that is interpreted to be a feeder dyke for the c. 2725 Ma Gidley Granophyre. The Gidley Granophyre is a major intrusion that outcrops in the Dampier area, 40 km to the northwest (see Locality 3.10). Traced northwestwards from Locality 3.5, the granophyre dyke is one of several that occur in a very narrow zone cutting both the Sholl Shear Zone and the c. 2925 Ma Sholl Intrusion. These xenolith-rich dykes contain angular to subrounded fragments of vein quartz and country rock up to 1 m in length.

The xenolith-rich dyke intrudes felsic tuff and agglomerate of the Tozer Formation, which locally dips steeply to the south-southeast. The formation is locally composed of rhyolite and dacite tuff, but elsewhere includes basalt and andesite. The well-preserved pyroclastic textures and wedge-shaped nature of this felsic section of the Tozer Formation indicate a locally developed explosive felsic volcanic centre. The Whundo Group is here interpreted to have originated in a rift system (Hickman, 2001) rather than a back-arc basin, but this rift could not have been floored by crust older than the Roebourne Group.

Locality 3.6: Traverse through upper Tozer Formation (MGA 0492500E 7678500N)

Drive south (no track) for 1 km to boulders of felsic tuff and agglomerate at Locality 3.6.

This locality provides the opportunity for a foot-traverse through the upper 100 m of the Tozer Formation. The traverse begins in good exposures of cleaved felsic pyroclastic rocks. Poorly sorted, angular clasts of chert and felsic lava up to 5 cm in diameter are set in a matrix of felsic tuff. The cleavage, which dips very steeply towards the southwest, is S_7 (see **Episodes of deformation**). Uphill, towards the southeast, the pyroclastic succession is terminated by a unit of quartz-rich felsic tuff or volcanogenic sedimentary rock. In thin section, this rock consists of approximately equal proportions of rounded quartz grains and subhedral plagioclase crystals embedded in a granular mosaic of quartz, plagioclase and chlorite. Overlying this unit is basalt of the Bradley Basalt. About 1 km to the west of this point, a flow of porphyritic rhyolite from the upper part of the Tozer Formation was dated at 3122 ± 7 Ma (Nelson, 1997).

Locality 3.7: Karratha Granodiorite and Ruth Well Formation (MGA 0477100E 7696100N)

Return to the railway crossing (MGA 0493300E 7679200N), cross the railway, and turn right on the main access road for the railway. Drive northwards for about 30 km (MGA 0477200E 7696100N). Turn left, immediately north of an east-west fence, and follow a faint track for about 100 m to a 5 m long outcrop of tonalite north of the track.

This locality provides one of the few fresh outcrops of the Karratha Granodiorite in the Mount Regal area. A sample blasted from the exposure in 1996 (GSWA sample no. 142433) was dated at 3270 ± 2 Ma using SHRIMP U–Pb zircon geochronology (Nelson, 1998). This result is consistent with U–Pb data (Smith et al., 1998, sample JS17) from tonalite 8 km to the east, and indicates that the age of crystallization 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 slightly older granitoids or greenstones in the area.

Because the tonalite at this locality extends as sheets and veins into the basal amphibolite of the adjacent Ruth Well Formation, the latter must clearly be older than 3270 Ma. This conclusion is supported by a SHRIMP U–Pb zircon age of 3269 ± 2 Ma from felsic schist (GSWA 136819) of the Nickol River Formation (Nelson, 1998). No stratigraphic components of the Ruth Well Formation have yet been found to be suitable for zircon dating.

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°, parallel to bedding in the Ruth Well Formation.

About 500 m west of the tonalite exposure the amphibolite of the Ruth Well Formation is overlain by a stratiform serpentinite unit, about 50 m thick, which in turn is overlain by a thin grey-white banded chert dipping 20° towards 310°. A fine-grained amphibolite (metabasalt) unit about 300 m thick overlies this to the west. These components of the Ruth Well Formation succession are interpreted to represent a metamorphosed sequence of ultramafic to mafic lavas containing silicified interflow sediments. The same lithological association occurs 3 km south of Karratha, and in the Ruth Well area, 10 km east-southeast of Mount Regal.

