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The Jack Hills greenstone belt, Western Australia Part 2: Lithological relationships and implications for the deposition of ≥4.0 Ga detrital zircons

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Abstract

The Jack Hills greenstone belt is situated along the southern margin of the Narryer Terrane in the northwest of the Yilgarn Craton, in fault contact with Archean granitic gneiss and granitic rocks. The belt has endured a long deformation history and the effects of recrystallisation, alteration, and the occurrence of a substantial proportion of lithologically similar metasedimentary rocks makes it difficult to correlate units and assign a stratigraphy. We have therefore divided the lithological units into four associations. These are: (1) an association of banded iron formation, chert, quartzite, mafic and ultramafic rocks; (2) an association of pelitic and semi-pelitic schist, quartzite, and mafic schist; (3) an association of mature clastic rocks including pebble metaconglomerate; (4) an association of Proterozoic metasedimentary rocks. Some of the second association was probably part of the same succession as association 1. The first two associations are of Archean age, were probably deposited at c. 3000 Ma, and have been intruded by Neoarchean granitic rocks. The mature clastic rocks of association 3 host ≥4.0 Ga detrital zircons that have been the focus of most previous work on the belt because they provide insight into early Earth processes. The mature clastic association may also have been deposited at c. 3000 Ma, but there is no direct evidence of intrusion by the Neoarchean granites, and it has a different, simpler, structural history than the first two associations. The maximum depositional age and deformation history of the Proterozoic metasedimentary rocks of association 4 suggest it may have been deposited during the late stages of the Capricorn Orogeny. This could have occurred in small transpressional pull-aparts within the Cargarah Shear Zone, which is a major, dextral, transpressional east-trending shear zone that was active during the Capricorn Orogeny between c. 1830 and c. 1780 Ma. Reactivation of the shear zone and overprinting by semi-brittle faulting has produced a complex network of fault slivers, and further sedimentation may have occurred after the Capricorn Orogeny.

Keywords: Ancient zircons; Jack Hills; Greenstones; Lithological associations; Narryer Terrane; Western Australia

1. Introduction

The Jack Hills greenstone belt and the Mount Narryer region (Fig. 1) are both well known for their siliciclastic associations that contain ≥4.0 Ga detrital zircons (e.g. Froude et al., 1983; Compston and Pidgeon, 1986; Wilde et al., 2001). The detrital zircons provide a rare record

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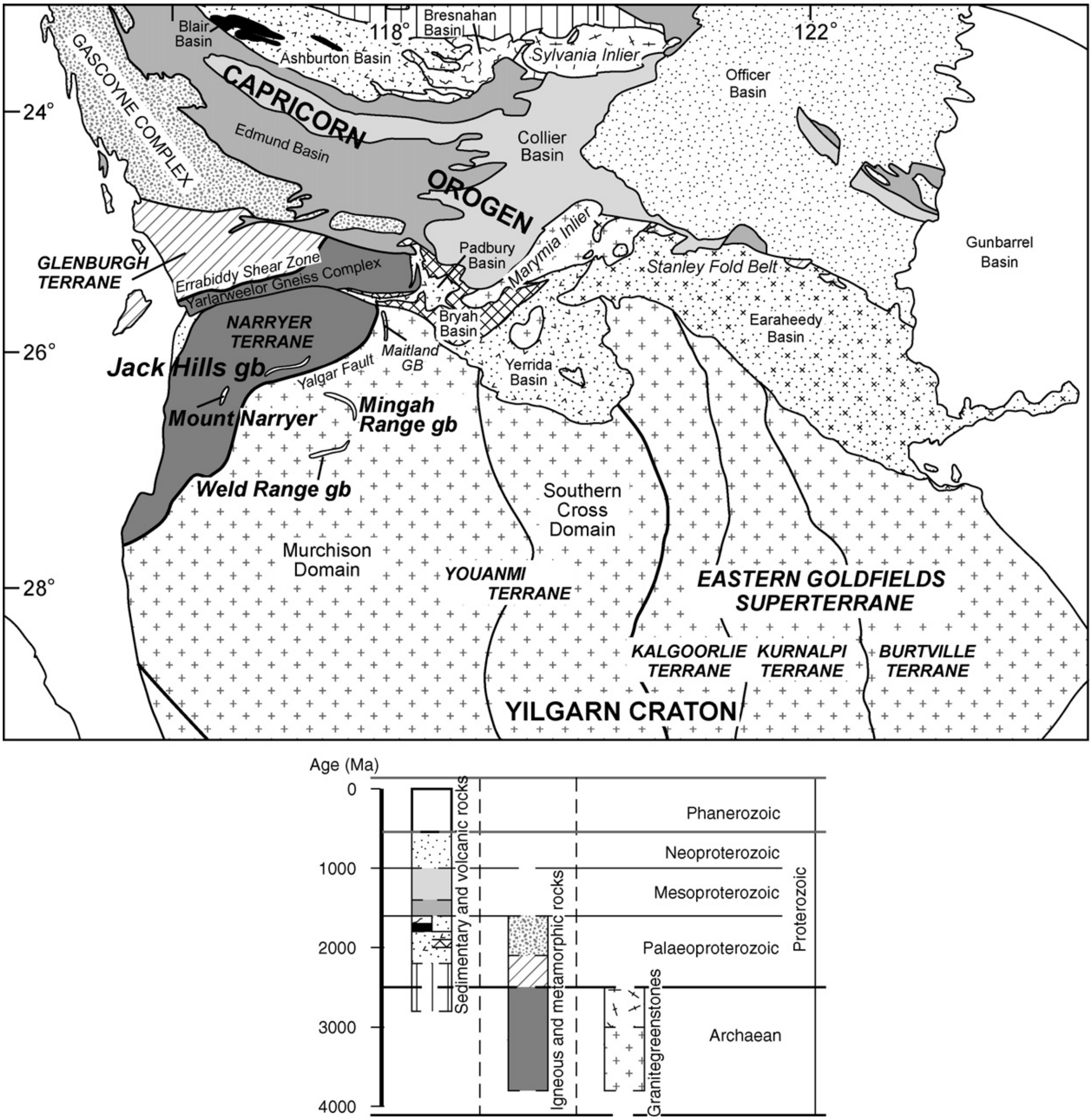

Fig. 1. Map of the Capricorn Orogen and northern Yilgarn Craton showing the location of the Jack Hills greenstone belt in the Narryer Terrane. The Youanmi Terrane is an amalgamation of what were previously the Southern Cross and Murchison Terranes (modified from Cawood and Tyler, 2004; Cassidy et al., 2006). gb, greenstone belt.

of early Earth processes, and are older than any known rocks on Earth. These zircons have been analysed for their U–Pb systematics and age data, $\delta\,^{18}$O isotopes, Hf isotopes, and geochemistry (e.g. Mojzsis et al., 2001; Valley et al., 2002; Cavosie et al., 2004, 2005; Harrison et al., 2005; Nemchin et al., 2006). The mature clastic association that hosts the ancient detrital zircons is inferred to have been deposited at c. 3000 Ma, based on the majority of detrital zircons having ages between 3.7 and 3.0 Ga (Compston and Pidgeon, 1986; Nutman et al., 1991; Cavosie et al., 2004; Dunn et al., 2005). However, the depositional age and provenance of the mature clastic association is unknown, as is its relationship to other units and associations in the belt. This is important as knowledge of the source of the $\geq$4.0 Ga detrital zircons is vital towards understanding early Earth processes.

The Jack Hills greenstone belt comprises several associations of metasedimentary and mafic–ultramafic rocks that are mostly strongly deformed, recrystallised, and tectonically disrupted. Because of this, and a lack of geochronological data, a formal stratigraphy has not been assigned to the belt. Recently, it has become clear that the belt contains both Archean and Proterozoic rocks (Cavosie et al., 2004; Dunn et al., 2005) but the nature, tectonic significance, and extent of the Proterozoic rocks are unknown. Similarly, there are insufficient geochronological data to determine the depositional ages of the Archean rocks, and whether they are all part of one succession.

This paper describes the field and petrological characteristics of all units in the Jack Hills greenstone belt, and the geological relationships between the units. The results have allowed the units to be grouped into four associations, some of which clearly differ in age. Particular emphasis is given to the mature clastic association and the geology surrounding the ancient detrital zircon sample site (W74), where most of the published geochronological data has come from (e.g. Wilde et al., 2001). Detailed maps and cross-sections of five key areas (Fig. 2), and reconnaissance mapping in other areas of the belt (Spaggiari, 2007b) have allowed the characterization and extent of specific units to be determined, provided greater context for understanding the geochronological data, and helped constrain age relationships between the units and associations. A companion paper (Spaggiari, 2007a) describes the structural evolution of the belt, and its tectonic significance in terms of the regional geology.

2. Regional geology

2.1. Narryer Terrane

The Jack Hills greenstone belt is situated along the southern margin of the Narryer Terrane, which forms the northwestern part of the Yilgarn Craton (Fig. 1). The Narryer Terrane is dominated by Paleoarchean through to Neoarchean granitic rocks and granitic gneisses (e.g. Kinny et al., 1988, 1990; Nutman et al., 1991; Pidgeon and Wilde, 1998, and references therein), interlayered with minor occurrences of deformed and metamorphosed banded iron-formation (BIF), mafic and ultramafic intrusive rocks, and metasedimentary rocks (Williams and Myers, 1987; Myers, 1988a; Kinny et al., 1990). The Jack Hills greenstone belt is the only greenstone belt within the Narryer Terrane, but an extensive belt of metasedimentary rocks occurs at Mount Narryer (Fig. 1).

The oldest component of the Narryer Terrane is a dismembered layered igneous complex known as the Manfred Complex (Myers and Williams, 1985; Myers, 1988b). Leucogabbro and meta-anorthosite from the complex have a magmatic SHRIMP U–Pb zircon age of 3730 ± 6 Ma (Fig. 3; Kinny et al., 1988). The oldest orthogneiss (Meeberrie Gneiss, Myers and Williams, 1985) is a complex mixture of various lithological associations (migmatite) and its oldest component has a similar, maximum, SHRIMP U–Pb zircon age to the Manfred Complex of 3730 ± 10 Ma (Kinny and Nutman, 1996). Both the Meeberrie gneiss and the Manfred Complex were intruded by various felsic magmas (Dugel and Eurada gneisses) until a major magmatic event at c. 3300 Ma, accompanied by deformation and high-grade (amphibolite–granulite facies) metamorphism (Kinny et al., 1988; Myers, 1988a; Nutman et al., 1991, 1993; Pidgeon and Wilde, 1998, and references therein). Further felsic magmatism is believed to have occurred at c. 3100 Ma (Nutman et al., 1991). The supracrustal associations in the Jack Hills greenstone belt and the Mount Narryer region have been considered to be deposited after these events (e.g. Compston and Pidgeon, 1986; Maas and McCulloch, 1991).

The Narryer and Youanmi Terranes are stitched by Neoarchean granites that were emplaced between c. 2750 and 2620 Ma throughout the western Yilgarn Craton (Fig. 3; Myers, 1990; Nutman et al., 1991; Pidgeon and Wilde, 1998). Although there is evidence of deformation prior to the intrusion of the Neoarchean granitic rocks, major deformation is believed to have occurred at amphibolite facies between c. 2750 and 2600 Ma (Myers, 1990). Between c. 2005 and 1960 Ma the Glenburgh Terrane was accreted to the northern Yilgarn Craton margin, during the Glenburgh Orogeny, forming the Errabiddy Shear Zone (S.J. Williams et al., 1983; Occhipinti et al., 2004). The Paleoproterozoic Capricorn Orogeny (1830–1780 Ma, Occhipinti et al., 2003) affected much of the Narryer Terrane, including the Jack Hills greenstone belt (Spaggiari et al., 2004; Spaggiari, 2007a; Spaggiari et al., submitted).

