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NORTHWEST PILBARA CRATON: A RECORD OF 450 MILLION YEARS IN THE GROWTH OF ARCHEAN CONTINENTAL CRUST

by AH Hickman

GEOLOGICAL SURVEY WESTERN AUSTRALIA

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Cover photograph: Pillow structures in the c. 3118 Bradley Basalt, Whundo Group, 2.6 km northwest of Harding Dam. The pillows were formed by submarine eruption of basalt lava within the world's oldest well-preserved island arc succession.

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Northwest Pilbara Craton: a record of 450 million years in the growth of Archean continental crust

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Abstract

The northwest Pilbara Craton provides an exceptionally well-preserved record of Archean crustal evolution from c. 3280 to 2830 Ma. A major change in tectonic processes took place at c. 3220 Ma when mantle plume-related volcanism and vertical deformation was abruptly replaced by plate tectonic evolution. This change, commencing with deep rifting of the 3530–3223 Ma volcanic plateau of the East Pilbara Terrane, caused northwest separation of the Karratha Terrane and established a northeast–southwest-trending rift basin floored by juvenile basaltic crust (Central Pilbara Basin). Much thinner, less rigid, and significantly hotter than the adjacent continental plates of the Karratha and East Pilbara Terranes, this basin became the focus of almost all tectonic and magmatic activity in the Pilbara Craton until c. 3068 Ma. A c. 3160 Ma metamorphic event in the Karratha Terrane coincided with the northwest separation of the Karratha Terrane being halted and a change to convergence back towards the East Pilbara Terrane. It is inferred that this reversal was due to c. 3160 Ma collision of the Karratha Terrane with another plate northwest of the Pilbara Craton. Convergence resulted in obduction of juvenile basaltic crust of the Central Pilbara Basin (Regal Formation) onto the Karratha Terrane to form the Regal Terrane. It also resulted in the development of a subduction zone within the Central Pilbara Basin producing the 3130–3093 Ma arc volcanism and granitic intrusion of the Sholl Terrane. At c. 3068 Ma the convergence culminated in accretion of three northwest Pilbara terranes (Karratha, Regal, and Sholl) to form the West Pilbara Superterrane and collision of this superterrane with the East Pilbara Terrane, resulting in the 3068–3066 Ma Prinsep Orogeny.

Crustal relaxation and extension following the Prinsep Orogeny initiated development of a shallow-water intracontinental basin (3066–3015 Ma Gorge Creek Basin) that covered the entire northern half of the Pilbara Craton. By c. 3023 Ma this extensional phase had been replaced by renewed northwest–southeast convergence between the Pilbara Craton and the inferred plate to the northwest. During convergence a continental volcanic arc (3009–2991 Ma Whim Creek Basin) was developed immediately southeast of a 3006–2982 Ma belt of granitic intrusions (Maitland River Supersuite). Southeast of the volcanic arc, a back-arc basin (Mallina Basin) evolved from c. 3015 Ma until its development was terminated by 2954–2919 Ma plate collision (North Pilbara Orogeny). Geochronological evidence of the second phase of Mesoarchean convergence and subduction in the northwest Pilbara is provided by a southeasterly migration of granitic intrusion from c. 3023 to 2919 Ma. The final event in the crustal evolution of the northwest Pilbara Craton was the intrusion of 2851–2831 Ma post-orogenic granites (Split Rock Supersuite).

Most Archean mineralization in the northwest Pilbara was related to plate tectonic processes and included volcanic-hosted massive sulfide (VHMS), Cu–Zn mineralization within arc volcanism, Ni–Cu, PGE, and V–Ti in layered ultramafic–mafic intrusions, and lode Au in and adjacent to major strike-slip shear zones.

KEYWORDS: Archean, convergence, mineralization, obduction, orogeny, plate tectonics, subduction

Introduction

Pilbara Craton

The Pilbara Craton is a 400 000 km^2 remnant of Eoarchean to Mesoarchean crust that underlies the northwestern part of Western Australia (Figs 1 and 2). Approximately 80% of the craton is concealed by Neoarchean and Proterozoic volcano-sedimentary successions and Phanerozoic basins but a ~60 000 km^2 exposure of the craton is present in the north (Fig. 2). This section of the craton, referred to as the northern Pilbara Craton, has been the subject of extensive geological mapping by the Geological Survey of Western Australia (GSWA) and other organizations and researchers over the past 50 years. Figure 3 shows the distribution of tectonic units in this northern section of the craton.

Crust underlying the Pilbara Craton ranges in age from c. 3800 to 2830 Ma (Hickman et al., 2010). Evidence that the presently exposed craton was constructed on older crust is provided by xenocrystic and detrital zircon, Nd model ages, and Hf isotopic data (Van Kranendonk et al., 2007a,b; Hickman, 2012; Kemp et al., 2015a,b). Some of this evidence comes from Paleoarchean rocks and because the craton evolved through the Mesoarchean, including major granitic intrusion, the amount of surviving older crust must have greatly diminished. The oldest rock identified in outcrop is c. 3650 Ma banded tonalite gneiss (Warrawagine gneiss, informal name), which is preserved in outcrop-scale xenoliths within Paleoarchean granodiorite (GSWA 142870, Nelson, 1999f). However, similar but undated gneisses elsewhere in the craton could be equally old and only geochronology can establish this. The oldest extensively exposed rocks are

3530–3490 Ma volcanics of the Coonterunah Subgroup of the Warrawoona Group and 3484–3462 Ma granites of the Callina Supersuite (Van Kranendonk et al., 2004, 2006). Exposures of these c. 3500 Ma rocks are confined to the East Pilbara Terrane (EPT) of the east Pilbara. The preserved record of crustal evolution in the northwest part of the craton spanned 450 million years, from 3280 to 2830 Ma, although, like in the east Pilbara isotopic evidence indicates that older crust, at least 3500 Ma old, once also existed here.

Craton boundaries

Geophysical evidence reveals that the Pilbara Craton is considerably larger than suggested by present-day exposures, as indicated by the interpreted concealed craton boundaries shown on Figures 1 and 2. A seismic survey between the southern Pilbara Craton and Gascoyne Province suggests that the concealed southern section of the Pilbara Craton extends up to 200 km beyond the inliers shown on Figure 2 (Johnson et al., 2011, 2013; Korsch et al., 2011; Thorne et al., 2011). Thus, the southern Pilbara Craton underlies the Proterozoic Ashburton, Edmund, and Collier Basins. Farther east, to the southeast from the Sylvania Inlier (Fig. 2), gravity data suggest that the Pilbara Craton continues at depth beneath the Proterozoic Collier, Salvation, and Officer Basins. Still farther east, Kirkland et al. (2013) used isotopic data to argue that the sedimentary successions of the Rudall Province were deposited on the east margin of the Pilbara Craton, implying that the craton underlies at least the western section of the Paterson Orogen. In the northeast Pilbara, Reading et al. (2012) interpreted passive seismic data in terms of a deepening of the Moho. In combination with a change in wavespeed contrast from the Pilbara Craton to the Paterson Orogen it was interpreted that the Pilbara Craton might terminate relatively close to its exposed northeastern margin. Alternatively, the Pilbara Craton might underlie the Paterson Orogen in this area, as interpreted in the Rudall Province. Along the Pilbara coast, the northeast margin of the craton underlies a zone of Proterozoic thrusting on the west margin of the Paterson Orogen. North and northwest from the Pilbara coast, gravity data and numerous seismic lines undertaken during petroleum exploration indicate that the Pilbara Craton extends between 100 and 150 km offshore beneath Phanerozoic sedimentary successions of the Northern Carnarvon Basin (Geoscience Australia, 2015). Within 30 to 100 km of the Pilbara coast, the successions of the Lambert Shelf and Peedamullah Shelf are less than 1 km thick, whereas farther offshore the Pilbara Craton is covered by thicker Phanerozoic successions in sub-basins of the Northern Carnarvon Basin (Fig. 2).

Fragment of a much larger Archean continent

Many geological features of the northern Pilbara Craton and of the overlying Fortescue and Hamersley Basins (Fig. 2) provide evidence that prior to Paleoproterozoic orogenic events along its eastern, southern, and western margins the craton formed part of a far more extensive section of Archean crust (Krapez and Eisenlohr, 1998; Thorne and Trendall, 2001). From 2775 Ma onwards this crust became fractured by late Neoarchean and early Paleoproterozoic crustal extension and rifting, and continental breakup eventually took place in the south Pilbara after 2590 Ma (Martin et al., 1998).

Indications of the minimum Archean extent of the Pilbara Craton plate can be obtained from the lateral thickness variations in the overlying Neoarchean and Paleoproterozoic successions. The 2775–2629 Ma Fortescue Group reveals no evidence of stratigraphic thinning towards the present margins of the craton (Blake, 1993; Thorne and Trendall, 2001). Likewise, one of the best exposed and most intensely studied formations of the 2629–2445 Ma Hamersley Group, the Brockman Iron Formation, thins only very gradually from west to east (Morris and Horwitz, 1983). Over a distance of 600 km this formation changes from a thickness of c. 645 m on the Yarraloola 1:250 000 map sheet area in the northwest (Williams, 1968) to a thickness of 450 m on the Robertson 1:250 000 map sheet area in the east (Williams and Tyler, 1991). Since in all areas of the Pilbara, both the Fortescue and the Hamersley Group, are known to overlie the Pilbara Craton these observations require that between 2775 and 2445 Ma the Pilbara Craton extended over a very much larger area than it does today. Martin et al. (1998) concluded that the Pilbara and Kaapvaal Cratons experienced breakup at the same time (c. 2590 Ma), and that they were aligned along a common ocean margin.

The eastern margin of the Pilbara Craton is a Paleoproterozoic belt of intense folding, thrusting, and metamorphism formed by the 1795–1760 Ma Yapungku Orogeny. The western margin is a relatively wide belt of deformation and metamorphism produced during the 1820–1770 Ma Capricorn Orogeny. The southern margin of the craton was deformed by the 2215–2145 Ma Ophthalmian Orogeny, which was a result of north–south collision with the Glenburgh Terrane (Johnson et al., 2010, 2011, 2013). There is no evidence of a Proterozoic orogenic margin on the northern side of the craton, implying either that this margin is simply a rifted passive margin or that any northern orogenic margin is far more distant.

In the extreme southern part of the Pilbara Craton there is evidence of initial rifting of the Archean crust at c. 2775 Ma (Blake, 1984, 1993; Blake and Groves, 1987; Thorne and Trendall, 2001). However, regional continuity of the Fortescue and Hamersley successions establishes that breakup along the southern margin is unlikely to have taken place until after 2445 Ma. Along the eastern margin of the craton the basal 2765–2740 Ma formations of the Fortescue Group contain a series of margin-parallel faults that may have originated as syndepositional extensional structures later reactivated as Proterozoic thrusts (Trendall, 1991; Thorne and Trendall, 2001). Evidence of syndepositional rifting is also recorded in the basal Fortescue Group of the northwest Pilbara (Blake, 1993; Hickman and Strong, 2003), where some of the early normal faults, in addition to earlier faults such as the Sholl Shear Zone (SSZ), were later reactivated as thrusts during the Capricorn Orogeny (Hickman and Strong, 1999).

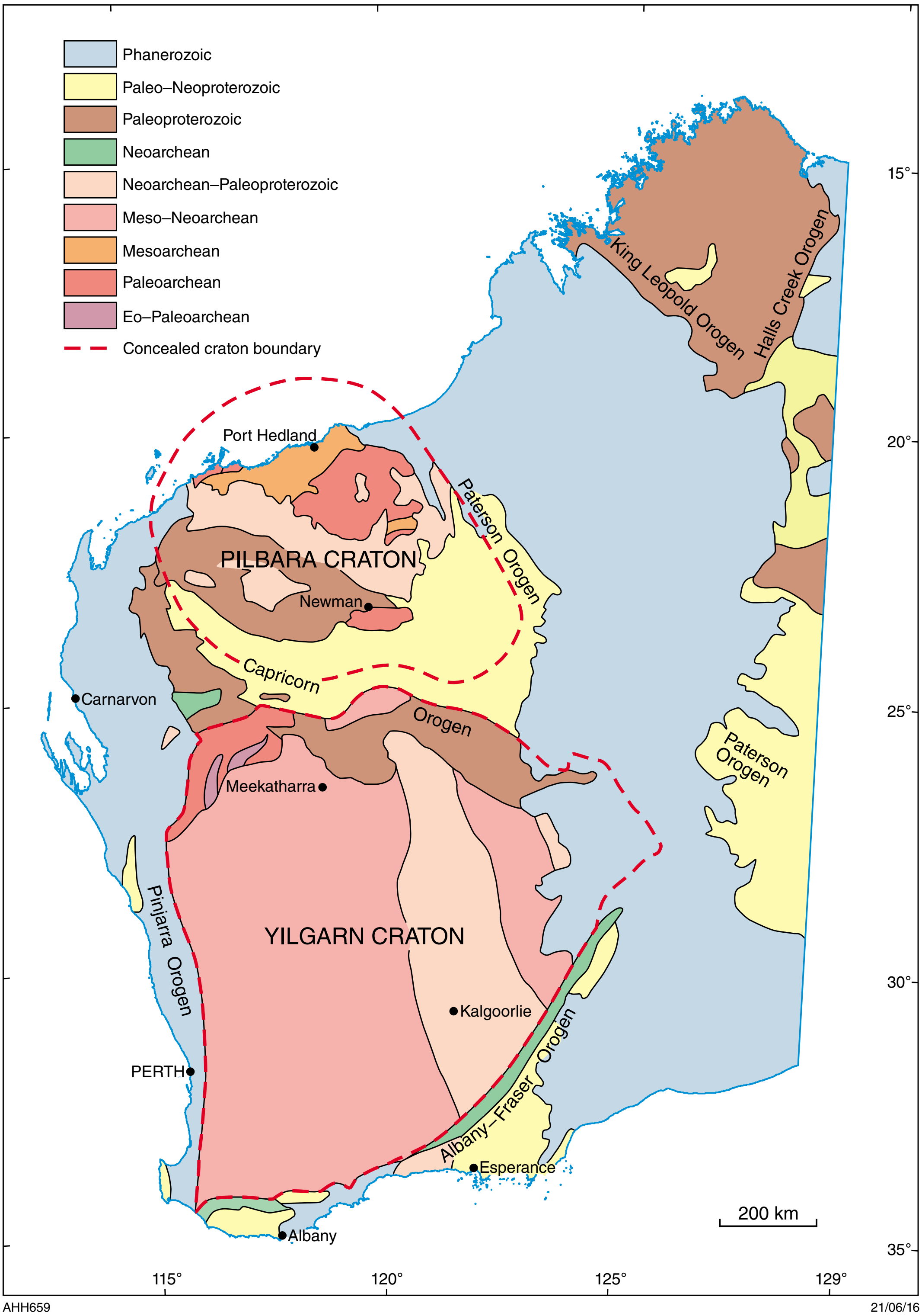


Figure 1. **Simplified Archean and Proterozoic chronological divisions of Western Australia, showing an interpretation of the concealed extents of the Pilbara and Yilgarn Cratons**

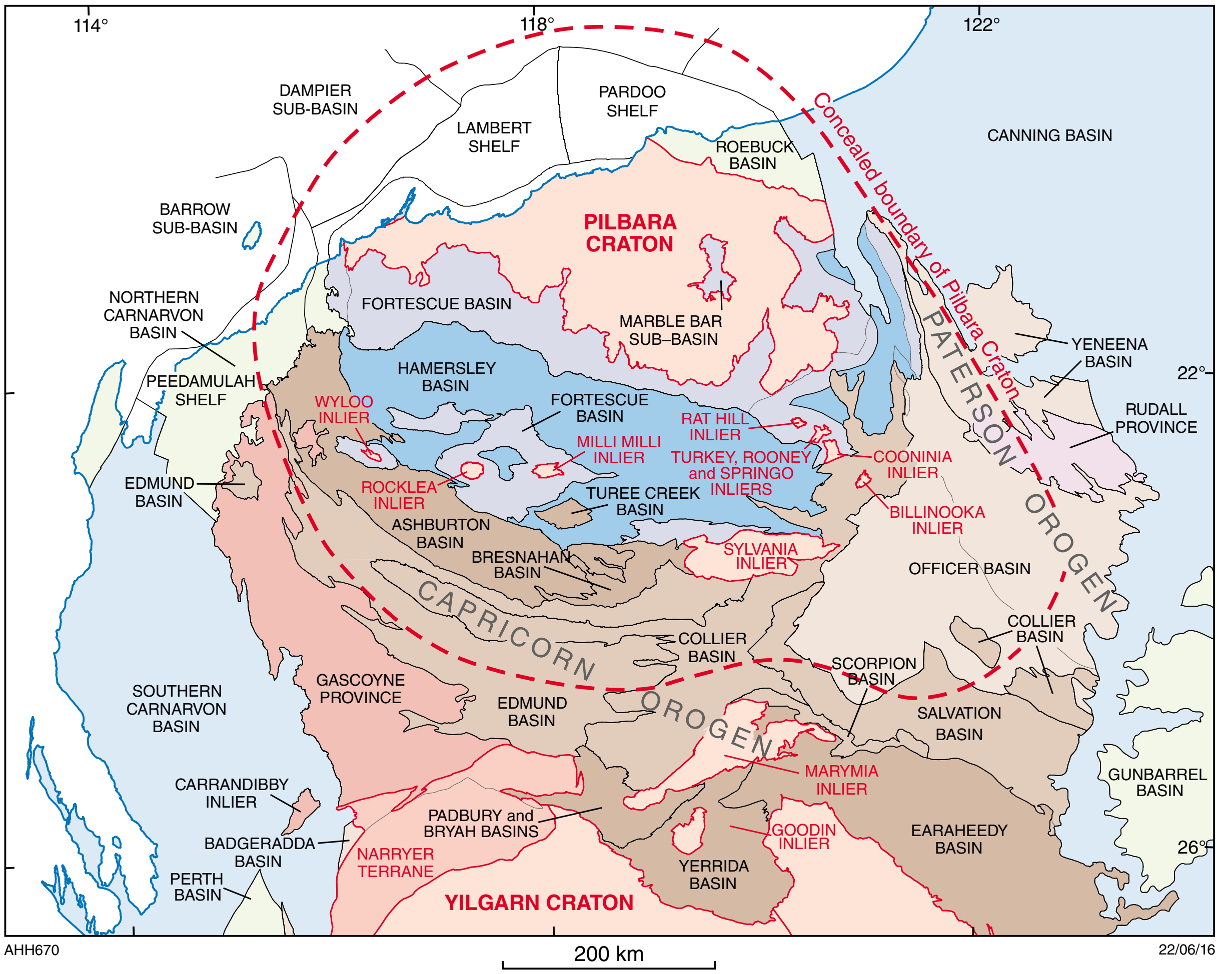


Figure 2. **Tectonic units of northwestern Western Australia, showing the setting of the Pilbara Craton. The southern half of the craton is concealed by Neoarchean and Proterozoic rocks except for rare exposures within inliers.**

Recent workers have attributed the rifting and continental breakup of the Archean Pilbara crust to uplift and extension above one or more Neoarchean mantle plumes (Arndt et al., 2001; Condie, 2001, 2004; Ernst et al., 2004; Pirajno, 2007; Pease et al., 2008; Hickman, 2012). One factor contributing to the breakup may have been that between orogenic events in the northwest and southeast Pilbara at c. 2900 Ma and first deposition of the Fortescue Group at c. 2750 Ma the Pilbara Craton experienced post-orogenic crustal relaxation accompanied by deep erosion. The relationship between the angular erosional unconformity at the base of the Fortescue Group and the underlying geology of the Pilbara Craton indicates deep erosion during this period. For example, by c. 2775 Ma the c. 2900 Ma orogenic fold belts along the Mosquito Creek and Mallina Basins (Fig. 3) had been almost entirely reduced to peneplains. Likewise, domes of the EPT, which had been increased greatly in amplitude during the North Pilbara Orogeny at c. 2950 Ma (Hickman and Van Kranendonk, 2004) had been 'unroofed' by deep erosion to produce a c. 2775 Ma landscape of low granitic hills, separated by valleys underlain by greenstones (Hickman, 1975a, 1983). The present estimate is that at least 5 to 10 km of Pilbara crust was eroded between 2900 and 2775 Ma. The combined thickness of the Pilbara Craton and the overlying Fortescue and Hamersley Groups is approximately 30 km (Reading et al., 2012; Yuan, 2015), which is approximately 10 km thinner than the average crustal thickness of the Yilgarn Craton (>40 km).

The conclusion that the Pilbara Craton was once part of a much larger Archean continental plate suggests that other sections of that plate might be preserved on other continents. Buoyant continental crust as thick as that of most Archean cratons is difficult to destroy by plate tectonic processes, such as subduction and it is therefore very likely that the other fragments of the original 3800–2445 Ma crust still exist. Several workers have proposed that one such fragment, much larger than preserved in the Pilbara, is the Kaapvaal Craton of southern Africa. A number of early stratigraphic comparisons between the Pilbara and Kaapvaal Cratons (Trendall, 1968; Button, 1976; Blake, 1984; Hickman and Harrison, 1986) laid the foundations for this idea. More recent stratigraphic, geochronological, and paleomagnetic studies (Nelson et al., 1992, 1999; Zegers et al, 1998b; Eriksson et al., 2002; Zhao et al., 2002, 2004; Pickard, 2003; de Kock et al., 2009, 2012) and evidence from asteroid impact spherule beds (Lowe and Byerly, 1986; Byerly et al., 2002; Rasmussen et al., 2005; Glikson and Vickers, 2006; Jones-Zimberlin et al., 2006; Simonson et al., 2009; Hassler et al., 2011; Glass and Simonson, 2012, 2013) provide strong grounds for stratigraphic correlation between the two cratons. Based on the early stratigraphic evidence, Cheney et al. (1988) and Cheney (1996) interpreted the two cratons to be fragments of an Archean continental plate, and named it 'Vaalbara'. Recently improved geological data from Archean cratons, such as those in Zimbabwe, Antongil (Madagascar), Tanzania, and parts of southern India, suggest that other fragments of Vaalbara may also be preserved in these areas.

Recognition that the present Pilbara Craton is merely a fragment of a much larger Archean continental plate improves our understanding of the likely scale of the Archean tectonic processes and settings recorded in its geology. This point is illustrated by the work of Zegers et al. (1998b) who interpreted a match between the strike-slip fault system seen in the northwest and central Pilbara Craton to a similar system in the Kaapvaal Craton.

This Report

Previous GSWA documentation of the geology of the northwest Pilbara Craton has been provided in Reports to accompany geological maps, field guides compiled for geological excursions, and in a number of journal papers. All these publications have focused on particular areas within the northwest Pilbara Craton or on specific fields of study. The purpose of this Report is to provide a more comprehensive geological account of the area, with reviews of previous work and interpretations where appropriate. Mineralization is reviewed and discussed within the context of the current interpretation of the region's crustal evolution.

Data acquisition

The largest single source of geological data used in the current interpretation of the northwest Pilbara has been the Pilbara Mapping Project. This project, which extended across the northern half of the Pilbara Craton and ran from 1994 to 2002, was undertaken by the Geological Survey of Western Australia (GSWA) and Geoscience Australia (GA) under the North Pilbara National Geoscience Mapping Accord (NPNGMA). At its peak, between 1997 and 2002, up to 12 geoscientists were employed annually on the project. Field mapping was preceded by regional aerial surveys to obtain extensive aeromagnetic and radiometric imagery. Results included the release of 1:100 000 and 1:250 000 Geological Series Maps and accompanying Reports. A particularly important feature of the project was the acquisition of U–Pb zircon geochronology obtained by means of sensitive high-resolution ion microprobe (SHRIMP) analysis of several hundred samples.

Northwest Pilbara Craton

Tectonic units

Tectonic units provide the geological framework used in writing this Report. The northwest Pilbara Craton is divided into four types of tectonic unit: terranes (including one superterrane), basins (including one superbasin), supersuites, and a 'tectonic zone', within which most of the first three types of unit developed (Figs 3–6). A terrane is defined as a fault-bounded body of rock of regional extent, characterized by a geological history different from that of contiguous bodies of rock. A superterrane consists of accreted terranes. A basin is an area underlain by a substantial thickness of sedimentary or volcanic rocks, which has unifying characteristics of stratigraphy and structure due to deposition in a regionally restricted area. Basins are bounded by unconformities except where major faults have juxtaposed a basin with another tectonic unit. A superbasin is a connected series of basins. Supersuites, consisting of multiple igneous suites and intrusions, were first established in the Pilbara as lithostratigraphic units (Van Kranendonk et al., 2006). However, they were emplaced by tectonomagmatic events, which in some instances (Elizabeth Hill and Split Rock Supersuites) are not known to be represented by any volcanic successions. Accordingly, supersuites are now also used by GSWA as tectonic units.

Figure 4 includes reference to isotopic data, indicating that certain major faults and shear zones of the northwest Pilbara (e.g. Sholl, Loudens, and Tabba Tabba) separate segments of crust formed at different stages in the geological evolution of the craton. These differences suggest minimum relative strike-slip displacements of tens of kilometres. Sedimentary and volcanic basins shown in Figure 5 are intruded by the predominantly granitic supersuites. Stratigraphic and tectonic relationships of the individual terranes, basins, and supersuites are illustrated diagrammatically in Figure 6. Each of the basins contains a unique volcanic or sedimentary succession. With two exceptions (Elizabeth Hill and Split Rock), the supersuites were contemporaneous with individual basins, where they may be partly represented by volcanic rocks, but they were also emplaced into older terranes and basins.

The two major tectonic units of the northwest Pilbara Craton are the West Pilbara Superterrane (WPS) (3280–3066 Ma) and the De Grey Superbasin (3066–2919 Ma) (Figs 3 and 6). However, the northeast–southwest-trending Central Pilbara Tectonic Zone (CPTZ) (Fig. 3), which runs through both these units, was integral to their development (Hickman, 1999; Hickman et al., 2001). Whereas the WPS and De Grey Superbasin are defined by lithostratigraphy, the CPTZ is a belt of tectonic and magmatic activity between two independently moving plates. In the early Mesoarchean these plates were the EPT and Karratha Terrane (KT), but from c. 3066 Ma they were the EPT and the WPS.

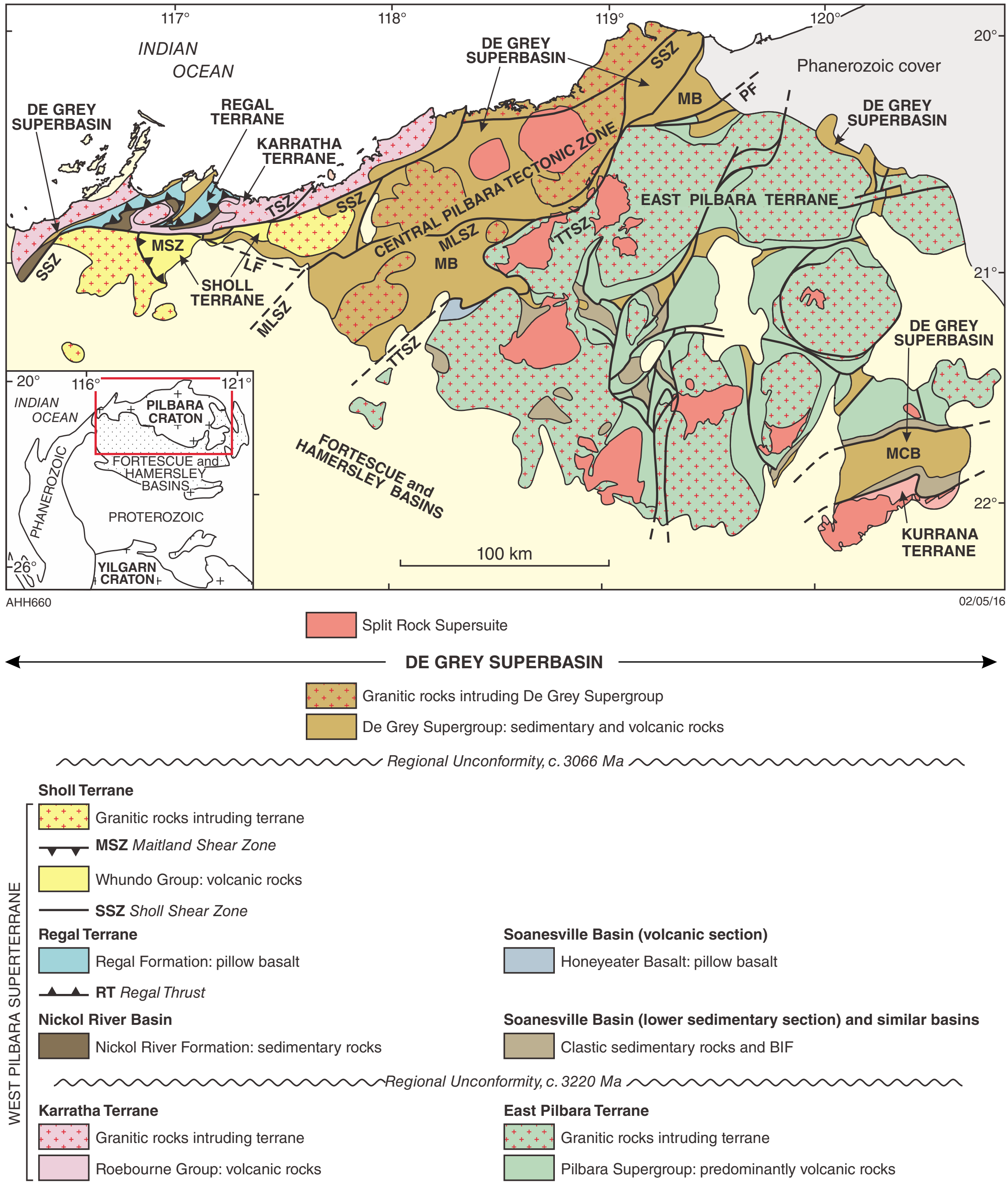
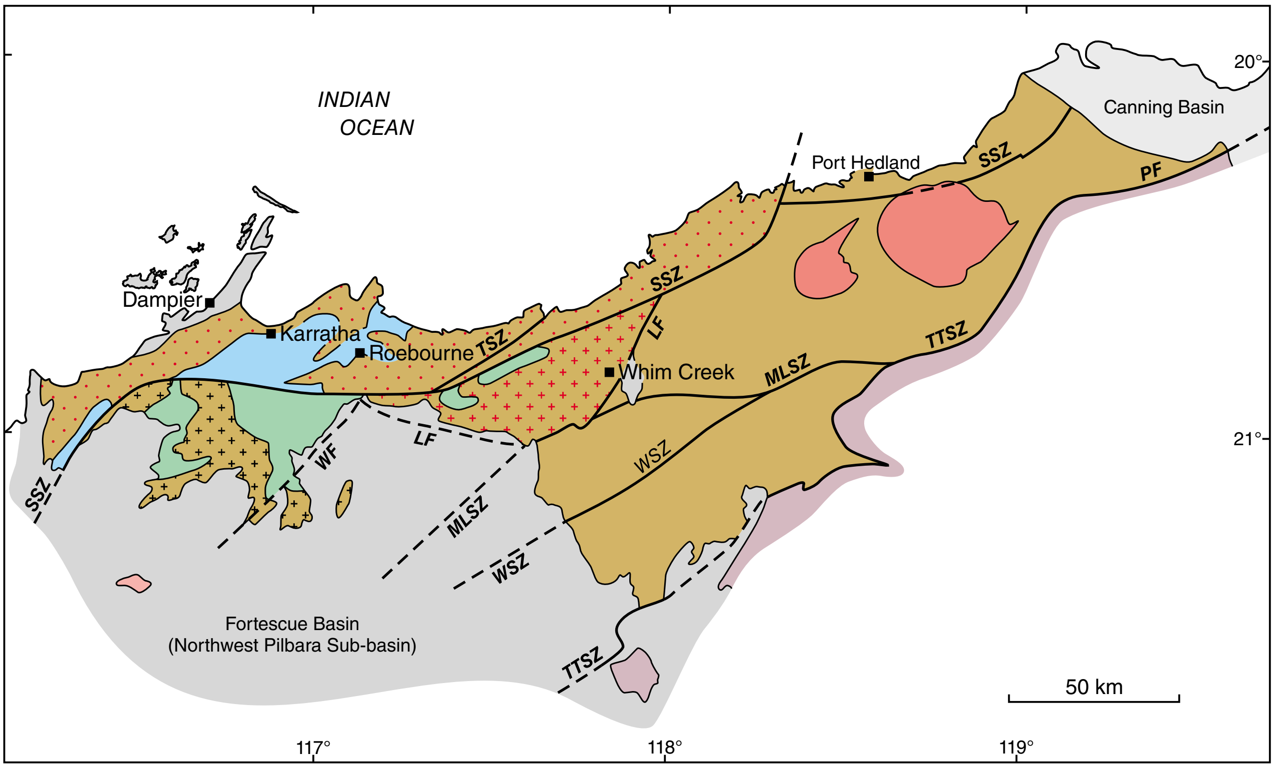


Figure 3. **Major tectonic units of the northern Pilbara Craton. Abbreviations: LF, Loudens Fault; MB, Mallina Basin; MCB, Mosquito Creek Basin; MLSZ, Mallina Shear Zone; MSZ, Maitland Shear Zone; PF, Pardoo Fault (part of TTSZ); SSZ, Sholl Shear Zone; TSZ, Terenar Shear Zone; TTSZ, Tabba Tabba Shear Zone**



Fortescue Basin (2775–2629 Ma): unconformably overlies Pilbara Craton

Pilbara Craton

Split Rock Supersuite (2851–2831 Ma): intrudes De Grey Superbasin

CENTRAL PILBARA TECTONIC ZONE

De Grey Superbasin of Cherratta Granitic Complex (3006–2924 Ma): overlies or intrudes Sholl Terrane. Xenocrystic zircon Pb–Pb ages <3280 Ma; Nd T_{DM} model ages <3240 Ma.

De Grey Superbasin in central part of Mallina Basin (3015–2931 Ma): overlies or intrudes Mesoarchean crust. Nd T_{DM} model ages typically <3230 Ma.

De Grey Superbasin of Whim Creek greenstone belt and Caines Well Granitic Complex (3066–2925 Ma): overlies or intrudes Sholl Terrane. Nd T_{DM} model ages <3400 Ma.

West Pilbara Superterrane south of Sholl Shear Zone (3130–3066 Ma); Sholl Terrane and Elizabeth Hill Supersuite. Nd T_{DM} model ages <3300 Ma.

De Grey Superbasin north of Sholl Shear Zone (3066–2931 Ma): overlies or intrudes Karratha and Regal Terranes. Nd T_{DM} model ages <3520 Ma.

Mingah Granitic Complex (c. 3236 Ma): Nd T_{DM} model age 3360 Ma.

West Pilbara Superterrane north of Sholl Shear Zone (3280–3160 Ma); Karratha and Regal Terranes and Nickol River Basin. Nd T_{DM} model ages <3520 Ma.

Undivided East Pilbara Terrane (3530–3220 Ma) and associated younger units. Xenocrystic zircon Pb-Pb ages 3725–3550 Ma. Nd T_{DM} model ages <3760 Ma.

Major fault or shear zone; concealed where dashed

Figure 4. Geological map showing the principal tectonostratigraphic divisions and structures of the northwest Pilbara Craton, including an interpretation of underlying crustal ages. Note that units within the Central Pilbara Tectonic Zone are underlain by relatively young crust, but that isotopic data indicate Paleoarchean crust between the Loudens Fault and Sholl Shear Zone. Abbreviations: LF, Loudens Fault; MLSZ, Mallina Shear Zone; PF, Pardoo Fault; SSZ, Sholl Shear Zone; TSZ, Terenar Shear Zone; TTSZ, Tabba Tabba Shear Zone; WF, Woodbrook Fault; WSZ, Wohler Shear Zone

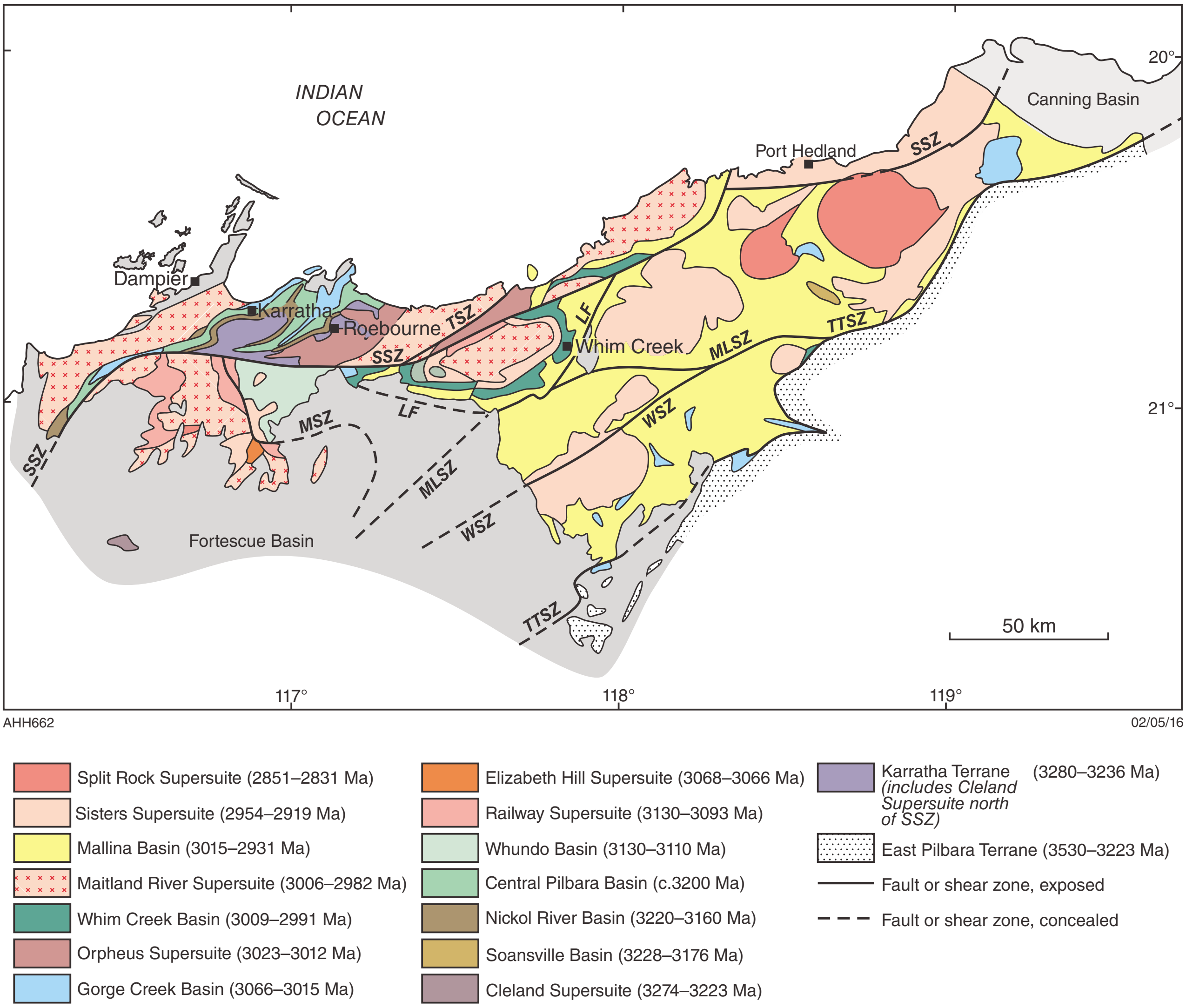


Figure 5. Geological map showing basins and supersuites of the northwest Pilbara Craton. Ages and contact relationships are summarized in Figure 6. Note that the supersuites (predominantly granitic) are distributed in east-northeast-trending linear zones with decreasing intrusive ages towards the southeast. Abbreviations: LF, Loudens Fault; MLSZ, Mallina Shear Zone; SSZ, Sholl Shear Zone; TSZ, Terenar Shear Zone; TTSZ, Tabba Tabba Shear Zone; WSZ, Wohler Shear Zone

The WPS is a tectonic collage of the KT, Regal Terrane (RT), and Sholl Terrane (ST) and the Nickol River Basin (Figs 4 and 6). Positioned between the KT and RT, the stratigraphic base of the Nickol River Basin marks the regional unconformity produced by c. 3220 Ma continental breakup of the EPT. The De Grey Superbasin unconformably overlies the WPS (Fig. 3), and its stratigraphic base is the regional unconformity produced by erosion following uplift associated with terrane collision (3068–3066 Ma Prinsep Orogeny).

Greenstone belts and granitic complexes

As in all Archean cratons worldwide, early descriptions of the Pilbara Craton referred to ‘greenstone belts’ and ‘granites’ (or ‘granite batholiths’). These simple lithological divisions required no prior knowledge of an area’s stratigraphy or structure. The term ‘greenstone belt’ was applied to any tract of country, commonly linear, that was underlain by supracrustal assemblages of metamorphosed volcanic, sedimentary, and mafic intrusive rocks, and which contained no large granitic intrusions. Geological investigations revealed most greenstone belts to be broadly synclinal or synformal or, in some instances, volcano-sedimentary packages contained within detached fold limbs. The ‘granites’, on the other hand, were first interpreted by many workers to be major plutons or batholiths that had intruded the greenstones. An increased use of geochronology in the Pilbara Craton established that most of the ‘batholiths’ are actually composed of multiple intrusions differing in age by several hundred million years.

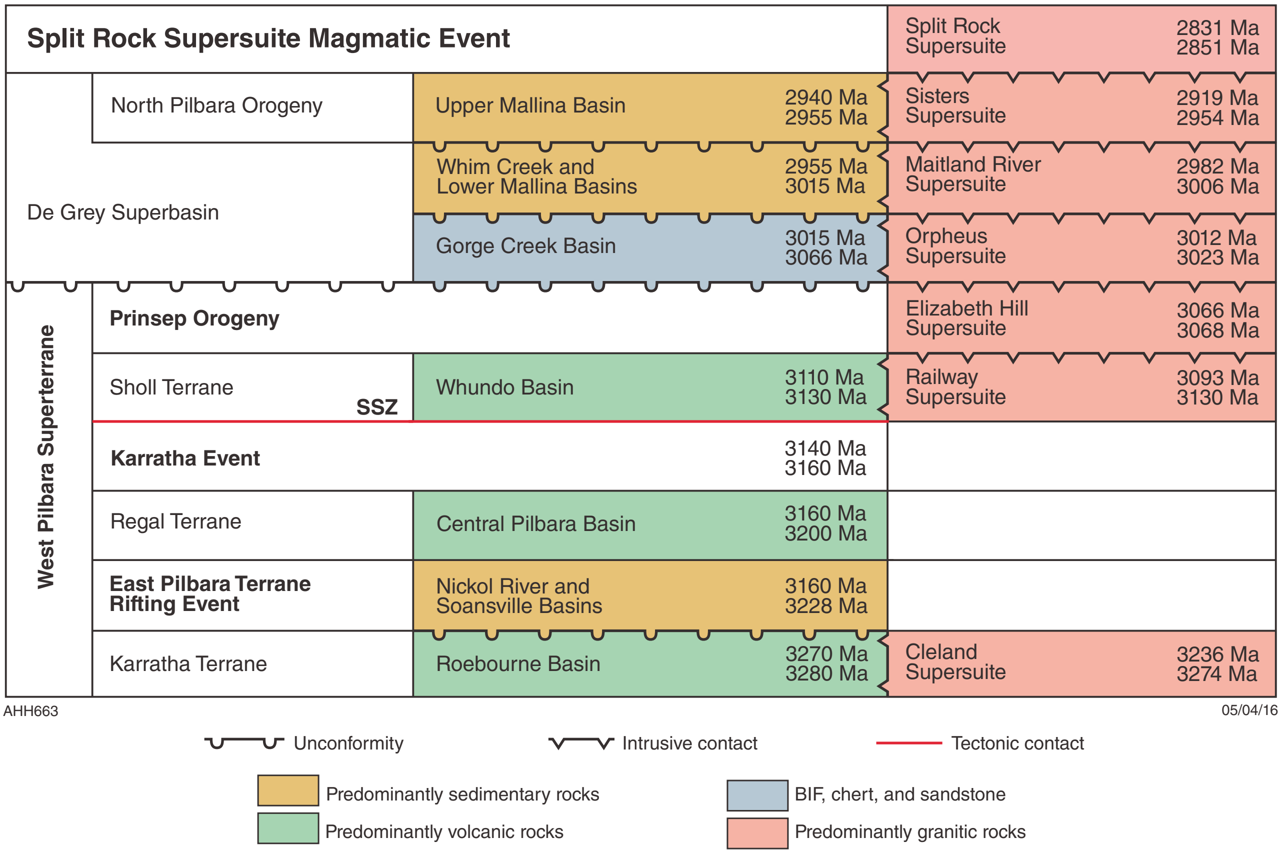


Figure 6. Diagrammatic illustration of the ages and contact relationships of terranes, basins, supersuites, and events in the northwest Pilbara Craton. The tectonic contact between the Nickol River Basin and the Central Pilbara Basin (Regal Terrane) is the Regal Thrust, whereas that between the Regal Terrane and the Sholl Terrane is the Sholl Shear Zone. Regional unconformities at c. 3220 Ma, 3066 Ma and c. 2955 Ma coincide with periods of erosion during major rifting or following collision orogeny. Deposition in volcanic and sedimentary basins was generally accompanied by supersuite intrusion.

The term batholith was therefore replaced by 'granitoid complex' (Griffin, 1990), later becoming 'granitic complex' (Hickman and Van Kranendonk, 2004). The boundaries of the granitic complexes, some of which are up to 100 km in diameter, control the overall dome-and-keel pattern of granites and greenstone belts in the northern Pilbara. This classically Archean pattern, best developed in the EPT (Fig. 3), owes its origin to the fact that successive granitic intrusions, emplaced over hundreds of millions of years, intruded only the pre-existing granitic domes, thus progressively accentuating the pattern (Hickman and Van Kranendonk, 2004). Greenstone belts were rarely intruded, except locally in extensional settings immediately above rising domes.

Greenstone belts and granitic complexes are essentially structurally controlled lithological packages which, unlike terranes and basins, are not distinguished by unique stratigraphic successions or separate structural history. Even so, naming them provides a convenient way to refer to geographically defined regional-scale outcrops of either supracrustal rocks or granites. It is therefore useful to outline the various greenstone belts and granitic complexes that make up the northwest Pilbara Craton (Fig. 7).

Greenstone belts

A preliminary division of the greenstone belts in the northern Pilbara Craton was provided by Hickman (1980b). In the northwest Pilbara the author distinguished the Regal, Sholl, Sherlock, Whim Creek, and Boodarie belts, plus two units which were given structural names, the Roebourne Syncline and the Mallina Synclinorium. More detailed geological mapping and new geophysical data have subsequently identified several important faulted contacts that require an extensive revision of the 1980 interpretation. Figure 7 shows the locations of the greenstone belts and Table 1 outlines the ages and relations of their component formations within the regional stratigraphy of the northern Pilbara Craton.

Devil Creek greenstone belt

The Devil Creek greenstone belt (Fig. 7) is a narrow belt of greenstones, up to 5 km wide, and extends 40 km southwest from the Yanyare River Bridge on the North West Coastal Highway (Strong et al., 2000; Hickman et al., 2006b). This greenstone belt is bounded on its southeast side by the SSZ and on its northwest margin is intruded by the Dampier Granitic Complex. The main structure of the belt is dominated by a northeast–southwest-trending syncline composed, in ascending stratigraphic order, c. 3220 Ma metasedimentary rocks of the Nickol River Formation, c. 3200 Ma metabasalt of the Regal Formation, and c. 3020 Ma banded iron-formation (BIF) and chert of the Cleaverville Formation in the De Grey Superbasin (Table 1, Fig. 3). Geophysical data indicate that the Devil Creek greenstone belt may be separated from the Roebourne greenstone belt by a combination of granite intrusion and shearing over a 10 km strike length west from the Maitland River Bridge. It is therefore concluded that remnants of the Roebourne Group and Karratha Granodiorite, locally intruded by Mesoarchean granites, underlie the entire area of the Roebourne greenstone belt. Localities not shown on figures are listed in Appendix 1.

Roebourne greenstone belt

The Roebourne greenstone belt (Fig. 7) occupies approximately 1400 km^2 of the onshore northwest Pilbara and geophysical data indicate that it extends at least 20 km northeast offshore. The southern margin of this greenstone belt is defined by the SSZ, whereas in the northwest and southeast it is intruded by granitic complexes (Dampier and Harding). Additionally, the greenstone belt is unconformably overlain by the Fortescue Group. The metamorphosed volcanic and sedimentary stratigraphic units of the greenstone belt are, in ascending order, the c. 3280 Ma Roebourne Group, c. 3220 Ma Nickol River Formation, c. 3200 Ma Regal Formation, and c. 3020 Ma Cleaverville Formation (Fig. 8, Table 1). Along its northwestern and eastern margins the base of the greenstone belt is intruded by the Dampier and Harding Granitic Complexes (see below) and it is centrally intruded by the 3270–3260 Ma Karratha Granodiorite. The greenstone belt contains three large c. 2940 Ma northeast-trending folds (Cleaverville Synform, Prinsep Dome, and Roebourne Synform, Figs 8 and 9) that deform both the greenstones and the granitic rocks. The Karratha Granodiorite outcrops in the core of the Prinsep Dome and on the southeast limb of the Roebourne Synform, and most likely extends at depth to the northwest limb of the Cleaverville Synform. In that area the Dampier Granitic Complex is extensively intruded by the younger Maitland River Supersuite and is very poorly exposed, but there are isotopic data indicating the presence of Paleoarchean rocks (see Dampier Granitic Complex).

The Cleaverville Formation unconformably overlies the Regal Formation, and the Regal Formation is separated from the Nickol River Formation and the Roebourne Group by the bedding-parallel Regal Thrust (see below). This thrust is a terrane boundary between the RT and the underlying KT (Figs 8 and 9).

Sholl greenstone belt

The Sholl greenstone belt (Fig. 7) is exposed over an area of approximately 400 km^2 south of the SSZ. The largest exposure of the greenstone belt lies between Whundo and Bradley Well, 50 km south of Karratha, but other exposures are present near Mount Fisher adjacent to the Caines Well Granitic Complex (Fig. 8). Geophysical evidence indicates that the Sholl greenstone belt is unconformably overlain by the <2775 Ma Fortescue Group over a large area east of Whundo (Hickman, 2004b). The belt is composed of the 3130–3110 Ma Whundo Group and several intrusions of the Sisters and Orpheus Supersuites. Its boundaries are the SSZ in the north, the Maitland Shear Zone (MSZ) in the west and south, and the Fortescue Group to the east (Fig. 8). The MSZ originated from thrusting of the Whundo Group over the Cherratta Granitic Complex to the west and south, and therefore the Sholl greenstone belt is structurally underlain by that granitic complex, at least in the west.

Weelarra greenstone belt

The small Weelarra greenstone belt (Fig. 7), 10 km south of Mount Leopold and surrounded by the Cherratta Granitic Complex and outliers of the Fortescue Group, was first identified by GSWA mapping (Hickman and Kojan, 2003). It is composed of mafic and ultramafic rocks metamorphosed to amphibolites facies, and is extensively intruded by granitic rocks and pegmatite. The greenstone belt strikes east–northeast and lies within an anomalously magnetic zone, 25 km long and parallel to the Yanyare River Fault. More field data are required to establish if the Weelarra greenstone belt is a remnant of the Whundo Group or if it is an ultramafic–mafic intrusive unit, metamorphosed and extensively broken up by granitic intrusion.

Whim Creek greenstone belt

The Whim Creek greenstone belt (Fig. 7) is essentially a ring-shaped belt of metamorphosed volcanic and sedimentary rocks surrounding an elongate uplifted core of granitic rocks (Caines Well Granitic Complex) within the c. 2940 Ma Whim Creek Anticline (Fig. 8). The outer northern, eastern, and southern margins of the greenstone belt are major strike-slip faults (SSZ in the north and the Loudens Fault [LF] in the south). The southwest outer margin is an unconformity with the Fortescue Group, but geophysical data indicate that LF continues west beneath the Fortescue Group to rejoin the SSZ 10 km south of Roebourne. In that scenario, all the outer margins of the Whim Creek greenstone belt are strike-slip faults. Prior to erosion and exposure of the central core of granitic rocks the total area of the belt would have been at least 1200 km^2. The stratigraphy of the Whim Creek greenstone belt is, in ascending stratigraphic order, the 3130–3110 Ma Whundo Group, c. 3020 Ma Cleaverville Formation, 3010–2990 Ma Whim Creek Group, and the c. 2950 Ma Bookingarra Group (Table 1, redefined unit Appendix 2). Additionally, a major ultramafic–mafic layered sill of the Langenbeck Suite, the Sherlock Intrusion, intrudes most of the central granite–greenstone contact (Fig. 8).

Peawah Hill greenstone belt

The poorly exposed Peawah Hill greenstone belt (Fig. 7) is interpreted to be the faulted northern section of the Whim Creek greenstone belt. The two belts were separated by

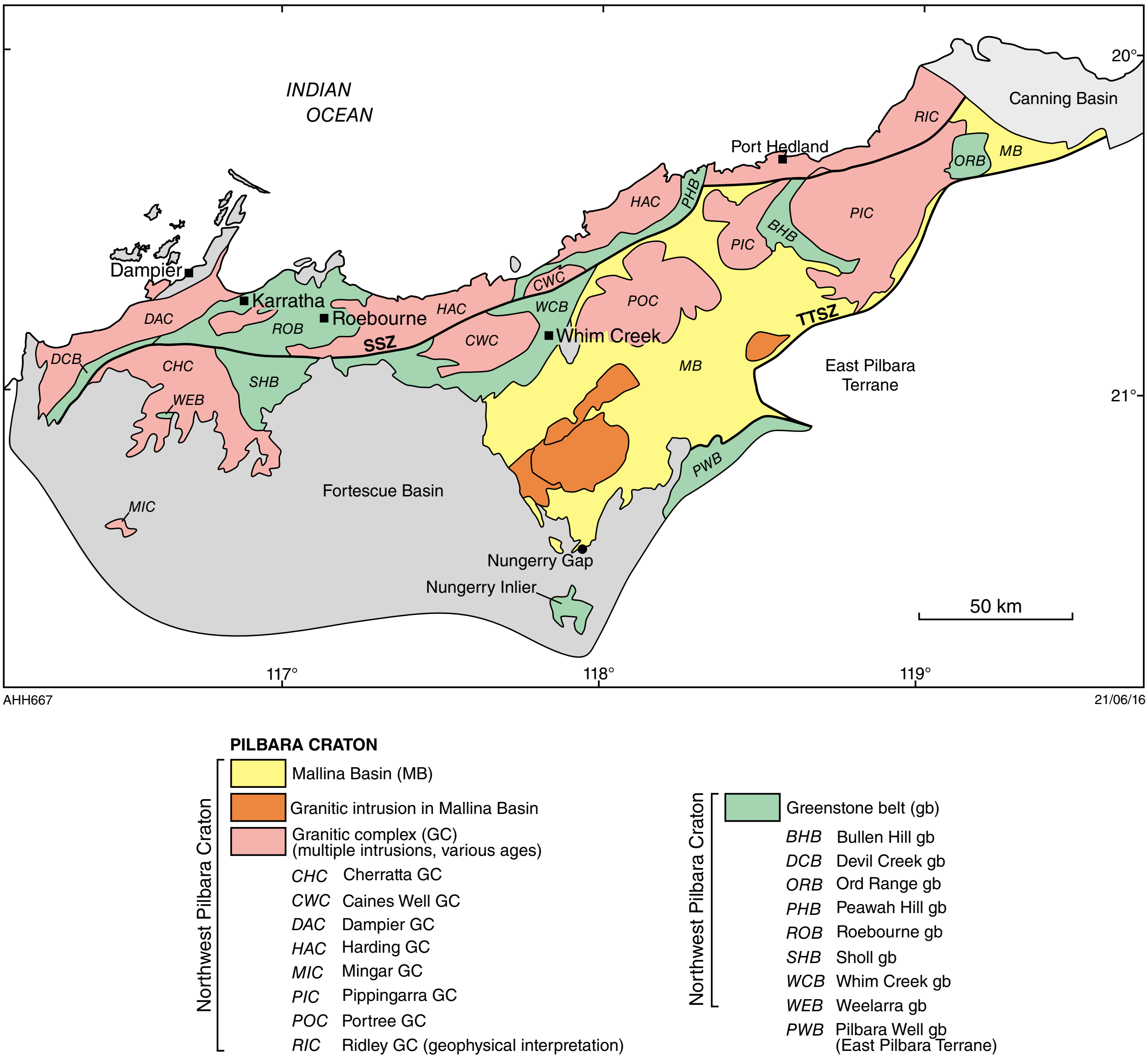


Figure 7. **Greenstone belts and granitic complexes of the northwest Pilbara Craton. Each of the granitic complexes contains more than one supersuite (Table 1). Abbreviations: SSZ, Sholl Shear Zone; TTSZ, Tabba Tabba Shear Zone**

30–40 km of dextral strike-slip movement along the SSZ between 2940 and 2920 Ma. The northern margin of the belt is concealed and could be either unconformable or intrusive with the Harding Granitic Complex. The Peawah Hill greenstone belt is c. 300 km^2 in area and is interpreted to contain very similar stratigraphy to the Whim Creek greenstone belt.

Bullen Hill greenstone belt

The 600 km^2 Bullen Hill greenstone belt (new name) outcrops intermittently in the central part of the Mallina between 10 and 60 km south of Port Hedland (Fig. 7). The BIF and chert at Bullen Hill are interpreted to be part of the Cleaverville Formation (Table 1), in which case the predominantly mafic volcanic succession of the belt overlies the Gorge Creek Group. The succession could be stratigraphically equivalent to metamorphosed mafic volcanic and intrusive rocks in the lower part of the Croydon Group.

Ord Range greenstone belt

The Ord Range greenstone belt is an inlier of the Gorge Creek Group overlain and surrounded by rocks of the Mallina Basin succession (Fig. 7). A poorly exposed succession of mafic and intermediate volcanic rocks that unconformably overlies the Gorge Creek Group east of the Ord Range is correlated with the basaltic Salt Well Member of the Lalla Rookh Sandstone and with basalt within the Mallina Basin approximately 50 km to the west.

Table 1. Generalized lithostratigraphy of the Pilbara Craton. In some instances, stratigraphic units in different terranes are interpreted to be laterally equivalent.

Age (Ma)	Terrane/Superbasin	Basin	Group	Formation
< 2980	De Grey Superbasin (De Grey Supergroup)	Mosquito Creek Basin	Nullagine Group	Mosquito Creek (EP)
c. 2940		Mallina Basin	Croydon Group (?)	Mallina (upper)
c. 2955			Bookingarra Group	Mount Negri Volcanics
				Louden Volcanics
				Rushall Slate
				Cistern Formation
			Unconformity	
< 2990			Croydon Group	Cattle Well (EP)
				Mallina (lower)
c. 3010				Constantine Sst.
			Tectonic contact	
c. 3010		Whim Creek Basin	Whim Creek Group	Red Hill Volcanics
				Warambie Basalt
			Unconformity	
		Gorge Creek Basin	Gorge Creek Group	Cundaline (EP)
c. 3020				Cleaverville
				Farrel Quartzite (EP)
			Regional Unconformity	
> 3110	Sholl Terrane	Whundo Basin	Whundo Group	Woodbrook
c. 3120				Bradley Basalt
c. 3125				Tozer
< 3130				Nallana
			Tectonic Contact —Sholl Shear Zone	
	Regal Terrane			Port Robinson Basalt
c. 3200				Regal
			Tectonic Contact — Regal Thrust	
< 3220	Rift-related successions	Soanesville Basin	Soanesville Group	Honeyeater Basalt (EP)
				Paddy Market (EP)
				Corboy (EP)
				Cardinal (EP)
< 3220		Budjan Creek Basin		Budjan Creek (EP)
< 3220		Coondamar Basin		Coondamar (EP)
3220–3160		Nickol River Basin		Nickol River/Dixon Island
c. 3280	Karratha Terrane		Roebourne Group	Weerianna Basalt
				Ruth Well
			No Contact	
c. 3235	East Pilbara Terrane (Pilbara Supergroup)		Sulphur Springs Group	Kangaroo Caves (EP)
				Kunagunarrina (EP)
< 3270				Leilira (EP)
			Unconformity	
			Kelly Group	Charteris Basalt (EP)
c. 3320				Wyman Formation (EP)
c. 3350				Euro Basalt (EP)
			Disconformity	
c. 3420				Strelley Pool (EP)
			Regional unconformity	
c. 3430			Warrawoona Group	Panorama (EP)
				Apex Basalt (EP)
c. 3465				Duffer (EP)
				Mount Ada Basalt (EP)
c. 3480				Dresser/McPhee (EP)
> 3490				North Star Basalt (EP)
				Double Bar (EP)
c. 3515				Coucal (EP)
c. 3525				Table Top (EP)

Note: (EP) — East Pilbara only

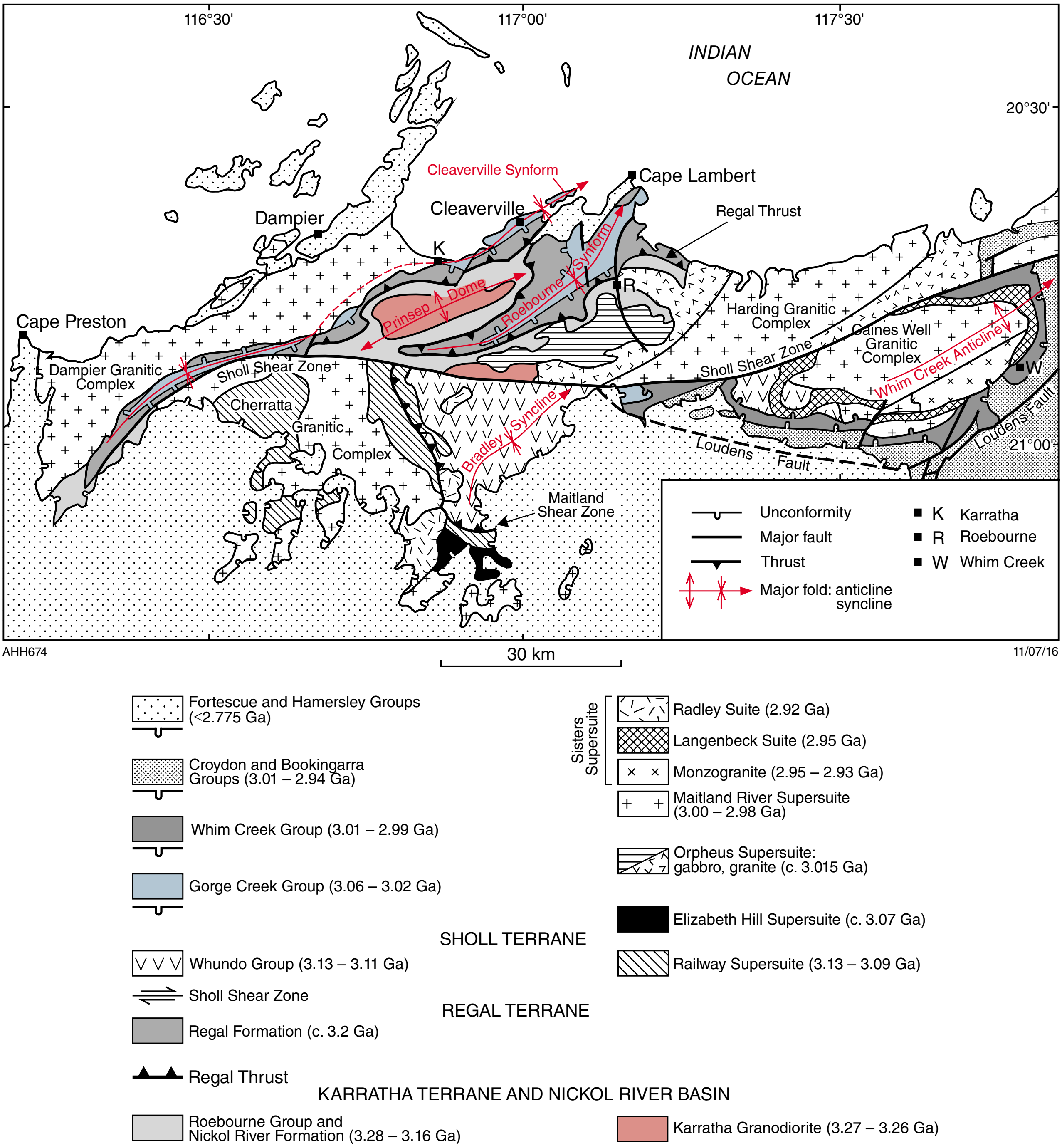
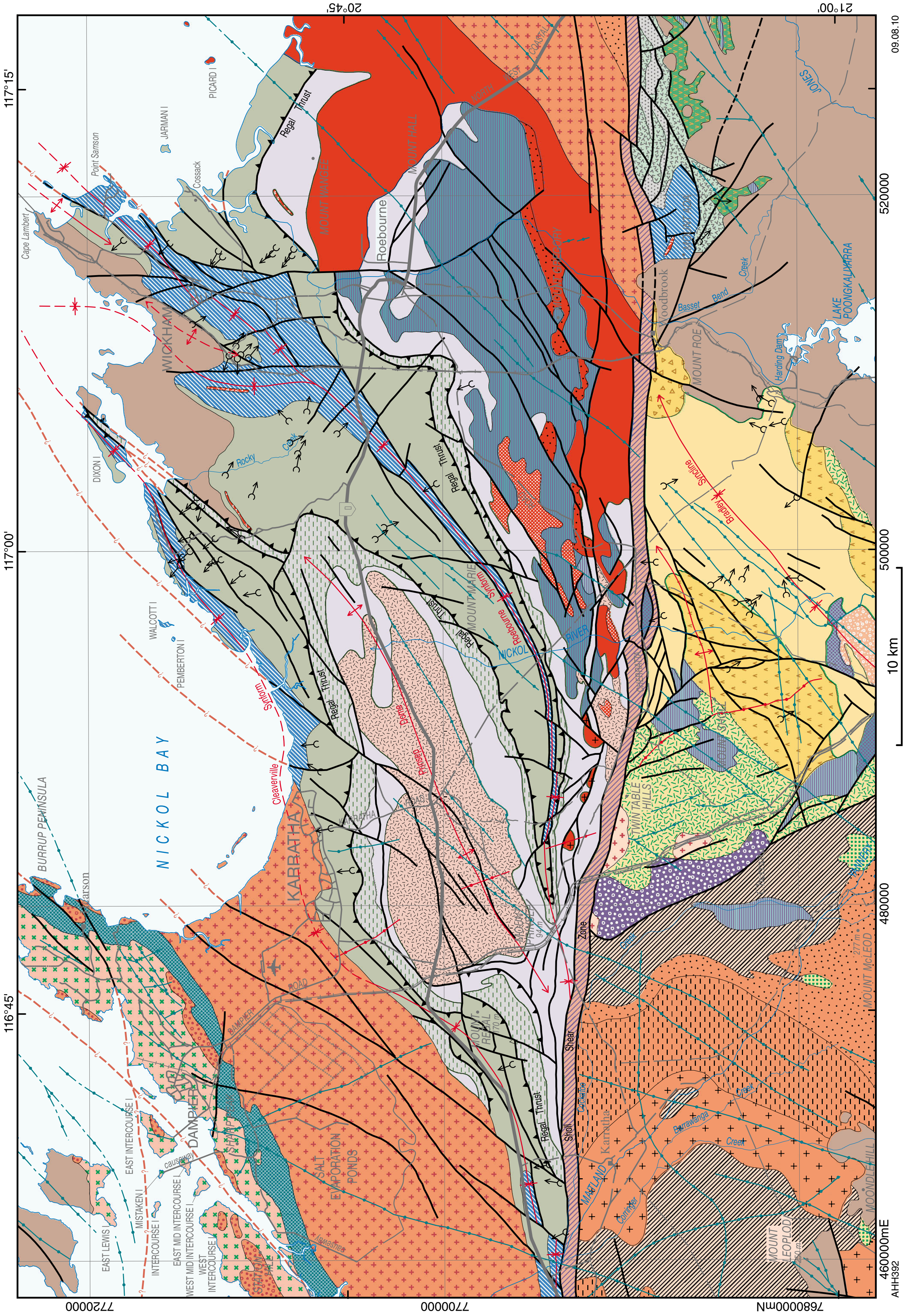


Figure 8. Simplified geological map of the northwest Pilbara Craton between Cape Preston and Whim Creek, showing lithostratigraphy, tectonic units, and major structures



20°45'
21°00'
117°15'
117°00'
116°45'
520000
500000
480000
460000mE
7720000
7700000
7680000mN
09.08.10
AHH392
10 km
NICKOL BAY
KARRATHA
Roebourne
Wickham
DAMPIER
BURRUP PENINSULA
Cossack
Point Samson
Cape Lambert
JARMAN I
PICARD I
DIXON I
WALCOTT I
PEMBERTON I
EAST LEWIS I
MISTAKEN I
EAST INTERCOURSE I
WEST INTERCOURSE I
SALT EVAPORATION PONDS
Regal Thrust
Sholl Shear Zone
Roebourne Synform
Cleaverville Syncline
Bradley Syncline
Prinsep Dome
MOUNT WANGEE
MOUNT HALL
MOUNT MARIE
MOUNT REGAL
MOUNT ROE
MOUNT McLEOD
MOUNT LEOPOLD
MOONDINE HILL
TWIN TABLE HILLS
LAKE POONGKALIARRA
Harding Dam
Woodbrook
Karratha
MAITLAND
NICKOL RIVER
Rocky Creek
Basset Bend Creek

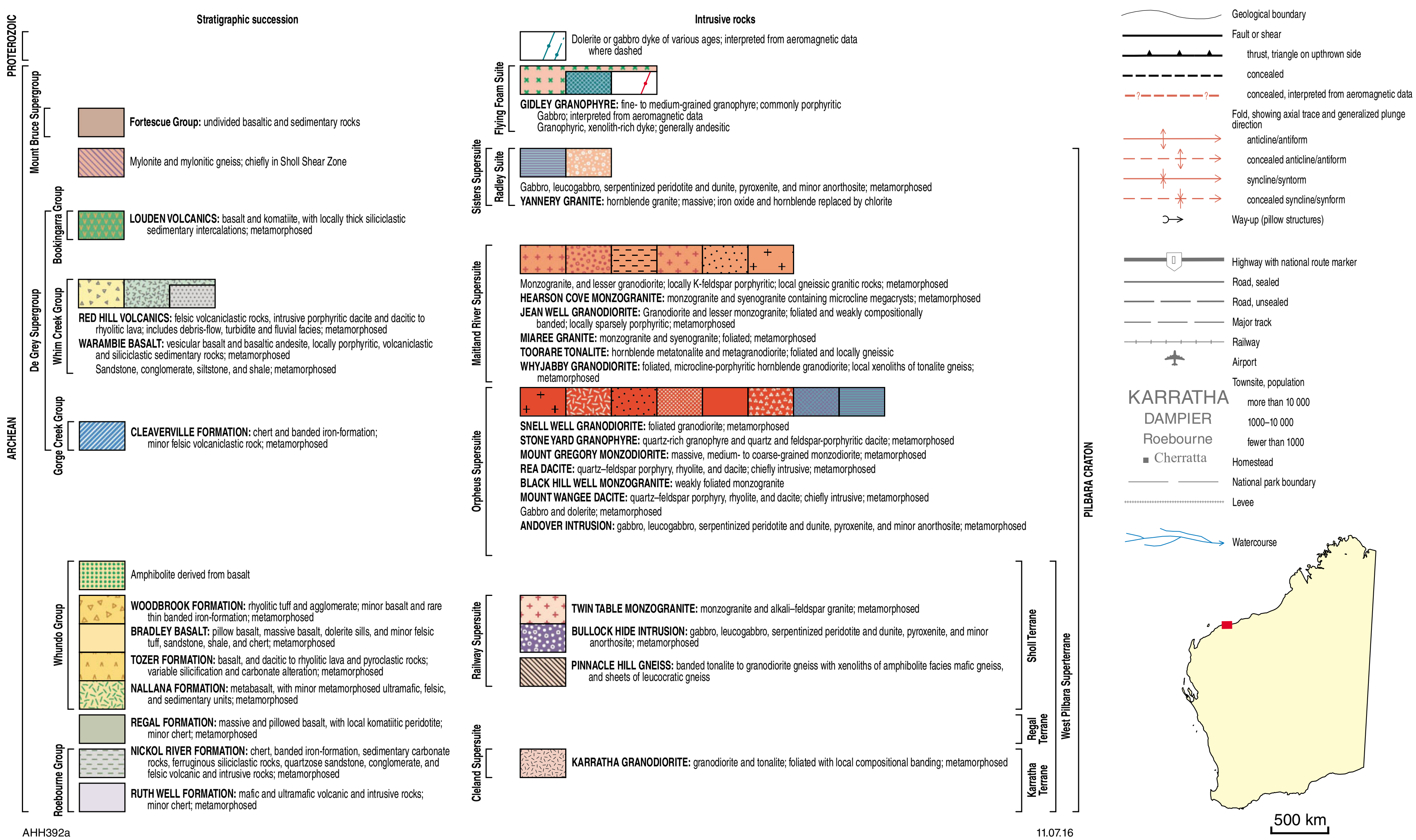


Figure 9. Geological map of the Roebourne–Karratha area, showing major structures and way-up evidence (from pillow structures) in the Regal Formation. Note that the c. 3160 Ma Regal Thrust is folded by the c. 2940 Ma Prinsep Dome, Roebourne Synform, and Cleaverville Synform (modified from Hickman et al., 2006b).