Locality 3.8: Regal Thrust and Regal Formation, (MGA 0492000E 7696800N)

Return to the main road, drive northwards for about 5 km to the North West Coastal Highway, and turn right. Drive eastwards about 10 km to the turnoff to Karratha. Opposite this turnoff, take the road to the right that follows the water pipeline southeastwards. Follow this road for 7 km (MGA 0491200E 7696800N). Leave the track about 100 m southeast of the creek, and drive east about 600 m to Locality 3.8.

Locality 3.8 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 reveals that it is a finely laminated silicic mylonite that has been isoclinally folded. Figure 27 shows a particularly good outcrop of the mylonite, and Figures 28 and 29 provide close-ups of the isoclinal folding. Parts of the outcrop (Fig. 29) show that the isoclinal folds have been refolded by later tight to isoclinal folds. Plunge of the early isoclines is generally low (up to 30°) east or west, and the prevailing dip of the mylonite is 60–80° south.

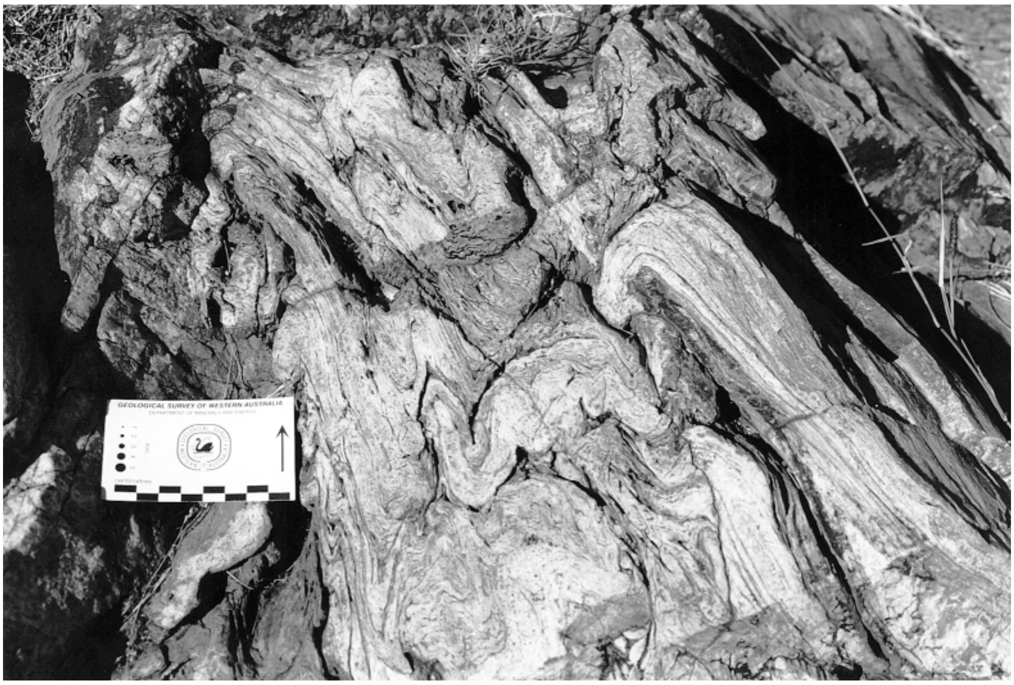
Above the mylonite, to the south is the 2 km-thick metabasalt succession of the Regal Formation. Underlying the mylonite, to the north, is a mixed lithological assemblage of ultramafic schist, felsic schist, volcanogenic metasedimentary rocks, quartzite, and chert in the Nickol River Formation. Outcrops of the mylonite and schist between these formations have been mapped over a distance of 25 km on the southeastern limb of the Prinsep Dome. Between Karratha and the Mount Regal area the Regal Thrust has been traced for 30 km. Although this continuity of outcrop is broken by a combination of structural complexity and limited exposure at the northeastern and southwestern ends of the dome, the present interpretation is that the Regal Thrust is folded around the dome. Detailed measurement of small-scale structures would reveal the primary orientation of isoclinal folds and lineations in the Regal Thrust. Preliminary observations

4IAS AHH FIG27 10.08.01

Figure 27. Intensely folded silicic mylonite at Locality 3.8

4IAS AHH FIG28 10.08.01

Figure 28. Close-up view of isoclinal folds in the mylonite at Locality 3.8



4IAS AHH FIG29 10.08.01

Figure 29. Close-up view of a refolded isoclinal fold in the mylonite at Locality 3.8

are that lineations, which in some outcrops are visibly stretching lineations, generally plunge northeast or southwest, but exceptions occur near Mount Regal where the plunge is northwesterly. Sun and Hickman (in press) suggested that this tectonic contact may have been produced by obduction of MORB-like crust (Regal Formation) across the granite–greenstone assemblage of the Karratha domain.