2.1.1. Mount Narryer area metasedimentary rocks

Amphibolite to granulite facies metasedimentary rocks, mainly quartzites, occur in the vicinity of Mount Narryer in tectonic contact with the granitic gneisses, and, as in the Jack Hills greenstone belt, contain detrital zircons that are ≥4.0 Ga (Williams and Myers, 1987; Froude et al., 1983; Compston and Pidgeon, 1986; Wilde et al., 2001). Rocks in the southern section of the Mount Narryer area contain a well-defined lithostratigraphy comprising five, quartzite-dominated

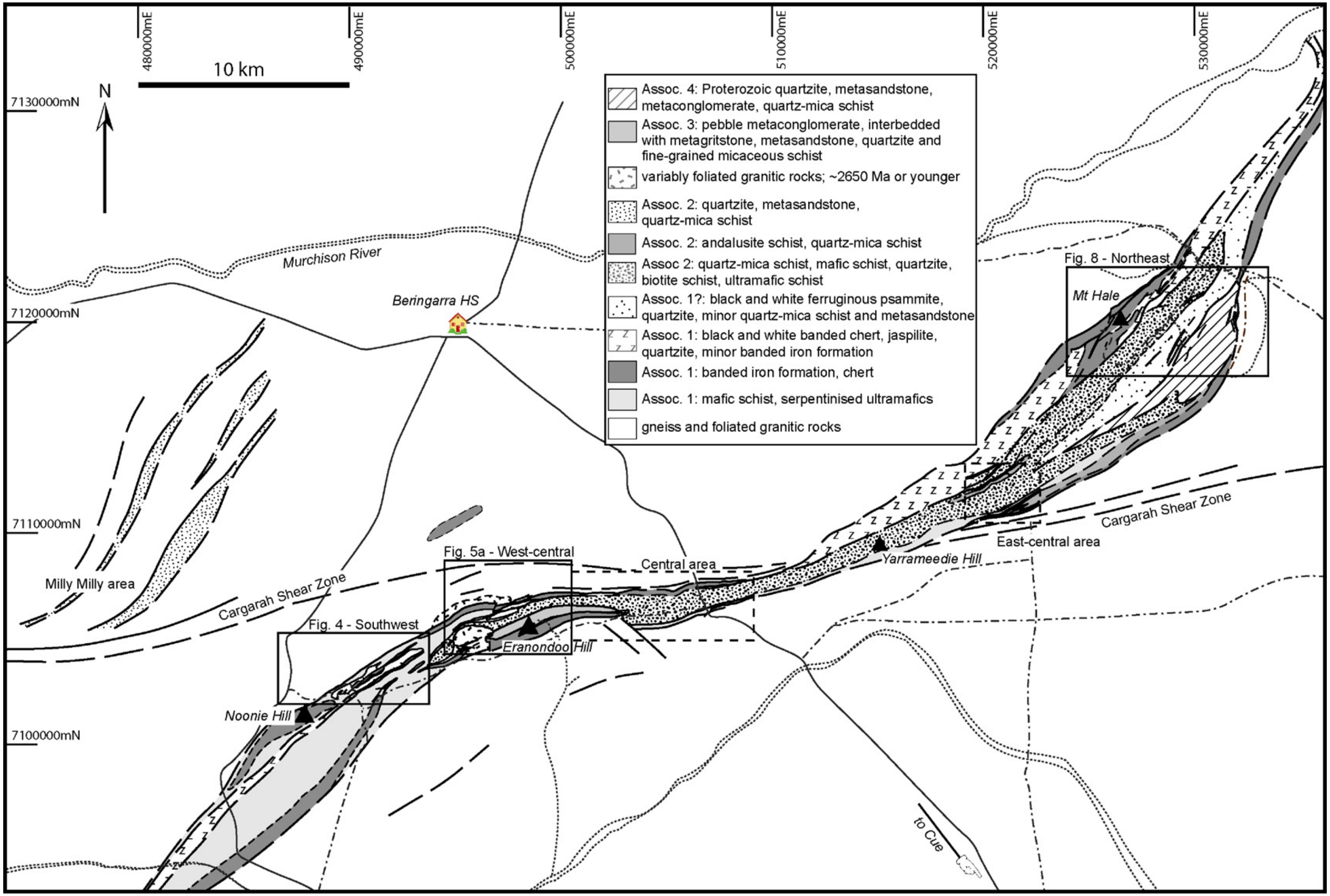

Fig. 2. Simplified geological map of the Jack Hills greenstone belt showing locations of simplified maps in Figs. 4, 5a, and 8. The lithological units are divided into associations, but it should be noted that the associations are not necessarily restricted to the areas shown. The locations of the central and east-central map areas shown in Spaggiari (2007a) are shown as dashed rectangles.

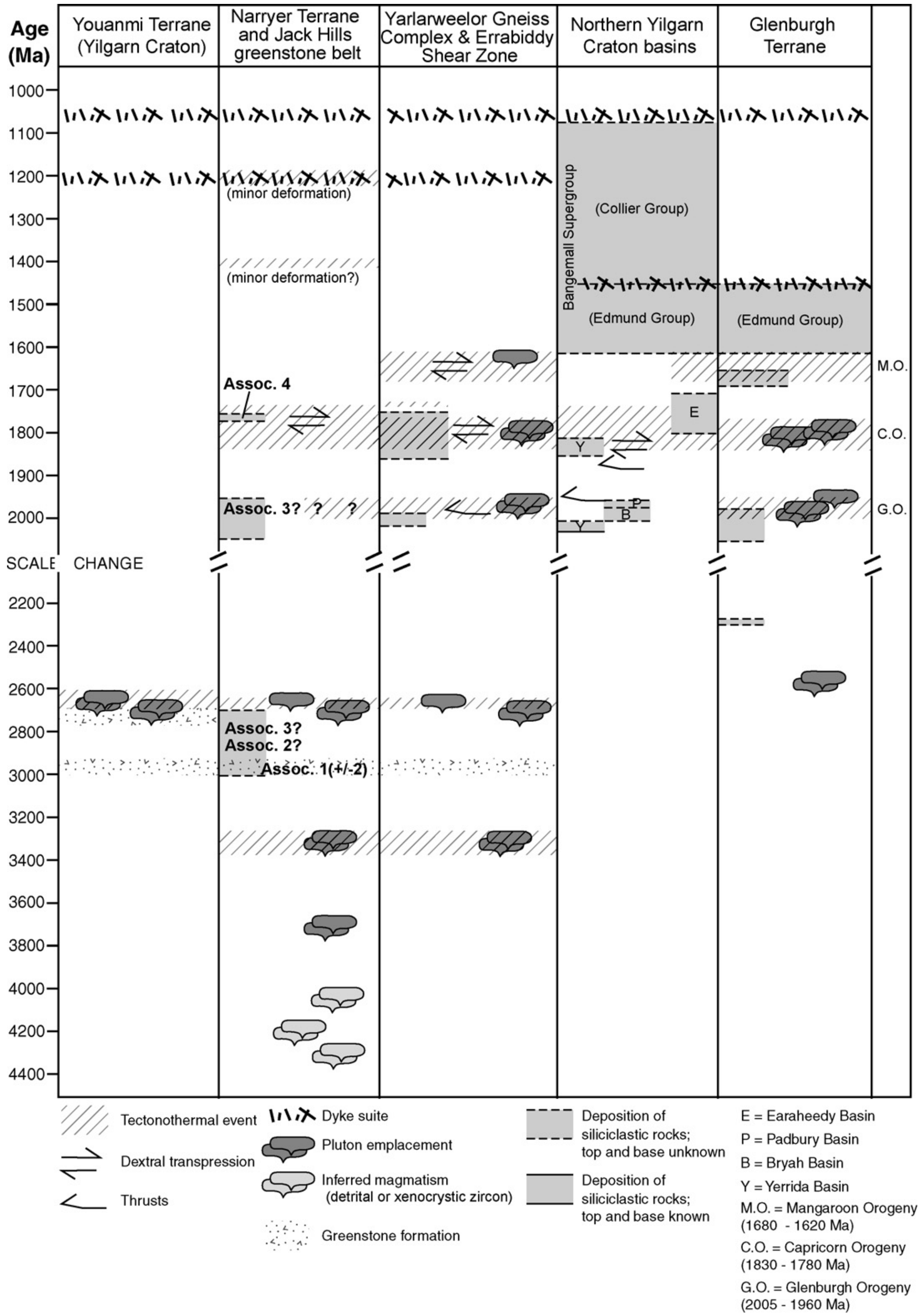

Fig. 3. Time–space diagram showing the regional relationships of greenstone formation, tectonothermal events, plutonism, and deposition of siliciclastic rocks. Associations 1–4 indicate approximate times of deposition of Jack Hills rocks. Modified from Occhipinti (2004), with additional information from Sheppard et al. (2005), and references in the text.

conformable units with tectonic contacts at the base and top (Williams and Myers, 1987). The units comprise feldspathic quartzite, garnet-sillimanite quartzite, and clinopyroxene quartzite, various gneisses including cordierite gneiss, feldspathic gneiss and calc-silicate gneiss, and both polymictic and oligomictic metaconglomerates (Williams and Myers, 1987). All of the rocks are interpreted as derived from sedimentary precursors. The metaconglomerates contain either clasts of quartz pebbles, or mixtures of subangular to subrounded clasts of banded quartz-magnetite rock, banded chert, quartzite, vein quartz, garnet-rich rock, and biotite-garnet schist (Williams and Myers, 1987). The quartzites contain abundant detrital zircons with similar ages to the surrounding granitic gneisses (Froude et al., 1983; Schärer and Allègre, 1985; Compston and Pidgeon, 1986; Kinny et al., 1988; Maas et al., 1992; Pidgeon and Nemchin, 2006), which suggests the granitic gneisses were basement to the clastic metasedimentary rocks. Lenses or stringers of quartzite, BIF, and calc-silicate gneiss occur throughout the Narryer gneisses in the vicinity of Mount Narryer (Williams and Myers, 1987).

2.2. *Paleoproterozoic basins*

Several Paleoproterozoic basins (Yerrida, Bryah, Padbury, and Earaheedy) overlie the Yilgarn Craton along its northern margin, and consist of predominantly siliciclastic and chemical sedimentary rocks, and some mafic igneous rocks (Figs. 1 and 3; Cawood and Tyler, 2004, and references therein). The Windplain Group of the intracratonic Yerrida Basin lies unconformably on Neoarchean granitic rocks of the Yilgarn Craton, and is interpreted to have been deposited at c. 2.17 Ga, based on a Pb–Pb isochron age of stromatilitic carbonate rocks (Woodhead and Hergt, 1997; Pirajno et al., 2004). The Bryah Basin is in fault contact with both the Yerrida Basin and Yilgarn Craton, and consists of a volcano-sedimentary association that includes mafic and ultramafic igneous rocks (Pirajno et al., 2004). The basin has a maximum age of 2020 Ma, and a minimum age of 1920 ± 35 Ma, determined by a Pb–Pb isochron from syngenetic pyrite (Windh, 1992; Cawood and Tyler, 2004). The Bryah Basin is unconformably overlain by the Padbury Basin, interpreted as a foreland basin, which has a maximum age of c. 2000 Ma (Nelson, 1997; Martin, 1999). It consists of dominantly clastic and chemical sedimentary rocks (Pirajno et al., 2004). Both the Bryah and Padbury Basins formed in response to the accretion of the Glenburgh Terrane, during the Glenburgh Orogeny (Occhipinti et al., 2004; Sheppard et al., 2004). The Earaheedy Basin contains clastic and chemical sedimentary rocks and unconformably overlies the Yilgarn Craton, Yerrida and Bryah Basins (Cawood and Tyler, 2004). Detrital zircons as young as 1808 ± 36 Ma from the upper part of the basin indicate deposition after that time (Halilovic et al., 2004).

3. Lithological associations in the Jack Hills greenstone belt

The dominant rock types of the Jack Hills greenstone belt are BIF, chert, and quartzite, mafic and ultramafic rocks, and siliciclastic rocks including quartz-mica schist, andalusite schist, metasandstone, and pebble metaconglomerate (Elias, 1982; I.R. Williams et al., 1983; Wilde and Pidgeon, 1990). Wilde and Pidgeon (1990) divided the supracrustal rocks of the Jack Hills greenstone belt into three informal associations: (1) BIF, chert, mafic schist (amphibolite) and minor ultramafic intrusions; (2) pelitic and semi-pelitic schists associated with mafic schists that were possibly part of a turbidite association; (3) a more restricted association of mature clastic sedimentary rocks comprising conglomerate, sandstone, quartzite, and siltstone. Association 3 was interpreted by Wilde and Pidgeon (1990) to have been deposited as an alluvial-fan, and are where the majority of $\geq$4.0 Ga detrital zircons have been found. It is not clear whether these associations were all part of the same sedimentary succession, or whether they are separate successions juxtaposed by deformation (Wilde and Pidgeon, 1990). BIF, chert, and some quartzite, mafic and ultramafic schist, and some pelitic and semi-pelitic rocks in the Jack Hills greenstone belt have been intruded by monzogranites and muscovite granite and pegmatite (Figs. 4 and 5a). The monzogranites have SHRIMP U–Pb zircon ages of 2654 ± 7 and 2643 ± 7 Ma (Pidgeon and Wilde, 1998).

The supracrustal rocks of the Jack Hills greenstone belt have not undergone the same high-grade (granulite facies) metamorphism as the gneissic rocks of the Narryer Terrane. The presence of grunerite in BIF, and the association of calcic plagioclase and hornblende in mafic schists indicate at least some supracrustal rocks were metamorphosed under amphibolite facies conditions (Wilde and Pidgeon, 1990). Hornblende in the mafic schists is commonly overprinted by actinolite, suggesting retrogression to greenschist facies. The majority of pelitic rocks are semi-pelites that lack diagnostic mineral assemblages, other than andalusite. Quartz–biotite–chloritoid assemblages in some siliciclastic rocks indicate upper greenschist facies metamorphism (Wilde and Pidgeon, 1990).