Granitic complexes

Localized accumulations of granitic intrusions belonging to more than one supersuite are referred to as granitic complexes. In the east Pilbara the age ranges of supersuites, suites, and intrusions in granitic complexes are up to 650 million years. However, in the northwest Pilbara, age ranges do not exceed 300 million years (Table 2). Granitic complexes are almost invariably restricted to uplifted areas of the crust, e.g. in the cores of domes or anticlines. This is consistent with the general layering of the Pilbara crust, in which greenstones overlie granitic rocks. Just as adjacent Pilbara greenstone belts normally share a common supracrustal succession, so adjacent Pilbara granitic complexes typically contain some of the same supersuites. Table 2 reveals that within the northwest Pilbara individual granitic complexes contain many of the same supersuites. The Mingar, Portree, and Ridley Granitic Complexes are not included in Table 2 owing to insufficient data.

It is significant that the zircon data from the listed complexes provide no evidence for the Railway or Elizabeth Hill Supersuites being present north of the SSZ, i.e. within the Dampier or Harding Granitic Complexes. Later in this Report it is emphasized that the Railway Supersuite, which is part of the ST (Fig. 3), originated in an entirely different tectonic setting from the KT north of the SSZ.

There is a considerable volume of isotopic evidence, including zircon data outlined briefly in Table 2 to suggest that, with the exception of the Portree Granitic Complex, the granitic complexes of the northwest Pilbara were initiated by intrusion of the Cleland Supersuite between 3274 and 3223 Ma. This is directly established for the Harding Granitic Complex, which includes c. 3265 Ma granodiorite, and is inferred for the Dampier, Cherratta, and Pippingarra Granitic Complexes, which contain xenocrystic zircon of this age. Inherited zircons in granites may have been recycled from detrital zircon in older sedimentary rocks, but for the granitic complexes involved such potential sources are limited to the Cleaverville and Nickol River Formations.

Dampier Granitic Complex

Available geochronology on the 1500 km^2 onshore part of the Dampier Granitic Complex (Fig. 7) indicates that it is predominantly composed of the Maitland River Supersuite, but older banded gneiss of unknown age is locally exposed. Nd T_{DM} model ages (T_{DM} model age indicating the time at which the isotopic composition of a sample was equivalent to that of the depleted mantle [DM] model) between 3387 and 3247 Ma indicate that the supersuite intruded, and may have been partly derived from, Paleoarchean crust. Analysis of the core of a zircon grain in the Miaree Granite (GSWA 136844, Nelson, 1998e) gave a $^{207}Pb/^{206}Pb$ age of c. 3255 Ma, which is similar to the crystallization age of the 3270–3261 Ma Karratha Granodiorite less than 5 km to the south of the Dampier Granitic Complex. As noted above (Roebourne greenstone belt), the present interpretation is that the Karratha Granodiorite extends northwestwards from the Prinsep Dome beneath the narrow belt of greenstones in the core of the Cleaverville Synform (Fig. 8), to form part of the Dampier Granitic Complex.

Cherratta Granitic Complex

The most westerly granitic complex of the northwest Pilbara Craton is the Cherratta Granitic Complex which, including sections concealed by the overlying Fortescue Basin, occupies an area of approximately 4000 km^2. The complex is made up of various intrusions of the Railway, Elizabeth Hill, Maitland River, and the Sisters Supersuites, establishing growth by a series of intrusive events over at least 200 million years (3130–2920 Ma). Banded gneiss (Fig. 10) is common in the Cherratta Granitic Complex, testifying to multiple magmatic events. Xenocrystic zircon grains in two intrusions in the eastern part of the complex are of Cleland Supersuite age. The Weelarra greenstone belt containing mafic volcanic and intrusive rocks (Hickman and Kojan, 2003) outcrops in the southwestern part of the complex, and geophysical data indicate that concealed greenstones underlie the Fortescue Group to the west. The Cherratta Granitic Complex is bounded by SSZ on its northern and northwestern sides and by the MSZ to the east. The southeast part of the complex is almost entirely concealed by the Fortescue Group, except for isolated exposures in two areas, one near Mardeburra Pool on Fish Creek (GSWA 168932, Nelson, 2001e) and a larger area near Waloo Waloo Pool (GSWA 168934, Nelson, 2001f). SHRIMP analysis revealed that the Maitland River Supersuite of this southeastern area contains considerably older xenocrystic zircon than in the main exposure to the northwest. Over 50% of zircon in the Mardeburra Granodiorite was dated as older than 3100 Ma, with the oldest grain being 3279 Ma.

Table 2. Supersuites in granitic complexes in the northwest Pilbara Craton

Supersuite	*Cherratta (S)*	*Dampier (N)*	*Harding (N)*	*Caines Well (S)*	*Pippingarra (S)*
Sisters	2924±5 Ma	2952 ±6 Ma	2931±8 Ma	2925±4 Ma	2946±5 Ma
Maitland River	2995±11 Ma	2997±3 Ma	z. 2997 Ma	2990±5 Ma	z. 3005 Ma
Orpheus	3013±4 Ma	z. 3026 Ma	3016±4 Ma	z. 3011 Ma	3018±9 Ma
Elizabeth Hill	3068±4 Ma			z. 3066 Ma	
Railway	3130±4 Ma	z. 3106 Ma		3093±4 Ma	z. 3114 Ma
Cleland	z. 3279 Ma	z. 3255 Ma	3265±4 Ma		z. 3246 Ma

NOTE: z = crystallization ages of supersuites interpreted to be present based on xenocrystic zircon ages. The other dates are SHRIMP published ages.

Figure 10. Banded gneiss in the Cherratta Granitic Complex at Toorare Pool, Maitland River. The gneiss contains layers of amphibolite and intrusive sheets of hornblende–granodiorite and monzogranite (Zone 50, MGA 479060E 7678550N).

The zircon groups in this sample suggest intrusion through the Cleland and Railway Supersuites.

Mingar Granitic Complex

A pronounced gravity low in the area of the Fortescue River indicates the existence of a large concealed granitic complex beneath the Fortescue Group. The northern part of this geophysical anomaly coincides with the Mingar Dome in the Fortescue Group (Williams, 1968). By analogy with deformation of the Fortescue Group in the east Pilbara (Hickman, 1983; Van Kranendonk, 2003) thick granitic crust in this area may explain domal uplift of the overlying Fortescue Group. The interpreted concealed granitic complex, approximately 4000 km^2 in area, is here named the Mingar Granitic Complex. Using the geophysical interpretation of Blewett et al. (2000), the Mingar Granitic Complex is separated from the Cherratta Granitic Complex by a narrow greenstone belt but this is entirely concealed. The Mingar Granitic Complex is exposed in several large outcrops along the valley of the Fortescue River. These exposures are composed of foliated hornblende–biotite tonalite assigned to the Tarlwa Pool Tonalite. SHRIMP dating indicated a crystallization age of 3236 ± 3 Ma (GSWA 142535, Nelson, 1998l), and the Nd T_{DM} model age was 3360 Ma (Smithies et al., 2007a).

Harding Granitic Complex

The Harding Granitic Complex lies north of the SSZ and extends over 130 km along the Pilbara coast (Fig. 7). The onshore width of the complex is generally less than 20 km, but geophysical data indicate that it could extend between 50 and 100 km offshore towards the northern margin of the Pilbara Craton. Mapping, geochronology, and aeromagnetic imagery suggest that the Harding Granitic Complex is mainly composed of the Orpheus and Maitland River Supersuites. Zircon geochronology has also revealed intrusions of Sisters Supersuite age (GSWA 142430, Nelson, 1999a). Additionally, 3270–3260 Ma granitic rocks correlated with the Karratha Granodiorite (Cleland Supersuite) outcrop on the southeast limb of the Roebourne Synform (Figs 8 and 9) (JS43, Smith et al., 1998; N3214, Smith, 2003). The Karratha Granodiorite south of Karratha has been interpreted as a northwest continuation of the Harding Granitic Complex, and that granites link the two units beneath the Roebourne Synform. Nd T_{DM} model ages from the Karratha Granodiorite range from 3520 Ma (age recalculated by Smithies et al. [2007a] from data of Smith et al. [1998]) to 3380 Ma, suggesting that parts of the Harding Granitic Complex intruded Paleoarchean crust north of the SSZ. Xenocrystic zircon grains from the Karratha Granodiorite have $^{207}Pb/^{206}Pb$ ages between 3332 and 3287 Ma (Smith et al., 1998).

Caines Well Granitic Complex

Prior to SHRIMP U–Pb zircon geochronology, the Caines Well Granitic Complex was informally referred to as the 'Caines Well Granite' (Fitton et al., 1975; Hickman, 1983). Fitton et al. (1975) interpreted the elongate domal structure of the complex and the overlying volcanics of the Whim Creek Group as evidence that its structure formed by diapirism, as had been proposed for several granitic complexes of the east Pilbara (Hickman and Lipple, 1975). However, whereas the east Pilbara granitic domes are steep sided with well-developed radial stretching lineations, the domal structure of the Caines Well Granitic Complex has gently dipping limbs that lack radial lineations. The northeast–southwest elongate dome of the Caines Well Granitic Complex was named by Smithies (1997) as the Whim Creek Anticline, and Krapez and Eisenlohr (1998) described the fold as a non-cylindrical antiform. Following structural interpretation of the Prinsep Dome farther west (Hickman, 2001b), the Whim Creek Anticline (Fig. 8) was interpreted to be one of several northeast-trending transpositional folds oblique to the strike of the SSZ. These folds developed during c. 2940 Ma northwest–southeast compression of the North Pilbara Orogeny (Van Kranendonk et al., 2006).

The Caines Well Granitic Complex contains granitic rocks of the Railway, Orpheus, Maitland River, and Sisters Supersuites. Exposure of the complex is extremely poor and interpretation of its geology largely depends on a few isolated outcrops combined with aeromagnetic imagery and SHRIMP dating. Towards the western end of the complex, between Mount Fisher and Red Hill, older granitic components of the complex are unconformably overlain by the 3015–2990 Ma Whim Creek Group and intrude the 3130–3110 Ma Whundo Group. Although no dating of the granitic rocks was undertaken along this southwest margin of the Caines Well Granitic Complex, the unconformity indicates the presence of either the Railway or Orpheus Supersuite. Farther north, in the centre of the complex, banded granodiorite–monzogranite gneiss contains zircon age groups dated at 3111 ± 13 Ma and 3093 ± 4 Ma (GSWA 118965, Nelson, 1997b). Mapping and geophysical data indicate that this gneiss is representative of much of the western section of the Caines Well Granitic Complex, supporting a correlation with the Railway Supersuite. The existence of the Orpheus Supersuite beneath or adjacent to the Whim Creek Group at Red Hill is indicated by 3019–3013 Ma inherited zircons in the Red Hill Volcanics (GSWA 141936, Nelson, 1998g). Nd T_{DM} model ages from the Caines Well Granitic Complex range from 3350 to 3150 Ma (Sun and Hickman, 1998; Smithies et al., 2007a). Nd T_{DM} model ages from basalts of the Whim Creek and Bookingarra Groups (Bookingarra Group redefined in Appendix 2), unconformably overlying the complex, range from 3490 to 3270 Ma (Arndt et al., 2001; Smithies et al., 2004a, 2007a). These model ages were obtained from 37 individual samples and provide reasonably good evidence of Paleoarchean to Early Mesoarchean crust between the SSZ and the LF (Fig. 4).

The eastern section of the Caines Well Granitic Complex is mainly composed of massive to weakly foliated granitic rocks of the Maitland River Supersuite. A sample of porphyritic monzogranite collected 5 km northwest from Whim Creek was dated at 2990 ± 5 Ma (GSWA 142950, Nelson, 2000i). Two xenocrystic zircon grains in this sample suggest minor inheritance from the Orpheus Supersuite. The distribution of the Maitland River Supersuite within the Caines Well Granitic Complex appears to coincide with the thickest development of the Red Hill Volcanics and related porphyritic dacite intrusions in the Whim Creek Group.

A very minor component of the complex is leucocratic monzogranite along the southeast margin of the complex. This rock, dated at 2925 ± 4 Ma (GSWA 118964, Nelson, 1997a), outcrops adjacent to the Sherlock Intrusion and is interpreted to be late granite of the Sisters Supersuite.

Portree Granitic Complex

East of the Whim Creek greenstone belt, and between the SSZ and Mallina Shear Zone, the Mallina Formation was intruded by a series of discrete, nested plutons collectively referred to as the Portree Granitic Complex (Smithies, 1999). The c.1000 km^2 complex (Fig. 7) is interpreted to contain at least six intrusions (Smithies and Hickman, 2003) but only two are exposed. The exposed rocks are high-temperature alkaline granites interpreted to have been derived from partial melting of a metasomatized basalt crust (Smithies and Champion, 2000). Some granitic rocks of the Cherratta Granitic Complex also show alkaline affinity but the Portree granites show Gallium enrichment which is a feature of A-type magmas (Smithies and Champion, 1998).

Pippingarra Granitic Complex

The Pippingarra Granitic Complex underlies an area of c. 2500 km^2 southeast of Port Hedland (Fig. 7). Evidence from geological mapping and geochronology indicates that it is entirely composed of granitic intrusions of the Sisters and Split Rock Supersuites. This interpretation is consistent with intrusive contacts against the Mallina Formation in the south, whereas in the west it intrudes concealed metamorphosed volcanic rocks of the Bullen Hill greenstone belt. Along its eastern margin the granitic complex intrudes the Gorge Creek Group in the Ord Range, whereas its southeast margin is bounded by sheared rocks within the Tabba Tabba Shear Zone (TTSZ). Interpretation of aeromagnetic data indicate that the northern limit of the complex is the SSZ, and that beyond that the coastal section of the Pilbara Craton is underlain by the totally concealed 'Ridley Granitic Complex' (new name).

During the early stages of the North Pilbara National Geoscience Mapping Accord (NPNGMA) the Pippingarra Granitic Complex was included in the EPT (Van Kranendonk et al., 2002). At that time, it was thought that the complex might include gneissic rocks representing part of the basement to the Mallina Basin. Rocks with the appearance of deformed banded granitic gneiss had been recorded near Mallindra Well within the southwest part of the Pippingarra Granitic Complex (Hickman, 1983). However, Smithies et al. (2002) reinterpreted the compositional layering and fabric of the rocks in this area to be a late-stage feature of granitic intrusion.

Geochronology on the Pippingarra Granitic Complex has revealed no granites older than the Sisters Supersuite, although xenocrystic zircon is rarely up to 300 million years older than the host rock crystallization age (GSWA 160727, Nelson, 2001c). A key feature of the Pippingarra Granitic Complex is that it includes several high-Mg diorite intrusions. These are restricted to the Mallina Basin and this feature, along with an absence of granites older than c. 2955 Ma, distinguish this granitic complex from those of the EPT.

Ridley Granitic Complex

Geophysical data indicate that the coastal plain from Port Hedland to Ridley River is underlain by granitic rocks that are here assigned to the Ridley Granitic Complex (Fig. 7). West from Port Hedland the complex is in faulted contact with the Peawah Hill greenstone belt, whereas in the south it is separated from the Pippingarra Granitic Complex by a fault interpreted to be the eastern extension of the SSZ. North of the Ord Range the complex is unconformably overlain by Phanerozoic rocks of the Canning Basin.

West Pilbara Superterrane

The oldest major tectonic unit of the northwest Pilbara Craton is the West Pilbara Superterrane (WPS) (Fig. 3). This is composed of the 3280–3236 Ma Karratha Terrane (KT), 3220–3160 Ma Nickol River Basin, 3200–3160 Ma Regal Terrane (RT), 3130–3093 Ma Sholl Terrane (ST), and the northwest Pilbara section of the 3068–3066 Ma Elizabeth Hill Supersuite (Fig. 6). The 3228–3160 Ma Soanesville Basin of the east Pilbara is laterally equivalent to the Nickol River Basin, but the Soanesville Basin is not part of the WPS because it is interpreted to be restricted to the basement that underlies the southeast margin of the Mallina Basin.

The KT (Figs 3, 5 and 6), comprising the Roebourne Group and the Cleland Supersuite, represented in the northwest Pilbara by the Karratha Granodiorite and Tarlwa Pool Tonalite, is interpreted to be a rifted fragment of the 3530–3223 Ma EPT (Sun and Hickman, 1998; Van Kranendonk et al., 2006). Unconformably overlying the KT, the Nickol River Basin (Hickman, 2012) is composed of the Nickol River Formation (Figs 3, 5, 6), interpreted to be a passive margin clastic succession deposited during rifting of the EPT. The RT is mainly composed of the Regal Formation, a thick basaltic succession, interpreted to be juvenile crust from the c. 3200 Ma Central Pilbara Basin (CPB) that was obducted across the Nickol River Basin and KT between 3160 and 3066 Ma. In the Cleaverville area (Fig. 8), Kiyokawa et al. (2012) named the part of the Regal Formation that overlies the Dixon Island Formation (Table 1) as the Port Robinson Basalt. Although stratigraphic relations at Cleaverville are complicated by shearing, the present interpretation is that the Dixon Island Formation, dated at c. 3195 Ma (Kiyokawa et al., 2002), can be correlated with the upper part of the Nickol River Formation.

The 3130–3093 Ma ST (Figs 3 and 6) was interpreted to be an oceanic island arc succession by Smithies et al. (2005a, 2007c), whereas Krapez and Eisenlohr (1998), Smith et al. (1998), and Smith (2003) interpreted a back-arc setting southeast of a magmatic arc. A back-arc setting for the Whundo Group was questioned by Hickman (2004a) because no evidence of a c. 3130–3093 Ma magmatic arc is preserved northwest of the ST. Smithies et al. (2005a) opposed the back-arc interpretation on geochemical grounds (see Whundo Group). In the intra-oceanic arc scenario (Smithies et al., 2007c) the ST formed along a 3130–3093 Ma subduction zone within oceanic crust. However, an oceanic plate of Phanerozoic proportions was not formed by separation of the EPT and KT. The present interpretation is that the basaltic crust on which the Whundo Group developed was chemically ‘oceanic-like’ but actually lay in a relatively narrow rift basin (Central Pilbara Basin) between the EPT and KT (Hickman, 2001b). This interpretation is supported by the preservation of c. 3260 Ma crust immediately northwest of the SSZ and immediately southeast of the Tabba Tabba Shear Zone (TTSZ) because these shear zones represent the main terrane separation zone at c. 3220 Ma.

At approximately 3070 Ma, north–south convergence between the EPT and KT caused the KT, RT, and ST to collide and become accreted forming the WPS. The collision resulted in the Prinsep Orogeny (Van Kranendonk et al., 2006) with associated granitic intrusion of the 3068–3066 Ma Elizabeth Hill Supersuite.

Karratha Terrane

The oldest rocks in the northwest Pilbara Craton are exposed in the Roebourne greenstone belt (Fig. 7). This greenstone belt is part of the Karratha Terrane (KT) (Figs 3 and 5), which is a Paleoarchean tectonic unit restricted to the coastal area north of the SSZ. The total southwest–northeast onshore extent of the terrane is interpreted to be approximately 450 km, but over most of this length it is either concealed by coastal regolith or obscured by Mesoarchean granitic intrusions. The most easterly exposures of the KT are at Mount Wangee, northeast of Roebourne, where the Roebourne Group is intruded by younger granites (concealed). About 30 km farther east, airborne magnetic data indicate the existence of a significant northeast-trending shear zone, branching off the SSZ in the area of Terenar Pool west of Depuch Island. This entirely concealed structure is interpreted to be a 2 km-wide shear zone and to contain linear belts of strongly attenuated greenstone units, most likely parts of the Roebourne Group. It is possible that this structure, named on Figures 3–5 as the Terenar Shear Zone (TSZ), marks the eastern limit of the KT, but without more geological evidence this is not the presently preferred interpretation.

In the main outcrop area, between Roebourne and Mount Regal, the KT comprises volcanic formations of the c. 3280 Ma Roebourne Group and the intrusive 3270–3261 Ma Karratha Granodiorite (Fig. 9).

Unconformably overlying the KT are clastic sedimentary rocks of the 3220–3160 Ma Nickol River Formation. In the first stratigraphic revision of the Roebourne greenstone belt (Hickman, 1997a), this formation was included in the Roebourne Group, but following recognition of a basal unconformity (Kiyokawa et al., 2002) it was linked to a c. 3220 Ma rifting event in the evolution of the craton (Hickman et al., 2010; Van Kranendonk et al., 2010; Hickman, 2012).

Tectonic setting

Correct interpretation of the depositional environment of the Roebourne Group is essential to understand the crustal evolution of the KT and its relationship to other stratigraphic units of the Roebourne greenstone belt. A number of previous workers have interpreted all or parts of the Roebourne greenstone belt to be Mesoarchean oceanic crust accreted onto the northwest margin of the Pilbara Craton (Isozaki et al., 1991; Kiyokawa and Taira, 1998; Kiyokawa, 1993; Krapez, 1993; Barley et al., 1994; Kiyokawa et al., 2002; Ohta et al., 1996; Barley, 1997; Kato et al., 1998; Shibuya et al., 2007). However, the interpretation in this Report is that the only formation of this greenstone belt to be composed of oceanic-like basaltic crust is the Regal Formation (Hickman, 1999, 2001a,b, 2004a, 2012; Van Kranendonk et al., 2002, 2006; Hickman et al., 2006a, 2010). Differences of interpretation are partly explained by earliest research being concentrated on the Regal Formation at Cleaverville. At this locality, the Regal Formation does lithologically and chemically resemble mid-ocean ridge basalt (MORB) (Ohta et al., 1996; Sun and Hickman, 1999; Smithies et al., 2007a). However, this ophiolite interpretation cannot be extended to all stratigraphic units of the Roebourne greenstone belt (Kiyokawa, 1993; Kiyokawa et al., 2002) because stratigraphy, geochemistry, and geochronology provide evidence to the contrary. For example, the basaltic succession of the c. 3280 Ma Roebourne Group does not have an oceanic-like composition and several lines of evidence indicate that it was deposited in a very different tectonic setting:

Stratigraphic evidence

- The Roebourne Group is intruded by the 3270–3261 Ma Karratha Granodiorite. Although small volumes of contemporaneous granitic rocks very rarely are located in mid-ocean settings (e.g. Iceland), isotopic evidence indicates that the Karratha Granodiorite was derived from partial melting of c. 3500 Ma crust or enriched lithospheric mantle (Hickman et al., 2010). Additionally, the Karratha Granodiorite is a regionally voluminous unit compared to the Roebourne Group.

- The Ruth Well Formation, in the lower part of the Roebourne Group, contains a 1000 m-thick succession of komatiite flows. Most workers interpret komatiites to be products of deep, high-temperature mantle melting (Arndt et al., 1998), commonly where hot mantle plumes underlie volcanic plateaux (Van Kranendonk et al., 2015). Komatiites are absent from Phanerozoic oceanic crust that was formed at spreading centres.

- The Roebourne Group is overlain across a c. 3220 Ma regional erosional unconformity by metamorphosed boulder conglomerate, pebbly sandstone, quartz-sandstone, shale, and carbonate sedimentary rocks of the Nickol River Formation. Unconformities of this age are present across the northern Pilbara Craton and mark erosion of the craton during the early stages of rifting and continental breakup (East Pilbara Terrane Rifting Event; Van Kranendonk et al., 2010; Hickman, 2012).

Trace element composition

- The trace element compositions of tholeiitic basalt, komatiitic basalt, and komatiite in the Ruth Well Formation of the Roebourne Group (Fig. 11a) suggest significant interaction with continental crust. Concentrations of Th in the basalts are up to 42 times primitive mantle values and the light rare earth elements (LREE) are strongly fractionated, with La at least 20 times primitive mantle values. La/Nb, La/Sm, and La/Yb ratios are high and negative Nb anomalies are strongly developed. Enrichments in Th, U, and LREE clearly distinguish these basalts from those of the Regal Formation (Fig. 11b), in which tholeiite has a chemical composition similar to that of MORB (Ohta et al., 1996; Sun and Hickman, 1999; Smithies et al., 2007a).

- The trace element composition of the Ruth Well Formation shows greater LREE enrichment than for any of the Paleoarchean basaltic formations in the east Pilbara (data in Smithies et al., 2007a), all of which were deposited on continental crust (Smithies et al., 2003, 2007b; Hickman, 2004a, 2012; Van Kranendonk et al., 2007a,b, 2015).

Sm–Nd data

- Nd model ages on samples of komatiite, komatiitic basalt, and tholeiitic basalt of the Ruth Well Formation (Arndt et al., 2001; Smithies et al., 2007a; Tessalina, written comm., 2012) vary between c. 3480 Ma and c. 3400 Ma (Tables 3 and 4). These results are inconsistent with the Ruth Well Formation being juvenile oceanic crust, but instead suggest melting of, or contamination by, much older crust or lithospheric mantle.

- ε_{Nd} values between –1.92 and 0.88 (Table 4) suggest magma derivation from older continental crust rather than direct derivation from the mantle.

In summary, there is compelling evidence that the Roebourne Group was deposited on continental crust. This is interpreted to have formed part of the Paleoarchean Pilbara Craton (EPT) before c. 3220 Ma crustal extension, rifting, and breakup. The intrusive Karratha Granodiorite was part of the same Paleoarchean crust and was derived from partial melting of much older (c. 3500 Ma) crust. In combination with the interpretation in this Report that the KT underlies the entire area of the Roebourne greenstone belt (see Roebourne greenstone belt) this evidence indicates that north of the SSZ, the northwest Pilbara Craton is predominantly composed of continental

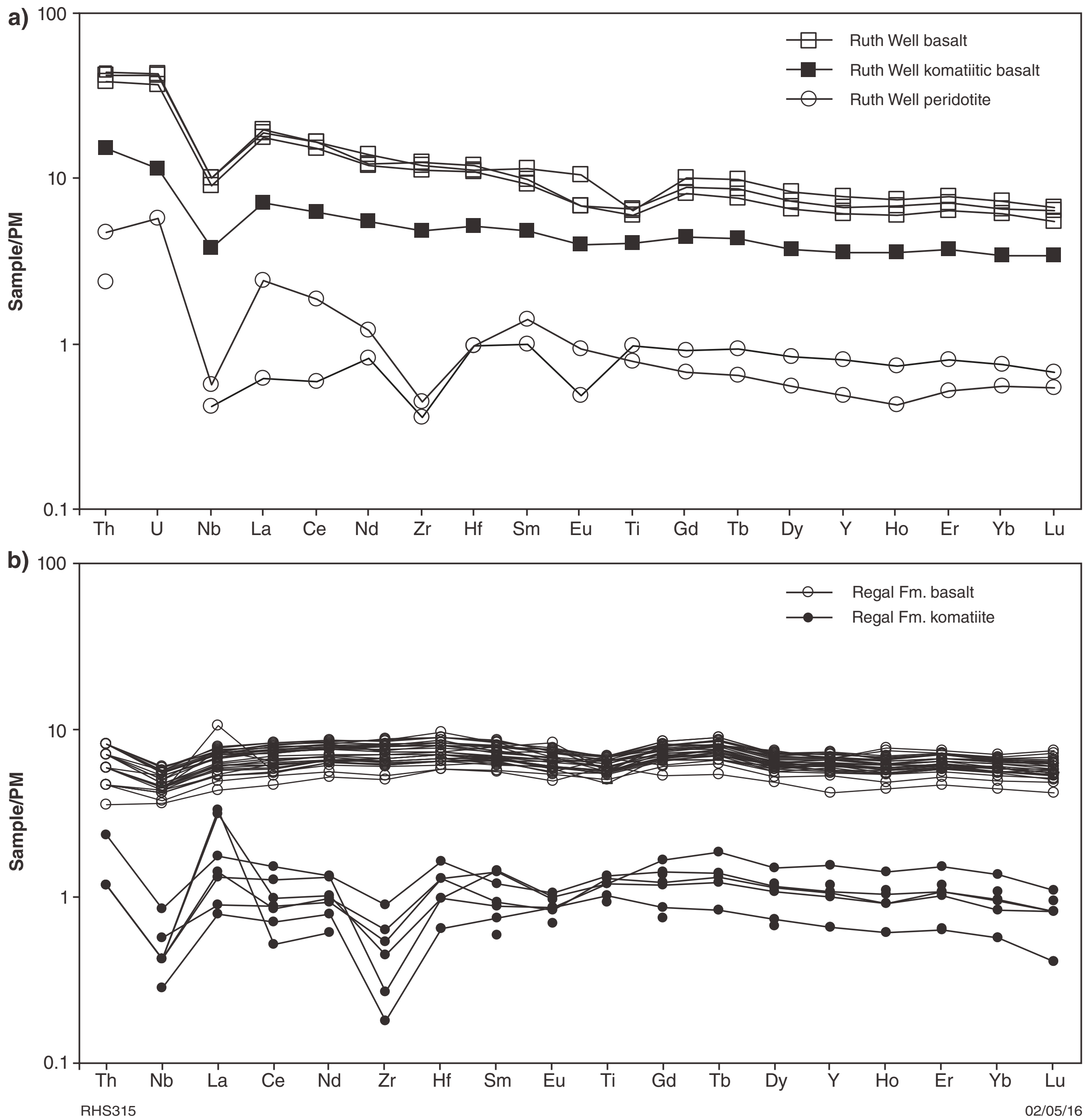


Figure 11. Trace element plots, normalized to primitive mantle, for basalts of the Ruth Well Formation and Regal Formation (after Smithies et al., 2007a). Normalizing factors after Sun and McDonough (1989).

crust. The possibility that the tectonically emplaced Regal Formation (RT) originated as oceanic crust is discussed later in the Report.

Roebourne Group

The Roebourne Group forms much of the Roebourne greenstone belt between Roebourne and the Maitland River (Fig. 7). First defined to comprise the Ruth Well Formation and the Nickol River Formation (Hickman, 1997a), the group is now redefined to comprise only the Ruth Well Formation and the 'Weerianna Basalt' (new name, Appendix 2). The Nickol River Formation is now known to unconformably overlie both the Roebourne Group and the Karratha Granodiorite (see below). The Ruth Well Formation is a mixed assemblage of komatiite, komatiitic basalt, tholeiite, chert, and sills of peridotite and dolerite whereas the Weerianna Basalt is composed of only tholeiitic and komatiitic basalt. Most of the group has been metamorphosed to lower amphibolite facies.

Field relations and geochronological data have established that the ultramafic and mafic volcanic rocks of the Ruth Well Formation are the oldest preserved components of the group. The absence of felsic volcanic rocks has precluded dating their depositional age by the SHRIMP zircon U–Pb method. However, the Ruth Well Formation is intruded by the Karratha Granodiorite, samples of which have been dated between 3270 Ma and 3261 Ma (GSWA 142433, Nelson, 1998j; Smith et al., 1998). Nd T_{DM} model ages obtained on samples of the Ruth Well Formation range from 3480 Ma to 3400 Ma (Tables 3 and 4). These data indicate mantle extraction ages of the sources for the Ruth Well Formation similar to those established for volcanic formations in the EPT (Smithies et al., 2007a).

Ruth Well Formation

The Ruth Well Formation is a c. 1500 m-thick volcanic succession composed of metamorphosed peridotitic komatiite (variably serpentized), talc–chlorite schist, metabasalts (locally pillowed), and thin (<10 m) units of metamorphosed bedded chert. No stratigraphic base is preserved owing either to intrusion by granitic rocks or to tectonic contacts with other units. However, quartzite and iron-rich chert in the Ruth Well Formation at Mount Wangee (Hickman, 2000) may lie close to the stratigraphic base. The quartzite is finely granular, pale grey to cream, and laminated. No clasts or sedimentary structures were observed during mapping and it is likely that the quartzite and chert are tectonically attenuated. The base of the formation is not locally exposed owing to a coastal lagoon immediately south of Mount Wangee. Rock mapped as quartzite overlies komatiite of the Ruth Well Formation at Mount Hall, 6 km east of Roebourne (Hickman, 2000). The quartzite is fine to medium grained and banded, but as at Mount Wangee 7 km to the north, it is possible that the rock is recrystallized mylonite.

Chert units within the formation are variously black, green, or grey and white banded (Fig. 12). Many of them are silicified interflow sedimentary rocks that were deposited as carbonaceous shale (black chert, and grey and white banded chert) or fine-grained volcaniclastic sedimentary rock. Green chert is typically located in sections containing ultramafic lava or intrusive ultramafic rock and the colour of the chert is attributed to the presence of finely disseminated chromian muscovite (fuchsite). Chert outcropping on the summit of Mount Wangee is 80 m thick and includes iron-rich chert and minor banded grey white chert. Chert outcrops 500 m east from the summit of Mount Wangee include red–white banded chert and brecciated iron-rich chert. It is possible that some of the chert at Mount Wangee is recrystallized mylonite. Even so, the quartzite and chert units in the vicinity of Mount Wangee and Mount Hall are considered to provide good opportunities for petrographic studies and possibly zircon U–Pb geochronology. If detrital zircon can be identified, dating should provide constraints on the maximum depositional age and provenance of the Ruth Well Formation.

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Figure 12. Banded grey white chert in the Ruth Well Formation west of the Karratha Granodiorite. The chert is interpreted to be a unit of silicified carbonaceous shale between flows of komatiite and basalt (Zone 50, MGA 476560E 7696400N).

The Ruth Well Formation is tectonically disrupted by faults related to the overlying Regal Thrust and, in the southern part of the Roebourne greenstone belt, it is faulted close to the SSZ (Fig. 9). To the south of Roebourne, the Ruth Well Formation is extensively intruded by 3270–2930 Ma granitic rocks and by mafic and felsic intrusions of the Orpheus Supersuite. The most complete sections through the formation are preserved between Mount Regal and Ruth Well, but intrusion of the c. 3270 Ma Karratha Granodiorite has removed the lower stratigraphy.

Diamond drilling during exploration for nickel mineralization in c. 1970 revealed that the ultramafic rocks of the Ruth Well area consist of numerous flows of komatiite, individually up to 40 m thick, and that these locally make up a 500–1000 m-thick succession (Nisbet and Chinner, 1981). Most flows contain a central zone with olivine spinifex texture that is underlain and overlain by chilled margins (Tomich, 1974). Excellent exposures showing olivine spinifex texture between aphanitic flow tops and bases are available at Mount Hall, 6 km east of Roebourne (Fig. 13). This locality shows well-preserved sheaf and random spinifex textures and individual olivine plates (pseudomorphed by serpentine and tremolite) are up to 30 cm long. Layers with different spinifex texture indicate that the ultramafic flows are up to 2 m thick, with random spinifex texture overlying sheaf spinifex. Near Tom Well, 4 km south of Karratha, a notable feature of the Ruth Well Formation is the interlayering of komatiite and komatiitic basalt units (Hickman, 2001b). Nisbet and Chinner (1981) attributed the same bimodal composition at Ruth Well to derivation of the basalt from komatiite in transient magma chambers.

Weerianna Basalt

The type area of the Weerianna Basalt (new name, first used in this Report and defined in Appendix 2) is immediately west of Roebourne at Weerianna Hill, where metamorphosed basalt, including fine-grained komatiitic basalt, is intruded by irregular sills and dykes of gabbro and dolerite, most of which are interpreted to belong to the Andover Intrusion. In areas near intrusive contacts the metabasalt appears as xenoliths within the gabbro. Brecciation and gabbroic veining of the basalt is most common in the area immediately southwest of Roebourne (Fig. 9). In this area the stratigraphic thickness of the Weerianna Basalt is up to 1000 m and the formation outcrops over a strike length of 20 km. Pillow structures are more common than in the Ruth Well Formation, for example 2.5 km east of Carlow Castle (Zone 50, MGA 509330 E 7698380 N). Flow top breccias are exposed west of Roebourne, approximately 1.5 km south of the Weerianna gold mine. The Weerianna Basalt conformably overlies the Ruth Well Formation and is distinguished by its lack of peridotitic komatiite and thin chert units. Chert is absent except close to the top of the formation southeast and south of the Weerianna gold mine. The upper contact of the Weerianna Basalt in the Roebourne area is marked by intensely sheared sedimentary rocks of the Nickol River Formation within the Regal Thrust.

The metamorphic grade of the formation varies between greenschist and lower amphibolite facies. In thin sections, komatiitic basalt consists of fibrous-textured amphibole intergrown with epidote, chlorite, and elongate randomly oriented plagioclase crystals (largely replaced by epidote). The amphibole is pale green actinolite and shows remnant spinifex texture and finer grained feathery spinifex texture.

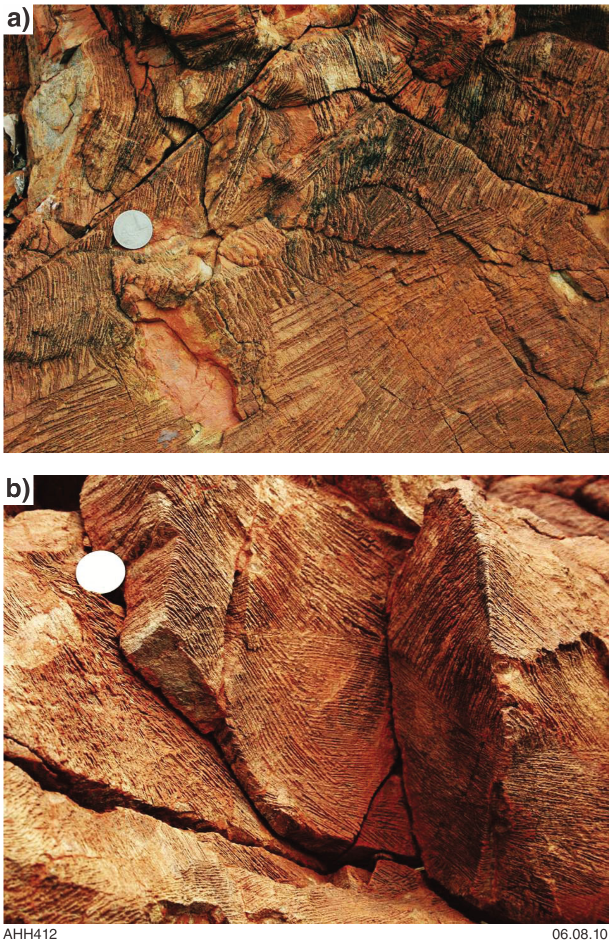

Figure 13. Bladed olivine-spinifex texture in a komatiite flow of the Ruth Well Formation at Mount Hall: a) platy olivine crystals, pseudomorphed by serpentine, tremolite, and chlorite, up to 50 cm long; b) close-up of interlocking olivine crystals. Scale in both images: 3 cm-diameter coin (Zone 50, MGA 520770E 7701337N)

Cleland Supersuite

Karratha Granodiorite

The 3270–3260 Ma Karratha Granodiorite varies from tonalite to granodiorite (GSWA 142433, Nelson, 1998j; Smith et al., 1998; Smith, 2003). Much older Nd T_{DM} model ages of 3520–3380 Ma from this intrusion (Tables 3 and 4) indicate that magma generation involved older crust or enriched lithospheric mantle, as for the 3274–3223 Ma Cleland Supersuite in the EPT (data in Smithies et al., 2007a). A sample of the granodiorite (JS17) dated by Smith et al. (1998) contained six xenocrystic zircon grains with $^{207}Pb/^{206}Pb$ ages between 3332 and 3287 Ma. This closely coincides with the age range of the Emu Pool Supersuite in the EPT, and it is notable that the granodiorite contains xenoliths of gneiss of unknown age (Fig. 14).

The type area of the Karratha Granodiorite is immediately south of Karratha but geochronology (JS43, Smith et al., 1998; N3214, Smith 2003) supports structural evidence from the Roebourne greenstone belt (Hickman, 2001a) that the intrusion extends into the Harding Granitic Complex on the southern limb of the Roebourne Synform (Fig. 9). The Karratha Granodiorite is therefore represented in both the KT and in the Harding Granitic Complex. Thus, the Harding Granitic Complex, which extends east almost to Port Hedland (Fig. 7), includes Paleoarchean crust. The Karratha Granodiorite intrudes the Roebourne Group but has no intrusive contacts with the Nickol River Formation.

Table 3. Published geochronological data indicating Paleoarchean crust >3350 Ma in the northwest Pilbara Craton. Zircon $^{207}Pb/^{206}Pb$ ages were obtained on detrital grains.

Age (Ma)	*Method*	*Material*	*Rock unit, age (Ma)*	*Sample (Ref)*
3494 ± 15	Nd model	Granodiorite	Karratha Granodiorite, c. 3261	JS17 [a]
3480–3420	Nd model	Komatiite	Ruth Well Formation, c. 3280	n = 2 [b]
3480–3420	Nd model	Granodiorite	Karratha Granodiorite, c. 3270	n = 3 [c]
3479 ± 13	Nd model	Granodiorite	Karratha Granodiorite, c. 3265	JS43 [a]
3461 ± 8	$^{207}Pb/^{206}Pb$	Zircon*	Cleaverville Formation, c. 3022	127330 [d]
3449 ± 5	$^{207}Pb/^{206}Pb$	Zircon*	Nallana Formation, c. 3125	114350 [e]
c. 3430	Nd model	Rhyolite	Nickol River Formation, <3251	118975 [f]
3430–3330	Nd model	Basalt	Mount Negri Volcanics, c. 2950	n = 3 [b]
3410–3340	Nd model	High-Mg basalt	Louden Volcanics, c. 2950	n = 14 [b]
3391 ± 15	Nd model	Granodiorite	Snell Well Granodiorite, c. 3024	JS25 [a]
c. 3380	Nd model	Basalt	Comstock Member, c. 2955	97045027A [c]
c. 3370	Nd model	Basalt	Warambie Basalt, c. 3010	78 [b]
c. 3368	Nd model	High-Mg basalt	Louden Volcanics, c. 2950	331/338 [f]
c. 3360	Nd model	Granite	Miaree Granite, c. 2997	136844 [c]
3360–3310	Nd model	Basalt	Mount Negri Volcanics, c. 2950	n = 4 [g]
c. 3360	Nd model	Tonalite	Tarlwa Pool Tonalite, c. 3236	142535 [c]
c. 3350	Nd model	Monzogranite	Caines Monzogranite, c. 2990	160502 [c]

NOTES: n – number of samples; * – detrital zircon in sandstone

REFERENCES: (a) Smith et al. (1998); (b) isotopic data from Arndt et al. (2001), model ages calculated by Smithies et al. (2007a); (c) Smithies et al. (2007a); (d) detrital zircon, Nelson (1998b); (e) xenocrystic zircon, Nelson (1996); (f) Sun and Hickman (1998); (g) Smithies et al. (2004a)

Tarlwa Pool Tonalite

The Karratha Granodiorite is restricted to the area north of the SSZ but another intrusion assigned to the Cleland Supersuite (Van Kranendonk et al., 2006) is exposed on the Fortescue River 40 km southwest of Karratha. The c. 3236 Ma Tarlwa Pool Tonalite (GSWA 142535, Nelson, 1998l) outcrops in several inliers surrounded by the unconformably overlying Fortescue Group. Gravity data indicate that the tonalite forms the northern part of a granitic complex distinct from and immediately to the south of the Cherratta Granitic Complex. The pronounced gravity low, which indicates the existence of this almost entirely concealed granitic area coincides with the Mingar Dome (Williams, 1968), which is about a 70 km diameter low-amplitude structure in the Fortescue Group. This concealed granitic unit, which occupies an area of approximately 3000 km^2 beneath the Fortescue Group, is here named as the 'Mingar Granitic Complex'. The depositional age of the deformed stratigraphy establishes that domal uplift included post-2500 Ma movement, but the dome of the Mingar Granitic Complex probably originated as a Paleoarchean structure. Using the geophysical interpretation of Blewett et al. (2000) the Mingar Granitic Complex is separated from the Cherratta Granitic Complex by a narrow greenstone belt.

Correlations with the East Pilbara Terrane

Sun and Hickman (1998) suggested that the Roebourne Group was originally linked to the Sulphur Springs Group, which is now best exposed around the Strelley Monzogranite in the EPT. This correlation, in combination with the present location of the group in the KT, might be explained by major west-southwest sinistral strike-slip displacement along the SSZ. Alternatively, because both the Sulphur Springs Group and the Cleland Supersuite are exposed along the central section of the TTSZ, the present geographic separation of the outcrops might be substantially due to northwest-southeast separation during 3220–3160 Ma rifting. In either scenario the KT is a fragment of the Paleoarchean EPT.

The 3270–3261 Ma Karratha Granodiorite is correlated with the Cleland Supersuite of the EPT (Van Kranendonk et al., 2006), which includes c. 3274 Ma granitic rocks in the Yilgalong Granitic Complex. The Roebourne Group is older than 3270 Ma because it is intruded by the Karratha Granodiorite, and unpublished Sm–Nd data suggest a possible depositional age c. 3280 Ma. If the Roebourne Group was deposited between 3280 and 3270 Ma it is approximately the same age as the lower part of the Sulphur Springs Group (Table 2). According to Buick et al. (2002), the lowest dated stratigraphic interval in the Sulphur Springs Group is a thin felsic volcaniclastic arenite between the felsic volcanic Kangaroo Caves Formation (uppermost formation of the group) and the mafic volcanic Kunagunarrina Formation. The zircon population of the sample dated was detrital, with 19 of 20 near-concordant zircon grains dated at 3520–3410 Ma. A 95% concordant grain was dated at c. 3255 Ma, and this was interpreted to provide the maximum depositional age at this stratigraphic level. However, the maximum age of the lower part of Sulphur Springs Group is probably c. 3275 Ma. The volcanic section of the lower Sulphur Springs Group, the Kunagunarrina

Table 4. Summary of published Nd data for 3280–3060 Ma igneous rocks of the West Pilbara Superterrane

Tectonic/stratigraphic unit	*Age (Ga)*	ε_{Nd}	T_{2DM} *(Ga)*	*Data source*
North of Sholl Shear Zone				
KARRATHA TERRANE				
Ruth Well Formation (n = 2)	c. 3.28	0.88 to 0.43	3.45 to 3.42	Arndt et al. (2001)
Ruth Well Formation, 160240	c. 3.28	0.43	3.48	Smithies et al. (2007a)
Ruth Well Formation (n = 6)	c. 3.28	–0.93 to –1.92	3.47 to 3.40	S. Tessalina (written comm.)
Karratha Granodiorite (n = 2)	c. 3.27	–0.16 to –0.55	3.52 to 3.49	Smith et al. (1998)*
Karratha Granodiorite (n = 4)	c. 3.27	1.34 to 0.00	3.48 to 3.38	Smithies et al. (2007a)
NICKOL RIVER BASIN				
Mount Regal dacite, 118975	c. 3.25	0.31	3.44	Sun and Hickman (1998)
REGAL TERRANE				
Regal Formation (n = 3)	c. 3.20	3.46 to 0.56	3.38 to 3.16	Smithies et al. (2007a)
Regal Formation (n = 2)	c. 3.20	1.01 to 0.91	3.31	Champion (2013)
South of Sholl Shear Zone				
SHOLL TERRANE				
Railway Supersuite (n = 2)	c. 3.11	1.61 to 0.62	3.31 to 3.23	Smith et al. (1998)
Railway Supersuite, 142534	c. 3.11	1.63	3.22	Smithies et al. (2007a)
Whundo Group				
Woodbrook Formation, 127378	c. 3.12	1.63	3.23	Smithies et al. (2007a)
Bradley Basalt (n = 5)	c. 3.12	2.65 to 0.75	3.30 to 3.15	Smithies et al. (2007a)
Tozer Formation (n = 4)	c. 3.12	2.11 to 1.34	3.26 to 3.20	Smithies et al. (2007a)
Nallana Formation (n = 4)	c. 3.13	2.04 to 1.40	3.25 to 3.20	Smithies et al. (2007a)
MINGAR GRANITIC COMPLEX				
Tarlwa Pool Tonalite, 142535	c. 3.24	1.26	3.36	Smithies et al. (2007a)
ELIZABETH HILL SUPERSUITE				
Cliff Pool Tonalite, 142661	c. 3. 07	1.15	3.23	Smithies et al. (2007a)

NOTES: *recalculated Nd model ages by Smithies et al. (2007a)

Formation, is lithologically similar to the Roebourne Group, being a succession of komatiitic basalt and basalt up to 2.5 km thick. Both the Ruth Well Formation and the Kunagunarrina Formation include units enriched in LREE and were deposited above older Paleoarchean crust. Nd T_{DM} model ages for the Sulphur Springs Group and the related Strelley Monzogranite are c. 3500 Ma (Smithies et al., 2007a).

Nickol River Basin

The Paleoarchean–Mesoarchean Nickol River Basin unconformably overlies the KT (Hickman, 2012) and is exposed in the Roebourne and Devil Creek greenstone belts (Fig. 7). The basin is entirely composed of metasedimentary rocks assigned to the Nickol River Formation (Hickman, 1997a), and is confined to the northwest Pilbara where it outcrops over a northeast–southwest strike length of 100 km. The formation is interpreted to be one of four passive margin successions deposited in separate basins on the rifted margins of the EPT from c. 3220 Ma onwards. The three other basins formed at this time were the Soanesville, Budjan Creek,

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Figure 14. Outcrop of the Karratha Granodiorite south of Karratha. The c. 3270 Ma granodiorite contains scattered xenoliths of older granitic rocks (undated), as seen on the upper, 1 m-diameter boulder (Zone 50, MGA 480520E 7695920N).

and Coondamar basins (Hickman, 2012). Very similar lithologies to those of the Nickol River Formation are located at the base of the Fig Tree Group in the Kaapvaal Craton of southern Africa where Lowe (2013) reported 3260–3225 Ma zircons from the Mapepe Formation. Although direct stratigraphic correlation between the Nickol River and the Mapepe Formations is not necessarily implied, these formations were deposited at the same time and each mark major long-term changes from plume-related volcanism to clastic basin sedimentation in their respective cratons.

Nickol River Formation

In areas where lower sections of the Nickol River Formation are well preserved, the formation is composed of either metaconglomerate or pebbly quartzite. The metaconglomerate contains rounded to angular boulders, pebbles, and smaller clasts of chert (black, banded, and green; Figs 15 and 16) and altered volcanic rocks derived by erosion of the underlying Roebourne Group. In most outcrops the formation is strongly deformed by shearing and the tectonic elongation of smaller clasts defines a stretching lineation. Granitic clasts have not been observed, but detrital zircon grains in a quartzite form a single age group dated at 3269 ± 2 Ma (GSWA 136819, Nelson, 1998d), indicating derivation from the underlying Karratha Granodiorite (3270–3261 Ma).

The maximum depositional age of the Nickol River Formation is inferred to be c. 3220 Ma based on several lines of evidence. Firstly, it unconformably overlies the Karratha Granodiorite (Kiyokawa et al., 2002), which establishes a maximum depositional age of c. 3261 Ma. Secondly, basal quartzite 1 km east of Mount Regal is intruded by 3251–3220 Ma porphyritic rhyolite (Hickman, 2001b). The reported emplacement age of this rhyolite is 3251± 6 Ma (GSWA 118975, Nelson, 1997g), but the rock contains zircon grains of more than one group. The older grains are similar in age to the Karratha Granodiorite, whereas the $^{207}Pb/^{206}Pb$ ages of three concordant to semi-concordant zircon grains are between 3220 and 3214 Ma. Therefore, an alternative interpretation is that all the older grains are inherited and that the c. 3220 Ma grains provide the closest estimate of the crystallization age of the intrusive rhyolite. Age constraints on the three other rift-related clastic basins of the northern Pilbara (Soanesville, Budjan Creek, and Coondamar Basins) indicate maximum depositional ages between 3230 and 3190 Ma (Hickman et al., 2010; Van Kranendonk et al., 2010; Hickman, 2012). If these basins formed at approximately the same time, in response to rifting of the EPT, the Nickol River Formation is unlikely to be significantly older than c. 3220 Ma.

The Nickol River Formation was originally named by Williams (1968) who described it as a metamorphosed succession of shale, siltstone, sandstone, conglomerate, dolomite, chert, and jaspilite with locally intercalated volcanic rocks. Williams (1968) stated that the Nickol River Formation is underlain by gneiss, amphibolite, and volcanic rocks and that it is overlain by the Regal Formation (metamorphosed basaltic volcanic rocks). The underlying 'gneiss, amphibolite, and volcanic rocks' are now assigned to the Karratha Granodiorite and Roebourne Group of the KT, whereas the overlying metabasalts belong to the Regal Formation of the RT.

Upper sections of the Nickol River Formation are composed of well-bedded black and ferruginous chert, metamorphosed carbonate rocks, quartz-sericite schist, ferruginous and carbonaceous shale, and BIF. These lithologies suggest deepening of the depositional basin with increased plate separation. The minimum depositional age of the upper part of the Nickol River Formation is that of the Regal Thrust, which based on metamorphic ages commenced formation at c. 3160 Ma, and may have been active until c. 3070 Ma (Hickman, 2004a; Van Kranendonk et al., 2006; Hickman et al., 2010). A more precise age on shale of the Nickol River Formation is available if the Dixon Island Formation (Table 1) is interpreted to be laterally equivalent. The Dixon Island Formation is a unit of black carbonaceous shale, chert, and minor felsic volcanic rocks exposed at Cleaverville and on nearby Dixon Island (Kiyokawa et al., 2002, 2012, 2014). At this locality the sedimentary succession underlies pillow basalt correlated with the Regal Formation (Hickman, 2001b). Kiyokawa et al. (2002) used the U–Pb zircon method to date felsic tuffaceous rock in the Dixon Island Formation at 3195 ± 15 Ma. Kiyokawa et al. (2012) interpreted deposition to have taken place during normal faulting of the sea floor and Kiyokawa et al. (2014) described evidence for hydrothermal activity, including numerous dykes of black chert. Dykes of black chert underlying bedded chert elsewhere in the northern Pilbara Craton were formed by circulation of hydrothermal fluids in normal faults under conditions of crustal extension (Nijman et al. 1998; Van Kranendonk, 2006; Hickman, 2008). Syndepositional normal faulting was particularly widespread in the northern Pilbara between 3235 and 3165 Ma (Eriksson, 1981; Wilhelmij and Dunlop, 1984; Zegers et al., 1996; Vearncombe et al., 1998; Van Kranendonk et al., 2010).

East Pilbara Terrane Rifting Event

As described above, the KT was separated from the EPT by continental breakup between 3235 and 3200 Ma (Hickman, 2004a, 2012; Van Kranendonk et al., 2006, 2010). The unconformable contact between the Sulphur Springs Group and the 3228–3176 Ma Soanesville Group in the EPT marks an abrupt change from the Paleoarchean magmatic cycles and gravity-driven doming to Mesoarchean plate tectonic settings. The equivalent unconformity in the northwest Pilbara is between the KT and the Nickol River Formation. In both terranes the unconformities separate thick Paleoarchean volcanic successions from Mesoarchean sandstone, conglomerate, shale, and BIF deposited on the passive margins of the separating terranes. The north–south or northwest–southeast crustal extension resulted in rifting, hydrothermal activity, and eventually basaltic volcanism and ultramafic–mafic intrusion. The original c. 3200 Ma rift structures in the northwest Pilbara have been largely obscured by granitic intrusion, although Kiyokawa et al. (2012) documented syndepositional rifting in the c. 3195 Ma Dixon Island Formation.

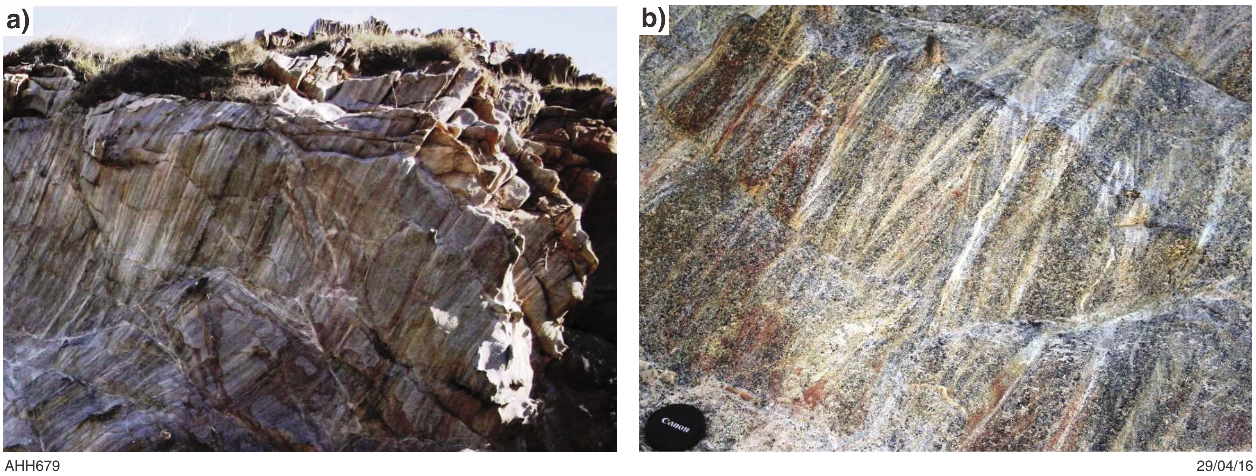

Figure 15. Breakaway outcrop of metaconglomerate and metasandstone in the Nickol River Formation south of Port Robinson (Zone 50, MGA 502725E 7713758N): a) steeply dipping beds of cross-bedded metasandstone are exposed in a 4 m-high cliff; b) close-up of cross-bedding indicating variable paleocurrent directions. Most of the dark grains in the metasandstone are fragments of black chert (scale: 5 cm-diameter lens cap).

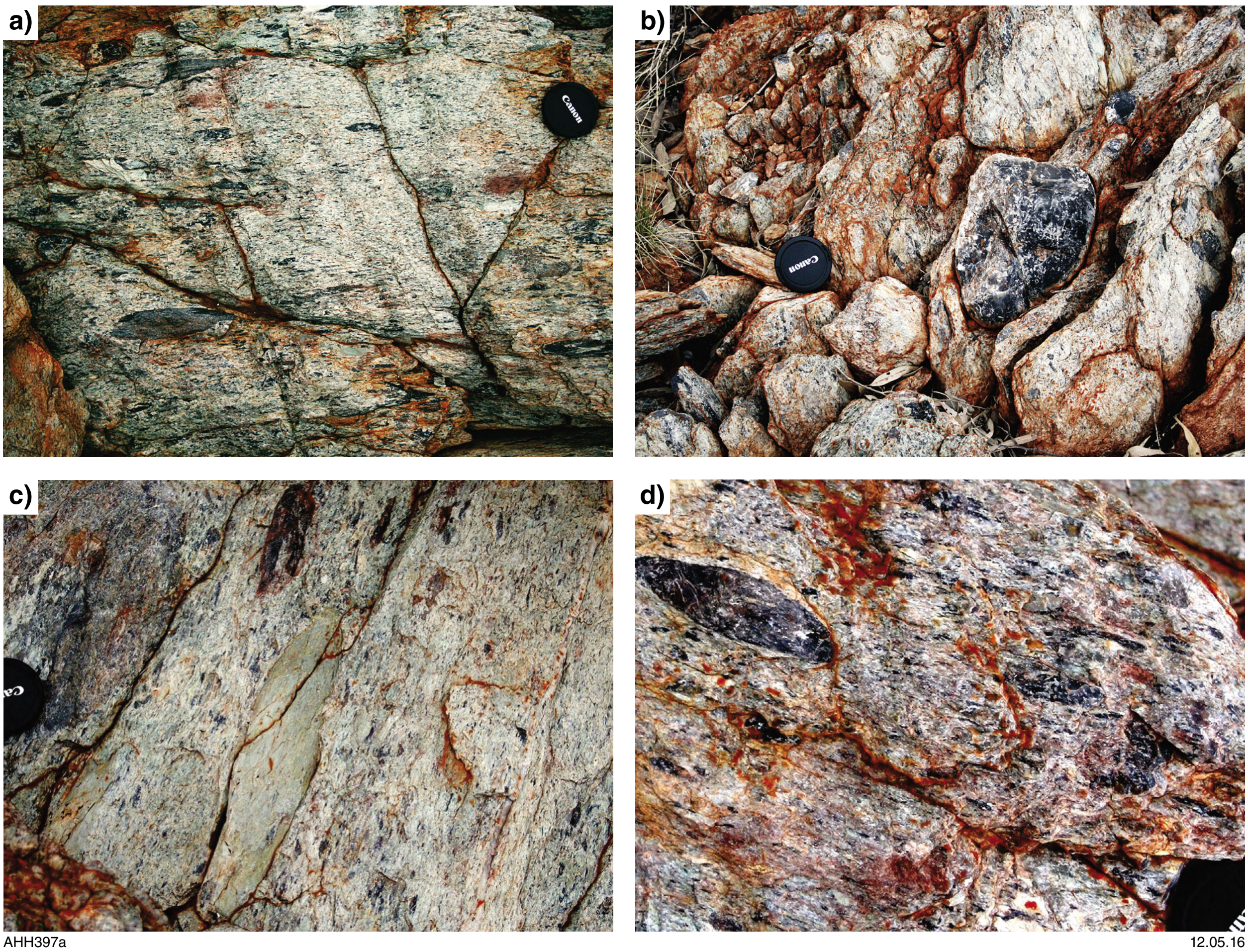

Figure 16. Outcrop of strongly sheared matrix-supported conglomerate in the Nickol River Formation 2 km east of Lydia gold mine (Zone 50, MGA 501505E 7707865N): a) acutely stretched clasts of grey chert and pelitic schist within mylonitized sandstone; b) small cobble of grey chert, largely undeformed, within sheared and lineated metasandstone matrix; c) stretched boulders of grey chert and fine-grained metasedimentary rock within strongly sheared metasandstone; d) numerous clasts of grey chert define a strongly oriented lineation in the metasandstone. Scale in a) to c): 5 cm-diameter lens cap

The SSZ and TTSZ, now located on opposite sides of the Mallina Basin, are interpreted to have first developed as two of the 3235–3200 Ma extensional faults. The only structural evidence of movement now preserved on these major shear zones is younger than 3200 Ma, and is either strike-slip movement related to 3070–3050 Ma terrane convergence or strike-slip associated with oblique convergence during the 2955–2920 Ma North Pilbara Orogeny (see De Grey Superbasin). Detailed descriptions of both shear zones are presented under 'Central Pilbara Tectonic Zone'.

By analogy with the c. 3240 Ma Strelley Monzogranite of the EPT, the Karratha Granodiorite may have been a low-amplitude dome by the end of its intrusion at c. 3260 Ma. Clastic material in the unconformably overlying Nickol River Formation includes numerous small fragments of black, grey, green chert and pebbles, and boulders of metamorphosed volcanic rock from the Roebourne Group plus feldspathic sandstone and sandstone containing c. 3270 Ma zircon grains from the Karratha Granodiorite (GSWA 136819, Nelson, 1998d). Derivation of such detritus could be explained by erosion of an uplifted area of the KT, possibly as a result of granitic doming similar to that which took place in the EPT. Erosion of the KT may also have been a response to a horst-and-graben topography produced by crustal extension preceding and during deposition of the Nickol River Formation.

Regal Terrane

The Regal Terrane (RT) tectonically overlies the Nickol River Basin and the KT. The terrane is mainly composed of the Regal Formation, which is a c. 3 km-thick basaltic succession metamorphosed to lower amphibolite facies, but in the Cleaverville area, on the northwest Pilbara coast, it also includes a greenschist-facies succession of basaltic units that includes the Port Robinson Basalt (Kiyokawa et al., 2012).

Regal Formation

The Regal Formation is composed of pillowed and massive tholeiitic basalt, dolerite and gabbro sills, and minor komatiite and komatiitic basalt. Clastic sedimentary rocks and felsic volcanic rocks are absent. Well-preserved pillow structures are exposed in many basalt outcrops (Fig. 17) but in areas of higher strain, such as the ridge immediately south of Karratha, the pillows are flattened and stretched to a degree that it is difficult to interpret stratigraphic way-up. The lowest preserved stratigraphic sections of the formation immediately overlying the Regal Thrust, are locally composed of spinifex-textured komatiite and komatiitic basalt. Good examples of komatiite are exposed 1 km north of Mount Regal and in an east–west striking section of the Roebourne greenstone belt 2 km south of Karratha. The presence of komatiite is difficult to explain if the Regal Formation is interpreted as Phanerozoic-style oceanic crust (Isozaki et al., 1991; Kiyokawa and Taira, 1998; Kiyokawa, 1993, Ohta et al., 1996; Kato et al., 1998; Shibuya et al., 2007, 2012), because komatiite is absent from such crust. On the other hand, if the Regal Formation originated in an Archean rift basin, as interpreted by GSWA authors, basal komatiite would not be unusual. Chemical analyses of the Regal Formation komatiites have revealed MgO contents of 27.0 – 36.9 wt%, Cr contents of 1720–5229 ppm, and Ni contents of 1489–2487 ppm (Smithies et al., 2007a). The trace element composition of the komatiites (Fig. 11b) suggests a degree of interaction with either continental crust or underlying clastic sedimentary rocks. Concentrations of Th and La are relatively high and the komatiites show negative Nb anomalies. Some degree of crustal contamination is also indicated by Nd T_{DM} model ages ranging up to c. 3380 Ma in the Regal Formation (Table 5).

Based on Sm–Nd data, Smithies et al. (2007a) interpreted the depositional age of the Regal Formation to be c. 3200 Ma. The minimum depositional age of the Regal Formation is indicated by metamorphism in the KT at c. 3160 Ma (Kiyokawa, 1993; Kiyokawa and Taira, 1998; Smith et al., 1998). This c. 3160 Ma metamorphic event is significant in the tectonic evolution of the northwest Pilbara because it coincides with a change from separation of the KT and EPT to their convergence. This might be explained by collision of the KT with another plate northwest of the Pilbara Craton. The maximum age of the Port Robinson Basalt, interpreted to correlate with the Regal Formation at Cleaverville (see below), is suggested by a date of 3195 ± 15 Ma on the underlying Dixon Island Formation (Kiyokawa et al., 2012), although the Dixon Island Formation – Port Robinson Basalt contact is sheared. The theoretical maximum depositional age of the Regal Formation is constrained by the timing of the East Pilbara Rifting Event at c. 3220 Ma.

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Figure 17. Pillow structures in basalt of the Regal Formation exposed on a wave-cut platform near Cleaverville (Zone 50, MGA 503290E 7716644N). The local succession of pillow basalt flows is 1000 m thick, with most pillow structures being between 1.0 and 1.5 m wide in cross-section. Convex pillow tops and cuspate tail structures developed above adjoining underlying pillows provide conclusive evidence of stratigraphic way-up direction.

Table 5. Summary of published Nd data for 3000–2920 Ma igneous rocks within and adjacent to the Mallina Basin

Tectonic/stratigraphic unit	*Age (Ga)*	ε_{Nd}	T_{2DM} *(Ga)*	*Data source*
MALLINA BASIN				
Portree Granite, 141986	c. 2.95	1.07	3.13	Smithies et al. (2004a)
Indee Suite (n = 5)	c. 2.95	0.8 to –0.1	3.26 to 3.14	Smithies et al. (2004a)
Satirist Monzogranite, 141977	c. 2.93	-0.27	3.22	Smithies et al. (2007a)
Kialrah Rhyolite, 169025	c. 2.94	–1.53	3.33	Smithies et al. (2007a)
Langenbeck Suite, low-Ti (n = 2)	c. 2.95	0.47 to 0.40	3.20	Smithies et al. (2007a)
Langenbeck Suite, 142263	c. 2.95	–0.29	3.25	Smithies et al. (2007a)
South Mallina Basalt Member (n = 4)	c. 2.95	–0.6 to –1.0	3.29 to 3.24	Smithies et al. (2004a)
Yareweeree Boninite Member (n = 2)	c. 2.95	–0.9 to –1.7	3.34 to 3.28	Smithies et al. (2004a)
CPTZ NORTHWEST FROM LOUDENS FAULT				
Comstock Member, 97045027A	c. 2.95	–1.93	3.38	Smithies et al. (2007a)
Louden Volcanics (n = 14)	c. 2.95	–1.41 to –2.30	3.41 to 3.34	Arndt et al. (2001)
Louden Volcanics (n = 4)	c. 2.95	–1.7 to –4.3	3.49 to 3.31	Smithies et al. (2004a)
Mount Negri Volcanics (n = 3)	c. 2.95	–1.31 to –2.64	3.43 to 3.33	Arndt et al. (2001)
Mount Negri Volcanics (n = 4)	c. 2.95	–1.7 to –2.4	3.36 to 3.32	Smithies et al. (2004a)
Warambie Basalt (n = 3)	c. 3.00	0.04 to –1.35	3.37 to 3.26	Arndt et al. (2001)
Warambie Basalt, 176737	c. 3.00	–0.11	3.27	Smithies et al. (2007a)
Caines Well Granitic Complex, 118964	c. 2.93	0.11	3.19	Smithies et al. (2007a)
Caines Well Granitic Complex (n = 2)	c. 2.99	–0.81 to –1.34	3.35 to 3.31	Smithies et al. (2007a)
CPTZ SOUTH OF SHOLL SHEAR ZONE				
Cherratta Granitic Complex (n = 4)	c. 2.99	0.2 to 0.9	3.22 to 3.15	Sun and Hickman (1998)
Radley Suite (n = 19)	c.2.93	–0.06 to –0.71	3.25 to 3.20	Sun and Hoatson (1992)

It is unknown if oceans of Phanerozoic size existed in the Archean. At least some juvenile basaltic crust probably originated in relatively small extensional basins between continental microplates. The current interpretation is that the Regal Formation originated in this type of setting, and was deposited in a basin formed by the c. 3200 Ma breakup of the EPT (Van Kranendonk et al., 2006). This basin, formed by separation of the KT from the EPT, is now referred to as the Central Pilbara Basin (CPB). The juvenile basaltic crust of this basin formed the basement beneath much of the CPTZ, and with time became progressively overlain by younger Mesoarchean basins such as the Whundo, Whim Creek, and Mallina Basins.

