Age constraints on the timing of iron ore mineralisation in the southeastern Gawler Craton

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SHRIMP U–Pb zircon data obtained from a magnetite gneiss deposit in the southeast Gawler Craton indicate that the protoliths were deposited between 1750 and 1735 Ma. Previously, these iron-rich gneisses were thought to have been a metamorphosed banded iron formation and part of the Neoarchean to early Paleoproterozoic Sleaford Complex. Detrital zircon data provide maximum depositional ages between 1765 Ma and 1740 Ma, with \( {e^{175}} \) values between -4.6 and -3.3. Metamorphic zircon ages are 1735–1725 Ma, indicating that the magnetite gneiss was formed during the 1730–1690 Ma Kimban Orogeny. The pelitic mineralogy of the magnetite gneiss and presence of detrital zircon demonstrate that the protoliths were not a pure chemical sediment. Correlation with the greenschist facies Price Metasediments in the southern Gawler Craton suggests a series of iron-bearing basins were developed prior to the onset of the Kimban Orogeny and suggest that metamorphism played a role in ore formation.

KEY WORDS: detrital zircon, Nd isotopes, iron ore, Gawler Craton, SHRIMP geochronology, magnetite.

INTRODUCTION

Most large magnetite deposits are formed through primary, magmatic processes, e.g. mafic large igneous provinces, Kiruna-type and lava flow deposits (Cliff et al. 1990; Hitzman et al. 1992; Groves et al. 2010); or secondary hydrothermal processes, e.g. hypogene banded iron formation (BIF) and skarn development (Klein 2005; Bekker et al. 2010; Duuring & Hagemann 2011, 2013). Comparatively few mag- netite deposits are known to have formed in metamorphic terranes, in which the iron predates metamorphism (Mucke & Amnor 1993; Mucke et al. 1996). The Warramboo magnetite deposit, in the southern Gawler Craton, South Australia (Figure 1), is one such example. The deposit consists of magnetite-rich pelitic gneisses as the ore, interleaved with barren felsic gneisses. A mineral resource of 3.7 billion tonnes at 16% Fe has been established with a 67% Fe beneficiation product (IronRoad 2013).

Regional aeromagnetic images of the southern Gawler Craton (Figure 1) show the eastern Eyre Peninsula is dominated by north–south-trending structural fabrics that parallel the Kalinjala Shear Zone and which formed during the ca 1730–1690 Ma Kimban Orogeny (Parker et al. 1993; Vassallo & Wilson 2002; Hand et al. 2007; Dutch & Hand 2009). In the northwestern Eyre Peninsula, the region around the Warramboo deposit is characterised by broadly east–west-trending structures that appear to be truncated by the north–south fabrics (Figure 1). This apparent structural discordance suggests that the structural history of Warramboo may pre-date the Kimban Orogeny. This interpretation was supported by earlier work based on a whole-rock Rb–Sr age of 2520 ± 163 Ma from the adjacent felsic gneisses (Webb et al. 1986) that proposed the Warramboo deposit to be a metamorphosed Archean BIF (CRA 1984; Daly & Fanning 1993). Thus, in published stratigraphic columns of the Gawler Craton, the Warramboo magnetite gneiss has been considered to be Neoarchean in age (Daly & Fanning 1993; Daly et al. 1998).

A typical metamorphosed Archean BIF that has not undergone any other type of secondary alteration has a total Fe content of ~20–40 wt% and generally contains few silicate minerals (Klein 2005). In contrast, the Warramboo magnetite gneiss is coarse-grained, has abundant garnet, quartz and aluminosilicate gangue minerals, and has a moderate total Fe content. This calls into question the classification of the Warramboo deposit as a metamorphosed BIF. The significance of the Archean Rb–Sr whole-rock age of the felsic gneiss to the Warramboo magnetite gneiss is also uncertain, since it only provides a constraint on the adjacent gneiss, and not the ore horizon. Furthermore, recent work in the southern Gawler Craton has indicated that there is greater stratigraphic complexity than originally recognised, as both Archean and Paleoproterozoic sequences have been identified within rocks
that were previously thought to be a single stratigraphic unit (Szpunar et al., 2011).

In order to place the deposit into the context of new constraints on the regional geology of the southern Gawler Craton, this study investigated the age and nature of the protoliths to the Warramboo magnetite gneiss through SHRIMP U–Pb zircon geochronology and whole-rock Sm–Nd isotopic composition. These

Figure 1 (a) Interpreted solid geology of the southern Gawler Craton, showing location of the Warramboo iron formation and the Price Metasediments. (b) TMI aeromagnetic image. White box corresponds to Figure 2.
results suggest that the Warramboo magnetite deposit consists of a Paleoproterozoic cover sequence overlying Archean–earliest Paleoproterozoic rocks of the Sleaford Complex. These have subsequently been deformed and metamorphosed during the Kimban Orogeny to form migmatitic magnetite-rich gneisses within a series of east–west-trending folds that are truncated by the dominant north–south Kimban trends.

GEOLOGICAL SETTING

The Gawler Craton preserves a record of Mesoproterozoic to Mesozoic geological history. In general, it is poorly outcropping, overlain by extensive Neoproterozoic to Mesozoic sedimentary basins, Cenozoic marine sediments and Neogene to Pleistocene eolian dune fields (Daly et al. 1998; Hand et al. 2007; Fraser et al. 2010; Reid & Hand 2012).

The oldest rocks in the craton are Mesoarchean ca 3250–3150 Ma granitic gneisses that outcrop in the southeastern Gawler Craton (Fraser et al. 2010). These gneisses occur within a shear zone-bounded tectonostratigraphic domain (Figure 1a) but may underlie much of the craton (Fraser et al. 2010).

The Neoarchean to earliest Paleoproterozoic domains of the Gawler Craton are the volcano-sedimentary Sleaford and Mulgathing complexes (Figure 1a). The southern domain, the Sleaford Complex, comprises sedimentary and volcanic units deposited during the period 2560–2480 Ma and the supracrustal intrusives of the Dutton Suite, including S-type granites, intruded between ca 2519–2420 Ma (Daly & Fanning 1993; Fanning 2002; Fanning et al. 2007; Dutch et al. 2008; Fraser & Neumann 2010; Jagodzinski et al. 2012). The Sleaford Complex was deformed and metamorphosed during the 2465–2420 Ma Sleaford Orogeny (Fanning 1997; Swain et al. 2005; Jagodzinski et al. 2012; Reid et al. 2014), with low pressure–high temperature conditions of 5–7.5 kbar and 770–810°C recorded in metapelites in the southern Gawler Craton (Dutch et al. 2010).

The next phase of geological activity began at ca 2000 Ma with the intrusion of felsic and mafic magmas, which are interpreted to signal the onset of extension that provided the accommodation space for deposition of Paleoproterozoic volcanoclastic sedimentary sequences across the Gawler Craton (Fanning et al. 1988, 2007). In the southern Gawler Craton, this included a series of sedimentary, volcanic and magmatic rocks, with the Darke Peak Group deposited at ca 1865 Ma (Hoek & Schaefer 1998; Reid et al. 2008). The Donington Suite was associated with the ca 1850–1847 Ma Corinian Orogeny, a cycle of shortening and high-grade metamorphism followed by extension (Reid et al. 2008). In the eastern Eyre Peninsula, the Cleve Group was deposited at ca 1790 Ma (Szpunar et al. 2011) and further east the rhyolitic Myola Volcanics (Szpunar & Fraser 2010). On the southern Eyre Peninsula a sequence of magnetite-bearing phyllitic schist, known as the Price Metasediments, was deposited at ca 1765 Ma and has been correlated with phyllites in the Cape Hunter region of Antarctica (Oliver & Fanning 1997). The ca 1755 Ma McGregor Volcanics (Fraser & Neumann 2010; Szpunar & Fraser 2010), ca 1763–1741 Ma Wallaroo Group (Fanning et al. 2007) and ca 1750 Ma Moonabie Formation (Fraser & Neumann 2010) were also deposited just prior to a major tectonothermal event, the 1730–1690 Ma Kimban Orogeny.

The Kimban Orogeny was a craton-wide event with metamorphic grades ranging from greenschist to granulite facies (Hoek & Schaefer 1998; Tong et al. 2004; Hand et al. 2007; Payne et al. 2008; Dutch & Hand 2009; Dutch et al. 2010). It has mostly been studied in the southern Gawler Craton where the Archean–Paleoproterozoic domains were extensively reworked within transpressional shear zones (Vassallo & Wilson 2002; Dutch et al.

