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Cover photograph: Cathodoluminescence images of three zoned zircon grains analysed in the study. Hafnium isotopes have provided more information on the >3530 Ma Pilbara crust that preceded volcanic rocks of the Pilbara Supergroup. The middle zircon, with two laser holes, is a 3580 Ma grain from the Warrawagine tonalite gneiss; the other grains are detrital zircons from c. 3650 Ma rocks that are no longer exposed. The landscape view shows outcrops of the c. 3500 Ma North Star Basalt at a locality (MGA Zone 50, 795650E 7673000N) 20 km north-northeast of Marble Bar.

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Early evolution of the Pilbara Craton from hafnium isotopes in detrital and inherited zircons

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Abstract

Various geological reviews of the Pilbara Craton, one of the best preserved Paleoarchean crustal blocks on Earth, have interpreted a >3500 Ma continental basement beneath the East Pilbara Terrane. Evidence has included 3800–3530 Ma detrital zircons in the post-3530 Ma sedimentary formations of the greenstone succession, <3730 Ma xenocrystic zircons in felsic volcanic and intrusive rocks, and whole-rock Sm–Nd isotopic data, suggesting crustal developments as early as c. 4000 Ma. Such evidence supports the previous existence of Hadean–Eoarchean crust in the Pilbara Craton, and therefore is consistent with the possibility of a globally widespread felsic crust in the Hadean. However, the concept of voluminous stabilized Hadean–Eoarchean continental crust contrasts with the very restricted amount of pre-3600 Ma rock presently identified on Earth's surface. To seek additional evidence for the existence or otherwise of Hadean–Eoarchean crust during the Archean evolution of the Pilbara Craton, we obtained Hf isotope data from previously dated >3550 Ma detrital zircons in sandstones, and inherited zircons hosted by granitic gneisses. All zircons had previously been dated by the U–Pb (SHRIMP) method. The Hf isotope composition of the ancient zircons in our study suggests that most of the earliest components of the Pilbara Craton were extracted from near-chondritic mantle between c. 3700 and 3600 Ma. If significantly older crust was ever present in the Pilbara it did not have input to the samples used in our study. Our new data suggest that the Pilbara Craton either developed remote from the isotopic influence of >3700 Ma continental masses, or that the stabilized volumes of the earliest continents have been overestimated.

KEYWORDS: Eoarchean, continental crust, detrital zircon, Hadean, Hafnium isotopes

Introduction

A number of workers have inferred the existence of large volumes of continental crust on the early Earth (Armstrong, 1981; McLennan and Taylor, 1982; Reymer and Schubert, 1984). For example, recent studies based on global compilations of detrital zircon age and Hf ± O isotope data predict that 30–40% of the present continental volume was stabilized by 3500 Ma (Belousova et al., 2010; Dhuime et al., 2012). However, this interpretation is not supported by the miniscule amount of exposed ancient crust that has been identified at the present day. Rocks older than 3500 Ma occupy substantially less than 1% of the currently exposed continents (Condie, 2011). It may be that large amounts of continental crust did exist in the Eoarchean to Paleoarchean although it has since been destroyed by any of a range of crustal recycling processes. Alternatively, the discrepancy between the amount of ancient crust predicted by isotope models and that identified in the geological record could reflect difficulties in recognizing the oldest components of cratons because of younger thermal overprints or extensive cover by vegetation, soil, or ice.

The Paleoarchean to Mesoarchean Pilbara Craton of Western Australia presents an intriguing case in this regard. The ancient (c. 3520 Ma) and extremely well-preserved ultramafic–mafic–felsic lavas of this craton have long been interpreted by some workers to have erupted onto an older continental substrate (Hickman, 1981, 1983, 1984, 2004; Van Kranendonk et al., 2002, 2007) or in rifts above or between older sialic crust (Glikson, 1972, 1979). Nowhere, however, has this putative crystalline basement been found to outcrop as a coherent tract of crust. Instead, the existence of ancient continental crust, perhaps older than 4000 Ma (Tessalina et al., 2010), is largely inferred from indirect evidence, such as trace element modelling (Green et al., 2000) and the Sm–Nd isotope systematics of basaltic (Van Kranendonk et al., 2007; Tessalina et al., 2010) and granitic rocks (Bickle et al., 1989, 1993;

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Smithies et al., 2003). Another line of evidence comes from Paleoarchean sedimentary rocks of the Pilbara Craton, which contain detrital zircons with crystallization ages that pre-date the most ancient exposed rocks in the area by up to 300 million years (Van Kranendonk et al., 2007). Zircon crystals significantly older than 3530 Ma also form xenocrysts in felsic igneous rocks of the craton (Thorpe et al., 1992), or discrete populations in gneissic enclaves in granitic complexes (McNaughton et al., 1988; GSWA 142870, Nelson, 1999). These zircons potentially provide a record, albeit fragmentary, of the involvement of much older crustal material in the evolution of the Pilbara Craton.

This study aims to unravel the earliest prehistory of the Pilbara Craton through Hf isotope analysis of a large selection of these >3530 Ma detrital and inherited zircons — particularly to test for the sampling of even more ancient, possibly Hadean, crust, as would be predicted by some recent continental growth models.

Pilbara Craton

Approximately 60 000 km^2 in area, the northern Pilbara Craton comprises three main lithotectonic elements, the East Pilbara Terrane (3530–3220 Ma), the West Pilbara Superterrane (3280–3070 Ma), and the De Grey Superbasin (3050–2930 Ma), clastic sedimentary rocks of which separate and unconformably overlie the older basement (Fig. 1; review by Hickman and Van Kranendonk, 2012). The Pilbara Craton is concealed in the southeast and southwest by volcanic and sedimentary rocks of the Fortescue (2770–2630 Ma) and Hamersley (2630–2450 Ma) Basins.

The East Pilbara Terrane is the archetypal Archean ‘granite–greenstone’ terrane (Van Kranendonk et al., 2007). It consists of large (35–120 km diameter), dome-like granitic complexes flanked by curvilinear belts of low- to medium-grade metavolcanic and metasedimentary rocks (‘greenstones’). The greenstone belt successions, the older parts of which are collectively known as the Pilbara Supergroup, encompass eight volcanic cycles of 5–30 million years duration that accumulated over c. 300 million years (3530–3220 Ma) (Hickman, 2011; Fig. 2). These cycles, which are largely stratigraphically coherent, are dominated by tholeiitic basalt although they also include komatiite, komatiitic basalt, and felsic volcanic rocks. Relatively thin sedimentary units (sandstone/conglomerate, chert, carbonate, and banded iron-formation) separate the volcanic cycles, typically above erosional unconformities (Hickman, 2012). By virtue of their detrital zircons, the sandstones are potentially important in revealing the ages of felsic crust exposed at different times during the Paleoarchean. This situation is equally true of detrital zircons in Archean sandstones, which were deposited in much larger sedimentary basins (see below).

Rhyolitic volcanic layers in the stratigraphically lowest unit, the Coonterunah Subgroup of the Warrawoona Group, have yielded zircon ^{207}Pb*/^{206}Pb* ages (Pb* indicates radiogenic Pb) of 3515 ± 3 Ma (Buick et al., 1995) and 3498 ± 2 Ma (GSWA 168995, Nelson, 2002). Zircon ages between c. 3530 and 3490 Ma are common in volcaniclastic sandstones of the Warrawoona Group, including the overlying Dresser Formation (Van Kranendonk et al., 2008). This clustering of ages is consistent with the earliest Warrawoona Group volcanic rocks erupting at c. 3530 Ma, although the base of the group is now everywhere in intrusive or tectonic contact against younger granitic rocks.

Granitic complexes of the East Pilbara Terrane are composite bodies that were constructed by multiple magmatic pulses over several hundred million years. Voluminous granite emplacement occurred at 3490–3420, c. 3300, 3240, and 2950–2930 Ma (Fig. 2). The oldest components may be represented by metre-sized tonalite gneiss enclaves contained by 3420–3240 Ma granodiorite and monzogranite of the Warrawagine Granitic Complex (Fig. 1). One of these enclaves contains zircon age components of c. 3660 Ma maximum age (see below). Zircon xenocrysts of similar antiquity are also present, albeit very sparsely, in younger felsic igneous rocks, such as in c. 3460 Ma rhyolite of the Duffer Formation (e.g. Thorpe et al., 1992). The presence of these older zircons is consistent with Sm–Nd isotope data, which imply that the granitic rocks of the East Pilbara Craton were derived from heterogeneous crustal sources with an average age typically >3500 Ma (Bickle et al., 1993; Smithies et al., 2009).