Regal Formation at Cleaverville

Excellent exposures of the Regal Formation, and BIF and shale of the Cleaverville Formation are present on the coast at Cleaverville (Figs 17 and 18) and offshore on Dixon Island. At Cleaverville, Kiyokawa et al. (2012) formalized two formation names additional to names previously used by GSWA: the Dixon Island Formation and the Port Robinson Basalt. The c. 3195 Ma Dixon Island Formation is here thought to correlate with the upper pelitic section of the Nickol River Formation, whereas the overlying Port Robinson Basalt is correlated with the Regal Formation. Amongst various informal units mentioned by Kiyokawa et al. (2012), the lagoon pillow basalt', 'basaltic rock member', and 'Regal – Nickol Well unit', are all mapped as part of the Regal Formation by GSWA.

Karratha Event (3160–3140 Ma) and the Regal Thrust (D_1)

Kiyokawa (1993) identified a metamorphic event in the Karratha Granodiorite at 3160 ± 158 Ma (K–Ar, hornblende); Smith et al. (1998) obtained zircon ages of 3156 ± 4 Ma and 3152 ± 5 Ma in the c. 3270 Ma Karratha Granodiorite and attributed these to a thermotectonic event; Beintema (2003) recorded a 3144 ± 35 Ma $^{40}Ar/^{39}Ar$ age in the Ruth Well Formation, which she interpreted as the cooling age of amphibolite-facies metamorphism. Granites adjacent to the TTSZ near Nunyerry Gap have crystallization ages of 3164 ± 4 Ma (GSWA 142946, Nelson, 2000k) and 3166 ± 5 Ma (GSWA 142948, Nelson, 2000l). Although no c. 3160 Ma granites have been identified within the KT, several Mesoarchean igneous and sedimentary rocks of the northwest Pilbara contain c. 3160 Ma zircon xenocrysts or detrital zircons. For example, the c. 3009 Ma Red Hill Volcanics includes inherited zircon dated at 3154 ± 8 Ma and 3141 ± 5 Ma (GSWA 141936, Nelson, 1998g). The 3160–3140 Ma event coincides with the time that the northwestward separation of the KT from the EPT was halted and reversed, resulting in their subsequent convergence. This change to convergence was later than c. 3170 Ma because rift-related volcanics of the Soanesville Group in the Pilbara Well and Wodgina greenstone belts include rocks dated at 3176 ± 3 Ma (GSWA 180098, Wingate et al., 2009c) and 3180 ± 6 Ma (GSWA 180039, Wingate et al., 2009a). In this Report, it is suggested that the most likely explanation of the change to convergence was c. 3160 Ma collision of the KT with another plate

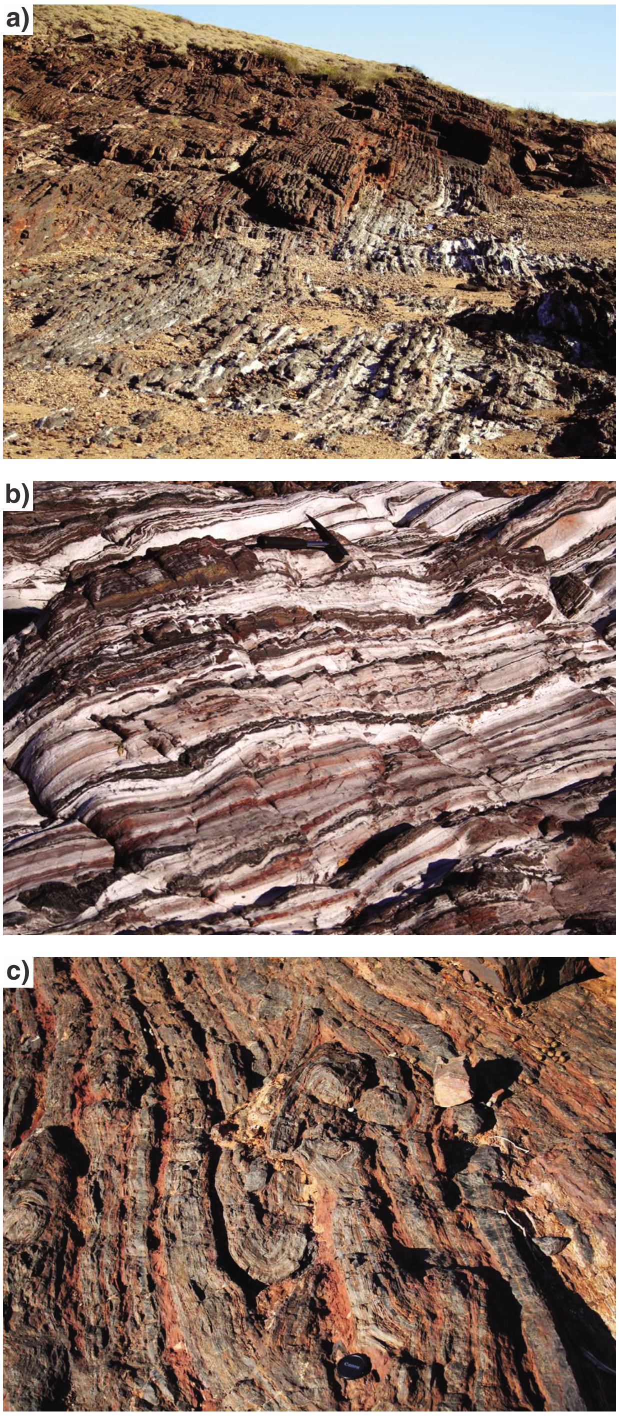

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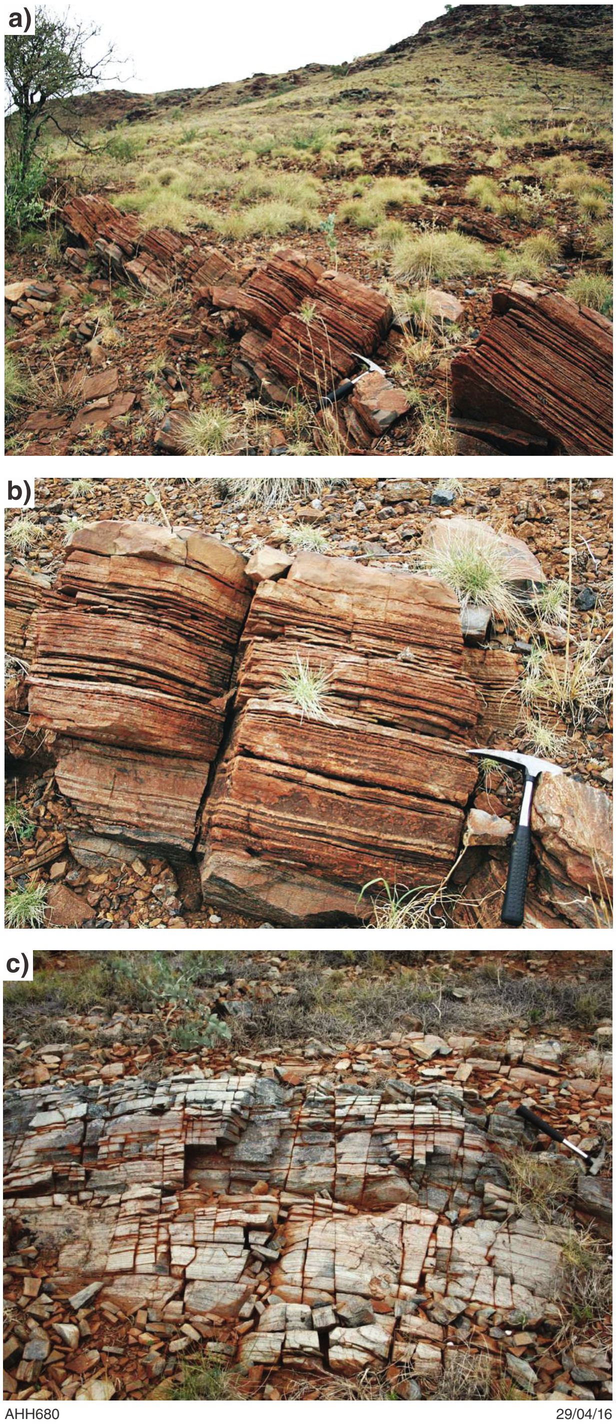
Figure 18. Outcrops of the Cleaverville Formation at Cleaverville Beach (Zone 50, MGA 503290E 7716644N): a) the weathered aspect of the Cleaverville Formation differs markedly between the ridge and the wave-cut platform (scale: most beds are 10–15 cm thick); b) outcrop in the tidal zone, where bleached fine-grained clastic rocks predominate over chert and iron formation (hammer for scale); c) outcrop on the adjacent ridge, where the main rock type is banded iron formation, locally deformed by tight folding and faulting (scale: 5 cm-diameter lens cap).

northwest of the Pilbara Craton. Subsequent convergence of this northwestern plate with the Pilbara Craton, and therefore also the KT with the EPT, is indicated by events in the northwest Pilbara between 3160 Ma and 3068 Ma. These included obduction of juvenile basaltic crust of the Central Pilbara Basin (Regal Formation) onto the KT to form the RT and development of a subduction zone within the Central Pilbara Basin, producing the 3130–3093 Ma arc volcanism and granitic intrusion of the ST (see below). At c. 3068 Ma the convergence culminated in accretion of three northwest Pilbara terranes (KT, RT, and ST) to form the WPS and collision of this superterrane with the EPT, resulting in the 3068–3066 Ma Prinsep Orogeny.

The Regal Thrust is a 1–2 km-thick belt of mylonite and faulted stratigraphy across which the Regal Formation was obducted onto the KT (Hickman, 2001b, 2004a). The obducted slab now forms the RT. The basal contact of the RT is everywhere tectonic (Hickman, 1997a) and is composed of schist, mylonite (Figs 19 and 20), and tectonic breccia. The fine-scale layering in the mylonite is planar in many outcrops (Fig. 19), but at some localities it is strongly deformed by later folding. Excellent exposures of complexly folded mylonite are present 12.5 km south of Karratha, where the primary mylonite foliation S_1 (Hickman, 2001b) is deformed by isoclinal folds and sheaf folds (Fig. 20). Refolded isoclines (Fig. 20b) establish a third phase of deformation within the Regal Thrust, but this may be part of ongoing deformation during D_1, rather than belonging to a later deformation event such as D_2. Sheaf folds, which are characteristic of shear zones, are exposed in some outcrops (Fig. 20c) and cross-sections through this type of fold structure display ‘eye structures’ (Reber et al., 2013).

In most outcrops of the Regal Thrust, the underlying Nickol River Formation is extensively deformed by recumbent folding and tectonic interleaving of various rock types between layers of mylonite. Metamorphosed sandstone, carbonate rocks, and shales of the Nickol River Formation are sheared and flattened with the development of a strong tectonic foliation (S_1). Extreme flattening and stretching of chert and vein quartz clasts in the conglomerates is a feature of many outcrops, including particularly good exposures east of the Lydia gold mine (Fig. 16). East and west from Carlow Castle, the Nickol River Formation is so intensely fractured and sheared that it is almost completely replaced by mylonitic chert and breccia. The mylonite zone includes brown ferruginous chert, black chert (partly in veins), thinly layered green chert, and locally red and black banded chert. Silicified lenses of ultramafic schist are locally present in this tectonic assemblage. South from Weerianna gold mine the Nickol River Formation is replaced by black chert breccia and lenses of silicified ultramafic and mafic schist, tectonically removed from the base of the overlying Regal Formation. The preserved regional extent of the Regal Thrust is approximately 2000 km^2, with exposures east–west from the Roebourne area to Devil Creek and north–south from Cleaverville to the SSZ. The original extent of the thrust cannot be determined due to widespread intrusion by later granites, in particular the 3006–2982 Ma Maitland River Supersuite. However, it is probable that thrusting of the northwest margin of the Central Pilbara Basin onto the KT extended far beyond the Karratha area.



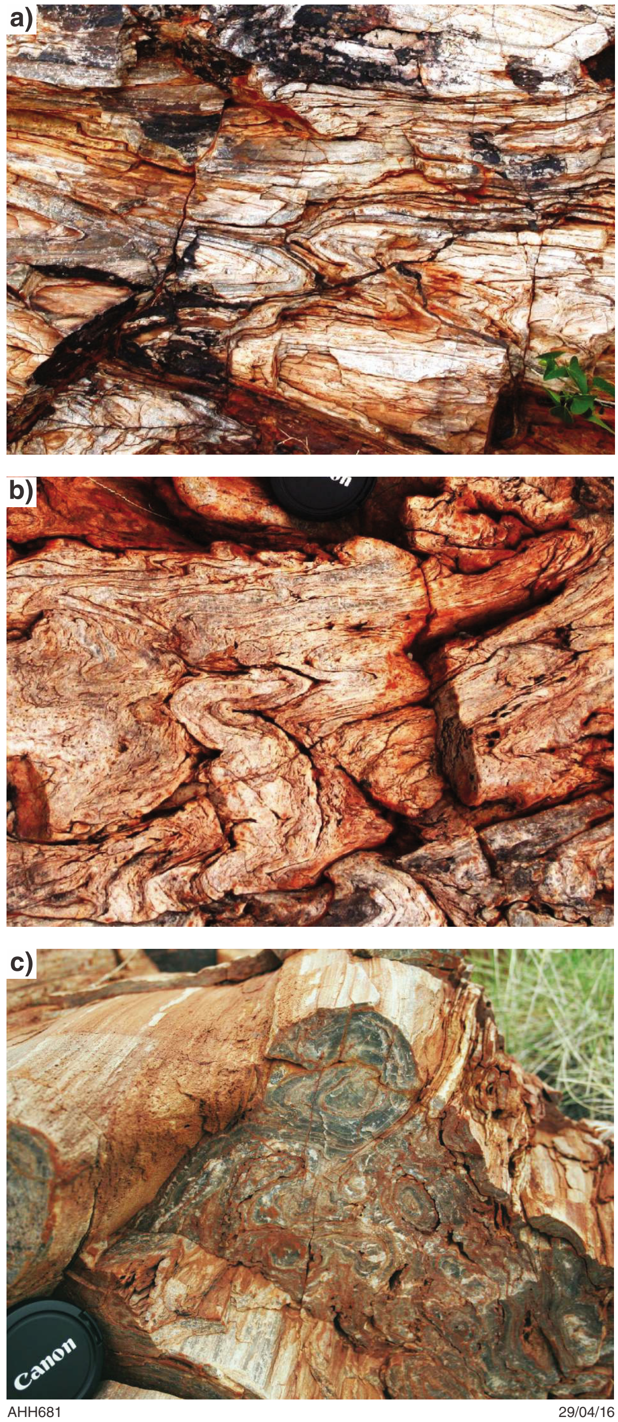


Figure 19. Mylonite in the Regal Thrust near Mount Regal (Zone 50, MGA 474800E 7698290N): a) mafic mylonite dips northwestwards beneath the metabasalts of the Regal Formation (top right), but there are intervening metasedimentary rocks, including BIF (Nickol River Formation) and sheared ultramafic rocks (base of Regal Formation); b) close-up view of mafic mylonite showing its strongly planar tectonic foliation; c) siliceous mylonite of uncertain protolith northeast of Mount Regal (Zone 50, MGA 474130E 7699525N). The pale green colour of the mylonite is suggestive of an extremely silicified ultramafic rock.

Figure 20. Isoclinally folded mylonite in the Regal Thrust 12.5 km southeast of Karratha (Zone 50, MGA 492000E 7696800N): a) finely laminated silicified mylonite with ubiquitous secondary isoclinal folding (field of view: 1 m); b) close-up of refolded isoclines; c) extremely non-cylindrical sheath folds that deform the primary mylonite foliation S_1, showing a mineral lineation parallel to the elongation direction and well-developed 'eye structures' (Reber et al., 2013) in a cross-section perpendicular to the shear direction (scale: 5 cm-diameter lens cap).

Minor structures in the Roebourne Group and Nickol River Formation suggest that movement on the Regal Thrust was to the north or northwest (Kiyokawa et al., 2002; Blewett, 2002; Beintema, 2003). This conclusion is consistent with far better preservation of the Nickol River Formation succession between Karratha and Mount Regal, compared to its southern area of outcrop between Roebourne and Ruth Well. The Nickol River Formation is more deformed and tectonically disrupted in southern areas, suggesting thrusting was from the south. This direction of thrusting also supports derivation of the Regal Formation from the c. 3200 Ma Central Pilbara Basin between the KT and EPT (Van Kranendonk et al., 2006).

Geological mapping of the Roebourne greenstone belt (Fig. 9) has shown that the Regal Thrust was formed as a horizontal to subhorizontal fault zone that was later folded by the c. 2940 Ma Roebourne Synform, Prinsep Dome, and Cleaverville Synform (Figs 8 and 9) (Hickman, 1997b, 2001b; Hickman et al., 2010). The minimum age of the thrust is 3068–3066 Ma, constrained by the D_2 Prinsep Orogeny (see below). Closer dating of the thrust is suggested by the observation that the 3130–3093 Ma ST contains no structures of D_1 type. Thus, D1 is interpreted to coincide with the 3160–3140 Ma amphibolite facies metamorphism of the Karratha Event.

Sholl Terrane

The Sholl Terrane (ST) outcrops over an area of 4000 km^2 immediately south of the SSZ (Fig. 8), and most likely unconformably underlies much of the Mallina Basin and adjacent parts of the Fortescue Basin. The terrane comprises the 3130–3110 Ma Whundo Group and the 3130–3093 Ma Railway Supersuite. The Whundo Group is a 10 km-thick succession of mafic and felsic lavas and felsic volcaniclastic rocks, and is exposed in two main areas of the northwest Pilbara: around Mount Sholl, adjacent to the Cherratta Granitic Complex, and near Mount Fisher on the western margin of the Caines Well Granitic Complex. The depositional extent of the Whundo Group is likely to have been considerably greater because the subduction zone, which is interpreted to have formed it (Smithies et al., 2005a, 2007c) probably extended at least several hundred kilometres northeast–southwest along the length of the Central Pilbara Basin. Geochemical evidence supports the conclusion that the Whundo Basin underlay the Mallina Basin northwest of the EPT (Smithies et al., 2007c).

The same age as the Whundo Group, the Railway Supersuite is composed of tonalite, granodiorite, monzogranite, and at least one large ultramafic–mafic intrusion (Bullock Hide Intrusion). Many parts of the Railway Supersuite consist of banded orthogneiss, containing greenstone enclaves. The bulk of the supersuite is exposed in the Cherratta Granitic Complex, but it also forms part of the Caines Well Granitic Complex. Small felsic intrusions of the supersuite intruding the Nallana Formation and dated at c. 3120 Ma (Smith et al., 1998; Smith, 2003) very likely represent feeders to overlying felsic volcanics of the Tozer and Woodbrook Formations.

SHRIMP U–Pb dating of igneous and sedimentary rocks younger than the ST has revealed 3130–3090 Ma xenocrystic and detrital zircons across a large section of the CPTZ. This supports the conclusion from geochemical data that the ST underlies much of the Mallina Basin (Smithies et al., 2007c). The southwest extent of the ST includes the Cherratta Granitic Complex where the Whundo Group may be represented by the amphibolites of the Weelarra greenstone belt (Fig. 7) (Hickman et al., 2006b).

Whundo Group

The 3130–3110 Ma Whundo Group is a 10 km-thick volcanic succession comprising four lithostratigraphic formations: from base to top, Nallana Formation, Tozer Formation, Bradley Basalt, and Woodbrook Formation (Hickman, 1997a). The metamorphic grade of the Whundo Group is lower greenschist-facies in the northern part of the Whundo greenstone belt, but southwards and eastwards the grade increases to amphibolite facies. The northern limit of the Whundo Group is defined by the SSZ, northwards across which metamorphic grade changes to amphibolite facies in the KT.

Based on geochemical traverses in the northern part of the Whundo greenstone belt, Smithies et al. (2005a) recognized three chemically distinct packages in the succession: lower, middle, and upper (Fig. 21). They interpreted the depositional environment of the Whundo Group to have been similar to that of an intra-oceanic arc. Previous workers (Krapez and Eisenlohr, 1998; Smith et al., 1998; Smith, 2003) interpreted a back-arc setting southeast of an inferred magmatic arc, but Smithies et al. (2005a) argued against a back-arc setting on geochemical evidence (see below).

Island arc or intracontinental rift basin?

Sun and Hickman (1998) reviewed Nd T_{DM} model ages for seven samples from the ST, with ages ranging between 3276 and 3118 Ma. They interpreted the data to indicate that the Whundo Group and the Railway Supersuite were generated as juvenile crust and not derived by melting of crust as old as the KT. Subsequent Nd data (Table 4) generally supports this interpretation although Nd T_{DM} model ages to 3310 Ma have been calculated. Van Kranendonk et al. (2002) commented that the relatively juvenile ε_{Nd} values (epsilon notation, indicated by the ‘ε’ symbol, is used to indicate deviations in parts per 10 000 from CHUR [Chondritic Uniform Reservoir]) of the Whundo Group suggest it represents newly formed uncontaminated crust. They concluded that the Whundo Group formed either in a rift basin or in an island arc setting. However, Smith (2003) used other geochemical data, mainly from the felsic igneous rocks, to argue for involvement of felsic crustal sources and concluded that the tectonic setting was either a back-arc basin or an intracontinental rift.

Following a number of geochemical traverses across the northern part of the Whundo greenstone belt, Smithies et al. (2005a) rejected an extensional tectonic environment.

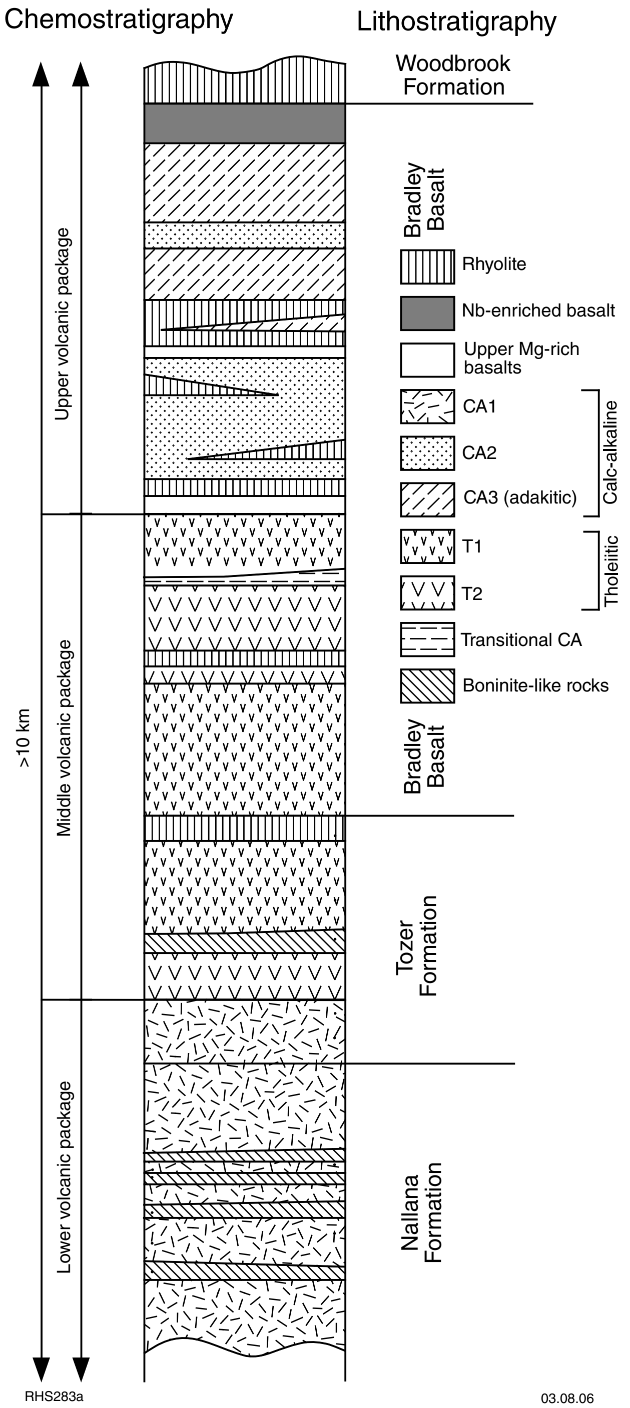


Figure 21. Chemostratigraphic column of the Whundo Group, with lithostratigraphic formation boundaries shown for comparison (after Smithies et al., 2005a)

Their interpretation, which focused on the geochemistry of the basaltic rocks of the Whundo Group, highlighted features suggestive of a modern-style convergent margin setting:

- fine- and broad-scale intercalation of discrete tholeiitic and calc-alkaline volcanics (Fig. 21)
- Nd isotopic compositions that lie very close to depleted mantle values
- rocks with a strong boninite affinity near the base of the sequence (Fig. 22)
- Nb-enriched basalts near the top of the sequence and the close association of these with lavas of adakitic affinity
- evidence that assimilation of crust is an unlikely cause of LREE-enrichments in the boninite-like lavas and calc-alkaline lavas
- a range of source regions from depleted to undepleted and enriched
- mixing of source regions and/or primitive magmas derived from undepleted, depleted, calc-alkaline, and tholeiitic (MORB) sources, with distinct periods characterized by enhanced interaction and magmatic diversity
- evidence that the calc-alkaline lavas are the result of flux melting
- absence of any intervening sequences containing exotic 'continental' material
- absence of evidence for continental (felsic) basement material and the possibility that the sequence was developed over arc-like, mid-level crust of similar age.

The present conclusion is that the Whundo Group and the contemporaneous Railway Supersuite most likely formed above a subduction zone, but not within an ocean of Phanerozoic proportions (see introduction to West Pilbara Superterrane). Nd T_{DM} model ages as old as 3310 Ma may be inherited from the Regal Formation which, as suggested above, might locally have overlain clastic sediments containing detritus from older crust.

Maitland Shear Zone

South and west of the Whundo area the Nallana Formation, at the base of the 10 km-thick group, is separated from the Cherratta Granitic Complex by the Maitland Shear Zone (MSZ). This shear zone is steeply dipping, up to 1 km wide, and includes thick zones of mylonite tectonically interleaved with amphibolite and granitic gneiss. Although the MSZ is steeply dipping it is interpreted to have been formed as a low-angle thrust at c. 2940 Ma when the Whundo Group was thrust southwards across the northeast part of the Cherratta Granitic Complex (Hickman, 2001b; Hickman and Kojan, 2003).

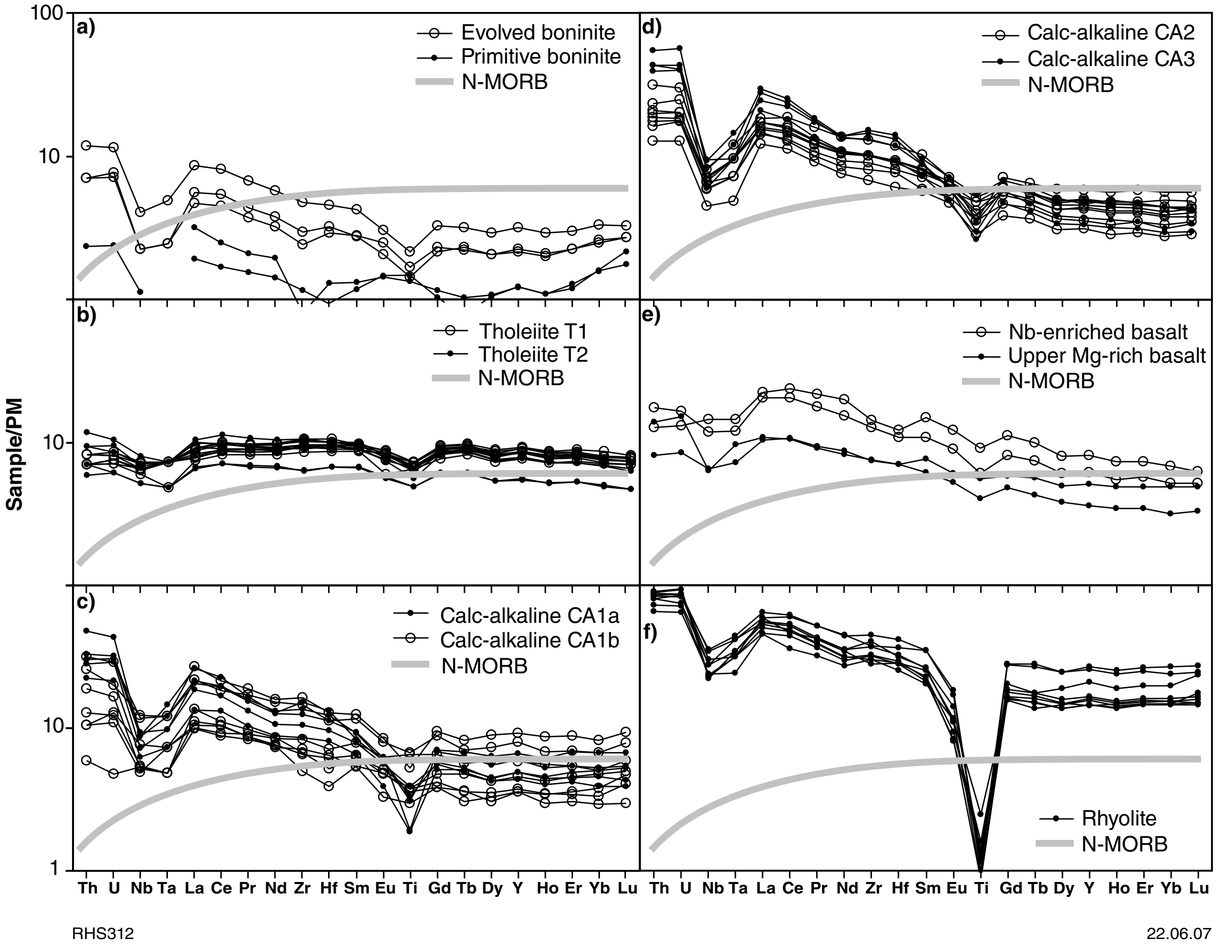


Figure 22. Trace element plots normalized to primitive mantle for various volcanic rocks of the Whundo Group (from Smithies et al., 2007a). Normalizing factors after Sun and McDonough (1989).

Nallana Formation

The Nallana Formation comprises a >2000 m-thick succession of metamorphosed ultramafic and mafic volcanic rocks, containing sills of gabbro and dolerite. Close to the MSZ it includes quartzo-feldspathic gneiss, interpreted as sheared granitic rock, and minor calc-silicate schist, containing abundant scapolite (Hickman and Kojan, 2003). Some of the metavolcanic units have been so extensively silicified that primary composition is difficult to establish in the field. These altered rocks locally resemble dacite in outcrop but most had basaltic protoliths. No felsic volcanic rocks were identified during mapping, preventing direct dating of the formation. The Nallana Formation is older than the basal part of the conformably overlying Tozer Formation, which has been dated at 3128 ± 7 Ma (N4325, Smith, 2003). Near the SSZ the Nallana Formation is intruded by rhyolite dykes dated at 3123 ± 2 Ma (N4413, Smith, 2003) and by the ultramafic–mafic Bullock Hide Intrusion, dated at 3122 ± 3 Ma (GSWA 178164, Wingate and Hickman, 2009b). About 2 km south of the MSZ, banded gneiss of the Railway Supersuite was dated at 3130 ± 4 Ma (GSWA 142835, Nelson, 1999e), whereas granitic rock interlayered with amphibolite within this shear zone was dated at 3114 ± 5 Ma (sample JS33, Smith et al., 1998). Because the ages of the Railway Supersuite and felsic volcanics of the Whundo Group closely coincide, and because the oldest recorded date from the overlying Tozer Formation is c. 3128 Ma, the depositional age of the Nallana Formation is interpreted to be c. 3130 Ma or older.

Based on geochemical data from northern outcrops of the Nallana Formation, Smithies et al. (2005a, 2007c) described this section of the Whundo Group as being composed of calc-alkaline basaltic to andesitic lavas containing thin flows with boninite-like compositions.

Tozer Formation

The Tozer Formation is 2500 m-thick and is composed of metamorphosed tholeiitic basalt, andesite, dacite, rhyolite, and thin units of sedimentary rocks, including volcaniclastic turbidite deposits, quartz-rich volcaniclastic sandstone, chert, and BIF (Hickman et al., 2001, 2006a). SHRIMP U–Pb zircon dating of felsic volcanic rocks from the Tozer Formation has revealed no evidence of significantly older underlying or adjacent felsic crust as would be expected in a continental rift or back-arc basin. Apart from one exceptionally old zircon grain

(c. 3449 Ma, GSWA 114350, Nelson, 1996), the significance of which is uncertain, the oldest of 123 zircon grains analysed from seven samples was younger than c. 3160 Ma. Similarly, of 84 zircon grains from four samples of the Railway Supersuite, the oldest was c. 3151 Ma.

Felsic volcanic rocks, including thick massive deposits of rhyolite breccia, bedded volcaniclastic rocks, and flow-banded rhyolite, make up approximately 30% of the Tozer Formation. However, the felsic units, though up to 1000 m thick in some areas, are of variable thickness and most lens out along strike over distances of 5–10 km. No evidence of thrusting has been observed within the Tozer Formation, and the wedge-shaped felsic units are therefore interpreted to represent local volcanic piles separated by more extensive basalt. Distal sections of the felsic piles, as exposed in outcrops immediately north of Whundo mine, show well-developed, graded bedding in reworked tuffaceous deposits. Volcanic-hosted massive sulfide (VHMS) deposits of copper and zinc have been mined from altered and deformed fine-grained felsic units near the stratigraphic base of the Tozer Formation at Whundo, West Whundo, and Yannery Hill.

Bradley Basalt

The Bradley Basalt conformably or paraconformably overlies the Tozer Formation and is more than 4000 m thick. In many areas there is an abrupt upward change from thick deposits of rhyolite at the top of the Tozer Formation to massive and pillow basalt at the base of the Bradley Basalt. In one area, 3.5 km southwest from Tozer Well, felsic volcaniclastic rocks of the Tozer Formation are separated from overlying basalt of the Bradley Basalt by a unit of quartz-rich volcaniclastic sandstone (Hickman et al., 2006a). This suggests erosion and reworking of the felsic volcanic pile prior to mafic volcanism. However, data from a geochemical traverse in a different area showed no significant lithological or chemical change at the mapped contact between the two formations (Smithies et al., 2005a). Such differences are here attributed to the irregular volcanic composition of the Tozer Formation and either variable topography prior to eruption of the Bradley Basalt or lateral interfingering of mafic and felsic volcanic piles. However, if parts of the Tozer Formation were exposed to erosion before deposition of the >4000 m-thick Bradley Basalt (subaqueous) rapid syndepositional subsidence (probably rift-related) is indicated.

Pillow basalt is common throughout the Bradley Basalt and some excellent examples are exposed west of Harding Dam (Fig. 23). Lower basalts of the formation locally exhibit pyroxene-spinifex texture, indicating more mafic compositions than higher in the formation. Even so, thin units of felsic tuffaceous material are also locally present near the base of the formation. Unless these felsic volcaniclastic lenses were derived from erosion of the Tozer Formation, this suggests a continuation of the bimodal volcanism into the Bradley Basalt. On one traverse (Smithies et al., 2005a) geochemistry defines an important volcanic change mid-way through the formation: tholeiites, like those of the underlying Tozer Formation, give way to calc-alkaline volcanics, including adakite-like rocks (CA3 in Fig. 21). Adakites are felsic igneous rocks characterized by high La/Yb and Sr/Y and low heavy rare earth elements (HREE). Phanerozoic adakites are typically associated with arcs above subduction zones, and are the product of partial melting of subducting hydrous oceanic crust (oceanic slab) or of the mantle metasomatized by slab melt.

In the Bradley Syncline, approximately 7 km northwest of Harding Dam, the change to calc-alkaline volcanics coincides with several thin units of felsic volcaniclastic rocks within the otherwise basaltic succession. Where fine grained, these felsic rocks include exceptionally well-preserved sedimentary structures, such as graded bedding, cross-bedding, flame structures, and slump folds (Fig. 24).

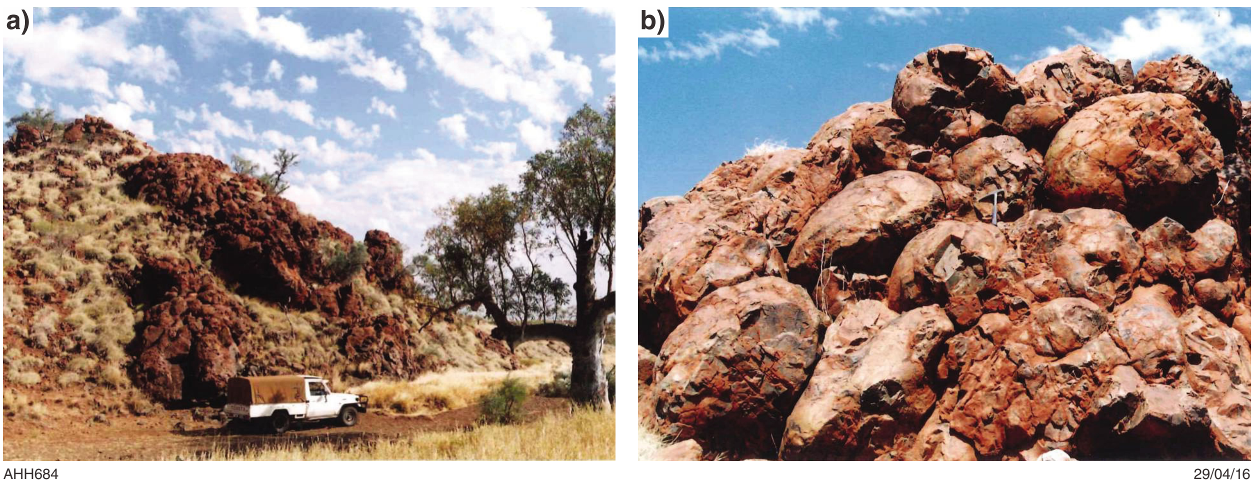


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Figure 23. Pillow lava in the Bradley Basalt 2.5 km northwest of Harding Dam (Zone 50, MGA 508900E 7682000N): a) small hill entirely composed of pillow basalt; b) close-up of pillow structures (1.5 to 2 m in diameter) near the top of the hill. Interpillow material has been removed by weathering and erosion.

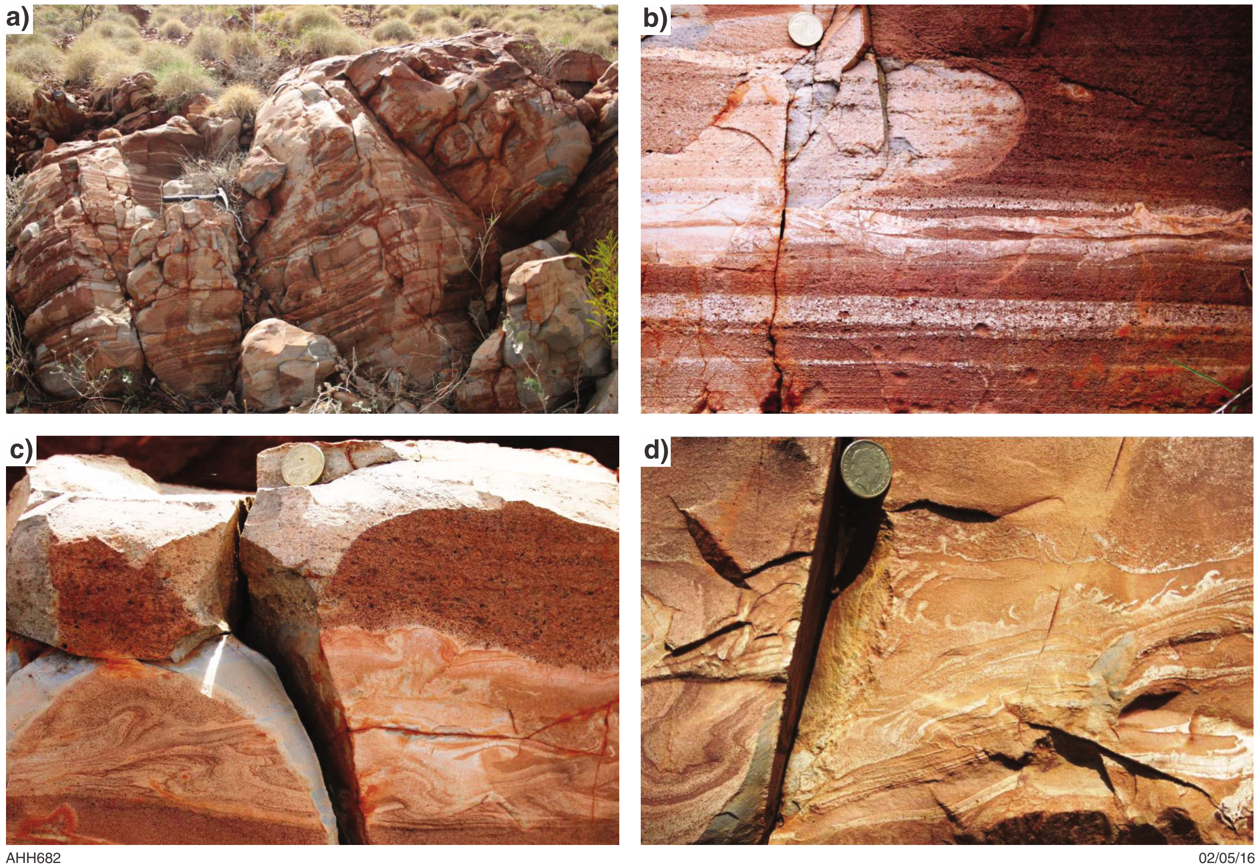


Figure 24. Sedimentary structures exposed in an outcrop of a 20 m-thick felsic volcaniclastic unit within the Bradley Basalt (Zone 50, MGA 505426E 7684197N): a) view of the outcrop showing well-developed bedding; b) fine-scale cross-bedding revealed in a weathered vertical section through eroded ripples; c) slump folding and local diapiric injection of fine-grained felsic volcaniclastic sediment into overlying coarser grained units (basal parts of upward-fining graded beds); d) well-developed flame structures (centre and right) and syndepositional slump folding (left). Scale in b) to d): 2 cm-diameter coin

The lithology of these units suggests distal deposition, possibly from turbidity currents, during erosion of one or more felsic volcanic centres. However, it is unknown if these distant felsic sources were contemporaneous with the Bradley Basalt or merely exposed areas of the Tozer Formation.

Woodbrook Formation

The Woodbrook Formation is composed of calc-alkaline volcanic rocks similar to those in the Bradley Basalt, but it is distinguished by its predominantly felsic composition. In the type area west of Woodbrook the formation comprises a 1000 m-thick succession of metamorphosed rhyolite tuff and agglomerate. Fragments within the tuff include pumice, flow-banded lava, and porphyritic and spherulitic lava. This rock was dated at 3117 ± 3 Ma (GSWA 127378, Nelson, 1998c), which is a very similar age to most dated samples from the Bradley Basalt and Tozer Formation. This suggests that the middle and upper parts of the Whundo Group were deposited within 5 million years. The lower part of the Woodbrook Formation includes minor basalt and a single thin BIF unit was mapped near the top of the formation (Hickman, 2000).

Railway Supersuite

The second major component of the ST, the Railway Supersuite, is mainly composed of metamorphosed tonalite, trondhjemite, granodiorite, and minor monzogranite, dated between 3130 and 3093 Ma (Van Kranendonk et al., 2006). The supersuite forms large parts of the Cherratta Granitic Complex but also forms small intrusions within the Whundo Group (Twin Table Monzogranite, Van Kranendonk et al., 2006). In the Caines Well Granitic Complex it is represented by unnamed 3111–3093 Ma banded biotite–trondhjemite gneiss (GSWA 118965, Nelson, 1997b; Smithies and Champion, 1998). Minor components of the supersuite are mafic and felsic intrusions within the lower part of the Whundo Group, and these include the 25 km^2 Bullock Hide Intrusion, dated at 3122 ± 3 Ma (GSWA 178164, Wingate and Hickman, 2009) and intrusive rhyolite dykes dated at 3123 ± 2 Ma (N4413, Smith, 2003). Granitic gneiss boulders in conglomerate at the base of the 3010–2990 Ma Whim Creek Group near Red Hill were probably eroded from the Railway Supersuite in the Caines Well Granitic Complex.

The oldest zircons dated from the supersuite in the northwest Pilbara Craton are c. 3150 Ma, and Nd T_{DM} model ages for the supersuite in this area range between 3310 and 3210 Ma (Smith et al., 1998; Sun and Hickman, 1998; Smithies et al., 2007a). Smith (2003) described the rare earth element (REE) geochemistry of two samples from the supersuite, noting LREE enrichment, negative Eu anomalies, flat HREE patterns, and high levels of high field strength elements (HFSE). This REE pattern is similar to that for felsic volcanic rocks of the Whundo Group and Smith (2003) concluded that the intrusive and extrusive rocks had a similar source. Smithies and Champion (1998) reported that the Railway Supersuite of the Cherratta and Caines Well Granitic Complexes is Sr-undepleted, has high ε_{Nd} between +2.5 and +2.8, and Nd-model ages very similar to the crystallization ages. They concluded that these granitic rocks represent juvenile crust and Sun and Hickman (1998) reached the same interpretation. Two additional samples of the supersuite (GSWA 142534 and 142536) reported by Smithies et al. (2007a) have moderately high ε_{Nd} (+1.63 and +1.38) and Nd model ages of 3220 to 3240 Ma. One of the samples (GSWA 142534) was collected close to the Weelarra greenstone belt, supporting field mapping evidence that the supersuite extends into the western part of the Cherratta Granitic Complex.

Pinnacle Hill Gneiss

The Pinnacle Hill Gneiss now occupies a substantial part of the Cherratta Granitic Complex (Hickman et al., 2006b), although its precise extent remains to be tested by more detailed mapping, geochemistry, and geochronology. The composition of the gneiss is mainly tonalite to trondhjemite (Smithies and Champion, 1998), and it is typically banded with elongate greenstone xenoliths, some of which are up to several hundred metres in length. In view of the juvenile chemistry of the gneiss and the lack of contamination by significantly older crust, the greenstones were most likely derived from the lower section of the Whundo Group, either by igneous injection and stoping or by tectonic interleaving during thrusting related to formation of the MSZ (see Tectonic events). Excellent exposures of the Pinnacle Hill Gneiss are present in the bed of the Maitland River 6 km south-southwest of Whundo copper mine, and in outcrops immediately beneath the unconformity of the Fortescue Group between Mount Leopold and Byong Creek.

Bullock Hide Intrusion

The Bullock Hide Intrusion (Hickman, 1997b) intrudes the base of the Nallana Formation over an area of 25 km^2 area south of the SSZ. The intrusion is composed of layered units of serpentinite (metaperidotite) and metamorphosed gabbro, dolerite, and leucogabbro. Northern outcrops of the intrusion are mainly composed of gabbro, containing numerous angular blocks of metabasalt. This suggests initial fragmentation and net-veining of the basalt followed by its incorporation as xenoliths within the gabbro. The sample dated at 3122 ± 3 Ma (GSWA 178164, Wingate and Hickman, 2009) was a coarse-grained leucogabbro from the upper part of the intrusion.

Tectonic events (D_2)

Whereas the D_1 Regal Thrust and associated minor structures are interpreted to have dated from c. 3160 Ma, the earliest tectonic structures preserved in the ST formed during c. 3070 Ma collision with the KT along the SSZ. Deformation related to this event is assigned to D_2 and included sinistral movement along the SSZ, and thrusting of the Whundo Group onto the Cherratta Granitic Complex upon the MSZ.

Maitland Shear Zone

The Maitland Shear Zone (MSZ) was first recognized from GSWA mapping between 1994 and 1995 (Hickman, 1997a,b). The shear zone is a belt of mafic mylonite and sheared granitic rocks that extends 35 km southwards from the SSZ to an area southeast of Whundo copper mine. In this area the MSZ is unconformably overlain by the Fortescue Group, but geophysical data indicate that it continues east-northeast for another 20 km (Hickman, 2004b). The MSZ is interpreted to have originated by southerly directed low-angle D_2 thrusting of the Whundo Group over the Railway Supersuite (Hickman and Kojan, 2003). The present variable orientation of the zone and its relatively steep inclinations, either to the northeast or north, is the result of 2945–2930 Ma folding of the ST about northeast-trending axes (Hickman and Kojan, 2003). About 5 km south of Whundo copper mine several subsidiary thrusts in the Nallana Formation are truncated by the South Whundo Monzogranite dated at 3013 ± 4 Ma (JS35, Smith et al., 1998). Late movement on the MSZ took place at c. 2930 Ma (D5, Hickman and Kojan, 2003) and established a strong tectonic foliation that cuts the folds, but is intruded by the c. 2925 Ma Radio Hill Intrusion. The northern end of the shear zone is abruptly truncated by the east–west-striking SSZ, which has a substantial component of 2920 Ma dextral movement (Hickman, 2001b).

Sholl Shear Zone

The Sholl Shear Zone (SSZ) is one of the major faults of the Pilbara Craton and is interpreted to have played a key role in the crustal evolution of the northwest Pilbara between 3200 and 2920 Ma. Accordingly, its description is delayed until later in the Report (De Grey Superbasin).

Prinsep Orogeny

Plate convergence from c. 3160 Ma onwards led to accretion of three terranes (KT, RT, and ST) to form the WPS at c. 3070 Ma. This collision resulted in the Prinsep Orogeny (D_1), which included low-angle thrusting on the MSZ and major sinistral strike–slip movement on the SSZ. The prevalence of sinistral strike-slip on the generally east-northeast-striking SSZ suggests oblique north–south convergence. During the same period, the EPT converged with the ST and RT. Crustal melting associated with collision resulted in intrusion of the 3068–3066 Ma Elizabeth Hill Supersuite, both in the WPS and in the western part of the EPT. The age of intrusions of the Elizabeth Hill Supersuite in the northwest and east Pilbara is inferred to define the age of the orogeny.

However, indirect isotopic evidence from inherited zircons in younger igneous and sedimentary rocks suggests the orogeny spanned a longer period, possibly c. 3080 Ma to c. 3060 Ma.

Elizabeth Hill Supersuite

The Elizabeth Hill Supersuite was first identified from geochronology in the northwest Pilbara (GSWA 142661, Nelson, 1998m), despite the fact that it appears to be restricted to a relatively small section (70 km^2) of the ST. Subsequent mapping and geochronology in the Yule Granitic Complex of the EPT indicates a more widespread distribution and a derivation by disequilibrium melting of Paleoarchean granitic protoliths (Smithies, 2003).

Cliff Pool Tonalite

The Cliff Pool Tonalite, the only named intrusion of the Elizabeth Hill Supersuite in the northwest Pilbara, was dated at 3068 ± 4 Ma (GSWA 142661, Nelson, 1998m). The tonalite includes xenocrystic zircons of Railway Supersuite age, supporting field evidence that it intrudes the c. 3130 Ma Pinnacle Hill Gneiss. Its northern contact with this gneiss is diffuse and interlayered, with sheets of the tonalite intruding rafts of the gneiss. The southern contact of the tonalite is intruded by granitic rocks of the 3006–2982 Ma Maitland River Supersuite. The Cliff Pool Tonalite was intruded within the same north-northeast-trending zone as the 2930–2924 Ma Radley Suite, and this coincides with a significant geophysical lineament. Later in this Report it is suggested that this lineament may be due to the presence of an early branch of the SSZ (see Sholl Shear Zone). Intrusion of the Cliff Pool Tonalite may therefore have been on one of the boundary faults of the CPTZ, reactivated by the Prinsep Orogeny. Alternatively, the tonalite might have been intruded beneath the MSZ when the Whundo Group was thrust across the Railway Supersuite in the Cherratta Granitic Complex. In the Yule Granitic Complex, in the northwest section of the EPT, the Cockeraga Leucogranite was dated at 3066 ± 4 Ma (GSWA 169016, Nelson, 2002b) and 3068 ± 22 Ma (GSWA 169014, Nelson, 2002a).

De Grey Superbasin

Introduction

The 3066–2919 Ma De Grey Superbasin (Van Kranendonk et al., 2006) is a major tectonic unit that unconformably overlies Paleoarchean and early Mesoarchean crust across the northern Pilbara Craton. Initial formation of the superbasin was in response to crustal relaxation, extension, and subsidence following the 3068–3066 Ma Prinsep Orogeny. This orogeny was produced by collision between the Karratha Terrane (KT), Regal Terrane (RT), and Sholl Terrane (ST) to form the West Pilbara Superterrane (WPS), and collision of the WPS with the East Pilbara Terrane (EPT). The stratigraphy of the superbasin spans almost 150 million years and it is a geologically complex unit which, along with older units within the Central Pilbara Tectonic Zone (CPTZ), contains the most economically important mineral deposits of the northern Pilbara Craton. The lithostratigraphic interpretation of the superbasin was revised and briefly summarized by Van Kranendonk et al. (2006). This Report follows the stratigraphic nomenclature of that revision except for some units of the Mallina Basin for which a modified interpretation is presented.

In the northwest Pilbara, the supracrustal succession of the De Grey Superbasin is composed of three volcano-sedimentary basins (Gorge Creek, Whim Creek, and Mallina) separated either by unconformities or major faults. The Gorge Creek and Whim Creek Basins contain the Gorge Creek and Whim Creek Groups, respectively. In the present revision of the Mallina Basin this contains the Croydon and Bookingarra Groups (Bookingarra Group is redefined in Appendix 2). The Whim Creek Group overlies the Gorge Creek Group across an erosional unconformity, whereas in the central part of the Mallina Basin the base of the Croydon Group overlies the Gorge Creek Group via a transition sequence (Fig. 25). Elsewhere other parts of the Croydon Group unconformably overlie the Gorge Creek Group. Three igneous supersuites (Orpheus, Maitland River, and Sisters) were intruded during formation of the superbasin and now make up the granitic complexes of the northwest Pilbara. Geochronology reveals that all but one of these complexes are composed of more than one supersuite and include granitic rocks older than the De Grey Superbasin (up to 3270 Ma, Table 2).

In the southeast Pilbara, the Mosquito Creek Basin, containing the Nullagine Group, separates the EPT from the Kurrana Terrane (Fig. 3). Poor exposure, structural complexity, and an apparent absence of felsic igneous rocks (for dating purposes) in the Nullagine Group have hindered interpretation of its stratigraphy and crustal evolution. The upper sedimentary fill of the basin, the Mosquito Creek Formation, is lithologically similar to the Croydon Group of the Mallina Basin and detrital zircon ages indicate it is of similar depositional age (Bagas et al., 2008).

Tectonic setting

The greater part of the De Grey Superbasin, containing the De Grey Supergroup, lies within the east-northeast-trending CPTZ (Hickman, 1999, 2001a, 2004a; Smithies and Farrell, 2000; Hickman et al., 2001). As already discussed, this section of the Pilbara Craton was tectonically and magmatically active from the continental breakup of the EPT between 3235 and 3200 Ma (see EPT Rifting Event) to the Prinsep Orogeny at 3068–3066 Ma, and from this event to the end of the North Pilbara Orogeny at c. 2919 Ma. The De Grey Superbasin was deposited in the CPTZ between c. 3066 Ma and c. 2919 Ma. The Gorge Creek Basin was not significantly influenced by the existence of the underlying CPTZ because it extended well outside this zone north of the SSZ and even more so in the east Pilbara. In the northwest Pilbara, the Cleaverville Formation of the Gorge Creek Group was deposited on top of the SSZ, establishing that the greatest strike-slip movement on that fault had taken place before c. 3020 Ma (Hickman, 1997a). However, deformation and erosion of the Cleaverville Formation prior to deposition of the Whim Creek Group provides evidence of renewed compression at c. 3015 Ma. As described below, the SSZ was reactivated to form a south-facing fault scarp during c. 3010 Ma deposition of the Warambie Basalt at the base of the Whim Creek Group. From that time on deposition of the Whim Creek, Croydon, and Bookingarra Groups (Bookingarra Group is redefined in Appendix 2) was largely within the CPTZ.

The oldest basin of the De Grey Superbasin, the Gorge Creek Basin, was formed by north–south or northwest–southeast crustal extension. Shallow-water sedimentary deposition of the Gorge Creek Group took place both within and on both sides of the CPTZ, and apparently was not influenced by the thickness or composition of the underlying crust. Two alternative interpretations have been reached concerning the tectonic settings of the Whim Creek and Mallina Basins from c. 3015 Ma onwards. One suggests that these basins evolved in continental arc and back-arc settings related to northwest–southeast plate convergence and subduction (Krapez and Eisenlohr, 1998; Smith et al., 1998; Blewett, 2002; Van Kranendonk et al., 2002; Beintema, 2003; Smith, 2003; Pike et al., 2006; Hickman, 2012), whereas the other argues that the basins evolved in the absence of subduction in continental rift settings (Barley, 1987; Smithies et al., 1999, 2001b; Smithies and Champion, 2000; Smithies, 2002a; Van Kranendonk et al., 2006). As noted by Pike et al. (2006) and Smithies et al. (2007a), the geochemistry of the Whim Creek Group is consistent with either a continental arc above a subduction zone or with a continental rift setting. In other words, geochemistry alone is inadequate to resolve this question and must be interpreted in combination with geological evidence.

Evidence consistent with subduction includes a southeast migration of granite ages from the northwest Pilbara (c. 3023 Ma) into the east Pilbara (c. 2919 Ma). This Report illustrates this migration in Figure 26. Granitic rocks dated between c. 3023 Ma and c. 3012 Ma (Orpheus Supersuite) are restricted to a northeast–southwest-trending zone on both sides of and within 30 km of the SSZ. The more voluminous 3006–2982 Ma Maitland River Supersuite intrudes the same area, but extends up to 50 km southeast and northwest from the SSZ. Granitic and mafic intrusions of the 2954–2919 Ma Sisters Supersuite extend from the SSZ approximately 200 km southeast as far as the Shaw Granitic Complex of the EPT. Granitic intrusions of the Sisters Supersuite emplaced within the EPT have a more restricted age range (2940–2919 Ma) than those in the northwest Pilbara (2954–2924 Ma).

An alternative to c. 2950 Ma subduction was proposed by Van Kranendonk et al. (2006). They attributed the southeast extent of the Sisters Supersuite to an influx of mantle material from northwest to southeast. They suggested that this influx was initiated by crustal relaxation after the Prinsep Orogeny and that it accompanied by breakoff of a previously subducted slab. The likely origin of this slab was interpreted to be the c. 3130 Ma Sholl subduction event. In this model, the Whim Creek Group was a result of magmatism related to rupture of a hinge in the slab, and later magmatic events (Sisters Supersuite) were related to influx of hot new mantle material during sinking of the detached slab. However, whereas the Sisters Supersuite outcrops over of the western 70% of the northern Pilbara Craton, the ST is restricted to part of the CPTZ.

In the interpretation of 3023–2919 Ma subduction the 3006–2982 Ma Maitland River Supersuite represents the core of a northeast–southwest-trending magmatic arc southeast of a subduction zone that lay to the northwest of the Pilbara Craton (Krapez and Eisenlohr, 1998; Smith et al., 1998; Blewett, 2002; Smith, 2003). The contemporaneous Whim Creek Group was the volcanic part of the arc and to the southeast the Mallina Basin was formed as a c. 3010 Ma back-arc basin. Whereas the Whim Creek Group lies within the CPTZ and therefore could have originated through ongoing processes following the Whundo subduction (noted above), the main belt of the Maitland River Supersuite intruded the KT northwest of the SSZ, and therefore cannot be explained in the same way. Krapez and Eisenlohr (1998) and Smith et al. (1998) inferred the existence of a subduction zone off the northwest Pilbara Craton from c. 3270 Ma to explain the origin of the Karratha Granodiorite. The present interpretation is that the 3270–3261 Ma Karratha Granodiorite and the c. 3280 Ma Roebourne Group were formed by a mantle plume, and that a subduction zone did not develop northwest of the Pilbara Craton until after the 3068–3066 Ma Prinsep Orogeny. This would be consistent with the northwest subduction zone replacing the Sholl subduction zone when that was destroyed by c. 3070 Ma accretion of the WPS.

Basin relationships

The stratigraphic succession of the De Grey Supergroup varies between different sections of the northwest Pilbara Craton (Fig. 27). Differences of stratigraphy and tectonic history exist across major faults and shear zones. Additionally, there is isotopic evidence that the composition of the basement to the supergroup is also variable across these faults (Fig. 4). Combined with mapping evidence of significant post-3066 Ma strike-slip displacements on the faults, the differences indicate that the present juxtapositions of the successions summarized in Figure 27 do not precisely correspond to their relative depositional positions. Most previous workers have recognized the importance of strike-slip faulting in the Mallina and Whim Creek Basins, and between these basins and the Paleoarchean terranes to the northwest and southeast (Barley, 1987; Krapez and Barley, 1987; Hickman, 1997a, 2001b, 2004a; Krapez and Eisenlohr, 1998; Sun and Hickman, 1998; Beintema, 2003). Such strike-slip faulting is typical of convergent margins.

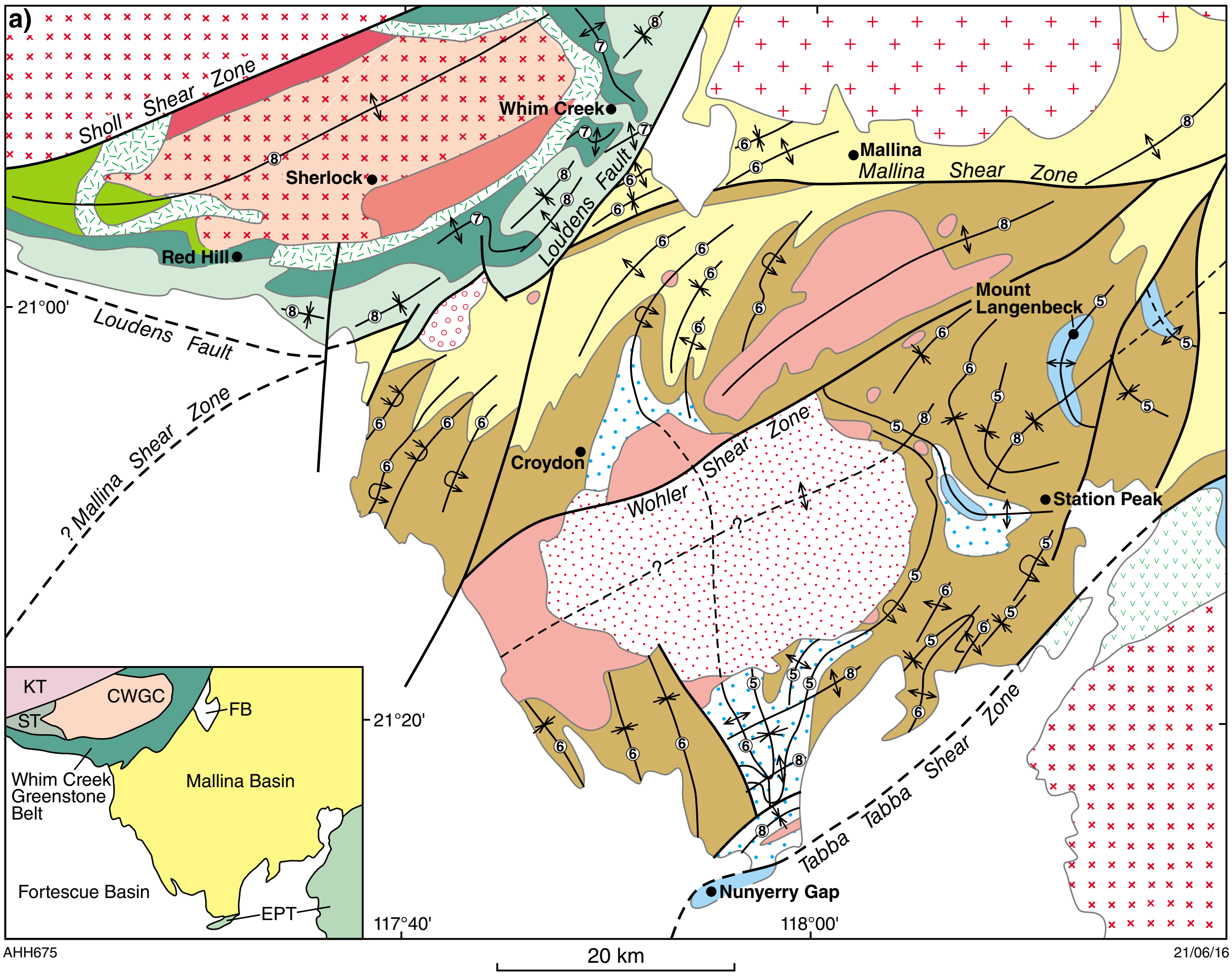


Figure 25. Stratigraphic and structural differences between the central Mallina Basin and the Whim Creek greenstone belt. The map and reference illustrate stratigraphic and structural differences across the Loudens Fault. Note that the Gorge Creek Group, which underlies the Croydon Group in the Mallina Basin, is exposed within the cores of several anticlines. Fold axial traces simplified from Smithies (1998b), Smithies and Farrell (2000), and Krapez and Eisenlohr (1998). Sample numbers, dates, and references are tabulated.