![Figure 2](image-url) Aeromagnetic image of Warramboo magnetite gneiss with sample locations and cross-section line A–A'.
In the southern Gawler Craton, two phases of deformation associated with the Kimban Orogeny are recognised; ca 1730 Ma D1 deformation developed upright, open-to-tight, shallowly south-plunging folds with P–T estimates of 8–9 kbar and 820–850°C (Dutch et al. 2010), followed by D2 deformation at ca 1700–1690 Ma where subhorizontal east–west shortening produced vertical, planar high strain zones and isoclinal folds associated with high-temperature decompression to metamorphic conditions of < 6 kbar and 790–850°C (Dutch et al. 2008, 2010). In the southeastern Gawler Craton syn-Kimban magmatic rocks of the monzonitic and leucogranitic Moody Suite intrude the Hutchison Group metasediments (Schwartz 1999), and magmatic rocks also intrude rocks of the Sleaford Complex and ca 2000 Ma Miltaltie Gneiss (Fanning et al. 2007).

Following the Kimban Orogeny, the geological record of the Gawler Craton is dominated by magmatic processes with the intrusion of the ca 1680–1670 Ma Tunkillia Suite (Ferris et al. 2004), the ca 1620–1610 Ma St Peter Suite (Flint et al. 1990) and the ca 1600–1570 Ma Hiltaba Suite and Gawler Range Volcanics (Fanning et al. 1988; Blissett et al. 1993; Daly et al. 1998). The early Mesoproterozoic in the Gawler Craton is also associated with a significant Iron Oxide Copper Gold (IOCG) province in the eastern Gawler Craton, including the Olympic Dam Cu–Au–U–REE deposit (Skirrow et al. 2002, 2007). Neoproterozoic rocks in the southern Gawler Craton are confined to the Polda Basin, in a narrow east–west-trending graben that extends for more than 350 km on the central Eyre Peninsula (Figure 1a), and consist of mixed clastics, evaporites and volcanics of the Kilroo Formation and Blue Range Beds (Preiss et al. 1993). The Polda Basin and younger Phanerozoic sediments likely contribute to the subdued magnetic response in the central Eyre Peninsula (Figure 1b).

**IRON MINERALISATION IN THE GAWLER CRATON**

Several periods of iron mineralisation have been recognised within the Gawler Craton (Davies 2000). The BIFs of the ca 2565 Ma Lower Middleback Jaspilite within the Middleback Range of northeastern Eyre Peninsula (Figure 1) are the oldest known iron formations in the Gawler Craton (Szpunar & Fraser 2010; Szpunar et al. 2011). A second period of BIF deposition at ca 2560–2470 Ma is recognised within the Christie Gneiss (Figure 1), where metamorphosed BIF and calc-silicate units are interleaved with migmatised metasedimentary gneiss of the Mulgathing Complex (Daly et al. 1978; Daly & Fanning 1993; Jagodzinski et al. 2009; Reid et al. 2014). Paleoproterozoic iron formations outcrop at Willigena Hill with a depositional age range of ca 1790–1740 Ma (Davies 2000).

**Warramboo Iron Deposit**

The Warramboo deposit is located in the central Eyre Peninsula, South Australia (Figure 1), and has a current mineral resource of 3.7 billion tonnes at 16% Fe (Iron-Road 2010). The magnetite gneiss is identified by a prominent high response on aeromagnetic imagery with east–west-trending fabrics and an isoclinal fold structure that appears to close to the east (Figure 2). The structures at Warramboo are markedly different from the dominant north–south-trending structures associated with the Kimban Orogeny that appear to truncate the east–west fabrics at Warramboo (Figure 1).

Iron mineralisation consists of coarse-grained magnetite and hematite grains intergrown with silicate minerals within a heterogeneous package of migmatitic gneisses. The major lithologies identified in the deposit can be divided into magnetite-poor host rocks and the magnetite-rich lithologies that constitute the ore horizon.

Warramboo iron-poor lithologies consist of:

1. Garnet–cordierite–biotite–quartz–feldspar migmatitic gneiss. Garnets are commonly porphyroblastic, up to 3 cm in size and surrounded by moats of 5–10 mm cordierite and feldspar within a gneissic fabric defined by biotite and quartz (Figure 4a). The gneissic fabric is generally high strain with boudins of low strain, weakly foliated, gneiss intersected in several drillholes.

2. K-feldspar–plagioclase–quartz–biotite±garnet migmatitic gneiss. This gneiss is dominated by quartz, K-feldspar augens and plagioclase with biotite, rare 1 mm garnets and minor magnetite (Figure 4b, c).

3. K-feldspar–cordierite–garnet weakly foliated gneiss. Porphyroblastic garnets (2–4 cm) are partially to completely replaced by cordierite+biotite within a weakly foliated K-feldspar matrix (Figure 4d).

The nature of the protoliths to the iron-poor lithologies is not always clear. Lithology one above is interpreted to be sedimentary in origin based on the aluminous composition, compositional heterogeneity and mineral textures. Lithology two is interpreted to be igneous in origin based on the dominantly felsic compositions and tabular K-feldspar phenocrysts. The interpretation of lithology three is less certain. This unit is generally more aluminous than lithology two but still contains a dominantly K-feldspar–plagioclase–quartz mineralogy. The tabular morphology of the K-feldspar phenocrysts and homogenous nature of the unit suggests it is derived from an igneous protolith, possibly an S-type granite similar to those in the Dutton Suite (Dutch et al. 2008). Warramboo magnetite-bearing lithologies can be divided into three groups

1. Magnetite±hematite–biotite–K-feldspar–quartz migmatitic gneiss. Abundant 5 mm magnetite±specular hematite is disseminated within a biotite–quartz matrix with coarse-grained 20 mm magnetite mineralisation developed adjacent to K-feldspar–quartz leucosomes. Magnetite–hematite–biotite is strongly foliated with deformed layer parallel leucosomes (Figure 4e).

2. Magnetite–garnet–sillimanite–cordierite–biotite–K-feldspar–quartz gneiss. Magnetite occurs as discrete 2–5 mm grains disseminated within a foliated matrix of garnet–cordierite–sillimanite–biotite. Sillimanite and cordierite are abundant. Garnets are euhedral 1 mm grains and very abundant. Foliation is defined by aligned biotite and sillimanite and
decimetre-scale variation of K-feldspar, biotite and sillimanite mineral abundance (Figure 4f).

(3) Magnetite–hematite–biotite–quartz–K-feldspar gneiss. Magnetite–hematite–biotite banding is abundant throughout with aligned grains defining a strong foliation with minor quartz and feldspar. Shear bands and S–C fabrics can be well developed within this unit (Figure 4g).

All magnetite-bearing lithologies have deformed K-feldspar–quartz leucosomes, varying abundances of specular hematite and orange–brown spessartine garnets; in contrast, magnetite-poor lithologies have large 30–50 mm pyrope–almandine garnets. We interpret the magnetite gneisses to be derived from sedimentary protoliths based on the aluminosilicate mineralogy and the heterogeneous nature of the magnetite gneisses in drill core (Figure 4e–g). This is consistent with the interpretation of Whitehead (1978), who described the Warramboo magnetite gneisses as most likely to be sedimentary in origin, with no evidence that they were derived from chemically precipitated sediments, although later work favoured a BIF interpretation (CRA 1984; Daly & Fanning 1993). Other minor lithologies include metamorphosed amphibolite mafics, calc-silicates, and flat lying mafic intrusives.

The mineralogy and migmatitic nature of the magnetite gneisses indicate the deposit has been metamorphosed to upper amphibolite or granulite facies, however there are no published metamorphic P–T constraints.