The West Pilbara Superterrane is a collage of three younger granite–greenstone terranes, the Karratha (3280–3250 Ma), Regal (c. 3200 Ma), and Sholl (3130–3110 Ma) Terranes (Fig. 2). These terranes are thought to have accreted to the East Pilbara Terrane at c. 3070 Ma during the Prinsep Orogeny (Van Kranendonk et al., 2010).

The Pilbara Craton was affected by two main periods of extension and basin formation. Initial rifting of the East Pilbara Terrane at c. 3200 Ma is interpreted to have led to separation of this terrane from a crustal fragment now represented by the Karratha Terrane (Van Kranendonk et al., 2010). This rifting was associated with deposition of the Soanesville Group (c. 3190 Ma) clastic succession (up to 3.5 km thick), overlain by an extensive basaltic volcanic succession. The second extensional period followed amalgamation of the East Pilbara Terrane and West Pilbara Superterrane, where sedimentary and volcanic rocks of the De Grey Supergroup (3050–2930 Ma) accumulated across the Pilbara Craton. Major depositional basins include the Gorge Creek Basin (3050–3015 Ma), which covered the whole craton, and three younger (3010–2930 Ma) coeval basins, the Whim Creek, Mallina, and Mosquito Creek Basins (Figs 1 and 2). The Mallina Basin occupies a northeasterly trending structural corridor between the East Pilbara Terrane and West Pilbara Superterrane, and the volcanic Whim Creek Basin is confined to the northwest margin of the Mallina Basin. The Mosquito Creek Basin is lithologically and tectonically similar to the Mallina Basin although it separates the East Pilbara Terrane from the Kurrana Terrane, a postulated rift fragment of the East Pilbara Terrane to the southeast (Fig. 1).

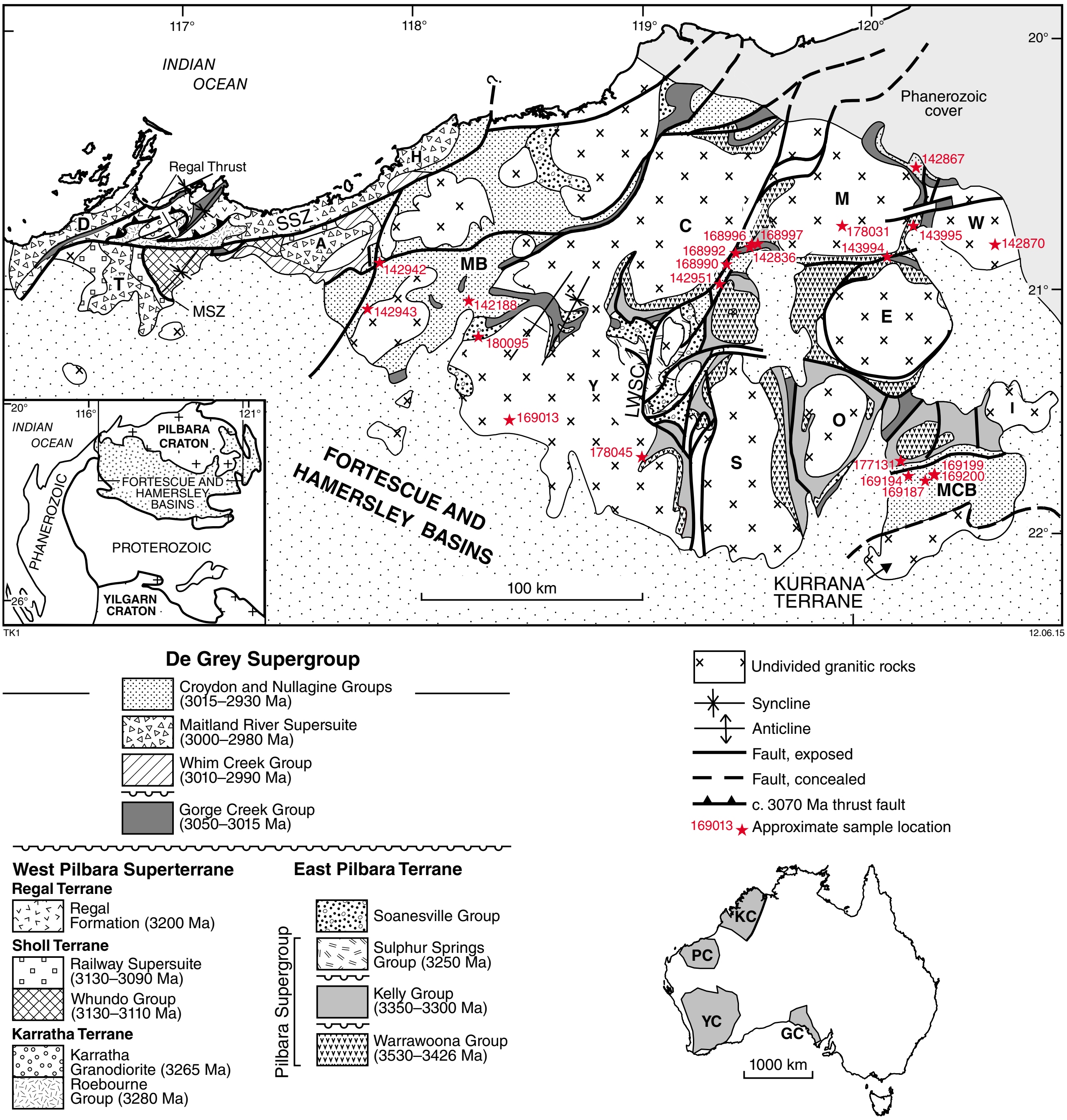

Figure 1. Simplified geological map of the exposed Pilbara Craton (modified after Smithies et al., 2007), showing the location of samples containing >3.53 Ga zircons. Zircons from most samples were analysed for Hf isotopes during this study. Tectonic units and features: LWSC, Lalla Rookh – Western Shaw structural corridor; MB, Mallina Basin; MCB, Mosquito Creek Basin; MSZ, Maitland Shear Zone; SSZ, Sholl Shear Zone. Granitic complexes: C, Carlindi; D, Dampier; E, Mount Edgar; H, Harding; I, Yilgalong; M, Muccan; O, Corunna Downs; S, Shaw; W, Warrawagine; Y, Yule. GC, Gawler Craton; KC, Kimberley Craton; PC, Pilbara Craton; YC, Yilgarn Craton.