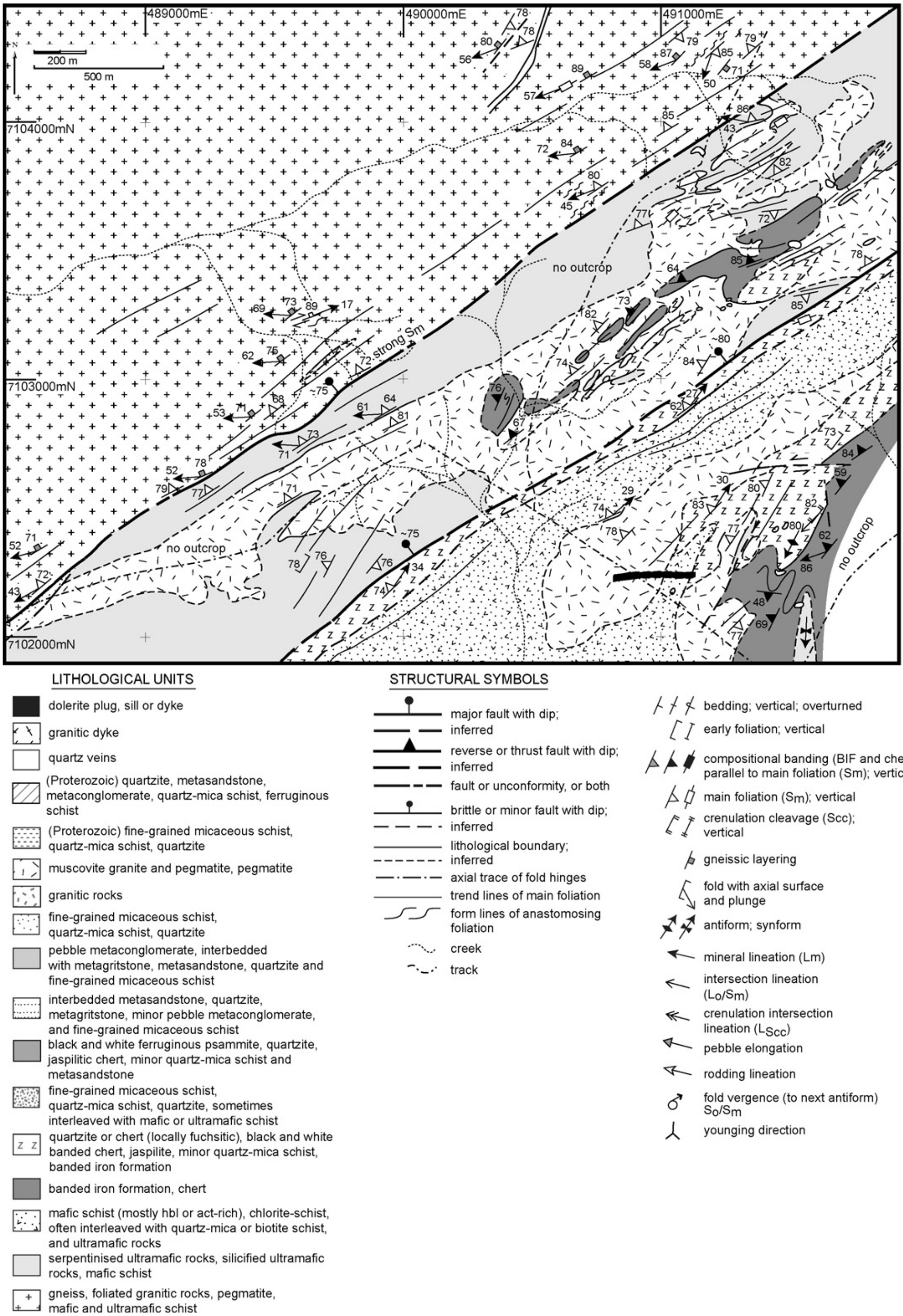

Fig. 4. Geological map of the southwestern area of the Jack Hills greenstone belt (see Spaggiari, 2007b, for more detailed maps). The legend shown also refers to the maps in Figs. 5 and 8.

(a)

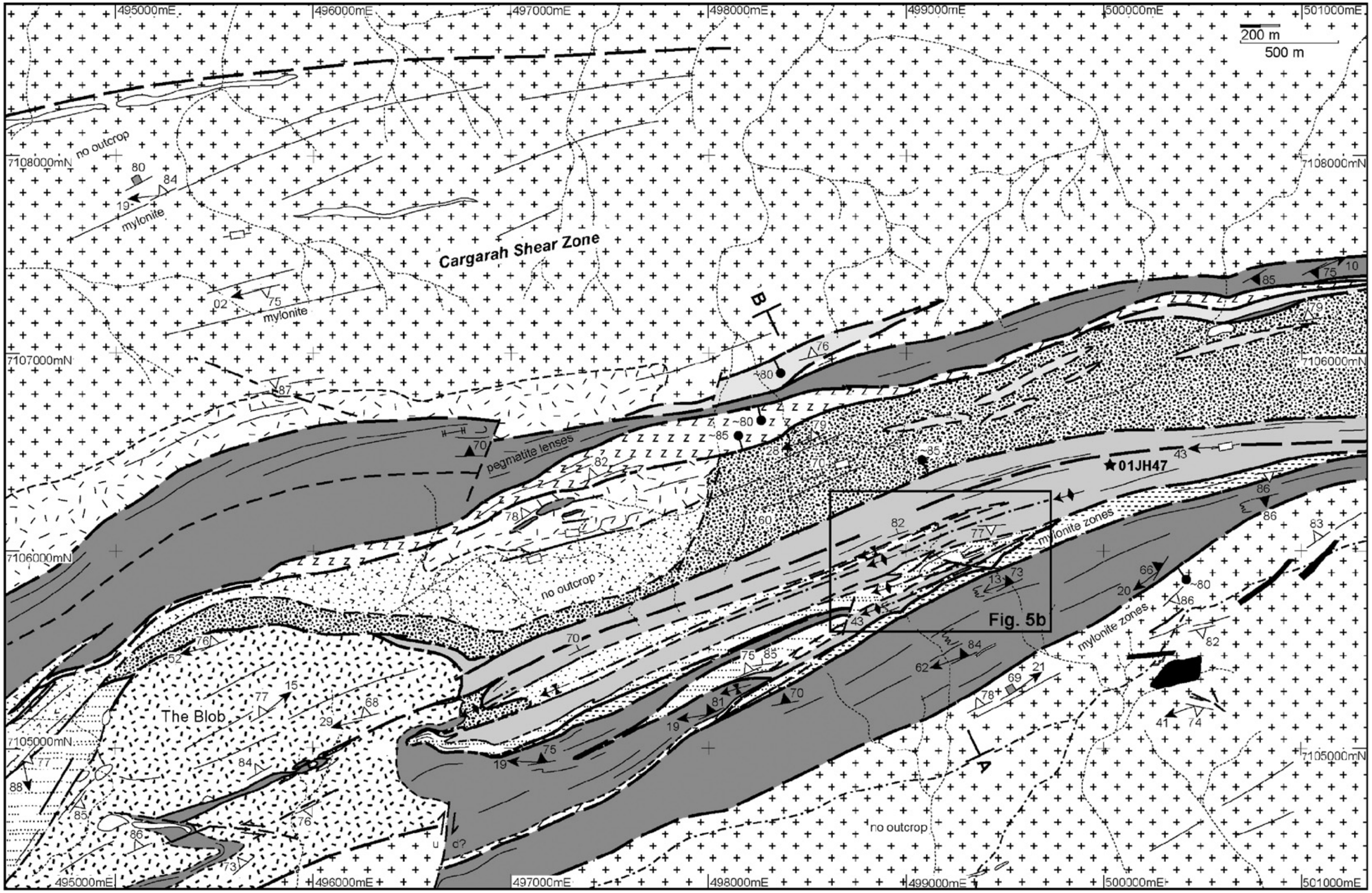

Fig. 5. (a) Map of the west-central area of the Jack Hills greenstone belt (see Spaggiari, 2007b, for more detailed maps). (b) Detailed geological map of the area including the W74 site, in the west-central area. Geochronological sample sites discussed in the text, and sites of cross-sections C–D and E–F, and lithological sections (Fig. 7), are shown. (c) Cross-sections C–D and E–F showing structural and lithological relationships in the W74 area.

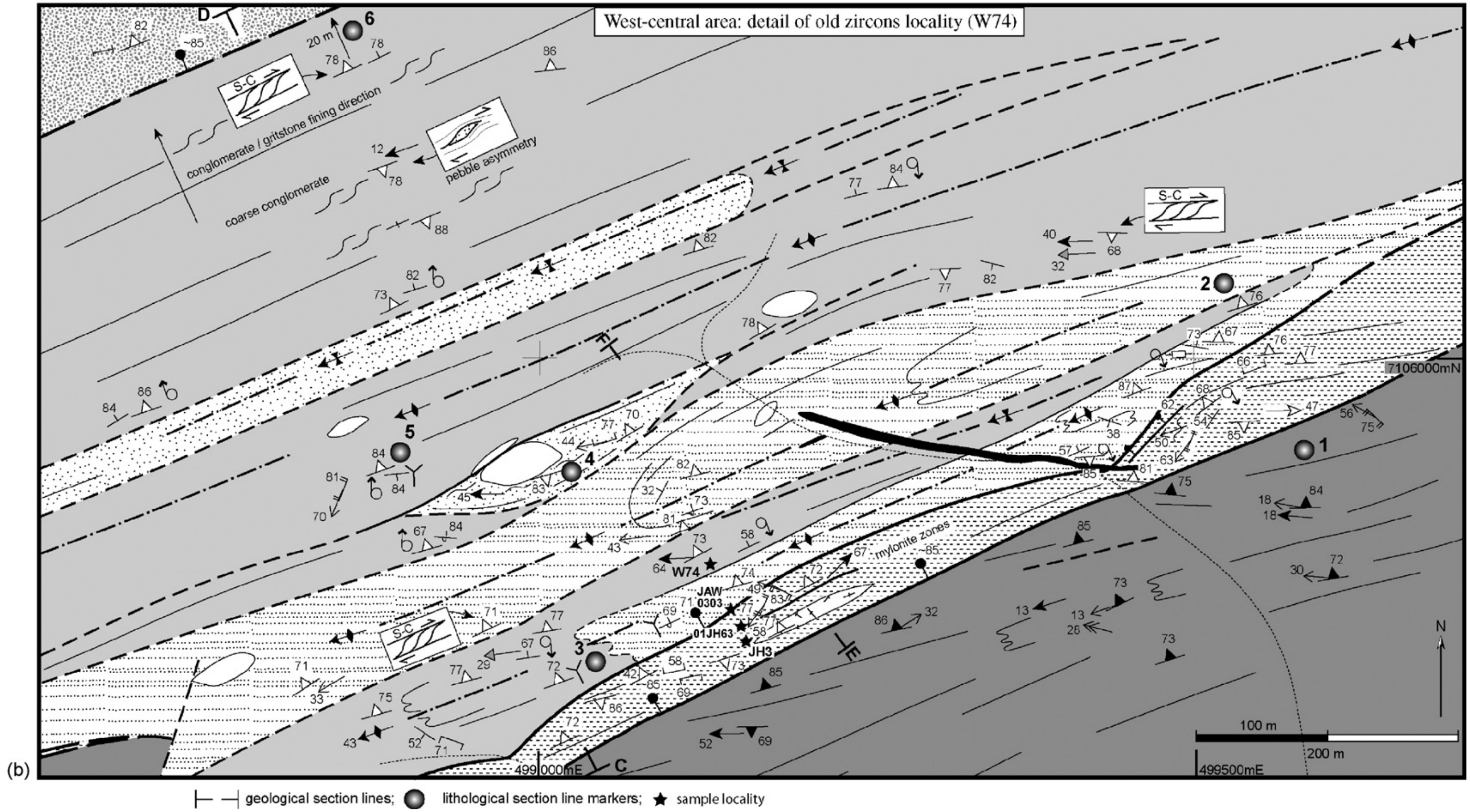

Fig. 5. (*Continued*).