Iron Road geochemistry (H. Pearce pers. comm. 2014) shows that the magnetite gneiss can be both manganese-rich and manganese-poor, and this compositional difference within the magnetite gneiss is also broadly parallel to the contact between the magnetite gneiss and the barren gneissic host rock in cross-sections of the deposit (Figure 3). These zones of Mn enrichment correlate across the deposit and likely represent a primary compositional layering (S0), possibly stratigraphic layering, within the magnetite gneiss. This primary layering is broadly parallel to the gneissic layering itself, suggesting the gneissosity is an S1 foliation developed parallel to the original S0 layering. This gneiss foliation dips moderately to gently south-southeast (Figures 2, 3). Small-scale fold hinges in drill core plunge shallowly to the east and west with a great circle fit to the fold hinges of ~20°/165° (Figure 3; H. Pearce pers. comm. 2014). Based on the geochemical layering, S1 gneissic foliation, and Iron Road structural logging data, it is apparent that both the primary layering and gneissic foliation dip between 40 and 50° south about an apparent isoclinal fold axis with a plunge to the west-southwest. In the upper sections of the deposit, magnetite-poor and magnetite-rich gneisses are interlayered on a decametre scale with layers of magnetite gneiss discontinuously interlayered with magnetite-poor gneiss, suggesting that zones of lithological or structural complexity exist within the deposit.

**GEOCHRONOLOGICAL SAMPLING RATIONALE**

We sampled magnetite-poor host gneiss and magnetite-rich ore gneiss within the Warramboo deposit in order to constrain the age of the protolith and investigate the nature of any stratigraphic or structural relationships.
Figure 4 Photomicrographs of selected samples. (a) Sample 2058320, leucocratic garnet-bearing quartz–feldspar–biotite–cordierite metasedimentary gneiss, similar to 2058314. (b) Sample 2058315, foliated quartz–feldspar–biotite granitic gneiss with rare garnet. (c) Sample 2058316, foliated quartz–feldspar–biotite granitic gneiss. (d) Sample 2058317, cordierite replacing garnet within weakly foliated feldspar–quartz–biotite granitic gneiss. (e) Sample 2058318, migmatitic magnetite–hematite–biotite–quartz–K-feldspar metasedimentary gneiss. (f) Sample 2058319, abundant spessartine garnets with sillimanite–biotite–cordierite–K-feldspar gneiss with elongate magnetite grains. (g) Sample 2058328, magnetite–hematite–biotite–quartz metasedimentary gneiss. (h) Sample 1839526, from Price Island, magnetite–quartz–biotite–chlorite schist. Bar at bottom of each image represents approximately 25 mm.
within the deposit. During the course of the investigation similarities between the age of the Warramboo magnetite gneiss and the Price Metasediments (Figure 1) became apparent. Previous reconnaissance data from the magnetite-bearing Price Metasediments suggested they were deposited at ca 1765 Ma (Oliver & Fanning 1997). We expanded our investigation to include samples of the Price Metasediments (Figure 4h) to investigate possible geochronological and Sm–Nd isotope correlations with the Warramboo magnetite gneiss. Warramboo sample locations are shown on Figure 2, while the location of the Price Metasediments is shown in Figure 1. Table 1 describes all analysed sample lithologies and locations as well as a summary of the geochronological results.

**ANALYTICAL METHODS**

**Zircon U–Pb SHRIMP geochronology**

Zircons were separated and concentrated by crushing, sieving, Franz magnetic separation, density separation in methylene iodide, and handpicking. Zircons were mounted in 25 mm epoxy discs, together with the U–Pb standard TEMORA-2 (416.8 ± 0.6 Ma; Black et al. 2004) and the 207Pb/206Pb standard OG1 (3465.4 ± 0.6 Ma; Stern et al. 2009). The U-concentration standard M257 (840 ppm U) was included on three of the four mounts. For the mount analysed in session 1, OG1 was utilised as a U standard using a reference value of 173 ppm derived from median values for 26 SHRIMP sessions published in Stern et al. (2009). The epoxy discs were polished to expose the interiors of the crystals. All grains were imaged with a cathodoluminescence (CL), and photographed in transmitted and reflected light to ensure that analyses were collected from discrete zircon growth phases. Mounts were then cleaned and thinly coated in gold to produce a resistivity of 10–20 fA across the disc.

U–Pb isotopic data for Warramboo samples were collected using SHRIMP IIA and IIB based in the John de Laeter Centre for Mass Spectrometry, Perth, Western Australia. U–Pb isotopic data for sample 1839526 from the Price Metasediments was collected on SHRIMP IIe at Geoscience Australia. Data were collected using a primary oxygen beam of ~2.5–3.0 nA producing a spot of approximately 20 microns diameter. Secondary ions were collected on an electron multiplier via cycling of the magnet through five scans across the mass range of interest. Deep cracks and inclusions were avoided during analysis.

Data from all samples were processed using the SQUID 2.5 software of Ludwig (2009) and plotted using Isoplot 4.15 (Ludwig 2012), and AgeDisplay 2.25 (Sircome 2004). Corrections for common Pb were estimated using the Concordia array function in Isoplot. All analyses with less than 5% discordance and reverse discordance (and a few with >±5% discordance) have concordia ages that closely match their 207Pb/206Pb age and a probability of concordance >0.05, indicating their error ellipses overlap concordia at the 95% confidence level. That is, these analyses are essentially concordant. Based on these results, a cutoff of ±5% discordance has been used to exclude poor analyses from age considerations.

For each analytical session an instrumental mass fractionation (IMF) factor was calculated by comparing the weighted mean 207Pb/206Pb ratio measured on OG1 with the thermal ionisation mass spectrometer (TIMS) reference value following the recommendations of Stern et al. (2009), who showed that IMF for zircon 207Pb/206Pb can be significant, and subject to variability between analytical sessions and instruments. The IMF can be used to correct 207Pb/206Pb ratios measured on concurrently analysed unknowns, allowing ages derived from different analytical sessions to be standardised, regardless of the specific causes of machine variability. Significant standard age variation was only observed in one session (Table 2), and pooled ages for the three samples analysed during this session have been corrected for IMF. In addition, the relative uncertainty in the weighted mean 207Pb/206Pb ratios have been augmented by adding in quadrature the relative uncertainty in the weighted mean 207Pb/206Pb ratio of OG1 measured in the session. For sessions 2–4, the IMF correction and uncertainty propagation makes trivial difference to the pooled ages and so is not applied. A summary of U–Pb zircon age results is presented in Table 1, and full data are presented in the Supplementary Papers.

The relatively small amount of drill core available for some samples from the Warramboo Iron Deposit limited the number of good quality zircon grains that could be separated to below recommended numbers for detrital studies (Andersen 2005), with metamict and discordant grains common.

**Whole-rock Sm–Nd isotopic analyses**

Whole-rock Sm–Nd analysis was undertaken at the University of Adelaide. Approximately 0.2 g of rock powder was evaporated to dryness in 7M HNO3/50%HF mixture on a hot plate at 140 °C then digested in hot 7M HNO3/50%HF in high pressure digestion jackets for at least 48 h in an oven, then evaporated to dryness. Samples were redissolved in 6M HCl for conversion to chlorides. Sm and Nd were extracted and separated by ion chromatography using AGW X8 (200–400 mesh) resin in Poly-prep columns followed by HDEHP (eichrom Ln resin SPS) coated Teflon columns. Nd was analysed on a Finnigan MAT262 TIMS in static and quadruple cup dynamic measurement modes. The 143Nd/144Nd ratio was normalised to I=146Nd/144Nd = 0.721903 with Nd concentrations corrected for a 200 pg blank. Sm was analysed on a Finnigan MAT262 TIMS in static measurement mode. Sm concentration was corrected for a 150 pg blank. Measurements of the JNd-1 standard was 143Nd/144Nd = 0.512081 ± 0.000008 (1σ, n = 8) and La Jolla reference material was 143Nd/144Nd = 0.511821 ± 0.000004 (1σ, n = 4). Long-term results for JNd-1 and La Jolla within this facility are: JNd-1 143Nd/144Nd = 0.512084 ±
Table 1 Summary of sample lithologies and U–Pb zircon ages obtained in this study.