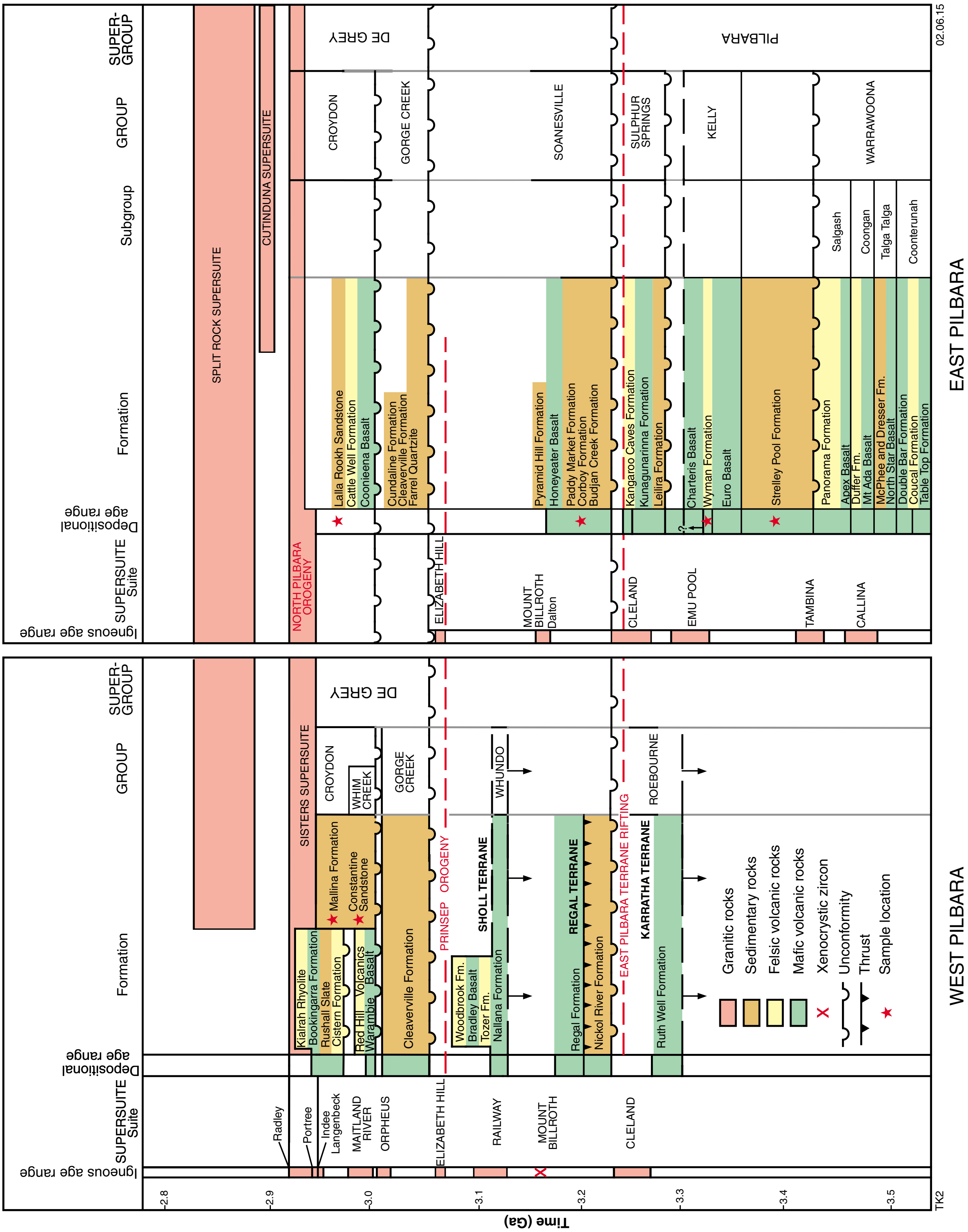

Figure 2. **Time–space diagram summarizing the evolution of the East Pilbara Terrane and the West Pilbara Superterrane (modified after Hickman et al., 2010). Stars indicate metasedimentary units from which detrital zircons have been analysed for Hf isotopes as part of this study.**

Samples

This study targeted zircons with common-lead-corrected $^{207}Pb^*/^{206}Pb^*$ ages in excess of c. 3530 Ma, as previously identified by the sensitive high-resolution ion microprobe (SHRIMP) geochronology program of the Geological Survey of Western Australia (see Table 1 for a summary of the samples used for Hf analysis). For convenience, these zircons are collectively referred to (below) as 'pre-Pilbara Supergroup zircons'. Most of the anomalously old detrital zircons in this sample set (Appendices 1 and 2a) exist as detrital grains in: (1) sandstone of the Strelley Pool Formation, deposited between c. 3430 and 3350 Ma above a major, East Pilbara Terrane – wide unconformity at the top of the Warrawoona Group; (2) volcaniclastic sandstone interstratified with rhyolites of the c. 3300 Ma Wyman Formation (Kelly Group) in the Warralong greenstone belt of the East Pilbara Terrane; (3) sandstones of the Corboy Formation (basal Soanesville Group) deposited in the c. 3190 Ma Soanesville Basin; and (4) sedimentary rocks of the 3050–2930 Ma De Grey Supergroup, as sampled from the Mallina Basin (Croydon Group) and Mosquito Creek Basin (Nullagine Group) (Fig. 1). The present study analysed 84 of these zircons (Appendix 2a), selected mainly on the basis of greatest age. It is important to note that an additional 90 old detrital zircons identified by the GSWA SHRIMP program, mainly dated between c. 3600 and 3530 Ma (Appendix 2b), were not analysed for Hf isotopes in this study. These additional 90 zircon grains nevertheless provide further information on the time and space distribution of pre-Pilbara Supergroup material. Sample information is summarized in Table 1 and Appendix 1 and the locations of the sampling sites are shown in Figure 1.

Also included in this study are >c. 3530 Ma zircons hosted by two igneous rocks, a banded tonalite gneiss enclave from the Warrawagine Granitic Complex (Figs 1 and 3), and the c. 3250 Ma Wolline Monzogranite of the adjacent Muccan Granitic Complex (Fig. 1).

Analytical techniques

The zircons analysed for Hf isotopes in this study (n = 108) were previously dated using a SHRIMP II at Curtin University, Perth, Western Australia, following analytical techniques described by Nelson et al. (2000) and Wingate and Kirkland (2015). The (U–Th)–Pb isotope data are available online at <www.dmp.wa.gov.au/geochron> and are reproduced in Appendix 2. All $^{207}Pb^*/^{206}Pb^*$ zircon ages referred to in this study have been corrected for common lead by the 204-method (denoted $^{207}Pb^*/^{206}Pb^*$, see Appendix 2). Monochromatic cathodoluminescence (CL) images were acquired for all zircons prior to Hf isotope analysis using a JEOL JSM5410LV scanning electron microscope with a Robinson CL detector in the Advanced Analytical Centre, James Cook University, Townsville, Queensland. Such images were used to verify SHRIMP spot placement relative to internal growth zoning.

The Hf isotope data were obtained with a Thermo-Scientific Neptune multi-collector inductively coupled plasma – mass spectrometer (ICP-MS) and Coherent GeoLas 193 nm ArF laser microsampling system with custom-built low-volume ablation cell in the Advanced Analytical Centre, James Cook University. Analytical protocols for this laboratory are described by Kemp et al. (2009) and Naeraa et al. (2012). Measurements were undertaken on polished, sectioned zircon grains mounted in epoxy resin, as guided by CL and transmitted light images and U–Pb age data, and by ablating the shallow pits generated by the preceding SHRIMP analysis. Sample zircons with <5% U–Pb age discordance (expressed as the percentage difference between the $^{238}U/^{206}Pb^*$ and $^{207}Pb^*/^{206}Pb^*$ dates) were preferentially targeted for Hf isotope analysis, although a few moderately to strongly discordant grains (which typically have elevated U contents) were analysed to investigate any differences in Hf isotope composition.

The laser was operated in energy constant mode, with output energy of 100 mJ. This laser was tuned to a fluence of 5–6 J/cm^2 using an optical attenuator in the beam path and an energy meter at the ablation site. This fluence translates into an ablation rate of 0.2 – 0.3 µm/s in zircon. A laser pulse rate of 4 Hz and beam diameter of 31, 42, or 58 µm was used, as dictated by the size of the targeted zircon growth phase. Care was taken to ensure that the analytical site was within the same CL-defined growth domain from where the age data were acquired. Ablation was carried out in He (about 0.8 L/min), which was combined with Ar (about 0.9 L/min) and a small (about 0.005 L/min, optimized daily) N_2 flow prior to transport into the ICP-MS. Each analysis consisted of a 30-s baseline ('on peak zero') followed by a 60-s ablation period, comprising 60 cycles of 1-s integration time. The masses of ^{171}Yb, ^{173}Yb, ^{175}Lu, $^{176}(Yb+Lu+Hf)$, ^{177}Hf, ^{178}Hf, and ^{180}Hf were measured in static mode on Faraday cups; amplifier gains and electronic baselines were calibrated daily. The isobaric interference of Lu and Yb on ^{176}Hf was corrected by monitoring the interference-free ^{171}Yb and ^{175}Lu intensities during the analysis, and then deriving ^{176}Yb and ^{176}Lu using $^{176}Yb/^{171}Yb$ = 0.897145 (Segal et al., 2003) and $^{176}Lu/^{175}Lu$ = 0.02655 (Vervoort et al., 2004). Yb isotope ratios were normalized to $^{173}Yb/^{171}Yb$ = 1.130172 (Segal et al., 2003) and Hf isotope ratios to $^{179}Hf/^{177}Hf$ = 0.7325 using an exponential mass bias law. The mass bias of Lu was assumed to follow that of Yb. Data were processed offline using a customized Microsoft Excel spread sheet. Within-run outlier rejection was set at three standard errors of the mean.

Reference zircons of variable (Yb+Lu)/Hf values analysed throughout this study yielded mean $^{176}Hf/^{177}Hf$ values identical to those determined by solution analysis (Table 2, Appendix 3). The Yb-doped synthetic zircons of Fisher et al. (2011) were analysed as an additional quality control measure, yielding 0.282133 ± 0.000013 (2σ) for the Yb-free zircon, and 0.282127 ± 0.000021 (2σ) for the Yb-rich zircon ($^{176}Yb/^{177}Hf$ up to about 0.35).