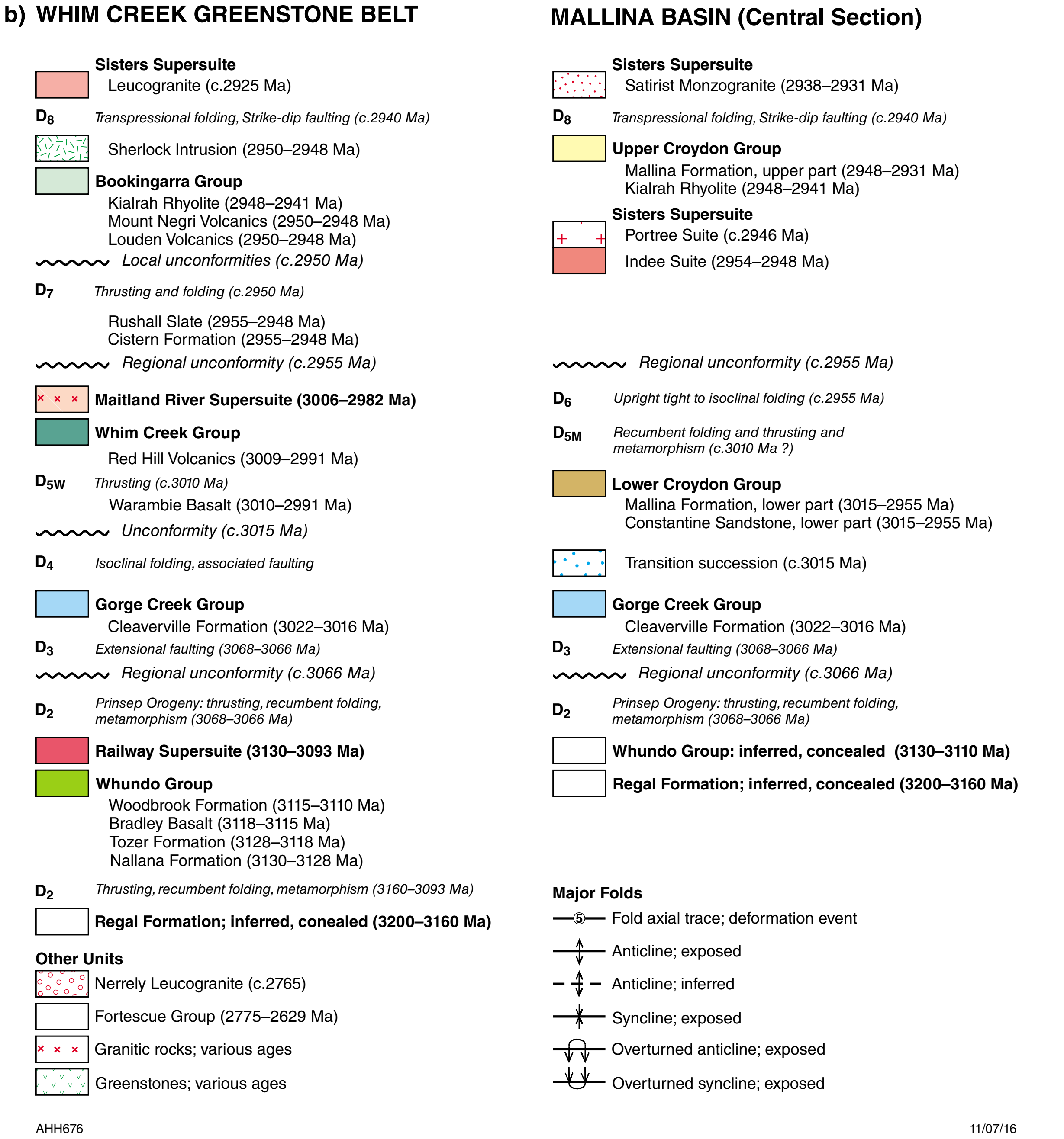
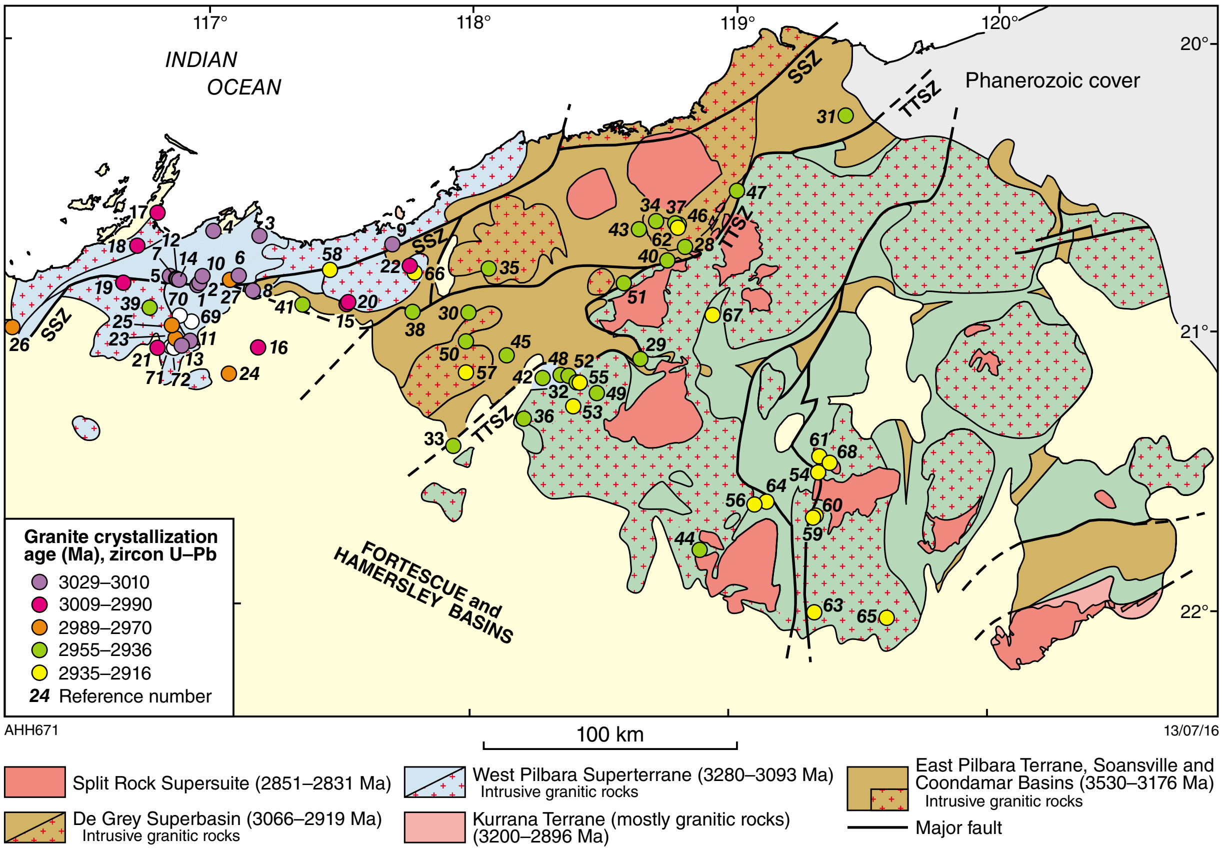


Figure 25. legend



No.	Sample	Sample details	Age (Ma)	Reference
1	JS25	tonalite	3024 ± 4	Smith (2003)
2	118976	porphyritic dacite	3023 ± 9	Nelson (1997h)
3	144224	porphyritic dacite	3021 ± 3	Nelson (1999j)
4	127327	porphyritic dacite	3018 ± 2	Nelson (1998a)
5	N4028	granodiorite	3017 ± 6	Smith (2003)
6	168936	monzodiorite	3016 ± 4	Nelson (2001j)
7	N3188	granodiorite	3016 ± 2	Smith (2003)
8	127320	quartz granophyre	3014 ± 6	Nelson (1997j)
9	118966	granodioritic gneiss	3014 ± 3	Nelson (1997c)
10	118979	quartz-feldspar porphyry	3014 ± 2	Nelson (1997i)
11	N4350	monzogranite	3013 ± 4	Smith (2003)
12	N4097	granodiorite	3013 ± 8	Smith (2003)
13	142832	intermediate rock	3012 ± 4	Nelson (2000a)
14	N4128	granodiorite	3012 ± 3	Smith (2003)
15	141936	welded tuff	3009 ± 4	Nelson (1998g)
16	168932	porphyritic granodiorite	3006 ± 12	Nelson (2001h)
17	178148	biotite granite	2999 ± 10	Wingate, Hickman (2009a)
18	136844	granite	2997 ± 3	Nelson (1998e)
19	118974	porphyritic granodiorite	2994 ± 2	Nelson (1997f)
20	UWA	porphyritic dacite	2991 ± 12	Barley et al. (1994)
21	142657	granodiorite	2990 ± 3	Nelson (1999d)
22	142950	porphyritic monzogranite	2990 ± 5	Nelson (2000n)
23	142438	granodiorite	2988 ± 4	Nelson (1999c)
24	168934	biotite monzogranite	2988 ± 7	Nelson (2001i)
25	N3132	granodiorite	2985 ± 7	Smith (2003)
26	142893	monzogranite	2982 ± 5	Nelson (1999i)
27	142430	monzogranite	2970 ± 5	Nelson (1999b)
28	142935	hornblende granodiorite	2954 ± 4	Nelson (2000e)
29	180038	metamonzodiorite	2952 ±7	Wingate et al. (2009a)
30	118967	tonalite	2948 ± 5	Nelson (1997d)
31	169025	rhyolite	2948 ± 3	Nelson (2002e)
32	142941	porphyritic dacite	2946 ± 4	Nelson (2000g)
33	142945	porphyritic andesite	2946 ± 20	Nelson (2000j)
34	160730	biotite granodiorite	2946 ± 3	Nelson (2001d)
35	142889	alkali granite	2946 ± 6	Nelson (1999g)
36	142938	tonalite	2945 ± 5	Nelson (2000f)
37	160744	granodiorite	2945 ± 2	Nelson (2001e)
38	160498	granodiorite	2945 ± 6	Nelson (2000a)
39	136826	tonalite gneiss	2944 ± 5	Nelson (1997k)
40	KB351	aplite	2944 ± 7	Beintema (2003)
41	144261	rhyolite	2943 ± 7	Nelson (1998u)
42	180097	dacite dyke	2942 ± 8	Wingate et al. (2009d)
43	142934	biotite monzogranite	2941 ± 4	Nelson (2000b)
44	169018	biotite monzogranite	2941 ± 5	Nelson (2002c)
45	142892	porphyritic rhyolite	2941 ± 4	Nelson (1999h)
46	160728	biotite monzogranite	2940 ± 2	Nelson (2001c)
47	160745	biotite granodiorite	2940 ± 3	Nelson (2001g)
48	KB746	granite	2939 ± 12	Beintema (2003)
49	142176	biotite monzogranite	2938 ± 3	Nelson (1999a)
50	141973	biotite monzogranite	2938 ± 4	Nelson (1998h)
51	KB770	granite	2938 ± 15	Beintema (2003)
52	142936	leucocratic monzogranite	2937 ± 7	Nelson (2000d)
53	142937	leucocratic monzogranite	2935 ± 3	Nelson (2000e)
54	T94/31	granite	2934 ± 2	Zegers (1996)
55	160442	biotite monzogranite	2933 ± 4	Nelson (2000r)
56	142884	biotite syenogranite	2933 ± 3	Nelson (1998q)
57	141977	granite	2931 ± 5	Nelson (1998i)
58	169080	quartz diorite, drillcore	2931 ± 8	Nelson (2004)
59	142965	monzogranite	2929 ± 4	Nelson (2000p)
60	142967	biotite monzogranite	2929 ± 5	Nelson (2000q)
61	142882	biotite monzogranite	2928 ± 2	Nelson (1998o)
62	160727	biotite granodiorite	2928 ± 6	Nelson (2001b)
63	178047	monzogranite	2927 ± 19	Wingate (in prep.)
64	142885	biotite monzogranite	2927 ± 3	Nelson (1998s)
65	178049	biotite monzogranite	2926 ± 6	Nelson (2005)
66	118964	foliated granite	2925 ± 4	Nelson (1997a)
67	169021	leucocratic syenogranite	2923 ± 3	Nelson (2002d)
68	142883	porphyritic syenogranite	2919 ± 3	Nelson (1998p)
69	N3162	granite, Radley Suite	2930 ± 4	Smith (2003)
70	N4450	granite, Radley Suite	2929 ± 13	Smith (2003)
71	103227	pegmatite, Radley Suite	2925 ± 16	Arndt et al. (1991)
72	142436	monzonite, Radley Suite	2924 ± 5	Nelson (1998k)

Figure 26. Southeasterly migration of granitic intrusion in the northwest Pilbara from 3023–2919 Ma, represented by the location of granite crystallization ages from all available zircon U–Pb geochronology (SHRIMP) data. Sample numbers, dates, and references are tabulated.

The Gorge Creek Group was deposited in an extensional continental basin following the 3068–3066 Ma Prinsep Orogeny, and is interpreted to have extended across all parts of the northwest Pilbara. Although U–Pb isotopic evidence from detrital zircons indicates that the Cleaverville Formation was deposited between c. 3022 and 3016 Ma, its maximum depositional age is not well constrained. The inferred maximum depositional age of the entire group is 3066 Ma, based on formation of the basin following the Prinsep Orogeny. The Cleaverville Formation is intruded by felsic sills of the Orpheus Supersuite, the oldest of which is dated at c. 3023 Ma (see below). The isotopic age range of the Gorge Creek Group is outside the ranges of the Whim Creek and Croydon Groups, both of which are observed to unconformably overlie the Gorge Creek Group in various areas. The oldest isotopic age recorded from a sandstone interpreted to be part of the Whim Creek Basin is 3016 ± 3 Ma (GSWA 168924, Nelson, 2001h). However, this is a maximum depositional age based on the ages of detrital zircons, presumed to have been derived from erosion of the Orpheus Supersuite. A more reliable age was obtained on volcaniclastic rocks of the Red Hill Volcanics and indicated deposition at 3009 ± 4 Ma (GSWA 141936, Nelson, 1998g). The maximum depositional age of the Croydon Group in the Mallina Basin was once interpreted to be 2997 ± 20 Ma (GSWA 118969, Nelson, 1997e). However, this result was obtained on a sandstone within the central part of the basin succession, leaving open the possibility that the maximum age of the lower part of the succession is greater than c. 2997 Ma. In this Report the maximum depositional age of the Mallina Basin is interpreted to be c. 3015 Ma based on apparently conformable relations with the Gorge Creek Group across a transition sequence in the central part of the basin (Figs 25 and 27).

Gorge Creek and Whim Creek Basins

The Whim Creek Group, which is restricted to the area northwest of the Loudens Fault (LF), overlies the Cleaverville Formation (Gorge Creek Group) across an angular unconformity (Hickman, 2002). This relationship contrasts with that between the Cleaverville Formation and the overlying Croydon Group in the central section of the Mallina Basin where no angular unconformity has been recorded (Fig. 25). Either the unconformity northeast of the LF reflects localized folding adjacent to the SSZ (active between 3015 Ma and 3010 Ma) or the LF has juxtaposed basins that were formed some considerable distance apart. In the latter case, the unconformity provides evidence of substantial strike-slip displacement along the LF after 3015 Ma. This is consistent with a number of previous interpretations (Barley, 1987; Krapez and Barley, 1987; Krapez, 1993; Krapez and Eisenlohr, 1998).

Gorge Creek and Mallina Basins

Southeast of the LF, in the central section of the Mallina Basin, the Croydon Group is apparently conformable with the Cleaverville Formation across a sedimentary transition sequence (Fig. 25). Units of BIF and chert correlated with the Gorge Creek Group are stratigraphically overlain by an alternating succession of sandstone, iron-rich shale, BIF, and chert in the lower part of the Croydon Group. On the southeast margin of the Mallina Basin the Cleaverville Formation is unconformably overlain by the Croydon Group (Fitton et al., 1975; Smithies and Farrell, 2000), but this margin is close to the TTSZ, which was active between c. 2955 Ma and c. 2940 Ma. The interpretation in this Report is that the conglomerates within the Mallina Basin along its southeast margin are younger than the lower succession of the Croydon Group in the central part of the basin.

Whim Creek and Mallina Basins

Early interpretations of the Mallina Basin described it as a deep-water trough or 'geosyncline' positioned between the volcanic succession of the Whim Creek greenstone belt to the northwest and granite–greenstones to the southeast (Ryan, 1964, 1965; Ryan and Kriewaldt, 1964). In this 'geosynclinal' model the sedimentary succession of the Mallina Basin was interpreted to be the same age as the adjacent volcanic successions. Subsequently, the sedimentary succession of the Mallina Basin was interpreted to be older than the volcanic succession of the Whim Creek Basin (Hickman, 1983, 1990; Barley, 1987; Krapez and Barley, 1987; Krapez, 1993; Barley et al., 1994; Krapez and Eisenlohr, 1998). There were four lines of evidence that led to this conclusion (Hickman, 1983): 1) the thick clastic sedimentary Mallina Basin succession was considered to be the upper part of the pre-3000 Ma Gorge Creek Group, and to conformably overlie the Cleaverville Formation in the central part of the group; 2) the Mallina Basin succession was deformed by major tight northerly trending folds that are not present in the adjacent Whim Creek Basin and which, based on Pb–Pb and Rb–Sr isotopic data from Oversby (1976), were interpreted to have formed during a regional metamorphic event at c. 2950 Ma (Hickman, 1983); 3) the Satirist Monzogranite, which intrudes the Mallina Basin was dated by the Rb–Sr method at c. 3040 Ma (Leggo et al., 1965); and 4) the Whim Creek Basin was interpreted to contain a conformable succession from the Warambie Basalt to the Rushall Slate (Fitton et al., 1975), which had unpublished Rb–Sr data, indicating deposition at between 2700 Ma and 2500 Ma (JR de Laeter, personal comm., in Hickman, 1983). Thus, the conclusion reached was that the Mallina Basin was older than c. 2950 Ma and the Whim Creek Basin was younger than this.

Krapez and Eisenlohr (1998) used sequence stratigraphy to estimate the age of the clastic succession in the Mallina Basin succession at between 3089 Ma and 2997 Ma. However, the only geochronological constraint was provided by SHRIMP U–Pb detrital zircon ages (GSWA 118969, Nelson, 1997e) that they interpreted to indicate the approximate minimum age of deposition. Smithies et al. (1999) also used zircon U–Pb geochronology and concluded that the sedimentary succession of the central Mallina Basin is essentially the same age as the volcanic succession of Whim Creek greenstone belt. Their evidence was that the Red Hill Volcanics

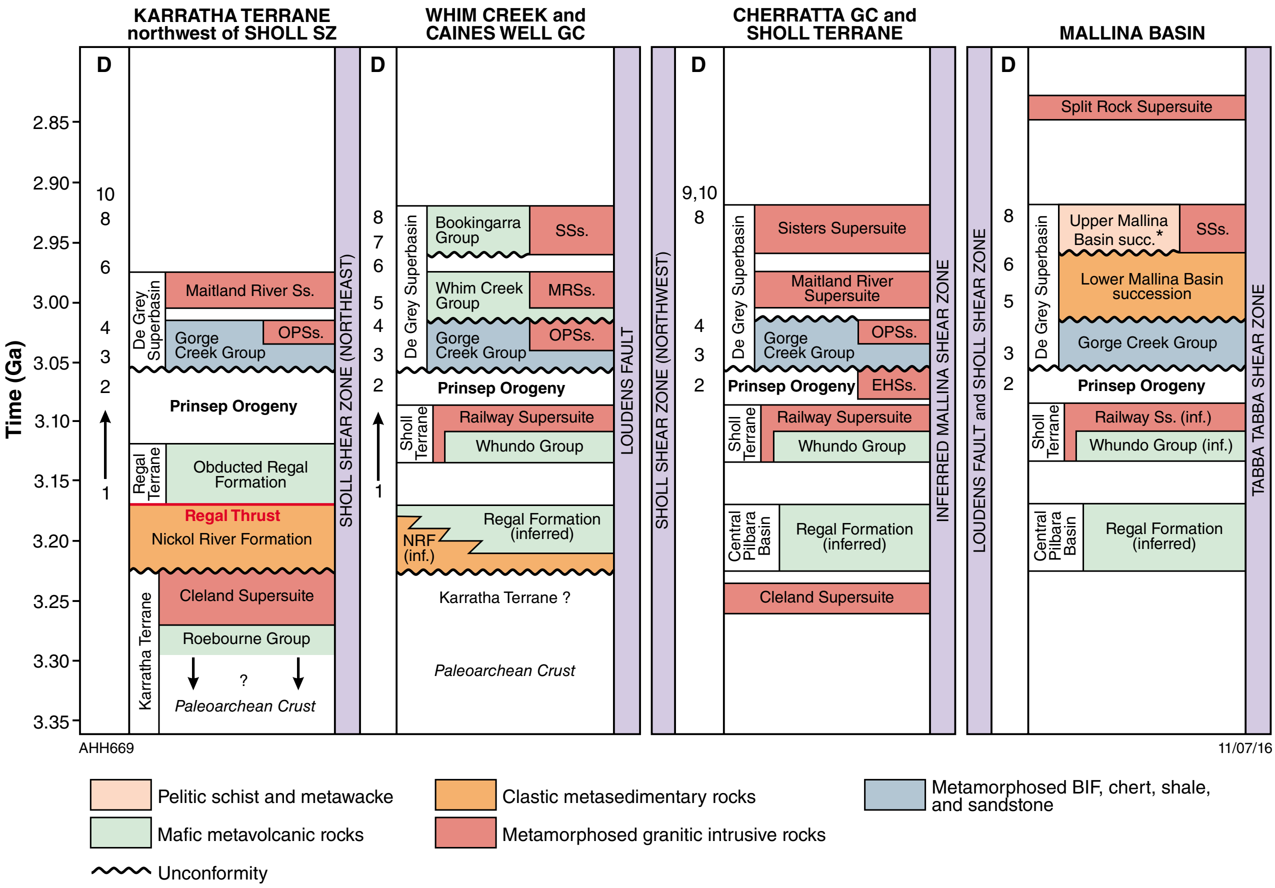


Figure 27. Time–space plot summarizing the stratigraphy and structural history of the four major fault-bounded terranes and basins of the northwest Pilbara Craton. Differences indicate significant strike-slip displacements along the major faults and regional variability of granitic intrusion, volcanism, and sedimentation following the Prinsep Orogeny (i.e. within the De Grey Superbasin). Abbreviations: GC, Granitic Complex; Ss, Supersuite; SSs, Sisters Supersuite; OPSs, Orpheus Supersuite; MRSs, Maitland River Supersuite; EHSs, Elizabeth Hill Supersuite; D, Deformation event. (* = succession)

had been dated at 3009–2991 Ma (Barley et al., 1994; GSWA 141936, Nelson, 1998g) and that zircon U–Pb dating in the Mallina Basin indicated a depositional age range spanning 3015–2940 Ma. Huston et al. (2000) estimated the maximum depositional age of the central Mallina Basin to be c. 3000 Ma. However, Smithies et al. (2001b) and Van Kranendonk et al. (2002, 2006) revised this conclusion, such that the sedimentary succession of the central part of the Mallina Basin was interpreted to span only 2970–2940 Ma. They interpreted that the c. 2990 Ma detrital zircons in the sedimentary succession indicated only the maximum depositional age of the sandstone dated, and that the depositional age was considerably younger than this. In support of the revised age interpretation, Smithies et al. (2001b) noted that some of the c. 2950 Ma granites that intruded the basin succession contain xenocrystic zircons dated at c. 2970 Ma. They argued that if the c. 2970 Ma zircons came from the lower Mallina Basin this must be c. 2970 Ma or younger. However, several other granites which intrude the Mallina Basin contain xenocrystic zircons dated at between 3015 Ma and 2997 Ma (GSWA 118967, Nelson, 1997d; GSWA 142889, Nelson, 1999g; GSWA 142934, Nelson 2000b; GSWA 142935, Nelson 2000c; GSWA 160744, Nelson, 2001f). Ages of xenocrystic zircon ages within the intrusive granites do not necessarily reflect the age of the lower part of the Mallina Basin succession because they might equally well have been recycled from detritus in younger sedimentary rocks with provenances outside the basin. For example, the c. 2970 Ma zircons might have originated from the Whim Creek greenstone belt where the Kialrah Rhyolite contains a large zircon group dated at c. 2975 Ma (GSWA 144261, Nelson, 1998v).

The interpretation that the maximum depositional age of the Croydon Group in the central Mallina Basin is c. 2970 Ma was based on (a) correlation of the Constantine Sandstone (central Mallina Basin) with the Cistern Formation (Whim Creek greenstone belt) and (b) an interpreted maximum depositional age of the Cistern Formation of c. 2978 Ma (GSWA 142949, Nelson, 2000m). However, the c. 2978 Ma age was calculated using a grouping of 22 detrital zircons spanning 2999–2956 Ma. This calculation did not take into account the main provenance of the Cistern Formation, which included the 3009–2991 Ma Red Hill Volcanics and the 3006–2982 Ma Maitland River Supersuite.

For geological reasons it can be argued that detrital zircons older than 2982 Ma should have been excluded from the calculation. Huston et al. (2002a) focused on the eight youngest concordant zircons in the sample to reinterpret the maximum depositional age as being 2964 ± 6 Ma.

Correlation of the Cistern Formation and Constantine Sandstone was key to the interpretation by Van Kranendonk et al. (2006) that the Constantine Sandstone is younger than c. 2970 Ma. This correlation also led these authors to move the Cistern Formation (and all overlying formations in the Whim Creek greenstone belt) from the Bookingarra Group (Pike, 2001; Pike et al., 2002; Van Kranendonk et al., 2002) into the Croydon Group of the Mallina Basin. However, several lines of evidence argue against the Cistern Formation – Constantine Sandstone correlation, and therefore for reinstatement of the Bookingarra Group (as redefined in Appendix 2):

1. Structural and geochronological data indicate there is a c. 2955 Ma unconformity within the Mallina Basin succession (Huston et al., 2000; Smithies and Farrell, 2000; Smithies et al., 2001b, 2002; Van Kranendonk et al., 2002; see Unconformity within the central Mallina Basin). Importantly, the Constantine Sandstone of the type area (Croydon Homestead, Fitton et al., 1975) is positioned stratigraphically beneath this unconformity. An unconformity underlies the Cistern Formation in the Whim Creek greenstone belt (Pike et al., 2002; see Unconformity with the northwest Mallina Basin), and in this Report this unconformity is also interpreted to be c. 2955 Ma. Both unconformities were most likely consequences of major deformation at the beginning of the 2955–2919 Ma North Pilbara Orogeny, and may therefore be parts of a regional unconformity in the northwest Pilbara Craton. Formations on opposite sides of a regional unconformity cannot be correlated.

2. The successions containing the Cistern Formation and Constantine Sandstone are separated by the subvertical LF over a distance of at least 100 km (Fig. 7) and there is no visible stratigraphic or structural continuity across this fault (Fig. 25). Immediately northwest of the LF, the Louden and Mount Negri Volcanics (members of the 'Bookingarra Formation' according to Van Kranendonk et al. (2006), but here reclassified as formations of the Bookingarra Group [Appendix 2]) form a succession over 2 km thick and are both abruptly terminated by the LF. Immediately southeast of the LF, the Mallina Formation, which is between 2.5 km and 5 km thick, is likewise terminated abruptly by the fault. Vertical movement on the LF cannot explain these differences because the successions above and below (a) the Louden and Mount Negri Volcanics, and (b) the Mallina Formation are very different (Figs 25 and 27). Therefore, the most likely explanation is major post-2950 Ma strike-slip displacement on the LF. This conclusion suggests considerable separation of the depositional basins during deposition of the c. 2955 Ma Cistern Formation.

3. There are major differences in the structural geology of the Whim Creek greenstone belt and the central Mallina Basin (Figs 25 and 27). Deformation events recorded in the Whim Creek greenstone belt are D_4, D_{5W}, D_7, and D_8, whereas structures in the Mallina Basin are assigned to D_{5M}, D_6, and D_8. As explained later in this Report (Tectonic events of the Whim Creek and Mallina Basins), D_{5W} and D_{5M} might not be the same age. Different structural histories suggest that the Whim Creek greenstone belt and the Mallina Basin were not adjacent prior to D_8 (c. 2940 Ma).

4. The predominant facies of the Cistern Formation is very different from that of the Constantine Sandstone. The Cistern Formation is composed of volcaniclastic conglomerate, volcanic breccia, and various facies of sandstone derived from erosion of the Whim Creek Group and the Maitland River Supersuite. Pike et al. (2006) interpreted much of the Cistern Formation to have been deposited in subaerial and alluvial environments. By contrast, the Constantine Sandstone is composed of siliciclastic rocks, principally quartzite, wacke, arkosic sandstone, and shale, deposited in relatively deep water (Eriksson, 1982, Smithies et al., 1999).

5. The Cistern Formation and the immediately overlying Rushall Slate were deposited in small isolated sub-basins (Pike at al., 2006). The Cistern Formation has a restricted distribution within the Whim Creek greenstone belt, and is closely associated with c. 2950 Ma VHMS Cu–Zn–Pb mineralization around individual volcanic centres (Collins and Marshall, 1999a,b). This association and the restricted distribution of the Cistern Formation is in marked contrast to the Constantine Sandstone, which comprises extensive submarine fan deposits that are unrelated to felsic volcanism.

6. U–Pb zircon geochronology indicates that the two formations had different provenances. Detrital zircon grains in two samples of Constantine Sandstone are dominated by age groups between 3631 and 3155 Ma (GSWA 142942, Nelson, 2000h; GSWA 142943, Nelson, 2000i). This suggests derivation of detritus from erosion of the EPT southeast of the Mallina Basin (Smithies et al., 1999). There are no detrital zircons from the Whim Creek Group in these samples, despite this group now being immediately adjacent across the LF. By contrast, almost all zircon grains in the Cistern Formation are younger than 3025 Ma, and most are clustered around 2970 Ma (GSWA 142949, Nelson, 2000m). These ages indicate that the majority of zircons in the Cistern Formation were derived from the underlying Red Hill Volcanics and Maitland River Supersuite.

7. In the type area of the Constantine Sandstone, in the western part of the central Mallina Basin, mapping indicates a sedimentary transition from the underlying Gorge Creek Group (Fig. 25). This suggests that the maximum depositional age of the lower part of the Mallina Basin succession, including the Constantine Sandstone in the type area around Croydon Homestead, is c. 3015 Ma.

Relations within the Whim Creek greenstone belt

The 3010–2990 Ma Whim Creek Group is composed of two volcanic formations, the Warambie Basalt and the Red Hill Volcanics, outcrops of which are restricted to the Whim Creek greenstone belt (Fig. 7). Prior to the work of Pike et al. (2002) the Whim Creek Group included formations stratigraphically overlying the Red Hill Volcanics (Smithies, 1997). In ascending order, these were the Cistern Formation, Rushall Slate, Louden Volcanics, Mount Negri Volcanics, and Kialrah Rhyolite. However, Pike (2001) and Pike et al. (2002) recognized an unconformity between the Red Hill Volcanics and the Cistern Formation, and therefore assigned the Cistern Formation and overlying units to a different group. This overlying succession (Cistern Formation\Rushall Slate\Louden Volcanics\Mount Negri Volcanics\Kialrah Rhyolite) was named the 'Bookingarra Group'. However, based partly on an interpreted Cistern Formation – Constantine Sandstone correlation (now rejected for reasons given above), Van Kranendonk et al. (2006) assigned all these formations to the Croydon Group of the Mallina Basin.

Based on galena Pb–Pb ages for VHMS mineralization within the Cistern Formation and overlying Rushall Slate, the formation's true depositional age is likely to be close to 2950 Ma (Huston et al. 2000). This is the interpretation favoured in this Report, which means that the Cistern Formation is approximately 50 million years younger than the Whim Creek Group and the lower part of the Croydon Group in the central Mallina Basin.

Supersuites

Deposition of each of the three basins of the De Grey Superbasin in the northwest Pilbara was accompanied by intrusion of an igneous supersuite (Fig. 6), i.e. in the Gorge Creek Basin, the Orpheus Supersuite comprises ultramafic–mafic and felsic intrusions; in the Whim Creek Basin, the Maitland River Supersuite is composed of tonalite, granodiorite, and monzogranite; and in the Mallina Basin, the Sisters Supersuite comprises a wide variety of ultramafic–mafic and felsic intrusions. The Orpheus and Sisters Supersuites each comprise intrusions ranging in composition from peridotite to granite and in size from large layered ultramafic–mafic or granitic intrusions to small dykes and sills. Although all three supersuites post-date the greatest sinistral strike-slip movement on the SSZ and therefore outcrop on both sides of it, each is concentrated in a different area. The Orpheus Supersuite was intruded along an approximately 40 km-wide belt straddling the SSZ, whereas the slightly younger and more voluminous Maitland River Supersuite occupies a >200 km-long, >50 km-wide belt, mainly northwest of the SSZ (Fig. 4). Geophysical data indicate that the Maitland River Supersuite may extend much farther offshore, possibly over 100 km to the northwest margin of the craton. By contrast, all intrusions within the Mallina Basin belong to either the Sisters Supersuite or the Split Rock Supersuite. Unlike the Orpheus and Maitland River Supersuites, the Sisters Supersuite extends into the EPT. The broadly linear distribution of the Orpheus, Maitland River, and Sisters Supersuites indicates a general trend of southeast migration of magmatic events from c. 3023 Ma to c. 2919 Ma (Fig. 26). If the 3006–2982 Ma Maitland River Supersuite was intruded in a continental arc southeast of a northwest subduction zone (Blewett, 2002; Smith, 2003), the broadly contemporaneous volcanic formations of the Whim Creek Group (see below) were most likely part of this arc. Given that the Croydon and Whim Creek Groups include rocks of similar age (discussed above), the Mallina Basin would be a back-arc basin.

Gorge Creek Basin

The Gorge Creek Basin (Figs 4 and 5) is composed of sedimentary rocks of the Gorge Creek Group, and the basin was intruded by contemporaneous sills and stocks of dacite, rhyolite, and granophyre of the Orpheus Supersuite. Associated felsic volcanism is established by the local occurrence of flow-banded rhyolite, thin beds of felsic tuff, and felsic volcaniclastic sedimentary rocks within the group. The larger intrusions of the Orpheus Supersuite are the Andover Intrusion, Mount Gregory Monzodiorite, Black Hill Well Monzogranite, Forestier Bay Granodiorite, and South Whundo Monzogranite. The group and the supersuite are closely related and are therefore described consecutively in this Report.

Gorge Creek Group

Development of the Gorge Creek Basin and deposition of the Gorge Creek Group is interpreted to have resulted from crustal relaxation, extension, and minor rifting following the 3068–3066 Ma Prinsep Orogeny. For this reason, the earliest deposition of the group across the northern Pilbara Craton might have commenced as early as c. 3066 Ma, but there is no geochronology on the lower part of the Gorge Creek Group to confirm this. In the east Pilbara, the Gorge Creek Group consists of three formations: a basal unit of metamorphosed sandstone and conglomerate named the Farrel Quartzite (Van Kranendonk et al., 2006); a central unit of chert, BIF, carbonaceous shale, siltstone, and sandstone, named the Cleaverville Formation (Ryan and Kriewaldt, 1964; Hickman, 1983, 1997a); and an upper unit of conglomerate, sandstone, and shale, assigned to the Cundaline Formation (Williams, 1999). However, the depositional environment and large aerial extent of the group suggest that its stratigraphic succession must vary laterally. In most areas of the northwest Pilbara, only the Cleaverville Formation can be identified although sandstone and rare units of conglomerate are intercalated with BIF, chert, and shale. Pilbara-wide, the basin overlies an erosion surface that developed across the WPS and the EPT, and it basal unconformity establishes a maximum age for terrane accretion. Deposition in the Gorge Creek Basin was initially shallow-water and at Point Samson the Cleaverville Formation includes evaporite and fluviatile deposits (Sugitani et al., 1998, 2003) (see below). A maximum thickness of the group of c. 2000 m indicates relatively minor basin subsidence. Lateral facies changes in the east Pilbara suggest that the basin was broken by islands, most likely over local horst blocks of the EPT. In the northeast Pilbara, there is evidence of erosion of the Cleaverville Formation during deposition of the overlying Cundaline Formation. The significance of this is uncertain. It may have been due to local areas of uplift or, as Horwitz (1990) interpreted, this disconformity may have more regional significance. Horwitz (1990) correlated the sedimentary rocks of the Cundaline Formation with the central Mallina Basin succession (Croydon Group).

Geochronology

The approximate depositional age of the Cleaverville Formation on the western margin of the EPT is c. 3016 Ma, based on SHRIMP U–Pb zircon dating of felsic volcanoclastic sandstone at Nunyerry Gap (GSWA 142842, Nelson, 1998o). This sandstone is interpreted to be part of the Cleaverville Formation and the ages of detrital zircons within it indicate derivation of detritus from the 3068–3066 Ma Elizabeth Hill Supersuite, the 3130–3093 Ma ST, and from 3228–3160 Ma rocks of either the Soanesville Group or the Nickol River Formation. These ages are consistent with post-3066 Ma erosion of the northeast-trending orogenic belt formed by collision of the EPT and WPS. The youngest zircon group, dated at c. 3016 Ma, probably represents ash from felsic volcanism related to the Orpheus Supersuite.

The Gorge Creek Group of the central and eastern parts of the east Pilbara has proved very difficult to date, most likely due to the absence of contemporaneous felsic volcanism or any potential felsic source stratigraphy between c. 3220 Ma and c. 3020 Ma. SHRIMP U–Pb dating of detrital zircons in the Farrel Quartzite was undertaken on four quartzite samples from widely spaced localities in the northeast Pilbara. The crystallization ages of the individual zircon grains (within 10% of concordia) vary between 3596 ± 7 Ma (GSWA 143995, Nelson, 1998u) and 3343 ± 13 Ma (GSWA 143994, Nelson, 1998t), indicating erosion of the EPT and older crust.

In the northwest Pilbara, where the Gorge Creek Basin was developed on terranes with a younger evolutionary history, the ages of zircons in four samples from the Cleaverville Formation dated by GSWA are about 300 million years younger than in the east. With the exception of one anomalously old grain dated at 3461 ± 8 Ma, the zircon ages range between 3287 ± 17 Ma (GSWA 127330, Nelson, 1998b) and 2999 ± 10 Ma (GSWA 136899, Nelson, 1998f). However, Aihara et al. (2012) reported that the formation contains detrital zircon grains with ages between c. 3700 Ma and c. 3200 Ma. The very old detrital zircons are consistent with the ages of detrital zircons in the Gorge Creek Group in the east Pilbara, and provide further evidence that the Cleaverville Formation of the northwest Pilbara was deposited in a continental basin. Aihara et al. (2012) also reported that a group of nine zircons had been dated at 3108 +14/-7 Ma and interpreted this as the depositional age. However, the present interpretation is that the c. 3108 Ma zircons are also detrital and were very probably derived from erosion of the deformed and uplifted ST, following the 3068–3066 Ma Prinsep Orogeny. For the four northwest Pilbara samples of the Cleaverville Formation dated by GSWA, half of the 76 zircon grains providing analytical data within 10% of concordia have ages falling in the 3025–3010 Ma range of the Orpheus Supersuite. The geochronology supports the interpretation that the Cleaverville Formation and the Orpheus Supersuite were closely contemporaneous, which may explain local hydrothermal activity in the Cleaverville Formation reported by Kato et al. (1998).

At Mount Ada, east of Woodbrook and immediately south of the SSZ, the Cleaverville Formation comprises BIF, jaspilite, sandstone, and shale intruded by sills of dolerite and granophyre and folded by upright isoclinal folds (D_4). SHRIMP U–Pb geochronology indicates that a 20 m-thick volcaniclastic sandstone within the BIF has a maximum depositional age of 3018 ± 3 Ma (GSWA 142830, Nelson, 1998n). The minimum age of this sandstone is tightly constrained by the Stone Yard Granophyre (Van Kranendonk et al., 2006) dated at 3014 ± 6 Ma (GSWA 127320, Nelson, 1997j), which locally intrudes the Cleaverville Formation. Between 1 and 5 km east of Mount Ada the isoclinally folded Cleaverville Formation is unconformably overlain by the c. 3010 Ma Warambie Basalt of the Whim Creek Group. Southwest of Mount Ada, the Cleaverville Formation is interpreted to unconformably overlie the 3130–3110 Whundo Group, but contacts are concealed by the Fortescue Group. Erosion of the Whundo Group before or during deposition of the Cleaverville Formation is indicated by detrital zircon at Cleaverville (GSWA 127330, Nelson, 1998b) and at Nunyerry Gap (GSWA 142842, Nelson, 1998o).

Cleaverville Formation

The name Cleaverville Formation is applied to a formally defined stratigraphic unit (Ryan and Kriewaldt, 1964; Hickman, 1983, 1997a). However, from 1991 onwards various researchers applied the name 'Cleaverville' to informally name a range of interpreted stratigraphic and tectostratigraphic units in the Roebourne greenstone belt. Kiyokawa (1993) combined the successions of the Cleaverville and Regal Formations to propose the existence of the 'Cleaverville group' (informal name). He also gave new informal names to three formations making up this group. The interpretation of Kiyokawa (1993) was not accepted by GSWA owing to its incompatibility with a range of geological and geochronological data (Appendix 3). Therefore, the informal name 'Cleaverville group' is obsolete, as is the name 'Cleaverville Supersequence' (Krapez, 1993; Krapez and Eisenlohr, 1998; Smith et al., 1998).

The name Cleaverville was also used to refer to various proposed tectonostratigraphic units, such as the 'Cleaverville accretionary complex' (Isozaki et al., 1991), 'Cleaverville unit' or 'Cleaverville supracrustal unit' (Kiyokawa, 1993; Kiyokawa and Taira,1998; Krapez and Eisenlohr, 1998; Smith et al., 1998), 'Cleaverville complex' (Ohta et al., 1996), and 'Cleaverville–Roebourne Supercomplex' (Kiyokawa and Taira,1998; Kiyokawa et al., 2002). North of the SSZ, these and others units have been replaced in GSWA publications by the RT and KT, the Nickol River and Gorge Creek Basins (Fig. 5), and the Dampier and Harding Granitic Complexes (Fig. 7).

The Cleaverville Formation is exceptionally well exposed in the type area at Cleaverville Beach (Fig. 18), at Point Samson, and northwest of Roebourne, where it is a 1500 m-thick succession of BIF, chert, shale, and sandstone, locally intruded by mafic and felsic sills (Hickman, 1997a). Magnetite, hematite, and goethite make up about 50% of the BIF, which in weathered surface exposures is typically almost black, locally with red jaspilitic layers. The layers rich in iron minerals alternate with layers of silica at 1–10 mm intervals. Such BIF units are typically 50–200 mm thick and are interbedded with mudstone or shale units of similar thickness. In drillcore and some creek exposures the shale is black or dark grey, carbonaceous and pyritic, but in most surface exposures it is pale grey, brown, or cream to off-white due to weathering (Fig. 18b). Black pyritic shale indicates deposition in a low-energy, reducing environment, whereas siltstone and sandstone units suggest alluvial or submarine fans. Thus, the dominant depositional environment was low energy, but there were periodic influxes of detrital material, possibly carried by turbidity currents or in migrating channels of fans. Near Point Samson Sugitani et al. (1998) presented sedimentological and geochemical evidence that BIF in that area was deposited in shallow water. Sedimentary structures include intraformational clasts, irregular fenestrae, edgewise breccia, and probable gypsum and halite pseudomorphs. Anomalously low Al_2O_3/TiO_2 values (<4), resulting from an influx of Ti-bearing minerals, such as rutile were interpreted as evidence of extensive chemical weathering in the source regions. This is consistent with slow deposition in a relatively shallow-water basin spanning a period of up to 50 million years.

Grey white banded chert makes up much of the Cleaverville Formation in the Point Samson – Wickham – Roebourne outcrops and is interpreted to be silicified fine-grained clastic sedimentary rocks, in some instances containing felsic ash. Alternating white and grey layers, typically 1–10 mm thick, generally have sharply defined contacts, but gradual colour transitions locally suggest graded bedding of fine-grained clastic rocks, possibly deposited from turbidity currents, resulting from extensional faulting within the basin. Thin units of metamorphosed sandstone and siltstone in association with BIF and ferruginous chert outcrop 5 km southwest of Wickham, 3 km south of Wickham, and on the southeast side of the Cleaverville promontory. Southwest of Wickham, felsic volcaniclastic sandstone in the centre of the formation was dated at 3015 ± 5 Ma (GSWA 136899, Nelson, 1998f). This sandstone is conformably underlain by a thick succession of metamorphosed shale and minor siltstone with interbedded chert and BIF.

The Cleaverville Formation of the Cleaverville area occupies the core of a faulted syncline northwest from the Prinsep Dome (Fig. 9). This syncline has been traced to Devil Creek, 80 km west-southwest of Cleaverville. Northeast from Cleaverville, aeromagnetic data indicate that the BIF in the Cleaverville Syncline is folded around the northeast plunging Prinsep Dome to outcrop on the west limb of the Roebourne Synform southwest of Wickham (Figs 8 and 9). All three northeast-trending folds are c. 2940 Ma transpressional structures (D_8) formed during the latter stages of the North Pilbara Orogeny and related to strike-slip movement on the SSZ (Hickman, 2001a,b). Kiyokawa (1993) and Kiyokawa and Taira (1998) did not interpret the Cleaverville Formation to be folded around the Prinsep Dome and Roebourne Synform, and therefore interpreted the BIF at Wickham to be a different unit from the BIF at Cleaverville. The Cleaverville–Wickham correlation is very important for two reasons. Firstly, it confirms the age of the Cleaverville Formation at Cleaverville as being approximately 3020 Ma, as indicated by a SHRIMP maximum depositional age of 3022 ± 12 Ma (GSWA 127330, Nelson, 1998b). This depositional age precludes stratigraphic continuity of the Cleaverville Formation with the c. 3200 Ma Regal and Dixon Island Formations. Secondly, it greatly increases the minimum area of deposition of the Cleaverville Formation from about 10 km^2 at Cleaverville and Dixon Island to over 1000 km^2 in the Roebourne and Devil Creek greenstone belts.

BIF and chert immediately beneath the Constantine Sandstone in the central part of the Mallina Basin (Fig. 25) are correlated with the Cleaverville Formation (Fitton et al., 1975; Smithies and Farrell, 2000; Smithies and Hickman, 2003, 2004), supporting stratigraphic continuity of the Cleaverville Formation between the northwest Pilbara and the western margin of the east Pilbara. Large outcrops of the Cleaverville Formation are exposed in the Ord Range (Hickman, 1980a, 1983; Van Kranendonk and Smithies, 2006) revealing the locally upfolded basement to the Croydon Group in the Mallina Basin. As at Mount Ada, 15 km south of Roebourne, the BIF is tightly to isoclinally folded. Mapping by Smithies (2002b) indicated that the lower part of the succession in the Ord Range consists of sandstone and siltstone overlain by grey and white chert and dark grey chert, most of which represents silicified shale. This lower succession, which may be equivalent to the Farrel Quartzite of the east Pilbara, is overlain by a thick unit of BIF and variably silicified black shale that passes upwards into alternating units of chert, shale, and BIF. Complex folding and faulting prevents accurate measurement of the original stratigraphic thickness of this succession, but the present estimate is between 1000 and 2000 m. Another area thought to contain the Cleaverville Formation lies between 10 and 60 km south of Port Hedland. This area forms part of the 600 km^2 Bullen Hill greenstone belt (new name, Fig. 6), which includes BIF and chert at Bullen Hill. The greenstone belt is very poorly exposed, but the predominantly mafic volcanic succession appears to overlie the Cleaverville Formation and underlie the Mallina Formation.

Orpheus Supersuite

The 3023–3012 Ma Orpheus Supersuite is composed of intrusions varying greatly in size and with a wide range of compositions from peridotite to granite. Geochronology indicates that its intrusion into the WPS was contemporaneous with deposition of the c. 3020 Ma Cleaverville Formation. The supersuite was intruded along an east-northeast-trending belt on both sides of the SSZ, probably over a strike length of at least 150 km. However, no other igneous rocks of this age have been identified in other parts of the northern Pilbara Craton. The fact that the Orpheus Supersuite is only marginally older than the oldest components of the 3006–2982 Ma Maitland River Supersuite, and the observation that both supersuites have a very similar distribution, suggests that the Orpheus Supersuite was a precursor to the Maitland River Supersuite. However, during the brief interlude between the supersuites there was a significant deformation event (D_4). This event of uplift, folding, and erosion is revealed by a regional angular unconformity between the Cleaverville Formation and the c. 3010 Ma Warambie Basalt (lower Whim Creek Group). This unconformity is exposed at a number of localities up to 20 km east from Mount Ada and south of the SSZ. This deformation event is evidence of renewed northwest–southeast or north–south convergence following the 3068–3066 Ma Prinsep Orogeny.

Andover Intrusion

Early GSWA mapping of the northwest Pilbara revealed the existence of large ultramafic–mafic layered intrusions near Roebourne (Ryan, 1965) and at Munni Munni Creek, 50 km southwest of Roebourne (Williams, 1966). The complex near Roebourne, described under the name 'Mount Hall – Carlow Castle Complex' by Hickman (1983), has since been formally named as the Andover Intrusion (Hickman, 1997b). This intrusion was initially inferred to belong to the same intrusive suite as the Munni Munni Intrusion (Hoatson et al., 1992) but subsequent geochronology (see below) has established that it forms part of the Orpheus Supersuite (Van Kranendonk et al., 2006).

The Andover Intrusion is the largest layered intrusion of the northern Pilbara Craton and occupies an area of about 200 km^2 southeast and southwest of Roebourne. Detailed studies were undertaken by Wallace (1992a,b), Hoatson et al. (1992), and Hoatson and Sun (2002). The minimum ages of particular sections of the Andover Intrusion are constrained by the crystallization ages of intrusive felsic units such as the Mount Gregory Monzodiorite, dated at 3016 ± 4 Ma (GSWA 168936, Nelson, 2001j) and the Rea Dacite, dated at 3014 ± 2 Ma (GSWA 118979, Nelson, 1997i). Dating of a leucogabbro unit on the northern margin of the intrusion indicated an age of 3007 ± 17 Ma (GSWA 178170, Wingate et al., 2015). This younger age for the leucogabbro supports field evidence that the intrusion is a complex body formed by more than one magmatic event (Hickman and Smithies, 2001). The maximum age of the intrusive complex is poorly constrained. It intrudes the c. 3280 Ma Ruth Well Formation and the c. 3265 Ma Karratha Granodiorite on the southeast limb of the Roebourne Synform. Nd T_{DM} model ages for the intrusion are in the range of 3300–3200 Ma (Hoatson et al, 1992), which is c. 200 million years younger than Nd T_{DM} model ages calculated for the Ruth Well Formation and the Karratha Granodiorite (Smithies et al., 2007a).

The lower ultramafic zone of the Andover Intrusion consists of rhythmically layered, peridotite, lherzolite, olivine websterite, clinopyroxenite, olivine orthopyroxenite, orthopyroxenite, and websterite with minor anorthosite, leucogabbro, and gabbro (Wallace, 1992b). Above this, there is a gabbroic zone composed of gabbro, leucogabbro and anorthosite. From GSWA mapping (Hickman, 2000) the intrusion appears be a lopolith or funnel-shaped body, up to 3 km thick, in which the layering of the lower ultramafic zone is partly obscured by late-stage discordant and irregular intrusion of gabbro and leucogabbro. It is concluded that the main body of the Andover Intrusion was emplaced in two or more intrusive events between 3023 and 3012 Ma, although some mafic intrusive rocks within the intrusion might be younger than this.

Granitic intrusions

Granitic intrusions of the Orpheus Supersuite are dated between 3025 and 3010 Ma. The largest intrusion is the Black Hill Monzogranite (Van Kranendonk et al., 2006), which intrudes the south and northeast margins of the Andover Intrusion and is interpreted to underlie an area of approximately 200 km^2 in the western part of the Harding Granitic Complex (Hickman et al., 2006b). In the type area, 3.5 km southeast of Black Hill Well, the rock is a weakly foliated monzogranite containing leucocratic veins. SHRIMP dating indicates two granitic components, the oldest with a crystallization age of 3018 ± 19 Ma and a younger granitic rock with an age of 2970 ± 5 Ma (GSWA 142430, Nelson, 1999b). Allowing for some isotopic disturbance associated with an adjacent mylonite zone, the younger group of zircons suggests the granitic rock of the Orpheus Supersuite was later intruded by granitic magma of the Maitland River Supersuite. Another intrusion of the Orpheus Supersuite is the Mount Gregory Monzodiorite, dated at 3016 ± 4 Ma (GSWA 168936, Nelson, 2001j). A few younger zircons within the rock indicate veining by fluids related to the 3000–2980 Ma Maitland River Supersuite. The Forestier Bay Granodiorite is an intrusion of granodiorite and monzogranite in the Harding Granitic Complex 20 km northwest of Whim Creek. The outcrop area is immediately north of the SSZ and occupies a total area about 30 km^2, but exposure is limited to rock pavements on intertidal and supratidal flats between the Sherlock River and Balla Balla rivers. The unit is composed of massive to foliated, and locally banded, biotite granodiorite and monzogranite. In some outcrops the rock contains large microcline phenocrysts. SHRIMP dating indicated a crystallization age of 3014 ± 3 Ma (GSWA 118966, Nelson, 1997c), but a small group of <3000 Ma zircon grains suggests addition of material from the Maitland River Supersuite.

Minor felsic intrusions

Greenstone successions adjacent to the Harding Granitic Complex are intruded by numerous minor felsic intrusions of the Orpheus Supersuite. The Snell Well Granodiorite (Van Kranendonk et al., 2006) intrudes the Ruth Well Formation at the western end of the complex. Smith (2003) informally referred to this foliated granodiorite as the 'Costean granodiorite' and she obtained SHRIMP U–Pb zircon crystallization ages of 3013 ± 8 Ma (sample N4097) and 3012 ± 3 Ma (sample N4128). In the same area, Smith (2003) dated small porphyritic dacite intrusions into the Ruth Well Formation at 3016 ± 2 Ma (sample N3188) and 3017 ± 6 Ma (sample N4028). Porphyritic dacite and rhyolite sills within and adjacent to the Andover Intrusion between Ruth Well and Carlow Castle are assigned to the Rea Dacite (Van Kranendonk et al., 2006). The largest outcrop, 5 km southeast of Mount Marie, occupies an area of 10 km^2. Samples at different localities were dated at 3014 ± 2 Ma (GSWA 118979, Nelson, 1997i) and 3023 ± 9 Ma (GSWA 118976, Nelson, 1997h). The Rea Dacite is locally flow-banded, suggesting that part of it may be extrusive, and felsic volcanism probably account for the felsic ash in parts of the Cleaverville Formation. The Mount Wangee Dacite (Van Kranendonk et al., 2006) is a very similar unit to the Rea Dacite but forms sills and small stocks farther to the north, between Mount Wangee and Cleaverville. These felsic intrusions intrude the Ruth Well, Regal, and Cleaverville Formations. The dacite sill at Mount Wangee is probably an apophysis related to the Black Hill Monzogranite, which underlies lagoonal and alluvial deposits immediately to the south. In the type area at Mount Wangee, the dacite was dated at 3021 ± 3 Ma (GSWA 144224, Nelson, 1999j). At Rocky Creek 4 km southeast of Cleaverville, a sill of porphyritic dacite intruding the base of the Regal Formation was dated at 3018 ± 2 Ma (GSWA 127327, Nelson, 1998a). This sill was mapped over a total strike length of 10 km and towards its faulted western end, it becomes thicker and grades into an altered and fractured fine-grained granitic rock.

The Stone Yard Granophyre (Van Kranendonk et al., 2006) is a 2 km-long sill of quartz granophyre that intrudes the Cleaverville Formation 500 m north of Mount Ada. The granophyre was dated at 3014 ± 6 Ma (GSWA 127320, Nelson, 1997j). A c. 3100 Ma xenocrystic zircon grain in the sample was probably derived from underlying rocks of the Whundo Group.

A group of small monzogranite intrusions into the Whundo Group south of Whundo are collectively assigned to the South Whundo Monzogranite (Van Kranendonk et al., 2006). The largest intrusion outcrops over an area of 5 km^2 and Smith et al. (1998) dated a sample (JS35) at 3013 ± 4 Ma.

Tectonic events (D_3 and D_4)

The Gorge Creek Basin developed as a post-collision extensional basin and therefore contained growth faults (Van Kranendonk et al., 2004). Direct evidence of such faults has not been observed in the northwest Pilbara, but many of the dolerite, dacite, and granophyre sills in the Cleaverville Formation were probably intruded along these syndepositional structures. Dating of the dacite and granophyre sills has confirmed that they belong to the 3025–3010 Ma Orpheus Supersuite. The growth faults are designated as D_3 structures as there are soft sediment slump folds and intraformational breccia units in the BIF of the Cleaverville Formation.

In many parts of the northwest Pilbara Craton the Cleaverville Formation is deformed by upright, tight to isoclinal folds assigned to D_4. At Mount Ada, 15 km south of Roebourne, east-southeast-trending upright folds deform the Cleaverville Formation and a sill of the c. 3014 Ma Stone Yard Granophyre (Orpheus Supersuite). The minimum age of these structures is constrained by the fact that they are unconformably overlain by the c. 3010 Ma Warambie Basalt (Whim Creek Group). Other upright east–west-trending tight to isoclinal folds in the Cleaverville Formation are present in the Ord Range. However, the D4 age of the Ord Range folds cannot be confirmed because they are not unconformably overlain by the Whim Creek Group. Instead, the Ord Range folds are unconformably overlain by the Salt Well Member of the Croydon Group, and the age of this unit is probably c. 2950 Ma.

The existence of D_4 deformation immediately prior to deposition of the Whim Creek Group and the east–west trend of the folds indicates that D_3 crustal extension was replaced by north–south or northwest–southeast D_4 compression at c. 3010 Ma.

Whim Creek Basin

The Whim Creek Basin is entirely composed of the Whim Creek Group, which comprises volcanic, intrusive, and volcaniclastic rocks of the Warambie Basalt and Red Hill Volcanics within the Whim Creek greenstone belt. Granitic rocks of the Maitland River Supersuite are closely contemporaneous with the Whim Creek Group, and now form large parts of several granitic complexes along and inland from the northwest Pilbara coast (Hickman et al., 2006a). As discussed below, geochronological data for the Whim Creek Group are far too limited to define a precise age range for its formation, but the best current estimate is between 3010 and 2990 Ma. Data from samples of the Maitland River Supersuite indicate crystallization ages between 3006 and 2982 Ma (Van Kranendonk et al., 2006). Both units outcrop in separate parallel east-northeast-trending belts, with the volcanic rocks of the Whim Creek Group lying immediately southeast of the Maitland River Supersuite. Combining this linear distribution with geochemical evidence, several workers have suggested that the group and the supersuite represent different parts of a continental magmatic arc, generated southeast of a convergent margin (Krapez and Eisenlohr, 1998; Smith et al., 1998; Blewett, 2002; Van Kranendonk et al., 2002; Beintema, 2003; Smith, 2003; Hickman, 2004a, 2012; Pike et al., 2006; Hickman et al., 2010; Hickman and Van Kranendonk, 2012). This interpreted relationship implies that the Whim Creek Basin originally occupied a larger area than is now preserved within the greenstone belt. Geochronology and outcrop relations (Maitland River Supersuite northwest of the Whim Creek Group) indicate a close relationship consistent with intrusive and volcanic sections of an arc.

Whim Creek Group

The Warambie Basalt and Red Hill Volcanics are restricted to the Whim Creek greenstone belt, as are the Cistern Formation, Rushall Slate, Louden Volcanics (redefined as a formation in Appendix 2), Mount Negri Volcanics (redefined as a formation in Appendix 2), and the Kialrah Rhyolite of the overlying Bookingarra Group (redefined in Appendix 2). The Whim Creek greenstone belt is bounded on its southeast side by the LF (Figs 3 and 6). At the northeast end of the Whim Creek greenstone belt the LF branches off the SSZ near Peawah Hill, whereas 15 km south of Whim Creek the LF merges with the Mallina Shear Zone. Aeromagnetic imagery indicates that beyond the Sherlock River the LF continues west as a concealed structure, previously referred to as the Jones River Fault (Hickman, 2004a). However, this fault is now interpreted as the western continuation of the LF. Near Woodbrook, approximately 50 km west of the Sherlock River, this western extension of the LF appears to rejoin the SSZ (Fig. 3).

Prior to 1975, geological investigations of the Whim Creek greenstone belt were focused on potentially economic VHMS mineralization at Whim Creek (Cu–Zn) and Mons Cupri (Cu–Zn–Pb), and ultramafic–mafic, intrusion-hosted mineralization at Balla Balla (V–Ti) and Sherlock Bay (Ni–Cu). The regional stratigraphy of the area was poorly understood at this time, but geological mapping by Fitton et al. (1975) and Hickman (1977) established a stratigraphic succession for the Whim Creek Group that included the Warambie Basalt and the Red Hill Volcanics – 'Mons Cupri Volcanics' prior to Pike et al. (2002) – and the overlying Cistern Formation and Rushall Slate. An ultramafic–mafic volcanic succession overlying the Rushall Slate, now assigned to the Louden Volcanics and Mount Negri Volcanics, was interpreted to be unconformable on this succession, and accordingly was not included in the Whim Creek Group. Subsequent work accepted this stratigraphic division (Barley, 1987; Krapez and Barley, 1987; Krapez, 1993; Smithies, 1996, 1997; Barley, 1997; Krapez and Eisenlohr, 1998; Smithies, 1998a; Smithies et al., 1999; Hickman et al., 2000). However, a detailed investigation of volcanic and sedimentary facies in the eastern half of the Whim Creek greenstone belt led to recognition of an unconformity between the Red Hill Volcanics and the Cistern Formation (Pike, 2001; Pike et al., 2002). This resulted in a redefinition of the Whim Creek Group to exclude the Cistern Formation and all overlying units. This unconformity was accepted by Van Kranendonk et al. (2006) because geochronology indicated a significant age difference between the successions (Red Hill Volcanics 3010–2990 Ma, Cistern Formation <2964 Ma, see Geochronology below). Pike et al. (2006) modified the stratigraphy of Pike et al. (2002) by reassigning widespread intrusive dacite within the greenstone belt to the Mons Cupri Dacite Member of the Red Hill Volcanics. However, the present Report notes that the precise stratigraphic relations between the Red Hill Volcanics and the Cistern Formation remain somewhat ambiguous, in particular with regard to the age and stratigraphic relations of the 'Mons Cupri Dacite Member' (discussed below).

Geochronology

Barley et al. (1994) determined a zircon U–Pb age of 2991 ± 12 Ma for porphyritic dacite of the 'Mount Brown Rhyolite' (obsolete name) at Red Hill. This unit was renamed the 'Mons Cupri Dacite Member' by Pike et al. (2006). They interpreted the dating of this member to also provide an age for its interpreted parent formation the Red Hill Volcanics (Pike et al., 2002). Previously the Red Hill Volcanics had been named the 'Mons Cupri Volcanics' (obsolete name). However, the significance of the c. 2991 Ma age obtained by Barley et al. (1994) was questioned by Smithies et al. (1999) on the basis that the zircon grains in the sample are discordant and that the grains used to calculate the age do not fall within error of a single mean $^{207}Pb/^{206}Pb$ value. Barley et al. (1994) interpreted the relatively large range in the radiogenic $^{207}Pb/^{206}Pb$ ages of the zircon grains to be evidence of early Pb loss, and therefore calculated the published crystallization age using only the three oldest of nine euhedrally zoned zircons. An alternative explanation would be that all the zircons are xenocrysts and that all have sustained different amounts of Pb loss. In this scenario, the crystallization age of the dacite would be less than 2991 Ma. Smithies et al. (1999) sampled the Red Hill Volcanics from a nearby locality containing volcaniclastic sedimentary rock. They reported a SHRIMP U–Pb age of 3009 ± 4 Ma (GSWA 141936, Nelson, 1998g) and interpreted this as the age of deposition. However, Pike et al. (2006) observed that the individual ages of the zircons selected to calculate this c. 3009 Ma age range from c. 3043 to 2995 Ma. At the time the geochronology was undertaken, it was not known that this zircon age range partly coincides with that of the Orpheus Supersuite. This knowledge now raises the possibility that many of the zircons used to calculate the 3009 Ma age were inherited. Based on current geological knowledge, zircons of Orpheus Supersuite age can reasonably be excluded from the age determination with the result that the true depositional age of the volcaniclastic sedimentary rock is between 3007 and 2995 Ma (11 zircons). However, this does not constrain the maximum depositional age of the entire group.

SHRIMP zircon U–Pb dating of a sheared unit of felsic volcaniclastic sandstone, approximately 2 km south-southeast of Whim Creek Hotel, indicated a maximum depositional age of 3016 ± 3 Ma (GSWA 168924, Nelson, 2001h). The sandstone underlies the Rushall Slate and it has been interpreted to belong either to the Red Hill Volcanics (Smithies, 1997) or the Cistern Formation (Pike et al., 2006). The zircon data support derivation of detritus from the Orpheus Supersuite, in which case 3016 Ma is the maximum depositional age of the Red Hill Volcanics.

Warambie Basalt

In the western 40 km of the Whim Creek greenstone belt, from Woodbrook to Red Hill (Hickman et al., 2006b), the basal formation of the Whim Creek Group is the Warambie Basalt, and felsic rocks that could be considered as part of the Red Hill Volcanics are either absent or very thin.

By contrast, in the eastern part of the belt over a strike length of 40 km between Red Hill and Whim Creek, the Whim Creek Group is almost entirely composed of felsic volcaniclastic of the Red Hill Volcanics and intrusive dacite of uncertain age. This major west–east change in the composition of the group coincides with three other stratigraphic changes: 1) the underlying basement rocks change from greenstones of the Whundo Group in the west to the Caines Well Granitic Complex in the east; 2) the Whim Creek Group outcrop is closer to the SSZ in the west; and 3) the lower part of the Whim Creek Group is intruded by mafic and felsic units of the Sisters Supersuite in the east.

Between Mount Ada and Mount Oscar, the stratigraphy of the Warambie Basalt indicates rapid deposition immediately south of the uplifted area along the SSZ. The Warambie Basalt is at least 500 m thick in this area and is chiefly composed of basaltic breccia, hyaloclastite, and vesicular and pillowed basalt, containing several thick intercalations of conglomerate, sandstone, and shale. The conglomerate contains boulders and pebbles of metabasalt and metadacite (most likely derived from erosion of the Roebourne and Whundo groups) and BIF and chert (from the Cleaverville Formation) (Hickman, 2002). Since the minimum depositional age of the Cleaverville Formation is 3015 Ma, this places a maximum depositional age on the Warambie Basalt. Restriction of coarse clastic sedimentary rocks to immediately south of the present shear zone suggests that at c. 3015 Ma the shear zone was expressed as an east–west-trending belt of south facing fault scarps. In this setting, the conglomerate and sandstone were deposited in alluvial fans due to active erosion of the uplifted area to the north. Dolerite intrusions and thick pyroclastic flows suggest extension of the crust with resulting fissure eruptions (Hickman, 2001b) as would take place in either a continental rift basin or a back-arc basin.

In the type area of the Warambie Basalt, around Warambie Homestead (Fitton et al., 1975), the formation is 200 m thick and composed of vesicular, amygdaloidal, and pillowed basalt and basaltic andesite with rare intercalations of chert and thin felsic volcaniclastic rocks. Farther east, near Red Hill, and immediately adjacent to the Caines Well Granitic Complex, the thickness of the Warambie Basalt varies from 40 to 150 m and the formation includes minor intercalations of felsic volcanic rocks that probably represent the early stages of deposition of the Red Hill Volcanics (Smithies, 1998a; Pike et al., 2006). The lower part of the Warambie Basalt in this area includes clast-supported conglomerate and breccia in which granitic and basaltic clasts are equally represented. The matrix of the conglomerate is composed of altered granitic and basaltic grains. The largest granitic boulders are up to 1 m in diameter, and are angular indicating proximal sources. Pike et al. (2006) interpreted the transport of such large granitic boulders to have been by gravitational sliding or rolling rather than suspension within a flow. Many of the basaltic clasts are vesicular and closely resemble the Warambie Basalt that is exposed in outcrops between10 and 40 km west of Red Hill. The Warambie Basalt is thickest in the west, indicating that uplift and erosion of older parts of the formation near Red Hill contributed material to the conglomerate and breccia in its upper part.

Geochemical data for the Warambie Basalt (Pike, 2001; Pike et al., 2006; Smithies et al., 2007a) indicate enrichment in K_2O, Rb, Ba, Th, and Ce with respect to MORB. This could be attributed to assimilation of continental crust in a rift system, but many workers would interpret it as indicating subduction-related basaltic magmas. Smithies et al. (2007a) noted that compared with otherwise similar subduction-related tholeiites, such as those of the Whundo Group, the Warambie Basalt is enriched in Nb and, in particular, Na_2O (up to 4.5 wt%), supporting magma interaction with the underlying continental crust. Negative ε_{Nd} values (Table 5) are consistent with contamination by underlying continental crust or epicontinental sediment. Possible sources for such sediment include the underlying Gorge Creek Group. Published Nd model ages (Table 5) are older than for the Whundo Group, suggesting Paleoarchean crust at depth beneath the Whim Creek Group.

Red Hill Volcanics

Early stratigraphic investigators of the Whim Creek Group (Fitton et al., 1975; Hickman, 1977, 1983) interpreted the Warambie Basalt to be older than but conformable with the felsic volcanic, volcaniclastic, and intrusive rocks of the overlying 'Mons Cupri Volcanics' (obsolete name), renamed by Pike et al., (2002) as the 'Red Hill Volcanics' (current name). Smithies (1998a) described the Whim Creek Group as being mainly composed of the 'Mons Cupri Volcanics' with minor intercalations of the Warambie Basalt, but his mapping area did not extend into the western half of the Whim Creek greenstone belt. As established by later mapping, in the west the Warambie Basalt is at least 500 m thick and felsic volcanic rocks are almost entirely absent. Stratigraphic logs through the eastern half of the Whim Creek greenstone belt (Pike et al., 2006) show that the Warambie Basalt underlies felsic volcaniclastic units of the Red Hill Volcanics. Pike et al. (2006) interpreted arkosic and mafic sandstone units within the basalt to be part of the Red Hill Volcanics. Based on evidence in the western half of the belt (Hickman, 2002), the interpretation preferred in this Report is that they are sedimentary units within the Warambie Basalt.

According to Pike et al. (2002), the Red Hill Volcanics comprise felsic volcaniclastic rocks, volcanic breccia, and subaqueous pyroclastic flows. They regarded intrusive porphyritic dacite, previously referred to as the 'Mount Brown Rhyolite' by Miller and Gair (1975) and the 'Mount Brown Rhyolite Member' by Fitton et al. (1975), as a separate formation, and they referred to this unit as the 'Mons Cupri dacite' (informal name). They described sedimentary facies as including well-bedded sandstone and siltstone, massive volcaniclastic sandstone, polymictic conglomerate, and breccia. At Red Hill, basal sandstone and siltstone of the Red Hill Volcanics overlie the Warambie Basalt and are overlain by rhyodacite pumice breccia (Pike et al., 2006). The basal sandstone and siltstone unit is up to 140 m thick and the pumice breccia is locally 145 m thick.

Pike et al. (2006) described the Mons Cupri Dacite Member as being composed of cream-coloured to grey spherulitic and feldspar-porphyritic dacite. In porphyritic dacite, white oligoclase phenocrysts up to 5 mm in length are set in a groundmass of grey to black microcrystalline quartzo-feldspathic material. Parts of the porphyritic dacite visibly crosscut stratigraphic contacts in the Red Hill Volcanics and Warambie Basalt and are obviously intrusive. Collins and Marshall (1999a) described three main areas where the Mons Cupri Dacite Member intrudes the eastern half of the Whim Creek greenstone belt: Mons Cupri – Whim Creek; Good Luck Well; and Ant Hill – Balla Balla. They interpreted these concentrations of the dacite to be domal structures marking large felsic volcanic centres, each approximately 15–25 km^2 in area. They also observed that VHMS deposits, such as those at Mons Cupri and Whim Creek, are situated on the flanks of these domes. The VHMS deposits are now interpreted to be contemporaneous with the c. 2955 Ma Cistern Formation (see below), in which case the Mons Cupri Dacite Member, which has not been dated, would actually be c. 30 million years younger than the Red Hill Volcanics and not part of the Whim Creek Group. An alternative interpretation is that intrusive dacite units of several ages have been mapped as the Mons Cupri Dacite Member.

Maitland River Supersuite

The 3006–2982 Ma Maitland River Supersuite forms a >200 km-long, >50 km-wide belt of tonalite, granodiorite, and monzogranite intrusions underlying the coastal plain of the northwest Pilbara (Fig. 5). The supersuite is the main constituent of the Dampier Granitic Complex, probably forms much of the Harding Granitic Complex (very poorly exposed), and also outcrops within the Cherratta and Caines Well Granitic Complexes (Fig. 7). Geophysical data suggest that the Orpheus and Maitland River supersuites extend northwest well beyond the Pilbara coast towards the northwest margin of the Pilbara Craton. North of the SSZ, the Maitland River Supersuite intrudes all parts of the WPS, including the 3280–3260 Ma KT. Smith (2003) interpreted the 3000–2980 Ma granitic rocks of the northwest Pilbara to be a tonalite–trondhjemite–granodiorite (TTG) series and GSWA mapping indicated that granodiorite was widespread. The apparent TTG composition and relatively young Nd model ages are consistent with, though not in themselves confirmation of, an interpretation that the Maitland River Supersuite was derived by partial melting of subducted oceanic crust. Smith (2003) interpreted the supersuite to be part of a 3000–2980 Ma continental magmatic arc, along what is now the northwest Pilbara coast. The age and distribution of the supersuite are consistent with it forming part of the same magmatic system as the Whim Creek Group (Hickman, 2004a), which Pike et al. (2002) interpreted to have originated in an ensialic back-arc basin. South of the SSZ, the Maitland River Supersuite intrudes the Whundo Group and the Railway Supersuite of the ST.

Individual intrusions of the Maitland River Supersuite are restricted to particular granitic complexes of the northwest Pilbara Craton. The Dampier Granitic Complex contains three named intrusions of the supersuite: the Erramurra Monzogranite, Miaree Granite, and Hearson Cove Monzogranite. In the western part of the complex, the Erramurra Monzogranite (2982 ± 5 Ma, GSWA 142893, Nelson, 1999i) is a seriate biotite monzogranite, containing mafic schlieren and veins of pegmatite. The intrusion is poorly exposed over an area of 10 km^2. Farther east, between the Maitland River and Karratha Airport, the Miaree Granite is composed of syenogranite and monzogranite that outcrops sporadically along the southeast shores of salt evaporation ponds. Relatively homogeneous monzogranite of the intrusion was dated at 2997 ± 3 Ma (GSWA 136844, Nelson,1998e), but the outcrops include large enclaves of banded granodiorite gneiss that remain undated. The Hearson Cove Monzogranite is intermittently exposed across an area of 600 km^2 on the Burrup Peninsula and Dampier Archipelago. This intrusion is composed of biotite monzogranite containing microcline megacrysts up to 5 cm in length. At Hearson Cove it was dated at 2999 ± 10 Ma (GSWA 178148, Wingate and Hickman, 2009a).

The most extensive units of the Maitland River Supersuite in the Cherratta Granitic Complex are the Toorare Tonalite, Jean Well Granodiorite, Whyjabby Granodiorite, Brill Monzogranite, Waloo Waloo Monzogranite, and Mardeburra Granodiorite. The Toorare Tonalite is composed of grey tonalite gneiss and foliated granodiorite containing layered enclaves of amphibolite and later crosscutting sheets of hornblende–biotite monzogranite. SHRIMP dating has revealed an age range from c. 2995 to 2985 Ma (GSWA 136826, Nelson, 1997k; N3132, Smith, 2003), with later intrusion at c. 2944 Ma (GSWA 136826, Nelson, 1997k). Good sections of the Toorare Tonalite are exposed between Toorare Pool and Old Cherratta Homestead. The Jean Well Granodiorite is foliated, weakly compositionally banded, and includes variably porphyritic granodiorite and monzogranite. It outcrops south and north of the c. 2925 Ma Munni Munni Intrusion and in the area east of Karratha Homestead. SHRIMP dating of two samples (GSWA 142438, Nelson, 1999c and GSWA 142657, Nelson, 1999d) indicates intrusion at c. 2990 to 2988 Ma. The Whyjabby Granodiorite is a foliated, microcline-porphyritic hornblende granodiorite, locally containing layers and blocks of banded, fine-grained tonalite gneiss and banded biotite-rich granodiorite gneiss. Near Whyjabby Pool, on the Maitland River, parts of the rock exhibit an orbicular texture. Excellent exposures of the Whyjabby Granodiorite are present in the type area of Baynton Hill, close to Karratha Homestead. SHRIMP dating indicates a crystallization age of 2994 ± 2 Ma (GSWA 118974, Nelson, 1997f). The Brill Monzogranite is lithologically similar to the Whyjabby Granodiorite but outcrops across 500 km^2 in the southern part of the Cherratta Granitic Complex. The intrusion is composed of coarsely microcline-porphyritic monzogranite and granodiorite, commonly containing minor hornblende. The eastern part of the Cherratta Granitic Complex is largely concealed by the Fortescue Group but inlier exposures are present approximately 40 km south of Roebourne. The Waloo Waloo Monzogranite is a foliated biotite monzogranite dated at 2988 ± 7 Ma (GSWA 168934, Nelson, 2001j). The sample included zircon as old as 3266 Ma, and a zircon group (c. 3050 Ma) of similar age to the Elizabeth Hill Supersuite. The Nd T_{DM} model age of this monzogranite was calculated as 3330 Ma

(Smithies et al., 2007a), supporting the zircon evidence of underlying Paleoarchean crustal material or sediments derived from Paleoarchean sources. The Mardeburra Granodiorite is a porphyritic biotite granodiorite that outcrops in a very small inlier within the Fortescue Group close to the eastern margin of the complex. The rock was dated at 3006 ±12 Ma (GSWA 168932, Nelson, 2001i), but it contained groups of significantly older xenocrystic zircons to 3279 Ma.