| Sample Number Easting Northing | Unit, Lithology, Location | Age type | Age (Ma) | MSWD | prob | n (used) | N (total) |
|--------------------------------|---------------------------|----------|----------|------|------|----------|----------|
| **Magnetite-poor metasedimentary samples** | |  |  |  |  |  |  |  |
| 2058314 562340 6321437 | garnet–cordierite–biotite leucocratic gneiss, IRD204, 90.7–92.6 m | Dmax | 2481 ± 9 | 72 |
| | | metamorphism | 2445 ± 7 |
| | | metamorphism | 2423 ± 7 |
| 2058320\(^1\) 560327 6321996 | garnet–cordierite–biotite leucocratic gneiss, IRD204, 90.7–92.6 m | Dmax | 2478 ± 12 | 65 |
| | | metamorphic range ca 2440–2420 |
| **Magnetite-poor met igneous samples** | |  |  |  |  |  |  |  |
| 2058315 562340 6321437 | quartzofeldspathic gneiss with minor garnet, IRD204, 430.0–440.2 m | magmatic | 2466 ± 5 | 69 |
| 2058316 564140 6321355 | quartzofeldspathic gneiss, IRD452, 372.0–373.7 m | metamorphic range ca 2440–2420 |
| | | metamorphism | 1734 ± 11 | 1.3 | 0.28 | 5 |
| | | magmatic | 2474 ± 5 | 0.51 | 0.77 | 7 | 48 |
| 2058317 561337 6321338 | cordierite–garnet–K-feldspar gneiss, IRD452, 141.00–141.0 m | metamorphic range ca 2440–2400 |
| | | magmatic | 2466 ± 9 |
| **Magnetite metasedimentary samples** | |  |  |  |  |  |  |  |
| 2058318\(^1\) 564940 6321940 | magnetite gneiss, IRD423, 76.0–78.91 m | Dmax | 1766 ± 9 | 1.15 | 0.33 | 7 | 16 |
| | | metamorphism | ca 1730 |
| 2058319 564737 6321339 | magnetite–garnet–sillimanite gneiss, IRD417, 141.76–143.7 m | Dmax | 1742 ± 7 | 1.4 | 0.14 | 33 |
| | | metamorphism | 1732 ± 10 | 0.73 | 0.6 | 6 | 45 |
| 2058328\(^1\) 560327 6321996 | magnetite–hematite–biotite gneiss, IR070, 250.5–255.24 m | Dmax | 1749 ± 8 | 0.69 | 0.65 | 7 | 45 |
| 1839526 525973 6136212 | Price Island, magnetite–quartz–chlorite schist | Dmax | 1749 ± 5 | 1.4 | 0.10 | 25 | 63 |

\(^1\)IMF corrected data. For ages obtained by weighted mean calculations MSWD, probability and n (used) is listed. Other ages are calculated by mixture modelling.
Table 2: Results for OG1 reference zircon analysed during SHRIMP sessions.

| Session no. | Instrument                      | No. of analyses | Weighted mean $^{207}$Pb/$^{206}$Pb age (Ma) $\pm$ Ma (MSWD) | Weighted mean $^{207}$Pb/$^{206}$Pb $\pm$ IMF factor |
|-------------|---------------------------------|-----------------|-------------------------------------------------------------|--------------------------------------------------|
| 1           | Curtin (SHRIMP IIB)             | 19              | 3465.3 $\pm$ 3.8 (1.20)                                   | 0.29903 $\pm$ 0.00073 (0.24%) 0.99987             |
| 2           | Curtin (SHRIMP IIA)             | 13              | 3459.9 $\pm$ 4.9 (1.60)                                   | 0.28600 $\pm$ 0.00093 (0.31%) 0.99642             |
| 3           | Curtin (SHRIMP IIA)             | 21              | 3464.0 $\pm$ 3.2 (1.02)                                   | 0.29879 $\pm$ 0.00061 (0.20%) 0.99906             |
| 4           | Geoscience Australia            | 21              | 3465.2 $\pm$ 1.6 (0.84)                                   | 0.29902 $\pm$ 0.00031 (0.10%) 0.99983             |

Instrumental mass fractionation (IMF) = $^{\cdot}$Pb/$^{\cdot}$PbOG1/$^{\cdot}$PbRM, where $^{\cdot}$Pb/$^{\cdot}$PbOG1 is the ratio measured in OG1 zircon via ion probe in the analytical session, and $^{\cdot}$Pb/$^{\cdot}$PbRM is the reference value of 0.29907 determined by multiple TIMS analyses (Stern et al. 2009). The reference age for OG1 zircon is 3465.4 $\pm$ 0.6 Ma (Stern et al. 2009). All uncertainties quoted at 95% confidence. In Electronic Appendix 1, each sample is referred to with the session number in which it was analysed.

RESULTS

Zircon U–Pb SHRIMP geochronology

MAGNETITE-POOR SAMPLES

Sample 2058314

This sample is a coarse-grained weak to moderately foliated garnet–cordierite–biotite–quartz–feldspar gneiss (Figure 4a). Zircons in this sample are 60–150 $\mu$m in diameter; clear and colourless with rare brown grains, round to subround and slightly elongate, with some elongate or ball-shaped grains (Figure 5a). CL images show that the internal structures of the grains are dominated by oscillatory zoned cores with dark diffuse rims (Figure 5a).

Seventy-two analyses were made on 68 grains; 25 analyses are $>\pm$5.0% discordant or have common $^{206}$Pb $>0.5$% and are not included in the following age calculations. Forty analyses of cores yield ages ranging from ca 3196 to 2397 Ma with the youngest group from ca 2650 Ma to 2397 Ma (Figure 6a). The unmixing algorithm of Sambridge & Compston (1994) calculates six age components at 2644 $\pm$ 8 Ma, 2559 $\pm$ 11 Ma, 2538 $\pm$ 10 Ma, 2508 $\pm$ 11 Ma, 2481 $\pm$ 9 Ma and 2425 $\pm$ 12 Ma. The youngest ca 2425 Ma cores have diffuse oscillatory zonation (Figure 5a) and are the same age as the 2445–2425 metamorphic rims. These zircon grains are interpreted to be protolith derived and affected by Pb loss and/or isotopic resetting during the 2445–2425 Ma metamorphic event. Our preferred interpretation is that the range of older zircon ages in this sample reflects source components in a metasedimentary rock. This is supported by the presence of cordierite and garnet in the metamorphic mineral assemblage, suggesting the protolith has an aluminous composition with the next oldest age grouping of 2481 $\pm$ 9 Ma providing a maximum age estimate for the timing of sedimentary deposition.

Seven analyses of zircon rims have apparent ages ranging from ca 2470 Ma to ca 2420 Ma (Figure 6a). The unmixing algorithm of Sambridge & Compston (1994) calculates two age components 2445 $\pm$ 7 Ma and 2423 $\pm$ 7 Ma and are interpreted to be periods of metamorphic zircon growth.

Sample 2058320

This sample is a coarse-grained moderately foliated garnet–cordierite–biotite–quartz–feldspar gneiss (Figure 4a). Garnet porphyroblasts, 2–4 cm in size, are wrapped by biotite and cordierite. Zircons in this sample are 50–150 $\mu$m in diameter, clear and colourless with rare brown grains, round to subround and slightly elongate (Figure 5b). Some grains are elongate with aspect ratios of 3:1. CL images show that the internal structures of the grains are dominated by oscillatory zoned cores truncated by thick dark homogenous rims (Figure 5b).

Sixty-five analyses were collected on 53 grains, targeting cores and rims. Over half the analyses are $>\pm$5% discordant and two discordant analyses (A18-11 and A18-51) are imprecise. Twenty concordant core analyses yield ages ranging from ca 3320 to 2470 Ma (Figure 6b) and include six grains with ages between ca 3320 and 2590 Ma, and a cluster of 14 analyses between ca 2565 and 2470 Ma. Mixture modelling was used to interpret possible ages within this younger cluster. The maximum number of possible age groups that can be modelled is 4, which gives an age of 2478 $\pm$ 12 Ma for the youngest peak, with older age groups at ca 2507 Ma, 2529 Ma and 2561 Ma. Thirty-one core analyses are discordant or imprecise. The older analyses (ca 2770–2470 Ma) produce age peaks that mimic almost identically the discordant age peaks in the same range, suggesting these are similar-aged grains that have experienced only recent Pb loss. Younger apparent ages (ca 2445–2170 Ma) have a similar age range as the dark CL metamorphic rims, with the oldest age peak (ca 2445 Ma) broadly coeval with regional Sleaford-aged metamorphism.