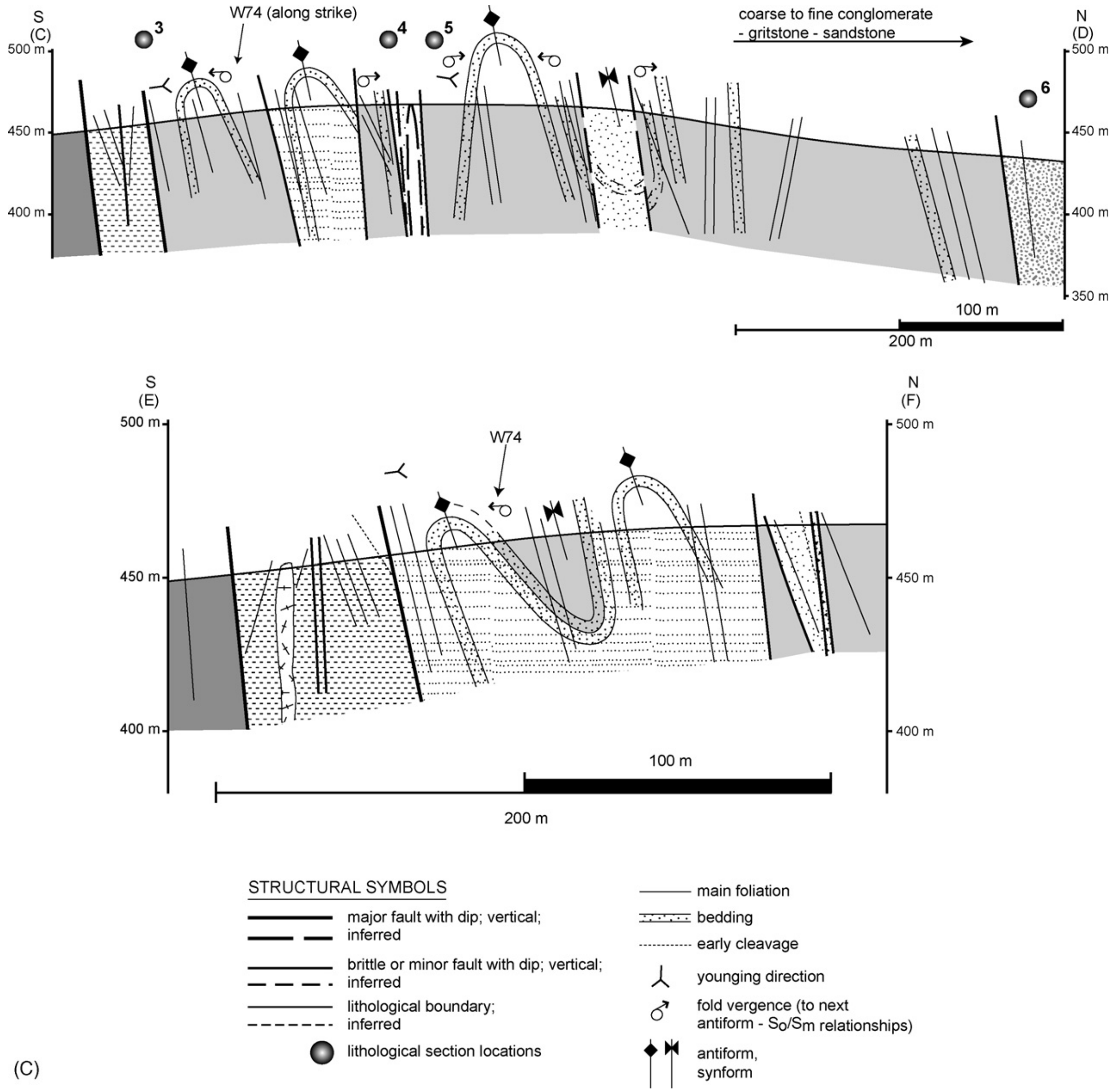

Fig. 5. (*Continued*).

3.1. Summary of SHRIMP U–Pb detrital zircon geochronology

Previously, all metasedimentary rocks in the Jack Hills greenstone belt were thought to be Archean, but it has now been recognised that some are Proterozoic (Cavosie et al., 2004; Dunn et al., 2005). However, the extent of the Proterozoic association within the belt is unknown, and it is generally difficult to distinguish it from other metasedimentary rocks in the belt due to lithological similarity, deformation, and recrystallisation. Although a significant number of detrital zircons have been analyzed from the belt, most of the work has focussed on zircons that are $\geq$4.0 Ga to study early Earth processes, and much of that material has come from a single pebble metaconglomerate outcrop in the mature clastic rocks at Eranondoo Hill in the west-central area (Figs. 2 and 5), commonly known as the W74, or discovery, site (e.g. Wilde et al., 2001). Therefore, the data are biased with respect to determination of a maximum depositional age of the association, and do not provide much insight into their provenance.

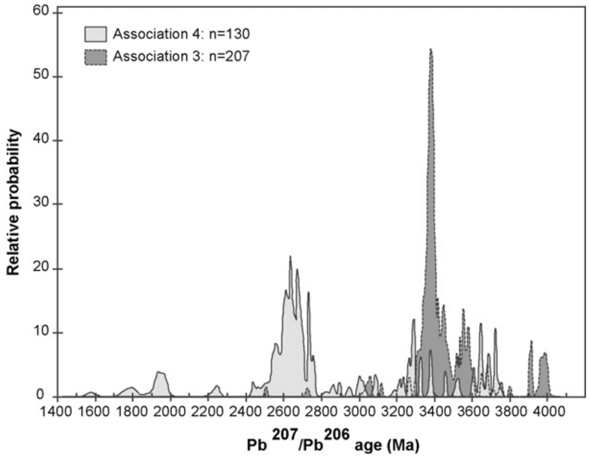

Fig. 6. Frequency histogram of detrital zircons from associations 3 and 4 that are <4.0 Ga. All analyses are SHRIMP U–Pb with 1σ errors, and are <15% discordant. Sample numbers, numbers of analyses, and data sources for association 3: W74 (also JH1), $n = 130$; JH2, $n = 63$; 01JH36, $n = 3$; 01JH54, $n = 8$; 01JH47, $n = 2$; 01JH60, $n = 1$ (Maas et al., 1992; Cavosie et al., 2004; Dunn et al., 2005; Pidgeon and Nemchin, 2006). Sample numbers, numbers of analyses, and data sources for association 4: 01JH63, $n = 6$; JH3, $n = 69$; JH4, $n = 55$ (Cavosie et al., 2004; Dunn et al., 2005).

The mature clastic rocks have yielded U–Pb detrital zircon ages up to 4.4 Ga (Wilde et al., 2001). However, from over 50,000 ≥4.0 Ga grains analyzed it has been shown that there is a complete spread of ages between 4.0 and 4.4 Ga, with the greatest percentage between 4.0 and 4.2 Ga (Cavosie et al., 2004; T.M. Harrison, pers. comm., 2005). The mature clastic rocks contain abundant detrital zircons dated between 3.7 and 3.0 Ga, indicative of deposition after that time (Fig. 6; Compston and Pidgeon, 1986; Nutman et al., 1991; Cavosie et al., 2004; Dunn et al., 2005). A few zircons younger than 3.0 Ga have been reported, but their significance is not clear (Fig. 6). Cavosie et al. (2004) analyzed detrital zircons from two transects, a western transect that included the W74 site, and an eastern transect located within more strongly deformed mature clastic rocks approximately 900 m to the east (Fig. 5a and b). Metaconglomerate sample 01JH47 (Fig. 5a) from the eastern transect yielded two grains with ages 2724 ± 7 Ma (12% discorcordant)[1] and 2504 ± 6 Ma. Metaconglomerate sample 01JH42 from the same transect also contained a single grain with an age of 2.3 Ga (A. Cavosie, written comm., 2005). Although more data are needed, these results suggest that the mature clastic rocks are a separate succession that was deposited after intrusion of the Neoarchean granites. This is consistent with a lack of evidence of intrusive relationships with Neoarchean granites, and with the structural history, which shows that the mature clastic association lacks evidence of an early deformation event that is present in banded iron-formation, chert, quartzite, mafic and ultramafic rocks, and pelitic and semi-pelitic schist of associations 1 and 2 (Spaggiari, 2007a).

Proterozoic detrital zircons have been reported from quartzite sample 01JH63 (Figs. 5b and 7b; Cavosie et al., 2004). It contained late Archean grains (2736 ± 6, 2620 ± 10, and 2590 ± 30 Ma), two Paleoproterozoic grains (1973 ± 11 and 1752 ± 22 Ma), and a single grain with an age of 1576 ± 22 Ma (Fig. 6). This quartzite is from the southern end of the western transect of Cavosie et al. (2004), close to the contact with the BIF (Fig. 5b), and is interpreted as part of a Proterozoic association of predominantly micaceous schist and quartz-mica schist. Dunn et al. (2005) analyzed a fine-grained micaceous schist from what appears to be the same association (sample JH3, Figs. 5b and 7b). That sample contained Paleoproterozoic zircons ranging in age between approximately 1944 and 1981 Ma, with a single, youngest grain dated at 1791 ± 21 Ma (Fig. 6). The sample also contained a significant proportion of late Archean grains. These data suggest the association was potentially deposited in the late stages of, or following, the Capricorn Orogeny, with a detrital contribution from rocks formed during the Glenburgh Orogeny to the northwest (Dunn et al., 2005). The single grain of 1576 ± 22 Ma (Cavosie et al., 2004) suggests the association may be even younger. However, this age does not correspond with any known sources of detritus from the region. Sample JH3 also contained a single zircon with an age of 4113 ± 3 Ma, plus several zircons in the range of 3500–3300 Ma (Fig. 6; Dunn et al., 2005). This shows that the source of the ≥4.0 Ga zircons was still present and being eroded during the Paleoproterozoic, or that these zircons were recycled from other, younger sources. Neoarchean granites could have provided this detritus as they locally contain xenocrysts of ≥4.0 Ga zircons (Nelson et al., 2000); however, zircons of this age from these granites are rare.

Paleoproterozoic zircons were also found in a metaconglomerate from the northeastern part of the belt, near the eastern margin (sample JH4, Figs. 2 and 8; Dunn et al., 2005). The metaconglomerate differs from the association of mature clastic rocks at Eranondoo Hill in that it is inferred to be a minor part of a thick succession of predominantly quartzite and metasandstone.

[1] Unless stated otherwise, all U–Pb analyses quoted in this paper are ≤10% discordant.

The sample contained a similar range of ages to JH3, with the youngest zircon dated at 1884 ± 32 Ma (Fig. 6). Older zircons range between 3725 and 3500 Ma (Dunn et al., 2005). Dunn et al. (2005) argued that at least two sedimentary associations are present in the Jack Hills greenstone belt; an older association (middle to late Archean) containing the metaconglomerate outcrop W74, and a younger association (Paleoproterozoic), represented by the outcrops of JH3 and JH4. They suggested the younger succession might have been part of either the Mount James Formation of the Capricorn Orogen, or part of the Earaheedy Basin sediments (Figs. 1 and 3).

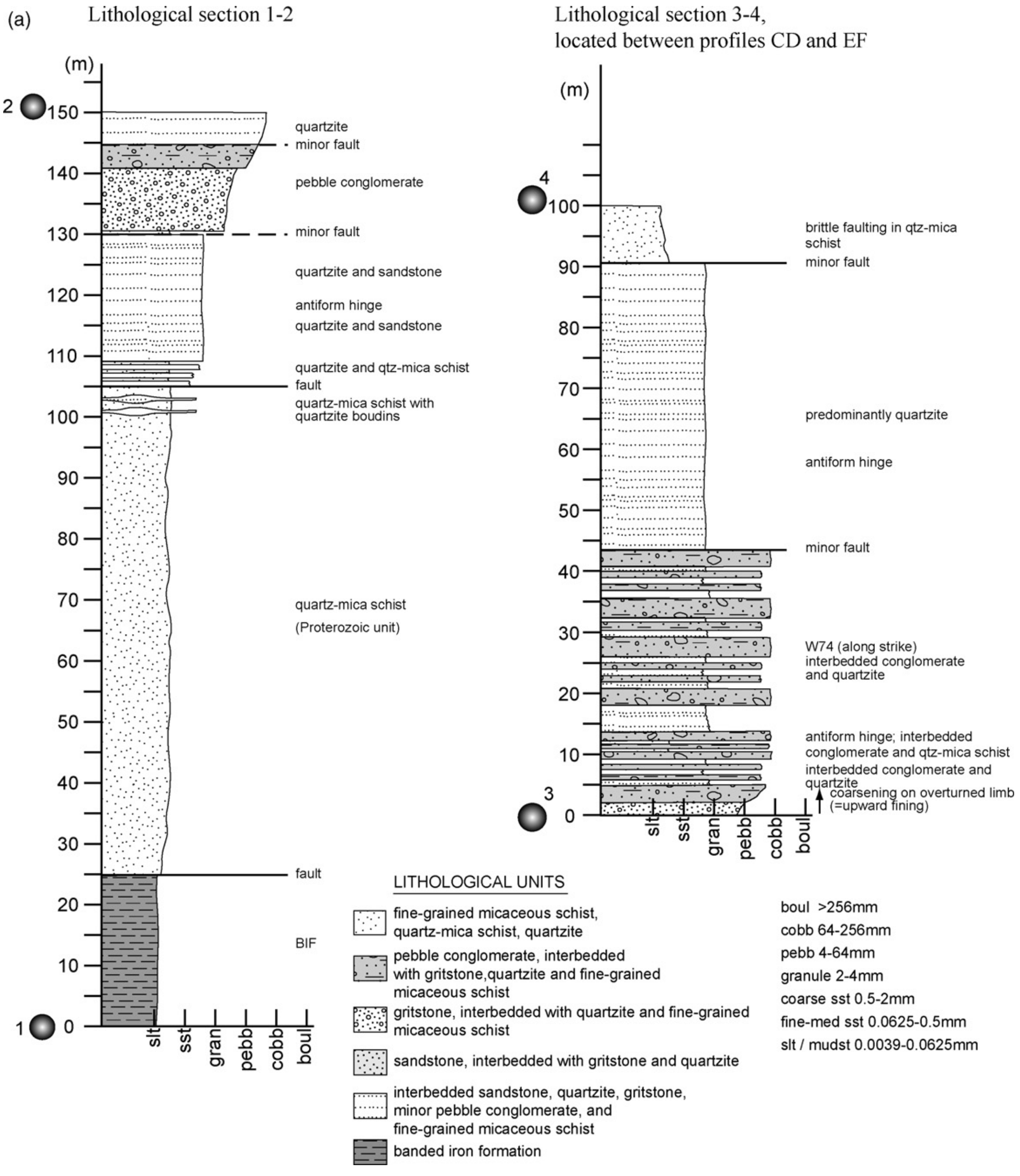

Fig. 7. (a) Lithological logs of the sections between points 1 and 2, and points 3 and 4 (see Fig. 5b and c for locations). (b) Lithological logs of the sections between points 5 and 6, and a 60 m section commencing with outcrop W74 (see Fig. 5b and c for locations). Sedimentary grain and lithic size definitions are after Boggs (1987).