Table 1. Details of samples from which zircons have been analysed for Hf isotopes as part of this study

Sample ID	*Zone*	*MGA coordinates*	*Stratigraphic age (Ma)*	*Group*	*Stratigraphic unit*	*Basin/greenstone belt*	*Lithology*
169199	51	222470E 7590790N	2980–2930	Nullagine	Mosquito Creek Fm	Mosquito Creek Basin	Lithic quartz sandstone
169200	51	222160E 7590280N	2980–2930	Nullagine	Mosquito Creek Fm	Mosquito Creek Basin	Lithic quartz metasandstone
178010	51	250130E 7615210N	2980–2930	Nullagine	Mosquito Creek Fm	Mosquito Creek Basin	Lithic quartz sandstone
142943	50	583800E 7664200N	c. 2980	Croydon	Constantine Sandstone	Mallina Basin	Quartz-rich metasandstone
142942	50	588900E 7685400N	c. 2980	Croydon	Constantine Sandstone	Mallina Basin	Quartz-rich metasandstone
142188	50	629470E 7667280N	2980–2930	Croydon	Mallina Fm	Mallina Basin	Subarkosic sandstone
142951	50	744865E 7675430N	2980–2940	Croydon	Lalla Rookh Sandstone	Lalla Rookh Basin	Feldspathic sandstone
178045	50	709240E 7597160N	c. 3190	Soanesville	Corboy Fm	Soanesville Basin	Feldspathic sandstone
180095	50	634435E 7651601N	3520–3190	Uncertain	Unassigned	Soanesville Basin	Volcaniclastic sandstone
169013	50	647130E 7613390N	c. 3190	Soanesville	Corboy Fm	Soanesville Basin	Feldspathic metasandstone
168992	50	752010E 7689400N	c. 3325	Kelly	Wyman Fm	Warralong Greenstone Belt[a]	Volcaniclastic metasandstone
142836	50	758500E 7692175N	3430–3350	Pilbara Supergroup	Strelley Pool Fm	Warralong Greenstone Belt[a]	Volcaniclastic sandstone
178031	50	800060E 7701040N	3250 ± 0.2	Cleland Supersuite	Wolline Monzogranite	Muccan Granitic Complex[a]	Granodiorite
142870	51	246941E 7696753N	3660–3420	Tambina Supersuite	n/a	Warrawagine Granitic Complex[a]	Tonalite gneiss

NOTES: (a) East Pilbara Terrane
n/a not applicable

These values overlap those of purified hafnium solutions from these zircons (0.282135 ± 0.000007, Fisher et al., 2011) and attest to the veracity of the ^{176}Yb interference correction on ^{176}Hf. All zircon Hf isotope data are normalized to the solution ^{176}Hf/^{177}Hf value of Mud Tank zircon (0.282507 ± 0.000006, Woodhead and Hergt, 2005, reported relative to JMC475 ^{176}Hf/^{177}Hf = 0.282160) using the laser-ablation data generated from this zircon in each analytical session. Adjustments were less than 0.5 ε units, except for the January 2014 session (0.9 ε units). This procedure effectively normalizes the laser data to the standard JMC475 Hf isotope solution. Analytical uncertainties combine the in-run error with the reproducibility of Mud Tank zircon analyses from the same session, added in quadrature. Hf isotope data for the sample zircons are listed in Appendix 4.

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Figure 3. Banded tonalite gneiss enclave enclosed by 3410 Ma and 3240 Ma granitic rocks of the Warrawagine Granitic Complex, East Pilbara Craton (GSWA 142870)

Table 2. Summary of Lu–Hf–Yb isotope data measured during this study from natural and synthetic reference zircons by laser-ablation MC-ICPMS

Zircon	*n*	*^{176}Hf/^{177}Hf $^{(a,b)}$*	*^{176}Lu/^{177}Hf*	*^{176}Yb/^{177}Hf*	*$\varepsilon_{Hf}(t)^{(c)}$*	*Solution values*		*Reference$^{(d)}$*
						^{176}Hf/^{177}Hf$^{(b)}$	*^{176}Lu/^{177}Hf*	
CZ3	6	0.281721 ± 0.000020	0.00003	0.0011	−25.1 ± 0.7	0.281732 ± 0.000007	–	1
Temora 2	13	0.282689 ± 0.000018	0.00127	0.0346	5.5 ± 0.5	0.282686 ± 0.000008	0.00109	2
FC1	70	0.282184 ± 0.000021	0.00114	0.0358	2.6 ± 0.6	0.282184 ± 0.000016	0.00126	2
R33	43	0.282763 ± 0.000030	0.00263	0.0798	8.2 ± 0.8	0.282761 ± 0.000006	0.00207	3
Synthetic 1	13	0.282133 ± 0.000013	0.00001	0.0001	−23.1 ± 0.5	0.282135 ± 0.000007	–	4
Synthetic 2	21	0.282127 ± 0.000021	0.01101	0.2447	−23.3 ± 0.8	0.282135 ± 0.000007	–	4

NOTES: (a) Hf isotope ratios are session-normalized to Mud Tank zircon ^{176}Hf/^{177}Hf = 0.282507 (Woodhead and Hergt, 2005), which is itself reported relative to JMC475 ^{176}Hf/^{177}Hf = 0.282160. Full analyses are given in Appendix 3.

(b) Uncertainties represent two standard deviations of the mean.

(c) Values are calculated at the inferred crystallization age of the zircon (zero Ma for the synthetics).

(d) Corresponding solution data are from: 1 Wu et al. (2006); 2 Woodhead and Hergt (2005); 3 Vervoort (2010); 4 Fisher et al. (2011), all reported relative to JMC475 ^{176}Hf/^{177}Hf = 0.282160.

Results

Zircon microstructures and U–Pb ages

The detrital zircons analysed for Hf isotopes in this study (n = 84) exhibit a variety of microstructures in CL images (Fig. 4). Most show well-developed oscillatory zoning (60% of studied zircons), although sector zoning, planar banding, and more complex irregular microstructures are also observed. In the sections below, weighted mean ages are quoted with 95% confidence intervals.

The apparent $^{207}Pb^*/^{206}Pb^*$ dates of these detrital zircons range from c. 3800 to 3550 Ma (Fig. 5). Most analyses are concordant within analytical uncertainty to slightly normally discordant. There is no correlation between degree of concordance and $^{207}Pb^*/^{206}Pb^*$ date, and pronounced discordance tends to be accompanied by increasing actinide contents, as typifies Archean zircons (Pidgeon et al., 1966; Fig. 6). This evidence, together with the distribution of data on the concordia diagram, is consistent with recent radiogenic lead loss from variably radiation-damaged zircon. Typically, a given sedimentary rock from this study contains detrital zircons with a spread of apparent $^{207}Pb^*/^{206}Pb^*$ ages >3520 Ma, which could indicate contributions from a range of ancient sources. This age range appears independent of the depositional age or spatial location of the sedimentary host (Fig. 7). Discrete zircon age components are, however, resolved in some samples. The oldest sedimentary unit, GSWA 142836 from the Strelley Pool Formation (c. 3420 Ma), has 24 variably discordant oscillatory zoned zircons that define a weighted mean $^{207}Pb^*/^{206}Pb^*$ age of 3602 ± 5 Ma (mean square of the weighted deviates, MSWD = 1.8), suggesting a single-age Archean source for this rock. An identical age peak is prominent in the second-oldest unit (GSWA 168992), a c. 3300 Ma volcaniclastic sandstone from the Warralong greenstone belt.

Zircons from the igneous host rocks are shown in Figure 8. The Wolline Monzogranite contains a single c. 3620 Ma xenocrystic core mantled by oscillatory zoned, weakly luminescent 3250 Ma zircon interpreted to date crystallization of the monzogranite. The core is rounded, with a relatively low CL response, and shows sector zoning that is sharply truncated by the enclosing zircon. Zircon age systematics are more complex in the tonalite gneiss (Fig. 5c, as befits the lithological heterogeneity of the rock. Analyses cluster into three broad $^{207}Pb^*/^{206}Pb^*$ age components, at 3660–3630 Ma (Group 1, n = 7), 3610–3570 Ma (Group 2, n = 8), and c. 3410 Ma (Group 3, n = 7)). The oldest ages (Group 1) are from cores that exhibit well-developed, fine oscillatory zoning. Group 2 analyses are also from cores, and although some of these have oscillatory zoning, a number are dark in CL images and exhibit planar or less-regular banding; one relatively high-U core (grain #11; 346 ppm U, Th/U 0.94, $^{207}Pb^*/^{206}Pb^*$ date 3580 Ma) shows patchy zoning (Fig. 8). Cores tend to be elongate with rounded ends, although equant or irregular shapes are also present. Analyses from Group 3 are all from fairly thin, mostly oscillatory zoned overgrowths on cores. These rims sharply truncate the zoning in the cores, and are commonly separated from the cores by a very thin shell of luminescent zircon (e.g. grains #8 and #10, Fig. 8). The youngest analysis (3370 Ma) is from a cracked tip of an elongate grain with vague concentric banding around a core that is relatively dark in the CL image (grain #19, Fig. 8).