Nine concordant analyses on dark homogenous zircon rims interpreted to have grown during metamorphism yield apparent ages ranging from ca 2420 Ma to 2265 Ma. As with the disturbed cores, the older end of the series coincides with the known age of the ca 2460–2410 Ma Sleaford Orogeny (Daly & Fanning 1993; Fanning...
Figure 5  CL images of representative zircons from magnetite-poor samples (right) along with location of SHRIMP analyses and their corrected $^{207}$Pb/$^{206}$Pb ages (left). (a) Sample 2058314. (b) Sample 2058320. (c) Sample 2058315. (d) Sample 2058316. (e) Sample 2058317. Black bar is approximately 100 μm.
Figure 6 Summary of SHRIMP U–Pb data from magnetite-poor samples presented as Terra-Wasserburg concordias (right), and probability density distributions (left). (a) Sample 2058314. (b) Sample 2058320. (c) Sample 2058315. (d) Sample 2058316. (e) Sample 2058317. Note that for metasedimentary samples (a & b) uncertainty ellipses on concordia diagrams are shown as blue for detrital zircon and red for metamorphic zircon analyses and reset cores; for meta-igneous samples (c–e) uncertainty ellipses are shown as green for magmatic zircon, purple for inherited zircon, and red for metamorphic zircon and reset cores. All error ellipses are 1σ. Probability density distributions show all zircon data with frequency histograms, bin width of 10 Ma. Light grey curve represents all zircon data. Dark grey curve represents detrital zircon data filtered for ≤5% discordance.
et al. 2007), and suggest that the metamorphic rims formed during this event that also isotopically reset the protolith zircon. The scattered discordia trend towards younger, post-Sleaford ages recorded in both the cores and metamorphic rims is difficult to interpret geologically, but is consistent with an episode of partial Pb loss during the Paleoproterozoic, perhaps during the Kimban Orogeny.

As with 2058314, our preferred interpretation is that the wide range of zircon ages in this sample reflects source components in a sedimentary rock and that the youngest concordant age grouping of 2478 ± 12 Ma provides a maximum age constraint on the timing of sedimentary deposition. Metamorphic zircon growth occurred in the later stages of the Sleaford Orogeny (ca 2445–2400 Ma), with further Pb loss from the cores and rims occurring during a later Paleoproterozoic event, most likely the Kimban Orogeny.

Sample 2058315

This sample is a medium-grained K-feldspar–plagioclase–quartz–biotite gneiss, dominated by 3–6 mm K-feldspar phenocrysts with rare 6–12 mm feldspar phenocrysts and small 1 mm garnets (Figure 4b). The foliation is defined by quartz–feldspar banding and orientated biotite. Zircons in this sample range from subangular to subrounded, with sizes ranging from 100 to 200 µm and some smaller 50–100 µm grains. Most grains have an aspect ratio of about 1:2 and long axis dimensions between 100 and 200 µm. Grains are clear and colourless with occasional brown coloured red staining. CL response shows the grains are dominated by oscillatory zoning; some grains have developed thick, dark rims (Figure 5c). Many cores exhibit sector zoning or broadening of oscillatory zoning with a light or dark grey patchiness.

Sixty-nine analyses were made on 68 grains, targeting both cores and rims. Forty analyses are >5.0% discordant. Three analysis (IR41-48, IR41-32 and IR41-04) were placed over a core–rim boundary, and are not considered further. Most of the concordant analyses cluster near concordia between ca 2495 and 2380 Ma. Four analyses are older; one concordant at ca 2580 Ma, and three discordant, with 207Pb/206Pb ages of ca 2895, 2600 and 2565 Ma. Within the main cluster of analyses, there is a dominant peak at ca 2465 Ma, with subordinate apparent age peaks at ca 2435 Ma and 2420 Ma and a few analyses ranging down to ca 2380 Ma in age. The 12 analyses defining the main peak yield a weighted mean age of 2466 ± 5 Ma (MSWD = 1.5, probability = 0.14) that we interpret to be the crystallisation age of the granite protolith. The oldest analysis in the group (IR41-30; 2495 Ma) is interpreted to be a slightly older inherited core. The younger concordant age peaks are difficult to interpret geologically. The concordant core and rim analyses all lie on a tight discordia with intercepts of 2494 ± 35 Ma and 1703 ± 80 Ma (MSWD = 0.47) that could suggest a single Pb loss event during the Kimban Orogeny with the apparent age peaks at ca 2435 Ma and 2420 Ma artefacts of variable amounts of lead Pb loss in the ca 2466 Ma zircon cores, and have no geological meaning. Alternatively, these peaks could also record older Pb loss during the Sleaford Orogeny. Given that there is also evidence that older metamorphic rims formed on the cores, this interpretation is preferred. From the high degree of scatter in the discordant core analyses, it is clear that the zircon cores experienced multiple episodes of Pb loss. The pattern of discordance is consistent with isotopic resetting during both the Sleaford and Kimban orogenies, overprinted by younger, perhaps recent Pb loss.

Nine analyses targeted metamorphic rims. Five concordant analyses have a weighted mean age of 1734 ± 11 Ma (MSWD 1.3, probability = 0.28), recording high-grade metamorphism during the Kimban Orogeny that is probably associated with the development of the gneissic fabric. Four discordant analyses are older and lie dispersed along the discordia between the concordant core analyses and the concordant ca 1735 Ma rims. It is possible these rims originally formed during the earlier Sleaford Orogeny and, owing to their high U, metamict nature, experienced Pb loss during the Kimban Orogeny, when new zircon growth also occurred on some grains.

Sample 2058316

This sample is a moderate to strongly foliated equigranular K-feldspar–quartz–plagioclase–biotite gneiss (Figure 4c). Zircons in this sample are subhedral to rounded and range in size from 80 to 300 µm but more commonly range from 100 to 150 µm. Grains are commonly equant with rounded terminations; ball-shaped round grains are also observed (Figure 5d). Grains are clear and colourless with occasional brown coloured grains, many with deep cracks and inclusions that are visible in transmitted light. Grains are dominated by oscillatory zoning with thin dark rims too narrow to accommodate a SHRIMP analysis. Grains commonly exhibit broadening of zonation or dark mottled patches (Figure 5d).

Forty-eight analyses were collected from 47 grains in this sample. Twenty-six analyses are >5.0% discordant; two discordant analyses with ages of ca 3233 Ma and 2573 Ma are interpreted as inheritance. Seventeen concordant analyses on cores give apparent ages ranging from ca 2480 Ma to 2290 Ma with a large proportion of analyses in the oldest population at ca 2475 Ma (Figure 6d). The seven analyses defining the oldest age peak of 2474 ± 5 Ma (MSWD = 0.5, probability = 0.77) are interpreted to be the crystallisation age of the granite protolith. Younger analyses ca 2450 Ma to 2290 Ma correspond to cores that exhibit a fine network of cracks, diffuse CL zonation and dark mottles, suggesting these grains were susceptible to Pb loss and/or isotopic resetting. Apparent ages between ca 2450 and 2415 Ma suggest some cores had their isotopic system reset during the Sleaford Orogeny; an interpretation supported by a single discordant analysis on a rim—with an age of 2401 ± 4 Ma (1σ) that suggests the granite was metamorphosed at this time. Younger apparent ages between ca 2370 and 2290 Ma suggest post-Sleaford Pb loss also affected the cores. These analyses lie on a shallow discordia trend consistent with an episode of partial Pb loss during the Paleoproterozoic, probably during the Kimban Orogeny, as this event is recorded in some of the other magnetite-poor samples in the study. Discordant core analyses plot on a discordia line trending between ca 2431 Ma and ca
679 Ma, suggesting more recent Pb loss has also affected the grains.

The oldest concordant age of ca 2474 ± 5 Ma is interpreted as the age of crystallisation for this granitic gneiss. The single concordant metamorphic rim at ca 2400 Ma suggests that the metamorphic zircon grew at or before ca 2400 Ma with subsequent ancient Pb loss and/or isotopic resetting resulting in the discordant scatter.

Sample 2058317

This sample is a weakly foliated K-feldspar–plagioclase–quartz–garnet–cordierite–biotite gneiss (Figure 4d). Cordierite partially replaces 10–30 mm garnet porphyroblasts with a foliation defined by biotite. Zircons in this sample are 50–100 μm in size and range in shape from elongate to equant. Grains are clear and colourless with light brown or yellowish tints to some grains. CL images show the grains to be dominated by oscillatory zoning that commonly exhibits some broadening or sector zoning (Figure 5e). Thin grey rims too narrow to accommodate a SHRIMP analysis truncate the zoning of some grains. Grains are commonly cut by deep and fine cracks, which are visible in transmitted light.