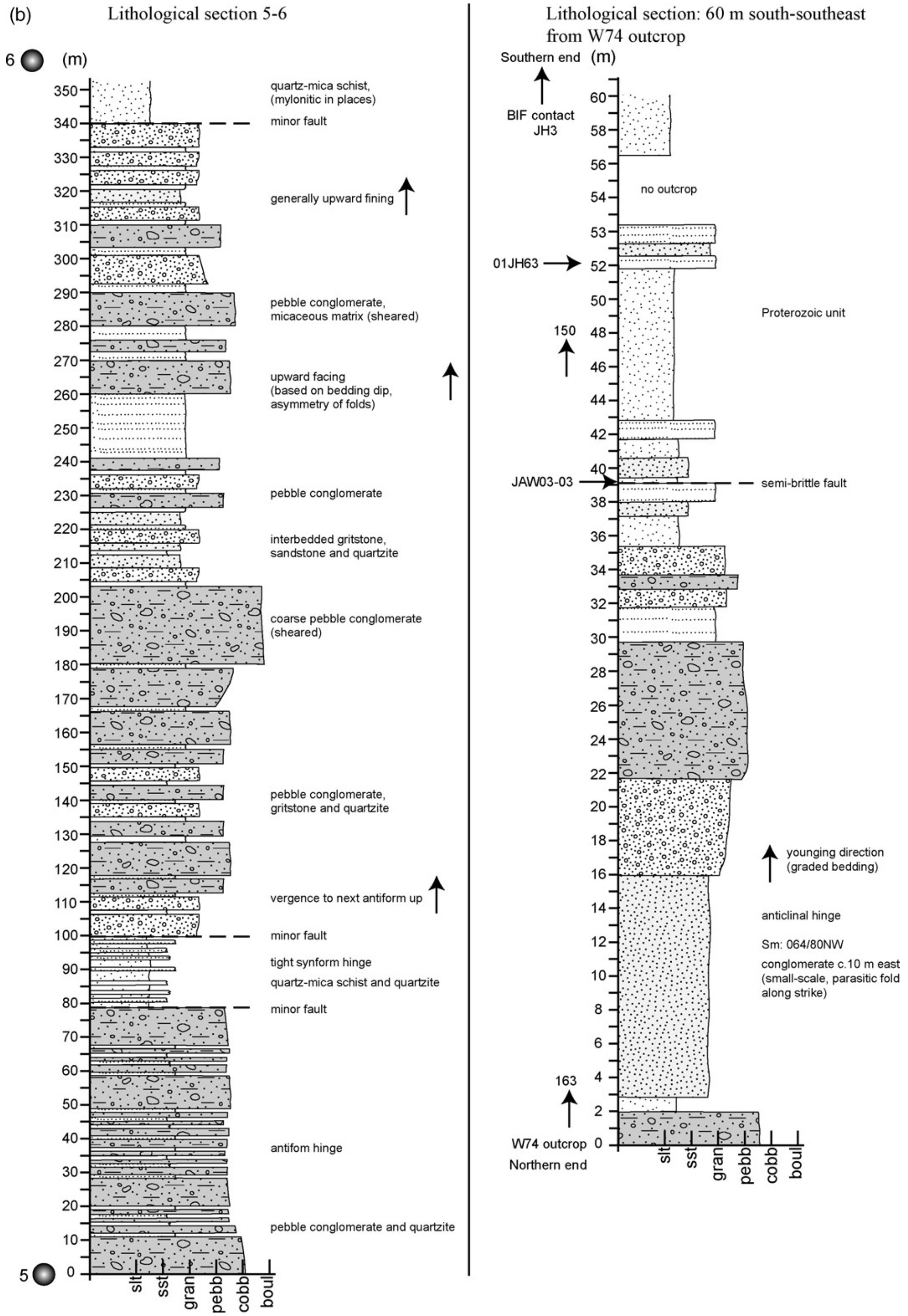

Fig. 7. (*Continued*).

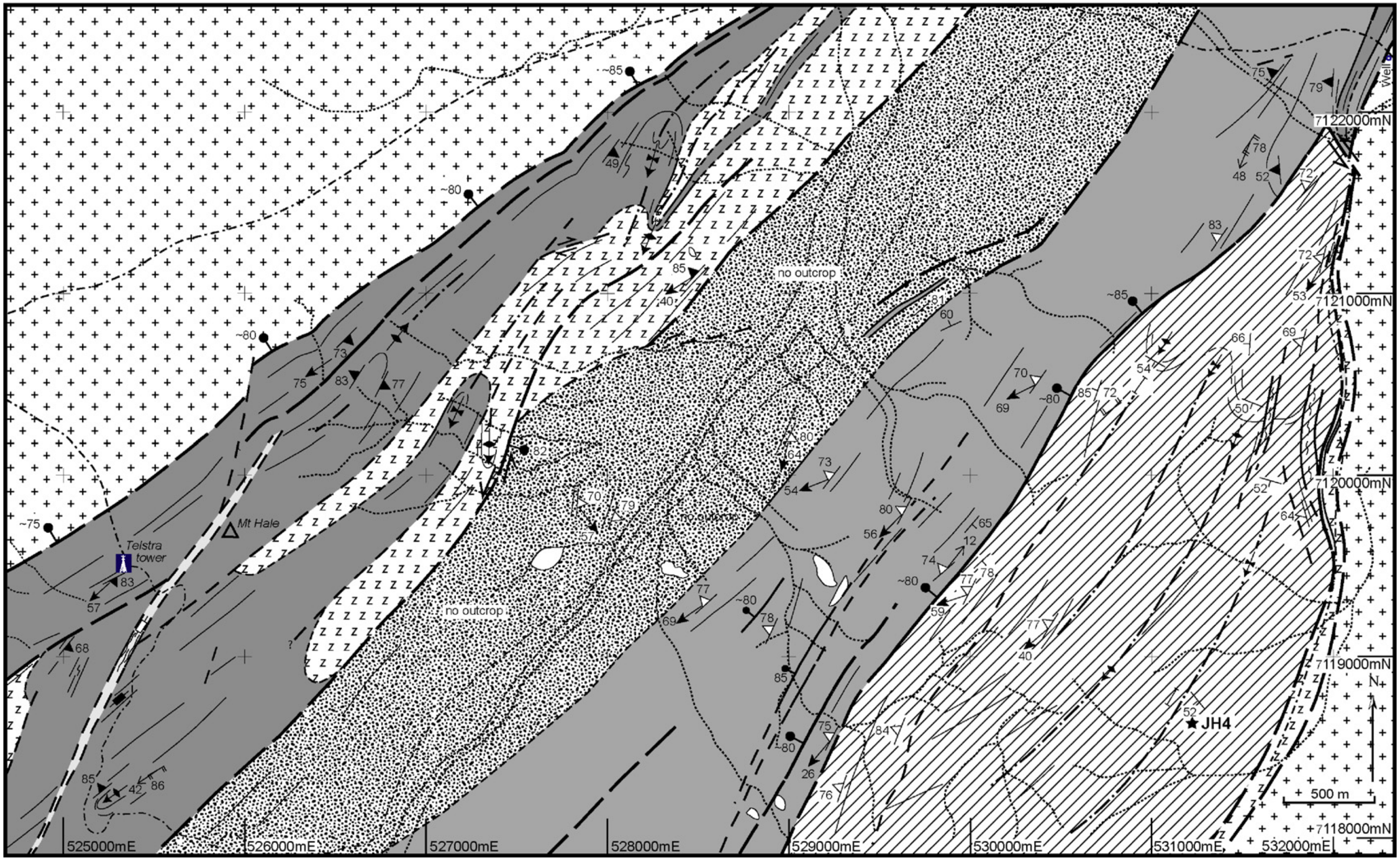

Fig. 8. Geological map of the northeastern area of the Jack Hills greenstone belt (see Spaggiari, 2007b, for more detailed maps).

4. Characteristics and relationships of the lithological units

The discovery of Proterozoic rocks within the Jack Hills greenstone belt has highlighted the difficulty of dividing lithologically similar, strongly deformed and extensively recrystallised rocks into separate successions. Because of this, no attempt has been made to construct a stratigraphy for the Jack Hills greenstone belt. The presence of numerous shear zones and faults, and extensive silicification, also makes stratigraphic subdivisions, and to some extent lithological unit subdivisions, problematic. Intrusive relationships are also difficult to determine for some units due to extensive tectonism.

The three associations described by Wilde and Pidgeon (1990) are similar to those described below. It is now evident that some of the rocks from association 2 of Wilde and Pidgeon (1990) are Proterozoic in age, whereas others are intruded by Neoarchean granitic rocks. Therefore, that association consists of at least two separate successions. Detailed descriptions of the rocks are presented below and in Spaggiari (2007b).

4.1. Association 1—BIF, chert, quartzite, mafic and ultramafic rocks

Interbedded BIF, chert, and quartzite are widespread and form a considerable component of the Jack Hills greenstone belt (Fig. 2). The BIF is locally interbedded with quartz-mica schist and quartzite. Extensive polydeformation and recrystallisation has masked much of the primary texture of the BIF, chert, and quartzite. These rocks typically exhibit strong banding delineated by Fe-oxide-rich (±stilpnomelane) and Fe-oxide poor layers. The banding may locally represent original bedding, but it is clear that, in many instances, it is a secondary layering, and was produced by deformation and metamorphism (Fig. 9a). Locally, the Fe-oxide layers have been remobilised into an oblique layering defining a crenulation cleavage (Fig. 9b). Where this is fully developed and accompanied by strong recrystallisation, it is impossible to define the original layering.

The BIF varies from fine-grained (approximately 1 mm grains) to coarse-grained (approximately 5 mm grains). These effects are largely due to deformation and metamorphism, where dynamic recrystallisation and mylonitisation has produced grain size reduction, or static recrystallisation has produced a coarse, equigranular texture. In quartzite, small, sparse, aligned white mica and/or fuchsite is locally present in the main foliation.

The BIF includes horizons of hematite-rich lenses with crystals up to 1 cm in diameter. These lenses are secondary and cross-cut the main layering of the BIF (Fig. 9c). Extensive quartz veins throughout the unit indicate a significant degree of silicification, which may have been associated with remobilisation of iron. Silicification has also produced some quartzites (recrystallised and metasomatised chert) that can be difficult to distinguish from the siliciclastic quartzites. They typically contain small, discontinuous bands of Fe-oxide, suggesting they were originally part of the BIF and chert unit.

In the central part of the northeastern region (Figs. 2 and 8), a unit of black and white quartzite is interbedded with minor amounts of chert, metasandstone and quartz-mica schist. The quartzite commonly has diffuse, often discontinuous, black and white banding (Fig. 9d). It is typically silicified, with extensive, sometimes massive, quartz veins. Epidote is common near faults. The black bands contain a high percentage of small Fe-oxide grains, which may relate to original bedding layers. Alternatively, the black Fe-oxide layers may be at least in part derived by metasomatism, potentially leached from BIF during silicification. The banding is locally folded into small-scale, isoclinal folds. The age of this unit, and its relationship to other lithological units in association 1 is unknown. Locally, the quartzite looks similar to banded chert within the BIF and chert unit. However, this unit differs from the BIF, chert, and quartzite unit by being dominated by quartzite, and interbedded with metasandstone. It is therefore interpreted as a separate unit.

The Jack Hills greenstone belt contains a significant component of mafic and ultramafic rocks, most of which are strongly deformed and schistose. These rocks are commonly interlayered with BIF, chert, and quartzite as metre-scale to 100 m-scale slivers or lenses. Mafic schists are also commonly interlayered with pelitic or semi-pelitic schists. The mafic schists commonly have assemblages of hornblende and/or actinolite, plagioclase, titanite, Fe-oxide, ±chlorite, ±biotite, ±epidote, ±quartz (Fig. 10a). Some contain abundant garnet and amphibole, with titanite, quartz, chlorite, and minor epidote and white mica (Fig. 10b). Ultramafic rocks are commonly serpentinized, and locally contain talc and/or amphibole.