Locality 3.9: Komatiite flows between the Nickol River Formation and pillow basalt of the Regal Formation (MGA 0482500E 7704900N)

Return along the water-pipeline road to the North West Coastal Highway, and take the road to Karratha. About 5 km from the highway turn left immediately before the outdoor cinema, and follow the track westwards for 2 km. Leave the track, driving 250 m northwards to the ridge at Locality 3.9.

Locality 3.9 contains outcrops of well-preserved, spinifex-textured flows of ultramafic lava. These flows are preserved within a tectonized lens of serpentinized peridotite within the Regal Thrust. This serpentinite unit is probably a tectonic slice from the base of the Regal Formation, but could also be part of the Nickol River Formation. The best exposures are located on the southern side of the ridge, and west of the end of a valley in the ultramafic succession.

Individual ultramafic lava flows are 1–2 m thick, and dip northwestwards at 10–30º. The most visually striking feature of the flows is the excellent development of sheaf-spinifex texture, in which blades of serpentinized olivine are up to 0.5 m in length. The sheaf-spinifex zones of each flow are overlain by zones of random-spinifex texture, and these underlie fractured, aphanitic, and vesicular flow tops. The basal parts of the flows are poorly exposed due to rubble from overlying flow components, but consist of massive serpentinite with locally visible olivine cumulate texture. Thin sections from similar rocks at nearby localities suggest that the olivine plates of the sheaf-spinifex zone are likely to have been completely replaced by serpentine and tremolite.

Locality 3.10: Dampier Granitoid Complex, Dampier (MGA 0472350E 7712100N)

Return to the road to Karratha, and follow this northwards and westwards, through Karratha, past Karratha Airport, and to the northern end of the causeway where the road crosses low exposures of compositionally banded granitoids.

Locality 3.10 includes outcrops on both sides of the road, about 200 m from the northwestern end of the causeway. Banded monzogranite of the Dampier Granitoid Complex contains melanocratic schlieren, and possible xenoliths of more mafic composition. About 1 km to the northwest the complex is intruded by the c. 2725 Ma Gidley Granophyre, and exposures of this are visible in a road cutting. Most of the Dampier Granitoid Complex consists of porphyritic granite–granodiorite and a more mixed assemblage of even-grained granite–granodiorite containing banded gneissic granitoids, syenogranite, and pegmatite. A sample of granite collected 7.5 km to the south was dated at 2997 ± 3 Ma (Nelson, 1998), but the cores of zircons in this rock gave $^{207}Pb/^{206}Pb$ ages ranging from 3106 to 3255 Ma. Neodymium T_{DM} model ages obtained on the Dampier Granitoid Complex range from 3387 to 3247 Ma (Hickman et al., in prep.).

The compositional banding at Locality 3.10 has a variable orientation, striking between southwest and southeast, and dipping moderately to steeply northwest or southwest. Regionally, the prevailing strike is southwesterly, with most exposures showing steep southeasterly dips of banding and the main tectonic foliation. This orientation indicates that the foliation is S_6 or S_8 (see **Episodes of deformation**). The irregular orientation of structures at Locality 3.10 is interpreted to result from its proximity to a post-Gidley Granophyre strike-slip fault about 100 m to the north.

Locality 3.11: Cleaverville Formation, Cleaverville (MGA 0497800E 7714700N)

From Dampier, return to the North West Coastal Highway, and drive about 20 km to the turnoff to Cleaverville Beach (MGA 0503600E 7705350N). Follow this road for about 13 km to the parking area at Cleaverville Creek (MGA 0497800E 7714500N). Locality 3.11 is 200 m north of this parking area (Fig. 30).