Thirty-six analyses were collected from cores. Common 206Pb was <0.5% in all analyses and 16 analyses are <5% discordant. Apparent ages range from ca 2872 to 2317 Ma (Figure 6e). Two older analyses are concordant at ca 2870 Ma and 2680 Ma. A single analysis at ca 2317 Ma on a zircon with indistinct CL oscillatory zonation and several cracks visible in transmitted light is interpreted to be metamict zircon affected by Pb loss. The remaining 13 cores range from ca 2580 to 2450 Ma. Applying mixture modelling to this group of analyses produces three ages of 2559 ± 11 Ma, 2513 ± 12 Ma and 2466 ± 9 Ma. Scattered discordant analyses suggest the zircons have been subject to multiple periods of Pb loss. The pattern is consistent with an episode of partial Pb loss during the Paleoproterozoic, perhaps during the Kimban Orogeny, overprinted by more recent Pb loss.

Owing to the small number of concordant analyses, it is difficult to draw conclusions as to the age and nature of this sample, with several histories possible. This sample may be interpreted as a granitic melt that crystallised at ca 2466 ± 9 Ma, with the older zircon representing a large amount of inherited zircon. Another possible interpretation is that this is a metasedimentary sample that has been affected during metamorphism to form a diatexite, and the youngest age peak of ca 2466 ± 9 Ma is a constraint on the maximum age of deposition for the metasedimentary protolith prior to metamorphism. A third interpretation is that the age of crystallisation of this sample is represented by the older age peak of ca 2559 ± 11 Ma, and the younger ages of ca 2513 Ma and 2466 Ma are zircon grains that have been variably affected by ancient Pb loss. This interpretation is considered the least likely, as the weakly developed foliation, well-preserved igneous textures and development of only very thin metamorphic zircon rims indicate that this sample was only weakly metamorphosed during the Seaford and Kimban orogenies. Therefore, the protolith zircon is less likely to be isotopically disturbed. The dominant K-feldspar–plagioclase–quartz mineralogy, tabular morphology of the K-feldspar phenocrysts and homogenous nature of the unit suggest this is granitic gneiss. The zircon CL images do not show the abundant rims that characterise metamorphic zircon growth that would support the formation of a diatexite during metamorphism. On the basis of the mineralogy, texture and zircon characteristics, the preferred interpretation is of a granitic gneiss that crystallised at ca 2466 ± 9 Ma, with the older ages representing a large inherited zircon component.

MAGNETITE GNEISS SAMPLES

Sample 2058318

This is a sample of the high ore grade magnetite gneiss. The sample is a coarse-grained (3–8 mm) magnetite–biotite–K-feldspar–quartz metasedimentary gneiss. Feldspar–quartz leucosome material makes up 15–20% of the rock with the rest composed of foliated magnetite–biotite±quartz (Figure 4e). The heterogeneous mineralogy within this unit may be a relic sedimentary feature. A small number of zircons were separated from this sample. Zircons are small, ranging in size from 40 to 100 μm. Grains are subround to round, clear to light brown and are round to slightly elongate in shape. CL imaging shows the grains are primarily oscillatory zoned to homogenous cores with thin bright rims (Figure 7a). Most cores appear affected by Pb loss or isotopic resetting with broadening of oscillatory zoning and cracks that are visible in transmitted light common.

Sixteen grains were analysed, targeting mainly cores and several rims thick enough to accommodate a SHRIMP analysis. All grains have common 206Pb <0.5%. The three oldest core analyses are discordant and yield 207Pb/206Pb ages of ca 1910, 1836 and 1809 Ma; these are interpreted as inherited zircon. Nine analyses on cores that are <5% discordant range from 1777 Ma to 1726 Ma. With a high MSWD of 2.6, the analyses are scattered beyond a single population at the 95% confidence level. The two youngest analyses are the same age as the bright CL zircon rims, suggesting these cores might have experienced Pb loss and/or isotopic resetting during metamorphism. When eliminated, the remaining seven analyses conform to a simple population with a weighted mean 207Pb/206Pb age of 1766 ± 9 Ma (MSWD 1.15, probability 0.33).

Two analyses on bright CL zircon rims have ages of 1742 ± 31 and 1726 ± 6 Ma (1σ errors). We interpret 1766 ± 9 Ma to be the best estimate for the maximum depositional age of this sample, with metamorphism occurring during the Kimban Orogeny, at ca 1730 Ma.

Sample 2058319

This sample is a magnetite–garnet– sillimanite–cordierite–biotite–K-feldspar–quartz metasedimentary gneiss with a foliation defined by biotite and elongate-fibrous sillimanite mats (Figure 4f). Euhedral garnets are 0.5 mm, brown and very abundant. Magnetite forms discrete aggregates that are sometimes aligned to the
Zircons in this sample are small and subhedral to rounded, generally 50–80 μm with some grains up to 100 μm. Grains are clear and colourless with fine and deep cracks visible in transmitted light. Most grains are dominated by cores of primary dark oscillatory zonation that are variably overgrown by grey or bright rims (Figure 7b). Cores are generally well preserved, with some cores exhibiting broadening of oscillatory zoning. Zoned cores are truncated by rim overgrowths. Dark homogenous zircons that do not show oscillatory zoning are also present in this sample.

Fifty-three analyses were made on 52 grains targeting cores and rims. Nineteen grains are >5.0% discordant and/or have >0.5% common Pb. Twenty-eight concordant analyses of cores have a spread of ages from ca 1842 to 1640 Ma (Figure 8b). The two youngest cores are significantly younger than the enveloping metamorphic rims, and as they exhibit cracks under transmitted light, they are interpreted to have experienced post-Kimban Pb loss, and the analyses are not considered to have any geological significance. The remaining 26 analyses form three age peaks at ca 1830 Ma (n = 2), 1781 ± 8 Ma

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**Figure 7** CL images of representative zircons from magnetite-bearing samples along with location of SHRIMP analyses and their corrected 207Pb/206Pb ages. (a) Sample 2058318. (b) Sample 2058319. (c) Sample 2058328. (d) Price Metasediments, sample 1839526. Black bar is approximately 100 μm.
Figure 8 Summary of SHRIMP U–Pb data from magnetite-bearing metasedimentary samples presented as Terra-Wasserburg concordia (right) and probability density distributions (left). (a) Sample 2058318. (b) Sample 2058319. (c) Sample 2058328. (d) Sample 1839526. Note that uncertainty ellipses on concordia diagrams are shown as blue for detrital zircon and red for metamorphic zircon analyses and reset cores. All error ellipses are 1σ. Probability density distributions show all zircon data with frequency histograms, bin width of 10 Ma. Light grey curve represents all zircon data. Dark grey curve represents detrital zircon data filtered for ≤5% discordance.
Ten analyses are oscillatory cores, rims and homogenous zircon grains. Twenty-eight analyses on oscillatory metamorphism. Zircon grains within the age range 1740 Ma to 1700 Ma have more distinct CL zoning and fewer cracks in transmitted light, and our preferred interpretation is that 1749 ± 8 Ma represents the maximum age of deposition of this sample.

Two concordant rim analyses and four dark homogenous zircons, with no oscillatory zoned cores that are interpreted as metamorphic zircon, give apparent ages ca 1740 Ma to 1708 Ma with a weighted mean age of 1732 ± 10 Ma (MSWD 0.73, probability 0.60, n = 6) and are interpreted as the age of metamorphic zircon growth in this sample.

Sample 2058328

This is a sample of the high-grade magnetite ore from the deposit. The rock is a magnetite–hematite–biotite–quartz–K-feldspar metasedimentary gneiss, defined by alternating bands of K-feldspar–quartz leucosomes and magnetite–hematite–biotite (Figure 4g). The strong foliation is defined by 3–10 mm biotite–magnetite–hematite on a metre to decametre scale and defines a shear fabric with top to the south normal movement defined by S–C fabric kinematics (Figure 4g). Zircon grains in this sample are round to subrounded and typically 50–80 μm in size with occasional larger grains 100 μm in size. Grains are clear and colourless. CL images show varied morphologies, with some grains dominated by oscillatory zoned cores with grey to white rims or sector zoning, while others exhibit homogenous CL from core to rim (Figure 7c).