4.2. Association 2—pelitic and semi-pelitic rocks

Quartz-mica schist and andalusite schist are common, particularly in the northeastern and central parts of the belt. Their primary relationship to other units in the belt is unknown and contacts are typically sheared or faulted. They are commonly associated with mafic

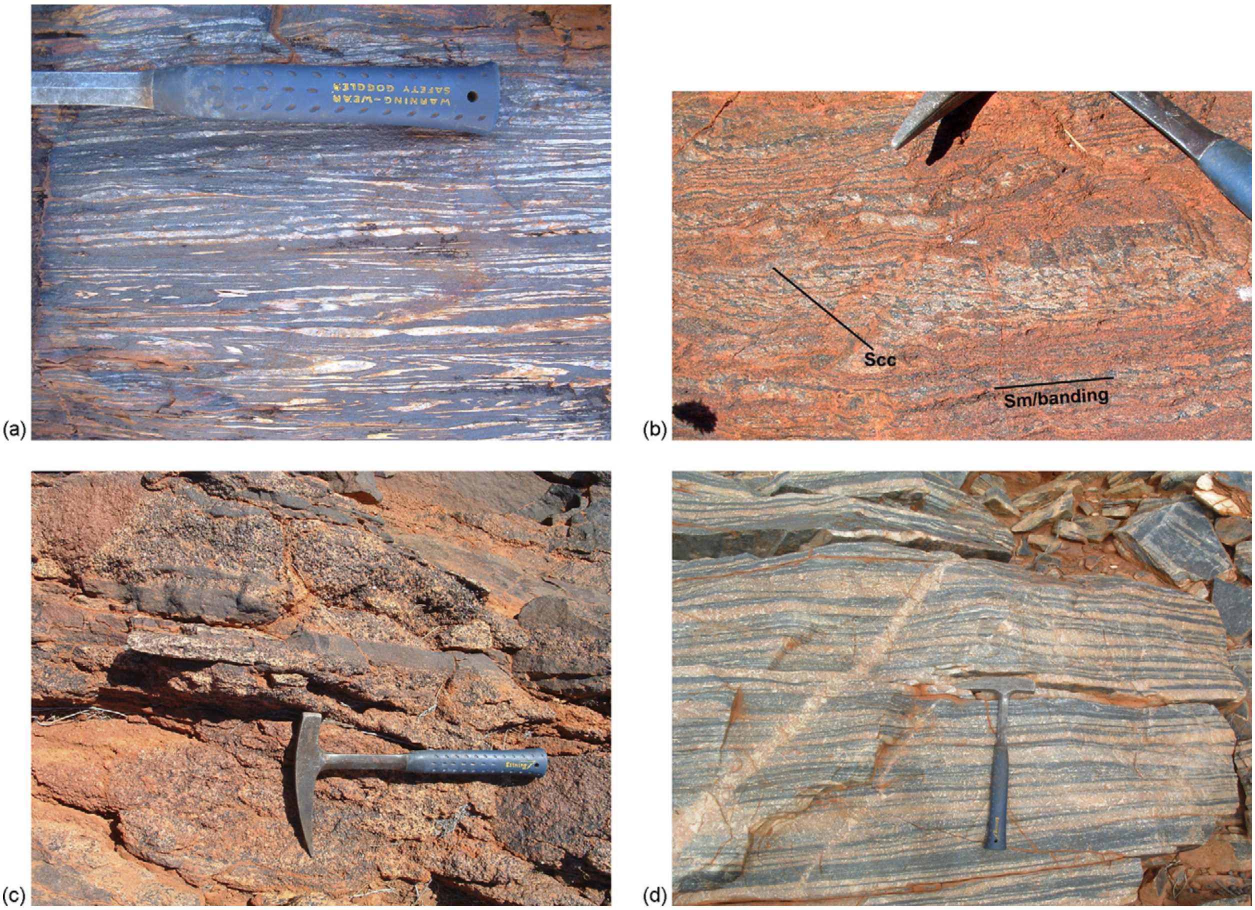

Fig. 9. (a) Transposed foliation in BIF from the northeastern map area (Fig. 8). (b) Incipient crenulation cleavage development in BIF, showing migration of Fe-rich layers into new cleavage planes. S_{cc}, crenulation cleavage; S_m/banding, a combination of the main foliation and lithological banding. Photograph taken in the central part of the Jack Hills greenstone belt. (c) BIF with horizons of hematite-rich lenses that cross-cut the main layering. Photograph taken in the northeastern map area (Fig. 8). (d) Photograph of black and white banded quartzite, taken in the northeastern map area (Fig. 8).

schists, chlorite-rich schists, and quartzite. Andalusite is a common component of most of the pelitic and semi-pelitic rocks, particularly in schists in the east-central area (Fig. 2). Quartzite is locally fuchsitic, or chert-like, for example, in the southwestern area (Fig. 4).

Quartz-mica schists are strongly foliated, with the foliation defined by white mica, quartz, and in some cases chlorite or biotite. Locally, the foliation is mylonitic. Andalusite porphyroblasts are commonly overgrown by sericite, white mica, and quartz that form part of the main foliation. This suggests that andalusite grew prior to development of the main foliation. Some quartz-mica schists contain two generations of white mica, with larger muscovite grains overgrown by smaller white mica grains that define the main foliation. Other schists are rich in biotite and contain elongate lenses of quartz and muscovite, and numerous small grains of Fe-oxide. Chlorite-rich schists have a strong foliation defined by chlorite and quartz, and locally, biotite. In the southwestern part of the belt, pelitic schist contains an early assemblage of garnet, staurolite and andalusite, overgrown by chloritoid and chlorite, consistent with retrogression from amphibolite facies to greenschist facies.

4.3. Association 3—mature clastic rocks

The association of mature clastic rocks is confined to the central and west-central areas of the belt (Figs. 2 and 5a). Similar rocks have been reported at Yarrameedie Hill (Fig. 2, Wilde and Pidgeon, 1990). The contacts of the mature clastic association with the other associations are faulted. The association consists of interbedded pebble metaconglomerate, metagritstone, metasandstone, metasiltstone, quartzite, and quartz-

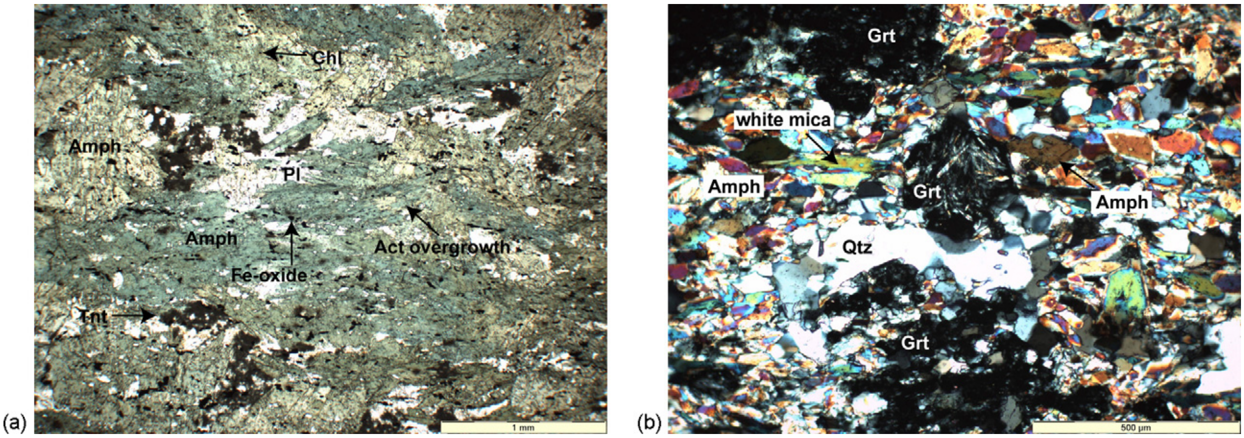

Fig. 10. (a) Photomicrograph, in plane-polarized light, of mafic schist from the east-central area of the belt (Fig. 2). Hornblende amphibole (Amph) is overgrown by actinolitic (Act) amphibole. Other phases are plagioclase (Pl), titanite (Tnt), Fe-oxide, and chlorite (Chl). Sample CS0378; map coordinates: MGA 520847mE, 7111151mN. (b) Photomicrograph, in cross-polarized light, of garnet-mafic schist from the central area of the belt (Fig. 2). Garnet (Grt) is rimmed and partially overgrown by amphibole (Amph). Other phases shown are quartz (Qtz), and white mica. Sample CS03100; map coordinates: MGA 505270mE, 7105938mN.

mica schist. Wilde and Pidgeon (1990) interpreted the association as a deltaic alluvial-fan deposit, based on well-preserved sedimentary structures (cross-bedding, graded bedding) and upward-fining sequences (Fig. 7). In the Eranondoo Hill area (Fig. 5b), pebble metaconglomerate, metagritstone, and metasandstone are the dominant lithologies. A faulted-out synform containing quartz-mica schist and minor quartzite (lithological section 5–6, Fig. 7b) suggests a succession of mudstone-dominated rocks may have been deposited above the coarser-grained, alluvial-fan sediments (see also Wilde and Pidgeon, 1990). This is supported by the gradual thinning and disappearance of conglomerate beds in quartz-mica schist towards the faulted contact with quartz-mica schist, quartzite, and mafic schist to the north (Fig. 5a and b). However, the facing direction near the faulted contact (assumed to be to the north, consistent with bedding dips) is not well constrained.

The metaconglomerate and metagritstone consist of rounded and flattened, mostly white quartz pebbles, minor chert and sandstone pebbles, from approximately 0.5–30 cm in length (Fig. 11a and b). Some of the pebbles have a similar appearance to the black and white quartzite described above from the northeastern area (Fig. 11c). Black and white quartzite also occurs locally as thin lenses within the association (Fig. 11d), suggesting the possibility that the conglomerate may include clasts of metasedimentary rocks derived from other units in the belt. The pebbles are supported by a sandy and micaceous matrix (Fig. 11e). The metaconglomerate typically has a strong foliation that anastomoses around aggregates of quartz (pebbles or clasts) (Fig. 11f). The foliation is defined by quartz, and strongly aligned micaceous and locally chloritic or fuchsitic layers that are part of the matrix.

4.4. Association 4—Proterozoic metasedimentary rocks

Two areas of Proterozoic rocks have been recognised by SHRIMP U–Pb detrital zircon dating (Cavosie et al., 2004; Dunn et al., 2005). Two thin lenses, approximately 50–100 m wide, of quartz-mica schist with minor interbedded, mostly thin beds of quartzite, crop out between BIF and the mature clastic rocks for a combined strike length of approximately 4 km (Fig. 5a–c). The lenses are fault-bounded and contain the sample sites 01JH63 and JH3 (Figs. 5b and 7). The schist is indistinguishable from other quartz-mica schists in the belt. They are mostly mylonitic (Fig. 12), and contain boudinaged quartz veins. Lithological section 1–2 (Fig. 7a) contains boudinaged quartzite beds at the northern end of the Proterozoic unit. This may indicate a change in depositional conditions, but it is not clear which way the unit faces.

In the northeastern part of the belt, a unit of interbedded quartzite and metasandstone, with minor interbedded metaconglomerate and quartz-mica schist is inferred to be of Proterozoic age, based on the presence of Paleoproterozoic detrital zircons from sample JH4 reported in Dunn et al. (2005) (Fig. 8). The quartzites in this unit are typically massive, and some are veined and silicified. The presence of andalusite indicates the rocks have been metamorphosed to at least greenschist


(a)
(b)
(c)
(d)
(e)
(f)
1 mm

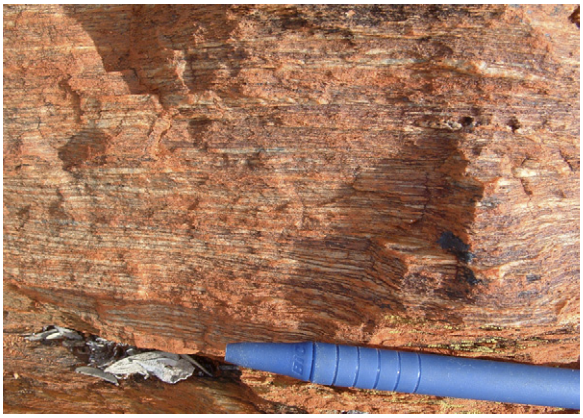

Fig. 12. Mylonitic Proterozoic quartz-mica schist from the west-central map area (Fig. 5a).

grade. Lenses of metaconglomerate occur in zones of tectonic disruption, and consist of flattened quartz pebbles in a micaceous matrix, however, it is not always clear whether these are true metaconglomerates, or zones of disrupted, interbedded and veined quartzite and quartz-mica schist (Fig. 13a). In the southeast-central part of the northeastern area (Fig. 8), one of these zones is interpreted as fault mélange because it has a variety of clast sizes in one, approximately 8 m wide horizon, and is adjacent to a major zone of shearing overprinted by brittle faulting (Fig. 13b). Rocks adjacent to this zone are coherent quartzite and metasandstone.

4.5. Granitic rocks and intrusive relationships

Neoarchean granitic rocks and pegmatite intrude granitic gneisses, BIF, chert, mafic and ultramafic rocks, and some quartzite and quartz-mica schist of associations 1 and 2. The granites are mostly medium to coarse-grained biotite monzogranites (Pidgeon and Wilde, 1998). Rafts of granitic gneiss are locally preserved in the monzogranite. Rafts of quartz-mica schist and thin-bedded quartzite (association 2) are preserved in the eastern edge of a large mass of monzogranite in the west-central area (Fig. 5a), locally known as "the Blob" (Fig. 14a). However, the mature clastic rocks (association 3), which appear to overlie association 2 there, do not show any evidence of intrusion by the granite, and its contact with the underlying rocks and granite may be unconformable, faulted, or both. The granite is predominantly weakly foliated, but has a strong foliation and is sheared and locally folded near its margins (Figs. 5a and 14b). The shears and strong foliation wrap around the granite, much like a large-scale porphyroclast. Low-grade, heterogeneously foliated, muscovite granite and pegmatite intrude BIF, chert, and quartzite, mafic and ultramafic rocks (association 1), and some quartzite and quartz-mica schist (association 2) in the southwestern and west-central areas of the belt (Figs. 2, 4 and 14c).