The Harding Granitic Complex lies north of the SSZ (Fig. 7) and extends over 130 km along the Pilbara coast from just east of Ruth Well to Cape Thouin, 40 km west from Port Hedland. The onshore width of the complex is generally less than 20 km, but geophysical data indicate that it could extend between 50 and 100 km offshore towards the northern margin of the Pilbara Craton. The Toorare Tonalite of the Maitland River Supersuite is tentatively mapped north of the SSZ, based on its distinctive hornblende-rich composition in the Cherratta Granitic Complex. Based on aeromagnetic imagery, unnamed monzogranite of the supersuite is interpreted to underlie much of the coastal plain between Roebourne and Cape Thouin (Hickman et al., 2006b), but there is almost no exposure. Between Balla Balla and Peawah Hill the Caines Monzogranite (previously 'Caines Well Granite') underlies a 50 km^2 area north of the SSZ. This part of the complex was displaced from the Caines Well Granitic Complex by 30–40 km dextral strike-slip movement at c. 2940 Ma. South of the shear zone, within the Caines Well Granitic Complex, the Caines Monzogranite occupies about 500 km^2 and was dated at 2990 ±5 Ma (GSWA 142950, Nelson, 2000n). The rock is a foliated, K-feldspar porphyritic biotite (–hornblende) monzogranite with local xenoliths of granitic gneiss (probably from the Railway and Orpheus Supersuites).

Mallina Basin and Croydon Group

Exposed over 15 000 km^2, the Mallina Basin is the largest tectonic unit of the northwest Pilbara Craton (Fig. 3). As part of a major revision of the lithostratigraphy of the Pilbara Craton, Van Kranendonk et al. (2006) assigned eight formations to the Croydon Group, including two formations in the central part of the Mallina Basin (Constantine Sandstone and Mallina Formation), four formations in the Whim Creek greenstone belt (Cistern Formation, Rushall Slate, Louden/Mount Negri Volcanics, and Kialrah Rhyolite, all previously forming the 'Bookingarra Group') and two formations in the east Pilbara (Cattle Well Formation and Lalla Rookh Sandstone, previously in the 'De Grey Group'). The Coonieena Basalt, previously of the Gorge Creek Group in the east Pilbara, was reinterpreted to be a member of the 'Bookingarra Formation' (Van Kranendonk et al., 2006), based mainly on geochemical criteria. The newly established Croydon Group included almost all formations that were interpreted by Van Kranendonk et al. (2006) to be between 2970 Ma and 2940 Ma. Only the Coondamar and Mosquito Creek Formations of the southeast Pilbara were excluded on the basis of deposition in a well-defined separate basin (Mosquito Creek Basin), and these formations were assigned to the Nullagine Group. The Mallina Basin was interpreted to be an intracontinental basin (Smithies et al., 2001b). However, evidence presented in this Report supports previous interpretations that the Mallina Basin of the northwest Pilbara evolved as a back-arc basin. For reasons given in this Report, the Croydon Group excludes the Coonieena Basalt of the east Pilbara and the five post-2955 Ma formations of the Whim Creek greenstone belt, which are now reassigned to the Bookingarra Group (redefined in Appendix 2).

The name Mallina Basin was used by Krapez and Eisenlohr (1998) to refer to most of the succession between the LF and the TTSZ. However, they excluded an underlying, more metamorphosed succession of ultramafic–mafic rocks, BIF, chert, and sandstone that they assigned to the 'Corboy-Paddy Market Basin' (now an obsolete name). Other workers have assigned the BIF and chert exposed in the anticlines in the Mallina Basin to the Gorge Creek Group (Fitton et al., 1975; Hickman, 1977, 1980b; Smithies and Farrell, 2000; Smithies and Hickman, 2003, 2004).

Divisions of the Mallina Basin

The definition of the Mallina Basin was changed by Van Kranendonk et al. (2006) when they included clastic sedimentary formations overlying the c. 3020 Ma Gorge Creek Group in the northwest part of the EPT. This change implied that the Mallina Basin, which occupies most of the CPTZ, evolved in three northeast-trending divisions or sectors: a central, relatively deep-water basin flanked by shallow-water margins to the northwest and southeast. However, the three sectors have different stratigraphic successions and are separated by two major faults. The LF separates the northwest part of the basin from the central part, and the TTSZ separates the central part from the southeast part (Fig. 25). The amounts of strike-slip movement on the faults are unknown, but there is evidence that lateral displacements were considerable (see Central Pilbara Tectonic Zone). The TTSZ is a terrane boundary (Van Kranendonk et al., 2006) and the LF was interpreted to be a tectonostratigraphic domain boundary (Krapez and Eisenlohr, 1998). Geochronology indicates that many formations of the three sectors were deposited during or after commencement of the North Pilbara Orogeny at c. 2955 Ma. On the other hand, evidence discussed in this Report indicates that the lower part of the central Mallina Basin succession is up to 50 million years older than the succession deposited after the c. 2955 Ma event. The central deep-water rift section of the basin, preserved between the TTSZ and the LF, is mainly composed of the Constantine Sandstone and the Mallina Formation. In the east Pilbara the far southeast margin of the entire Mallina Basin, separated from the central sector by the TTSZ, contains fluviatile, lacustrine, and shallow-water marine sedimentary successions of the Lalla Rookh Sandstone (Hickman, 1983; Krapez, 1984; Krapez and Barley, 1987; Williams, 1999; Van Kranendonk, 2000; Smithies, 2004). Geochronology on the Lalla Rookh Sandstone (GSWA 142951, Nelson, 2000o) revealed detrital zircons ranging in age from c. 3660 to 3233 Ma, evidently derived from erosion of the EPT and older underlying Eoarchean Pilbara crust. In the northeast Pilbara, the Lalla Rookh

Sandstone locally overlies the c. 2988 Ma Cattle Well Formation, but in most areas its maximum depositional age is constrained only by the 3066–3015 Ma depositional age of the underlying Gorge Creek Group. Thus, parts of the Lalla Rookh Sandstone may be of similar age to the Constantine Sandstone. Unconformably overlying the EPT, the southeast formations of the Mallina Basin are described in more detail in a later Report (Hickman, in prep.).

Continuity of basin stratigraphy

From evidence summarized in this Report, the age range of the Croydon Group is 3015–2931 Ma, which is an exceptionally long timeframe for a single lithostratigraphic group. Consideration of several factors indicates that the stratigraphy assigned to the Croydon Group (Van Kranendonk et al., 2006) requires subdivision into at least three groups. Firstly, and most importantly, a regional unconformity is now interpreted to exist at c. 2955 Ma in the central part of the Mallina Basin between the LF and TTSZ (see below). Secondly, and related to the unconformities, structures of four deformation events (D_5–D_8) have been recognized in the 3015–2931 Ma succession. This Report is primarily a regional review of northwest Pilbara geology and therefore, while identifying the issues, it has been decided to limit stratigraphic revision to that considered essential at this stage. The main revision, based on the absence of any apparent stratigraphic or structural continuity across the LF, is to reinstate the Bookingarra Group (Appendix 2), as originally defined by Pike and Cas (2002). The Bookingarra Group comprises all Mesoarchean formations unconformably overlying the Whim Creek Group. Other stratigraphic issues remaining for the Mallina Basin succession are related to unconformities.

Unconformity within the central Mallina Basin

The succession of the central (main) part of the Mallina Basin includes an unconformity at c. 2955 Ma (Huston et al., 2000; Smithies and Farrell, 2000; Smithies et al., 2001b, 2002; Van Kranendonk et al., 2002). The succession beneath this unconformity was deformed by D_5 and D_6 folds that are absent from the succession above the unconformity. The D_5 and D_6 folds were intruded by the 2954–2945 Ma Indee Suite (Sisters Supersuite), intrusion of which is interpreted to mark the earliest stage of the North Pilbara Orogeny (see below). The basal part of the lower succession of the Mallina Basin is underlain, apparently across a sedimentary transition sequence, by the 3022–3016 Cleaverville Formation of the Gorge Creek Group (Fitton et al., 1975; Hickman, 1977). The upper part of the Mallina Basin succession, above the c. 2955 Ma unconformity, includes clastic sedimentary rocks containing c. 2941 Ma detrital zircons (Smithies et al., 1999). However, far more sampling and geochronology would be required to provide a more reliable estimate of the age range of the upper succession.

Deposition of the central Mallina Basin succession was in two main stages. The first stage, from 3015 Ma to 2955 Ma, included deposition of most of the Constantine Sandstone and part of the Mallina Formation. This interpretation establishes that the present stratigraphic division of the central Mallina Basin requires revision, because metasedimentary units mapped as either Constantine Sandstone and Mallina Formation are interpreted to overlie the c. 2955 Ma unconformity, and this is interpreted to be a regional unconformity. The present problem is that, due to poor exposure and complex folding and faulting, the unconformity has not been mapped across the basin and both formations have been mapped entirely on the basis of lithology. The lower part of the Mallina Basin includes the succession around Croydon Homestead and it would therefore be most reasonable to use the name Croydon Group for this part of the succession. The second stage, from c. 2950 to 2930 Ma, included deposition of most of the Mallina Formation, rhyolite volcanism and intrusion at c. 2940 Ma (Kialrah Rhyolite), D_8 strike-slip faulting, and most mineralization in the Mallina Basin (e.g. lode Au and Cu–Zn; see Mineralization). It is recommended that this part of the Mallina Basin succession should in future be assigned to a new group separate from the Bookingarra Group.

Unconformities within the northwest Mallina Basin

Many workers have interpreted the ultramafic–mafic succession comprising the Louden and Mount Negri Volcanics to unconformably overlie the Rushall Slate and Cistern Formation (Fitton et al., 1975; Hickman, 1977, 1983, 1990; Barley, 1987; Krapez, 1993; Smithies, 1996, 1998a; Krapez and Eisenlohr, 1998; Collins and Marshall, 1999). Krapez and Eisenlohr (1998) assigned this ultramafic–mafic succession to a separate basin ('Negri basin'), recognizing an angular regional unconformity between it and the underlying Rushall Slate. In their sequence stratigraphy, this unconformity defines the boundary between two megasequences. Despite previous references to an unconformity above the Cistern Formation – Rushall Slate succession, Pike et al. (2002) could not confirm this interpretation. Instead, they used an observation of peperite at a contact between a dolerite sill (which they correlated with the 'Negri Volcanics') and the Rushall Slate as evidence that deposition of the protolith of the Rushall Slate and the overlying mafic volcanism were contemporaneous. However, the peperite evidence was observed only at Good Luck Well in the central part of the Whim Creek greenstone belt and its regional significance is dubious. In other areas, the Rushall Slate contains basalt and dolerite (e.g. Comstock Member) and both the Louden and Mount Negri Volcanics include clastic sedimentary units (Krapez and Eisenlohr, 1998; Pike et al, 2006). Consequently, the interaction of basalt and dolerite with soft sediment is likely to have taken place during deposition of several formations.

More reliable evidence on the regional stratigraphic relationship between the Rushall Slate and the overlying Louden Volcanics and Mount Negri Volcanics may be taken from the different tectonic settings of the formations. The Cistern Formation – Rushall Slate succession was deposited around felsic volcanic centres whereas the overlying ultramafic–mafic succession is generally interpreted to have been deposited during rifting, most likely associated with uplift and extension above a mantle plume (Sun et al., 1991; Krapez and Eisenlohr, 1998; Smithies and Champion, 2000; Arndt et al., 2001; Pike, 2001; Pike et al, 2006). A mantle plume origin for the Louden Volcanics and Mount Negri Volcanics is consistent with extensive komatiite and komatiitic basalt in these formations. Several authors have argued that Archean komatiite provides evidence of plume-related volcanism due to decompression melting of exceptionally hot mantle (Sossi et al., 2016). In this interpretation, there would be no continuity between deposition of the Cistern – Rushall Slate succession and eruption of the Louden Volcanics. From previous accounts there is evidence that unconformities locally exist at the base of the Mount Negri Volcanics, but it is not clear if these observed stratigraphic breaks form parts of a regional unconformity. If a regional unconformity were to be demonstrated the Bookingarra Group would require separation of the Cistern Formation and Rushall Slate into a separate group. However, the present interpretation is that all reported unconformities within the Bookingarra Group are local in nature and are consequences of deposition during crustal extension and rifting.

Stratigraphic correlations across the Loudens Fault

The controversial nature of one stratigraphic correlation (Cistern Formation with Constantine Sandstone) across the Loudens Fault (LF) has already been discussed (Whim Creek and Mallina Basins). A second significant correlation has involved the ultramafic–mafic succession of the upper Bookingarra Group. In the Whim Creek greenstone belt the Louden Volcanics, which overlies the Cistern Formation is approximately 2 km thick and is abruptly terminated by the LF. Immediately southeast of the LF the Mallina Formation, which is between 2.5 km and 5 km thick, is likewise terminated abruptly by the fault. Vertical movement on the LF cannot explain these differences because the successions above and below the Bookingarra and Mallina Formation are also different. Therefore, the most likely explanation is major post-2950 Ma strike-slip displacement on the LF.

Van Kranendonk et al. (2006) suggested that the ultramafic–mafic succession of the ‘Bookingarra Formation’ (Louden and Mount Negri Volcanics in this Report) can be correlated with mafic sills (named by them the Yareweeree Boninite Member [YBM]) and mafic volcanic rocks (named as the South Mallina Basalt Member [SMBM]) in the central Mallina Basin. The YBM is chemically distinct from the SMBM and is very close to the base of the central Mallina Basin (Smithies et al., 2004a). Smithies et al. (2007a) interpreted the YBM to be a sill, which in theory leaves open the possibility that it is the same age as the Louden Volcanics. However, if that is the situation the sill must have been intruded around the limbs of tight D_5 and D_6 folds that pre-date the c. 2955 Ma unconformity and this is unlikely. The SMBM occupies a higher stratigraphic level in the central Mallina Basin and therefore might be part of the succession above the c. 2955 Ma unconformity. This interpretation has been provisionally adopted in this Report.

Van Kranendonk et al. (2006) proposed that a thick unit of pillow basalt that overlies the Cleaverville Formation in the northeast Pilbara is laterally equivalent to the Louden and Mount Negri Volcanics. Williams (1999) previously named this basaltic unit the Coonieena Basalt and assigned it to the Gorge Creek Group. Correlation with the post-2955 Ma Bookingarra Group, which was based on geochemistry, has since been precluded by geochronology. Dating of the Cattle Well Formation (GSWA 180048, Wingate et al., 2009b), which stratigraphically overlies the Coonieena Basalt, indicates that the depositional age of the basalt is greater than c. 2990 Ma. The Coonieena Basalt, which is currently not assigned to a group, extends to the Bamboo Creek – Coppin Gap area of the east Pilbara (Williams, 1999), an area which is 250 km east-southeast of the Bookingarra Group in the Whim Creek greenstone belt.

Croydon Group, Central Mallina Basin

The central part of the Mallina Basin trends southwest–northeast and, as preserved today, is up to 100 km wide and 300 km long. Prior to orogenic northwest–southeast crustal shortening and closure of the basin during the 2955–2919 Ma North Pilbara Orogeny (see below) the width of the central Mallina Basin was most likely more than twice its present width. Regional gravity and aeromagnetic data indicate that the total length of the rift basin, including concealed sections in the southwest and northeast, is at least 600 km (Hickman, 2004a). As emphasized in the Introduction of this Report, the Mesoarchean extent of all linear tectonic units, such as the Mallina Basin would have been greater before Neoarchean–Paleoproterozoic continental breakup.

The regional stratigraphy of the central part of the Mallina Basin remains unclear due to poor exposure and limited geochronology. The c. 85 million year depositional history of this part of the basin (3015–2931 Ma) included post-2955 Ma intrusion of the Sisters Supersuite and deformation that divided the succession into lower and upper units (Smithies et al., 2001b, 2002; Van Kranendonk et al., 2002). The tectonic evolution of the Mallina Basin involved at least four deformation events (D_5 to D_8) defined by thrusting, folding, and metamorphism. Additionally, strike-slip and normal faulting involving local extension was probably ongoing (Smithies et al., 2001b; Hickman, 2004a; Van Kranendonk et al., 2006). Several workers have recorded evidence of consistent vertical movement on faults within the basin, with southeast blocks being uplifted relative to those to the northwest (Smithies et al., 1999, 2001b; Huston et al., 2001; Blewett, 2002; Beintema, 2003). Major faults to the northeast of the basin, such as the LF, Kents Bore Fault (KBF), and SSZ indicate predominantly northwest-side-up vertical movement although, at least for the LF and SSZ, vertical movement is minor compared to horizontal strike-slip movement. Syndepositional movement on the faults within the central part of the Mallina Basin is likely to have caused abrupt lateral facies changes, breaks in the stratigraphy, and possibly even the development of ephemeral sub-basins. Several workers have concluded that there must be as yet unrecognized unconformities within the succession (Fitton et al. 1975; Smithies and Farrell, 2000; Smithies et al., 2001b; Van Kranendonk et al., 2002, 2006; Beintema, 2003).

Basement rocks

Exposures within the central part of the Mallina Basin establish that it is underlain by the Gorge Creek Basin. Airborne magnetic data show the positions of these basement highs because the Gorge Creek Group includes BIF of the Cleaverville Formation. However, there is no evidence in terms of outcrops or isotopic data that the Whim Creek Group is present in this sector of the Mallina Basin. It is inferred that the Regal Formation of the CPB (basaltic crust formed by the c. 3200 Ma EPT Rifting Event) underlies the Gorge Creek Basin in the central Mallina Basin (Fig. 27). Evidence for this is provided by <3250 Ma Nd model ages from granites of the Sisters Supersuites (Van Kranendonk et al., 2007b) and the age range of xenocrystic zircons in these granites (maximum age c. 3246 Ma, GSWA 160727, Nelson, 2001c). Since the Whundo Group (ST) was formed within the Central Pilbara Basin it should also underlie part of the Mallina Basin (Fig. 27) and xenocrystic zircons in the Sisters Supersuite support this conclusion (c. 3127 Ma and c. 3114 Ma, GSWA 160727, Nelson, 2001c). The c. 3120 Ma ages of detrital zircons in the Cleaverville Formation on the southeast margin of the basin (GSWA 142842, Nelson, 1998o) suggest that the Whundo Group extended across the central Mallina Basin in the southwest. Isotopic data suggest that during deposition of the Croydon Group and intrusion of the Sisters Supersuite the basement was predominantly <3250 Ma crust (Table 5). This is an important conclusion because it signifies that accretion of the KT to the EPT during the 3068–3066 Ma Prinsep Orogeny (Van Kranendonk et al., 2006, 2010) was incomplete and left a belt of entirely Mesoarchean basement separating these older terranes. In other words, if evolution of the CPTZ is likened to the stages of a Wilson cycle (Van Kranendonk et al., 2010) the Prinsep Orogeny and subsequent erosion did not complete the cycle. Instead, after a brief interlude of post-orogenic extension (Gorge Creek Basin) convergence continued for 100 million years to finally culminate with major orogeny from 2955 to 2919 Ma.

Stratigraphy

Fitton et al. (1975) introduced the names Constantine Sandstone and Mallina Formation to distinguish relatively arenaceous from more argillaceous parts of the Croydon Group in the central part of the Mallina Basin. This distinction was based only on reconnaissance mapping, and even today the detailed stratigraphy of the basin is poorly understood. Fitton et al. (1975) and most subsequent workers made the generalization that throughout the central part of the basin the Mallina Formation stratigraphically overlies the Constantine Sandstone. Where the Constantine Sandstone is most thickly developed, particularly in the southwest, this generalization is essentially correct. Taking into account the tectonic environment in which this extensive basin evolved, including rifting, strike-slip faulting, granitic intrusion, and folding over a period of c. 100 million years, the lithological differences between the Constantine Sandstone and Mallina Formation most likely reflect a very complex depositional history. As currently mapped, medium- to coarse-grained, poorly sorted subarkose to wacke, locally with thick conglomerate layers, are assigned to the Constantine Sandstone, whereas interbedded well-graded, fine- to medium-grained wacke, and shale are mapped as the Mallina Formation.

Detrital zircon ages in the Constantine Sandstone and Mallina Formation include many grains between c. 3795 Ma and c. 3530 Ma that were derived from erosion of crust older than the Pilbara Supergroup of the EPT (Hickman, 2012; Kemp et al., 2015a,b). A similar situation is recorded in the Mosquito Creek Basin of the east Pilbara where Bagas et al. (2004, 2008) suggested transport of detritus by longitudinal basin currents from distant sources. However, the fact that such old zircon grains are widespread in other clastic sedimentary rocks of the De Grey Supergroup, some of which include proximal conglomerate and sandstone deposits (e.g. Farrel Quartzite of the Gorge Creek Group), suggests that the exceptionally old zircon grains were derived from early crust of the Pilbara Craton.

Constantine Sandstone

The name Constantine Sandstone has been applied to a particular sedimentary facies association rather than to a lithostratigraphic formation with well-defined lower and upper contacts. Uncertainty over the regional extent of the Constantine Sandstone is illustrated by comparing its distributions as mapped by Fitton et al. (1975), Smithies et al. (2001b), and Smithies and Hickman (2003). Deposited in submarine fans (Eriksson, 1982; Smithies et al., 1999), the Constantine Sandstone may be present as wedge-shaped units of more than one age. Additionally, the Constantine Sandstone facies is not the oldest part of the central Mallina Basin succession. In several areas it is underlain by intercalated chert, BIF, and clastic sedimentary rocks ranging from shale to conglomerate (Ryan, 1965; Fitton et al., 1975; Hickman, 1983; Smithies and Farrell, 2000; Smithies and Hickman, 2003, 2004). This lithological association is interpreted to be a sedimentary transition from the Cleaverville Formation of the Gorge Creek Group, implying that the lowest part of the basin succession is c. 3015 Ma.

On the southeast margin of the central Mallina Basin there is an angular erosional unconformity between the Cleaverville Formation and coarse clastic sedimentary rocks assigned to the Constantine Sandstone. Facies of the sandstone formation testify to extremely active erosion along a fault scarp or series of scarps. Adjacent to the Pilbara Well greenstone belt, between Teichmans gold mine and north of Hong Kong mining centre, basal clastic units of this part of the Croydon Group include polymictic conglomerate, boulder beds, and breccia, and are up to 500 m thick (Smithies and Farrell, 2000). Boulders of chert and BIF measure up to 5 m in diameter (Fitton et al., 1975) and angular blocks of chert are locally up to 50 m long (Smithies and Farrell, 2000). The unconformity is interpreted to closely follow an earlier rifted contact along the basin margin and is complicated by later faulting and shearing along the same zone (i.e. TTSZ). The conglomerate appears as lenses in upward-fining sequences that pass upwards and laterally into massive or graded beds of pebbly to fine-grained sandstone. Smithies et al. (1999) interpreted this facies in terms of channel deposits of a submarine fan. Elsewhere, lithologies assigned to the Constantine Sandstone were probably deposited in proximal lobe environments (Eriksson, 1982).

Yareweeree Boninite Member

This Report does not follow the stratigraphic revision proposed by Van Kranendonk et al. (2006) in which the Yareweeree Boninite Member (YBM) was defined as a member of the post-2955 Ma 'Bookingarra Formation' (equivalent to two formations of the Bookingarra Group in this Report). The main reason is that the YBM is stratigraphically located beneath the c. 2955 Ma regional unconformity within the Mallina Basin. A second important reason is that the YBM was folded by D_5 and D_6 structures (D_1 and D_2 using the classification by Smithies, 1998b and Smithies and Farrell, 2000), whereas the Bookingarra Group was folded only by D_8. Thirdly, descriptions by Smithies and Farrell (2000) indicate that the YBM was intruded and metamorphosed by the 2954–2945 Ma Indee Suite, indicating an age greater than 2954 Ma for the YBM. Smithies et al. (2007c) supported this interpretation by concluding that the YBM was intruded at c. 2970 Ma, the maximum age they interpreted for deposition of the lower succession.

Van Kranendonk et al. (2006) placed the YBM within the 'Bookingarra Formation' (now assigned to the Louden and Mount Negri Volcanics in the upper part of the Bookingarra Group). They estimated the depositional age of the 'Bookingarra Formation' to be c. 2955 Ma. However, in defining the YBM they gave its age as c. 2970 Ma. Evidence presented in this Report suggests that the age of the Louden Volcanics (Bookingarra Group) is 2955–2948 Ma (see Northwest Mallina Basin). The stratigraphic correlation of the YBM with the 'Bookingarra Formation' (Van Kranendonk et al., 2006), which was based on geochemistry, is not accepted in this Report and instead it is interpreted that mafic units of similar composition were erupted and intruded at different times. In this Report the YBM is considered to be a member of the Constantine Sandstone and its age is estimated to be 3015–2970 Ma.

The YBM intrudes the central part of the Mallina Basin succession beneath the main units of Constantine sandstone (Fig. 25). Due to metamorphism, the member now comprises serpentine-rich ultramafic schist, chlorite–actinolite–serpentine–epidote schist (metamorphosed komatiitic basalt or dolerite), and metamorphosed melanogabbro and pyroxenite. Smithies (1998) first assigned rocks of this member to the c. 2950 Ma Millindinna Intrusion but he later obtained geochemical data indicating a strong similarity to boninite (Smithies, 2002a; Smithies et al., 2004a). Boninites are rare, high-Mg basaltic to andesitic rocks, typically with low Ti concentrations, high large-ion lithophile element (LILE) concentrations, and with very high Al_2O_3/TiO_2, low Gd/Yb, and high La/Gd compared to primitive mantle. In the Phanerozoic record such rocks are confined to convergent margin settings (Crawford et al., 1989). However, Smithies et al. (1999, 2001b, 2004a, 2004c) and Smithies (2002a) interpreted the Mallina Basin to be an intracontinental basin and suggested that this formed at least 40 million years after the completion of subduction events in the northwest Pilbara. Smithies et al. (2004a) interpreted the origin of the YBM as a product of partial melting of older subduction-enriched mantle. Smithies et al. (2004c) suggested that sediment derived from >3300 Ma crust was partially melted during the subduction event that had been responsible for the Whim Creek Group and fluids generated by this subduction infiltrated a mantle wedge. The age of the wedge was uncertain in this interpretation but it was suggested that it might date from the c. 3120 Ma Whundo subduction. The metasomatism was thought to have been regionally extensive and provided a later source for the YBM.

The interpretation of the tectonic setting of the Mallina Basin is controversial (see Tectonic setting), and in this Report the Mallina Basin is interpreted to be a back-arc basin that evolved along with ongoing subduction of a northwest plate. In this interpretation the boninite-like rocks were generated on an active convergent margin. Presently available geochronology does not support a 40 million-year break between a final subduction event and emplacement of the YBM. Subduction responsible for the Whim Creek Group and the Maitland River Supersuite occurred from c. 3010 Ma to at least 2982 Ma. Additionally, a 2970 ± 5 Ma zircon group within the c. 3018 Ma Black Hill Well Monzogranite (GSWA 142430, Nelson, 1999b) suggests subduction to at least 2970 Ma. Other evidence of ongoing felsic magmatism is provided by post-2970 Ma felsic units in the Whim Creek greenstone belt and in c. 2950 Ma granitic intrusions within the Mallina Basin. Each of these felsic units contains groups of zircon ages between c. 2980 and 2960 Ma (GSWA 144261, Nelson, 1998v; GSWA 142949, Nelson, 2000m; GSWA 142889, Nelson, 1999g; GSWA 142935, Nelson, 2000c). Therefore, more or less continuous felsic magmatism is interpreted from 3010 to 2960 Ma, after which the Sisters Supersuite was intruded the northwest Pilbara between 2954 and 2919 Ma.

Trace element plots normalized to primitive mantle (Fig. 28) show that most mafic volcanic rocks of the Croydon Group (central Mallina Basin) and Bookingarra Group (Whim Creek greenstone belt) exhibit enriched Th, Zr, and LREE. The U-shaped profile of the middle REE, so characteristic of boninites, is particularly well displayed by the YBM.

Mallina Formation

The Mallina Formation is very poorly exposed over large areas of the central Mallina Basin and airborne magnetic data provide little information on its stratigraphy or the structures that deform it. Lower sections of the Mallina Formation locally include thickly bedded medium- to coarse-grained subarkose and arkose, but the formation is dominated by wacke and shale deposited in proximal to distal lobes of submarine fans. The wacke is chloritic

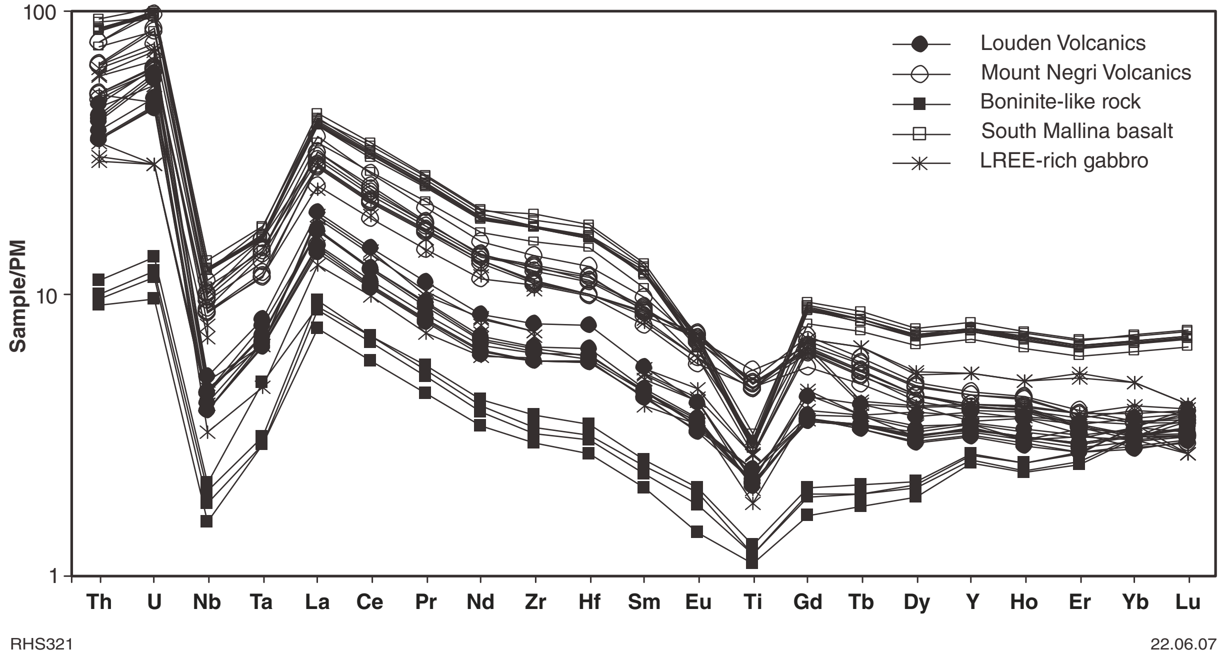


Figure 28. Trace element plots normalized to primitive mantle (PM) for basaltic units of the Mallina Basin (from Smithies et al., 2007a). Normalizing factors after Sun and McDonough (1989).

and contains abundant plagioclase, suggesting mafic to intermediate sources. The limited exposures available display well-preserved sedimentary structures of the type found in turbidite deposits, such as graded bedding, cross-bedding, and scour structures. A general regional trend to increasingly fine-grained sedimentary facies in the northern and eastern parts of the Mallina Basin suggests an overall transition to deeper water, basin plain environments (Smithies et al., 1999).

South Mallina Basalt Member

The South Mallina Basalt Member (SMBM) consists of thin units of siliceous basalt and high-Mg basalt within the sedimentary rocks in the central part of the Mallina Basin. The basalt is typically variolitic, commonly pyroxene spinifex-textured, locally pillowed, and hyaloclastic breccia is present in some areas. The stratigraphic position of the SMBM within the central Mallina Basin is not firmly established but it is present mainly within the Mallina Formation. The chemistry of the SMBM is distinct from that of the YBM (Smithies et al. 2004a). For example, the SMBM contains approximately ten times greater normalized concentrations of Th, Zr, and LREE compared to the YBM (data in Smithies et al., 2004a).

Salt Well Member (Lalla Rookh Sandstone)

Immediately east of the Ord Range greenstone belt (Fig. 7) and across the Pardoo Fault into the northern part of the EPT, the Salt Well Member is a unit of mafic volcaniclastic deposits, pillowed and vesicular basalt flows, and dolerite intrusions (Smithies et al., 2004b). Originally the member was included within the ‘Paradise Plains Formation’, but this name was made obsolete by Van Kranendonk et al. (2006) who correlated it with the Lalla Rookh Sandstone. In the EPT, the Lalla Rookh Sandstone was deposited unconformably on the Gorge Creek Group, probably for the most part across a discontinuous shallow-water shelf but also locally within small rift basins between the east Pilbara granitic domes that were actively rising at c. 2950 Ma (Hickman and Van Kranendonk, 2004).

Smithies et al. (2004b) correlated the Salt Well Member south of the Pardoo Fault with mafic volcanic rocks in the Mallina Basin east of the Ord Range. This correlation suggests a northwesterly transition from the shallow-water Lalla Rookh Sandstone to deeper-water clastic facies in the central part of the Mallina Basin. The presence of basalt, dolerite, basaltic volcaniclastic rocks, and mafic debris flows in the Lalla Rookh Sandstone of the Ord Range area suggests a phase of rifting, mafic volcanism, and dolerite intrusion along the Pardoo Fault. Notably, both the Pardoo Fault and its southwest continuation, the TTSZ, appear to have acted as conduits for ultramafic and mafic intrusions at c. 2950 Ma.

Bookingarra Group, Whim Creek greenstone belt

In this Report all volcanic and sedimentary units overlying the c. 2955 Ma unconformity in the Whim Creek greenstone belt are assigned to the redefined (Appendix 2) Bookingarra Group (Pike, 2001; Pike et al., 2002; Van Kranendonk et al., 2002). Available geochronology and regional geological considerations (discussed in this Report) indicate that the age of the unconformity beneath the Cistern Formation is c. 2955 Ma. This is the same age as the unconformity in the central part of the

Mallina Basin (see Basin relationships). The minimum depositional age of the lower part of the Bookingarra Group is suggested by a date of c. 2948 Ma for Cu–Zn mineralization in the Rushall Slate (Huston et al., 2002a).

Although both the Bookingarra Group and the upper succession in the central part of the Mallina Basin overlie a c. 2955 Ma unconformity the two successions are very different lithologically. The most obvious difference is that the Bookingarra Group is mainly composed of ultramafic–mafic volcanic rocks (Louden and Mount Negri Volcanics), whereas the Mallina Formation is almost entirely composed of metasedimentary rocks. The only formation recognized in both successions is the Kialrah Rhyolite, units of which has been dated between c. 2948 and 2941 Ma (see below). This overlies or intrudes the Louden Volcanics at the top of the Bookingarra Group but forms flow-banded rhyolite units in the lower part of the upper Mallina Basin succession. If the depositional age of the Louden Volcanics is taken as c. 2950 Ma it is likely that the LF, which separates the two successions, juxtaposed them shortly after 2950 Ma, most likely during D_8 in the CPTZ at c. 2940 Ma.

Cistern Formation and Rushall Slate

The Cistern Formation is composed of volcanoclastic rocks, including conglomerate and sandstone. The maximum depositional age of the Cistern Formation was interpreted to be 2964 ± 6 Ma (GSWA 142949, Nelson, 2000m; data reinterpreted by Huston et al., 2002a) and Pb isotopic data from mineralization in the Rushall Slate were interpreted by Huston et al. (2002a) to indicate an age of c. 2948 Ma. If the felsic volcanic centres around which the Cistern Formation was deposited (Collins and Marshall, 1999a, 1999b; Pike et al., 2006) were genetically related to early granites of the Sisters Supersuite a maximum age of c. 2955 Ma might be assumed.

The Rushall Slate (Hickman, 1977) is best developed in the Whim Creek area, but also outcrops to the south of Mount Negri, and between Good Luck Well and Opaline Well. Contacts with the Cistern Formation are conformable and gradational. Where the Cistern Formation is absent, Rushall Slate – Red Hill Volcanics contacts are disconformable or faulted. The formation reaches a maximum thickness of approximately 200 m and comprises slate (metamorphosed shale) with minor metamorphosed siltstone and sandstone. Graded bedding is common indicating deposition from turbidity currents. In many exposures the rock contains a prominent schistosity or slaty cleavage that is axial planar to folding. The slate is composed of very fine-grained sericite, chlorite, and quartz, and is heavily dusted with iron oxides and pyrite. Marston (1979) and Barley (1987) recorded the presence of tuffaceous beds and chert within the slate, with associated VHMS Cu–Zn mineralization at Whim Creek. Pike et al. (2002) interpreted the slate and metasiltstone protoliths of the Rushall Slate to have been deposited in a deepening basin with periodic influx of turbidity currents.

The Cistern Formation and Rushall Slate are economically important for containing VHMS deposits at Mons Cupri, Whim Creek, Balla Balla, and Salt Creek (see Mineralization). Documentation of the mineralization has been extensive (Low, 1963; Blockley, 1971; Miller and Gair, 1975; Marston, 1979; Collins and Marshall, 1999a; Ferguson, 1999; Huston et al., 2000; Huston, 2006) and the close spatial relationship between this mineralization and the restricted distribution of the Cistern Formation – Rushall Slate succession suggests a common depositional setting around volcanic centres. Several workers have linked the location of these volcanic centres to large intrusions of porphyritic dacite and rhyolite at Mons Cupri, Good Luck Well, and Ant Hill east of Balla Balla and Salt Creek (Sylvester and de Laeter, 1987; Collins and Marshall, 1999a). This suggested relationship is stratigraphically significant because, as noted above, it implies that at least some of the large rhyolite and dacite intrusions into the 3009–2991 Ma Red Hill Volcanics are not part of the Whim Creek Group but similar in age to the Cistern Formation.

Due to syndepositional rifting there are lateral variations in the thicknesses and facies of the Cistern Formation and Rushall Slate, and the Cistern Formation is almost entirely absent west of Gold Luck Well (Hickman, 2002; Hickman et al., 2006b). Between Mons Cupri and Whim Creek there is a basal unit of volcaniclastic conglomerate, volcanic breccia, and sandstone, but this unit wedges out at Whim Creek with the result that the Rushall Slate directly overlies the Red Hill Volcanics. In the Good Luck Well area, a lower association of volcanic breccia is interbedded with coarse-grained lithic arenite or granulestone and overlain by ripple-bedded sandstone. Pike et al. (2006) interpreted these facies as reflecting proximal fluviatile deposition. In most areas the upper levels of the formation are composed either of lithic-sandstone, containing basaltic clasts or quartz-rich sandstone derived from erosion of felsic igneous rocks. Another facies consists of planar-laminated sandstone and siltstone and wacke with graded bedding, indicative of deposition from turbidity currents. Pike et al. (2006) interpreted this upper part of the Cistern Formation as indicating deposition in subaqueous fans.

Comstock Member

A thin unit of vesicular basalt and high-Mg basalt exposed near the base of the Rushall Slate was referred to as the 'Comstock Andesite Member' (obsolete name) by Miller and Gair (1975), but it was later renamed the Comstock Member by Smithies (1997). Pike et al. (2006) interpreted the member to be a dolerite sill, in which case it could be part of the Opaline Well Intrusion. The rock is fine grained, locally vesicular, variolitic and pyroxene-spinifex textured, and is generally chloritized and ferruginized. The Comstock Member is chemically similar to basalt of Mount Negri Volcanics which overlies the Rushall Slate in several areas.

Louden and Mount Negri Volcanics

Where both formations are exposed together, as around Mount Negri northeast of Whim Creek, the more komatiitic Louden Volcanics are stratigraphically overlain by the dominantly tholeiitic Mount Negri Volcanics.

Hickman (1977) reported that the two units could be mapped separately based on degree of deformation combined with compositional differences. The Louden Volcanics (Hickman, 1983; redefined in Appendix 2) are far more extensive than the Mount Negri Volcanics and outcrop for 110 km along the southern and southeastern margins of the Whim Creek greenstone belt close to the LF. Based mainly on geophysical data, the formation extends an additional 70 km northeast of the merger of the LF with the SSZ, along the northern side of the SSZ almost to Cape Thouin (Hickman and Smithies, 2000). The Louden Volcanics are weakly metamorphosed with the dominant rock type being komatiitic basalt, containing coarse pyroxene-spinifex textures or randomly orientated, acicular pyroxene phenocrysts. Approximately 14.5 km south of Sherlock Homestead, komatiite consists of skeletal olivine (partly altered to serpentine) plates set in a groundmass of feathery pyroxene (replaced by amphibole and chlorite). South of Whim Creek at Loudens Patch Arndt et al. (2001) reported komatiitic flows in which pyroxene-spinifex texture includes individual crystals up to 1 m long and 3 mm wide. Olivine and pyroxene cumulate zones in the flows were found to contain up to 32 wt% MgO, 4500 ppm Cr, and 1150 ppm Ni whereas most spinifex-textured samples were of komatiitic basalt composition (<18 wt% MgO).

Aphyric basalt is common near the top of the Louden Volcanics, forming cooling units up to 10 m thick. Hyaloclastite and pillowed basalt are also present at the top of the sequence, where they are locally interlayered with and conformably overlain by chert and clastic rocks, including conglomerate. In the northeast section of the Whim Creek greenstone belt, adjacent to the LF and SSZ, metamorphosed clastic rocks are interbedded with the Louden Volcanics, suggesting syndepositional extensional faulting. The sedimentary units are dominated by poorly sorted, medium- to fine-grained sandstone, containing beds and lenses of shale and units of polymictic conglomerate up to 100 m thick.

The Mount Negri Volcanics outcrop in several outliers in the eastern part of the Whim Creek greenstone belt, most notably at Mount Negri and between Hill Well and Mons Cupri. The formation is mainly composed of variolitic and vesicular basalt, with pyroxene-spinifex textures locally preserved. Variolitic basalt forms individual flows up to 15 m thick and at Mount Negri reaches an accumulated thickness of greater than 150 m. Dark green, pea-sized varioles are abundant and consist of acicular clinopyroxene, interstitial plagioclase, and devitrified glass and lie in a light green groundmass of clinopyroxene, plagioclase, and glass. Euhedral clinopyroxene phenocrysts up to 2 mm in length are distributed randomly throughout varioles and groundmass.

Some workers have interpreted the Mount Negri Volcanics to unconformably overlie the Louden Volcanics (Hickman, 1983; Smithies, 1996, 1998a; Smithies et al., 1999) consistent with early observations that the Louden Volcanics are more deformed (Hickman, 1977). In several areas of the Whim Creek greenstone belt only one of the formations is present. For example, both on the southeast slopes of Mount Negri and in the Good Luck Well area the Rushall Slate is directly overlain by the Mount Negri Volcanics, whereas in the Warambie area the formation is represented only by the Louden Volcanics (Hickman, 2002). In view of the c. 2000 m thickness of the combined Louden Volcanics and Mount Negri Volcanics (Krapez and Eisenlohr, 1998) major lateral facies changes suggest either that deposition was on very uneven topography or that there was substantial syndepositional uplift and erosion (Barley, 1987).

No age determinations have been possible on the Louden and Mount Negri Volcanics, but the Kialrah Rhyolite, which stratigraphically overlies the Louden Volcanics (Hickman, 1997a, 2002), has a minimum age of 2943 ± 7 Ma (GSWA 144261, Nelson, 1998v). The minimum age of the Mount Negri Volcanics is less well constrained but in the type area of Mount Negri it contains Pb mineralization dated at c. 2922 Ma (data from Thorpe et al., 1992 interpreted by Huston et al., 2002a).

Several workers have confirmed significant lithological and chemical differences between the formations (Smithies, 1998a; Arndt et al., 2001; Smithies et al., 2004a, 2007a). Geochemical data from Arndt et al. (2001) indicate compositional gaps between the formations with respect to TiO_2, MgO, Ce, Sm, Nd, and Zr. These differences are confirmed by data in Smithies et al. (2007a), showing that, when compared to the Louden Volcanics, the Mount Negri Volcanics contain consistently higher concentrations of TiO_2, Na_2O, P_2O_5, REE (La, Ce, Pr, Nd, Sm, Eu, Gd, Tb, Dy), HFSE (Nb, Hf, Zr, Y, Th) and the chalcophile elements Cu and Zn, but consistently lower MgO and CaO contents and lower Mg#. Plots of Zr vs. Sr, Zr vs. La, Nb vs. La, and Sm vs. La show no overlaps in composition. Isotopic data (Table 5) indicate significant crustal components in both formations.

Arndt et al. (2001) argued that magmas for the Louden and Mount Negri Volcanics were derived by melting of continental lithosphere due to a mantle plume. The continental geochemical signature was a combination of enriched incompatible elements, negative Nb anomalies, negative ε_{Nd} (–1.1 to –2.4), and high SiO_2 versus Mg# (av. SiO_2 55.7%; av. Mg# 56, n = 34). Smithies et al. (2007a), on the other hand, suggested an alternative explanation, viz. that the mafic magmas of the Mallina Basin were derived from a mantle source, containing a homogeneous mix of Paleoarchean continental crust and crust chemically similar to the c. 3120 Ma Whundo Group. They argued this on the basis of relatively narrow ranges of La/Nb (2.7 to 3.7), La/Sm (4.8 to 5.8), La/Zr (0.13 to 0.16), and ε_{Nd} (–0.6 to –2.8) in basalts (exclude boninite-like rocks and gabbroic intrusions). Smithies et al. (2007a) considered that such relatively narrow ranges were unlikely to be the result of crustal contamination alone. They proposed that the mantle source would have required a previous subduction event, and that this might have been the c. 3010 Ma subduction suggested to explain the origin of the Whim Creek Group and Maitland River Supersuite.

Kialrah Rhyolite

In the western part of the Whim Creek greenstone belt, 2 km south of Warambie Homestead, the Kialrah Rhyolite is a 1000 m-thick unit of flow-banded and porphyritic rhyolite stratigraphically overlying or intruding the Louden Volcanics (Hickman, 1997a; 2002). The formation has a mapped strike length of 12 km but aeromagnetic data and field observations suggest that the formation may be laterally equivalent to the undated porphyritic rhyolite and dacite that intrudes the Whim Creek Group farther east. The Kialrah Rhyolite was initially dated at 2975 ± 4 Ma (GSWA 144261, Nelson, 1998v) but, based on the presence of a younger zircon group within the sample, Van Kranendonk et al. (2006) reinterpreted the crystallization age as 2943 ± 7 Ma and interpreted the c. 2975 Ma group to be xenocrysts. A sample of flow-banded rhyolite within the Mallina Formation east of the Ord Range was dated at 2948 ± 3 Ma (GSWA 169025, Nelson, 2002e), although based on the ages of the youngest concordant zircons in the group the true age of the rhyolite may be more similar to that of the formation in other areas and close to 2940 Ma. Flow-banded rhyolite is interlayered with the Mallina Formation between Mount Satirist and Station Peak, where a sample was dated at 2941 ± 4 Ma (GSWA 142892, Nelson, 2000b). The rhyolite units within the Mallina Formation are interpreted to correlate with the Kialrah Rhyolite in the Whim Creek greenstone belt, indicating a widespread event of rhyolite intrusion and volcanism at c. 2940 Ma. This age range coincides with intrusion of parts of the Sisters Supersuite and the interpreted age of D_8.

Tectonic events of the Whim Creek and Mallina Basins (D_5 to D_8)

The history of deformation in the Whim Creek and Mallina Basins is relatively complex and not fully understood. Structural interpretations of various parts of the Whim Creek and Mallina Basins have been provided by Krapez and Eisenlohr, (1998), Smithies (1998a, 1998b), Blewett (2000, 2002), Hickman et al. (2000), Smithies and Farrell (2000), Hickman and Smithies (2001), Huston et al. (2002b), Beintema (2003), and Hickman (2004a). Most of these accounts attempt to correlate deformation in the Whim Creek and Mallina Basins with late deformation in the WPS. For one event (D_8 below) correlation is relatively simple because large folds produced by this event have been mapped across the northwest Pilbara. These folds, which all trend northeast or east-northeast, include the Cleaverville Syncline, Prinsep Dome, Roebourne Synform, Sholl Anticline, Bradley Syncline, and Whim Creek Anticline (Fig. 8). Large folds of the same generation are interpreted to be present in the Mallina Basin (D_3, Smithies, 1998b), but owing to limited exposure have not been mapped in detail. This regional deformation event therefore provides a reference event for correlation of earlier and later structures.

D_5 deformation

The first structures to post-date deposition of the Gorge Greek Group and also deform the successions of the Mallina Basin and the Whim Creek Basin are assigned to the deformation event D_5. However, although D_5 structures in both these tectonic units were formed by north–south compression it is unknown if they are precisely the same age. For this reason they are distinguished as D_{5W} in the Whim Creek Basin and D_{5M} in the Mallina Basin. The observation that D_{5W} structures deform the Warambie Basalt but do not deform the overlying Red Hill Volcanics suggests that they formed at c. 3010 Ma. D_{5W} structures in the Warambie Basalt east of Mount Ada (Hickman, 2002) are low-angle thrust faults adjacent to the SSZ and might be local structures produced by c. 3010 Ma movement along the shear zone. In the Mallina Basin the minimum age of D_{5M} isoclinal folds are less evident because the depositional age of the succession deformed by them is poorly constrained. D_{5M} folds are deformed by c. 2955 Ma D_6 folds (absent from the Whim Creek greenstone belt), and are most easily distinguished in the lowest parts of the Mallina Basin succession, such as the transition succession between the Cleaverville Formation and the Constantine Sandstone and in some sections of the Constantine Sandstone (Fig. 25). They were interpreted by Smithies (1998b) as originally having steeply dipping axial planes striking east or northeast.

D_6 deformation

North-trending tight, upright, and locally overturned 'D_2 folds' (Smithies, 1998a, b; Smithies and Farrell, 2000) in the central part of the Mallina Basin are assigned to D_6 in this Report. These folds include the Powereena Anticline, Croydon Anticline, and large folds northwest of the Croydon Anticline and south of the Satirist Monzogranite. Farther east, various unnamed north-trending folds are also present in the Station Peak area. A steep axial plane schistosity (S_6) cuts across the refolded D_5 folds. This same schistosity is a prominent feature of the metamorphosed mafic and ultramafic rocks in the Croydon Group southeast of the Satirist Monzogranite (Smithies, 1998b). No major north-trending folds are present in the Whim Creek greenstone belt although much smaller scale D_7 folds in the Whim Creek area (see below) may be of similar age. The minimum age of D_6 is constrained by c. 2948 Ma intrusion by the Peawah Granodiorite, and the event is inferred to have taken place at c. 2955 Ma.

D_7 deformation

D_7 deformation (this Report) is assigned to thrusts and folds interpreted by Krapez and Eisenlohr (1998) to exist in the Whim Creek and Bookingarra Groups at Whim Creek, Mons Cupri, Good Luck Well, and Salt Creek. The evidence for the thrusts, which Krapez and Eisenlohr (1998) assigned to 'Phase 3' in their structural synthesis, is mainly stratigraphic and no outcropping thrust planes were described. Assuming the D_7 thrusting was a discrete deformation event later than D_6, it took place after deposition of the Cistern Formation at c. 2955 Ma, and most likely before deposition of the Louden and Mount Negri Volcanics, which were deposited in an extensional tectonic setting.

D_8 deformation

D_8 deformation was a significant and widespread tectonic event across the northwest Pilbara. Structures formed during D_8 include major northeast-trending folds in the WPS, principally the Roebourne Synform, Prinsep Dome, Cleaverville Syncline, and Bradley Syncline (Fig. 8), and large-scale structures in the De Grey Superbasin, for example the Whim Creek Anticline and northeast-trending folds in the Mallina Basin. In the central part of the Mallina Basin, D_8 folds were designated as 'D_3' by Smithies (1998a,b) and D_8 is interpreted to be equivalent to 'Phase 4' of Krapez and Eisenlohr (1998). However, geochronology in the Mallina Basin (Smithies, 1998a,b) establishes that the age of these structures must be 2955–2930 Ma, not 2906–2863 Ma as suggested by Krapez and Eisenlohr (1998). D_8 folds are oblique to regional-scale strike-slip faults and shear zones of the WPS and Mallina Basin, and are interpreted to be transpressional folds formed under regional northwest–southeast compression (North Pilbara Orogeny).

Minor D_8 structures in the WPS include a steeply dipping, east-northeast striking axial plane foliation (S_8) in the Prinsep Dome and in an anticline northwest of the Bradley Syncline. Minor D_8 folds deform S_1 in the Prinsep Dome. In the greenstones southwest of the Karratha Granodiorite (core of the elongate dome) they plunge southwest, whereas in greenstones at the other end of the dome they plunge northeast.

In the central Mallina Basin, D_8 deformation is characterized by large-scale, open to tight folds with steeply dipping, east-northeast-trending axial planes, and axes that typically plunge less than 45° east-northeast or west-southwest. D_8 structures in the Whim Creek greenstone belt mainly consist of east-northeast-trending upright folds with an axial plane cleavage dipping steeply southeast ('Phase 4A', Krapez and Eisenlohr, 1998). The folds are open to tight and plunge moderately east-northeast or west-southwest. Krapez and Eisenlohr (1998) interpreted the folds as second-order structures on the southeast limb of the Whim Creek Anticline, whereas on the northeast crest of the anticline, between Salt Creek and Mount Negri, minor D_8 folds have an M-fold profile (no prevailing vergence). Krapez and Eisenlohr (1998) considered that a number of major faults were reactivated during D_8, including the LF and the SSZ.

Geochronology on granitic intrusions of the Mallina Basin indicates that D_8 post-dated the c. 2948 Ma Peawah Granodiorite and either pre-dated or was contemporaneous with the 2938–2931 Ma Satirist Monzogranite (Smithies, 1998b). The age of D_8 is therefore interpreted as c. 2940 Ma.

Late Mesoarchean tectonic events of the West Pilbara Superterrane (D_8 to D_{10})

D_8 deformation

Major D_8 folds, including the Cleaverville Syncline, Prinsep Dome, Roebourne Synform, Sholl Anticline, and Bradley Syncline (Fig. 8), govern the outcrop pattern of the WPS and the overlying Gorge Creek Basin. Anticlinal structures expose the granitic complexes (Dampier, Cherratta, and Harding) and individual granitic intrusions (Karratha Granodiorite), whereas synclinal structures preserve the supracrustal succession within greenstone belts. Early interpretations of this pattern in the northwest Pilbara (Fitton et al., 1975; Hickman, 1981) concluded that it originated in the same way as the diapiric dome-and-syncline pattern of the east Pilbara, but that the northwest Pilbara domes had been modified by later horizontal deformation. Likewise, some structural interpretations of the Yilgarn Craton have interpreted formation of the numerous elongate domes in that craton to have involved solid state or magmatic diapiric doming (Archibald et al., 1978; Gee et al., 1981; Dalstra et al., 1998; Weinberg et al., 2003; Zibra, 2012; Caudery, 2014; Fenwick, 2014).

The granitic complexes of the northwest Pilbara are no longer interpreted to be the cores of diapiric domes deformed by later deformation. They lack the structural features characteristic of diapiric domes (e.g. radial stretching lineations and very steeply dipping sheared margins), and are now interpreted to be magmatic intrusions that, along with the greenstones, were folded by horizontal tectonic processes (Hickman, 2004a). All the major D_8 folds of the northwest Pilbara have northeast-trending axes, oblique to contemporaneous major shear zones, in particular the SSZ, which vary in strike from east–west to north-northeast–south-southwest. Geochronological evidence indicates that D_8 folds formed at 2940–2930 Ma, but the sense of strike-slip movement varied on different faults and shear zones. On the east–west striking SSZ and Black Hill Shear Zone the movement was dextral (Hickman, 2001b), whereas on north-northeast striking section of the LF and on the TTSZ it was sinistral (Krapez and Eisenlohr, 1998; Smithies, 1998a; Smithies et al., 2002; Beintema, 2003). In the EPT, Van Kranendonk (2008) used geochronology on synkinematic granites to estimate the age of sinistral strike-slip movement in the Lalla Rookh – Western Shaw Structural Corridor to be 2936–2928 Ma.

Minor D_8 folds are widespread in the central part of the Mallina Basin where they deform ultramafic–mafic sills of the c. 2950 Ma Langenbeck Suite. However, due to poor exposure many of these interpreted D_8 folds are visible only on aeromagnetic imagery, and some of them could be D_6 folds that have been rotated into a northeast alignment between D_8 strike-slip faults.

D9 deformation

The northern section of the MSZ, separating the Whundo Group from the Railway Supersuite, contains a steeply inclined tectonic foliation S_9, that truncates northeasterly trending D_8 folds of the Mount Sholl area. The same tectonic foliation is developed in the adjacent greenstones of the Whundo Group and in the Railway Supersuite on the east margin of the Cherratta Granitic Complex. The eastern side of the complex, within approximately 10 km of the MSZ, is dominated by the S_9 foliation trend, which is northwest to north. This is in contrast to the pattern of tectonic foliations in the central and western parts of the complex, which strike west-southwest parallel to the SSZ. It is concluded that D_9 shear in the eastern part of the complex and along the MSZ rotated the main west-southwest-striking foliation of the complex into a north-northwest–south-southeast alignment. Some shear zones within the complex include deformed granitic veins, containing zircons dated at 2944 ± 5 and 2925 ± 2 Ma (GSWA 136826, Nelson, 1997k). This suggests that D_9 took place at c. 2930 Ma.

D_{10} deformation

As noted by Krapez and Eisenlohr (1998), zircon geochronology on several rock units close to the SSZ has revealed a metamorphic disturbance event at about 2920 Ma. This event may have coincided with D_{10}. Minor D_{10} structures in the SSZ include dextral drag folding and isoclinal folding of mylonite lamination, and associated small-scale faulting and brecciation.

Sisters Supersuite

Across the northern Pilbara Craton the 2954–2919 Ma Sisters Supersuite is mainly composed of leucocratic high-K monzogranite interpreted to have been derived from partial melting of older crust, although Nd model ages indicate that crustal sources differed regionally. In the northwest Pilbara, where the supersuite intrudes the Mallina Basin, the supersuite includes alkaline granite of the Portree Suite, hornblende-granodiorite and high-Mg diorite (sanukitoid) of the Indee Suite, and c. 2950 Ma ultramafic–mafic layered intrusions of the Langenbeck Suite. Northwest of the Mallina Basin the Sisters Supersuite is largely restricted to a few small monzogranite intrusions in the Cherratta, Harding, and Caines Well granitic complexes (Table 2).

A suite of 2930–2925 Ma large ultramafic–mafic layered intrusions about 50 km south of Karratha was assigned to the Sisters Supersuite by Van Kranendonk et al. (2004, 2006) and named the Radley Suite. Inclusion of the Radley Suite in the Sisters Supersuite is questionable because there are reasons to suggest it may not be genetically related to the remainder of the supersuite. The Radley Suite is an ultramafic–mafic suite like the Langenbeck Suite, but it is approximately 25 million years younger. Additionally, the Radley Suite is restricted to a north-northeast-trending zone in the ST and Cherratta Granitic Complex, whereas almost all other intrusions of the supersuite are located in the Mallina Basin or western half of the EPT (Fig. 26). Under the present interpretation of southeast-migrating granitic intrusion from 3023 to 2919 Ma (Fig. 26), the Radley Suite is anomalous with regard to its age, composition, and location.

Geochronology has revealed that the first phase of the supersuite was the intrusion of ultramafic–mafic sills, high-Mg diorite (sanukitoid), and hornblende granodiorite between 2955 and 2945 Ma. The ultramafic–mafic sills, assigned to the Langenbeck Suite (Van Kranendonk et al., 2006) have not been dated, but their minimum age is constrained to 2948 Ma by the observation that they were deformed by north-trending D_6 folds in the central Mallina Basin. Intrusion of the Langenbeck Suite therefore took place at the same time as ultramafic–mafic volcanism of the Bookingarra Group in the Whim Creek greenstone belt (Van Kranendonk et al., 2002). This provides evidence of a significant c. 2950 Ma event of mafic intrusion and volcanism in the central and northwest parts of the Mallina Basin. In the southeast part of the basin this event is probably represented by the Salt Well Member of the Lalla Rookh Sandstone. Mafic volcanism of the Bookingarra Group was most likely associated with crustal extension and rifting in the basin. Some workers have interpreted the presence of komatiite and komatiitic basalt in the Bookingarra Group as evidence for a mantle plume event at this time (see Bookingarra Group), although no evidence of a c. 2950 Ma plume event has yet been recorded in the adjacent east Pilbara.

The high-Mg diorites and hornblende-granodiorites are assigned to the Indee Suite (Van Kranendonk et al., 2006) and mainly form relatively small intrusions outcropping over approximately 20 km^2. The high-temperature alkaline granites are restricted to the Portree Granitic Complex in the northern part of the central Mallina Basin and are assigned to the Portree Suite. Only one sample has been dated, and the zircon U–Pb data indicated an age of 2946 ± 6 Ma (GSWA 142889, Nelson, 1999g).

Langenbeck Suite

The Langenbeck Suite includes layered ultramafic–mafic sills of the Millindinna, Opaline Well, and Sherlock Intrusions and various unnamed gabbro intrusions in the central part of the Mallina Basin. Previously, the Sherlock Intrusion was included in the c. 2930 Ma Radley Suite (Van Kranendonk et al., 2006), but various geological features indicate that it was emplaced at c. 2950 Ma. Individual members of the Langenbeck Suite differ geochemically but they share relatively constant La/Sm and La/Zr ratios that cannot be accounted for through assimilation of any known Pilbara crustal component (Smithies et al., 2004a).

Sherlock Intrusion

The Sherlock Intrusion is a layered sill, locally over 1 km thick, that was intruded along the stratigraphic contact of the Caines Well Granitic Complex with the overlying Whim Creek Group. Prior to D_8 folding and later erosion the Sherlock Intrusion is estimated to have extended beneath an area of at least 1000 km^2.

Although the intrusion has not been dated, several factors indicate that it does not belong to the 2930–2924 Ma Radley Suite. Firstly, layering in the Sherlock Intrusion was folded by the c. 2940 Ma Whim Creek Anticline and is now inclined parallel to bedding in the adjacent 2960–2950 Ma stratigraphy. Secondly, the Sherlock Intrusion is a stratabound sill of great lateral extent, not a localized discordant vertical intrusion. In this respect it is very similar to the closely adjacent Opaline Well Intrusion. Thirdly, the Sherlock Intrusion underlies the ultramafic–volcanic succession of the c. 2950 Ma Bookingarra Group, suggesting it is a subvolcanic intrusion. Fourthly, at Sherlock Bay the Sherlock Intrusion is sheared out by major strike-slip movement on the SSZ, which is interpreted to be 2940–2920 Ma. Finally, mineralization in the Sherlock Intrusion is predominantly V–Ti and Ti–Fe, types not yet identified in the Radley Suite.

Millindinna Intrusion

Sills assigned to the Millindinna Intrusion are widespread in the lower part of the Croydon Group south of the Mallina Shear Zone, and particularly southeast of the Peawah Granodiorite in the vicinity of Millindinna Hill, Mount Langenbeck, and Mount Satirist. Most of the sills are deformed, along with the Croydon Group by D_6 and D_8 folds. Sills interpreted to belong to the Millindinna Intrusion also intrude the Croydon Group along the southeast margin of the central part of the Mallina Basin adjacent to the TTSZ. In the type area within a radius of 10 km from Millindinna Hill, the Millindinna Intrusion is a sill up to 400 m thick (Smithies and Farrell, 2000). A basal layer of serpentized lherzolite is overlain by a central unit of metapyroxenite (now actinolite–chlorite rock) and a thin upper layer of metamorphosed melanogabbro. Geological mapping has revealed that sills assigned to the Millindinna Intrusion do not all occupy a single stratigraphic level within the Croydon Group (Smithies and Farrell, 2000). Additionally, outside the type area many of the intrusions show no clear evidence of layering and may be either ultramafic or mafic. This may be due to intrusion at several stratigraphic levels or to local tectonic dismembering. Compositional variation in these sills is more varied than in the type area, and dominant rock types range from metamorphosed dunite to metamorphosed peridotite, pyroxenite, or gabbro.

Opaline Well Intrusion

In the Whim Creek greenstone belt the Opaline Well Intrusion (Smithies, 1998a) is present as sills up to 100 m thick within the Cistern Formation and as dykes up to 400 m wide in the underlying Whim Creek Group (Pike et al., 2006). Metamorphism is low grade and rock types include peridotite, peridotitic gabbro, olivine gabbro, gabbro, dolerite, and basalt. The basalt and dolerite typically contain acicular clinopyroxene, suggesting quenching of a crystal-rich magma, and many of the rocks are petrographically similar to mafic and ultramafic components of the overlying Louden Volcanics of the Bookingarra Group. The Opaline Well Intrusion is interpreted to be an intrusive component of the Bookingarra Group (Smithies, 1998a; Pike et al., 2006).

Indee Suite

Intrusions of the 2954–2945 Ma Indee Suite, which was once informally referred to as the 'Pilbara high-Mg diorite Suite' (obsolete name), intrude the Croydon Group in the central part of the Mallina Basin, and are also spatially associated with strike-slip faulting along the TTSZ. The distribution of the suite is shown by Smithies and Hickman (2003). Named intrusions of the suite are the Geemas Granodiorite, Jallagnoonina Granodiorite, Jones Well Granodiorite, Mallindra Well Granodiorite, Peawah Granodiorite, Wallareenya Granodiorite, and Toweranna Porphyry (Smithies and Champion, 2000). The Peawah Granodiorite is the largest intrusion, underlying an area of 180 km^2, and is a composite body that can be geochemically divided into eastern and western parts (Smithies and Champion, 2000). Geochronology indicates that the Indee Suite was coeval with the alkaline rocks of the Portree Suite.

The most common rock types of the Indee Suite are massive to moderately foliated, mesocratic, hornblende–biotite granodiorite and tonalite, and hornblende–biotite (–clinopyroxene) granodiorite. They range in texture from equigranular to seriate to porphyritic, with plagioclase phenocrysts up to 1 cm long. Fine-grained clinopyroxene- and orthopyroxene-bearing melanodiorite forms chilled margins to some intrusions.

The hornblende-granodiorite and high-Mg diorite were derived from melting of a mantle source along with a Th-, Zr-, and LREE-enriched crustal component (Smithies and Champion, 2000; Smithies, 2002a; Smithies et al., 2004a,b). Mantle enrichment could have taken place during a previous subduction event. Either the c. 3130 Ma subduction responsible for the oceanic arc succession of the Whundo Group (Smithies et al., 2005a) or the 3010–2980 Ma subduction that has been interpreted to have produced the Whim Creek Basin and Maitland River Supersuite (Blewett, 2002; Smith 2003; Hickman, 2004a, 2012; Pike et al., 2006). Melting of the enriched mantle source at c. 2950 Ma could have been due to ongoing subduction following the 3010–2980 Ma event, crustal extension, and thinning within a back-arc basin (decompression melting). Alternatively, it could be explained by the arrival of a mantle plume (Smithies and Champion, 2000). Other workers have also suggested a mantle plume origin for the c. 2950 Ma ultramafic–mafic succession in the upper part of the Bookingarra Group (Sun et al., 1991; Arndt et al., 2001; Pike, 2001; Pike et al, 2006).

Portree Suite

The high-temperature alkaline granitic intrusions that make up the Portree Suite are aeromagnetically expressed as a series of discrete nested plutons, referred to as the Portree Granitic Complex, north of the Mallina Shear Zone. Smithies and Champion (2000) attributed the Portree Suite to high-temperature melting of a metasomatized basalt crust, and suggested that the heat source was the same c. 2950 Ma mantle plume they inferred for generation of the Indee Suite.

Outcrop of the Portree Suite is very poor and the overall structure of the complex and the area it occupies (approximately 1000 km^2) are largely interpreted from aeromagnetic data. The complex contains at least six intrusions (Smithies and Hickman, 2003), but only two of these are exposed. Near the highway between Port Hedland and Roebourne the southern intrusion, named the Portree Granite by Smithies (1998a) is composed of massive, medium-grained leucocratic tonalite and granodiorite underlying an area of approximately 500 km^2. A sample of this intrusion was dated at 2946 ± 6 Ma (GSWA 142889, Nelson, 1999g), but also contained an older group of zircons dated at 2978 ± 4 Ma, plus a single inherited zircon grain dated at c. 3029 Ma. The southern intrusion is abnormally sodic in composition (>5 wt% Na_2O) and composed of quartz, microcline, sodic-plagioclase, sodic clinopyroxene, and minor biotite. Blue alkali-amphibole is a late-magmatic to subsolidus replacement of clinopyroxene. The rocks are characterized by steep REE profiles with no Eu anomaly (Smithies and Champion, 1998).

An unnamed intrusion of the Portree Granitic Complex adjacent to the SSZ is a coarse-grained alkali granite, containing tabular perthite up to 1 cm in size, locally crowded with inclusions of sodic-plagioclase. The compositions and ages of the remaining concealed intrusions are unknown.

High-K monzogranites

Nd isotopic data from c. 2940 Ma high-K monzogranites of the Sisters Supersuite indicate multiple sources. Southeast of the Mallina Basin, these sources include Paleoarchean rocks of the EPT, whereas within the Mallina Basin the sources were younger than c. 3200 Ma, and may have included the 3130–3093 Ma ST.

Radley Suite

Large mafic intrusions were first recognized in the northwest Pilbara as a result of GSWA mapping in the 1960s. Ryan (1965) described the Andover Intrusion near Roebourne, and Williams (1966) identified the Munni Munni and Maitland Intrusions 50 km southwest of Roebourne. Subsequent work led to descriptions of similar intrusions at Radio Hill (Richardson, 1976) and Mount Sholl and Bullock Hide Well (Hickman, 1983), and to the naming and description of intrusions at Dingo Well and North Whundo (Hickman, 1997b, 2001b; Hickman and Kojan, 2003). However, geochronology has indicated that not all the large mafic intrusions of the northwest Pilbara belong to the 2930–2924 Ma Radley Suite. The Bullock Hide Intrusion was dated at 3122 ± 3 Ma (Wingate and Hickman, 2009b) and geochronology on the Andover Intrusion and intrusions into it indicates that it belongs to the 3023–3012 Ma Orpheus Supersuite. Van Kranendonk et al. (2006) included the Sherlock Intrusion in the Radley Suite, but this Report presents evidence supporting assignment to the Langenbeck Suite (see Sherlock Intrusion, Langenbeck Suite).

Intrusions of the Radley Suite were emplaced along a 180 km-long, north-northeast-trending zone immediately south of the SSZ. The intrusions are either lopoliths or funnel-shaped bodies (Hoatson et al., 1992; Hoatson, 1998; Hoatson and Sun, 2002), the largest being the Munni Munni Intrusion, which is c. 5500 m thick (Donaldson, 1974). This intrusion outcrops over an area of 150 km^2 and gravity data indicate an additional southwest section that is concealed by the Fortescue Group over a similar area. Donaldson (1974) recognized more than 40 cyclic units in the ultramafic zone, in which individual peridotite layers grade upwards into clinopyroxenite. Fine-scale layering in the clinopyroxenite consists of olivine-rich layers alternating with pyroxenite at intervals of 2 to 200 mm. Donaldson (1974) stated that the upper gabbroic zone of the intrusion was not rhythmically layered, but the rock is locally laminated by virtue of oriented plagioclase laths.