Significance of the Cleaverville area

Cleaverville is the type area of the Cleaverville Formation (Ryan and Kriewaldt, 1964), which Ryan (1965) and Hickman (1983) correlated with the Gorge Creek Group of the EPGGT. They also interpreted the Cleaverville Formation to be a major stratigraphic marker that extends across most of the North Pilbara Terrain. SHRIMP U–Pb zircon geochronology (Nelson, 1998) has indicated that all outcrops correlated with the Cleaverville Formation in the WPGGT, and outcrops of similar banded iron-formation in the extreme northwestern part of the EPGGT, are the same age (c. 3020–3015 Ma). Several attempts to precisely date the major iron formations of the Gorge Creek Group in the EPGGT have been unsuccessful. If proven, extension of the Cleaverville Formation across the North Pilbara Terrain would have major implications for interpretations of the early crustal evolution of the Pilbara Craton. In particular, it would establish that cratonization was essentially complete by 3020 Ma. In the WPGGT the Cleaverville Formation is up to 1500 m thick, and is composed of banded iron-formation, ferruginous chert, grey-white and black chert, shale, siltstone, and minor volcanogenic sedimentary rocks. Magnetite-bearing iron-formation makes the Cleaverville Formation readily identifiable on regional aeromagnetic images. In the Roebourne–Dampier area these images confirm previous conclusions from geological mapping (Ryan et al., 1965; Hickman, 1980) that the Cleaverville Formation is folded around the Prinsep Dome (Fig. 2). Outcrops at Cleaverville are located on the northwestern limb of this fold whereas extensive outcrops of the formation in the Roebourne–Wickham area are situated on the southeastern limb. Southwest of Roebourne, almost continuous outcrops of the Cleaverville Formation extend 30 km along the axial region of the Roebourne Synform. Southwest from Cleaverville the formation can be traced, using a combination of mapping and aeromagnetic image interpretation, for a distance of 70 km through Karratha to Mount Regal, Maitland River (Fig. 3), and farther west to Devil Creek. The other major outcrop of the Cleaverville Formation in the WPGGT is located at Mount Ada, 20 km south of Roebourne, and south of the Sholl Shear Zone. Correlation of the Mount Ada unit with the Cleaverville Formation is based on similarities in lithology and thickness, and on precise SHRIMP U–Pb zircon geochronology.

In the EPGGT all thick banded iron-formations of the greenstone succession are within the Gorge Creek Group (Hickman, 1983; Van Kranendonk, 1998, 2000; Williams, 1999). The Gorge Creek Group stratigraphically overlies the c. 3260–3235 Ma Sulphur Springs Group (Van Kranendonk, 1998), and underlies the De Grey Group. The depositional age of the De Grey Group is poorly constrained between a maximum age of 3048 Ma (Nelson, 1999) and c. 2940 Ma. The maximum (3048 Ma) age of the De Grey Group is based on SHRIMP analyses on a single clastic zircon within a quartzofeldspathic sedimentary rock, and the depositional age of this rock may therefore be considerably younger.