Forty-five analyses were made on 45 grains targeting oscillatory cores, rims and homogenous zircon grains. Ten analyses are >5.0% discordant and are not included in age calculations. Twenty-eight analyses on oscillatory zoned cores have ages ranging from ca 1790 Ma to 1710 Ma, with two older grains at ca 1929 Ma and 1838 Ma. Mixture modelling resolves the main group of analyses into three peaks at 1780 ± 9 Ma, 1749 ± 8 Ma and 1714 ± 15 Ma. Zircon grains within this youngest age group of ca 1715 Ma are cut by fine and deep cracks, visible in transmitted light, that are interpreted to have been affected by Pb loss and/or isotopic resetting during metamorphism. Zircon grains within the age range ca 1780–1750 Ma have more distinct CL zoning and fewer cracks in transmitted light, and our preferred interpretation is that 1749 ± 8 Ma represents the maximum age of deposition of this sample.

Three concordant analyses on rims have ages of ca 1740–1710 Ma. Four concordant analyses on homogenous zircon cores have ages of ca 1740–1700 Ma. These zircon are interpreted as zircon rims and new zircon grains that grew during metamorphism with a weighted average age of 1727 ± 13 (MSWD 0.69, probability 0.65, n = 7) interpreted as the timing of metamorphism in this sample.

Sample 1839526, Price Metasediments

This sample is a magnetite–chlorite–biotite–quartz schist (Figure 4h) from rock outcrops on Price Island that form part of the Price Metasediments aeromagnetic high on southwest Eyre Peninsula (Figure 1). Zircons are small, about 50 μm in size, and euhedral to subhedral, many with crystal faces and pyramidal terminations preserved. The grains are clear and colourless, with few inclusions and cracks. CL images reveal that many grains have concentric, oscillatory zoning throughout the grains with a relatively bright CL response (Figure 7d). A few have an oscillatory core surrounded by a darker, grey CL rim.

Sixty-three grains were analysed. Twenty-one analyses contain >0.5% common 206Pb and/or are >5.0% discordant. Of the remaining 42 analyses, the youngest 25 define a weighted mean 207Pb/206Pb age of 1749 ± 5 Ma (MSWD 1.4, probability 0.10). This is a maximum depositional age estimate for this metasedimentary rock. The remaining grains record a range of ages between 2510 Ma and ca 1777 Ma, with interim ages at ca 2307 Ma, 1975 Ma, 1868 Ma, 1820 Ma and 1797 Ma (Figure 8d). No metamorphic zircon is observed in this sample.

SM–Nd WHOLE-ROCK DATA

Sm–Nd isotope data from whole-rock samples from the magnetite gneiss at Warramboo and the magnetite schist from the Price Metasediments are summarised in Table 3. εNd(t) is calculated at 1750 Ma for all samples based on zircon U–Pb age constraints. The Warramboo magnetite gneiss and Price Metasediments samples

| Sample | Age (Ma) | Sm (ppm) | Nd (ppm) | εNd/144Nd | εNd0 | εNd0 | TDM (Ma)b |
|--------|----------|----------|----------|-----------|------|------|-----------|
| Warramboo magnetite metasedimentary samples | 2058318 | 1766 ± 6 | 7.8 | 46.3 | 0.1013 | 0.510137 | −26.03 | −4.66 | 2475 |
| 2058319 | 1753 ± 8 | 6.7 | 37.8 | 0.1078 | 0.510169 | −23.95 | −4.04 | 2474 |
| 2058328 | 1719 ± 5 | 5.2 | 29.1 | 0.1073 | 0.510205 | −23.37 | −3.33 | 2419 |
| Price Metasediments | 2017472 | 1749 ± 5 | 6.8 | 36.8 | 0.1119 | 0.510201 | −22.41 | −3.41 | 2456 |
| 1839526 (2017371) | 1749 ± 5 | 7.4 | 41.9 | 0.1066 | 0.510222 | −23.18 | −3.00 | 2392 |
| 2017372 | 1749 ± 5 | 4.9 | 26.6 | 0.1117 | 0.510180 | −23.27 | −4.22 | 2514 |
| 2017473 | 1749 ± 5 | 10.0 | 57.5 | 0.1049 | 0.510224 | −23.54 | −2.97 | 2380 |

εNd/144Nd CHUR(0) = 0.51966.

TDM calculated using single stage model of Michard et al. (1985).
DISCUSSION

Geochronological constraints on the Warramboo deposit

There is a clear trend within the geochronology; magnetite-poor metasedimentary and meta-igneous samples record Archean to earliest Paleoproterozoic ages, while magnetite-rich metasedimentary samples record much younger protolith ages, down to ca 1745 Ma (Table 1). Samples 2058314 and 2058320 are interpreted as metasedimentary gneisses, have almost identical mineralogy and contain similar Archean and earliest Paleoproterozoic aged zircons, with the youngest detrital population at ca 2480–2470 Ma. These samples are interpreted to be part of the same metasedimentary unit at Warramboo. Detrital zircon ages constrain the protolith to this unit to be no older than 2470 Ma and the lower age limit is defined by the presence of metamorphic zircon rims at ca 2445 Ma. Dominant detrital populations range between ca 2560 and 2470 Ma with smaller peaks at ca 3220–3100, 2875 and 2660 Ma. A possible source within the Gawler Craton of 3200–3100 Ma zircon is the ca 3150 Ma Cooer- doo Granite (Fraser et al. 2010) on the eastern Eyre Peninsula. In the Sleaford Complex, sources of ca 2875 Ma zircon have not been identified, although similar ages have been recorded in inherited and detrital zircons from units of the Sleaford and Mulgathing complexes (Swain et al. 2005; Fraser & Neumann 2010; Szpunar et al. 2011; Reid et al. 2014). Additionally, ca 2560 Ma migmatic rocks are not identified in the Sleaford Complex, but are recognised within the Mulgathing Complex with the ca 2555 Ma Devils Playground Volcanics being a possible source (Cowley & Fanning 1991). Multiple sources of ca 2530–2500 Ma zircon are recorded within the Sleaford Complex (Fanning 1997, 2002; Fraser & Neumann 2010; Reid et al. 2014), indicating that zircons of this age were likely sourced from Sleaford Complex igneous rocks. Samples 2058315, 2058316 and 2058317 are interpreted on the basis of their mineralogy and zircon data to be granitic gneisses. These samples yield crystallisation ages ca 2475–2465 Ma (Table 1), with minor inherited components ca 2900–2500 Ma. The magnetite-poor units at Warramboo are likely part of the Sleaford Complex basement of the Southern Gawler Craton (Daly & Fanning 1993).

Metamorphic zircon from metasedimentary and meta-igneous samples falls into two populations at ca 2445 Ma and ca 2422 Ma, which are consistent with ages for the Sleaford Orogeny from the Gawler Craton (McFarlane 2006; Fanning et al. 2007; Jagodzinski et al. 2009; Dutch et al. 2010; Reid et al. 2014).

In contrast to the magnetite-poor lithologies at Warramboo, the magnetite-rich gneisses present a significantly younger depositional history. All three samples of Warramboo magnetite gneiss analysed record similar detrital zircon populations and maximum deposition ages. The youngest detrital zircon populations within the Warramboo magnetite gneisses range in age from 1766 Ma to 1742 Ma. The minimum depositional age for the protoliths is constrained by the growth of ca 1730 Ma rims. The time gap between the youngest detrital zircons and the metamorphic zircons suggests there was a time interval of possibly less than 20 Myr between deposition and high-grade metamorphism. These data also show that the metamorphism and deformation of the Warramboo deposit occurred during the Kimban Orogeny, in a region previously thought to have been only moderately affected by the Kimban Orogeny (Daly & Fanning 1993; Fraser & Neumann 2010).

The zircon provenance of the magnetite gneisses is dominated by ca 1790–1750 Ma zircon, with >1800 Ma zircon rare and commonly discordant (Figure 8; Table 1). Possible source rocks of similar age are known throughout the Gawler Craton, with ca 1790 Ma Wertigo Granite, ca 1755 Ma Burkitt Granite and unnamed granites ca 1790–1750 Ma found along the eastern Eyre Peninsula (Fraser & Neumann 2010). Similar-aged volcanics, the ca 1790 Ma Myola Volcanics (Fraser & Neumann 2010; Szpunar & Fraser 2010) and the ca 1755 McGregor Volcanics (Fraser & Neumann 2010), also occur in the southern Gawler Craton (Figure 1) and are possible sources for the Warramboo magnetite gneisses.