Most intrusions of the Radley Suite comprise a lower section of ultramafic layers (dunite, peridotite, and pyroxenite) overlain by layered units of gabbro, leucogabbro, norite, and more rarely anorthosite and granophyre. Gabbroic units of the Radio Hill and North Whundo intrusions are overlain by granitic rocks. In the case of the North Whundo Intrusion the upper granitic component, the Yannery Granite (Hickman and Kojan, 2003), is a hornblende-monzogranite dated at 2930 ± 4 Ma (N3162, Smith, 2003). The granite related to the Radio Hill Intrusion was dated at 2929 ± 13 Ma (N4450, Smith, 2003). Mineral exploration (summarized by Ruddock, 1999) has resulted in discoveries of significant mineralization in the Radio Hill Intrusion (Ni–Cu–Co) and in the Munni Munni Intrusion (platinum group elements [PGE], Au, and Ni–Cu). The Radio Hill Intrusion has been mined and the Munni Munni Intrusion has been extensively explored.

The Radley Suite is contained within a wide, north-northeast-trending zone that coincides with gravity and magnetic lineaments, with a number of exposed faults (e.g. Zebra Hill Fault; Hickman and Kojan, 2003), and with possible fault control of the lower Fortescue Group. This suggests a deep structural feature, and this may have originated as an early branch of the SSZ. Figure 3 shows an inferred fault, the newly named Woodbrook Fault, along the north-northeast-trending geophysical lineament. Previous authors have also suggested structural control over emplacement of the Radley Suite (Mathison and Marshall, 1981; Hoatson and Sun, 2002). A major fault zone on this alignments might also account for intrusion of the c. 3068 Ma Cliff Pool Tonalite (Elizabeth Hill Supersuite) in this zone during the Prinsep Orogeny (see SSZ).

The largest intrusion of the Radley Suite, the Munni Munni Intrusion, is aligned along the trend of the inferred Woodbrook Fault and its eastern margin is at least partly fault controlled. Most Radley Suite intrusions contain locally abundant breccia (Hickman, 1997b) in which large angular clasts of country rock, in most instances metabasalt, are set in a matrix of gabbro, dolerite, or ultramafic rock. This suggests intrusion during faulting or into existing zones of fault breccia.

North Pilbara Orogeny

The 2955–2919 Ma North Pilbara Orogeny was one of two separate episodes of major deformation and metamorphism in the Pilbara Craton between 2955 and 2900 Ma. This orogenic episode affected the northwest Pilbara and western section of the east Pilbara, whereas the c. 2900 Ma Mosquito Creek Orogeny was restricted to the southeast Pilbara. The North Pilbara Orogeny was the final tectonic stage in the 290-million-year evolution of the CPTZ (Hickman, 1999, 2004a; Hickman et al., 2001), and included all deformation from D_6 to D_{10}. The orogeny was accompanied by intrusion of ultramafic, mafic, and felsic intrusive rocks of the 2954–2919 Ma Sisters Supersuite.

Most workers who have studied the structural geology of the northwest Pilbara agree that deformation between 2955 and 2919 Ma was the result of north–south or northwest–southeast compression, and most of them have related this to plate convergence (Krapez and Eisenlohr, 1998; Smith et al., 1998; Blewett, 2002; Van Kranendonk et al., 2002; Beintema, 2003; Smith, 2003; Hickman, 2004a, 2012; Pike et al., 2006; Hickman et al., 2006a, 2010). The converging northwest plate was probably oceanic and is interpreted to have been subducted beneath the northwest margin of the Pilbara Craton. Structural evidence, including the widespread development of major strike-slip faults and transpressional folds, indicates that convergence was oblique, changing from north–south to northwest–southeast between 2955 to 2919 Ma. Syndepositional normal faults within the central part of the Mallina Basin (Smithies et al., 2001b) establish that crustal extension was locally important early in the orogenic phase of basin evolution, and much of the Mallina Formation was deposited at this time.

Structural and metamorphic evidence of the North Pilbara Orogeny is also evident in the western half of the EPT. The Pilbara Well, Wodgina, and Pincunah greenstone belts contain refolded isoclinal recumbent folds that are similar in age to the D_6 folds of the Mallina Basin. Termination of the North Pilbara Orogeny was marked by c. 2919 Ma dextral strike-slip on the SSZ accompanied by metamorphism. This effectively marked the completion of the cratonization process in the northwest part of the Pilbara Craton.

Central Pilbara Tectonic Zone

The Mesoarchean crustal evolution of the northwest Pilbara Craton included numerous episodes of deformation, involving extensional faulting, thrusting and folding, and differential strike-slip movement of the terranes and basins (Barley, 1987; Krapez and Barley, 1987; Krapez, 1993; Hickman, 1997a, 1999, 2001a,b, 2004a; Krapez and Eisenlohr, 1998; Smith et al., 1998; Sun and Hickman, 1998; Van Kranendonk and Collins, 1998; Van Kranendonk et al., 2002; Beintema, 2003). This deformation is concentrated within the east-northeast-trending CPTZ (Hickman, 1999, 2001a, 2004a; Smithies and Farrell, 2000; Beintema et al., 2001, 2003; Hickman et al., 2001; Van Kranendonk et al., 2002; Beintema, 2003). The tectonic evolution of the CPTZ, and the terranes and basins that formed within it (RT and ST; Whim Creek and Mallina Basins), are unique to this linear zone within the Pilbara Craton. Major strike-slip movements during the evolution of the CPTZ changed the relative positions of its terranes and basins with the result that present juxtaposition does not reflect depositional relationships.

From c. 3235 to 2920 Ma the CPTZ (Fig. 3) exercised major controls over the crustal evolution of the northwest Pilbara. Tectonically and magmatically active for c. 300 million years, and including two orogenic events, the CPTZ has certain features of an orogenic belt, but does not contain any exposed rocks metamorphosed above amphibolite facies. The zone initially developed as a post-3235 Ma rift system in the EPT, which quickly developed into a deep, northeast-southwest-trending rift basin (East Pilbara Terrane Rifting Event). Separation of the EPT by the rift established the KT in the northwest and the EPT in the southeast. The northwest boundary of the CPTZ is the SSZ as far east as Cape Thouin and its southeast boundary is the TTSZ (Fig. 3). East from the Ord Range the continuation of the TTSZ is named the Pardoo Fault. Both structural boundaries are interpreted to have originated as normal faults during the rifting but they were subsequently reactivated during a sequence of tectonic events until 2920 Ma. Nd T_{DM} model ages from intrusive and volcanic rocks within the CPTZ differ markedly from Nd T_{DM} model ages in the adjacent EPT and KT. Whereas the latter are typically >3400 Ma those within the CPTZ are almost invariably <3300 Ma (Van Kranendonk et al., 2007b). This indicates that for most of its evolutionary history the CPTZ contained little or no underlying Paleoarchean crust. Instead, the oldest crust beneath CPTZ was c. 3200 Ma basaltic crust that is now exposed only in the RT. Preservation and exposure in the RT, which lies northwest of the CPTZ, is attributed post-3160 Ma thrusting of a section of the c. 3200 Ma basaltic crust onto the KT.

The CPTZ crust was initially much thinner and less rigid than the continental crust of the two terranes. As a consequence, from 3200 to c. 2955 Ma the CPTZ was preferentially deformed and reworked (including by subduction) during all tectonic events affecting the northwest part of the Pilbara Craton. Between 2955 and 2920 Ma deformation, metamorphism, and igneous intrusion of the North Pilbara Orogeny marked final collision and accretion of the EPT to the WPS. From c. 2955 Ma onwards the EPT was no longer isolated from the dynamic effects of convergence along the northwest margin of the Pilbara Craton and it became tectonically active. This contrasts markedly to the preceding c. 250 million years when EPT recorded no significant deformation (Van Kranendonk et al., 2002; Hickman, 2004a).

Some indication of the magnitude of strike-slip movement within the CPTZ is provided by major stratigraphic mismatches across the faults and shear zones; for example between the KT and ST across the SSZ, and between the Mallina Basin and the EPT across the TTSZ (and its northeast continuation as the Pardoo Fault). Both these long-lived structures exercised considerable influence over the evolution of the CPTZ. A third fault, the LF, separates the central part of the Mallina Basin from the Whim Creek Basin and underlying Caines Well Granitic Complex. Isotopic data indicate significant differences in the ages of the crust underlying the basins. In many respects the isotopic data for the area between the LF and the SSZ are similar to data north of the SSZ.

Sholl Shear Zone

Hickman (1983) and Barley (1987) described the Sholl Shear Zone (SSZ) as a major sinistral strike-slip fault, and geophysical evidence indicates that its onshore extent is about 350 km (Hickman et al., 2001a). Its total length, including sections concealed by the Fortescue Basin in the southwest and offshore beneath the Northern Carnarvon Basin to the northeast is likely to be c. 600 km. In areas where the shear zone is best exposed, as in the Nickol River area 20 km southeast of Karratha, it is a subvertical unit of mylonite and schist between 1 and 2 km wide (Hickman, 1997b, 2004a; Hickman et al., 1998).

Multiple episodes of movement on the SSZ have been established by investigations between Port Hedland and Cape Preston (Hickman et al., 1998, 2000; Krapez and Eisenlohr, 1998; Smith et al., 1998; Smithies, 1998a; Hickman, 2001b, 2004a; Blewett, 2002, Van Kranendonk et al., 2002). Evidence from the Sulphur Springs Group (Vearncombe et al., 1998) and the unconformably overlying Soanesville Group (Eriksson, 1981; Wilhelmij and Dunlop, 1984) indicates that the earliest stages of the EPT Rifting Event commenced at c. 3235 Ma, although breakup of the EPT is considered to have begun at c. 3220 Ma, which marks final construction of the EPT. It is likely that the SSZ originated as one of many early normal faults that developed into a major break defining the northwest margin of the CPTZ. The crustal extension which produced the earliest rifting within the EPT has been attributed to the 3275–3223 Ma mantle plume responsible for the Sulphur Springs Group and the Cleland Supersuite (Van Kranendonk et al., 2006, 2007a,b, 2010; Hickman, 2012). The reason for the inferred predominance of northeast–southwest-striking faults in the EPT is uncertain. One possibility is that it was controlled by c. 3250 Ma regional heterogeneities in the crust. Alternatively, uplift and extension above the plume head may have moved across the EPT in a northwest–southeast direction. In either scenario, during rifting and separation of the terranes the progenitor of the SSZ would have remained an extensional structure until the microplates began to reconverge at c. 3160 Ma (see Current tectonic interpretation). Most of the sinistral movement on the SSZ took place between the c. 3160 and 3070 Ma collision of the ST with the KT (Prinsep Orogeny). The interpretation that part of the c. 3200 Ma basaltic crust between the two terranes is now preserved as the RT, and that the 3160–3070 Ma convergence included thrusting of this northwards across the KT, suggests that similar thrusting also occurred on the SSZ.

Between c. 3050 Ma and c. 2970 Ma, during deposition of the Gorge Creek, Whim Creek, and lower Croydon groups, the SSZ again occupied extensional settings, although local tight to isoclinal folding of the Gorge Creek Group at c. 3015 Ma prior to deposition of the Whim Creek Group establishes a brief episode of compression. Relative uplift on its northern side is indicated by the stratigraphy and sedimentology of the Whim Creek Group (Barley, 1987; Hickman, 2002; Pike et al., 2002, 2006). North–south plate convergence prior to 2950 Ma produced isoclinal recumbent folds in the Mallina Basin and on the western margin of the EPT. At the same time, sinistral strike-slip movement took place on all the major shear zones of the CPTZ, including the SSZ (Krapez and Eisenlohr, 1998; Smith et al., 1998; Blewett, 2002; Beintema, 2003; Hickman, 2004a). The final phase of the North Pilbara Orogeny included c. 2920 Ma dextral strike-slip movement of 30–40 km on the SSZ (Hickman et al., 1998, 2000; Smithies, 1998a; Hickman, 2001b; Blewett, 2002).

Geophysical data suggest the existence of another major splay fault or shear zone branching southwest from the SSZ near Woodbrook, but concealed by the Fortescue Group. This Report refers to the inferred concealed fault as the Woodbrook Fault (Fig. 3). Numerous minor northeast-striking faults deform the Whundo Group immediately west of this inferred major structure and previous authors have suggested structural control over emplacement of the Radley Suite (Mathison and Marshall, 1981; Hoatson and Sun, 2002). Apart from influencing emplacement of these major layered intrusions, the Woodbrook Fault might have constituted a boundary between the central Mallina Basin, underlain by little or no pre-3200 Ma crust, and an area to the northwest that does contain some Paleoarchean crust. A change of crustal composition across this inferred tectonic contact is supported by the presence of the c. 3236 Ma Tarlwa Pool Tonalite (GSWA 142535, Nelson, 1998l; Van Kranendonk et al., 2006) west of the interpreted Woodbrook Fault. This tonalite is the oldest rock dated from south of the SSZ and is correlated with the 3274–3223 Ma Cleland Supersuite of the EPT. Additionally, the Nd model age of the Tarlwa Pool Tonalite is 3360 Ma (Smithies et al., 2007a), which is similar to many granitic rocks north of the SSZ. Like the 3270–3261 Ma Karratha Granodiorite of the KT, the Tarlwa Pool Tonalite might have originated in the EPT.

Relative uplift on the northern side of the SSZ is suggested by an abrupt increase of metamorphic grade across it, from greenschist facies in the south to amphibolites facies in the north. Although this grade difference could result merely from strike-slip juxtaposition of different terranes, there is also sedimentological evidence that during deposition of the Whim Creek Group the shear zone was expressed as a south facing scarp. Alluvial and colluvial fans of conglomerate and sandstone are developed in the Whim Creek Group immediately south of the shear zone (Barley, 1987; Hickman, 2002). Smithies (1998a) reported similar conglomerate in the Louden Volcanics of the Bookingarra Group adjacent to the shear zone.

Excellent exposures of the SSZ are provided by rock pavements in the bed of the Nickol River (Fig. 29a). Along much of its length between Roebourne and Devil Creek the shear zone is defined by a belt of mylonite between 1 and 2 km wide. At Nickol River the shear zone is predominantly composed of felsic mylonite derived from extremely sheared granitic rocks, although there are also thin layers of amphibolite. The mylonitic lamination is folded by tight, west-plunging Z-folds, interpreted to be related to the late dextral movement, along with earlier isoclinal folds (Figs 29b,c). All these structures are displaced by late brittle fractures, which are locally filled by pseudotachylite.

Tabba Tabba Shear Zone

The southeast boundary of the CPTZ is defined by a zone of intense faulting and shearing that, form northeast to southwest, includes the Pardoo Fault, the Tabba Tabba Shear Zone (TTSZ), and the faulted northwest margin of the Pilbara Well greenstone belt (Figs 3, 4). The Pardoo Fault (Fig. 3) is very poorly exposed, and mainly interpreted from magnetic data. The fault zone on the northwest margin of the Pilbara Well greenstone belt (Fig. 7) is substantially concealed by Neoarchean volcanic of the Fortescue Basin. The central section of the TTSZ is better exposed and has been studied by several workers (Beintema et al., 2001, 2003; Hickman et al., 2001; Smithies et al., 2001a, 2002; Smithies and Champion, 2002; Beintema. 2003).

The TTSZ is up to 3 km wide and composed of foliated to mylonitic rocks with protoliths, including a wide age range of granitic and ultramafic–mafic intrusive rocks, felsic and mafic volcanic rocks, and sedimentary rocks, including the Cleaverville Formation and the Mallina Formation. The shear zone is interpreted to have originated as a late Paleoarchean extensional fault in the EPT. Evidence for early movement along the TTSZ includes gabbro and diorite intrusions dated at c. 3235 Ma (Beintema et al., 2001, 2003; Beintema, 2003). Sheared granitic rocks in the TTSZ, dated at c. 3250 Ma (Beintema, 2003), belong to the regionally extensive Cleland Supersuite and in this Report are interpreted to pre-date the TTSZ.

Farther east, within the EPT, rift-related intrusion of dolerite and gabbro continued to c. 3190 Ma (Van Kranendonk et al., 2010; Hickman, 2012). On the southeast side of the TTSZ the gabbros intruded granodiorite and granite dated between 3257 and 3250 Ma (Beintema, 2003; Beintema et al., 2003; GSWA 160745, Nelson, 2001g) and metamorphosed felsic volcanic rocks dated at 3253 ± 4 Ma (GSWA 160258, Wingate et al., 2010). The felsic volcanic rocks contain VHMS mineralization, as does the Sulphur Springs Group in the type area around the Strelley Monzogranite. Prior to rifting and separation along the TTSZ and SSZ, it is interpreted that the Sulphur Springs Group and Cleland Supersuite of the EPT were contiguous with the Roebourne Group and Karratha Granodiorite of the KT.

Geochronology has established that the Carlindi Granitic Complex includes the c. 3430 Tambina and c. 3465 Ma Callina Supersuites (Van Kranendonk et al., 2006; Hickman and Van Kranendonk, 2012). Xenocrystic zircon in the c. 3250 granites of the shear zone include zircon ages consistent with derivation from such older Paleoarchean sources (GSWA 160745, Nelson, 2001g; KB 263, KB 312, KB 770, KB 779, Beintema, 2003). By contrast, granitic intrusions in the Mallina Basin immediately northwest of the TTSZ contain no xenocrystic zircon older than c. 3246 Ma (GSWA 160727,

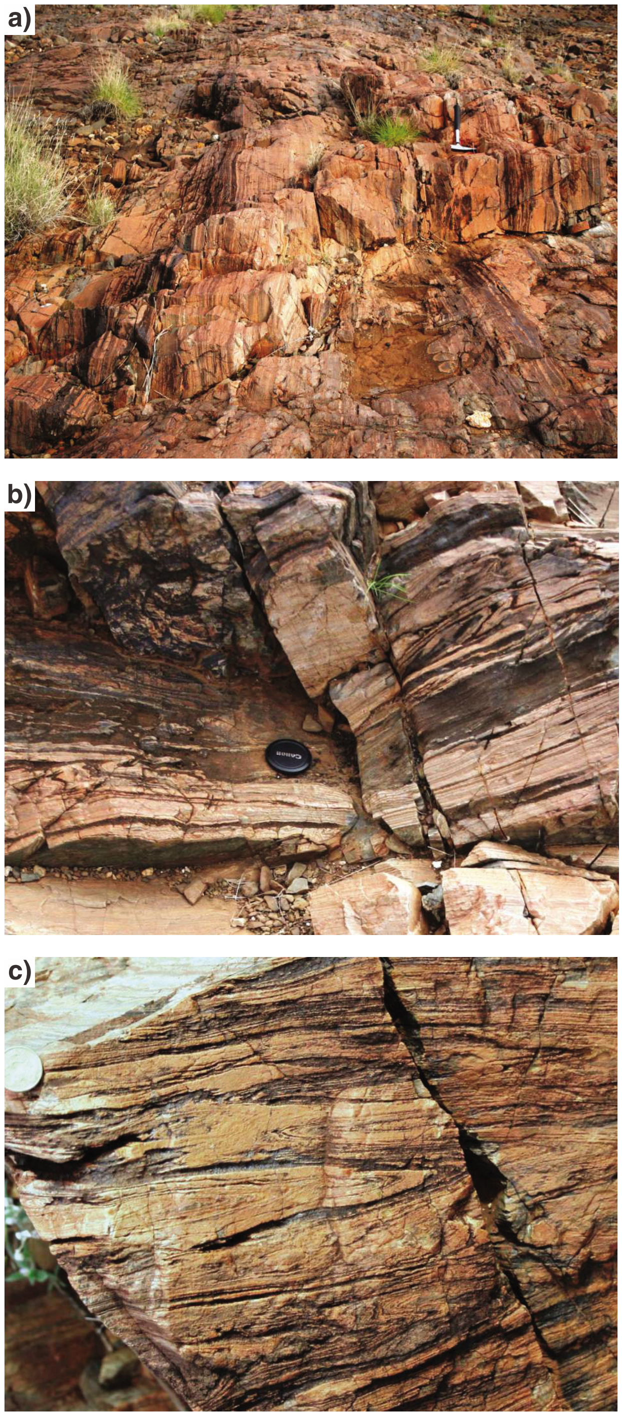


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Figure 29. Mylonite in the Sholl Shear Zone at Nickol River (Zone 50, MGA 494950E 7689700N): a) extensive outcrop in the Nickol River, showing vertically inclined layers of felsic and mafic mylonite. Felsic mylonite (pale colour) derived from intensely sheared granitic rocks and mafic mylonite, originating from similarly sheared greenstones; b) view of mylonite layering from above, showing minor folding of some layers indicating dextral shear sense (2940–2920 Ma), scale: 5 cm-diameter lens cap; c) isoclinal folding of mylonite fabric (scale: 2 cm-diameter coin)

Nelson, 2001c) and the single 3246 Ma grain detected was most likely inherited from detrital zircon in the Mallina Formation rather than from basement rocks. Thus, the TTSZ coincides with the northwest edge of the EPT, and marks a major change in the crustal architecture of the northern Pilbara Craton.

The second episode of deformation along the TTSZ took place at or shortly after c. 3160 Ma, when separation of the EPT and KT was replaced by convergence. A metamorphic event happened in the KT at this time (Kiyokawa, 1993; Beintema, 2003) and c. 3165 Ma tonalite (Flat rocks Tonalite) was intruded at the southwest end of the Pilbara Well greenstone belt. Subduction in the centre of the basaltic basin between the two converging terranes gave rise to the 3130–3110 Whundo Group and the 3130–3090 Ma Railway Supersuite. The southeast extent of these units across the CPTZ and beneath the Mallina Basin is uncertain, but the TTSZ is interpreted to record a 3160–3090 Ma event based on ages of xenocrystic zircon in 2950–2930 Ma granitic rocks (KB 746, KB 770, Beintema, 2003). Beintema et al. (2001, 2003) described evidence of early dextral movement in the shear zone and suggested that this happened at or before c. 3115 Ma. However, this timing is difficult to reconcile with the northeast strike of the shear zone and north–south compression responsible for 3160–3070 Ma and 3130–3110 Ma events in the RT and ST (Van Kranendonk et al., 2006). An alternative explanation is that dextral shear took place during the formation of north-trending folds in the Mallina Basin (D_6 in this Report).

The 3070–3060 Ma Prinsep Orogeny, caused by collision of the WPS with the EPT (Van Kranendonk et al., 2006), must have impacted on the TTSZ. Although neither intrusive rocks nor small-scale structures of this age have yet been recognized on the shear zone, Beintema (2003) reported c. 3050 Ma xenocrystic zircon in a c. 2930 Ma granite (KB 770) and $^{207}Pb/^{206}Pb$ overprint events between 3069 and 3045 Ma (KB 265, KB 779), affecting c. 3250 Ma granites. It is probable that structural evidence of c. 3060 Ma deformation on the TTSZ was destroyed by sinistral strike-slip and vertical movements between 2955 and 2930 Ma.

Most workers who have studied the structural geology of the TTSZ have reported evidence of major c. 2950 Ma sinistral strike-slip movement (Beintema et al., 2001, 2003; Hickman et al., 2001; Smithies and Champion, 2002; Smithies et al., 2002; Beintema. 2003). Minor structures, such as C–S fabrics and rotated feldspar phenocrysts, indicate a sinistral component of displacement. Additionally, northwest-plunging mineral and stretching lineations indicate normal movement consistent with juxtaposition of the Croydon Group in the Mallina Basin and Paleoarchean rocks of the EPT to the southeast. A concentration of 2955–2940 Ma granitic and mafic intrusions northeast of Wallareenya is attributed to a local change of strike in the TTSZ (northeast instead of east-northeast) that established an extensional releasing bend under sinistral shear (Smithies and Champion, 2002). Smithies et al. (2002) commented that the age of these intrusions closely coincided with the second depositional event in the Mallina Basin. The intrusion of gabbro and high-Mg diorite along the TTSZ reveals that it was a deep-crustal structure controlling the migration and emplacement of mantle-derived magmas (Smithies and Champion, 2000).

The Pardoo Fault is a northeast continuation of the TTSZ, and relative sinistral displacement of the Cleaverville Formation between the Ord Range and Goldsworthy is apparently at least 30 km. In this area the Pardoo Fault separates the Mallina Formation from the Warrawoona Group (Smithies, 2004), suggesting downthrow of at least 5 km on the northern side of the fault.

Evidence of late dextral movement on the TTSZ is provided by crenulations and steeply southwest-plunging folds that overprint the main shear foliation related to the c. 2950 Ma sinistral movement (Beintema et al., 2001; Hickman et al., 2001). Beintema et al. (2001) interpreted this to be post-2930 Ma, which suggests it could have taken place at the same time as late dextral movement on the SSZ, c. 2920 Ma.

Loudens Fault

Imagery from airborne magnetic data shows a major lineament extending 120 km along the southeast and southwest margins of the Whim Creek greenstone belt. The southeast section of this lineament, coinciding with the Loudens Fault (LF) recognized by previous workers (Barley, 1987; Krapez and Eisenlohr, 1998; Smithies, 1998a), is a boundary between metabasalt of the greenstone belt and metasedimentary rocks of the Mallina Basin. The southeast sector, coinciding with an interpreted concealed fault (Hickman, 2000), is a concealed boundary between metabasalt and BIF of the greenstone belt and interpreted metavolcanic rocks of the Whundo Group underlying the Fortescue Group. This fault, here interpreted to be a western continuation of the LF, joins the SSZ west of Mount Ada. Previously, the LF was interpreted to leave the southern margin of Whim Creek greenstone belt and continue south-southwest into the Mallina Basin. However, this fault has only limited geophysical expression, and because it extends farther south-southwest into the Fortescue Group was probably entirely formed during the 2775–2752 Ma Fortescue Rifting Event.

Evidence relating to the structural significance of the LF has been discussed earlier in this Report (e.g. Whim Creek and Mallina Basins). Previous stratigraphic correlations across the fault were based on inconclusive lithological or geochemical evidence, and the only correlation that is supported by both lithology and geochronology involves the c. 2940 Ma Kialrah Rhyolite, which is present on both sides of the LF. The Whim Creek greenstone belt and the Mallina Basin were most likely widely separated until c. 2940 Ma. A major deformation event (D_8), involving folding and strike-slip faulting, affected the entire northwest Pilbara Craton at c. 2940 Ma, and movement along the LF was most likely included in this deformation.

Isotopic data (Table 4) indicate that the area north of the LF is underlain by Paleoarchean crust (Fig. 3), whereas the central Mallina Basin to the south is not. This supports a conclusion that the LF forms a significant crustal boundary. However, present evidence is insufficient to determine if this boundary originated during the c. 3200 Ma EPT Rifting Event or was produced by strike-slip between 3010 Ma and 2940 Ma. Major strike-slip movement on the LF at c. 2940 Ma would explain the present juxtaposition of 3015– 2950 Ma successions with differing stratigraphy and structural history.

Post-orogenic events

Following the North Pilbara Orogeny, the Pilbara Craton was deeply eroded between 2895 Ma and 2775 Ma. There are no known sedimentary or volcanic formations from this period although some undated clastic sedimentary units at the base of the Fortescue Group, which are currently assigned to the Bellary Formation may have been produced by this erosion. Geochronology indicates the presence of c. 2870 Ma epithermal Pb and Ag mineralization immediately north of the Munni Munni Intrusion (Elizabeth Hill deposits; Marshall, 2000; Huston et al., 2002a,b) and c. 2890 Ma Ni–Cu mineralization in the Radio Hill Intrusion (Frick et al., 2001). However, the main event to take place after the North Pilbara Orogeny was the intrusion of the Split Rock Supersuite between 2851 Ma and 2831 Ma.

Split Rock Supersuite

The Split Rock Supersuite is composed of highly fractionated monzogranites that were emplaced in a linear, southeast–northwest-trending belt across the Kurrana Terrane and EPT and into the eastern part of the Mallina Basin. Champion and Smithies (2001) described the supersuite as consisting of silica-rich granitic rocks (>73 wt% SiO_2), with high LILE contents, moderate to large negative Eu anomalies, depleted in Sr, and enriched in Y. The moderate to high Rb, Rb/Sr, Rb/Ba, Ca/Sr, and K_2O/Na_2O, and low K/Rb ratios are consistent with fractionation of partial melts of older granitic crust (De Laeter and Blockley, 1972; De Laeter et al., 1975; Blockley, 1980; Davy and Lewis, 1986; Champion and Smithies, 2001). Nd model ages between 3700 Ma and 3400 Ma (Bickle et al., 1989) suggest the sources included Eoarchean crust underlying the Pilbara Supergroup.

Intrusions of the Split Rock Supersuite have been dated by the SHRIMP U–Pb zircon method at 2851–2831Ma. Previously, the age range of the supersuite was stated to extend from c. 2890 Ma (Van Kranendonk et al., 2006), based on U–Pb dates on cassiterite and tantalite from pegmatites in the granitic complexes (Kinny, 2000). However, the main group of cassiterite and tantalite ages was c. 2840 Ma (Kinny, 2000; Sweetapple and Collins, 2002). The present interpretation is that older ages on cassiterite and tantalite probably came from pegmatites intruded during the final stage of the North Pilbara Orogeny, and that these do not belong to the Split Rock Supersuite.

The supersuite is post-orogenic but the crustal processes involved in magma generation are uncertain. The 2851–2831 Ma age falls within the period when the Pilbara Craton underwent deep erosion following the North Pilbara Orogeny, and when the crust was probably subject to post-orogenic extension. Field evidence indicates that at least some of the granites and related pegmatites were intruded as subhorizontal sheets into the earlier granites and greenstones of the craton (Hickman, in prep.). These various features suggest reduced vertical pressures within the crust where the granites were intruded, and crustal thinning is likely to have increased temperatures, promoting melting of the lower crust. A second possibility is that orogenic crustal thickening was followed by melting of the lower crust, resulting from delamination. A third possibility is that the southeast–northwest alignment of the intrusions across the craton traces the course of 2851–2831 Ma drift of the craton across a hot spot. The aerial extent and age range of the Split Rock Supersuite supports an analogy with movement of the North American plate across the Snake River Plain – Yellowstone hot spot.

Three intrusions interpreted to belong to the Split Rock Supersuite intrude the Mallina Basin, all within the Pippingarra Granitic Complex: Myanna Leucogranite, Tabba Tabba Leucogranite, and Thelman Monzogranite. The Myanna Leucogranite is poorly exposed over an area of almost 1000 km^2 south of Port Hedland, and composed of massive to weakly foliated biotite–muscovite monzogranite, locally with quartz and K-feldspar phenocrysts. The Tabba Tabba Leucogranite is a biotite (–muscovite) granite, massive to weakly foliated with seriate to equigranular texture. It intruded the Mallina Formation immediately northwest of the TTSZ, and is moderately to strongly foliated along its southeast margin. This establishes reactivation of the shear zone after the North Pilbara Orogeny, and the movement could have taken place during 2775–2760 Ma rifting of the craton during deposition of the Fortescue Group. The Thelman Monzogranite occupies less than 10 km^2, and is a massive to weakly foliated muscovite–biotite (–garnet) monzogranite. The rock is fine to medium grained, locally feldspar porphyritic, and in many outcrops contains late-stage pegmatite dykes and sills. Many parts of the Pippingarra Granitic Complex are extensively intruded by pegmatite dykes and subhorizontal sills, most of which are interpreted to be related to the Split Rock Supersuite. Kinny (2000) dated tantalite from pegmatite in the Strelley mining area (northern section of TTSZ) at 2836 ± 26 Ma. Some of the pegmatite bodies are several hundred metres wide, and their composition ranges from relatively simple, unzoned, very coarse-grained garnet–muscovite–biotite granites to complexly zoned tourmaline–garnet–muscovite–biotite granites (Smithies et al., 2002).

Mineralization

Archean mineralization in the northwest Pilbara Craton was primarily governed by its crustal evolution through cycles of extension, compression, and terrane collision. However, the earliest mineralization (Ni–Cu in the Roebourne Group) is preserved in the Karratha Terrane (KT) and at c. 3280 Ma pre-dates the East Pilbara Terrane Rifting Event when this terrane separated from the East Pilbara Terrane (EPT). From c. 3200 Ma onwards, the first Mesoarchean mineralization was sandstone- and conglomerate-hosted Au and heavy mineral deposition on a passive margin (Nickol River Formation). This was followed by plate convergence and c. 3120 Ma subduction-related VHMS Cu–Zn(–Pb–Ag) mineralization within an oceanic arc (Whundo Group). Collision of the KT, Regal Terrane (RT) and Sholl Terrane (ST) to form the West Pilbara Superterrane (WPS) at 3070–3060 Ma (Prinsep Orogeny) produced orogenic Au and Cu mineralization (Regal Thrust between the Roebourne Group and Regal Formation). Following the Prinsep Orogeny, crustal relaxation and extension established the 3050–3015 Ma Gorge Creek Basin with extensive BIF deposition that later provided the protore for magnetite and supergene-enriched hematite–goethite Fe deposits. Rifting between 3015 and 2970 Ma established the Whim Creek Basin (volcanic) and the first stage of the Mallina Basin (sedimentary) with local conglomerate-hosted Au mineralization. From c. 2970 Ma, renewed convergence led to c. 2955 Ma subduction-related VHMS CU–Pb–Zn mineralization (Cistern Formation and Rushall Slate of the Bookingarra Group) in the northwest part of the Mallina Basin. At the same time, Ni–Cu, PGE, and Au mineralization was associated with sanukitoid intrusions (Indee Suite) in the central part of the Mallina Basin. The final collision event of the northwest Pilbara Craton between the WPS and the EPT (North Pilbara Orogeny) took place between 2960 and 2920 Ma when orogenic Au deposits were formed in major shear zones of the Mallina Basin, and Au–Cu deposits were formed in faults within the WPS. Late in the North Pilbara Orogeny Ni–Cu and PGE mineralization was formed as part of 2930–2925 Ma layered ultramafic–mafic intrusions (Radley Suite) in the ST. Following the North Pilbara Orogeny, crustal relaxation and extension resulted in pegmatite-hosted Sn–Ta, Nb, and Be mineralization on the margins of monzogranite and leucogranite intrusions of the 2851–2831 Ma Split Rock Supersuite. Crustal extension following the orogeny also resulted in epithermal Au mineralization in faults and shear zones and local epithermal polymetallic mineralization, including Pb–Ag.

In addition to consideration of the mineralization in relation to stages of crustal evolution, it is evident that styles of mineralization are different between the Central Pilbara Tectonic Zone (CPTZ) and the area north of the SSZ. There are major differences in tectonic setting between units formed on the thick and relatively stable Paleoarchean crust north of the shear zone compared to those developed in the far more tectonically active settings within the CPTZ. The mineralization within the CPTZ is all associated with specific plate tectonic environments. For example, VHMS Cu–Pb–Zn deposits with felsic volcanic centres of arcs or rift basins and lode Au deposits controlled by convergence-related strike-slip shear zones. This Report describes the northwest Pilbara mineralization both by location, north or south of the SSZ, and in relation to the stages of crustal evolution.

Mineralization north of the Sholl Shear Zone

The section of the northwest Pilbara Craton located north of the Sholl Shear Zone (SSZ) includes the 3280–3260 Ma KT, and this terrane is interpreted to have been built on older crust similar to that of the 30 km-thick Paleoarchean EPT. The c. 3220 Ma Nickol River Basin, the c. 3200 Ma RT, and the 3066–3015 Ma Gorge Creek Basin all overlie the KT. Remnants of the KT probably underlie the Dampier and Harding Granitic Complexes, but there are insufficient geochronological data to confirm this on a regional scale. It is significant that almost all known Archean mineral deposits north of the SSZ are located within the 2000 km^2 area known to be underlain by Paleoarchean crust. The remaining area, occupied by voluminous Mesoarchean tonalite, granodiorite, and monzogranite, appears to be largely unmineralized.

Although the area north of the SSZ is underlain by Paleoarchean crust it was affected by some of the deformation in the CPTZ, particularly during the 2955–2920 Ma North Pilbara Orogeny. Accordingly, the northern area also contains structurally controlled mineralization introduced at that time.

Karratha Terrane

The Roebourne greenstone belt of the KT contains c. 3280 Ma komatiite-hosted Ni–Cu mineralization in the Ruth Well Formation and numerous shear-hosted gold and copper deposits in the Roebourne Group and the Nickol River Formation. The granitic rocks of the terrane appear to be unmineralized.

Komatiite-hosted Ni–Cu

The Ruth Well Ni–Cu deposits are hosted by serpentinized spinifex-textured peridotite flows in the c. 3280 Ma Ruth Well Formation. Detailed reports were provided by Tomich (1974) and Marston (1984). The mineralization consists of violaritized pentlandite, pyrrhotite, gersdorffite, niccolite, chalcopyrite, and magnetite, which is a similar mineral association to komatiite-hosted nickel deposits in the Kambalda area of the Yilgarn Craton. One diamond drillhole intersected 8.38 m of mineralization averaging about 3.52 wt% Ni and 0.78 wt% Cu (Marston, 1984), but exploration indicated that the deposits are relatively small, with resources quoted at approximately 70 000 t at a grade of 3 wt% Ni (Marston, 1984).

Metamorphism of the Ni–Cu mineralization is lower amphibolite to greenschist facies, similar to metamorphism in the remainder of the Ruth Well Formation. The metamorphism took place either during the first metamorphic event recognized in the northwest Pilbara at c. 3160 Ma, or during the Prinsep Orogeny at c. 3070 Ma. However, the deposits were subjected to several other metamorphic events since their formation. Metamorphism is likely to have accompanied the c. 3015 Ma D_4 event that strongly deformed the Cleaverville Formation prior to deposition of the Whim Creek Group (see Tectonic events, D_3 and D_4). This event was contemporaneous with intrusion of the Orpheus Supersuite (3023–3012 Ma), and several small granodiorite intrusions of this age were emplaced next to the deposits. Beintema (2003) dated zircon in leucocratic veins within the SSZ at Balla Balla, using laser ablation ICP-MS and obtained a U/Pb concordia age of 3014 ± 16 Ma and a $^{207}Pb/^{206}Pb$ age of 3019 ± 12 Ma. She interpreted the results as an indication of the age of movement on the SSZ. The Ruth Well Ni–Cu mineralization is adjacent to 2950–2920 Ma strike-slip faults related to the SSZ. Zircon geochronology on rock units close to the SSZ indicates a metamorphic disturbance event at 2930–2920 Ma (Krapez and Eisenlohr, 1998; Beintema, 2003). The metamorphic history of the deposits will make direct isotopic dating of the mineralization difficult.

Nickol River Basin and Regal Thrust

The Regal Thrust extensively deformed basalt and komatiite at the base of the Regal Formation, the sedimentary Nickol River Formation, and basalt in the upper part of the Roebourne Group. Although the zone of shearing and faulting is only 1–2 km wide, most of the known gold and copper mineralization in the Roebourne greenstone belt is situated within this zone (Hickman and Strong, 2003). Accordingly, the distribution of known gold deposits within the Roebourne greenstone belt shows a strong correlation with the Regal Thrust, e.g. at Lower Nickol, Weerianna gold mine, Carlow Castle, and Sing Well (Hickman, 2002). This zone of gold mineralization was folded by the c. 2940 Ma Roebourne Synform and Prinsep Dome during the North Pilbara Orogeny. Gold mineralization is interpreted to have taken place between 3160 Ma and 3070 Ma, most likely during the Prinsep Orogeny at c. 3070 Ma. A potential source of the gold was probably conglomerate and sandstone of the Nickol River Formation. The Nickol River Basin is likely to have received large quantities of detrital gold during c. 3220 Ma erosion of the EPT. The EPT contains numerous gold deposits older than 3220 Ma (Huston et al., 2002) and erosion of this crust during rifting and plate separation must have shed detrital gold into the passive margin basins. Dark heavy mineral bands are locally present in the Nickol River Formation, and Cenozoic alluvial and colluvial deposits derived from weathering of the formation contain detrital PGE, including Os (Powell and Horwitz, 1994) and Au at Lower Nickol. The PGE were probably derived from erosion of peridotite in the Ruth Well Formation. There are no reported studies on background gold contents of quartzite and conglomerate in the Nickol River Formation but these units could potentially host placer deposits. Thrusting and metamorphism of the Nickol River Formation is likely to have mobilized any placer gold and concentrated it in shear zones and brittle faults.

Regal Terrane

The thick basaltic succession of the c. 3200 Ma Regal Formation appears to be unmineralized. Komatiite is locally preserved immediately above the Regal Thrust but no nickel–copper mineralization comparable to that in the Ruth Well Formation has been reported. This is probably explained by the different origins of the Ruth Well Formation (lower part of a plume-related volcanic cycle on continental crust) and the Regal Formation (oceanic-like basaltic crust in a rift basin).

Late mineralization north of the Sholl Shear Zone

Several gold deposits of the KT belt are controlled by east-northeast striking faults (Hickman, 2001b), and these structures were probably formed as strike-slip faults during the 2950–2920 Ma North Pilbara Orogeny. Copper mineralization is partly situated on the same east-northeast-striking fault system, as in the old Fortune and Goods Luck Cu–Au workings southwest of Roebourne (Marston, 1979). In this area, the mineralized fault system cuts the c. 3015 Ma Andover Intrusion.

Mineralization in the Gorge Creek Basin

The Gorge Creek Basin formed between 3050 and 3015 Ma when the CPTZ was inactive except for regional crustal extension following the 3070–3060 Ma Prinsep Orogeny. Following the EPT Rifting Event, the Gorge Creek Group was the first lithostratigraphic unit to be deposited on both sides of the SSZ, and its depositional basin also extended across the TTSZ at least 200 km onto the eroded EPT. The 3025–3010 Ma Orpheus Supersuite was partly contemporaneous with the Gorge Creek Group, but its intrusion was restricted to a c. 50 km-wide belt above the SSZ. This provides evidence of minor tectonic activity taking place in close proximity to the shear zone. Mineralization within the Gorge Greek Basin is limited to iron ore in the Cleaverville Formation, whereas the Andover Intrusion of the Orpheus Supersuite contains small deposits of titaniferous magnetite and chrysotile asbestos. Although this c. 3015 Ma intrusion is similar in size to c. 2930 Ma layered ultramafic–mafic intrusions of the Radley Suite south of the SSZ, for example the Munni Munni Intrusion, exploration of the Andover Intrusion has failed to identify similar Ni–Cu or PGE mineralization. Minor felsic intrusions of the Orpheus Supersuite may be one of the sources of copper mineralization in c. 2940 Ma east- and northeast-striking faults and quartz veins on both sides of the SSZ although no mineralization has been recorded within the felsic rocks of the supersuite.

Iron ore

In terms of past and current production, and demonstrated resources, iron ore is by far the most important mineral commodity in the northern Pilbara Craton. Mining of iron ore from deposits within the Cleaverville Formation commenced in the east Pilbara in the mid-1960s and has continued until the present day. However, mining of iron ore in the northwest Pilbara has so far been restricted to the Pardoo project in the Ord Range where supergene enrichment of BIF locally formed ore bodies composed of hematite, goethite, and martite. These deposits average 56–57 wt% Fe, although local hematite-rich zones can be up to 63 wt% Fe.

It is notable that almost all of the mined iron ore deposits in the northern Pilbara Craton are supergene enrichments of BIF immediately underlying the 'Hamersley Surface'. The Hamersley Surface is a Mesozoic–Cenozoic peneplain that was deeply dissected from the Neogene onwards, leaving isolated remnants on the tops of the highest hills and ridges in the region. This probably explains why BIF outcrops of the Cleaverville Formation in the Roebourne and Devil Creek greenstone belts of the northwest Pilbara, where the Hamersley Surface has been removed by erosion, show little evidence of supergene enrichment. The Cleaverville Formation of these areas locally contains large tonnages (> 500 Mt) of magnetite-rich BIF, but average grades do not exceed 34 wt% Fe. Earliest drilling was undertaken on large magnetic anomalies in the Roebourne Synform near Wickham, which resulted in definition of the Cape Lambert and Mt Anketell magnetite deposits. Similar deposits have since been reported at Mount Oscar, 17 km south of Roebourne, and at several localities between Mount Regal and Devil Creek.

Mineralization in the Central Pilbara Tectonic Zone

The CPTZ is the most mineralized section of the Pilbara Craton. Deposits within this zone include VHMS Cu–Zn and CU–Pb–Zn deposits (e.g. Whundo, Whim Creek, Salt Creek, and Turner River), epigenetic base metal deposits (e.g. Mons Cupri and Comstock), layered mafic-intrusion-hosted vanadium–titanium deposits (e.g. Balla Balla), layered mafic–ultramafic intrusion-hosted nickel–copper–PGE deposits (e.g. Radio Hill, Munni Munni, Mount Sholl, and Three Kings), and numerous orogenic gold and gold–antimony deposits related to hydrothermal alteration in shear zones (e.g. deposits along the Mallina Shear Zone and the TTSZ).

Apart from orogenic lode Au deposits in shear zones and Ni–Cu and PGE mineralization in late ultramafic–mafic intrusions of the Radley Suite, most mineral deposits in the CPTZ were formed between 2955 and 2940 Ma, and are hosted by stratigraphic units of either the Croydon Group (central Mallina Basin) or the Sisters Supersuite. The timing of this mineralization coincides with initial stages of the North Pilbara Orogeny when plate convergence controlled the crustal evolution.

CPTZ mineralization before 2955 Ma

Sholl Terrane VHMS

The ST contains VHMS copper–zinc deposits at Whundo, West Whundo, Yannery Hill, and Ayshia. The closely spaced Whundo and West Whundo deposits are hosted by quartz-chlorite schist and quartz–sericite–chlorite schist within metabasalt close to the stratigraphic contact between the Nallana and Tozer Formations of the Whundo Group. Copper was mined intermittently at Whundo from 1911, and prior to renewed mining in 2006 total historical production stood at 12 000 t of supergene ore grading 22.3 wt% Cu (Collins and Marshall, 1999a). Mining at West Whundo in 2006 produced 5611 t of Cu-in-concentrate (Fox Resources Limited, 2006) and remaining resources in January 2007 were reported to be 0.627 Mt at 1.6 wt% Cu and 1.7 wt% Zn (Fox Resources Limited, 2007a).

Roberts (1974) described two distinct types of primary sulfide mineralization at Whundo: fine- to medium-grained layered pyrite, sphalerite, and chalcopyrite, and massive medium- to coarse-grained pyrite and pyrrhotite with minor sphalerite and chalcopyrite. At West Whundo, layered pyrite–sphalerite–chalcopyrite with disseminated magnetite is overlain by massive pyrrhotite and pyrite, essentially devoid of sphalerite and chalcopyrite. Supergene mineralization is mainly massive chalcocite and goethite–limonite with minor malachite, cuprite, and native copper.

The mineralization at Whundo and West Whundo is primarily VHMS, but the thickness of the main deposits and their location as ore shoots was subsequently influenced by north-trending folding and north-northwest-striking faults (Collins and Marshall, 1999a). The tectonic foliation of the schist, which hosts the deposits was probably produced by earlier shearing associated with D_2 thrusting. This would have taken place preferentially along the major lithological contact between the Nallana and Tozer Formations at Whundo. Evidence that the shearing at Whundo and West Whundo was restricted to this narrow stratigraphic level is provided by the observation that overlying rocks of the Tozer Formation contain extremely well-preserved sedimentary structures. Hornblende hornfels contact metamorphism of the schist at Whundo happened during intrusion of the c. 3013 Ma South Whundo Monzogranite.

The Cu–Zn mineralization at Yannery Hill and Ayshia, 1–2 km northeast of Whundo, is hosted by variably pyritic chloritic schist within metamorphosed volcanogenic sedimentary rock in the lower part of the Tozer Formation. Other local lithologies include quartz–sericite schist and chlorite–quartz–sericite schist with local andalusite porphyroblasts. Primary copper–zinc sulfide mineralization is concentrated in one stratabound unit, 1–3 m thick, that has been extensively folded within a northwest-plunging syncline. Thickening of the unit in fold cores controlled the geometry of most stopes. Ore minerals are secondary and include malachite, chalcocite, and cuprite within massive limonite (Marston, 1979).

Approximately 3045 t of high-grade oxidized and supergene ore averaging 15.9 wt% Cu were produced prior to 1999. The Ayshia Cu–Zn mineralization was discovered 700 m northeast of Yannery Hill in 2006. Published resources (measured, indicated, and inferred) have been reported as 0.767 Mt at 2.41wt% Zn and 0.43 wt% Cu (Fox Resources Limited, 2012).

Concealed VHMS Cu–Zn mineralization at the same stratigraphic level as at Whundo and Yannery Hill may underlie a substantial area southwest and northeast of the Orpheus prospect, as far as Bradley Well south of the SST. The c. 2940 Ma fault system around Orpheus is mineralized by Cu, Zn, Ag, and Au (Ruddock, 1999) and GSWA mapping near Bradley Well revealed gossans and quartz veins containing up to 14.5 wt% Zn, 2.2 wt% Pb, 840 g/t Ag, and 1.1 g/t Au (Hickman, 2001b). The total east-northeast–west-southwest strike length of mineral occurrences is 7 km. Faulting related to the nearby SSZ and hydrothermal activity during the North Pilbara Orogeny may have combined to mobilize sections of the c. 3120 Ma VHMS mineralization.

Prinsep Orogeny

Collision and terrane accretion between 3070 and 3060 Ma resulted in orogenic deformation, metamorphism, and granitic intrusion. North of the SSZ, mineralization associated with this event probably includes gold and copper mineralization along the Regal Thrust (described above). Additionally, thrusting of the Whundo Group onto the Railway Supersuite of the Cherratta Granitic Complex is interpreted to have structurally modified earlier VHMS Cu–Zn mineralization in the Whundo Group. Various small fault-controlled Au and Cu deposits within the Whundo Group may have been derived from metamorphism of thrust-related 3070 Ma mineralization. Otherwise, the apparent lack of c. 3070 Ma orogenic gold deposits in the CPTZ is explained by the fact that only the Whundo Group could contain them and, for the most part, the exposed Whundo Group consists of relatively undeformed basaltic rocks.

Whim Creek Basin

Following stratigraphic revision of the Whim Creek Group (Pike et al., 2002), VHMS mineralization previously interpreted to be within the group is now assigned to the overlying Bookingarra Group. The Warambie Basalt is apparently unmineralized and the felsic succession of the Red Hill Volcanics is predominantly composed of intrusive dacite that contains no known mineralization.

Mallina Basin

This Report explains that the Mallina Basin evolved in three northeast-trending zones: a central, relatively deep-water rift basin flanked by shallow-water margins to the northwest and southeast. The southeast shelf facies of the basin evolved on stable, thick continental crust of the EPT and until c. 2940 Ma was essentially unmineralized except for conglomerate-hosted gold in the Lalla Rookh Sandstone (Carpentaria Exp Co Pty Ltd, 1983; Barnes, 1984; Krapez and Groves, 1984; Taylor, 1985; Hickman and Harrison, 1986). The depositional age of the Lalla Rookh Sandstone, and therefore also that of the placer gold, is constrained between 3015 and 2940 Ma (Hickman, in prep.). The peak phase of the North Pilbara Orogeny took place at c. 2940 Ma, at which time collision and accretion of the EPT and WPS was complete, and mineralization across the northern Pilbara Craton became dominated by orogenic gold in faults and shear zones.

In the central part of the Mallina Basin, the De Grey Group is interpreted to be in two parts, separated by a c. 2955 Ma unconformity. There is evidence that the lower part of the group is partly laterally equivalent to the Lalla Rookh Sandstone, implying that paleoplacer gold mineralization might also be present in this part of the Mallina Basin. However, the only known gold mineralization that might be syndepositional is located in the Constantine Sandstone west of the Sherlock River. At Croydon Top Camp gold has been mined from auriferous carbonate-rich wacke and ferruginous chert (Smithies, 1998b), although this mineralization might be hydrothermal.

CPTZ mineralization 2955–2940 Ma

The 2955–2940 Ma crustal evolution of the upper Mallina Basin succession in the CPTZ resulted in a greater diversity of mineralization than anywhere else in the northern Pilbara Craton. Coinciding with intrusion of the Sisters Supersuite and mafic volcanism of the Bookingarra Group, 2955–2940 Ma mineralization included base metal deposits of either VHMS or epigenetic origin (Mons Cupri and Comstock), VHMS Cu–Zn and CU–Pb–Zn deposits (Whim Creek, Salt Creek, and Balla Balla), V–Ti–Fe deposits in layered mafic intrusions (Balla Balla), and Ni–Cu, PGE, and Au mineralization related to high-Mg diorites (sanukitoids) as at Three Kings. The tectonic setting was extensional, with the central part of the Mallina Basin being a back-arc basin, whereas the northwest part of the basin, now within the Whim Creek greenstone belt, was on the southeast margin of a continental arc. As noted earlier in this Report (see Basin relationships), at 2955 Ma these two sections of the Mallina Basin were in different relative positions than today although the precise amount of c. 2940 Ma strike-slip displacement on the LF is unknown. The degree of shearing within the fault, and the major stratigraphic changes across it, suggest many kilometres of relative displacement.

Whim Creek greenstone belt

Mons Cupri VHMS

Two levels of mineralization are present in conglomerate of the Cistern Formation at Mons Cupri: a lower, funnel-shaped, Cu-rich disseminated stockwork overlain by a stratiform, 5–10 m-thick Pb–Zn–Ag zone (Low, 1963; Blockley, 1971; Miller and Gair, 1975; Marston, 1979; Smith, 1980; Collins and Marshall, 1999b; Ferguson, 1999; Hickman et al., 2000; Huston et al., 2000; Pike et al., 2002; Huston, 2006). The copper-rich stockwork contains chalcopyrite with minor sphalerite and galena, and is associated with intense chloritic alteration, whereas the stratiform Pb–Zn zone contains pyrite, sphalerite, and galena with minor chalcopyrite and tetrahedrite and has carbonate alteration. Venturex Resources Limited (2015) estimated total resources at Mons Cupri to be 4.607 Mt at 0.9 wt% Cu, 1.3 wt% Zn, 0.5 wt% Pb, and 24.1 g/t Ag.

The conglomerate host of the Mons Cupri mineralization is polymictic and poorly sorted with subangular to locally rounded clasts and blocks up to 10 m across. The clasts are mainly composed of rhyolite, granite, and lesser basalt. The matrix of the conglomerate contains rhyolite shards and has been intruded by felsic sills, dykes, and domal bodies (Huston et al., 2000), providing evidence of coeval volcanism during conglomerate deposition. The Cistern Formation was intruded by the Opaline Well Intrusion which is interpreted to be an intrusive component of the Louden Volcanics (Smithies, 1998a; Pike et al., 2006) and the conformably overlying Rushall Slate contains the 100 m-thick Comstock Member, interpreted by Pike et al. (2006) to be intrusive.

Although the Pb–Zn zone is stratiform, Huston et al. (2000) used textural evidence to conclude that the mineralization is epigenetic. Pb model ages of 2933–2920 Ma from galena in the Mons Cupri mineralization (Huston et al., 2000, 2002a) are approximately 20–30 million years younger than the depositional age of the Cistern Formation, providing additional evidence that the mineralization might be epigenetic (Huston et al., 2002a; Pike et al., 2002; Huston, 2006). However, the epigenetic interpretation is controversial, partly because there are no known 2933–2920 Ma potential source rocks, such as granites or felsic volcanic rocks in the Mallina Basin. Additionally, Pb model ages on apparently related deposits such as Whim Creek and Salt Creek are more in line with the interpreted depositional age of the Cistern Formation. At Whim Creek stratabound Cu–Zn deposit in the Rushall Slate, stratigraphically overlying or laterally equivalent to the Cistern Formation, has provided Pb model ages of 2948–2942 Ma (Huston et al., 2000, 2002a) and stratiform Pb–Zn mineralization in the Cistern Formation at Salt Creek has given Pb model ages of 2960–2950 Ma (Richards et al., 1981). The same data were recalculated by Huston et al. (2000) to either 2962–2959 Ma or 2945–2942 Ma, depending on the model used. An alternative explanation of the textures observed in the Mons Cupri mineralization is that they were formed by metamorphism of the deposits, either by heat and hydrothermal fluids derived from c. 2950 Ma ultramafic, mafic, and granitic intrusions or during peak metamorphism (2940–2930 Ma) of the North Pilbara Orogeny. It is probable that quartz veins carrying Pb–Ag mineralization in high-Mg basalt or dolerite of the nearby epigenetic Comstock deposit were derived by metamorphism of the Mons Cupri VHMS mineralization. Galena from the Comstock quartz veins, which immediately overlie the stratiform Pb–Zn–Ag mineralization at Mons Cupri, has provided Pb model ages of 2947–2921 Ma (Huston et al., 2000).

Whim Creek VHMS

The Whim Creek Cu–Zn VHMS deposit is stratabound within a particular stratigraphic horizon of the Rushall Slate. Primary minerals are pyrite, pyrrhotite, chalcopyrite, and sphalerite with minor galena, magnetite, and arsenopyrite, whereas supergene alteration minerals are chalcocite, covellite, and malachite, with minor azurite and chrysocolla (Collins and Marshall, 1999a). Collins et al. (2004b) reported that the mineralization extends at least 5 km along strike, except in the Whim Creek mine area where the sulfide zone is less than 0.5 m thick. In the mining area, the deposit is 10–15 m thick over a strike length of 600 m. Pb isotopic data were interpreted by Huston et al. (2002) to provide a model age of c. 2948 Ma.

The Rushall Slate is a more distal sedimentary facies than the Cistern Formation, and the two formations are probably at least partly laterally equivalent. Between 1891 and 2009 total production from the Whim Creek mine was approximately 46 500 t Cu, 35 000 t of which was produced by Straits Resources between 2005 and 2009 (Hickman, 1983; Venturex Resources Limited, 2009). Venturex Resources Limited (2015) estimated total resources at Whim Creek to be 0.97 Mt at 2.1 wt% Cu, 1.1 wt% Zn, 0.2 wt% Pb, and 13.9 g/t Ag.

Salt Creek and Balla Balla VHMS

The Pb–Zn (–Cu) deposits at Salt Creek and Balla Balla are situated 20 km northwest of Whim Creek, a short distance south of the SSZ. The local succession was described by Pike et al. (2002) and Pike et al. (2006). Massive sulfide lenses at Salt Creek contain sphalerite, pyrite, galena, and chalcopyrite within sandstone and shale overlying volcanic and volcaniclastic rocks (Huston et al., 2000). This is closely analogous to the stratigraphy of the Cistern Formation and Rushall Slate in the Mons Cupri – Whim Creek area, and Pb model ages (see Mons Cupri) also support a correlation.

The regional association of VHMS mineralization with the Cistern Formation – Rushall Slate succession has been linked to the location of volcanic centres with intrusions of porphyritic dacite and rhyolite (Sylvester and de Laeter, 1987; Collins and Marshall, 1999a). Salt Creek and Whim Creek are located on opposite limbs of the Whim Creek Anticline, establishing that the succession of the Whim Creek greenstone belt is underlain by the Caines Well Granitic Complex. Due to poor exposure, the detailed geology of this complex is uncertain but it is likely that the uppermost intrusions of c. 2950 Ma granites within it were the source of the VHMS mineralization in the Cistern Formation and Rushall Slate. Venturex Resources Limited (2015) reported indicated zinc resources at Salt Creek to be 0.475 Mt at 14.1 wt% Zn, 4.4 wt% Pb, and 107.1 g/t Ag. Indicated copper resources were separately reported as 0.423 Mt at 3.7 wt% Cu and 0.9 wt% Zn.

Balla Balla V–Ti–Fe

Approximately 10 km northwest of Whim Creek, the c. 2950 Ma Sherlock Intrusion includes a 20–30 m-thick layer of massive titanomagnetite with elevated vanadium. The ore zone extends over a strike length of almost 20 km from Balla Balla East to Don Well. Drilling has identified mineral reserves over the eastern 8 km of this strike length. The mineralization is tabular and positioned between layers of anorthositic gabbro and leucogabbro. Other layers of the intrusion include pyroxenite, norite, anorthosite, and granophyre. At Balla Balla, the Sherlock Intrusion intrudes the axial region of the Whim Creek Anticline between the Caines Well Granitic Complex and the Whim Creek Group. Layering is inclined 25° to the northeast in this area, confirming the sill was folded by the c. 2940 Ma D_8 anticline. At Don Well, southwest of Salt Creek, the intrusion lies on the northern limb of the anticline. Proposed products from mining are magnetite concentrate (containing V_2O_5) and ilmenite. A relatively minor occurrence of V–Ti mineralization in the Sherlock Intrusion has also been identified at Mount Fisher, 35 km west-southwest of Balla Balla.

Sherlock Bay Ni–Cu

The Sherlock Bay Ni–Cu deposits are in tectonic lenses within quartz–amphibole–magnetite–sulfide schist and metavolcanic and metasedimentary rocks on the southern margin of the SSZ. Deformed units of titaniferous magnetite outcrop in the same area (Miller and Smith, 1975), and it is evident that both the titaniferous magnetite and Ni–Cu mineralization were derived from extreme deformation of the Sherlock Intrusion. This intrusion rims the Caines Well Granitic Complex in all areas except at Sherlock Bay, where it is almost completely sheared out along the SSZ. The metamorphosed felsic volcanic and sedimentary units are almost certainly tectonic slices of the Red Hill Volcanics and Cistern Formation – Rushall Slate succession. The Ni–Cu mineralization has been drilled to a depth of 1000 m, and inferred resources are 16 Mt at 0.75 wt% Ni and 0.9 wt% Cu (Ruddock, 1999).

Central Mallina Basin

The central part of the Mallina Basin contains 2955–2940 Ma mineralization related to intrusion of the Indee and Langenbeck suites of the Sisters Supersuite.

Indee Suite PGE

The association of high-Mg diorites (sanukitoids) and ultramafic–mafic intrusive rocks in the Mallina Basin has the potential to host PGE mineralization (Hickman, 2004c). This style of mineralization is mined at Lac des Iles in the Wabigoon Subprovince of Ontario. The 2955–2945 Ma high-Mg diorites of the Mallina Basin were intruded along an east-northeast-trending belt that probably followed a zone of deep faulting in the crust beneath the Croydon Group. This fault zone was reactivated at c. 2940 Ma forming the presently exposed Mount Wohler, Tabba Tabba, and Mallina Shear Zones. Immediately north of the TTSZ, between the Turner and Yule Rivers, PGE mineralization was discovered along a peridotite–pyroxenite contact next to a large high-Mg diorite intrusion near Mount Dove. The best intersection reported from the prospect, known as Three Kings, was 6 m at 2.09 g/t PGE, which included 2 m at 3.18g/t PGE and 1 m at 4.56 g/t PGE (De Grey Mining Limited, 2005). Drilling 1 km to the east provided an intersection of 14 m at 1.32 PGE, whereas 3 km to the west an intersection of 6 m at 0.85 g/t was recorded (De Grey Mining, 2005). The Three Kings area is poorly exposed and drilling was mainly focused on areas of outcrop.

Langenbeck Suite Ni–Cu

The 2955–2940 Ma Langenbeck Suite of layered ultramafic–mafic sills is widespread in the central Mallina Basin, and probably extends northwest as the Sherlock Intrusion (see above). The Sherlock Intrusion intruded the contact between the Caines Well Granitic Complex and the Whim Creek Group over at least 1000 km^2. As noted above, the Sherlock Intrusion contains both V–Ti–Fe and low-grade Ni–Cu mineralization, but current evidence is that the Langenbeck Suite in the central Mallina Basin is essentially unmineralized. However, exceptions may happen where ultramafic–mafic intrusions of the suite are deformed by major shear zones. Approximately 120 km east-northeast of Port Hedland on the Pardoo Fault (northeast extension of the TTSZ) the Highway Ni–Cu deposit is the most likely example. The Highway Ni–Cu mineralization is contained within sheared ultramafic–mafic rocks of one or more intrusions. Exploration drilling by CRA Exploration Pty Ltd in 1991 resulted in intersections of c. 90 m at 0.35 wt% Ni and 0.14 wt% Cu and delineated a mineralized zone averaging 50–75 m thick, dipping 70° north, and extending along an east–west strike length of 700 m (CRA Exploration Pty Ltd, 1991). The most recently published inferred resources at Highway are 50 Mt at 0.3 wt% Ni and 0.13 wt% Cu (Segue Resources Limited, 2013). It is probable that the ultramafic–mafic intrusive rocks at Highway belong to the Langenbeck Suite. Farther southwest the Millindinna Intrusion is deformed along the TTSZ for a distance of more than 100 km.

CPTZ mineralization 2940–2920 Ma

Peak deformation and metamorphism of the North Pilbara Orogeny took place between 2940 and 2920 Ma and produced a large number of mineral deposits in the CPTZ. Principal among these are orogenic lode Au deposits along major shear zones and orthomagmatic Ni–Cu and PGE deposits in layered ultramafic–mafic intrusions of the Radley Suite. The TTSZ contains a wide range of mineral deposits (Au, Pb–Zn, Ni–Cu, PGE, magnetite iron ore, Sn, Ta, and Be) but only Au was evidently introduced between 2940 and 2920 Ma. Zn–Pb mineralization in the TTSZ probably originated as VHMS mineralization in the Sulphur Springs Group because the present structure and size of the deposits is controlled by c. 2940 Ma deformation. Elsewhere in the CPTZ, other 2940–2920 Ma mineralization included epigenetic Au, Cu, Cu–Zn, and Pb–Ag deposits in quartz veins and minor faults.

Orogenic lode Au and Au–Sb deposits

Detailed descriptions of lode Au deposits in the Mallina Basin known prior to 2002 were provided by Huston et al. (2002b). The two main deposits recognized at that time were Withnell and Camel, both on the Mallina Shear Zone close to its junction with the Mount Wohler Shear Zone. The two shear zones are separated by the c. 2948 Ma Peawah Granodiorite which, as a pre-existing rigid body, probably caused deflection of the two shear zones at c. 2940 Ma. An analogous structural setting exists for another group of lode Au deposits in the Mallina Basin east of the Turner River, where the Tabba Tabba and Mallina shear zones were deflected around a large c. 2950 Ma high-Mg diorite intrusion at Mount Dove. The largest gold deposit in this area is Wingina Well on the TTSZ and others include Amanda and Mount Berghaus (De Grey Mining Limited, 2012).

The Withnell and Camel mineralization is in quartz and quartz–carbonate vein systems on the northern margin of the Mallina Shear Zone. Both deposits contain pyrite and gold grades increase with pyrite content. Huston et al. (2002b) noted a difference between the two deposits in that Withnell contains sericite alteration and Camel has pyrophyllite alteration. They commented that the presence of pyrophyllite at Camel indicates involvement of magmatic-hydrothermal fluids. Sericite zones at Withnell are relatively high grade, although Au grades generally are not high compared to many other lode Au deposits in Western Australia. Prior to a decision to mine in 2005, combined resources of Withnell and Camel were reported as 10.46 Mt at 1.6 g/t Au for 529 000 oz Au (Range River Gold Limited, 2005). Mining during 2006 and 2007 produced approximately 30 000 oz Au, after which the mines were placed on care and maintenance.

The Wingina Well deposit contains total measured, indicated, and inferred resources of 5.1 Mt at 1.30 g/t for 268 000 oz Au (Polymetals Mining Limited, 2013). The host rocks are mainly sheared BIF and chert of the Cleaverville Formation, although sheared metabasalt is also present. Northeast plunging shoots of Au mineralization suggest combined structural and lithological controls of gold from hydrothermal fluids within the shear zone. Deflection of the TTSZ and Mallina Shear Zone around the high-Mg diorite intrusion at Mount Dove probably formed a triangular ‘pressure-shadow’. Wingina Well is near the eastern apex of this triangular area but other deposits and prospects on the shear zones either side of it include Lost Ark, Wingina Well 2, Last Crusade, Edkins, Amanda, Amanda East (all on the TTSZ), Brierly, and Mount Berghaus (on the Mallina Shear Zone). Published inferred resources at Amanda are 0.687 Mt at 1.6 g/t for 35 000 oz Au and at Mount Berghaus are 0.92 Mt at 1.4 g/t for 43 000 oz Au (De Grey Mining Limited, 2012).

Another type of Au mineralization in the CPTZ is seen in a number of Au–Sb deposits. Mines with historic production of antimony are located at Balla Balla, Mallina Homestead, Peawah Hill, and Sherlock Homestead (Finucane and Telford, 1939). With the possible exception of Balla Balla (old workings not relocated), the antimony was mined from quartz–stibnite(–cervantite) veins in metamorphosed shale or argillaceous greywacke of the Croydon Group. The Au–Sb association is common in orogenic gold deposits and it is assumed that Sb was derived by metamorphic devolatilization of the host rocks (Hickman, 1983).

Tabba Tabba Shear Zone Zn–Pb

The TTSZ (Fig. 3) contains intermittent Zn–Pb mineralization, with minor Cu, Ag, and Au over a strike length of at least 20 km. Two deposits, Orchard Tank (previously Orchard Well) and Discovery, have reported combined inferred resources (1% Zn cut-off) of 2.6 Mt at 2.7 wt% Zn, 1.1 wt% Pb, 0.1 wt% Cu, 89 g/t Ag, and 0.7 g/t Au (De Grey Mining Limited, 2012). However, individual intersections in the drilling at both deposits have suggested the presence of higher grade zones, for example at Orchard Tank, 5.4 m at 11.59 wt% Zn, 6.63% Pb, 300.55 g/t Ag, and 2.3 g/t Au; and at Discovery, 10 m at 6.27% Zn, 2.31% Pb, 154.6 g/t Ag, and 0.99 g/t Au (De Grey Mining Limited, 2014). The lenses of mineralization at Orchard Tank and Discovery strike east-northeast and are inclined steeply north-northwest or south-southeast, parallel to lithological boundaries and main tectonic foliation (S_8) within the shear zone. The Orchard Tank lenses are described as stacked (De Grey Mining Limited, 2014), suggesting possible duplication in tectonic slices. At both deposits and in other prospects along the shear zone, the Zn–Pb mineralization is hosted by felsic schist (sheared and metamorphosed felsic volcanic rocks) dated at 3253 ± 4 Ma (GSWA 160258, Wingate et al., 2010). This age supports a stratigraphic correlation with the 3275–3235 Ma Sulphur Springs Group, which in the type area also contains VHMS mineralization. Investigations of that mineralization concluded that it was associated with crustal extension and rifting (Vearncombe et al., 1998).

Radley Suite PGE and Ni–Cu

Some of the most significant mineralization in the CPTZ is contained in the large layered ultramafic–mafic intrusions of the 2930–2924 Ma Radley Suite (Sisters Supersuite). These intrusions, all of which were emplaced in a north-northeast-trending zone across the ST, host large Ni–Cu and PGE deposits at Radio Hill, Munni Munni, and Mount Sholl. Detailed descriptions were provided by Hoatson et al. (1992) and Hoatson and Sun (2002), but three intrusions (Bullock Hide, Andover, and Sherlock) — considered by these authors to be contemporaneous with the Radley Suite — are now known to be older (see Radley Suite, Sisters Supersuite). Intrusions of the Radley Suite typically comprise a lower section of ultramafic layers (dunite, peridotite, and pyroxenite) overlain by a layered mafic section of gabbro, leucogabbro, norite, and more rarely, anorthosite and granophyre. Additionally, gabbroic units of the Radio Hill and North Whundo intrusions are overlain by granitic rocks. In the case of the North Whundo Intrusion the upper granitic component, the Yannery Granite (Hickman and Kojan, 2003) is a hornblende-monzogranite dated at 2930 ± 4 Ma (N3162, Smith, 2003). The granite related to the Radio Hill Intrusion was dated at 2929 ± 13 Ma (N4450, Smith, 2003).