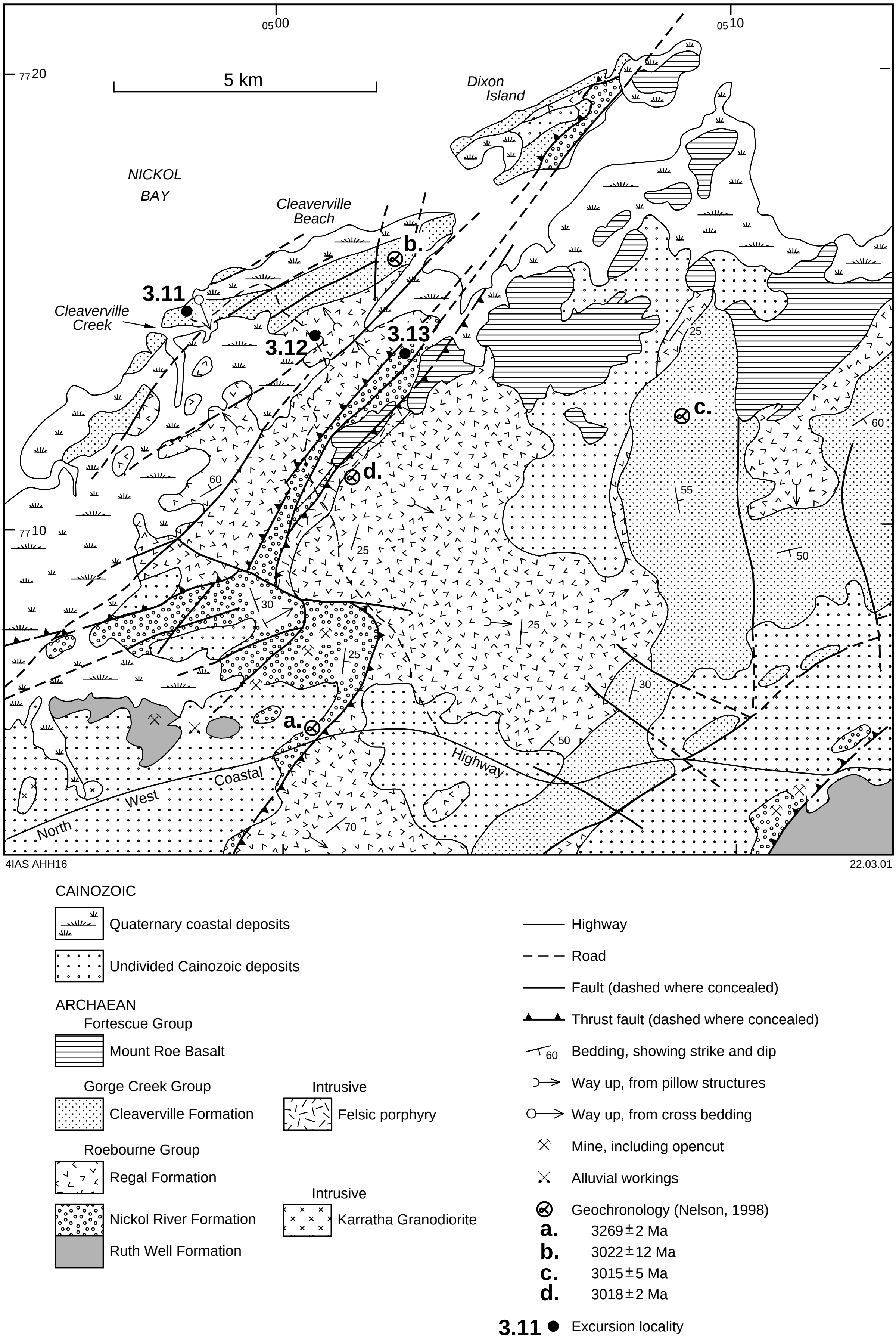


Figure 30. Geological map of the Cleaverville area, showing excursion localities 3.11–3.13

In recent years certain workers have challenged the interpretation that the Cleaverville area is part of the WPGGT. Ohta et al. (1996) claimed that the geology of the Cleaverville area indicates accretion of a MORB–trench succession onto the northwestern margin of the Pilbara Craton. Kiyokawa and Taira (1998) also interpreted the area as an accreted unit, but gave evidence supporting their conclusion that the Cleaverville succession formed in an oceanic island arc environment. A narrow belt of clastic sedimentary rocks lies along the southeastern margin of the Cleaverville area, and Kiyokawa and Taira (1998) interpreted this as a syn- to post-orogenic basin succession in a suture zone. Krapez 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.

Ohta et al. (1996) based their interpretation of the Cleaverville area on lithological associations of a tectonically repeated (seven tectonic slices) MORB–trench succession. They recognized a single succession of pillow basalt overlain by chert and banded iron-formation, with an upper assemblage of volcanic and clastic rocks. This sequence was interpreted to be diagnostic of an oceanic plate moving from mid-ocean ridge (basalt) through deep-sea pelagic (chert and banded iron-formation) to hemipelagic and trench settings. However, the interpretation given by Kiyokawa and Taira (1998) did not include the complex tectonic duplication proposed by Ohta et al. (1996), and they interpreted the Cleaverville succession as composed of three volcanosedimentary cycles. Each cycle was said to commence with basaltic volcanism, followed by rhyolite volcanism, and to be terminated by deposition of chemical sediments and volcanosedimentary rocks. The geological mapping and interpretation of the Cleaverville area published by Kiyokawa and Taira (1998) is substantially different from the interpretation by Ohta et al. (1996). Kiyokawa and Taira (1998) reject an oceanic ridge model because they interpret several basalt–rhyolite chemical-sediment cycles to be present, and because the basalts contain no oriented dyke complexes.