4.6. Quartz veins

Quartz veins and silicification are common throughout the Jack Hills greenstone belt. Veins range in size from cm-scale to 50 m or more wide. The large veins have strike lengths up to 2 km, and are commonly associated with brittle faults. Good examples are northwest trending, large quartz veins and brittle faults that offset the southern margin of the belt, with dextral sense, in the central area (Fig. 2). There are clearly at least two generations of quartz veins, because late, cross-cutting veins overprint folded veins in some outcrops. The late-stage veins are either weakly deformed or undeformed. Earlier-formed veins are commonly folded, or sheared. Some veins are also associated with epidote alteration of the host rocks, particularly where faults are present. Much of the fluid activity is likely to have taken place during major shearing across the belt.

4.7. Mesoproterozoic mafic dykes

Dolerite dykes that intrude the Jack Hills greenstone belt and surrounding granitic rocks cross-cut major struc-

Fig. 11. (a) Interbedded metasandstone, metagritstone and pebble metaconglomerate from the mature clastic association in the west-central map area (Fig. 5a). (b) Large quartz pebble in metaconglomerate from the west-central map area (Fig. 5a). This pebble is approximately the maximum size in the association. (c) Clast (upper centre) of black and white quartzite in metasandstone and metaconglomerate in the west-central map area (Fig. 5a). (d) Lenses of black banded quartzite interbedded with metasandstone in the mature clastic association. Small, west-northwest-trending brittle faults offset the layering. Photograph from the west-central map area (Fig. 5a). (e) Sandy and micaceous matrix supporting stretched quartz pebbles in metaconglomerate. The greenish colour is due to fuchsitic micas in the foliation. Photograph taken in the west-central map area (Fig. 5a, MGA 497261mE, 7105495mN). (f) Photomicrograph, in cross-polarized light, of the metaconglomerate showing the quartz, white mica and fuchsite foliation anastomosing around aggregates of quartz (from pebbles or clasts). Sample CS0322, from the west-central map area (Fig. 5a); map coordinates MGA 499135mE, 7105844mN. Note, this is from the same outcrop as W74 (Fig. 5b). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of the article.)

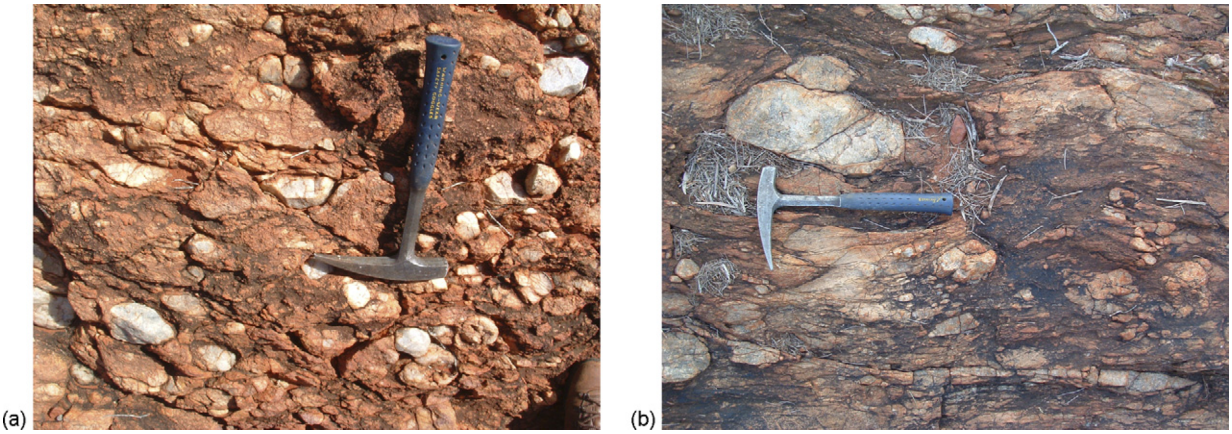

Fig. 13. (a) Possible metaconglomerate in a small shear zone in the northeastern map area (Fig. 6). (b) Tectonically disrupted quartzite and quartz-mica schist from the northeastern map area (Fig. 8; map coordinates MGA 529034mE, 7118818mN).

tures and most lithological units. Major deformation and folding is constrained to have taken place prior to the intrusion of these dykes. The dykes mostly have easterly trends and locally appear to have utilised pre-existing major structures. For example, in the central area, an undeformed, east-trending dolerite dyke trends parallel to the main foliation in sheared gneiss and granite, near the southern margin of the belt. At least two phases of intrusion are recognised: dolerite and gabbro that typically show some chlorite and epidote alteration; and fresh, unaltered dolerite. A leucogabbro dyke in the southwestern part of the belt has a SHRIMP U–Pb zircon age of 1211 ± 3 Ma (Wingate et al., 2005). This dyke, which trends 080°, is at a high angle to, and may cross-cut, this part of the belt, and link up with similarly trending gabbroic and dolerite dykes within the belt (Fig. 4, MGA 490876mE, 7102231mN). A similar dyke intruded into granite 30 km south of the belt has a SHRIMP U–Pb age of c.1213 Ma (M.T.D. Wingate, pers. comm., 2005). These dykes have been included in the 1210 Ma Marnda Moorn large igneous province by Wingate et al. (2005).

Other east-trending dolerite outcrops to the east and south of the Jack Hills greenstone belt have been dated at 1075 ± 10 Ma using combined SHRIMP U–Pb and paleomagnetic techniques, and are related to the Warakurna large igneous province (Wingate et al., 2004). A west-northwest trending dyke that cuts across fault and fold structures in associations 3 and 4 in the Eranondoo Hill area (Fig. 5b) is correlated with the 1075 Ma dyke suite (M.T.D. Wingate, pers. comm., 2004). Discontinuous outcrops of east-northeast trending dolerite dykes also occur in gneiss and granite just southeast of Eranondoo Hill, and are probably part of the same suite of dykes.

4.8. Summary of the structural history of the four associations

Associations 1 and 2 contain evidence of at least three phases of deformation. The first phase is defined by the presence of an early foliation, and recumbent folds that are evident in the BIF (Spaggiari, 2007a). This foliation, and the recumbent folds, are folded by a second generation of upright to inclined, variably plunging folds. These folds are also evident in associations 3 and 4, but are first generation folds; i.e. they do not fold an earlier foliation. This indicates that associations 3 and 4 are younger. The second-generation folds are locally cut by numerous shears, many of which are part of the Cargarah Shear Zone that cuts through the belt (Figs. 2 and 5a). These second-generation folds may have formed in the early stages of shearing, or may pre-date it. Neoarchean granites, such as the Blob (Fig. 5a) contain evidence of shearing, which clearly post-dates their intrusion, but do not appear to have been affected by the first phase of deformation and metamorphism evident in associations 1 and 2. Both the first and second generations of folds, and the shears, are overprinted by semi-brittle faults and kink folds (Spaggiari, 2007a).

5. Geology of the west-central area

The west-central area (Figs. 2 and 5a) contains the majority of outcrop of the mature clastic rocks (association 3), including the W74 sample site (Fig. 5b and c). The association is folded, and fault-bounded by lenses of Proterozoic rocks, BIF, chert, and quartzite to the south, and strongly deformed, steeply-dipping, mafic, ultramafic, pelitic and semi-pelitic schists to the north

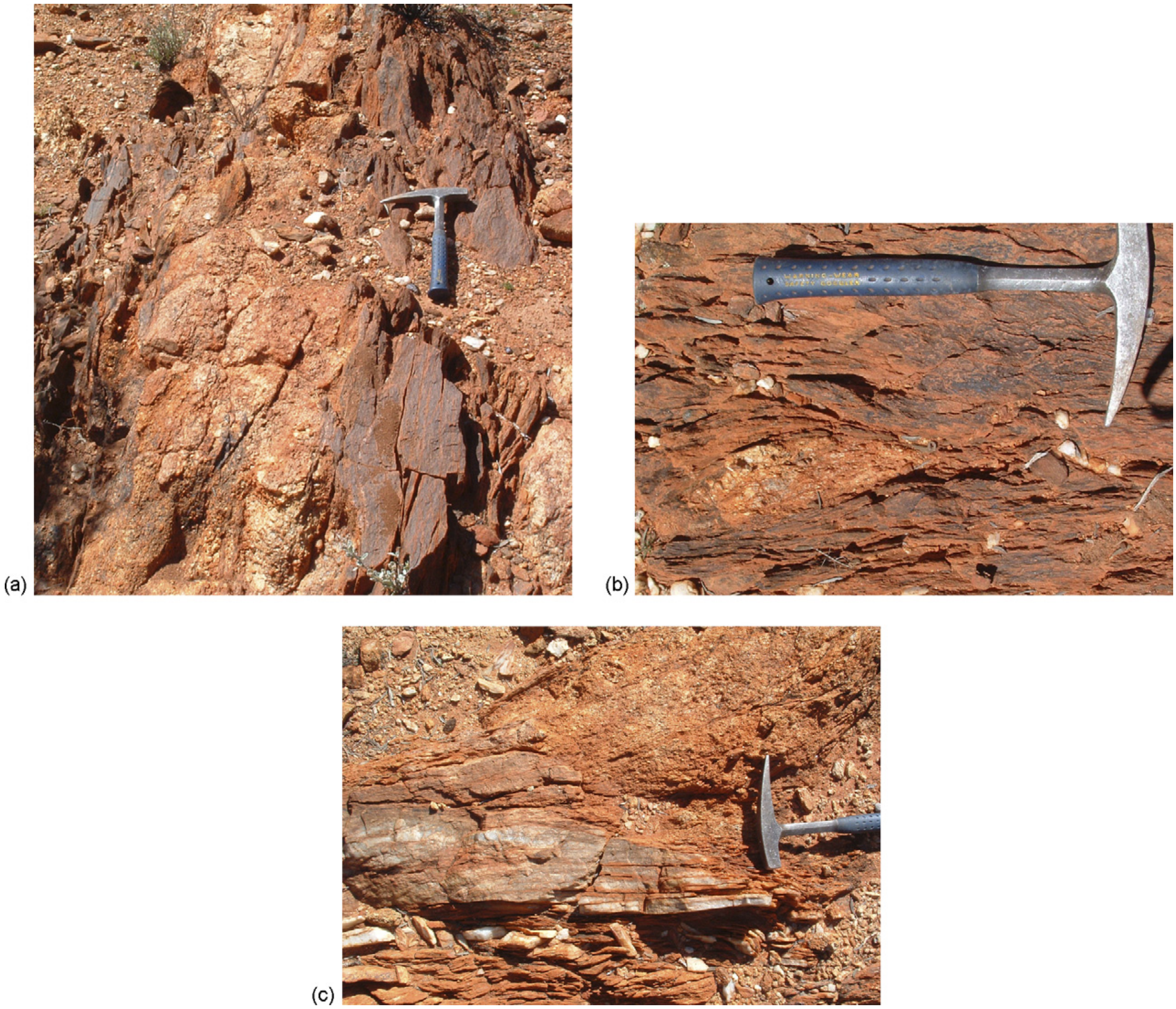

Fig. 14. (a) Quartz-mica schist intruded by granite near the eastern margin of the main granite body known as the Blob, from the west-central map area (Fig. 5a). (b) Granite lens within quartz-mica schist near the margin of the main granite body known as the Blob, from the west-central map area (Fig. 5a). Both the granite and quartz-mica schist are sheared, therefore shearing took place after the granite intrusion. (c) Muscovite granite (upper right) intruding quartzite and quartz-mica schist, in the southwestern map area (Fig. 3).

(Fig. 5a). The Cargarah Shear Zone (Spaggiari, 2007b) cuts through gneiss and granitic rocks just north of the belt in the west-central area, and is locally marked by mylonites, small-scale isoclinal folding, local high strain zones with predominantly dextral S-C foliations, and large quartz veins. In the west-central area, the southern margin of the belt is a steeply dipping, sheared contact between BIF, and granitic rocks and gneiss. Mylonite zones occur locally within granitic rocks and gneiss parallel to the belt. The BIF is internally folded into tight to isoclinal folds that are mostly second-generation folds. These are locally refolded into tight folds, parallel to the trend of the belt (e.g. just north of MGA 496838mE, 7106498mN, and MGA 496474mE, 7105003mN). Small, late-stage brittle faults cut all rocks in the west-central area, including the Proterozoic rocks.

The mature clastic association is folded into tight, south- to southeast-inclined folds that are cut by numerous shears and brittle faults (Fig. 5a and b). The W74 outcrop is interpreted to occur on the southern limb of a syncline, where it is underlain by beds of predominantly metasandstone and quartzite (Fig. 5b). The folds plunge moderately to the southwest, and because this outcrop occurs on a northeast-facing hill, the bed that it occurs in has limited exposed extent. However, similar pebble metaconglomerate beds are numerous and crop out over extensive areas. The northern part of the mature clastic association is steeply-dipping, more sheared than the southern part, and fines towards the north near the contact with quartz-mica schist (association 2, Fig. 5b).

The southern contact of the mature clastic association is complex and is folded and faulted in with BIF, chert, and quartzite. Two, fault-bounded lenses of Proterozoic quartz-mica schist and quartzite are mostly mylonitic, and contain slivers of ultramafic schist (MGA 497464mE, 7105183mN). Near the W74 outcrop (Fig. 5b), the faulted contact between the mature clastic association and the Proterozoic sequence shows evidence of late-stage, brittle or semi-brittle movement, such as a sharp, irregular contact, small duplexes that suggest sinistral displacement, and quartz veins. This fault may be either a reactivated structure, or a late structure formed after the main phase of shearing.