Hf isotope compositions

Figure 9 summarizes the Hf isotope composition of the pre-Pilbara Supergroup zircons. In no case was there any evidence from time-resolved signals for intragrain isotope variation that could arise from intersection of different age components during ablation. Most data from the detrital zircons cluster between Hf = +1.0 and ε_{Hf} = –1.5, with a few analyses dispersed to slightly higher and lower values. The single analysis from the 3620 Ma xenocryst in the Wolline Monzogranite falls within the detrital zircon cluster (ε_{Hf} = –1). The c. 3600 Ma zircons from the Strelley Pool Formation sandstone have very similar Hf isotope compositions (ε_{Hf} = +0.5 to –1.2), irrespective of the degree of age discordance. These grains were plausibly eroded from a single, homogeneous igneous source rock, followed by modern disturbance to their Pb–U systems. The three oldest detrital zircons (c. 3800, 3710, and 3680 Ma) are among the least radiogenic relative to chondritic uniform reservoir (CHUR) (ε_{Hf} = –2.4, –2.4, and –2.8, respectively).

There appears to be no systematic relationship between zircon Hf isotope composition and the depositional age of the host sedimentary rock, although isotope fine structure at sub-ε unit level is not resolvable with the existing in situ dataset. It is notable, however, that >3600 Ma detrital zircons from the c. 3300 Ma volcaniclastic sandstone of the Warralong greenstone belt (GSWA 168992) are slightly more radiogenic than zircons comprising the main ε_{Hf} – time cluster, and extend to the highest ε_{Hf} (+0.8 at c. 3670 Ma, average +0.2 ± 0.9 2σ) (Fig. 10). Conversely, >c. 3600 Ma zircons from GSWA 180095 (c. 3190 Ma depositional date) plot at the lower bounds of the detrital age–Hf isotope envelope (average ε_{Hf} = –1.3 ± 0.3 2σ, Fig. 10). Zircons from these samples could therefore have originated from isotopically different Archean source components. Higher precision solution Hf isotope data would be required to explore this option.

Hf isotope data from zircons of the Warrawagine tonalite gneiss define three tight clusters (Fig. 9), corresponding to the different age components in this rock. The Group 1 cores (3660–3630 Ma) have the highest ε_{Hf} (ε_{Hf} ≈ 0 to +1), and these values are slightly lower in the Group 2 cores (3610–3570 Ma, ε_{Hf} = –0.5 to –1.0). Both age–ε_{Hf} components plot within the broad cluster defined by the detrital zircons. Analyses of younger zircons are more scattered, with the Group 3 (c. 3410 Ma) rim domains approximately chondritic.

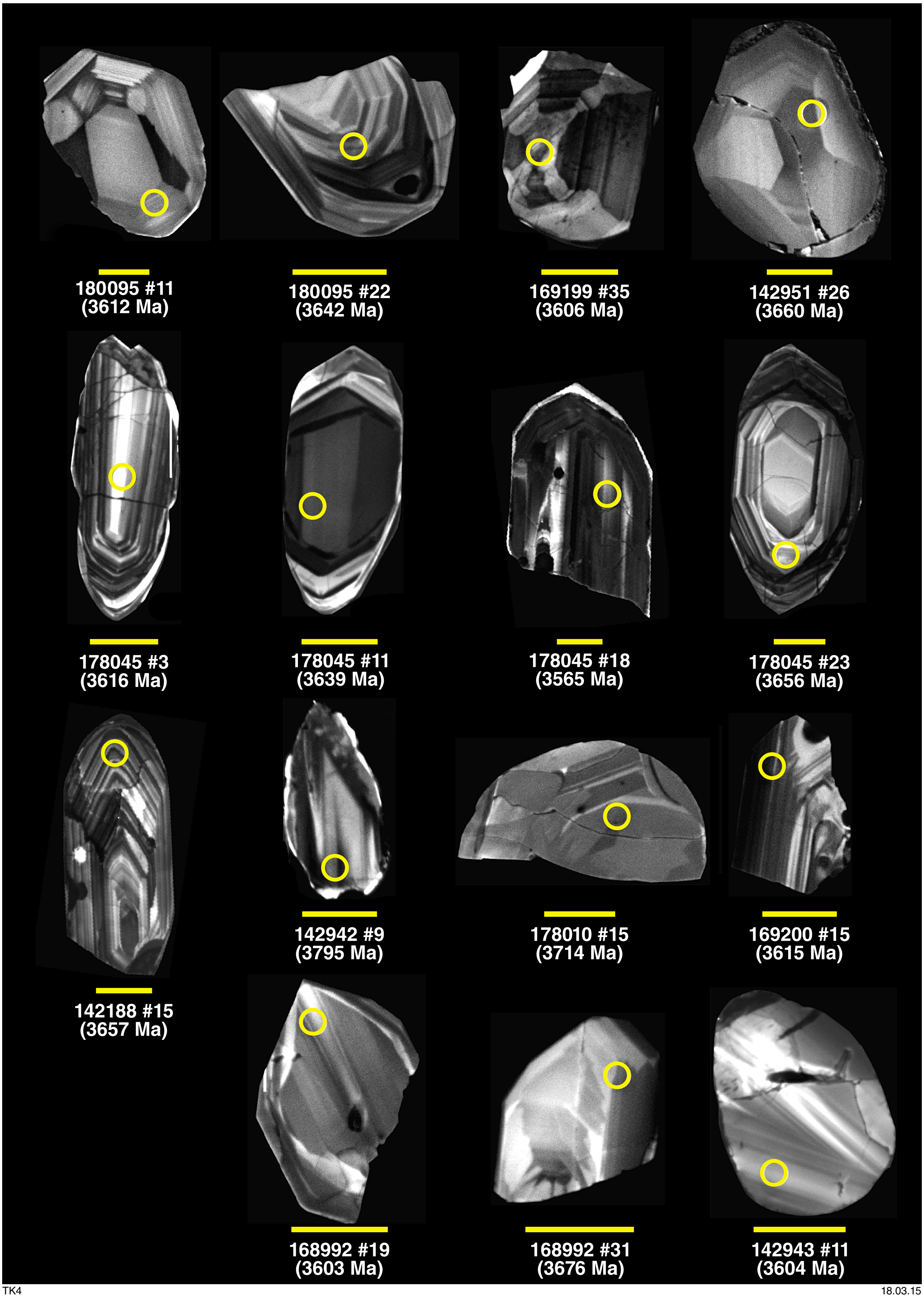


Figure 4. Selected cathodoluminescence (CL) images of pre-Pilbara Supergroup detrital zircons. The analytical site (open circle) and corresponding $^{207}Pb^{*}/^{206}Pb^{*}$ age for each zircon are indicated. The scale bar for each grain is 50 µm.

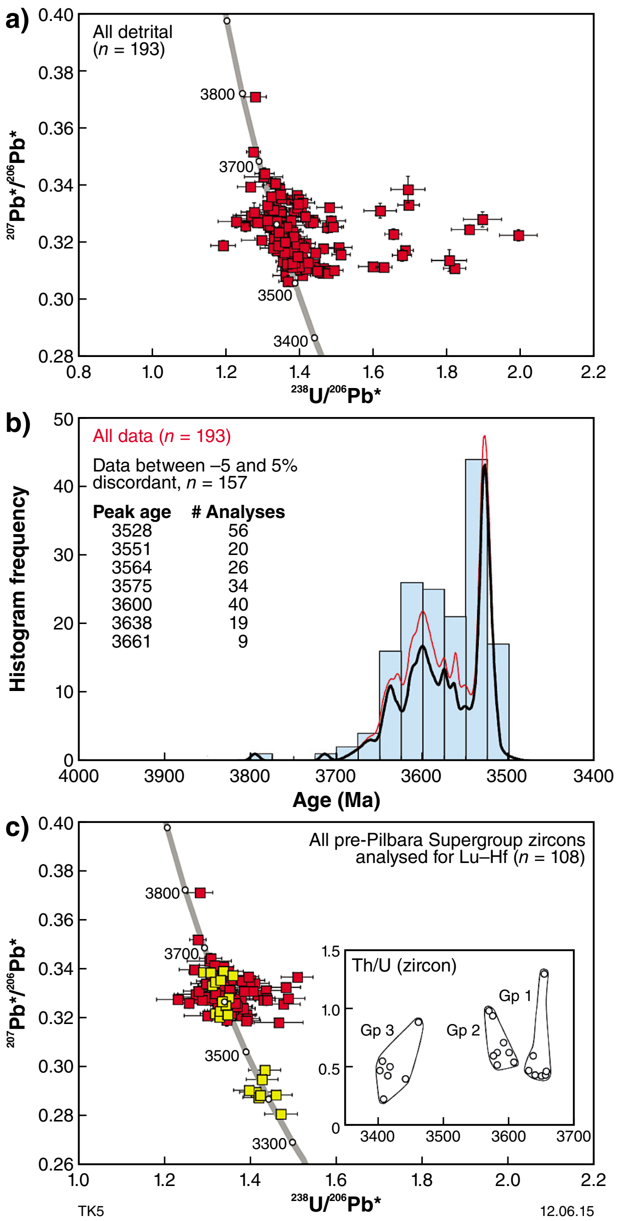


Figure 5. a) Tera–Wasserburg concordia diagram showing data for all pre-Pilbara Supergroup zircons (i.e. those with $^{207}Pb^*/^{206}Pb^*$ ages >3530 Ma) in the studied metasedimentary samples; b) probability density diagram and histogram of $^{207}Pb^*/^{206}Pb^*$ ages for all pre-Pilbara Supergroup zircons. The thick curve, maxima values, and frequency histogram (bin width 25 Ma) include only accepted data (discordance <5%), whereas the thin red curve is based on all data; c) ion microprobe U–Pb isotope systematics of the pre-Pilbara Supergroup detrital zircons analysed for Hf isotopes, also showing the analysed zircons from the Warrawagine tonalite gneiss (yellow). Most of the discordant analyses are from sample GSWA 142836. All plots were constructed using Isoplot Ex (Ludwig, 2001).