Ohta et al. (1996) and Kiyokawa and Taira (1998) assume that the chert and banded iron-formation units of the Cleaverville Formation were deposited in deep-water oceanic environments remote from any influx of continental material. However, this interpretation is not supported by evidence provided by sedimentary structures and mineralogy (Sugitani et al., 1998) that the Cleaverville Formation of the adjacent Roebourne area is a shallow-water deposit. Also significant is the fact that the Cleaverville Formation contains clastic zircons dated at 3461 ± 8 Ma and 3287 ± 17 Ma (Nelson, 1998). These ages coincide with those of the Warrawoona Group in the EPGGT and the Roebourne Group and Karratha Granodiorite in the WPGGT respectively, suggesting deposition adjacent to older continental crust.

The regional geology of the North Pilbara Terrain precludes interpretation of Gorge Creek Group depositional environments in terms of Phanerozoic oceanic and convergent margin settings. The stratigraphy of greenstone successions elsewhere in the North Pilbara Terrain establishes that pillow basalt, chert, banded iron-formation, and clastic sediment sequences in the Gorge Creek Group overlie lithologically diverse supracrustal successions that are over 10 km thick (Hickman, 1983; Williams, 1999; Van Kranendonk, 2000). In the northeast Pilbara the Gorge Creek Group rests directly on major erosional unconformities (Dawes et al., 1995; Williams, 1999). Such sequences were clearly not oceanic, but were deposited on continental crust. Moreover, the widespread distribution and structural geology of the Gorge Creek Group in the EPGGT is not consistent with deposition in convergent margin environments.

In the WPGGT the banded iron-formation of the Roebourne–Wickham area overlies an older granite–greenstone assemblage at least 8 km thick, and this includes sandstone and conglomerate (Hickman, 1997; 2001). As described above, aeromagnetic data and detailed mapping demonstrate that the chert–banded iron-formation units at Roebourne–Wickham and at Cleaverville are contiguous around the northeastern part of the Prinsep Dome. Kiyokawa and Taira (1998) recognized five formations in the Cleaverville succession, and named it as the 'Cleaverville Group'. However, the Cleaverville Group, as so defined, is not equivalent to the Cleaverville Formation because its lowest two formations equate to part of the Regal Formation (Ryan and Kriewaldt, 1964; Hickman, 1997). This change in lithostratigraphy has not been followed in the GSWA mapping of the west Pilbara because data from a far larger area of the WPGGT indicates that the Regal Formation should remain in the Roebourne Group (Hickman, 1997).

In summary, the Cleaverville Formation is not confined to the Cleaverville area, as would be required by the accretionary models that have been proposed. The pillow basalt unit that underlies the Cleaverville Formation is still assigned to the Regal Formation (Hickman, 1997), and is further discussed under **Locality 3.12.**

Traverse at Locality 3.11

From the parking area, walk about 200 m northwards to shoreline exposures of banded iron-formation and chert forming part of the Cleaverville Formation.

Nelson (1998) dated a thin sandstone unit, 5 km east of Cleaverville Creek, at 3022 ± 12 Ma. Geochronology on samples of the formation from elsewhere in the WPGGT has established a depositional age of 3020–3015 Ma.

Locality 3.11 (Fig. 30) provides excellent exposures of the formation in an extensive wave-cut platform and in cliffs along the shore. On the platform, beds of banded iron-formation and chert, up to 50 cm thick, form resistant ridges separated by recessive beds of deeply weathered fine-grained rock. Less altered exposures of the formation north of Roebourne indicate that the fine-grained rocks were originally probably shale or siltstone, but Kiyokawa and Taira (1998) interpreted them as altered felsic tuff. Towards the eastern end of the traverse, about 400 m from the parking area, banded mudstone and siltstone north of the banded iron-formation shows fine-scale cross-bedding that indicates way-up to the south.

At Cleaverville, the Cleaverville Formation is deformed by major, northeasterly striking, vertical faults that are almost parallel to bedding. The result is that small-scale faulting, folding and boudinage of the beds are common features in the exposures along the shore. Way-up indicators to the northwest and southeast of Cleaverville Creek section indicate that the formation locally occupies the core of an upright, isoclinal syncline.