The significant difference in zircon provenance between the magnetite gneisses and the felsic magnetite-barren gneisses is interpreted to reflect a contact between Sleaford Complex basement and a ca 1750 Ma cover sequence. The boundary between the magnetite gneiss and the older felsic gneisses was therefore most likely originally an unconformity. Interestingly, however, the lack of detritus older than ca 2400 Ma in the magnetite gneisses, a common age in the felsic gneisses at Warramboo and within the Gawler Craton, suggests that these regions were not being eroded into the basin at the time of deposition. However, the restricted populations of the Warramboo samples could also be explained by the limited number of zircons analysed, as the spectra do not necessarily represent statistically robust detrital zircon populations (Andersen 2005). In drill hole IR070, where felsic and magnetite gneiss was sampled (samples 2058320 and 2058328, respectively), the unconformity was identified at a depth of 238.5 m as a zone of leucocratic gneiss with abundant coarse-grained garnet and feldspar terminating at a K-feldspar–quartz–biotite gneiss with no magnetite that transitions sharply into magnetite gneiss (Figure 9). Structurally the older felsic gneiss is above the magnetite gneiss; therefore, in this drill hole, the stratigraphy is inverted.

Prior to this study, the magnetite gneisses at Warramboo were thought to be a BIF formed during the interval ca 2555–2460 Ma and therefore to be part of the Sleaford Complex (CRA 1984; Webb et al. 1986; Daly & Fanning 1993). SHRIMP zircon geochronology from the magnetite-poor lithologies at Warramboo confirms that these units are earliest Paleoproterozoic in age and likely part of the Sleaford Complex. However, SHRIMP geochronology from the magnetite gneisses indicates that these units were deposited between ca 1750 and 1730 Ma, and are therefore significantly younger than the surrounding magnetite-poor host rocks. The presence of detrital zircons also suggests a significant clastic component in the original sediments and therefore indicates that the
magnetite gneisses are unlikely to be true chemical sediment, supporting the early work of Whitehead (1978).

Structure of Warramboo deposit

The recognition of the presence of stratigraphy at Warramboo allows for an interpretation of the structural style in the deposit. The isoclinal fold structure observed in the aeromagnetics and fold hinge data from drill core indicate that the large-scale fold hinge is shallowly south dipping and plunges to the southwest. The ca 1750 Ma Warramboo magnetite gneiss lies structurally between rocks of the Sleaford Complex (Figure 3). Therefore, the large-scale structure of the Warramboo deposit is interpreted to be a shallowly southwest-plunging syncline (Figure 3).

The east–west fold axis orientation at Warramboo indicates the major compressional stress was orientated roughly north–south. Elsewhere in the southern Gawler Craton, the early Kimban Orogeny involved north–south stretching and dextral top-to-the-north shearing followed by east–west compression and folding (Dutch et al. 2010). The structural fabric at Warramboo likely preserves the early phases of deformation during the Kimban Orogeny that has been significantly reworked elsewhere on the southern Eyre Peninsula.

Correlation with Price Metasediments

U–Pb detrital geochronology indicates that the Warramboo magnetite gneisses were deposited at ca 1750 Ma. Similar sedimentary sequences occur across the southern Gawler Craton, including the Wallaroo Group (Cowley et al. 2003; Fanning et al. 2007), the Moonabie Formation (Fraser & Neumann 2010), and significantly, the magnetite-bearing Price Metasediments, described by Oliver & Fanning (1997) as iron-rich phyllites on the southern Eyre Peninsula (Figure 1). Compositionally the Price Metasediments comprise magnetite–quartz–biotite–muscovite–chlorite±spessartine garnet with total Fe 16.8 wt%, MnO 0.89 wt% and SiO$_2$ 59.81 wt% (Oliver & Fanning 1997). Oliver & Fanning (1997) presented only a summary of their zircon geochronology, which indicated deposition at ca 1766 ± 14 Ma, and correlated with the Cape Hunter phyllite located in Commonwealth Bay, Antarctica, interpreting a large narrow basin several hundred kilometres long.

Our new zircon data from the Price Metasediments indicate a maximum depositional age of ca 1749 ± 5 Ma (Figure 8d) that correlates very closely with the maximum depositional ages for the magnetite gneisses at Warramboo. Detrital zircon peaks at ca 1790 and 1750 Ma are found in both sequences, and although the Price Metasediments record singular older grains at ca 2510 and 2300 Ma, not found in the Warramboo magnetite gneiss, the age spectra are virtually identical (Figure 8). The presence of spessartine garnets in both units also suggests a common sedimentary source. The Sm–Nd isotope data from the Warramboo magnetite gneisses range from $\varepsilon_{\text{Nd}}$ 4.46 to 3.34, closely overlapping the range for the Price Metasediments with $\varepsilon_{\text{Nd}}(1750)$ values 4.23 to 2.98 (Table 3). The similarity in the $\varepsilon_{\text{Nd}}(1750)$ values also supports a common source region. This $\varepsilon_{\text{Nd}}(1750)$ range is more juvenile than the average Archean and early Paleoproterozoic >2450 Ma Gawler Craton with an $\varepsilon_{\text{Nd}}(1750)$ value of approximately −11 (Howard et al. 2011), suggesting input from a more juvenile source region for both the Warramboo magnetite gneiss and the Price Metasediments.

On the southern Eyre Peninsula, the Price Metasediments occupy an isoclinal synform ~3 km wide that is bounded by Sleaford Complex granitic gneisses (Figure 1; Dutch et al. 2008, 2010). The current structural architecture of the Price Metasediments is remarkably similar to the Warramboo deposit, where the magnetite gneiss is now isoclinal folded and interlayered with Sleaford Complex metasedimentary and meta-igneous gneisses.
Furthermore, the aeromagnetic trends of the Price Metasediments taper northward, becoming less distinct in the region beneath the Polda Basin (Figure 1). However, despite the lower overall magnetic response in this region, it appears that the magnetic trend re-emerges to the north of the Polda basin in approximately the location of the large magnetic anomaly associated with the Warramboo magnetite gneiss. Indeed, there may even be a physical link between the Warramboo magnetite gneiss and the Price Metasediments.

The geochronology and Sm–Nd isotope values for both the Warramboo magnetite gneisses and the Price Metasediments are remarkably similar, and we conclude that they have a common sedimentary origin and likely formed within the same basin system. By extension, the Cape Hunter phyllite in Antarctica is also a correlative of the Warramboo magnetite gneiss. The Price Metasediments may also be considered as low-metamorphic grade equivalents of the magnetite gneisses. The metamorphism experienced by the Warramboo gneisses has produced a coarse-grained gneiss, and given the abundance of partial melting within the Warramboo sequences it is likely that the metamorphism has effectively upgraded the deposit. This deposit style, where the metamorphism directly upgrades the iron formation to an economic deposit, is in contrast to the BIF and IOCG deposits known in the Gawler Craton (Skirrow et al. 2007; Szpunar et al. 2011; Reid et al. 2014). However, similar-aged sediments are known in metamorphic terranes across the Gawler Craton (Cutts et al. 2011) and represent an under-explored iron potential. There is also potential for similar magnetite deposits to be hosted in middle Paleoproterozoic basins across the world, particularly in regions that have undergone high-grade metamorphism.

**CONCLUSIONS**

- Zircon geochronology on the Warramboo magnetite gneisses provide maximum depositional ages of ca 1760–1745 Ma and minimum depositional age constraint of ca 1730 Ma.
- The magnetite gneisses at Warramboo are not an Archean BIF; they are a package of iron-rich sediments deposited between ca 1750-1730 Ma onto rocks of the ca 2480 Ma Seaforth Complex and were subsequently deformed and metamorphosed to granulite facies during the Kimban Orogeny. The large-scale structural architecture of the Warramboo deposit appears to be a syncline with metamorphism and melt loss significantly increasing the iron content of the magnetite gneiss.
- The Price Metasediments on the southern Eyre Peninsula and the Cape Hunter phyllite in Antarctica may represent basin correlates to the Warramboo magnetite gneiss.

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**SUPPLEMENTARY PAPERS**

Appendix 1: U–Pb SHRIMP data and Sm–Nd data.

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