6. Discussion

6.1. Relationships between lithological associations, and timing of deposition

Lithological similarities, deformation, metamorphism, and silicification have made identification of complete successions from individual rock units problematic. Extensive shearing and brittle faulting has also juxtaposed similar rock types of different ages. It is evident that at least three, but possibly four, metasedimentary successions are present in the Jack Hills greenstone belt. These have been divided into the following associations: (1) an older succession of BIF, chert, quartzite, quartz-mica schist, and associated mafic and ultramafic rocks; (2) an association of pelitic and semi-pelitic rocks, quartzite and mafic schists; (3) a mature clastic association that hosts the majority of $\geq$4.0 Ga detrital zircons; (4) a Proterozoic association of quartz-mica schist, quartzite, and metasandstone with local metaconglomerate. Association 1, and at least part of association 2, were deposited prior to Late Archean granite intrusion, and may be part of the same succession (Fig. 3).

The mafic–ultramafic rocks and BIF (greenstone component) of the Jack Hills greenstone belt may be similar in age to some of the other greenstones in the Murchison Domain of the Youanmi Terrane, such as the c. 2.95 Ga Yalgoo-Singleton greenstone belt (dated by U–Pb in zircons from felsic volcanics; Pidgeon and Wilde, 1990; Wang et al., 1998), and the nearby Weld Range greenstone belt which contains lithic sandstone interbedded with BIF that was deposited close to c. 2970 Ma (Fig. 3; sample 184112; Bodorkos and Wingate, 2007; Spaggiari, 2006).

The depositional age of the mature clastic rocks (association 3) is unknown. Associations 1 and 2 contain an earlier foliation and folding phase that is not present in association 3, which suggests it is younger. The youngest detrital zircons in most samples of association 3, including the W74 outcrop, are c. 3.0 Ga, which indicates the association was deposited after this time, possibly between c. 3.0 and 2.7 Ga if it was prior to the intrusion of Neoarchean granites (Figs. 3 and 6). Alternatively, the association could be Paleoproterozoic, but not necessarily deposited as late as the Proterozoic rocks of association 4. The absence of Neoarchean detrital zircons from most samples dated from association 3 suggests deposition prior to Late Archean magmatism. However, it is possible that either the Neoarchean granites were not fully exposed and eroding at the time of deposition, or that the rocks sampled show some bias in their detrital zircon populations. The first rocks in the region to record a component of Neoarchean zircons are the c. 2.17 Ga sedimentary rocks of the Yerrida Basin (Figs. 1 and 3; Pirajno et al., 2004), indicating exposure of the Neoarchean granites, at least locally, at that time. The Camel Hills metamorphic rocks, preserved in the Errabiddy Shear Zone, were deposited by c. 2.0 Ga and also contain a component of Neoarchean detritus (Figs. 1 and 3; Occhipinti et al., 2004). This suggests that the Neoarchean granitic rocks were uplifted by at least 2.2–2.0 Ga. It is conceivable that the mature clastic rocks of the Jack Hills greenstone belt were deposited at about this time, synchronous with the lower parts of the Yerrida Basin, or possibly during tectonism associated with the Glenburgh Orogeny. Alternatively, they may have accumulated after intrusion, but prior to exposure of, the Neoarchean granites, potentially as passive margin sediments. It is unlikely that they were deposited after the Glenburgh Orogeny, as they do not appear to contain detrital zircons of that age, unlike the Proterozoic rocks of association 4 (Fig. 6).

The relationship of the lithological units preserved in the Jack Hills greenstone belt to the Mount Narryer metasedimentary rocks is unknown, but there are some distinct differences. The Mount Narryer succession lacks the "greenstone" component of the Jack Hills greenstone belt, i.e. there are no large tracts of mafic–ultramafic rocks and BIF, but BIF does occur as lenses within the gneisses flanking the metasedimentary succession. The Mount Narryer metasedimentary succession itself is also different to that at Jack Hills. It is more aluminous (more pelitic), contains calc-silicate gneiss, and some of the metaconglomerate contains a variety of clasts such as biotite-garnet schist, unlike the quartz-pebble metaconglomerate at Jack Hills. Although there is a link in that

≥4.0 Ga detrital zircons also occur in the Mount Narryer quartzites, this does not mean they are part of the same succession, as detritus of this age also occurs in the Southern Cross Domain of the Youanmi Terrane (Wyche et al., 2004).

6.1.1. Proterozoic metasedimentary rocks

The youngest recorded zircon in the Proterozoic rocks (association 4) is 1576 ± 22 Ma (Cavosie et al., 2004). However, only a single zircon of this age has been recorded and it does not match any known rocks in the region. Other Proterozoic zircons in the association indicate a maximum depositional age of 1791 ± 21 Ma (Fig. 6; Dunn et al., 2005). All of the Proterozoic metasedimentary rocks are deformed and must have been deposited prior to the cessation of any major deformation, interpreted to have been prior to intrusion of a second suite of dolerite dykes at 1075 Ma (Spaggiari, 2007b; Spaggiari, 2007a). The Proterozoic rocks of association 4, particularly in the west-central area (Fig. 5a), appear to have been affected by major shearing related to the Capricorn Orogeny (1830–1780 Ma, Occhipinti et al., 2003). This, and their maximum depositional age of 1791 ± 21 Ma, suggests deposition may have occurred during the late stages of the Capricorn Orogeny (Fig. 3).

One explanation of how this may have occurred is that the rocks of association 4 were deposited in small, transpressional, pull-apart basins associated with the Cargarah Shear Zone. The major compressive stress direction is inferred to have been northwest-southeast and oblique to the belt, which could have allowed approximately northeast-southwest oriented pull-aparts to have formed (Fig. 15). Note that the strain rate on the fault system is likely to have been variable over time, and the jogs and associated extensional faults would be overprinted by ongoing shearing. Transpressional deformation may also have produced folding and uplift within the shear zone, causing sedimentation during fault movement. Structures associated with formation of the Cargarah Shear Zone are overprinted by semi-brittle faults and folds, and many of the shears are reactivated or relatively late; e.g. the fault separating the Proterozoic rocks and the mature clastic rocks of association 3 south of the W74 outcrop (Fig. 5b and c; Spaggiari, 2007a). ^{40}Ar–^{39}Ar dating of white mica from this fault indicates it was active at c. 1200 Ma, close to the time the first suite of mafic dykes was emplaced (sample JAW-03-03, Figs. 5b and 7b; Spaggiari et al., submitted). Therefore, it is also conceivable that further deposition may have occurred at a later stage during reactivation.

The Proterozoic rocks (association 4) in the Jack Hills greenstone belt appear to have limited extent, but strong lithological similarities between the semi-pelitic rocks and quartzites in association 2 makes the spatial extent of both associations 2 and 4 difficult to determine. Additional detrital zircon dating is required to identify whether a greater proportion of the metasedimentary rocks are part of association 4. It is also not clear if the Proterozoic rocks are a local occurrence confined to the belt, or whether they relate to other Proterozoic rocks in

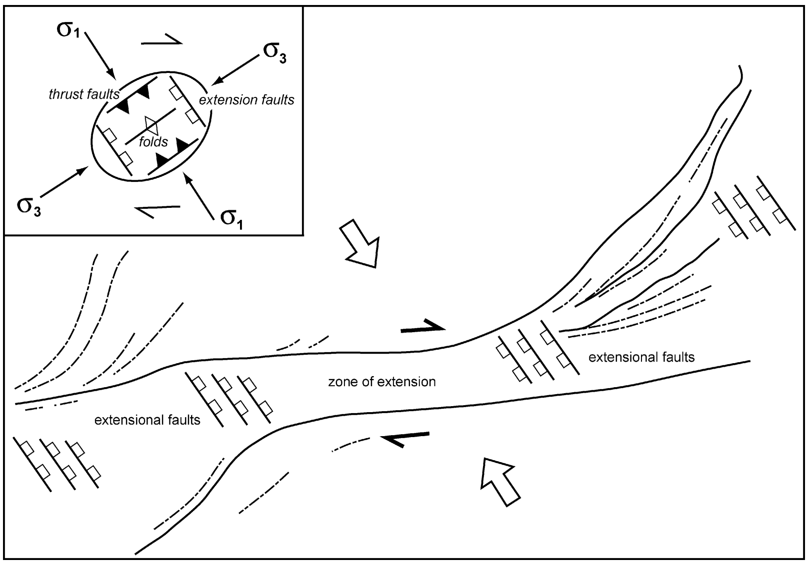

Fig. 15. Schematic diagram showing formation of small pull-apart basins or fault jogs in the Jack Hills greenstone belt during dextral strike-slip deformation, in an inferred right-stepping (upper fault to lower fault) wrench system (modified from McClay, 1987). Inset shows the inferred regional strain ellipse associated with the system (modified from McClay, 1987).

the region, for example, the Paleoproterozoic basins that cover much of the northern margin of the Yilgarn Craton (Figs. 1 and 3). Outliers of Proterozoic rocks of unknown association also occur throughout the northern part of the Murchison Domain of the Youanmi Terrane (Belele and Glengarry 1:250,000 scale geological map series, Geological Survey of Western Australia) in close proximity to the Jack Hills greenstone belt. The lower part of the Yerrida Basin, and the Bryah and Padbury Basins were deposited prior to the Jack Hills Proterozoic rocks (association 4). The only constraint on the age of the upper Yerrida Basin is a monazite age of 1843 ± 14 Ma from shale at the base of the Maraloou Formation, the uppermost unit of the basin (Rasmussen and Fletcher, 2002; Pirajno et al., 2004). This suggests the upper Yerrida Basin may also have been deposited prior to association 4 in the Jack Hills greenstone belt.

The Earaheedy Basin was deposited after 1.8 Ga (Halilovic et al., 2004), but its original extent is unknown (Figs. 1 and 3). It was deposited prior to formation of the Stanley Fold Belt, now interpreted as having formed during the Mangaroon Orogeny between c. 1680 and 1620 Ma (Sheppard, 2005; Sheppard et al., 2005). The detritus is interpreted to have come largely from the west and southwest, based on detrital zircon age patterns that include sources from both the Narryer Terrane and Gascoyne Complex, and limited paleocurrent data (Halilovic et al., 2004). If the Proterozoic rocks in the Jack Hills greenstone belt are the same age as some of those in the Earaheedy Basin, they may have had a similar siliciclastic provenance, but been closer to the source (*cf.* Dunn et al., 2005).

6.2. Implications for tectonic history

The recognition of various lithological associations within the Jack Hills greenstone belt, although not clearly defined in terms of stratigraphy, allows speculation to be made on the tectonic history. Rocks that are part of association 1 are interpreted to have largely formed synchronously, most likely with some of those from association 2. This includes a succession of mafic and ultramafic rocks, some of which were likely to have been mafic (basaltic) volcanics and peridotites. The close association of these with BIF, chert, and some pelitic and semi-pelitic rocks suggests eruption and intrusion into a basin of unknown extent. The nearby Mingah Range and Weld Range greenstone belts (Fig. 1) may also have been part of the same basin, or part of a series of small basins, perhaps related to a rift or marginal basin either adjacent to, or deposited on, the older gneissic rocks that formed the basement (Fig. 3). Whether the basin included a near-shore, fluvial sedimentary environment where the mature clastic rocks of association 3 were deposited is unclear. Alternatively, the mature clastic rocks may have been deposited after late Archean magmatism, either as a passive margin sequence, or in the hinterland of the accreted Glenburgh Terrane during the Glenburgh Orogeny. In contrast to the above, those rocks that are clearly Proterozoic in age were deposited in an intracratonic setting, during and/or following the Capricorn Orogeny.

7. Conclusions

The Jack Hills greenstone belt contains four rock associations. Association 1 was deposited prior to intrusion of Neoarchean granites and includes BIF, chert, quartzite, quartz-mica schist, and mafic and ultramafic rocks. Association 2 contains pelitic and semi-pelitic rocks, quartzite and mafic schists. At least part of this association was intruded by Neoarchean granites, and, along with association 1, may be part of one succession. Mature clastic rocks of association 3 that host $\geq$4.0 Ga detrital zircons may be part of association 1 and/or 2, but their depositional age is unconstrained, as they do not show intrusive relationships with the Neoarchean granites. They also lack first generation folds and a foliation related to an early deformational history that is present in rocks of associations 1 and 2 (Spaggiari, 2007a). Proterozoic rocks of association 4 consist of quartz-mica schist, quartzite, and metasandstone with local metaconglomerate, and were deposited after 1791 ± 21 Ma (Dunn et al., 2005). It is conceivable that they were deposited in small pull-apart basins formed during transpressional shearing related to the later stages of the Capricorn Orogeny. Although outliers of Proterozoic metasedimentary rocks and several Paleoproterozoic basins are in close proximity to the Jack Hills greenstone belt, it is not clear whether the Jack Hills Proterozoic metasedimentary rocks are a local occurrence, or part of a more extensive succession. Deformation, extensive dynamic and static recrystallisation, and lithological similarities make division of the four associations into successions problematic, and preclude determination of a complete stratigraphy.

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