Locality 3.12: Pillow basalt, Regal Formation (MGA 0500800E 7714100N)

From the parking area at Cleaverville Creek, drive 4 km back along the road to the northern end of a short causeway. Locality 3.12 is a rock pavement on the eastern side of the road.

At Locality 3.12, pillow basalt of the Regal Formation dips 80º towards the northwest. Here, a rock pavement on the northern edge of the small tidal inlet provides excellent cross sections through pillow structures up to 2 m across. The pillows demonstrate that the basaltic flows young towards the ridge of the Cleaverville Formation to the northwest. Similar exposures of the basalt over an area of about 20 km^2 south of the Cleaverville Formation consistently show this younging direction. Ohta et al. (1996) stated that the petrography of the basaltic succession at Cleaverville indicates greenschist, prehnite–pumpellyite and zeolite facies metamorphism. Hickman (2001) noted that this contrasted with the amphibolite–greenschist metamorphic grade of the Regal Formation in the Roebourne Synform and south of Karratha. Hickman attributed this to late-stage down-faulting of the Cleaverville area. The contact between the basalt succession and the Cleaverville Formation is exposed 2 km to the northwest where basalt is intercalated with ferruginous chert. However, it is unclear if this alternating sequence marks a transitional zone or is due to tectonic interleaving. About 2 km to the east-northeast the basalt–chert contact contains a deformed sandstone unit that has a maximum depositional age of 3022 ± 12 Ma (Nelson, 1998). About 6 km west of Wickham, thin units of clastic sedimentary rock at the base of the Cleaverville Formation provide some evidence of a disconformity.

The geochemistry of the basalt succession at Cleaverville has been studied by Ohta et al. (1996) and by Sun and Hickman (1999) to make a comparison with modern MORB. Ohta et al. (1996) found that the major-element chemistry of the basalts is tholeiitic, but differs from modern MORB in that there is little TiO_2 enrichment with increasing FeO/MgO ratio. The Cleaverville basalts also have much higher Fe contents than modern MORB. Sun and Hickman (1999) reported MORB-like chemistry with respect to Nb/Th (7–8), Nb/La (0.7 – 0.9), and slight light-REE depletion, but a difference in that Th is enriched relative to Nb. The data were inconclusive, and similar studies are required on basalts elsewhere in the North Pilbara Terrain, particularly those with better constraints on depositional environments.

Locality 3.13: Sheared contact between the Cleaverville and Karratha domains (MGA 0502750E 7713800N)

Drive 300 m southeast from Locality 3.12, and take an old track leading westwards. After about 700 m take the right fork onto another track and continue 1.5 km to an old quarry at the top of the ridge.

Locality 3.13 provides exposures of an altered shear zone between the basalts of the Cleaverville area and a narrow unit of sandstone and conglomerate. Kiyokawa and Taira (1998) named this clastic unit the ‘Lizard Hills Formation’, and interpreted it as unconformably overlying the pillow basalt succession. According to Kiyokawa and Taira (1998) at Locality 3.13 the formation includes conglomerate, sandstone, dolomite, volcanic rocks, and chert, and youngs towards the southeast. The writer’s (AHH) observations are that the sandstone dips 50º south-southeast but youngs north-northwest. Pillow basalt to the northwest dips very steeply south-southeast, but also youngs to the north-northwest. The basalt and the sandstone are separated by a sheared and brecciated zone of basalt. This zone is best exposed at the end of the ridge, 1 km northeast of the quarry. The geology to the south of the ridge is complicated by northeasterly striking faults and by partial concealment by unconformably overlying basalt of the Fortescue Group. Chert in this area is here interpreted to be part of the Nickol River Formation.

Kiyokawa and Taira (1998) interpreted this unit of sandstone and conglomerate as being of continental derivation, and deposited in a collision-zone basin after accretion of the Cleaverville succession. However, the clastic succession at Locality 3.13 outcrops continuously for a distance of 8 km to the southwest where it merges with outcrops of the Nickol River Formation. Although this continuity is complicated by faulting, the lithological similarity of the two units, and way-up evidence, suggest that the sedimentary rocks at Locality 3.13 belong to the Nickol River Formation.

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