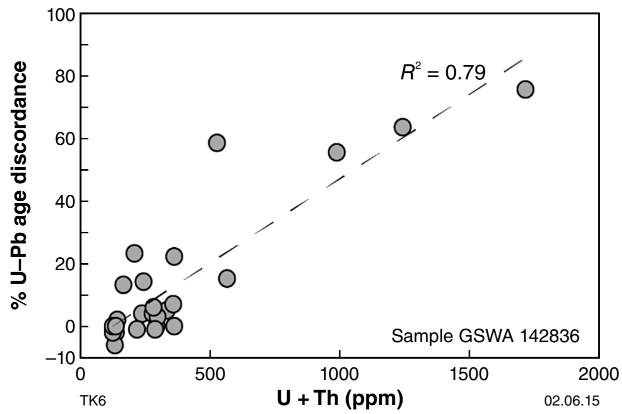


Figure 6. U–Pb age discordance vs combined U and Th contents for zircons from sample GSWA 142836

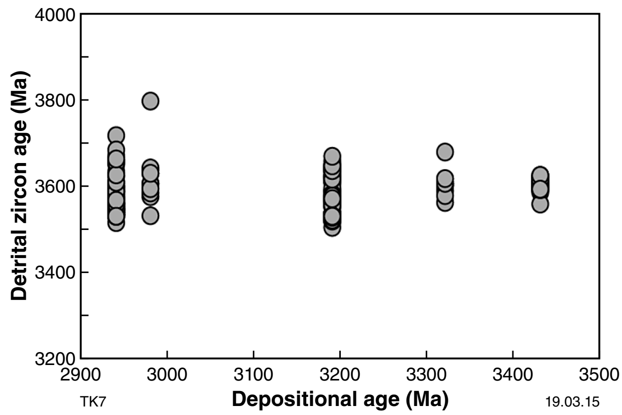
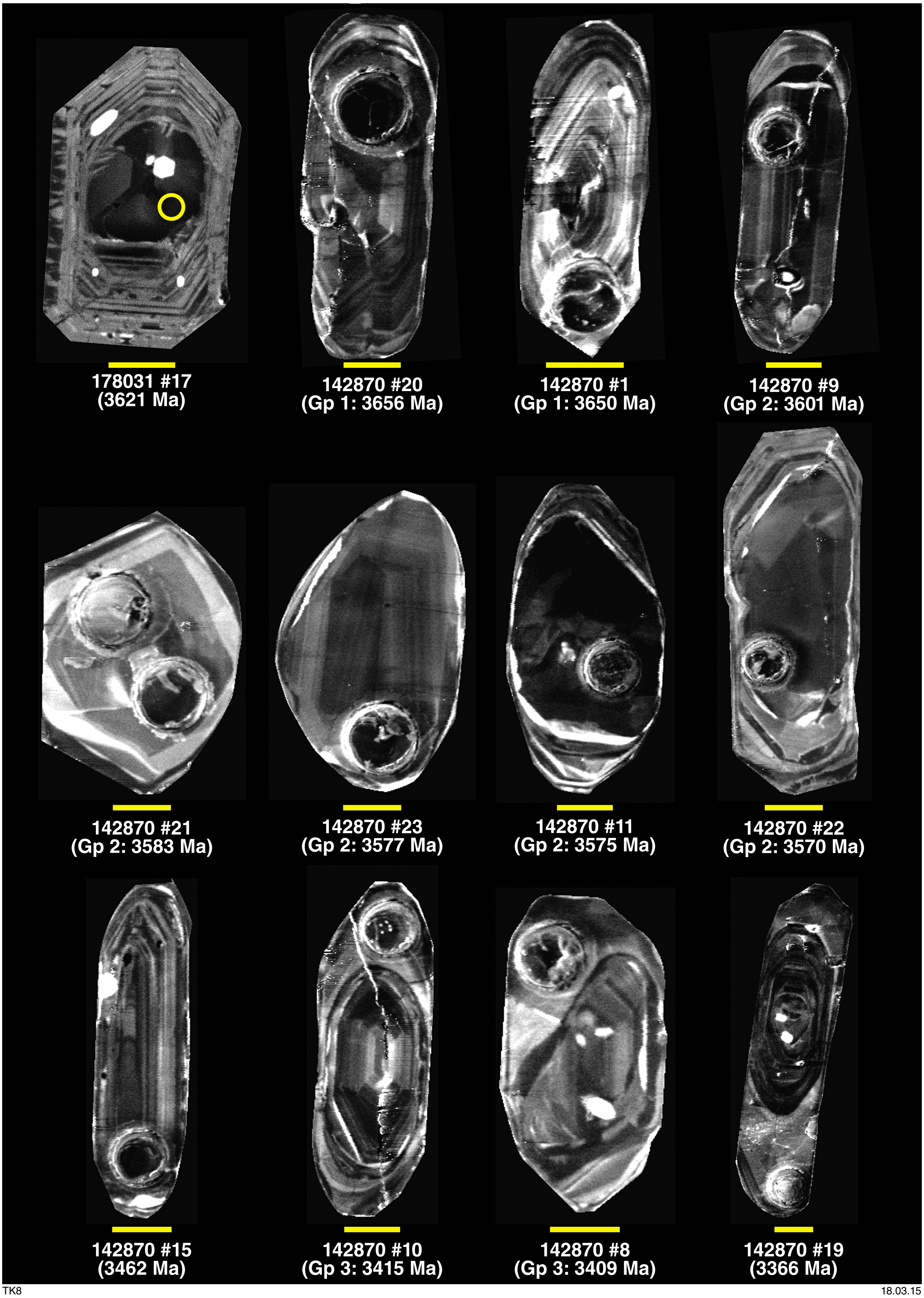


Figure 7. Detrital zircon ages vs depositional age of the host sedimentary rock for the studied samples. Only pre-Pilbara Supergroup zircons (i.e. those older than 3530 Ma) are shown on this diagram.

Figure 8. (right) Cathodoluminescence (CL) images of igneous-hosted zircons analysed in this study, including the inherited core in Wolline Monzogranite (GSWA 178031) and zircons from the Warrawagine tonalite gneiss (GSWA 142870). Laser-ablation pits resulting from Hf isotope analysis are sited directly over the SHRIMP analytical sites. The scale bar shown for each grain is 50 µm.


178031 #17
(3621 Ma)
142870 #20
(Gp 1: 3656 Ma)
142870 #1
(Gp 1: 3650 Ma)
142870 #9
(Gp 2: 3601 Ma)
142870 #21
(Gp 2: 3583 Ma)
142870 #23
(Gp 2: 3577 Ma)
142870 #11
(Gp 2: 3575 Ma)
142870 #22
(Gp 2: 3570 Ma)
142870 #15
(3462 Ma)
142870 #10
(Gp 3: 3415 Ma)
142870 #8
(Gp 3: 3409 Ma)
142870 #19
(3366 Ma)
TK8
18.03.15