Hunter Resources commenced a program of PGE exploration in the early 1980s. The exploration initially targeted all known ultramafic–mafic layered intrusions in the northwest Pilbara, but later focused on the ultramafic–mafic contact in the Munni Munni Intrusion. Hunter based its PGE exploration on the Bushveld and Stillwater models, and in 1984 discovered the main layer of PGE mineralization which became known as 'Hunter's reef'.

Radio Hill Intrusion Ni–Cu

The Radio Hill Intrusion consists of a basal ultramafic zone comprising lherzolite, dunite, and websterite and an overlying gabbroic zone, including quartz gabbro, gabbro, and gabbronorite. The basal layer of the ultramafic zone contains massive, disseminated, veined and brecciated pyrrhotite, and pentlandite–chalcopyrite–magnetite at or near the contact with the Nallana Formation. Underground mining has been intermittent since the late 1980s and has mainly produced Ni and Cu with Co, Pd, Pt, and Ag as byproducts. Between 1998 and 2002 the mine produced 21 000 t Ni and 16 000 t Cu (Fox Resources Limited, 2007b). A second period of mining between early 2007 and mid-2008 resulted in production of an additional 2605 t Ni and 4120 t Cu (Quarterly reports to ASX, Fox Resources Limited). Co production was 589 t, and Pd was 159 kg (MINEDEX). From 2008 the mine was placed on care and maintenance.

Mount Sholl Intrusion Ni–Cu

The Mount Sholl Intrusion contains Ni–Cu mineralization at a number of localities referred to as the A1, B1, and B2 zones. Abeysinghe and Flint (2008) referred to drilling data that indicate B2 as the most prospective zone, with indicated and inferred resources of 5.987 Mt at 0.5 wt% Ni, 0.6 wt% Cu. The mineralization consists of aggregates of pentlandite, pyrrhotite, and chalcopyrite within a thin gabbroic marginal layer along the northwestern margin of the intrusion. This gabbroic layer underlies a thicker ultramafic layer of peridotite and pyroxenite that forms the main basal section of the intrusion. Mathison and Marshall (1981) and Marston (1984) provided additional information on the Mount Sholl deposits. The known mineralization at Mount Sholl is relatively low grade, but future mining may be possible if processing is by heap leaching.

Munni Munni Intrusion PGE and Ni–Cu

The c. 2925 Ma Munni Munni Intrusion, which outcrops over an area of 150 km^2 and is concealed by the Fortescue Group over a similar area contains one of the most significant PGE deposits in Australia. The indicated and inferred resources are 23.6 Mt at 2.9 g/t PGE + Au (Platina Resources Limited, 2006). These resources were calculated for one 7.5 km-long section of the mineralized layer but its total known length is 22 km. Best intersections reported by Hoatson and Sun (2002) are 5–8 ppm combined Pt+Pd+Au over a vertical thickness of 0.5 m.

PGE mineralization in the Munni Munni Intrusion takes place near the top of a layer of porphyritic (large euhedral hypersthene grains) plagioclase websterite orthocumulate. The porphyritic websterite overlies a 1850 m-thickness of ultramafic rocks and underlies a >3630 m-thick upper zone of gabbro (Hoatson, 1986; Hoatson and Sun, 2002). The ultramafic zone contains cyclic layers of dunite, lherzolite, wehrlite, olivine websterite, clinopyroxenite, and websterite with orthopyroxenite and norite. The base of the gabbro zone is marked by the first upward appearance of cumulus plagioclase and inverted pidgeonite in the intrusion (Hoatson and Sun, 2002). Hoatson and Sun (2002) interpreted the porphyritic websterite to be a product of mixing between the ultramafic and gabbroic zones, and described an erosional contact with the underlying ultramafic zone. The total thickness of the mineralized orthocumulate layer is 30–80 m, and it extends 22 km around the western northern and eastern sides of the intrusion. The highest PGE grades are concentrated in a 7.5 km-long section of porphyritic websterite in the northern part of the intrusion. Mineralization consists of Pd and Pt accompanied by Au, Cu (0.3 wt%), and Ni (0.2 wt%). Disseminated sulfides are principally chalcopyrite, pentlandite, and pyrrhotite.

Other 2940–2920 Ma mineralization

The CPTZ contains scattered, in most cases small, deposits of Au, Cu, Cu–Zn, or Pb–Ag that are not associated with shear zones, but were most likely formed during or shortly after peak metamorphism at 2940–2930 Ma. The largest Au deposits in this class is at Station Peak and Toweranna, both in the central Mallina Basin. At Station Peak Au mineralization is in east–west-striking quartz veins in a large dolerite sill of the Langenbeck Suite. Model Pb ages of 1916 Ma and 1548 Ma (applying different models) obtained on galena from a quartz vein in this area (Huston et al., 2002a) are unlikely to indicate the age of the Au mineralization, and may reflect local Proterozoic hydrothermal activity on the local fault system. At present it seems most likely that Au mineralization was associated with 2940–2920 Ma faulting and hydrothermal fluids in the dolerite. The Egina Fault is situated only 3 km east of Station Peak, and there are related splay faults immediately to the east of the old mining area. Au mineralization at Toweranna is in auriferous quartz veins through a small intrusion of porphyritic high-Mg granodiorite in the axial region of the Croydon Anticline. This and other small felsic intrusions along the axis of the anticline are apophyses from the c. 2948 Ma Peawah Granodiorite. The age of the Toweranna mineralization may be similar to that of Cu mineralization at Evelyn and Quamby, also in the axial region of the Croydon Anticline. Huston et al. (2002a) interpreted this Cu mineralization to have a VHMS origin, but the host rock is tremolite–chlorite schist (possibly a metamorphosed intrusion of the Langenbeck Suite). Smithies (1998b) interpreted the Evelyn Cu deposits to be structurally controlled and either syn- or post-D_8 (D_3 in his local structural scheme). This interpretation is consistent with the model Pb ages from the deposits which are younger than 2945 Ma (Huston et al., 2002a). It is likely that the Toweranna and Evelyn–Quamby mineralization took place at 2940–2920 Ma and was introduced with hydrothermal fluids during metamorphism. Venturex Resources Limited (2015) estimated total resources at Evelyn (Liberty-Indee Project) to be 0.657 Mt at 1.8 wt% Cu, 3.7 wt% Zn, 0.3 wt% Pb, and 35.9 g/t Ag.

Another significant area of Cu mineralization is located immediately west of the Egina Fault at Egina. The Egina Fault is a major southern splay structure off the Mallina Shear Zone. Most movement probably took place at 2940–2920 Ma although, as in the case of the LF, the structure was reactivated during early deposition of the Fortescue Group. Marston (1979) reported that the near-surface Cu mineralization at Egina takes the form of malachite and atacamite in east-southeast dipping layers of massive limonite within carbonaceous phyllite of the Mallina Formation. Minor quartz veining is present in the mineralization and larger quartz veins are present nearby. Marston (1979) refers to adjacent units of 'iron-poor banded iron-formation'. These may be silicified sheared fine-grained pyritic sedimentary rocks. There are no reports of Au in the Egina deposits but 19th century prospectors found numerous Au nuggets and finer eluvial and alluvial Au in the Egina area (Smithies and Farrell, 2000).

Post-orogenic mineralization

Split Rock Supersuite, Sn, Ta, and pegmatite minerals

As described earlier in this Report, intrusions of the 2851–2831 Ma Split Rock Supersuite were emplaced along a wide, northwest-trending belt across the northern Pilbara Craton. Unlike many older granitic intrusions of the craton these post-orogenic intrusions are not confined to pre-existing granitic complexes and associated mineralization varies depending on the geological setting (Hickman, 1975b). The highly fractionated intrusions are typically fringed by pegmatite veins and many of these have sustained small-scale mining operations for various minerals including cassiterite, tantalite, columbite, scheelite, wolframite, beryl, lithium minerals, mica, and K-feldspar (Miles et al., 1945; Ellis, 1950; Blockley, 1980; Hickman, 1983; Sweetapple and Collins, 2002). At many localities, most production of Sn and Ta has come from Cenozoic alluvial channels adjacent to the pegmatites.

Principal mining areas for pegmatite minerals in the northwest Pilbara have been Tabba Tabba and Strelley (both on the TTSZ), and Pippingarra. Minor production has come from Mount Hall and Bore Creek. Sn and Ta mineralization at Pippingarra and Bore Creek are interpreted to be related to the Myanna Leucogranite. The deposits are described by Blockley (1980), Hickman (1983), Ferguson and Ruddock (2001), and Sweetapple and Collins (2002).

Epithermal vein deposits

The northwest Pilbara Craton contains a number of post-2900 Ma low-temperature, low-pressure vein deposits that occupy extensional fractures related either to post-orogenic relaxation or crustal extension and rifting during deposition of the 2775–2630 Ma Fortescue Group. In the absence of geochronology, relationships to the Fortescue Rifting Event are interpreted where the mineralized fractures penetrate or displace the Fortescue Group (Marshall, 2000). The only significant example of epithermal mineralization dated between 2900 and 2830 Ma is at Elizabeth Hill, close to the northeast margin of the Munni Munni Intrusion. Elizabeth Hill is a high grade Ag deposit that also contains Pb, Ni, Cu, and PGE. Marshall (2000) provided a detailed description in which he cited textures and other features supporting a low-temperature, low-pressure polymetallic deposit. Galena from Elizabeth Hill was dated by Huston et al. (2002a) at 2867–2861 Ma. The calcite–quartz vein system cuts the basal zone of the c. 2930 Ma Munni Munni Intrusion, which accounts for the Ni–Cu and PGE within the c. 2860 Ma mineralization.

Future investigations

Stratigraphic correlations

Additional investigations are required to test a number of potential stratigraphic correlations between the northwest and east Pilbara.

Ruth Well Formation and Sulphur Springs Group

The depositional age of the Ruth Well Formation is known to be >c. 3270 Ma because it is intruded by the 3270–3261 Ma Karratha Granodiorite. However, the maximum age of the formation is only very loosely constrained by Nd model ages between 3480 Ma and 3400 Ma (Table 4). Unpublished Nd isotopic data (S. Tessalina, written comm.) suggest a depositional age between 3330 Ma and 3280 Ma, and c. 3280 Ma is currently inferred to be the most probable age. However, 3280 Ma is older than the depositional age of any dated formations in the Sulphur Springs Group, whereas the older c. 3330 Ma age would closely match that of the 3350–3335 Ma Euro Basalt in the east Pilbara. Opportunities to better constrain the depositional age of the Ruth Well Formation may be available from detrital zircons in quartzite and secondary chert units.

Karratha Granodiorite

Additional geochronology on the Karratha Granodiorite is needed to establish if it includes older Paleoarchean granitic rocks and to better constrain the maximum age of its sources. Banded gneiss enclaves within the Maitland River Supersuite of the Dampier Granitic Complex also require dating to establish if such gneiss is related to the Karratha Granodiorite or older granites.

Stratigraphy of the Mallina Basin

The Croydon Group in the central Mallina Basin is divided by a c. 2955 Ma unconformity. This unconformity was produced by deformation in the early stages of the 2955–2919 Ma North Pilbara Orogeny and is therefore likely to be a regional unconformity. A lithostratigraphic group cannot include a regional unconformity. The name Croydon Group could be retained for the lower succession but a new group name should be introduced for the upper succession, along with more lithostratigraphic subdivision into formations and members. This will require more detailed mapping and structural analysis of the better exposed sections of the central Mallina Basin.

Whim Creek Group and Cattle Well Formation

A potential stratigraphic correlation exists between 3015–2990 Ma felsic volcanic rocks of the Whim Creek Group and the c. 2988 Ma Cattle Well Formation in the east Pilbara. The current spatial separation of the units might be largely due to c. 2940 Ma strike-slip faulting. The age range of the Whim Creek Group is poorly constrained. More geochronology is needed, not only in the Red Hill area, but also in other parts of the Whim Creek greenstone belt. As discussed in this Report, neither of the two published dates on the succession provided definitive results. Additionally, the intrusive age of the Mons Cupri Dacite Member (Red Hill Volcanics) is unknown, apart from being younger than c. 3009 Ma. This dacite, if it does represent a single intrusive event, is more volumetrically important than either the Red Hill Volcanics of the Whim Creek Group or the Cistern Formation of the overlying Bookingarra Group. The close spatial association between some units of this dacite and VHMS mineralization in the Cistern Formation (Sylvester and de Laeter, 1987; Collins and Marshall, 1999a) suggests an age close to c. 2955 Ma rather than c. 2990 Ma that might be interpreted from previous geochronology (Barley et al., 1994).

Myanna Leucogranite with the Split Rock Supersuite

Correlation of the Myanna Leucogranite, Tabba Tabba Leucogranite, and Thelman Monzogranite with the Split Rock Supersuite requires testing by geochronology. This is important not only for economic reasons (potential of these granites to host Sn, Ta, and pegmatite mineralization close to the existing infrastructure of Port Hedland), but also to provide evidence on the age of the crust that underlay the Mallina Basin at c. 2830 Ma, after the final terrane collision of the North Pilbara Orogeny.

Conclusions

Paleoarchean–Mesoarchean evolution of a continental margin

The geology of the northwest Pilbara Craton records evidence that plate tectonic processes operated from c. 3220 Ma. A major rifting event in the 3530–3223 Ma EPT caused northwest-southeast separation of the KT and the development of a rift basin composed of c. 3200 Ma juvenile basaltic crust CPB. From c. 3220 Ma to c. 3170 Ma passive margin successions formed along the margins of the EPT and KT as the two plates moved apart. The passive margin succession of the northwest Pilbara is represented by the Nickol River Formation, which unconformably overlies the KT. Plate separation was halted and reversed at c. 3160 Ma, most likely due to collision of the KT with another plate to the northwest. The change from separation to convergence coincided with a 3160–3140 Ma metamorphic event in the KT and with c. 3165 Ma granitic intrusion on the northwest margin of the EPT. Within the CPB convergence of the EPT and KT was accommodated by northwest obduction of the c. 3200 Ma juvenile basaltic crust (Regal Formation) onto the KT to form the Regal Terrane, and by the establishment of a northeast–southwest-trending subduction zone along the axis of the basin (possibly along the earlier spreading centre). From c. 3130 to 3093 Ma, volcanism above this subduction zone developed a >10 km-thick island arc succession (Whundo Group) and a belt of granitic intrusions (Railway Supersuite). Together, the Whundo Group and the Railway Supersuite are now exposed as the ST. Ongoing convergence finally culminated with 3068–3066 Ma collision and accretion of the KT, RT, and ST to form the WPS. Collision of the WPS and EPT resulted in the 3068–3066 Ma Prinsep Orogeny.

Between c. 3165 and 3070 Ma, the EPT had remained remote from the volcanism, granitic intrusion, and tectonic activity taking place in the northwest Pilbara. However, following the 3068–3066 Ma collision the east and northwest Pilbara shared many features of crustal evolution, including: a) intrusion of the 3068–3066 Ma Elizabeth Hill Supersuite during the Prinsep Orogeny; b) post-orogenic extension with development of the 3066–3015 Ma Gorge Creek Basin; c) post-3015 Ma deformation related to ongoing plate convergence, with associated sedimentation (Croydon Group); d) intrusion of the 2954–2919 Ma Sisters Supersuite during the North Pilbara Orogeny; and e) intrusion of the post-orogenic 2851–2831 Ma Split Rock Supersuite. An important difference between the east and northwest Pilbara following their collision was the development of a 3023–2982 Ma magmatic arc in the northwest Pilbara. This produced igneous and sedimentary stratigraphic units found only in the northwest Pilbara: the 3023–3012 Ma Orpheus Supersuite, the 3015–2990 Ma Whim Creek Group, the 3006–2982 Ma Maitland River Supersuite, and the 3015–2931 Ma back-arc basin succession in the Mallina Basin.

Evolution of the northwest Pilbara from c. 3220 Ma closely followed the sequence of stages in a Wilson cycle (Van Kranendonk et al., 2010). However, collision and orogeny at 3068–3066 Ma (Prinsep Orogeny) did not complete the cycle because after a brief interlude of post-orogenic extension (Gorge Creek Basin), plate convergence continued for another 100 million years, finally culminating in another collision (North Pilbara Orogeny) between c. 2955 Ma and 2919 Ma. This history is interpreted to be evidence of two stages of plate convergence. The first, responsible for CPB subduction and formation of the ST, was active from c. 3130 to at least c. 3093 Ma, whereas the second was active from c. 3025 Ma. This second stage of convergence is interpreted to have been accommodated by a southeasterly inclined subduction zone northwest of the Pilbara Craton. From c. 3025 to 2982 Ma this subduction produced granitic intrusion and volcanism in the northwest WPS, but from c. 2982 to 2919 Ma igneous activity migrated southeast into the Mallina Basin and the northwestern part of the EPT. The present location of the suture marking the northwest subduction zone may coincide with a linear gravity high about 100–150 km off the Pilbara coast.

How do these conclusions contribute to the ongoing debate over when plate tectonic processes commenced on Earth? The evidence from the Pilbara Craton is that plate tectonics operated from continental breakup of the EPT at c. 3220 Ma, although the Pilbara plates were small in comparison to many Phanerozoic plates. Evidence from the Kaapvaal Craton indicates similar processes of crustal evolution to those in the Pilbara (Van Kranendonk et al., 2015). The Paleoarchean–Mesoarchean stratigraphic similarity between the Pilbara and Kaapvaal Cratons (Zegers et al, 1998b), and common features of unconformably overlying Neoarchean–Paleoproterozoic successions in the Pilbara and South Africa (Trendall, 1968; Button, 1976; Blake, 1984; Nelson et al., 1992, 1999; Cheney, 1996; Martin et al., 1998; Thorne and Trendall, 2001; Eriksson et al., 2002; Pickard, 2003; de Kock et al., 2009, 2012) supports a conclusion that both cratons are fragments of the same Paleoarchean–Paleoproterozoic supercontinent, Vaalbara. Relatively poor preservation of geological evidence in the few other cratons that contain early Archean crust leave open the question as to whether the early crustal evolution of Vaalbara was representative of the Paleoarchean–Mesoarchean on a global scale.

References

Abeysinghe, PB and Flint, DJ 2008, Copper, lead, and zinc in Western Australia: a commodity review for 2006–07: Geological Survey of Western Australia, Record 2008/4, 34p.

Aihara, Y, Kiyokawa, S, Takehara, M and Horie, K 2012, Zircon U–Pb dating of the Cleaverville Formation, Pilbara, Australia: 34th International Geological Congress, Brisbane, Australia, 5–10 August 2012, Abstracts, p. 1658.

Archibald, NJ, Bettenay, LF, Binns, RA, Groves, DI and Gunthorpe, RJ 1978, The evolution of Archaean greenstone terrains, Eastern Goldfields Province, Western Australia: Precambrian Research, v. 6, p. 103–131.

Arndt, N, Albarede, F, Cheadle, M, Ginibre, C, Herzberg, C, Genner, G, Chauvel, C and Lahaye, Y 1998, Were komatiites wet?: Geology, v. 26, p. 739–742.

Arndt, N, Bruzak, G and Reischmann, T 2001, The oldest continental and oceanic plateaus: geochemistry of basalts and komatiites of the Pilbara Craton Australia, *in* Mantle Plumes: Their Identification Through Time *edited by* RE Ernst and KL Buchan: Geological Society of America Special Publication 352, Boulder, Colorado, USA, p. 359–87.

Bagas, L, Bierlein, FP, Bodorkos, S and Nelson, DR 2008, Tectonic setting, evolution, and orogenic gold potential of the late Mesoarchaean Mosquito Creek Basin, North Pilbara Craton, Western Australia: Precambrian Research, v. 160, p. 237–44.

Bagas, L, Farrell, TR and Nelson, DR 2004, The age and provenance of the Mosquito Creek Formation: Geological Survey of Western Australia, Annual Review 2003–04, p. 62–70.

Barley, ME 1987, The Archaean Whim Creek Belt, an ensialic fault-bounded basin in the Pilbara Block, Australia: Precambrian Research, v. 37, p. 199–215.

Barley, ME 1997, The Pilbara Craton, *in* Greenstone belts *edited by* MJ de Wit and L Ashwal: Oxford University Monographs on Geology and Geophysics, no. 35, p. 657–664.

Barley, ME, McNaughton, NJ, Williams, IS and Compston, W 1994, Age of Archaean volcanism and sulphide mineralization in the Whim Creek Belt, West Pilbara: Australian Journal of Earth Sciences, v. 41, p. 175–177.

Barnes, GB 1984, Soanesville/Tambourah: Preliminary Geological Report for Racomea Pty Ltd: Geological Survey of Western Australia, Statuary mineral exploration report, A13752 (unpublished), 46p.

Beintema, KA 2003, Geodynamic evolution of the west and central Pilbara Craton: a mid-Archean active continental margin: Geologica Ultraiectina, Utrecht University, Utrecht, The Netherlands, PhD thesis (unpublished), 248p.

Beintema, KA, de Leeue, GAM, White, SH and Hein, KAA 2001, Tabba Tabba Shear: a crustal scale structure in the Archaean Pilbara Craton, WA, *in* Extended Abstracts *edited by* KF Cassidy, JM Dunphy and MJ Van Kranendonk: 4th International Archaean Symposium, Perth, Western Australia, September 2001; AGSO, Record 2001/37, p. 285–287.

Beintema, KA, Mason, PRD, Nelson, DR, White, SH and Wijbrans, JR 2003, New constraints on the timing of tectonic activity in the Archaean central Pilbara Craton, Western Australia: Journal of the Virtual Explorer, v. 13, p. 16.

Bickle, MJ, Bettenay, LF, Chapman, HJ, Groves, DI, McNaughton, NJ, Campbell, IH and de Laeter, JR 1989, The age and origin of younger granitic plutons of the Shaw batholith in the Archaean Pilbara Block, Western Australia: Contributions to Mineralogy and Petrology, v. 101, p. 361–376.

Blake, TS 1984, The lower Fortescue Group of the northern Pilbara Craton — stratigraphy and paleogeography, *in* Archaean and Proterozoic basins of the Pilbara, Western Australia — evolution and mineralization potential *edited by* JR Muhling, DI Groves and TS Blake: The University of Western Australia, Geology Department and University Extension, Publication no. 9, p. 123–143.

Blake, TS 1993, Late Archaean crustal extension, sedimentary basin formation, flood basalt volcanism, and continental rifting: The Nullagine and Mount Jope Supersequences, Western Australia: Precambrian Research, v. 60, p. 185–241.

Blake, TS and Groves, DI 1987, Continental rifting and the Archean–Proterozoic transition: Geology, v. 15, p. 229–232.

Blewett, RS 2002, Archaean tectonic processes: a case for horizontal shortening in the North Pilbara granite–greenstone terrane, Western Australia: Precambrian Research, v. 113, p. 87–120.

Blewett, RS, Wellman, P, Ratajkoski, M and Huston, DI 2000, Atlas of North Pilbara Geology and Geophysics, 1:1.5 million scale: Geological Survey of Western Australia, Record 2000/04, 36p.

Blockley, JG 1971, Geology and mineral resources of the Wodgina district: Geological Survey of Western Australia, Annual Report for 1970, p. 38–42.

Blockley, JG 1980, The tin deposits of Western Australia, with special reference to the associated granites: Geological Survey of Western Australia, Mineral Resources Bulletin 12, 184p.

Button, A 1976, Transvaal and Hamersley Basins – review of basin development and mineral deposits: Mineral Science Engineering, v. 8, p. 262–293.

Byerly, GR, Lowe, DR, Wooden, JL and Xiaoogang, XIA 2002, A meteorite impact layer 3470 Ma from the Pilbara and Kaapvaal Cratons: Science, v. 297, p. 1325–1327.

Carpentaria Exp Co Pty Ltd 1983, Carpentaria Exp Co Pty Ltd Report 1983 E45/50 Mount Grant: Geological Survey of Western Australia, Statutory mineral exploration report, A13799 (unpublished).

Caudery, JN 2014, Structural evolution of the Yalgoo Dome, Yilgarn Craton, Western Australia: Geological Survey of Western Australia, Record 2014/4, 87p.

Champion, DC 2013, Neodymium depleted mantle model age map of Australia: explanatory notes and user guide: Geoscience Australia, Record 2013/44, 209p, doi:10.11636/Record.2013.044.

Champion, DC and Smithies, RH 2001, The geochemistry of the Yule Granitoid Complex, East Pilbara Granite – Greenstone Terrane: evidence for early felsic crust: Geological Survey of Western Australia, Annual Review 1999–2000, p. 42–48.

Cheney, ES 1996, Sequence stratigraphy and plate tectonic significance of the Transvaal succession of southern Africa and its equivalent in Western Australia: Precambrian Research, v. 79, p. 3–24.

Cheney, ES, Roering, C and Stettler, E 1988, Vaalbara: Geological Society of South Africa, Geocongress 88, Extended Abstracts, p. 85–88.

Collins, PLF and Marshall, AE 1999a, Volcanic-hosted massive sulfide deposits at Whundo–Yannery in the Sholl belt, *in* Lead, zinc, and silver deposits of Western Australia *edited by* KM Ferguson: Geological Survey of Western Australia, Mineral Resources Bulletin 15, p. 73–79.

Collins, PLF and Marshall, AE 1999b, Volcanogenic base-metal deposits of the Whim Creek Belt, *in* Lead, zinc and silver deposits of Western Australia *edited by* KM Ferguson: Geological Survey of Western Australia, Mineral Resources Bulletin 15, p. 44–72.

Condie, KC 2001, Mantle plumes and their record in Earth history: Cambridge University Press, Cambridge, United Kingdom, 306p.

Condie, KC 2004, Supercontinents and superplume events: distinguishing signals in the geological record: Physics of the Earth and Planetary Interiors, v. 146, p. 319–332.

CRA Exploration Pty Ltd 1991, CRA Exploration Pty Ltd Annual Report 1990–1991 E45/691–692, E 45/697/699, E 45/1025 Worthy: Geological Survey of Western Australia, Statutory mineral exploration report, A34890 (unpublished).

Crawford, AJ, Falloon, TJ and Green, DH 1989, Classification, petrogenesis and tectonic setting of boninites, *in* Boninites *edited by* AJ Crawford: Unwin Hyman, London, United Kingdom, p. 1–49.

Dalstra, HJ, Bloem, EJM, Ridley, JR and Groves, DI 1998, Diapirism synchronous with regional deformation and gold mineralisation, a new concept for granitoid emplacement in the Southern Cross Province, Western Australia: Geologie en Mijnbouw, v. 76, p. 321–338.

Davy, R and Lewis, JD 1986, The Mount Edgar Batholith, Pilbara area, WA – geochemistry and petrography: Geological Survey of Western Australia, Report 17, 42p.

De Grey Mining Limited 2005, Quarterly operations report to Australian Securities Exchange, three months ending September 2005.

De Grey Mining Limited 2012, 40% Increase in Turner River gold resources: Report to Australian Securities Exchange, 11 April 2012, 7p.

De Grey Mining Limited 2014, Turner River – base metals project update: Report to Australian Securities Exchange, 16 July 2014, 27p.

De Kock, MO, Beukes NJ and Armstrong, RA 2012, New SHRIMP U–Pb zircon ages from the Hartswater Group, South Africa: Implications for correlations of the Neoarchean Ventersdorp Supergroup on the Kaapvaal craton and with the Fortescue Group on the Pilbara craton: Precambrian Research, v. 204–205, p. 66–74.

De Kock, MO, Evans, DA and Beukes, NJ 2009,Validating the existence of Vaalbara in the Neoarchean: Precambrian Research v. 174, p. 145–54.

De Laeter, JR and Blockley, JG 1972, Granite ages within the Pilbara Block, Western Australia: Geological Society of Australia Journal, v. 19, p. 363–370.

De Laeter, JR, Lewis, JD and Blockley, JG 1975, Granite ages within the Shaw Batholith of the Pilbara Block: Geological Survey of Western Australia, Annual Report 1974, p. 73–79.

Donaldson, MJ 1974, Petrology of the Munni Munni Complex, Roebourne, Western Australia: Geological Society of Australia Journal, v. 21, p. 1–16.

Ellis, HA 1950, Some economic aspects of the principal tantalum-bearing deposits of the Pilbara Goldfield, Northwest Division: Geological Survey of Western Australia, Bulletin 104, 93p.

Eriksson, KA 1981, Archaean platform-to-trough sedimentation, East Pilbara Block, Australia, *in* Archaean Geology *edited by* JE Glover and DI Groves: Geological Society Australia, Second International Archaean Symposium, Perth, Western Australia, 1980: Special Publication 7, p. 235–244.

Eriksson, KA 1982, Geometry and internal characteristics of Archaean submarine channel deposits Pilbara Block Western Australia: Journal of Sedimentary Petrology, v. 52, no. 2 p. 383–393.

Eriksson, PG, Condie, KC, van der Westhuizen, W, van der Merwe, R, de Bruiyn, H, Nelson, DR, Altermann, W, Catuneanu, O, Bumby, AJ, Lindsay, J and Cunningham, MJ 2002, Late Archaean superplume events: a Kaapvaal–Pilbara perspective: Journal of Geodynamics, v. 34, p. 207–247.

Ernst, RE, Buchan, KL and Prokoph, A 2004, Large igneous province record through time, *in* The Precambrian Earth: tempos and events *edited by* PG Eriksson, W Altermann, DR Nelson, WU Mueller and O Catuneanu: Developments in Precambrian Geology 12, p. 173–180.

Fenwick, MJ 2014, Structural evolution of the Yalgoo Dome, Yilgarn Craton, Western Australia: Geological Survey of Western Australia, Record 2014/16, 93p.

Ferguson, KM 1999, Lead, zinc and silver deposits of Western Australia: Geological Survey of Western Australia, Mineral Resources Bulletin 15, 314p.

Finucane, KJ and Telford, RJ 1939, The antimony deposits of the Pilbara Goldfield: Aerial, Geological and Geophysical Survey of Northern Australia, Western Australia, Report 47, 6p.

Fitton, MJ, Horwitz, RC and Sylvester, G 1975, Stratigraphy of the Early Precambrian in the West Pilbara: CSIRO Minerals Research Laboratories, Division of Mineralogy, Report FP 11, 41p.

Fox Resources Limited 2006, West Whundo production update: report to Australian Securities Exchange, 14 September 2006, 2p.

Fox Resources Limited 2007a, Whundo copper resource increases 600%: report to Australian Securities Exchange, 30 January 2007, 5p.

Fox Resources Limited 2007b, Nickel Production at Radio Hill: report to Australian Securities Exchange, 30 May 2007, 6p.

Fox Resources Limited 2012, Significant Copper Mineralisation at First Aysia Drill Hole: report to Australian Securities Exchange, 29 February 2012, 5p.

Frick, LR, Lambert, DD and Hoatson, DM 2001, Re–Os dating of the Radio Hill Ni–Cu deposit, west Pilbara Craton, Western Australia: Australian Journal of Earth Sciences, v. 48, no. 1, p. 43–47.

Gee, RD, Baxter, JL, Wilde, SA and Williams, IR 1981, Crustal development in the Yilgarn Block, *in* Archaean Geology *edited by* JE Glover and DI Groves: Geological Society of Australia, Second International Archaean Symposium: Perth, Western Australia, Special Publication 7, p. 43–56.

Glass, BP and Simonson, BM 2012, Distal impact ejecta layers: spherules and more: Elements, v. 8, p. 43–48.

Glass, BP and Simonson, BM 2013, Distal impact ejecta layers: a record of large impacts in sedimentary deposits: Springer Verlag, Heidelberg, Germany, 716p.

Glikson, A and Vickers, J 2006, The 3.26 – 3.24 Ga Barberton asteroid impact cluster: tests of tectonic and magmatic consequences, Pilbara Craton, Western Australia: Earth and Planetary Science Letters, v. 241, no. 1–2, p. 11–20.

Griffin, TJ 1990, North Pilbara granite–greenstone terrane, in Geology and Mineral Resources of Western Australia: Geological Survey of Western Australia, Memoir 3, p. 128–158.

Hassler, SW, Simonson, BM, Sumner, DY and Bodin, L 2011, Paraburdoo spherule layer (Hamersley Basin, Western Australia): Distal ejecta from a fourth large impact near the Archean–Proterozoic boundary: Geology, v. 39, p. 307–310.

Hickman, AH 1975a, Explanatory notes on the Nullagine 1:250 000 geological sheet: Geological Survey of Western Australia, Record 1975/5, 50p.

Hickman, AH 1975b, Precambrian structural geology of part of the Pilbara region: Geological Survey of Western Australia, Annual Report 1974, p. 68–73.

Hickman, AH 1977, Stratigraphic relations of rocks within the Whim Creek Belt, in Annual report 1976: Geological Survey of Western Australia, p. 53–56.

Hickman, AH 1980a, Archaean geology of the Pilbara Block: Excursion guide: Geological Society of Australia, WA Division, Second International Archaean Symposium, Perth, Western Australia, 1980, 55p.

Hickman, AH 1980b, Lithological map and stratigraphic interpretation of the Pilbara Block, in Geology of the Pilbara Block and its environs by AH Hickman: Geological Survey of Western Australia, Bulletin 127, Plate 1.

Hickman, AH 1981, Crustal evolution of the Pilbara Block, in Archaean Geology *edited by* JE Glover and DI Groves: Geological Society Australia, Second International Archaean Symposium, Perth, Western Australia, 1980: Special Publication 7, p. 57–69.

Hickman, AH 1983, Geology of the Pilbara Block and its environs: Geological Survey of Western Australia, Bulletin 127, 268p.

Hickman, AH 1984, Archaean diapirism in the Pilbara Block, Western Australia, *in* Precambrian tectonics illustrated *edited by* A Kröner and RE Greiling: Schweizerbart'sche Verlagsbuchhandlung, Stuttgart, Germany, p. 113–127.

Hickman, AH 1990, Geology of the Pilbara Craton, *in* Excursion Guidebook No. 5, Pilbara and Hamersley Basin, Third International Archaean Symposium Perth, Western Australia *edited by* SE Ho, JE Glover, JS Myers and JR Muhling: The University of Western Australia, Geology Department and University Extension, Publication 21, p. 2–13.

Hickman, AH 1997a, A revision of the stratigraphy of Archaean greenstone successions in the Roebourne–Whundo area, west Pilbara, *in* Geological Survey of Western Australia Annual Review 1996–97: Geological Survey of Western Australia, p. 76–82.

Hickman, AH 1997b, Dampier, WA Sheet 2256: Geological Survey of Western Australia, 1:100 000 Geological Series.

Hickman, AH 1999, New tectono-stratigraphic interpretations of the Pilbara Craton, Western Australia, *in* GSWA 99 extended abstracts: new geological data for WA explorers: Geological Survey of Western Australia, Record 1999/6, p. 4–6.

Hickman, AH 2000, Roebourne, WA Sheet 2356: Geological Survey of Western Australia, 1:100 000 Geological Series.

Hickman, AH 2001a, The West Pilbara Granite – Greenstone Terrane, and its place in the Pilbara Craton, *in* Extended Abstracts *edited by* KF Cassidy, JM Dunphy and MJ Van Kranendonk: 4th International Archaean Symposium, Perth, Western Australia, September 2001; AGSO, Record 2001/37, p. 319–321.

Hickman, AH 2001b, Geology of the Dampier 1:100 000 sheet: Geological Survey of Western Australia, 1:100 000 Geological Series Explanatory Notes, 39p.

Hickman, AH 2002, Geology of the Roebourne 1:100 000 sheet: Geological Survey of Western Australia, 1:100 000 Geological Series Explanatory Notes, 35p.

Hickman, AH 2004a, Two contrasting granite–greenstones terranes in the Pilbara Craton, Australia: evidence for vertical and horizontal tectonic regimes prior to 2900 Ma: Precambrian Research, v. 131, p. 153–172.

Hickman, AH 2004b, Geology of the Cooya Pooya 1:100 000 sheet: Geological Survey of Western Australia, 1:100 000 Geological Series Explanatory Notes, 25p.

Hickman, AH 2004c, Using new-generation geological maps of the Pilbara to guide mineral exploration, *in* GSWA 2004 extended abstracts: promoting the prospectivity of Western Australia: Geological Survey of Western Australia, Record 2004/5, p. 1–4.

Hickman, AH 2008, Regional review of the 3426–3350 Ma Strelley Pool Formation, Pilbara Craton, Western Australia: Geological Survey of Western Australia, Record 2008/15, 27p.

Hickman, AH 2012, Review of the Pilbara Craton and Fortescue Basin, Western Australia: Crustal evolution providing environments for early life: Island Arc, v. 21, p. 1–31.

Hickman, AH in prep., East Pilbara Craton: 750 million years in the growth of an Archean continent: Geological Survey of Western Australia.

Hickman, AH and Harrison, PH 1986, A review of the occurrence of and potential for Precambrian conglomerate-hosted gold mineralization within Western Australia, *in* Geocongress '86 – Extended Abstracts: Geological Society of South Africa; University of Witwatersrand, Johannesburg, Republic of South Africa, p. 301–319.

Hickman, AH, Huston, DL, Van Kranendonk, MJ and Smithies, RH 2006a, Geology and mineralization of the west Pilbara — a field guide: Geological Survey of Western Australia, Record 2006/17, 50p.

Hickman, AH and Kojan, CJ 2003, Geology of the Pinderi Hills 1:100 000 sheet: Geological Survey of Western Australia, 1:100 000 Geological Series Explanatory Notes, 36p.

Hickman, AH, and Lipple, SL 1975, Explanatory notes on the Marble Bar 1:250 000 Geological Sheet, W.A.: Geological Survey of Western Australia, Record 1974/20, 24p.

Hickman, AH and Lipple, SL 1974, Explanatory notes on the Marble Bar 1:250 000 geological sheet, Western Australia: Geological Survey of Western Australia, Record 1974/20, 90p.

Hickman, AH and Smithies, RH 2000, Roebourne, WA Sheet SF 50-3 (2nd edition): Geological Survey of Western Australia, 1:250 000 Geological Series.

Hickman, AH and Smithies, RH 2001, Roebourne, Western Australia (2nd edition): Geological Survey of Western Australia, 1:250 000 Geological Series Explanatory Notes, 52p.

Hickman, AH, Smithies, RH and Huston, DL 1998, Excursion guide to the geology of the granite–greenstone terrane of the west Pilbara: Geological Survey of Western Australia, 51p.

Hickman, AH, Smithies, RH, Pike, G, Farrell, TR and Beintema, KA 2001, Evolution of the West Pilbara granite–greenstone terrane and Mallina Basin, Western Australia — a field guide: Geological Survey of Western Australia, Record 2001/16, 65p.

Hickman, AH, Smithies, RH and Huston, DL 2000, Archaean geology of the West Pilbara Granite–Greenstone Terrane and Mallina Basin, Western Australia — a field guide: Geological Survey of Western Australia, Record 2000/9, 61p.

Hickman, AH, Smithies, RH and Strong, CA 2006b, Interpreted bedrock geology of the northwestern Pilbara Craton (1:250 000): Geological Survey of Western Australia, Plate 1.

Hickman AH, Smithies, RH and Tyler, IM 2010, Evolution of active plate margins: West Pilbara Superterrane, De Grey Superbasin, and the Fortescue and Hamersley Basins — a field guide: Geological Survey of Western Australia Record 2010/3, 74p.

Hickman, AH and Strong, CA 1999, Structures in the Cape Preston area, northwest Pilbara region — new evidence for a convergent margin: Geological Survey of Western Australia, Annual Review 1997–98, p. 77–84.

Hickman, AH and Strong, CA 2003, Dampier – Barrow Island, Western Australia (2nd edition): Geological Survey of Western Australia, 1:250 000 Geological Series Explanatory Notes, 75p.

Hickman, AH and Van Kranendonk, MJ 2004, Diapiric processes in the formation of Archean continental crust, East Pilbara Granite – Greenstone Terrane, Australia, *in* The Precambrian Earth: tempos and events *edited by* PG Eriksson, W Altermann, DR Nelson, WU Mueller and O Catuneanu: Developments in Precambrian Geology 12, p. 54–75.

Hickman, AH and Van Kranendonk, MJ 2012, Early Earth evolution: evidence from the 3.5–1.8 Ga geological history of the Pilbara region of Western Australia: Episodes, v. 35, p. 283–297.

Hoatson, DM 1986, Geology of the Munni Munni layered intrusion: Preliminary Edition 1:20 000 geological map: Australian Bureau of Mineral Resources, Geology and Geophysics.

Hoatson, DM 1998, Platinum-group element mineralisation in Australian Precambrian layered mafic–ultramafic intrusions: AGSO Journal of Australian Geology and Geophysics, v. 17, no. 4, p. 139–151.

Hoatson, DM and Sun, S-S 2002, Archean layered mafic–ultramafic intrusions in the west Pilbara Craton, Western Australia: a synthesis of some of the oldest orthomagmatic mineralizing systems in the world: Economic Geology, v. 97, p. 847–872.

Hoatson, DM, Wallace, DA, Sun, S-S, Macias, LF, Simpson, CJ and Keays, RR 1992, Petrology and platinum-group element geochemistry of Archaean layered mafic–ultramafic intrusions, west Pilbara Block, Western Australia: Australian Geological Survey Organisation, Bulletin 242, 319p.

Horwitz, RC 1990, Palaeogeographic and tectonic evolution of the Pilbara Craton, Northwestern Australia: Precambrian Research, v. 48, p. 327–340.

Huston, DL 2006, Mineralization and regional alteration at the Mons Cupri stratiform Cu–Zn–Pb deposit, Pilbara Craton, Western Australia: Mineralium Deposita, v. 41, p. 17–32.

Huston, DL, Smithies, RH and Sun, S-S 2000, Correlation of the Archaean Mallina – Whim Creek Basin: implications for base-metal potential of the central part of the Pilbara granite–greenstone terrane: Australian Journal of Earth Sciences, v. 47, no. 2, p. 217–230.

Huston, DL, Blewett, RS, Sweetapple, M, Brauhart, C, Cornelius, R and Collins, PLF 2001, Metallogenesis of the north Pilbara granite–greenstones, Western Australia — a field guide: Geological Survey of Western Australia, Record 2001/11, 87p.

Huston, DL, Sun, S-S, Blewett, R, Hickman, AH, Van Kranendonk, MJ, Phillips, D, Baker, D and Brauhart, C 2002a, The timing of mineralisation in the Archaean Pilbara Craton, Western Australia: Economic Geology, v. 97, p. 733–755.

Huston, DL, Blewett, Keillor, B, Standing, J, Smithies, RH, Marshall, A, Mernaugh, TP and Kamprad, J 2002b, Lode gold and epithermal deposits of the Mallina Basin, North Pilbara Terrain, Western Australia: Economic Geology, v. 97, p. 801–818.

Isozaki, Y, Maruyama, S and Kimura, G 1991, Middle Archean (3.3 Ga) Cleaverville accretionary complex in Northwestern Pilbara Block, Western Australia: Eos, v. 72, p. 542.

Johnson, SP, Cutten, HN, Tyler, IM, Korsch, RJ, Thorne, AM, Blay, O, Kennett, BLN, Blewett, RS, Joly, A, Dentith, MC, Aitken, ARA, Goodwin, JA, Salmon, M, Reading, A,Boren, G, Ross, J, Costelloe, RD and Fomin, T 2011, Preliminary interpretation of deep seismic reflection lines 10GA–CP2 and 10GA–CP3: crustal architecture of the Gascoyne Province, and Edmund and Collier Basins, *in* Capricorn Orogen seismic and magnetotelluric (MT) workshop 2011: Extended Abstracts *edited by* SP Johnson, AM Thorne and IM Tyler: Geological Survey of Western Australia, Record 2011/25, p. 49–60.

Johnson, SP, Sheppard, S, Rasmussen, B, Wingate, MTD, Kirkland, CL, Muhling, JR, Fletcher, IR and Belousova, E 2010, The Glenburgh Orogeny as a record of Paleoproterozoic continent–continent collision: Geological Survey of Western Australia, Record 2010/5, 54p.

Johnson, SP, Thorne, AM, Tyler, IM, Korsch, RJ, Kennett, BLN, Cutten, HN, Goodwin, J, Blay, O, Blewett, RS, Joly, A, Dentith, MC, Aitken, ARA, Holzschuh, J, Salmon, M, Reading, A, Heinson, G, Boren, G, Ross, J, Costelloe, RD and Fomin, T 2013, Crustal architecture of the Capricorn Orogen, Western Australia and associated metallogeny: Australian Journal of Earth Sciences, v. 60, p. 681–705.

Jones-Zimberlin, S, Simonson, BM, Kreiss-Tomkins, D and Garson, D 2006, Using impact spherule layers to correlate sedimentary successions: a case study of the Neoarchean Jeerinah layer (Western Australia): South African Journal of Geology, v. 109, p. 245–261.

Kato, Y, Ohta, I, Tsunematsu, T, Watanabe, Y, Isozaki, Y, Maruyama, S and Imai, N 1998, Rare earth element variations in mid-Archean banded iron formations: implications for the chemistry of ocean and continent plate tectonics: Geochimica et Cosmochimica Acta, v. 62, p. 3475–3497.

Kemp, AIS, Hickman, AH, Kirkland, CL and Vervoort, JD 2015a, Hf isotopes in detrital and inherited zircons of the Pilbara Craton provide no evidence for Hadean continents: Precambrian Research, v. 261, p. 112–126.

Kemp, AIS, Hickman, AH and Kirkland, CL 2015b, Early evolution of the Pilbara Craton from hafnium isotopes in detrital and inherited zircons: Geological Survey of Western Australia, Report 151, 26p.

Kinny, PD 2000, U–Pb dating of rare-metal (Sn–Ta–Li) mineralized pegmatites in Western Australia by SIMS analysis of tin and tantalum-bearing ore minerals: Beyond 2000, New Frontiers in Isotope Geoscience, Lorne, Victoria, 30 January to 4 February 2000, Abstracts and Proceedings, p. 113–116.

Kirkland, CL, Johnson, SP, Smithies, RH, Hollis, J, Wingate, MTD, Tyler, IM, Hickman, AH, Cliff, JB, Belousova, EA, Murphy, R and Tessalina, S 2013, The crustal evolution of the Rudall Province from an isotopic perspective: Geological Survey of Western Australia, Report 122, 30p.

Kiyokawa, S 1993, Stratigraphy and structural evolution of a Middle Archaean greenstone belt, northwestern Pilbara Craton: University of Tokyo, Tokyo, Japan, PhD thesis (unpublished).

Kiyokawa, S, Ito, T, Ikehara, M, Yamaguchi, KE, Koge, S and Sakamoto, R 2012, Lateral variations in the lithology and organic chemistry of a black shale sequence on the Mesoarchean seafloor affected by hydrothermal processes: the Dixon Island Formation of the coastal Pilbara Terrane, Western Australia: Island Arc, v. 21, no. 2, p. 118–147.

Kiyokawa, S, Koge, S, Ito, T and Ikehara, M 2014, An ocean-floor carbonaceous sedimentary sequence in the 3.2 Ga Dixon Island Formation, coastal Pilbara terrane, Western Australia: Precambrian Research, v. 255, no. 1, p. 124–143.

Kiyokawa, S and Taira, A 1998, The Cleaverville Group is the west Pilbara coastal granite–greenstone terrane of Western Australia: an example of a mid-Archaean immature oceanic island-arc succession: Precambrian Research, v. 88, p. 102–142.

Kiyokawa, S, Taira, A, Byrne, T, Bowring, S and Sano, Y 2002, Structural evolution of the middle Archean coastal Pilbara terrane, Western Australia: Tectonics, v. 21, no. 1044, p. 1–24.

Korsch, RJ, Johnson, SP, Tyler, IM, Thorne, AM, Blewett, RS, Cutten, HN, Joly, A, Dentith, MC, Aitken, ARA, Goodwin, JA and Kennett, BLN 2011, Geodynamic implications of the Capricorn deep seismic survey: from the Pilbara Craton to the Yilgarn Craton, *in* Capricorn Orogen seismic and magnetotelluric (MT) workshop 2011: extended abstracts *edited by* SP Johnson, AM Thorne and IM Tyler: Geological Survey of Western Australia, Record 2011/25, p. 107–114.

Krapez, B 1984, Sedimentation in a small, fault-bounded basin: the Lalla Rookh sandstone, east Pilbara Block, *in* Archaean and Proterozoic basins of the Pilbara, Western Australia: Evolution and Mineralization potential *edited by* JR Muhling, DI Groves and TS Blake: The University of Western Australia, Geology Department and University Extension, Publication 9, p. 89–110.

Krapez, B 1993, Sequence stratigraphy of the Archaean supracrustal belts of the Pilbara Block, Western Australia: Precambrian Research, v. 60, p. 1–45.

Krapez, B and Barley, ME 1987, Archean strike-slip faulting and related ensialic basins: evidence from the Pilbara Block, Australia: Geological Magazine, v. 124, p. 55–567.

Krapez, B and Groves, DI 1984, Gold mineralization potential of Archaean clastic sequences in the east Pilbara Block, *in* Archaean and Proterozoic basins of the Pilbara, Western Australia: Evolution and Mineralization potential *edited by* JR Muhling, DI Groves and TS Blake: The University of Western Australia, Geology Department and University Extension, Publication 9, p. 111–122.

Krapez, B and Eisenlohr, B 1998, Tectonic settings of Archaean (3325–2775 Ma) crustal–supracrustal belts in the West Pilbara Block: Precambrian Research, v. 88, p. 173–205.

Leggo, PJ, Compston, W and Trendall, AF 1965, Radiometric ages of some Precambrian rocks from the North-West Division of Western Australia: Journal of the Geological Society of Australia, v. 12, p. 53–65.

Low, GH 1963, Copper deposits of Western Australia: Geological Survey of Western Australia, Mineral Resources Bulletin 8, 202p.

Lowe, DR 2013, Crustal fracturing and chert dyke formation triggered by large meteorite impacts, ca. 3260 Ma, Barberton greenstone belt, South Africa: Geological Society of America Bulletin, v. 125, p. 894–912.

Lowe, DR and Byerly, GR 1986, Early Archean silicate spherules of probable impact origin, South Africa and Western Australia: Geology, v. 14, p. 83–86.

Marshall, AE 2000, Low temperature–low pressure (‘epithermal’) vein deposits of the North Pilbara granite–greenstone terrane, Western Australia: Australian Geological Survey Organization, Record 2000/1, 40p.

Marston, RJ 1979, Copper mineralisation in Western Australia: Geological Survey of Western Australia, Mineral Resources Bulletin 13, 208p.

Marston, RJ 1984, Nickel mineralisation in Western Australia: Geological Survey of Western Australia, Mineral Resources Bulletin 14, 271p.

Martin, DMcB, Clendenin, CW, Krapez, B and McNaughton, MJ 1998, Tectonic and geochronological constraints on late Archaean and Palaeoproterozoic stratigraphic correlation within and between the Kaapvaal and Pilbara cratons: Journal of the Geological Society of London, v. 155, p. 311–322.

Mathison, CI and Marshall, AE 1981, Ni–Cu sulphides and their host mafic–ultramafic rocks in the Mount Sholl intrusion, Pilbara region, Western Australia: Economic Geology, v. 76, p. 1581–1596.

Miles, KR, Carroll, D and Rowledge, HP 1945, Tantalum and niobium: Geological Survey of Western Australia, Mineral Resources Bulletin 3, 150p.

Miller, LJ and Gair, HS 1975, Mons Cupri copper-lead-zinc deposit, *in* Economic Geology of Australia and Papua New Guinea — Metals *edited by* CL Knight: Australasian Institute of Mining and Metallurgy, Monograph 5, p. 195–202.

Miller, LJ and Smith, ME 1975, Sherlock Bay nickel-copper deposit, *in* Economic Geology of Australia and Papua New Guinea — Metals *edited by* CL Knight: Australasian Institute of Mining and Metallurgy, Monograph 5, p. 168–174.

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

Nelson, DR 1996, 114350: metadacite, Mount Sholl; Geochronology Record 478: Geological Survey of Western Australia, 4p.

Nelson, DR 1997a, 118964: foliated granite, Caines Well; Geochronology Record 456: Geological Survey of Western Australia, 4p.

Nelson, DR 1997b, 118965: equigranular biotite monzogranite gneiss, old highway – Sherlock River crossing; Geochronology Record 457: Geological Survey of Western Australia, 4p.

Nelson, DR 1997c, 118966: porphyritic granodioritic gneiss, Forestier Bay, Geochronology Record 458: Geological Survey of Western Australia, 4p.

Nelson, DR 1997d, 118967: equigranular hornblende-biotite tonalite, Ten Foot Well, Geochronology Record 459: Geological Survey of Western Australia, 4p.

Nelson, DR 1997e, 118969: fine-grained greywacke sandstone, May Bore, Geochronology Record 445: Geological Survey of Western Australia, 4p.

Nelson, DR 1997f, 118974: foliated porphyritic hornblende granodiorite, Baynton Hill, Geochronology Record 431: Geological Survey of Western Australia, 4p.

Nelson, DR 1997g, 118975: porphyritic rhyolite, Mount Regal; Geochronology Record 432: Geological Survey of Western Australia, 4p.

Nelson, DR 1997h, 118976: porphyritic dacite, Cherratta Road–Nickol River Crossing, Geochronology Record 433: Geological Survey of Western Australia, 4p.

Nelson, DR 1997i, 118979: quartz-feldspar porphyry, No. 6 Well, Geochronology Record 434: Geological Survey of Western Australia, 4p.

Nelson, DR 1997j, 127320: quartz granophyre, Mount Ada; Geochronology Record 440: Geological Survey of Western Australia, 4p.

Nelson, DR 1997k, 136826: equigranular biotite–hornblende tonalite gneiss, Toorare Pool; Geochronology Record 414: Geological Survey of Western Australia, 5p.

Nelson, DR 1998a, 127327: dacite porphyry, Rocky Creek, Geochronology Record 441: Geological Survey of Western Australia, 3p.

Nelson, DR 1998b, 127330: volcaniclastic sedimentary rock, Cleaverville; Geochronology Record 442: Geological Survey of Western Australia, 4p.

Nelson, DR 1998c, 127378: welded tuff, Woodbrook homestead, Geochronology Record 443: Geological Survey of Western Australia, 3p.

Nelson, DR 1998d, 136819: quartz–mica schist, Lydia Mine; Geochronology Record 413: Geological Survey of Western Australia, 3p.

Nelson, DR 1998e, 136844: granite, Dampier Salt Ponds; Geochronology Record 415: Geological Survey of Western Australia, 3p.

Nelson, DR 1998f, 136899: volcanigenic sedimentary rock, Wickham; Geochronology Record 416: Geological Survey of Western Australia, 3p.

Nelson, DR 1998g, 141936: welded tuff, Red Hill; Geochronology Record 396: Geological Survey of Western Australia, 3p.

Nelson, DR 1998h, 141973: biotite monzogranite, Wakeman Well; Geochronology Record 397: Geological Survey of Western Australia, 3p.

Nelson, DR 1998i, 141977: granite, Manyon Well, Geochronology Record 398: Geological Survey of Western Australia, 3p.

Nelson, DR 1998j, 142433: tonalite, Mount Regal; Geochronology Record 403: Geological Survey of Western Australia, 3p.

Nelson, DR 1998k, 142436: micromonzonite dyke, Munni Munni; Geochronology Record 404: Geological Survey of Western Australia, 3p.

Nelson, DR 1998l, 142535: foliated hornblende-biotite tonalite, Tarlwa Pool, Geochronology Record 406: Geological Survey of Western Australia, 3p.

Nelson, DR 1998m, 142661: foliated biotite tonalite, Zebra Hill; Geochronology Record 409: Geological Survey of Western Australia, 3p.

Nelson, DR 1998n, 142830: volcanigenic sedimentary rock, Mount Ada; Geochronology Record 390: Geological Survey of Western Australia, 3p.

Nelson, DR 1998o, 142842: volcaniclastic sedimentary rock, Nunyerry Gap, Geochronology Record 379: Geological Survey of Western Australia, 4p.

Nelson, DR 1998p, 142882: biotite monzogranite, west of Mulgandinna Hill, Geochronology Record 351: Geological Survey of Western Australia, 3p.

Nelson, DR 1998q, 142883: foliated porphyritic syenogranite dyke, south of Mulgandinna Hill, Geochronology Record 352: Geological Survey of Western Australia, 3p.

Nelson, DR 1998r, 142884: schlieric biotite syenogranite, Mount Webber, Geochronology Record 353: Geological Survey of Western Australia, 3p.

Nelson, DR 1998s, 142885: biotite monzogranite, Mundine Well, Geochronology Record 354: Geological Survey of Western Australia, 3p.

Nelson, DR 1998t, 143994: quartzite, Kittys Gap, Geochronology Record 266: Geological Survey of Western Australia, 4p.

Nelson, DR 1998u, 143995: quartzite, Friendly Stranger Mine; Geochronology Record 267: Geological Survey of Western Australia, 4p.

Nelson, DR 1998v, 144261: rhyolite, Bradley Well, Geochronology Record 272: Geological Survey of Western Australia, 4p.

Nelson, DR 1999a, 142176: megacrystic foliated biotite monzogranite, Yandeyarra Homestead, Geochronology Record 400: Geological Survey of Western Australia, 3p.

Nelson, DR 1999b, 142430: monzogranite, Black Hill Well, Geochronology Record 402: Geological Survey of Western Australia, 4p.

Nelson, DR 1999c, 142438: granodiorite, Pinderi Hills, Geochronology Record 405: Geological Survey of Western Australia, 3p.

Nelson, DR 1999d, 142657: granodiorite, Cadgerina Pool, Geochronology Record 407: Geological Survey of Western Australia, 4p.

Nelson, DR 1999e, 142835: tonalitic gneiss, Whundo, Geochronology Record 392: Geological Survey of Western Australia, 3p.

Nelson, DR 1999f, 142870: biotite banded gneiss, 6 Mile Well, Geochronology Record 345: Geological Survey of Western Australia, 4p.

Nelson, DR 1999g, 142889: foliated alkali granite, Roberts Hill; Geochronology Record 355: Geological Survey of Western Australia, 4p.

Nelson, DR 1999h, 142892: porphyritic rhyolite, Two Mile Well; Geochronology Record 356: Geological Survey of Western Australia, 4p.

Nelson, DR 1999i, 142893: schlieric, pegmatite-veined monzogranite, Eramurra Creek, Geochronology Record 357: Geological Survey of Western Australia, 5p.

Nelson, DR 1999j, 144224: dacite porphyry, Mount Wangee, Geochronology Record 270: Geological Survey of Western Australia, 3p.

Nelson, DR 2000a, 142832: metamorphosed intermediate rock, Whundo; Geochronology Record 391: Geological Survey of Western Australia, 4p.

Nelson, DR 2000b, 142934: biotite monzogranite, Florrie Well; Geochronology Record 321: Geological Survey of Western Australia, 4p.

Nelson, DR 2000c, 142935: hornblende granodiorite, Wallareenya Homestead; Geochronology Record 322: Geological Survey of Western Australia, 4p.

Nelson, DR 2000d, 142936: leucocratic monzogranite, Yandeyarra Pool; Geochronology Record 323: Geological Survey of Western Australia, 4p.

Nelson, DR 2000e, 142937: leucocratic monzogranite, Numbana Pool; Geochronology Record 324: Geological Survey of Western Australia, 4p.

Nelson, DR 2000f, 142938: biotite–hornblende tonalite, Cheearra Hill; Geochronology Record 325: Geological Survey of Western Australia, 4p.

Nelson, DR 2000g, 142941: feldspar–hornblende porphyry, Black Gin Well; Geochronology Record 326: Geological Survey of Western Australia, 3p.

Nelson, DR 2000h, 142942: metasandstone, Croydon Well; Geochronology Record 327: Geological Survey of Western Australia, 4p.

Nelson, DR 2000i, 142943: metasandstone, Croydon Homestead; Geochronology Record 311: Geological Survey of Western Australia, 4p.

Nelson, DR 2000j, 142945: plagioclase–hornblende–pyroxene andesite porphyry, Jigimining Pool; Geochronology Record 296: Geological Survey of Western Australia, 4p.

Nelson, DR 2000k, 142946: foliated biotite tonalite, Flat Rocks, Geochronology Record 297: Geological Survey of Western Australia, 4p.

Nelson, DR 2000l, 142948: tonalite, Flat Rocks, Geochronology Record 298: Geological Survey of Western Australia, 4p.

Nelson, DR 2000m, 142949: metasandstone, Whim Creek, Geochronology Record 299: Geological Survey of Western Australia, 4p.

Nelson, DR 2000n, 142950: porphyritic biotite monzogranite, OT Well, Geochronology Record 300: Geological Survey of Western Australia, 4p.

Nelson, DR 2000o, 142951: sandstone, Lalla Rookh Well, Geochronology Record 301: Geological Survey of Western Australia, 5p.

Nelson, DR 2000p, 142965: monzogranite, Unices Well, Geochronology Record 305: Geological Survey of Western Australia, 4p.

Nelson, DR 2000q, 142967: biotite monzogranite, Unices Well, Geochronology Record 307: Geological Survey of Western Australia, 3p.

Nelson, DR 2000r, 160442: foliated biotite monzogranite, Yandeyarra Pool, Geochronology Record 239: Geological Survey of Western Australia, 4p.

Nelson, DR 2001a, 160498: hornblende–biotite granodiorite, Geemas Well, Geochronology Record 240: Geological Survey of Western Australia, 4p.

Nelson, DR 2001b, 160727: biotite granodiorite, Chillerina Well, Geochronology Record 241: Geological Survey of Western Australia, 3p.

Nelson, DR 2001c, 160728: biotite monzogranite, Chillerina Well, Geochronology Record 243: Geological Survey of Western Australia, 3p.

Nelson, DR 2001d, 160730: foliated biotite granodiorite, MalindraWell, Geochronology Record 244: Geological Survey of Western Australia, 3p.

Nelson, DR 2001e, 160744: foliated biotite-hornblende granodiorite, Chillerina Well; Geochronology Record 227: Geological Survey of Western Australia, 3p.

Nelson, DR 2001f, 160745: foliated biotite granodiorite, old Marble Bar Road – Tabba Tabba Creek crossing; Geochronology Record 213: Geological Survey of Western Australia, 3p.

Nelson, DR 2001g, 168924: volcaniclastic sedimentary rock, Whim Creek Hotel, Geochronology Record 203: Geological Survey of Western Australia, 3p.

Nelson, DR 2001h, 168932: porphyritic granodiorite, Mardeburra Pool, Geochronology Record 204: Geological Survey of Western Australia, 4p.

Nelson, DR 2001i, 168934: biotite monzogranite, Waloo Waloo Pool, Geochronology Record 205: Geological Survey of Western Australia, 4p.

Nelson, DR 2001j, 168936: monzodiorite, Andover Mine, Geochronology Record 207: Geological Survey of Western Australia, 3p.

Nelson, DR 2002a, 169014: foliated biotite-hornblende quartz diorite, Mount Gratwick, Geochronology Record 139: Geological Survey of Western Australia, 4p.

Nelson, DR 2002b, 169016: foliated biotite quartz diorite, Mount Gratwick, Geochronology Record 140: Geological Survey of Western Australia, 3p.

Nelson, DR 2002c, 169018: biotite monzogranite, Cunmagnunna Hill, Geochronology Record 141: Geological Survey of Western Australia, 3p.

Nelson, DR 2002d, 169021: leucocratic syenogranite gneiss, Birthday Gift Mine, Geochronology Record 143: Geological Survey of Western Australia, 3p.

Nelson, DR 2002e, 169025: rhyolite, Knaptons Well, Geochronology Record 145: Geological Survey of Western Australia, 3p.

Nelson, DR 2004, 169080: quartz diorite, Terenar Pool, Geochronology Record 115: Geological Survey of Western Australia, 4p.

Nelson, DR 2005, 178049: biotite monzogranite, Bamboo Springs Homestead, Geochronology Record 568: Geological Survey of Western Australia, 4p.

Nelson, DR, Trendall, AF and Altermann, W 1999, Chronological correlations between the Pilbara and Kaapvaal cratons: Precambrian Research, v. 97, p. 165–189.

Nelson, DR, Trendall, AF, de Laeter, JR, Grobler, NJ and Fletcher, IR 1992, A comparative study of the geochemical and isotopic systematics of late Archaean flood basalts from the Pilbara and Kaapvaal Cratons: Precambrian Research, v. 54, p. 231–256.

Nijman, W, de Bruijne, CH, and Valkering, ME 1998, Growth fault control of early Archaean cherts and barite mounds and chert-barite veins, North Pole Dome, Eastern Pilbara, Western Australia: Precambrian Research, v. 88, p. 25–52.

Nisbet, EG and Chinner, GA 1981, Controls on the eruption of mafic and ultramafic lavas: Ruth Well Cu–Ni prospect, western Pilbara: Economic Geology, v. 76, p. 1729–1735.

Ohta, H, Maruyama, S, Takahashi, E, Watanabe, Y and Kato, Y 1996, Field occurrence, geochemistry and petrogenesis of the Archaean Mid-Oceanic Ridge Basalts (AMORBs) of the Cleaverville area, Pilbara Craton, Western Australia: Lithos, v. 37, p. 199–221.

Oversby, VM 1976, Isotopic ages and geochemistry of Archaean acid igneous rocks from the Pilbara, Western Australia: Geochimica et Cosmochimica Acta, v. 40, p. 817–829.

Pease, V, Percival, J, Smithies, H, Stevens, G and Van Kranendonk, M 2008, When did plate tectonics begin? Evidence from the orogenic record, *in* When Did Plate Tectonics Begin on Planet Earth? *edited by* KC Condie and V Pease: Geological Society of America, Special Paper 440, p. 199–228.

Pickard, AL 2003, SHRIMP U–Pb zircon ages for the Palaeoproterozoic Kuruman Iron Formation, Northern Cape Province, South Africa: evidence for simultaneous BIF deposition in the Kaapvaal and Pilbara Cratons: Precambrian Research, v. 125, p. 275–315.

Pike, G 2001, The facies architecture of two contrasting volcanosedimentary basin successions from the Archaean Whim Creek Belt, North Pilbara Terrain, Western Australia: The intra-continental arc-related Whim Creek Group and plume-related, continental rift-hosted Bookingarra Group: Monash University, Melbourne, Australia, PhD thesis (unpublished).

Pike, G and Cas, RAF 2002, Stratigraphic evolution of Archaean volcanic rock-dominated rift basins from the Whim Creek Belt, west Pilbara Craton, Western Australia, *in* Precambrian Sedimentary Environments: A Modern Approach to Depositional Systems *edited by* W Altermann and P Corcoran: International Association of Sedimentologists, Special Publication 33, Blackwell Science, Oxford, UK, p. 213–34.

Pike, G, Cas, RAF and Hickman, AH 2006, Archean volcanic and sedimentary rocks of the Whim Creek greenstone belt, Pilbara Craton, Western Australia: Geological Survey of Western Australia, Report 101, 104p.

Pike, G, Cas, RAF and Smithies, RH 2002, Geological constraints on base metal mineralization of the Whim Creek greenstone belt, Pilbara Craton, Western Australia: Economic Geology, v. 97, p. 827–845.

Pirajno, F 2007, Ancient to modern Earth: the role of mantle plumes in the making of continental crust, *in* Earth's Oldest Rocks *edited by* MJ Van Kranendonk, RH Smithies and VC Bennett: Developments in Precambrian Geology 15, p. 1037–1064.

Platina Resources Limited 2006, Prospectus: Australian Securities Exchange, 2006, 108p.

Polymetals Mining Limited 2013, Turner River Gold Project: Mineral Resources increased to 345 koz Gold: Announcement to the Australian Securities Exchange, 13 March 2013.

Powell, CMcA and Horwitz, RC 1994, Late Archaean and Early proterozoics and basin formation of the Hamersley Ranges: Geological Society of Australia (WA Division); 12th Australian Geological Convention, Excursion Guidebook 4, 53p.

Range River Gold Limited 2005, Indee Gold Project to be developed: report to Australian Securities Exchange, 16 September 2005.

Rasmussen, B, Blake, TS and Fletcher, IR 2005, U–Pb zircon age constraints on the Hamersley spherule beds: Evidence for a single 2.63 Ga Jeerinah–Carawine impact ejecta layer: Geology, v. 33, p. 725–728.

Reading, AM, Tkalcic, H, Kennett, BLN, Johnson, SP and Sheppard, S 2012, Seismic structure of the crust and uppermost mantle of the Capricorn and Paterson Orogens and adjacent cratons, Western Australia, from passive seismic transects: Precambrian Research, v. 196–197, p. 295–308.

Reber, JE, Dabrowski, M, Galland, O and Schmid, DW 2013, Sheath fold morphology in simple shear: Journal of Structural Geology, v. 53, p. 15–26.

Richards, JR, Fletcher, IR and Blockley, J 1981, Pilbara galenas: precise assay of the oldest Australia leads; model ages and growth-curve implications: Mineralium Deposita, v. 16, p. 7–30.

Roberts, DL 1974, Geology and copper-zinc mineralization of the Whundo area, West Pilbara Goldfield, Western Australia: The University of Western Australia, BSc (Hons) thesis, unpublished.

Ruddock, I 1999, Mineral occurrences and exploration potential of the west Pilbara: Geological Survey of Western Australia, Report 70, 63p.

Ryan, GR 1964, A reappraisal of the Archaean of the Pilbara Block: Geological Survey of Western Australia, Annual Report 1963, p. 25–28.

Ryan, GR 1965, The geology of the Pilbara Block: Australasian Institute of Mining and Metallurgy, Proceedings, v. 214, p. 61–94.

Ryan, GR and Kriewaldt, MJB 1964, Facies changes in the Archaean of the West Pilbara Goldfield: Geological Survey of Western Australia, Annual Report 1963, p. 28.

Segue Resources Limited 2013, Pardoo Exploration Update: Australian Securities Exchange Announcement, 10 September 2013.

Shibuya, T, Kitajima, K, Komiya, T, Terabayashi, M and Maruyama, S 2007, Middle Archean ocean ridge hydrothermal metamorphism and alteration recorded in the Cleaverville area, Pilbara Craton, Western Australia: Journal of Metamorphic Geology, v. 25, p. 751–767.

Shibuya, T, Tahata, M, Kitajima, K, Ueno, Y, Komiya, T, Yamamoto, S, Igisu, M, Terabayashi, M, Sawaki, Y, Takai, K, Yoshida, N and Maruyama, S 2012, Depth variation of carbon and oxygen isotopes of calcites in Archean altered upper oceanic crust: Implications for the CO_2 flux from ocean to oceanic crust in the Archean: Earth and Planetary Science Letters, 321–322, p. 64–73.

Simonson, BM, Sumner, DY, Beukes, NJ, Johnson, S and Gutzmer, J 2009, Correlating multiple Neoarchean-Paleoproterozoic impact spherule layers between South Africa and Western Australia: Precambrian Research, v. 169, p. 100–111.

Smith, JB 2003, The episodic development of intermediate to silicic volcano–plutonic suites in the Archaean West Pilbara, Australia: Chemical Geology, v. 194, p. 275–295.

Smith, JB, Barley, ME, Groves, DI, Krapež, B, McNaughton, NJ, Bickle, MJ and Chapman, HJ 1998, The Sholl Shear Zone, West Pilbara; evidence for a domain boundary structure from integrated tectonostratigraphic analysis, SHRIMP U–Pb dating and isotopic and geochemical data of granitoids: Precambrian Research, v. 88, p. 143–172.

Smithies, RH 1996, Refinement of the stratigraphy of the Whim Creek belt, Pilbara granite–greenstone terrain: new evidence from the Sherlock 1:100 000 sheet: Geological Survey of Western Australia, Annual Review 1995–96, p. 118–123.