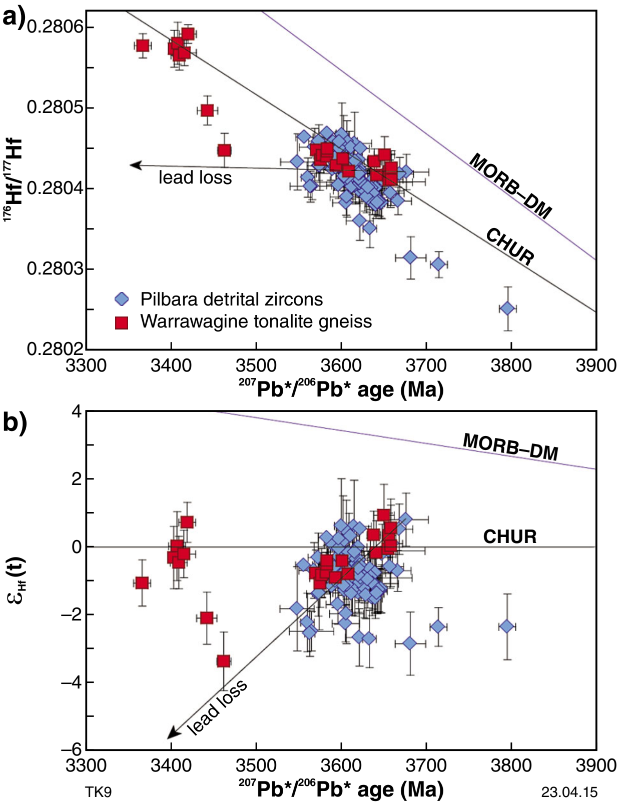


Figure 9. **Hf isotope composition of pre-Pilbara Supergroup zircons analysed in this study: a) $^{176}Hf/^{177}Hf$ vs $^{207}Pb^{*}/^{206}Pb^{*}$ age. The arrowed line on both panels shows the trajectory expected from ancient lead loss. MORB-DM, average composition of modern MORB extrapolated to $\varepsilon_{Hf} = 0$ at 4500 Ma; b) ε_{Hf} vs $^{207}Pb^{*}/^{206}Pb^{*}$ age. The steep ancient lead loss trajectory reflects the rapid isotopic evolution of the chondritic uniform reservoir (CHUR) reference ($^{176}Lu/^{177}Hf$ = 0.0336) relative to that of zircon ($^{176}Lu/^{177}Hf \approx 0.001$). The CHUR parameters of Bouvier et al. (2008) and the ^{176}Lu decay constant of Söderlund et al. (2004) were used.**

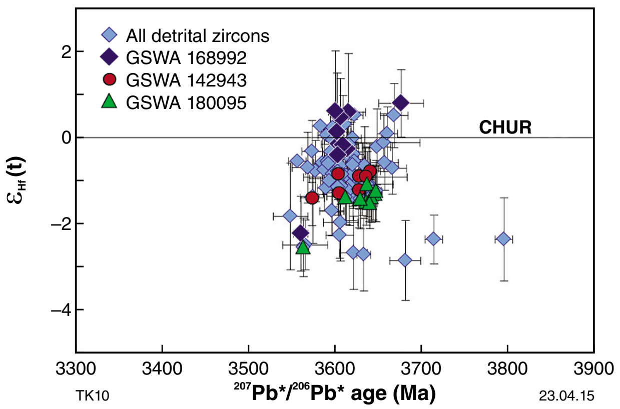


Figure 10. **Comparison of the Hf isotope compositions of detrital zircons from GSWA 168992 of the Warralong greenstone belt (stratigraphic age c. 3320 Ma) with those from GSWA 180095 of the Soanesville Basin (c. 3190 Ma) and GSWA 142943 of the Mallina Basin (3010–2930 Ma)**

Discussion

Source of detrital zircons

Before exploring the implications of the Hf isotope data it is important to consider the limitations of detrital zircons as a record of crustal evolution. The obvious drawback is that because the original geological context of each zircon is lost, the nature, composition, and location of the source rock are matters of deduction. Zircon, as a mechanically and chemically robust detrital mineral, may be far-travelled and/or recycled through multiple episodes of erosion and sedimentary transport — hence, there is no guarantee that detrital zircons in younger sedimentary rocks relate in any way to underlying crystalline basement, or provide an unequivocal record of when older basement rocks were exposed at the Earth's surface. For example, the age spectra of detrital zircons in Ordovician and Silurian metasedimentary rocks of the Paleozoic Lachlan Orogen (southeastern Australia) show a prominent peak at c. 1100 Ma, with individual zircons at c. 3300 Ma maximum age (e.g. Kemp et al., 2005), yet this orogen is floored by Cambrian oceanic crust (Gray and Foster, 2004). On the other hand, there are instances where a direct relationship can be inferred between the detrital zircons in sandstone and the magmatic zircons in nearby igneous rocks. The c. 3050 Ma Farrel Quartzite of the Gorge Creek Group contains a dominant group of c. 3430 Ma detrital zircons (GSWA 143996, Nelson, 1998) and granodiorite of the Tambina Supersuite, which directly underlies the sandstone beneath an erosional unconformity, was dated at c. 3440 Ma (Dawes et al., 1995; GSWA 124755, Nelson, 1996; Williams, 1999).

Although the precise sources of >c. 3530 Ma detrital zircons in metasedimentary units of the Pilbara Craton cannot be conclusively determined, several lines of evidence suggest that the provenance might have been relatively local. Sources of evidence include sedimentary facies (e.g. immature, poorly sorted, feldspathic or lithic sandstone, commonly interbedded with boulder conglomerate suggest local provenance), tectonic setting (e.g. proximity to growth-faults in areas of crustal extension indicate derivation of detritus from local erosion), paleocurrent data (which can provide evidence of deposition in high-energy alluvial or submarine fans), or location directly above erosional unconformities. Several sedimentological studies in the Pilbara Craton have provided interpretation of depositional environments and tectonic settings (Eriksson, 1981, 1982; Buick and Barnes, 1984; Krapez, 1984; Wilhelmij and Dunlop, 1984; DiMarco and Lowe, 1989a,b; Allwood et al., 2006; Wacey et al., 2010). Most of these studies have concluded that the clastic sediments were locally derived.

Many of the sandstones are immature and poorly sorted, indicative of limited detrital transport. For example, sample GSWA 178045 (Corboy Formation) was collected from a 20 m-thick coarse feldspathic sandstone unit that directly overlies undated banded granitic gneiss of the eastern Yule Granitic Complex. Twenty-one of the 35 zircons dated from this sample are older than c. 3530 Ma, and 10 are older than c. 3600 Ma

(3656–3605 Ma). The feldspathic composition of the sandstone, and its erosional contact with the underlying gneiss, suggest local provenance and therefore the possible existence of outcropping c. 3650 Ma gneiss in this part of the Yule Granitic Complex. The tectonic setting of this formation (Appendix 1) is also significant because the Corboy Formation was deposited in a passive margin basin following c. 3200 Ma uplift and rifting of the East Pilbara Terrane (Van Kranendonk et al., 2010). Another sample interpreted to have local provenance is GSWA 142951, from the Lalla Rookh Sandstone in the central section of a small intracratonic rift basin. Sedimentological studies have indicated that the clastic sediments in this basin were derived from directly adjacent granites and greenstones (Krapez, 1984). The Lalla Rookh Sandstone contains numerous units of polymictic conglomerate interbedded with feldspathic sandstone that were deposited in alluvial fans and talus slopes adjacent to active growth-faults. Several >c. 3600 Ma zircons (as old as c. 3660 Ma) were dated in this sample, which might indicate exposed 3660–3600 Ma crust in the Pilbara Craton at c. 2950 Ma. Another conglomerate–sandstone unit related to rifting is represented by GSWA 178010 (Nelson, 2005) from the Mosquito Creek Basin. This sample contains old zircon grains dated between c. 3714 and 3548 Ma, whereas younger zircon groups in the same sample have ages of 3461 ± 7, 3424 ± 12, and 3294 ± 14 Ma (Nelson, 2000). The ages of these three groups indicate derivation from the Callina, Tambina, and Emu Pool Supersuites (or their volcanic equivalents), respectively, which outcrop in the East Pilbara Terrane to the north of the basin margin. Thus, the pre-3530 Ma zircons in this sample may also came from the East Pilbara Craton north of the Mosquito Creek Basin (Bagas et al., 2005).

The oldest clastic unit known to contain pre-Pilbara Supergroup zircons is the 3470–3460 Ma Duffer Formation. Sample GSWA 168996, collected from the Gorge Range between Port Hedland and Marble Bar, contains six zircons dated between c. 3570 and 3560 Ma (Appendix 2b). The rock is a feldspathic sandstone that overlies polymictic conglomerate containing boulders of chert up to 20 cm in diameter. This facies indicates local derivation of the clastic or volcaniclastic material in the unit. The tight cluster of ages suggests a single source, possibly in the Carlindi Granitic Complex directly to the northwest.

The clastic sedimentary samples used in this study also contain zircons younger than c. 3530 Ma and these contain zircon age components consistent with erosion of the granitic supersuites that make up the domed complexes. Progressive uplift of the domes through several events (Hickman and Van Kranendonk, 2004) is estimated to have exceeded 20 km, and so could have exposed sections of the pre-3530 Ma crust.