Smithies, RH 1997, The Mallina Formation, Constantine Sandstone and Whim Creek Group: a new stratigraphic and tectonic interpretation for part of the western Pilbara Craton: Geological Survey of Western Australia, Annual Review 1996–97, p. 83–88.

Smithies, RH 1998a, Geology of the Sherlock 1:100 000 sheet: Geological Survey of Western Australia, 1:100 000 Geological Series Explanatory Notes, 29p.

Smithies, RH 1998b, Geology of the Mount Wohler 1:100 000 sheet: Geological Survey of Western Australia, 1:100 000 Geological Series Explanatory Notes, 19p.

Smithies, RH 1999, Geology of the Yule 1:100 000 sheet: Geological Survey of Western Australia, 1:100 000 Geological Series Explanatory Notes, 15p.

Smithies, RH 2002a, Archaean boninite-like rocks in an intracratonic setting: Earth and Planetary Science Letters, v. 197, p. 19–34.

Smithies, RH 2002b, De Grey, WA Sheet 2757 (Version 1.0): Geological Survey of Western Australia, 1:100 000 Series.

Smithies, RH 2003, Geology of the White Springs 1:100 000 sheet: Geological Survey of Western Australia, 1:100 000 Geological Series Explanatory Notes, 16p.

Smithies, RH 2004, Geology of the De Grey and Pardoo 1:100 000 sheets: Geological Survey of Western Australia, 1:100 000 Geological Series Explanatory Notes, 24p.

Smithies, RH and Champion, DC 1998, Secular compositional changes in Archaean granitoid rocks of the west Pilbara: Geological Survey of Western Australia, Annual Review 1997–98, p. 71–76.

Smithies, RH and Champion, DC 2000, The Archaean high-Mg diorite suite: links to tonalite–trondhjemite–granodiorite magmatism and implications for Early Archaean crustal growth: Journal of Petrology, v. 41, p. 1653–1671.

Smithies, RH and Champion, DC 2002, Assembly of a composite granite intrusion at a releasing bend in an active Archaean shear zone: Geological Survey of Western Australia, Annual Review 2000–01, p. 63–68.

Smithies, RH, Champion, DC and Blewett, RS 2001a, Wallaringa, WA Sheet 2656: Geological Survey of Western Australia, 1:100 000 Geological Series.

Smithies, RH, Champion, DC and Blewett, RS 2002, Geology of the Wallaringa 1:100 000 sheet: Geological Survey of Western Australia, 1:100 000 Geological Series Explanatory Notes, 27p.

Smithies, RH, Champion, DC and Cassidy, KF 2003, Formation of Earth's early Archaean continental crust: Precambrian Research, v. 127, p. 89–101.

Smithies, RH, Champion, DC and Sun, S-S 2004a, Evidence for Early LREE-enriched Mantle Source Regions: Diverse Magmas from the ca. 3.0 Ga Mallina Basin, Pilbara Craton, NW Australia: Journal of Petrology, v. 45, p. 1515–1537.

Smithies, RH, Champion, DC, Van Kranendonk, MJ and Hickman, AH 2007a, Geochemistry of volcanic units of the northern Pilbara Craton: Geological Survey of Western Australia, Report 104, 47p.

Smithies, RH, Champion, DC and Van Kranendonk, MJ 2007b, The oldest well-preserved volcanic rocks on Earth: geochemical clues to the early evolution of the Pilbara Supergroup and implications for the growth of a Paleoarchean continent, *in* Earth's Oldest Rocks *edited by* MJ Van Kranendonk, RH Smithies and VC Bennett: Developments in Precambrian Geology 15, p. 339–367.

Smithies, RH, Champion, DC, Van Kranendonk, MJ, Howard, HM and Hickman, AH 2005a, Modern-style subduction processes in the Mesoarchaean: geochemical evidence from the 3.12 Ga Whundo intraoceanic arc: Earth and Planetary Science Letters, v. 231, p. 221–237.

Smithies, RH and Farrell, T 2000, Geology of the Satirist 1:100 000 sheet: Geological Survey of Western Australia, 1:100 000 Geological Series Explanatory Notes, 42p.

Smithies, RH, Hickman, AH and Nelson, DR 1999, New constraints on the evolution of the Mallina Basin, and their bearing on relationships between the contrasting eastern and western granite–greenstone terrains of the Archaean Pilbara Craton, Western Australia: Precambrian Research, v. 94, p. 11–28.

Smithies, RH and Hickman, AH 2003, Archaean geology of the Mallina and Whim Creek basins (1:250 000 scale): Geological Survey of Western Australia.

Smithies, RH and Hickman, AH 2004, Pyramid WA Sheet SF 50-7 (2nd edition): Western Australia Geological Survey, 1:250 000 Geological Series.

Smithies, RH, Nelson, DR and Pike, G 2001b, Detrital and inherited zircon age distributions – implications for the evolution of the Archaean Mallina Basin, Pilbara Craton, northwestern Australia: Sedimentary Geology, v. 141–142, p. 79–94.

Smithies, RH, Van Kranendonk, MJ and Champion, DC 2007c, The Mesoarchaean emergence of modern style subduction, *in* Island Arcs: Past and Present *edited by* S Maruyama and M Santosh: Gondwana Research, v. 11, p. 50–68.

Smithies, RH, Van Kranendonk, MJ and Hickman, AH 2004b, De Grey, WA Sheet 2757 (Version 2.0): Geological Survey of Western Australia, 1:100 000 Series.

Sossi, PA, Eggins, SM, Nesbitt, RW, Nebel, O, Hergt, JM, Campbell, IH, St.C. O'Neill, H, Van Kranendonk, MJ and Rhodri Davies, D 2016, Petrogenesis and Geochemistry of Archean Komatiites: Journal of Petrology, v. 57, p. 147–184, doi:10.1093/petrology/egw004.

Strong, CA, Hickman, AH, and Kojan, CJ 2000, Preston, WA Sheet 2156: Geological Survey of Western Australia, 1:100 000 Geological Series.

Sugitani, K, Horiuchi, Y, Adachi, M and Sugisaki, R 1996, Anomalously low Al_2O_3/TiO_2 values for Archean cherts from the Pilbara Block, Western Australia – possible evidence for extensive chemical weathering on the early Earth: Precambrian Research, v. 80, 49–76.

Sugitani, K, Yamamoto, K, Adachi, M, Kawabe, I and Sugisaki, R 1998, Archaean cherts derived from chemical, biogenic, and clastic sedimentation in a shallow restricted basin: examples from the Gorge Creek Group in the Pilbara Block: Sedimentology, v. 45, p. 1045–1062.

Sugitani, K, Mimura, K, Suzuki, K, Nagamine, K and Sugisaki, R 2003, Stratigraphy and sedimentary petrology of an Archean volcanic-sedimentary succession at Mount Goldsworthy in the Pilbara Block, Western Australia: implications of evaporite (nahcolite) and barite deposition: Precambrian Research, v. 120, 55–79.

Sun, S-S and Hickman, AH 1998, New Nd-isotopic and geochemical data from the west Pilbara — implications for Archaean crustal accretion and shear zone development: Australian Geological Survey Organisation, Research Newsletter, no. 28, p. 25–29.

Sun, S-S and Hickman, AH 1999, Geochemical characteristics of ca 3.0-Ga Cleaverville greenstones and later mafic dykes, west Pilbara: implication for Archaea crustal accretion: Australian Geological Survey Organisation, Research Newsletter, no. 31, p. 23–29.

Sun, S-S and Hoatson, DM 1992, Chemical and isotopic characteristics of parent magmas of the west Pilbara layered intrusions: implications for petrogenesis, magma mixing, and PGE mineralization in layered intrusions: Australian Geological Survey Organisation, Bulletin 242, p. 141–149.

Sun, S-S and Mc Donough, WF 1989, Chemical and isotopic systematics of oceanic basalts: implications for mantle compositions and processes, *in* Magmatism In Ocean Basins *edited by* AD Saunders and MJ Norry: Geological Society of London, Special Publication, v. 42, p. 313–345.

Sun, S-S, Wallace, DA, Hoatson, DM, Glikson, AY and Keays, RR 1991, Use of geochemistry as a guide to platinum group element potential of mafic–ultramafic rocks: Examples from the West Pilbara and Halls Creek mobile zone, Western Australia: Precambrian Research, v. 50, p. 1–35.

Sweetapple, MT and Collins, PLF 2002, Genetic framework for classification and distribution of Archean rare metal pegmatites in the North Pilbara Craton, Western Australia: Economic Geology, v. 97, p. 873–895.

Sylvester, GC and de Laeter, JR 1987, Geochronology of the Mons Cupri Archean volcanic centre, Pilbara Block, Western Australia: Journal of the Royal Society of Western Australia, v. 70, Part 2, p. 29–34.

Taylor, T 1985, De Grey Project – Mt. Grant EL 45/50, Pilbara Mineral Field, WA: Sedimentological appraisal of the Amphitheatre Gold Prospect: Geological Survey of Western Australia, Statuary mineral exploration report, A15836 (unpublished), 21p.

Thorne, AM and Trendall, AF 2001, The geology of the Fortescue Group, Pilbara Craton, Western Australia: Geological Survey of Western Australia, Bulletin 144, 249p.

Thorne, AM, Tyler, IM, Korsch, RJ, Johnson, SP, Brett, JW, Cutten, HN, Blay, O, Kennett, BLN, Blewett, RS, Joly, A, Dentith, MC, Aitken, ARA, Holzschuh, J, Goodwin, JA, Salmon, M, Reading, A and Boren, G 2011, Preliminary interpretation of deep seismic reflection line 10GA–CP1: crustal architecture of the northern Capricorn Orogen, *in* Capricorn Orogen seismic and magnetotelluric (MT) workshop 2011: extended abstracts *edited by* SP Johnson, AM Thorne and IM Tyler: Geological Survey of Western Australia, Record 2011/25, p. 19–26.

Thorpe, RI, Hickman, AH, Davis, DW, Mortensen, JK and Trendall, AF 1992, Constraints to models for Archaean lead evolution from precise U–Pb geochronology from the Marble Bar region, Pilbara Craton, Western Australia, *in* The Archaean: Terrains, processes and metallogeny *edited by* JE Glover and SE Ho: The University of Western Australia, Geology Department and University Extension, Publication no. 22, p. 395–408.

Tomich, BNV 1974, The geology and nickel mineralisation of the Ruth Well area, Western Australia: The University of Western Australia, BSc (Hons) thesis (unpublished).

Trendall, AF 1968, Three great basins of Precambrian iron formation deposition: a systematic comparison: Geological Society of America Bulletin 79, p. 1527–1544.

Trendall, AF 1991, Progress report on the stratigraphy and structure of the Fortescue Group in the Gregory Range area of the eastern Pilbara Craton: Geological Survey of Western Australia, Record 1990/10.

Van Kranendonk, MJ 2000, Geology of the North Shaw 1:100 000 sheet: Geological Survey of Western Australia, 1:100 000 Geological Series Explanatory Notes, 86p.

Van Kranendonk, MJ 2003, Stratigraphic and tectonic significance of eight local unconformities in the Fortescue Group, Pear Creek Centrocline, Pilbara Craton, Western Australia: Geological Survey of Western Australia, Annual Review 2001–02, p. 70–79.

Van Kranendonk, MJ 2006, Volcanic degassing, hydrothermal circulation and the flourishing of early life on Earth: a review of the evidence from the c. 3490–3240 Ma rocks of the Pilbara Supergroup, Pilbara Craton, Western Australia: Earth-Science Reviews, v. 74, p. 197–240.

Van Kranendonk, MJ 2008, Structural geology of the central part of the Lalla Rookh – Western Shaw structural corridor, Pilbara Craton, Western Australia: Geological Survey of Western Australia, Report 103, 29p.

Van Kranendonk, MJ and Collins, WJ 1998, Timing and tectonic significance of Late Archaean, sinistral strike-slip deformation in the Central Pilbara Structural Corridor, Pilbara Craton, Western Australia: Precambrian Research, v. 88, p. 207–232.

Van Kranendonk, MJ, Hickman, AH, Smithies, RH, Nelson, DR and Pike, G 2002, Geology and tectonic evolution of the Archean North Pilbara Terrain, Pilbara Craton, Western Australia: Economic Geology, v. 97, p. 695–732.

Van Kranendonk, MJ, Hickman, AH, Smithies, RH, Williams, IR, Bagas, L and Farrell, TR 2006, Revised lithostratigraphy of Archean supracrustal and intrusive rocks in the northern Pilbara Craton, Western Australia: Geological Survey of Western Australia, Record 2006/15, 57p.

Van Kranendonk, MJ, Smithies, RH, Hickman, AH, Bagas, L, Williams, IR, and Farrell, TR 2004, Event stratigraphy applied to 700 million years of Archaean crustal evolution, Pilbara Craton, Western Australia: Geological Survey of Western Australia, Annual Review 2003–04, p. 49–61.

Van Kranendonk, MJ, Smithies, RH, Hickman, AH and Champion, DC 2007a, Paleoarchean Development of a Continental Nucleus: the East Pilbara Terrane of the Pilbara Craton, *in* Earth's Oldest Rocks, *edited by* MJ Van Kranendonk, RH Smithies and VC Bennett: Developments in Precambrian Geology 15, p. 307–337.

Van Kranendonk, MJ, Smithies, RH, Hickman, AH and Champion, DC 2007b, Secular tectonic evolution of Archaean continental crust: interplay between horizontal and vertical processes: Terra Nova, v. 19, p. 1–38.

Van Kranendonk, MJ, Smithies, RH, Hickman, AH, Wingate, MTD and Bodorkos, S 2010, Evidence for Mesoarchean (~3.2 Ga) rifting of the Pilbara Craton: the missing link in an early Precambrian Wilson cycle: Precambrian Research, v. 177, p. 145–161.

Van Kranendonk, MJ, Smithies, RH, Griffin, WF, Huston, DL, Hickman, AH, Champion, DC, Anhaeusser, CR and Pirajno, F 2015, Making it thick: A volcanic plateau origin of Paleoarchean continental lithosphere of the Pilbara and Kaapvaal cratons, *in* Continent Formation Through Time *edited by* NMW Roberts, MJ Van Kranendonk, S Parman, S Shirley and PD Clift: Geological Society, London, Special Publications, v. 389, p. 83–111.

Vearncombe, S, Vearncombe, JR and Barley, ME 1998, Fault and stratigraphic controls on volcanogenic massive sulphide deposits in the Strelley belt, Pilbara Craton, Western Australia: Precambrian Research, v. 88, p. 67–82.

Venturex Resources Limited 2009, Venturex to acquire Whim Creek and Salt Creek base metal projects from Straits Resources: report to Australian Securities Exchange, 20 August 2009, 8p.

Venturex Resources Limited 2015, 2015 Annual Report: report to Australian Securities Exchange, 30 September 2015, 49p.

Wallace, DA 1992a, Andover mafic–ultramafic complex, West Pilbara, Western Australia: Chemical and petrographic data, and summary of geology: Australian Bureau of Mineral Resources, Geology and Geophysics, Record 1992/13, 29p.

Wallace, DA 1992b, Field relationships, petrography, mineralogy, and geochemistry of the Andover Complex, *in* Petrology and platinum-group element geochemistry of Archaean layered mafic–ultramafic intrusions, west Pilbara Block, Western Australia: Australian Geological Survey Organisation, Bulletin 242, p. 126–140.

Weinberg, RF, Moresi, L and van der Borgh, P 2003, Timing of deformation in the Norseman-Wiluna Belt, Yilgarn Craton, Western Australia: Precambrian Research, v. 120, p. 219–239.

Wilhelmij, HR and Dunlop, JSR 1984, A genetic stratigraphic investigation of the Gorge Creek Group in the Pilgangoora syncline, *in* Archaean and Proterozoic Basins of the Pilbara, Western Australia: evolution and mineralization potential *edited by* JR Muhling, DI Groves and TS Blake: The University of Western Australia Geology Department and University Extension, Publication no. 9, p. 68–88.

Williams, IR 1968, Geology of the Yarraloola 1:250 0000 sheet: Geological Survey of Western Australia, 1:250 000 Geological Series Explanatory Notes.

Williams, IR 1999, Geology of the Muccan 1:100 000 sheet: Geological Survey of Western Australia, 1:100 000 Geological Series Explanatory Notes, 39p.

Williams, IR and Tyler, IM 1991, Robertson WA: Geological Survey of Western Australia, 1: 250 000 Geological Series Explanatory Notes, 36p.

Wingate, MTD, Bodorkos, S and Van Kranendonk, MJ 2009a, 180039: felsic volcaniclastic rock, Wodgina; Geochronolohy Record 806: Geological Survey of Western Australia, 4p.

Wingate, MTD, Bodorkos, S and Van Kranendonk, MJ 2009b, 180048: felsic metavolcanic rock, Cattle Well; Geochronology Record 807: Geological Survey of Western Australia, 4p.

Wingate, MTD, Bodorkos, S and Van Kranendonk, MJ 2009c, 180098: welded rhyolite tuff, Hong Kong Mine; Geochronology Record 814: Geological Survey of Western Australia, 4p.

Wingate, MTD and Hickman, AH 2009a, 178148: biotite granite, Hearsons Cove; Geochronology Record 800: Geological Survey of Western Australia, 4p.

Wingate, MTD and Hickman, AH 2009b, 178164: leucogabbro, Bottom Bore; Geochronology Record 803: Geological Survey of Western Australia, 4p.

Wingate, MTD, Kirkland, CL, Bodorkos, S and Hickman, AH 2010, 160258: felsic metavolcanic rock, Orchard Well; Geochronology Record 840: Geological Survey of Western Australia, 4p.

Wingate, MTD, Kirkland, CL and Hickman, AH 2015 178170: leucogabbro, Little Fortune mine; Geochronology Record 1227: Geological Survey of Western Australia, 4p.

Yuan, H 2015, Secular change in Archaean crust formation recorded in Western Australia: Nature Geoscience, v. 8, p. 808–813.

Zegers, TE 1996, Structural, kinematic and metallogenic evolution selected domains of the Pilbara granitoid–greenstone terrain: Geologica Ultraiectina, Mededelingen van de Faculteit Aardwetenschappen, Universiteit Utrecht, No. 146, 208p.

Zegers, TE, de Wit, MJ, Dann, J and White, SH 1998b, Vaalbara, Earth's oldest assembled supercontinent? A combined structural, geochronological, and palaeomagnetic test: Terra Nova, v. 10, p. 250–259.

Zegers, TE, White, SH, de Keijzer, M and Dirks, P 1996, Extensional structures during deposition of the 3460 Ma Warrawoona Group in the eastern Pilbara Craton, Western Australia: Precambrian Research, v. 80, p. 89–105.

Zhao, G, Cawood, PA, Wilde, SA and Sun, M 2002, Review of global 2.1–1.8 Ga orogens: implications for a pre-Rhodinia supercontinent: Earth-Science Reviews, v. 59, p. 125–162.

Zhao, G, Sun, M, Wilde, SA and Li, S 2004, A Paleo-Mesoproterozoic supercontinent: assembly, growth and breakup: Earth-Science Reviews, v. 67, p. 91–123.

Zibra, I 2012, Syndeformational granite crystallisation along the Mount Magnet Greenstone Belt, Yilgarn Craton: evidence of large-scale magma-driven strain localisation during Neoarchean time: Australian Journal of Earth Sciences, v. 59, p. 793–806.

Appendix 1

Gazetteer of place names

Locality	*Latitude*	*Longitude*
Ant Hill	21°29'22"S	116°45'57"E
Ayshia mine	21°04'00"S	116°56'08"E
Balla Balla	20°41'40"S	117°47'19"E
Baynton Hill	20°53'11"S	116°40'10"E
Black Hill Well	20°51'28"S	117°02'48"E
Bore Creek	20°33'53"S	118°38'20"E
Bradley Well	20°57'31"S	117°21'47"E
Bullen Hill	20°34'21"S	118°34'44"E
Byong Creek	20°56'42"S	116°29'58"E
Cape Preston	20°49'57"S	116°12'21"E
Cape Thouin	20°20'10"S	118°10'55"E
Carlow Castle mine	20°48'34"S	117°03'42"E
Comstock mine	20°53'07"S	117°48'09"E
Cleaverville	20°39'05"S	117°01'59"E
Croydon Homestead	21°07'13"S	117°48'24"E
Dampier	20°39'46"S	116°42'23"E
Depuch Island	20°38'00"S	117°43'11"E
Devil Creek	20°54'59"S	116°25'16"E
Dixon Island	20°37'45"S	117°03'40"E
Egina mine	21°05'59"S	118°13'19"E
Elizabeth Hill mine	21°05'23"S	116°52'30"E
Evelyn mine	21°05'49"S	117°50'44"E
Fish Creek	21°06'51"S	117°11'11"E
Fortune mine	20°49'39"S	117°04'32"E
Goldsworthy	20°20'31"S	119°31'15"E
Good Luck mine	20°49'16"S	117°04'36"E
Good Luck Well	20°56'42"S	117°40'47"E
Hearson Cove	20°37'03"S	117°47'35"E
Highway prospect	20°15'53"S	119°33'41"E
Hill Well	20°55'48"S	117°45'27"E
Hong Kong mining area	21°11'21"S	118°17'07"E
Karratha	20°43'50"S	116°51'26"E
Karratha Homestead	20°53'04"S	116°40'18"E
Lilly Blanche mine	20°46'27"S	117°06'53"E
Lower Nickol mining area	20°44'44"S	116°58'16"E
Maitland River Bridge	20°51'03"S	116°36'39"E
Mallina Homestead	20°52'58"S	118°02'22"E
Mardeburra Pool	21°07'27"S	117°11'21"E
Millindinna Hill	21°02'02"S	118°07'31"E
Mons Cupri	20°52'49"S	117°48'25"E
Mount Ada	20°55'22"S	117°09'36"E
Mount Dove	20°56'16"S	118°27'35"E
Mount Fisher	20°55'01"S	117°27'43"E
Mount Hall	20°47'20"S	117°11'56"E
Mount Leopold	20°59'02"S	116°36'54"E

Locality	*Latitude*	*Longitude*
Mount Marie	20°48'49"S	116°59'59"E
Mount Negri	20°47'09"S	117°50'46"E
Mount Oscar	20°54'45"S	117°18'38"E
Mount Regal	20°49'00"S	116°45'00"E
Mount Satirist	21°05'16"S	118°08'21"E
Mount Sholl	20°56'16"S	116°55'09"E
Mount Wangee	20°44'02"S	117°12'45"E
Munni Munni prospect	21°07'05"S	116°49'46"E
Nickol River	20°54'08"S	116°57'37"E
Nunyerry Gap	21°29'24"S	117°55'41"E
Old Cherratta Homestead	21°00'47"S	116°48'18"E
Opaline Well	20°58'34"S	117°42'44"E
Ord Ranges	20°17'54"S	119°09'17"E
Peawah Hill	20°38'26"S	117°56'07"E
Pippingarra	20°26'02"S	118°45'14"E
Point Samson	20°37'35"S	117°11'48"E
Port Hedland	20°18'37"S	118°36'03"E
Quamby mine	21°03'49"S	117°51'18"E
Radio Hill mine	20°59'04"S	116°52'10"E
Red Hill	20°57'49"S	117°31'39"E
Rocky Creek	20°41'31"S	117°02'39"E
Ruth Well	20°51'15"S	116°51'12"E
Salt Creek prospect	20°45'25"S	117°42'16"E
Sherlock Bay prospect	20°48'45"S	117°32'40"E
Sherlock Homestead	20°53'43"S	117°38'40"E
Sherlock River	20°56'42"S	117°36'50"E
Sing Well	20°50'49"S	116°58'21"E
Station Peak	21°09'30"S	118°10'59"E
Teichman's gold mine	21°16'16"S	118°12'15"E
Terenar Pool	20°49'06"S	117°27'43"E
Three Kings prospect	20°54'34"S	118°31'09"E
Tom Well	20°46'58"S	116°49'51"E
Toorare Pool	20°58'55"S	116°46'43"E
Toweranna mine	20°58'38"S	117°52'45"E
Tozer Well	20°58'05"S	116°57'08"E
Waloo Waloo Pool	21°13'23"S	117°05'39"E
Warambie Homestead	20°56'52"S	117°22'20"E
Weerianna mine	20°45'42"S	117°06'38"E
Weerianna Hill	20°45'56"S	117°07'11"E
West Whundo mine	21°04'45"S	116°55'26"E
Whim Creek mine	20°50'48"S	117°49'53"E
Whundo mine	20°04'45"S	116°55'39"E
Whyjabby Pool	20°53'56"S	116°40'40"E
Wickham	20°40'43"S	117°08'17"E
Wingina Well	20°50'34"S	118°33'52"E
Woodbrook	20°54'31"S	117°07'07"E
Yannery Hill mine	21°004'23"S	116°55'58"E
Yanyare River Bridge	20°52'25"S	116°31'02"E
Zebra Hill	21°11'35"S	116°50'30"E

Appendix 2

Definitions of new and revised lithostratigraphic names

Weerianna Basalt (new name)

Derivation of name

Weerianna Hill (Lat. 20°45’56” S, Long. 117°07’11” E), 2 km west of Roebourne

Distribution

From Mount Wangee, through Roebourne, southwest to Nickol River, outcropping over an area of 100 km^2

Stratigraphic thickness

Approximately 1000 m at Weerianna Hill and Mount Wangee, becoming thinner southwest towards Nickol River

Southwest thinning is attributed to deformation beneath the Regal Thrust

Type area

Weerianna Hill

Lithology

Metamorphosed basalt, including fine-grained komatiitic basalt

Relationships

Formation of the Roebourne Group and conformably overlies the Ruth Well Formation

Overlain, across a faulted unconformity, by the c. 3220 Ma Nickol River Formation

Intruded by irregular sills and dykes of gabbro and dolerite related to the c. 3015 Ma Andover Intrusion

Intruded by rhyolite and dacite sills and stocks of the c. 3015 Ma Orpheus Supersuite

Age

Minimum age c. 3270 Ma

The Ruth Well Formation, conformably underlying the Weerianna Basalt, is intruded by the 3270–3261 Ma Karratha Granodiorite. Maximum age c. 3330 Ma based on unpublished Nd isotopic data from the Ruth Well Formation

Bookingarra Group (redefinition)

Derivation of name

Bookingarra Creek (Lat. 20°55’47” S, Long. 117°45’20” E), 12 km southwest of Whim Creek mine

Unit name history

The name Bookingarra Group was first used between 2001 and 2006 to formally name a succession of five formations that unconformably overlies the Whim Creek Group (Pike and Cas, 2002). Van Kranendonk et al. (2006) included these five formations in the Croydon Group, based on an interpretation of stratigraphic continuity between the successions of the Whim Creek greenstone belt and the adjacent Mallina Basin. This reinterpretation is no longer accepted because structural and geochronological data indicate that the units correlated have different ages and are located on opposite sides of a c. 2955 Ma regional unconformity.

The name ‘Bookingarra Formation’ used by Van Kranendonk et al. (2006) to refer only to the succession of the Louden Volcanics and Mount Negri Volcanics is now obsolete.

Constituent units

The Bookingarra Group comprises five formations:

Kialrah Rhyolite (youngest)

Mount Negri Volcanics

Louden Volcanics

Rushall Slate

Cistern Formation (oldest)

Parent unit

De Grey Supergroup

Distribution

Exposed along the entire 110 km length of the Whim Creek greenstone belt from Mount Ada, 15 km south of Roebourne to Peawah River, 20 km northeast of Whim Creek mine

The total outcrop area is approximately 500 km^2 and the depositional extent exceeded 2000 km^2.

The group is also the main constituent of the poorly exposed c. 300 km^2 Peawah Hill greenstone belt.

Stratigraphic thickness

Maximum thickness approximately 3000 m in the eastern half of the Whim Creek greenstone belt

Type area

The type area is Bookingarra Creek to Mount Negri, 5 km northeast of Whim Creek mine

No single type section can be nominated for the group

Lithology

Predominantly composed of volcanic and subvolcanic intrusive rocks. Volcanic rocks include komatiite, komatiitic basalt, tholeiitic basalt, andesite, dacite, and rhyolite. Subvolcanic sills of peridotite, pyroxenite, gabbro, and dolerite are common and there are local stocks of dacite and rhyolite. Metamorphosed volcaniclastic conglomerate, sandstone, siltstone, shale, and minor chert are present in the lower part of the group, but also form wedge-shaped units at higher stratigraphic levels.

Relationships

The Bookingarra Group unconformably overlies the 3009–2991 Ma Whim Creek Group across a c. 2955 Ma regional unconformity, and the group is overlain across a c. 2775 Ma regional unconformity by the 2775–2629 Ma Fortescue Group. The Bookingarra Group is separated from the Croydon Group by the Loudens Fault, a major strike-slip fault that juxtaposed the depositional basins of the two groups.

Age

A maximum depositional age of c. 2955 Ma is indicated by the interpreted age of the regional unconformity that separates the Bookingarra Group from the underlying Whim Creek Group (Pike and Cas, 2002). Additionally, the Cistern Formation has a maximum isotopic depositional age of 2964 ± 6 Ma (Huston et al., 2002), and the true depositional age is likely to be younger because several near-concordant detrital zircons were dated between c. 2962 and c. 2956 Ma. The minimum age of the Louden Volcanics is constrained by SHRIMP zircon U–Pb dating of the overlying Kialrah Rhyolite at 2943 ± 7 Ma (Nelson, 1998).

Depositional environment

The Bookingarra Group was deposited in an extensional tectonic setting immediately following orogenic deformation at c. 2955 Ma (D_6 in this Report). Clastic and volcaniclastic rocks of the Cistern Formation and Rushall Slate were deposited adjacent to local volcanic centres, closely contemporaneous with earliest granitic intrusions of the Sisters Supersuite. Rifting of the depositional basin continued during basaltic volcanism of the Louden and Mount Negri Volcanics.

Based on eruption of komatiites and komatiitic basalts, several workers have interpreted the Louden and Mount Negri Volcanics to be products of a mantle plume. The Kialrah Rhyolite is the same age as c. 2940 Ma granites (Sisters Supersuite) that intruded the Mallina Basin, and rhyolite of the same age within the Mallina Basin suggests that by c. 2940 Ma the basins of the Bookingarra and Croydon Groups were close together.

Louden Volcanics (redefinition)

Derivation of name

Louden Creek (Lat. 20°52’40” S, Long. 117°50’20” E), 4 km south of Whim Creek mine

Synonymy, unit name history

First defined as Louden Volcanics (Hickman, 1983), subsequently renamed as Louden Volcanic Member (Van Kranendonk et al., 2006)

Parent unit

Bookingarra Group

Distribution

Exposed along the entire 110 km length of the Whim Creek greenstone belt from Mount Ada, 15 km south of Roebourne to Peawah River, 20 km northeast of Whim Creek mine. The total outcrop area is approximately 400 km^2 and the depositional extent exceeded 2000 km^2. Based mainly on geophysical data, the formation extends an additional 70 km northeast of the merger of the Loudens Fault with the Sholl Shear Zone (SSZ), along the northern side of the SSZ almost to Cape Thouin.

Stratigraphic thickness

Maximum stratigraphic thickness approximately 2000 m (Hickman, 1983), but regionally variable due to syndepositional normal faulting and local erosion prior to deposition of the Mount Negri Volcanics

Type area

The type area is 6 km south-southwest of Mons Cupri (Hickman, 1983).

Lithology

Ultramafic and mafic volcanic rocks, including komatiite, komatiitic basalt, pillowed and massive tholeiite, and local lenses of conglomerate and sandstone

Relationships

Where the Louden and Mount Negri Volcanics are exposed in close proximity, the Louden Volcanics are overlain by the Mount Negri Volcanics. Both these formations overlie the Rushall Slate and Cistern Formation, and the contacts are locally unconformable. The Louden Volcanics are overlain by the Kialrah Rhyolite, parts of which may intrude the Louden Volcanics. The Louden Volcanics are unconformably overlain by the c. 2775 Ma Mount Roe Basalt of the Fortescue Group.

Age

The maximum depositional age of the Louden Volcanics is c. 2955 Ma and the true age is probably c. 2950 Ma. The minimum age is constrained by SHRIMP zircon U–Pb dating of the overlying Kialrah Rhyolite, which was dated at 2943 ± 7 Ma in the Whim Creek greenstone belt (Nelson, 1998v).

Mount Negri Volcanics (redefinition)

Derivation of name

Mount Negri (Lat. 20°47’09” S, Long. 117°50’46” E)

Synonymy, unit name history

The name Mount Negri Volcanics was informally used by Miller and Gair (1975). Previously Fitton et al. (1975) included units now separated as the Louden Volcanics and Mount Negri Volcanics within the ‘Negri Volcanics’. The name Negri Volcanics was first formally defined by Hickman (1983) to refer only to the unit later renamed as the Mount Negri Volcanics (Hickman, 1990). More recently the Mount Negri Volcanics was renamed, but not formally redefined, as the Mount Negri Volcanic Member (Van Kranendonk et al., 2006).

Parent unit

Bookingarra Group

Distribution

Generally unconformably overlying older stratigraphic units in the Whim Creek greenstone belt, the Mount Negri Volcanics is exposed in a number of outliers, most notably at Mount Negri and between Mons Cupri and Hill Well.

Stratigraphic thickness

The maximum stratigraphic thickness of the formation is approximately 500 m (Hickman, 1983), but in most areas the thickness is between 100 and 200 m.

Type area

Mount Negri

Lithology

The Mount Negri Volcanics are predominantly composed of variolitic and vesicular tholeiitic basalt, although pyroxene-spinifex textures are locally present.

Relationships

The Mount Negri Volcanics overlies the Louden Volcanics, locally apparently unconformably. However, most contacts between the formations are faulted. North of Whim Creek in the type area around Mount Negri, the Mount Negri Volcanics unconformably overlies the Rushall Slate and Cistern Formation. Nowhere are the Mount Negri Volcanics in stratigraphic contact with the c. 2775 Ma Mount Roe Basalt.

Age

The maximum depositional age of the Mount Negri Volcanics is c. 2955 Ma. The minimum age of the formation is less well constrained, but in the type area of Mount Negri it contains Pb mineralization dated at c. 2922 Ma (data from Thorpe et al., 1992 interpreted by Huston et al., 2002).

References

Fitton, MJ, Horwitz, RC and Sylvester, G 1975, Stratigraphy of the early Precambrian in the west Pilbara, Western Australia: Aust. CSIRO Miner. Res. Lab., Invest. Rep, 41p.

Hickman, AH 1983, Geology of the Pilbara Block and its environs: Geological Survey of Western Australia, Bulletin 127, 268p.

Hickman, AH 1990, Geology of the Pilbara Craton, *in* Excursion Guidebook No. 5, Pilbara and Hamersley Basin, Third International Archaean Symposium Perth W.A. 1990 *edited by* SE Ho, JE Glover, JS Myers and JR Muhling: The University of Western Australia Geology Department and University Extension, Publication 21, p. 2–13.

Huston, DL, Sun, S-S, Blewett, R, Hickman, AH, Van Kranendonk, MJ, Phillips, D, Baker, D and Brauhart, C 2002, The timing of mineralization in the Archean North Pilbara Terrain, Western Australia: Economic Geology, v. 97, p. 733–755.

Miller, LJ and Gair, HS 1975, Mons Cupri copper–lead–zinc deposit, *in* Economic geology of Australia and Papua New Guinea — metals *edited by* CL Knight: Australasian Institute of Mining and Metallurgy, Monograph 5, p. 195–202.

Nelson, DR 1998, 144261: rhyolite, Bradley Well; Geochronology Record 272: Geological Survey of Western Australia, 4p.

Pike, G and Cas, R 2002, Stratigraphic evolution of Archaean volcanic rock-dominated rift basins from the Whim Creek Belt, west Pilbara Craton, Western Australia, *in* Precambrian sedimentary environments: a modern approach to ancient depositional systems *edited by* W Altermann and P Corcoran: Blackwell Science, Oxford, United Kingdom, International Association of Sedimentologists Special publication 33, p. 213–234.

Thorpe, R, Hickman, AH, Davis, DW, Mortensen, JK and Trendall, AF 1992, Constraints to models for Archaean lead evolution from precise U-Pb geochronology from the Marble Bar region, Pilbara Craton, Western Australia, *in* The Archaean: Terrains, processes and metallogeny *edited by* JE Glover and SE Ho: Geology Department and University Extension, The University of Western Australia, Perth, Western Australia, Publication 22, p. 395–408.

Van Kranendonk, MJ, Hickman, AH, Smithies, RH, Williams, IR, Bagas, L and Farrell, TR 2006, Revised lithostratigraphy of Archean supracrustal and intrusive rocks in the northern Pilbara Craton, Western Australia: Geological Survey of Western Australia, Record 2006/15, 57p.

Appendix 3

Previous tectonic models

Several tectonic models have been applied to the northern Pilbara Craton since regional-scale geological investigations first commenced in the 1950s. This Report presents a tectonic interpretation of the northwest Pilbara Craton that, although it incorporates evidence from previous work, is largely based on information from the Pilbara Mapping Project (see Introduction). Some of the main features of previous tectonic models are not consistent with current data.

Geosynclinal and rift models (pre-1978)

Prior to the advent of the plate tectonic theory in the 1970s the sedimentary and volcanic successions of Earth's Phanerozoic orogenic belts were usually attributed to deposition in geosynclines – subsiding linear depositional basins on or at the margins of continental crust (Knopf, 1948). Therefore, when the first detailed geological investigations of Archean granite–greenstone terranes commenced in the 1950s, most tectonic interpretations were strongly influenced by previously accepted geosynclinal models. In most of these early models, the greenstones were interpreted as metamorphosed ophiolite, molasse, or flysch successions, deposited in troughs in approximately the same positions as the currently preserved greenstone belts (Anhaeusser et al., 1968, 1969; Glikson, 1970, 1971, 1972; Anhaeusser, 1971a, 1971 b; Stowe, 1971). Some envisaged underlying continental crust, whereas others did not (Glikson and Lambert, 1976; Glikson, 1977, 1979).

Reconnaissance geological mapping of granite–greenstones of the northwest Pilbara in the 1960s provided stratigraphic interpretations based on the 'geosynclinal cycle' (Ryan and Kriewaldt, 1964; Kriewaldt, 1964; Ryan, 1964, 1965). The Mallina Basin was initially regarded as a northeast-trending geosynclinal trough, containing a 12 km-thick sedimentary succession. Influenced by the geosynclinal model, Ryan (1965) interpreted all volcanic successions northwest and southeast of the basin (now assigned to the Whim Creek, Whundo, Roebourne and Soanesville Groups, the Regal Formation, and the Pilbara Supergroup) to be the same age as the sedimentary rocks within the basin. Miller (1975) provided much the same type of geosynclinal interpretation and interpreted the volcanic and granitic rocks of the northwest coastal area to be parts of a volcanic arc northwest of the Mallina trough. These stratigraphic interpretations, which are now superseded, resulted from applying a pre-existing tectonic model to the northwest Pilbara. Subsequent work has revealed that the formations of the northwest Pilbara were deposited over approximately 400 million years.

Volcano-sedimentary plateau model (1975–93)

GSWA remapping of the east Pilbara between 1972 and 1975 was undertaken with the aim of providing detailed lithological maps to assist mineral exploration that was extremely active in 1972. The resulting geological maps and various geological observations made during the mapping were subsequently used to deduce the area's stratigraphy and crustal evolution (Hickman, 1975, 1977, 1981, 1983). This established a new tectonic model that was substantially different from models being applied to other Archean granite–greenstone terranes (Hickman, 1983). Previous models for the evolution of granite–greenstone terranes that were rejected for the east Pilbara included:

- deposition of the greenstone successions as oceanic crust (Glikson, 1972)
- greenstone derivation from volcanism within large asteroid impact craters (Green, 1972)
- deposition of supracrustal rocks in separate rift zones (Windley, 1973)
- Phanerozoic-style plate tectonic models (Goodwin and Ridler, 1970).

Gravitational subsidence of east Pilbara greenstone belts relative to adjacent granitic crust (a feature of the geosynclinal model) had taken place in the east Pilbara, but stratigraphic continuity between the different greenstone belts established that the greenstone succession had not been deposited in separate basins. The conclusion was that the present separation of the greenstone belts by areas of granite was simply the result of deep erosion of the dome-and-syncline structures. Although there were lateral changes in the succession due to internal unconformities and removal of lower sections by granitic intrusion, the greenstone stratigraphy was found to be essentially continuous around and across adjacent granitic domes (Hickman, 1975, 1984). Present thicknesses of the successions in different belts varied between 10 and 30 km (Hickman, 1980c) and prior to deformation the greenstone succession formed a volcano-sedimentary platform or plateau (Hickman, 1981). Based on field observations on exposures of granitic gneiss in the domes, it was interpreted that these rocks had a more complex structural history than the greenstones, and are therefore probably included in the pre-greenstone basement (Hickman, 1983). On the other hand, the domes were known to include large volumes of granite that had intruded the greenstone succession. Geochronology supported the interpretation that the domes were composed of granitic intrusions of various ages from c. 3490 to 2840 Ma (Hickman, 1984, Fig. 2).

Two main stratigraphic divisions of the east Pilbara greenstones were recognized in 1980:

- the Warrawoona Group, which is now equivalent to the Paleoarchean Pilbara Supergroup (Warrawoona, Kelly and Sulphur Springs Groups)
- the Gorge Creek Group, which is now equivalent to the Mesoarchean Soanesville and Gorge Creek Groups, and part of the Croydon Group.

A brief survey of the northwest Pilbara in 1976 was used to interpret an extension of most of the east Pilbara stratigraphy to the northwest (Hickman, 1980a,b, 1981, 1983, 1990, in agreement with Fitton et al., 1975). However, it was recognized that several important stratigraphic units of the northwest Pilbara were not present in the east: the Whim Creek Group, Louden Volcanics, and Negri Volcanics (Hickman, 1980b, 1980c).

Weaknesses of the regional stratigraphic interpretation in this tectonic model included:

- geochronological data were either too limited or too inaccurate to satisfactorily constrain stratigraphic correlations
- time constraints imposed by the mapping schedule meant that all observations were essentially of a reconnaissance nature. The style of geological mapping, along spaced traverses joined by air photo interpretation, did not permit detailed stratigraphic analysis.

Acquisition of greatly improved geochronology in the 1990s revealed that the majority of c. 1980 stratigraphic correlations across the east Pilbara had been substantially correct, whereas most correlations made between the east and northwest Pilbara were not correct.

Accretion model (1993–2002)

Studies of accretionary complexes along the Japanese margin of the Pacific Ocean (Taira et al., 1980, 1989, 1992; Taira, 1985; Maruyama and Seno, 1986; Matsuda and Isozaki, 1989; Isozaki et al., 1990; Isozaki and Maruyama, 1992) suggested a basis for re-examination of some of the world's Archean greenstone successions. Located close to the Pilbara coast, the Roebourne greenstone belt (Fig. 7) was selected to determine if its geology was consistent with that of an accretionary complex. Resulting studies concluded that the Roebourne greenstone comprises parts of an oceanic plate that collided with the Pilbara Craton at c. 3000 Ma (Isozaki et al., 1991; Kiyokawa and Taira, 1998; Kiyokawa, 1993, Ohta et al., 1996; Kato et al., 1998; Kiyokawa et al., 2002). Kiyokawa and his co-workers interpreted most of the Roebourne greenstone stratigraphy to be that of an oceanic island arc, whereas other workers (Ohta et al., 1996; Kato et al., 1998) interpreted it as MORB and trench-related sediments.

Arguably the most influential plate-tectonic interpretation was that made by Kiyokawa and Taira (1992, 1998) and Kiyokawa (1993). Their accretionary interpretation for the origin of the Roebourne greenstone belt was released when other researchers were considering if Phanerozoic-style plate-tectonic processes might explain the crustal evolution of the entire northern Pilbara Craton (Kimura et al., 1991; Krapez, 1993; Barley, 1997; Krapez and Eisenlohr, 1998; Smith et al., 1998).

Kiyokawa (1993) interpreted the Roebourne greenstone belt as a stack of nappes, containing six tectonostratigraphic units. He proposed that at c. 3200 Ma, a subduction zone off the northwest margin of the craton produced an oceanic island arc and that this eventually collided with the Pilbara Craton at c. 3050 Ma. Developing this model, Kiyokawa and Taira (1998) and Kiyokawa et al. (2002) interpreted the Roebourne greenstone belt to be an accretionary complex produced by thrusting of oceanic crust onto the 3270–3261 Ma Karratha Granodiorite. Several lines of evidence subsequently argued against this interpretation.

Regional evidence

The Roebourne greenstone belt is situated approximately 150 km inland from the northwest margin of the Pilbara Craton (Hickman, 2004) and the concealed offshore section of the craton is interpreted to be composed of the c. 3000 Ma Dampier and Harding Granitic Complexes and older greenstones. Onshore exposures of these granitic complexes have provided evidence that they include Paleoarchean crust (Tables 2 and 4). This presents a problem for the coastal accretion model because if the offshore section has a similar composition to the onshore section it would be necessary for the proposed c. 3000 Ma accretion complex to have been formed at least 150 km inland of the craton margin.

Structural evidence

Most of the lithological contacts interpreted to be thrusts within an accretionary complex (Kiyokawa and Taira, 1998) have been described as intrusive contacts by other workers (Hoatson et al., 1992; Wallace, 1992a, b; Krapez and Eisenlohr, 1998; Smith et al., 1998; Hoatson and Sun, 2002; Smith, 2003). Additionally, the 3270–3261 Ma Karratha Granodiorite, which in the accretion model is interpreted to be part of an exotic terrane pre-dating and unrelated to the later island arc succession, is actually intrusive into the Ruth Well Formation (part of the interpreted oceanic succession). At c. 2940 Ma the Roebourne Synform (Kriewaldt et al., 1964; Biggs, 1979; Hickman, 1983, 1997a,b; Powell and Horwitz, 1994) and Prinsep Dome folded the Regal Thrust, causing it to be structurally repeated across the Roebourne greenstone belt. In the accretion model, these structural repetitions of a single thrust plane were interpreted to be separate thrusts, giving the impression of a tectonically stacked succession.

Geochronology

All units within a single island arc assemblage should be approximately the same age. In discussing Archean plate tectonics, Komiya and Maruyama (2007) concluded that assuming the Archean mantle was 120–150 ºC hotter than the present-day mantle, oceanic plates would have been much smaller and probably had a life span of only 15 million years. However, the ages of the units interpreted by Kiyokawa and Taira (1998) to make up the inferred island arc assemblage range from c. 3280 to 3015 Ma. Additionally, an oceanic island arc succession should not contain detritus from older continental crust. However, at Cleaverville the c. 3020 Ma Cleaverville Formation contains detrital zircon of various ages between c. 3700 and 3022 Ma (Aihara et al., 2012; GSWA 127330, Nelson, 1998). The older zircon ages indicate derivation from Eoarchean–Paleoarchean crust of the Pilbara Craton. A c. 3108 Ma zircon group identified by Aihara et al. (2012) probably originated from the Sholl Terrane (ST) south of the Sholl Shear Zone (SSZ) after its deformation and uplift by the 3068–3066 Ma Prinsep Orogeny.

Geochemistry

The accretion model requires that the c. 3280 Ma Ruth Well Formation originated as oceanic crust. However, published Nd model ages for the Ruth Well Formation range from c. 3480 to 3420 Ma and ε_{Nd} values are between –1.92 and +0.88 (Table 4). The ε_{Nd} values are more negative than for Paleoarchean basaltic formations of the east Pilbara, which are known to have been deposited on continental crust (e.g. Euro Basalt, Smithies et al, 2007). Smithies et al. (2007) found that tholeiitic basalts of the Ruth Well Formation have Th concentrations up to 42 times that of primitive mantle values and that the LREE are fractionated, with La at ~20 times primitive mantle values. They noted that enrichments in Th, U, and LREE clearly distinguish the Ruth Well Formation from oceanic basaltic crust.

Plate-tectonic model applying sequence stratigraphy (1993–2002)

Research from 1983 onwards tested various interpretations reached in the volcano-sedimentary plateau model. Of concern to many workers was that the model did not explain the crustal evolution of the northern Pilbara Craton, using conventional plate-tectonic theory. One line of evidence suggesting that the northern Pilbara Craton might be composed of accreted terranes was provided when Landsat imagery indicated a number of northeast-trending lineaments across the northern Pilbara. Initially, these lineaments were interpreted to coincide with late-stage faults that had exerted very little influence over the early crustal evolution of the craton. However, one of the first applications of zircon U–Pb dating in the northwest Pilbara (Horwitz and Pidgeon, 1993) indicated that a felsic volcanic unit previously correlated with the Duffer Formation of the east Pilbara was approximately 300 million years too young for this correlation. Although this result involved only one of the previous east–west correlations, some workers used it as evidence that the entire lithostratigraphy of the northern Pilbara required re-examination. In the absence of adequate geochronology, Krapez (1993) proposed that an entirely new stratigraphic scheme should be adopted. His concept was that all the greenstone successions of the northern Pilbara Craton had evolved in plate-tectonic settings, and that their stratigraphy could be related to four megacycles of continental breakup and reassembly. Krapez (1993) suggested that each megacycle had produced a megasequence and that these spanned between 230 and 150 million years. In the absence of geochronological data on the successions, he argued that a scheme based on sequence stratigraphy would provide more reliable stratigraphic correlations. The sequence stratigraphy proposed by Krapez (1993) was subsequently revised by Krapez and Eisenlohr (1998). In this revision, the duration of the megacycles was changed to between 190 and 175 million years.

Issues for Archean sequence stratigraphy

Sequence stratigraphy was devised to assist interpretation of concealed stratigraphy in Phanerozoic sedimentary basins, particularly in combination with seismic profiles. It aimed to define unconformity-bound sequences and relate these to conceptual worldwide cycles of eustatic changes in sea level. These, in turn, were interpreted to be related to plate-tectonic processes, such as spreading at mid-ocean ridges and convergence of plate margins. Sequence stratigraphy is widely used in the Phanerozoic, but there are important issues questioning its application in the Archean where Phanerozoic-scale plate-tectonic processes are unproven. Some workers consider that plate tectonics could not have operated in the early Archean due to Archean Earth having a much hotter mantle than present-day Earth. Pilbara Paleoarchean and early Mesoarchean successions were predominantly volcanic, and no large sedimentary basins were formed until at least 3066 Ma. Paleoarchean evolution of the East Pilbara Terrane was governed by an irregular series of melting events in the mantle (Van Kranendonk et al., 2002, 2007a; Pirajno, 2007; Hickman, 2011, 2012). Whether produced by mantle plumes or crustal delamination, such events were relatively local. They were also irregular, commencing at intervals varying between 100 and 30 million years (i.e. 3530, 3480, 3450, 3350, and 3270 Ma). If inappropriate for the Paleoarchean successions of the East Pilbara Terrane, to what extent might sequence stratigraphy have application to the Mesoarchean successions of the northwest Pilbara?

The Mesoarchean crustal evolution of the northwest Pilbara involved plate-tectonic processes, but Archean plate size is interpreted to have been much smaller than in the Phanerozoic (Komiya and Maruyama, 2007; Van Kranendonk et al., 2007b). For example, breakup of the East Pilbara Terrane at c.3220 Ma is known to have produced relatively small plates. With far more individual plates in the Archean than in the Phanerozoic, any global periodicity of tectonic cycles is less likely. Geochronological evidence from the northwest Pilbara has established a sequence of events that does not match the megacycles proposed by Krapez and Eisenlohr (1998). Key events, such as the East Pilbara Terrane Rifting Event at c. 3220 Ma (taking place between deposition of the Roebourne Group and the Nickol River Formation) and the Prinsep Orogeny at 3068–3066 Ma (taking place between formation the ST and the Whim Creek Basin) fall within, not between, the proposed megacycles.

Westward growth model (1994–97)

Features of the plate-tectonic model (Krapez, 1993) and the accretion model (Kiyokawa, 1993) led to an interpretation that the northwest Pilbara Craton is composed of exotic terranes successively accreted onto the northwest margin of an older craton (Barley, 1997). In this ‘westward growth model’ it was suggested that the terranes were composed of progressively younger rocks from southeast to northwest. However, geochronological data obtained since 1997 indicates that the most northwesterly terrane (West Pilbara Superterrane [WPS]) contains stratigraphy up to 300 million years older than the Mallina Basin succession (Ruth Well Formation, c. 3280 Ma: Croydon Group, 3015–2931 Ma).

Wilson cycle model for the northwest Pilbara (2010)

Van Kranendonk et al. (2010) argued that the Mesoarchean evolution of the northern Pilbara Craton is consistent with a plate tectonic ‘Wilson cycle’. In one respect the 3066–2919 Ma evolution of the northwest Pilbara described in this Report differs from the interpretation of Van Kranendonk et al. (2010). They suggested that the Prinsep Orogeny marked complete accretion of the WPS to the East Pilbara Terrane (EPT) and that all subsequent events to c. 2919 Ma were the result of post-orogenic extensional collapse. However, this Report discusses evidence that final accretion of the WPS and EPT did not take place until the 2955–2919 Ma North Pilbara Orogeny. The 3068–3066 Ma Prinsep Orogeny was due to accretion of the Karratha Terrane (KT), Regal Terrane (RT), and ST, forming the WPS, but this left a belt of post-3200 Ma basaltic crust (Central Pilbara Basin [CPB] and Whundo Basin) between the WPS and the EPT. Amalgamation of the ST into the WPS prevented any further subduction by the Sholl subduction zone. As a result, continued northwest–southeast convergence between the Pilbara Craton and the plate inferred northwest of the craton was accommodated by formation of a new subduction zone northwest of the Pilbara Craton. Evidence for the 3023–2919 Ma existence of this northwest subduction zone is provided by the linear distribution of the Orpheus, Maitland River, and Sisters Supersuites, by the volcanic arc of the Whim Creek Group, and by the southeast migration of granitic intrusion until the close of the North Pilbara Orogeny at c. 2919 Ma.

References

Aihara, Y, Kiyokawa, S, Takehara, M and Horie, K 2012, Zircon U–Pb dating of the Cleaverville Formation, Pilbara, Australia: Australian Geoscience Council: 34th International Geological Congress: Brisbane, Queensland, 5–10 August 2012, Abstracts no. 2089.

Anhaeusser, CR 1971a, Cyclic volcanicity and sedimentation in the evolutionary development of Archaean greenstone belts of shield areas, *in* Symposium on Archaean Rocks (3rd edition) *edited by* JE Glover: Geological Society of Australia, Perth, Western Australia, Special Publication No. 3, p. 57–70.

Anhaeusser, CR 1971b, The Barbeston Mountain Land, South Africa — a guide to the understanding of the Archaean geology of Western Australia, *in* Symposium on Archaean Rocks (3rd edition) *edited by* JE Glover: Geological Society of Australia, Perth, Western Australia, Special Publication No. 3, p. 103–119.

Anhaeusser, CR, Roering, C, Viljoen, MJ and Viljoen, RP 1968, The Barberton Mountain Land: a model of the elements and evolution of an Archaean fold belt: Geological Society of South Africa, v. 71, p. 225–254.

Anhaeusser, CR, Mason, R, Viljoen, MJ and Viljoen, RP 1969, A reappraisal of some aspects of Precambrian shield geology: Geological Society of America Bulletin, v. 80, no. 11, p. 2175–2200.

Barley, ME 1987, The Archaean Whim Creek Belt, an ensialic fault-bounded basin in the Pilbara Block, Australia: Precambrian Research, v. 37, p. 199–215.

Barley, ME 1997, The Pilbara Craton, *in* Greenstone belts *edited by* MJ de Wit and L Ashwal: Oxford University Monographs on Geology and Geophysics, no. 35, p. 657–664.

Biggs, ER 1979, Karratha Urban Geology: 2356 II (first edition): Geological Survey of Western Australia.

Fitton, MJ, Horwitz, RC and Sylvester, G 1975, Stratigraphy of the Early Precambrian in the West Pilbara: CSIRO Minerals Research Laboratories, Division of Mineralogy, Report FP 11, 41p.

Glikson, AY 1970, Geosynclinal evolution and geochemical affinities of Early Precambrian systems: Tectonophysics, v. 9, p. 397–433.

Glikson, AY 1971, Archaean geosynclinal sedimentation near Kalgoorlie, Western Australia, *in* Symposium on Archaean Rocks *edited by* JE Glover: Geological Society of Australia, Special Publication No. 3, p. 443–460.

Glikson, AY 1972, Early Precambrian evidence of a primitive ocean crust and island nuclei of sodic granite: Geological Society of America Bulletin, v. 83, p. 3323–3344.

Glikson, AY 1977, On the basement of Canadian greenstone belts: Geoscience Canada, v. 5, p. 3–12.

Glikson, AY 1979, Early Precambrian tonalite-trondhjemite sialic nuclei: Earth Science Reviews, v. 15, p. 1–73.

Glikson, AY and Lambert, IB 1976, Vertical zonation and petrogenesis of the early Precambrian crust in Western Australia: Tectonophysics, v. 30, p. 55–89.

Goodwin, AM and Ridler, RH 1970, The Abitibi orogenic belt: Canada Geological Survey Paper, 70–40, p. 1–24.

Green, DH 1972, Archaean greenstone belts may include terrestrial equivalents of lunar maria? Earth and Planetary Science Letters, v. 15, p. 263–270.

Hickman, AH 1975, Precambrian structural geology of part of the Pilbara region: Geological Survey of Western Australia, Annual Report 1974, p. 68–73.

Hickman, AH 1977, Stratigraphic relations of rocks within the Whim Creek Greenstone Belt: Geological Survey of Western Australia Annual Report 1976, p. 53–56.

Hickman, AH 1980a, Archaean geology of the Pilbara Block: Excursion guide: Geological Society of Australia, W.A. Division, Second International Archaean Symposium, Perth, Western Australia, 1980, 55p.

Hickman, AH 1980b, Lithological and Stratigraphic Interpretation of the Pilbara Block (1:1 000 000 scale): Geological Survey of Western Australia, Bulletin 127, Plate 1.

Hickman, AH 1980c, Archaean stratigraphic successions in various parts of the Pilbara Block: Geological Survey of Western Australia, Bulletin 127, Plate 2.

Hickman, AH 1981, Crustal evolution of the Pilbara Block, *in* Archaean Geology, Second International Archaean Symposium, Perth, 1980 *edited by* JE Glover and DI Groves: Geological Society of Australia, Special Publication 7, Perth, Western Australia, p. 57–69.

Hickman, AH 1983, Geology of the Pilbara Block and its environs: Geological Survey of Western Australia, Bulletin 127, 268p.

Hickman, AH 1984, Archaean diapirism in the Pilbara Block, Western Australia, *in* Precambrian Tectonics Illustrated *edited by* A Kröner and R Greiling: E. Schweizerbarts'che Verlagsbuchhlung, Stuttgart, Germany, p. 113–27.

Hickman, AH 1990, Geology of the Pilbara Craton, *in* Excursion Guidebook No. 5, Pilbara and Hamersley Basin, Third International Archaean Symposium, Perth, Western Australia, 1990 *edited by* SE Ho, JE Glover, JS Myers and JR Muhling: The University of Western Australia Geology Department and University Extension, Publication 21, p. 2–13.

Hickman, AH 1997a, A revision of the stratigraphy of Archaean greenstone successions in the Roebourne–Whundo area, west Pilbara: Geological Survey of Western Australia Annual Review 1996–97, p. 76–82.

Hickman, AH 1997b, Dampier, WA Sheet 2256: Geological Survey of Western Australia, 1:100 000 Geological Series.

Hickman, AH 2004, Two contrasting granite–greenstones terranes in the Pilbara Craton, Australia: evidence for vertical and horizontal tectonic regimes prior to 2900 Ma: Precambrian Research, v. 131, p. 153–172.

Hickman, AH 2011, Pilbara Supergroup of the East Pilbara Terrane, Pilbara Craton: updated lithostratigraphy and comments on the influence of vertical tectonics: Geological Survey of Western Australia, Annual Review 2009–10, p. 50–59.

Hickman, AH 2012, Review of the Pilbara Craton and Fortescue Basin, Western Australia: crustal evolution providing environments for early life: Island Arc, v. 21, p. 1–31.

Hickman, AH, and Lipple, SL 1975, Explanatory notes on the Marble Bar 1:250 000 Geological Sheet, W.A.: Geological Survey of Western Australia, Record 1974/20, 24p.

Hoatson, DM and Sun, S-S 2002, Archean layered mafic–ultramafic intrusions in the west Pilbara Craton, Western Australia: A synthesis of some of the oldest orthomagmatic mineralizing systems in the world: Economic Geology, v. 97, p. 847–872.

Hoatson, DM, Wallace, DA, Sun, S-S, Macias, LF, Simpson, CJ and Keays, RR 1992, Petrology and platinum-group element geochemistry of Archaean layered mafic–ultramafic intrusions, west Pilbara Block, Western Australia: Australian Geological Survey Organisation, Bulletin 242, 319p.

Horwitz, R and Pidgeon, RT 1993, 3.1 Ga tuff from the Sholl Belt in the west Pilbara: further evidence for diachronous volcanism in the Pilbara Craton: Precambrian Research, v. 60, p. 175–183.

Isozaki, Y, Maruyama, S and Furuoka, F 1990, Accreted oceanic materials in Japan: Tectonophysics, v. 181, p. 179–205.

Isozaki, Y, Maruyama, S and Kimura, G 1991, Middle Archean (3.3 Ga) Cleaverville accretionary complex in Northwestern Pilbara Block, Western Australia: Eos, v. 72, p. 542.

Isozaki, Y and Maruyama, S 1992, Ocean plate stratigraphy: the prime criterion of accretionary complex: 29th International Geological Congress, Kyoto, Japan, August 24 – September 3 1992, Abstracts 2/3, p. 414.

Kato, Y, Ohta, I, Tsunematsu, T, Watanabe, Y, Isozaki, Y, Maruyama, S and Imai, N 1998, Rare earth element variations in mid-Archean banded iron formations: implications for the chemistry of ocean and continent plate tectonics: Geochimica et Cosmochimica Acta, v. 62, p. 3475–3497.

Kimura, G, Maruyama, S and Isozaki, I 1991, Early Archean accretionary complex in the eastern Pilbara Craton, Western Australia — detailed field occurrence in the Marble Bar area: Eos, v. 72, p. 542.

Kiyokawa, S 1993, Stratigraphy and structural evolution of a Middle Archaean greenstone belt, northwestern Pilbara Craton: University of Tokyo, Tokyo, Japan, PhD thesis (unpublished).

Kiyokawa, S and Taira, A 1992, Deformation of the upper oceanic crust in Archean – West Pilbara district of Western Australia: 29th International Geological Congress, Kyoto, Japan, 24 August – 3 September, Abstracts, v. 2, p. 461.

Kiyokawa, S and Taira, A, 1998, The Cleaverville Group in the west Pilbara coastal granite–greenstone terrane of Western Australia: an example of a mid-Archaean immature oceanic island-arc succession: Precambrian Research, v. 88, p. 102–142.

Kiyokawa, S, Taira, A, Byrne, T, Bowring, S and Sano, Y, 2002, Structural evolution of the middle Archaean coastal Pilbara terrane, Western Australia. Tectonics, v. 21, 1044, doi:10.1029/2001TC001296.

Knopf, A 1948, The Geosynclinal Theory: Bulletin of the Geological Society of America, v. 59, p. 649–670.

Komiya, T and Maruyama, S 2007, A very hydrous mantle under the western Pacific region: Implications for formation of marginal basins and style of Archean plate tectonics: Gondwana Research, v. 11, p. 132–147.

Krapez, B 1993, Sequence stratigraphy of the Archaean supracrustal belts of the Pilbara Block, Western Australia: Precambrian Research, v. 60, p. 1–45.

Krapez, B and Eisenlohr, B 1998, Tectonic settings of Archaean (3325–2775 Ma) crustal–supracrustal belts in the West Pilbara Block: Precambrian Research, v. 88, p. 173–205.

Kriewaldt, MJB 1964, The Fortescue Group of the Roebourne region, North-west Division: Geological Survey of Western Australia, Annual Report for 1963, p. 30–34.

Matsuda, T and Isozaki, Y 1989, Well-documented travel history of Mesozoic pelagic chert in Japan: from remote ocean to subduction zone: Tectonics, v. 10, p. 475–499.

Miller, LJ 1975, Volcanogenic CU–Pb-Zn mineralization in the Mons Cupri District, West Pilbara: Australasian Institute of Mining and Metallurgy, Western Australia Conference 1973, p. 95–96.

Nelson, DR 1998, 127330: volcaniclastic sedimentary rock, Cleaverville; Geochronology Record 442: Geological Survey of Western Australia, 4p.

Ohta, H, Maruyama, S, Takahashi, E, Watanabe, Y and Kato, Y 1996, Field occurrence, geochemistry and petrogenesis of the Archaean Mid-Oceanic Ridge Basalts (AMORBs) of the Cleaverville area, Pilbara Craton, Western Australia: Lithos, v. 37, p. 199–221.

Pirajno, F 2007, Ancient to modern Earth: the role of mantle plumes in the making of continental crust, *in* Earth's Oldest Rocks *edited by* MJ Van Kranendonk, RH Smithies and VC Bennett: Developments in Precambrian Geology 15, p. 1037–1064.

Powell, CMcA and Horwitz, RC 1994, Late Archaean and Early Proterozoic basin formation of the Hamersley Ranges: Geological Society of Australia (WA Division); 12th Australian Geological Convention, Excursion Guidebook 4, 53p.

Ryan, GR 1965, The geology of the Pilbara Block: Australasian Institute of Mining and Metallurgy, Proceedings, v. 214, p. 61–94.

Smith, JB, Barley, ME, Groves, DI, Krapež, B, McNaughton, NJ, Bickle, MJ and Chapman, HJ 1998, The Sholl Shear Zone, West Pilbara; evidence for a domain boundary structure from integrated tectonostratigraphic analysis, SHRIMP U–Pb dating and isotopic and geochemical data of granitoids: Precambrian Research, v. 88, p. 143–172.

Smithies, RH, Champion, DC, Van Kranendonk, MJ and Hickman, AH 2007, Geochemistry of volcanic units of the northern Pilbara Craton: Geological Survey of Western Australia, Report 104, 47p.

Stowe, CW 1971, Summary of the Tectonic Development of the Rhodesian Archaean Craton, *in* Symposium on Archaean Rocks *edited by* JE Glover: Geological Society of Australia, Special Publication No. 3, p. 377–383.

Taira, A, Katto, J, Tashiro, M and Okamura, M 1980, The geology of the Shimanto belt in Kochi Prefecture, Shikoku, *in* Geology and Palaeontology of the Shimanto Belt *edited by* A Taira and M Tashiro: Rinja-Kosaikai Press, Kochi, p. 319–389.

Taira, A 1985, Pre-Neogene accretion tectonics in Japan: a synthesis: Recent Progress of Natural Science in Japan, v. 10, p. 51–63.

Taira, A, Pickering, KT, Windley, BF and Soh, W 1992, Accretion of Japanese Island arcs and implications for the origin of Archean greenstone belts: Tectonics, v. 11, p. 1224–1244.

Taira, A, Tokuyama, H and Soh, W 1989, Accretion tectonics and evolution of Japan, in The Evolution of the Pacific Ocean Margins edited by Z Ben-Avraham: Oxford University Press, Monograph Geology and Geophysics 8, p. 100–123.

Van Kranendonk, MJ, Hickman, AH, Smithies, RH, Nelson, DR and Pike, G 2002, Geology and tectonic evolution of the Archean North Pilbara Terrain, Pilbara Craton, Western Australia: Economic Geology, v. 97, p. 695–732.

Van Kranendonk, MJ, Smithies, RH, Hickman, AH and Champion, DC 2007a, Paleoarchean Development of a Continental Nucleus: the East Pilbara Terrane of the Pilbara Craton, *in* Earth's Oldest Rocks *edited by* MJ Van Kranendonk, RH Smithies and VC Bennett: Developments in Precambrian Geology 15, p. 307–337.

Van Kranendonk, MJ, Smithies, RH, Hickman, AH and Champion, DC 2007b, Secular tectonic evolution of Archaean continental crust: interplay between horizontal and vertical processes: Terra Nova, v. 19, p. 1–38.

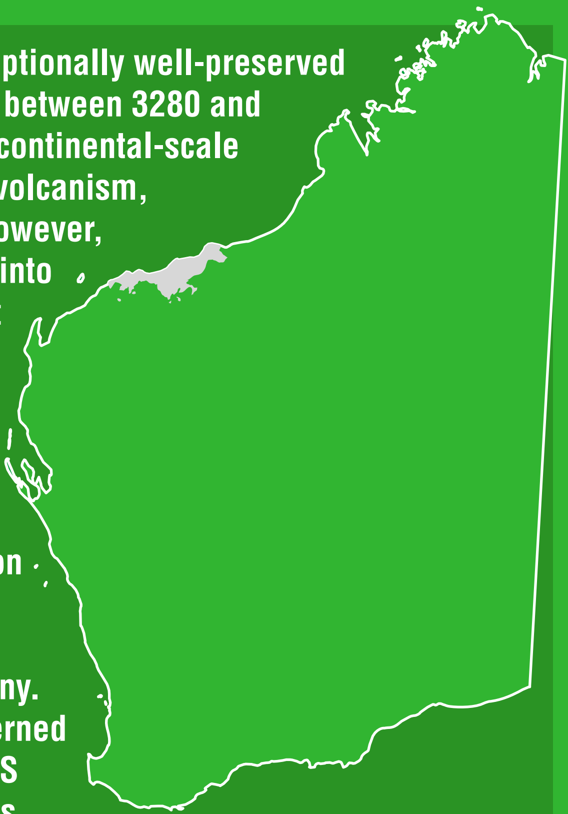
Van Kranendonk, MJ, Smithies, RH, Hickman, AH, Wingate, MTD and Bodorkos, S 2010, Evidence for Mesoarchean (~3.2 Ga) rifting of the Pilbara Craton: the missing link in an early Precambrian Wilson cycle: Precambrian Research, v. 177, p. 145–161.

Wallace, DA 1992a, Andover mafic–ultramafic complex, West Pilbara, Western Australia: Chemical and petrographic data, and summary of geology: Australian Bureau of Mineral Resources, Geology and Geophysics, Record 1992/13, 29p.

Wallace, DA 1992b, Field relationships , petrography, mineralogy, and geochemistry of the Andover Complex, *in* Petrology and platinum-group element geochemistry of Archaean layered mafic–ultramafic intrusions, west Pilbara Block, Western Australia *edited by* DM Hoatson et al.: Australian Geological Survey Organisation, Bulletin 242, p. 126–140.

Windley, BF 1973, Crustal development in the Precambrian: Philosophical Transactions of the Royal Society of London, A273, p. 321–341.

The northwest Pilbara Craton provides an exceptionally well-preserved geological record of Archean crustal evolution between 3280 and 2830 Ma. At 3280 Ma the craton was part of a continental-scale volcanic plateau dominated by plume-related volcanism, granitic intrusion, and vertical deformation. However, minor rifting of the plateau at c. 3235 evolved into deep rifting followed by continental breakup at c. 3220. One northeast trending zone of rifting and plate separation, now exposed as the Tabba Tabba Shear Zone (TTSZ), became a continental margin, the remnants of which now define the southeast boundary of the northwest Pilbara. Subsequent crustal evolution northwest of the TTSZ occurred through plate-tectonic processes including obduction, subduction, arc migration, and collision orogeny. Mineralization in the northwest Pilbara is governed by its plate tectonic history, and includes VHMS within Mesoarchean volcanic arcs, base metals within layered ultramafic–mafic intrusions, and lode gold along thrusts and strike-slip shear zones.



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