Together, the geological context of the studied metasedimentary samples is consistent with a proximal source for most >c. 3600 Ma zircons in the Pilbara Craton, rather than ad hoc introduction of detrital grains by long-range sedimentary transport. Moreover, although the overall $^{207}Pb^{*}/^{206}Pb^{*}$ date range >c. 3600 Ma is limited, the relationships between these ages and the accompanying Hf isotope data hint at the sampling of isotopically distinct crustal sources in different places at different times, and this could also imply a heterogeneous local source.

An additional piece of evidence in support of local derivation is that zircons of similar antiquity and Hf isotope composition to the >c. 3600 Ma detrital zircons form xenocrystic cores in igneous rocks of the Pilbara Craton (e.g. the c. 3621 Ma inherited grain in the c. 3240 Ma Wolline Monzogranite). The source of these old zircons was sampled by magmas from at least c. 3460 Ma onwards (Thorpe et al., 1992). Notwithstanding this, the possibility remains that the >c. 3600 Ma inherited zircons, and the detrital zircons in the Archean sedimentary units, were not simply recycled from older sedimentary units in the greenstone belts but genuinely reflect the existence of crystalline basement at the time of magmatism or sedimentary deposition. However, it is important to emphasize that gneissic enclaves with zircons older than c. 3500 Ma have been documented from granitic complexes of the East Pilbara Craton (e.g. the Warrawagine tonalite gneiss GSWA 142870 studied here, and the c. 3580 Ma meta-anorthosite pods in the Shaw Granitic Complex). These gneissic outcrops are widespread throughout the Warrawagine Granitic Complex, and are probably significantly under-represented in geochronological databases. The Hf isotope data support that the >c. 3600 Ma detrital zircons (at least those to c. 3660 Ma) were derived from source rocks that are isotopically similar to the Warrawagine tonalite gneiss.

Veracity and significance of crystallization ages

Radiogenic isotope tracer data are of little use unless paired with robust crystallization ages. Determining the true crystallization age of a single zircon, whether detrital or inherited, may, however, be problematic without whole-rock information or constraints from field relations. It is difficult, in particular, to evaluate the effects of ancient lead loss, an issue that plagues geochronological studies of old zircons, in the same way that can be done with multi-grain zircon fractions from the same igneous rock (e.g. Compston and Kröner, 1988; Bowring and Williams, 1999). This difficulty arises because the ancient lead loss trajectory may be parallel to concordia, and so ion microprobe analyses from disturbed zircon domains appear concordant within the limits of analytical uncertainty. This effect can be cryptic in that there may be no accompanying microstructural evidence for isotope disturbance (e.g. Cavosie et al., 2004). If the geological context of the host rock is unknown, as is the case for single detrital and inherited zircons, it may become impossible to unambiguously determine the true crystallization age of the grain.

Combined age and Hf isotope tracer information can, under certain circumstances, reveal the effects of ancient lead loss within zircon populations (e.g. Amelin et al., 2000; Gerdes and Zeh, 2009; Whitehouse and Kemp, 2010). The Hf isotope ratio of zircon is typically unaffected by simple $^{207}Pb^{*}/^{206}Pb^{*}$ age resetting, and the very low Lu/Hf value of zircon means that the calculated initial $^{176}Hf/^{177}Hf$ value is insensitive to age (typical zircon evolution is about 0.000002 per 100 Ma). For these reasons, variable ancient lead loss characteristically results in a horizontal array on a Hf isotope evolution diagram, but a steep trend on an ε_{Hf}–time diagram (Amelin et al., 2000), which reflects the relatively rapid evolution of the chondritic reference frame (about 2.3 ε_{Hf} units per 100 Ma). Underestimation of crystallization ages caused by ancient lead loss leads to spuriously low ε_{Hf} values, and skews model crust-formation ages to values that are too old — an important effect that may not have been adequately taken into account in the calculation of global crustal growth curves.

There is some indication of ancient lead loss in the age–Hf isotope systematics of zircons of the Warrawagine tonalite gneiss. The oscillatory zoned rim on grain #19 (c. 3360 Ma) has $^{176}Hf/^{177}Hf_{3360\ Ma} = 0.280576 \pm 0.000015$, identical to that of the microstructurally similar c. 3410 Ma component (mean $^{176}Hf/^{177}Hf = 0.280575$ at c. 3410 Ma; Fig. 9). Similarly, the core of grain #15 returned a $^{207}Pb^*/^{206}Pb^*$ date of c. 3460 Ma, although in CL this domain resembles cores yielding >c. 3600 Ma ages. The Hf isotope composition of the grain #15 core plots near the lead loss trajectory from Group 1 zircons (Fig. 11), suggesting that it may have originally been c. 3650 Ma and was reset to a younger age. Most of the Group 2 zircon cores also plot close to this trajectory. Instead of being a discrete source component, Group 2 domains may also be c. 3650 Ma zircons that have lost variable amounts of radiogenic lead. In this case, the lower ε_{Hf} values are an artefact of assuming a crystallization age that is too young, rather than indicating a greater crustal residence prehistory for the crustal source.

By contrast, the clearly higher $^{176}Hf/^{177}Hf$ values of the c. 3410 Ma Group 3 zircons (Fig. 9a) means that these zircons cannot have formed simply by reconstitution of the older zircon cores they enclose. These growth domains have incorporated relatively radiogenic Hf, either from the bulk rock matrix (e.g. Patchett, 1983) or from an external source. The fine oscillatory zoning is consistent with crystallization from melt — although given the strongly deformed and recrystallized nature of the host tonalite gneiss it is now not possible to say whether this melt was generated by in situ anatexis of a c. 3650 Ma tonalite (of which the older zircon cores would now be remnants), injected from an external source, or the c. 3410 Ma zircon crystallized from an intrusive magma in which the older zircon cores were inherited components. Similar uncertainties have fuelled debates around the age and origin of other Eoarchean to Paleoarchean tonalite gneisses (e.g. Whitehouse et al., 1999 vs Nutman et al., 2001).

Assessing the effects of ancient lead loss in the pre-Pilbara Supergroup detrital zircons is difficult because individual grains are not necessarily related through a common source rock. Nonetheless, several younger zircons (3570–3550 Ma) yield low $^{176}Hf/^{177}Hf$ values that are similar to those of the 3670–3650 Ma zircons, and the corresponding ε_{Hf} values lie along predicted ancient lead loss trajectories from the older zircons (see Fig. 9). It is therefore plausible that the scattering to lower Hf values evident in the detrital zircon cluster is, at least in part, a manifestation of ancient lead loss.

Hadean crustal substrate to the Pilbara Craton?

Whether the exposed volcano-sedimentary successions of the Pilbara Craton developed on older protocontinental basement or formed in an oceanic setting has implications for the volume, composition, and mode of formation of the earliest continental nuclei. Evidence cited in support of a continental substrate includes neodymium model ages for the granitic rocks that are >c. 3500 Ma and, in some cases, >c. 3700 Ma (Bickle et al., 1993); the occurrence of >c. 3530 Ma detrital zircons in sedimentary units (Van Kranendonk et al., 2007); and the low initial $^{143}Nd/^{144}Nd$ ratios in c. 3490 Ma metabasalts and komatiites reported by Tessalina et al. (2010). The authors of the latter study attribute the apparently unradiogenic Nd ($\varepsilon_{Nd} = -3.3$) to extensive contamination by Hadean (c. 4300 Ma) continental crust. Green et al. (2000) favoured a similar model for the Coonterunah Subgroup basalts, based on trace element data.

The Hf isotope compositions of the detrital and igneous rock-hosted zircons presented in this study are considerably more radiogenic than evolutionary trends defined by generalized Hadean crustal reservoirs (Fig. 12). These data therefore provide no indication of the presence of Hadean crust in the Pilbara Craton. The distinctly subchondritic Hf isotope composition of the three oldest detrital zircons (>c. 3680 Ma) accords with reworking of significantly older rocks, although the location of that crust and its relationship to the Pilbara Craton is speculative. It may be that >c. 3680 Ma crust does underlie the Pilbara Craton although the c. 15 km-thick cover formed by the Pilbara Supergroup prevented its exposure until the first major orogenic event at 3070–3060 Ma (Van Kranendonk et al., 2010). The extreme scarcity of >3680 Ma zircons (only four have been identified in the Pilbara Craton) and the clearly different Hf isotope composition could, however, also be interpreted to indicate that these grains are exotic. Potential sources on the Australian mainland include quartzofeldspathic gneisses and migmatites of the Narryer Terrane (Yilgarn Craton) farther south, which contain zircon age populations of 3760–3730, c. 3680, and c. 3620 Ma (Kinny et al., 1988; Kinny and Nutman, 1996; Pidgeon and Wilde, 1998). Hf isotope data from these rocks, although sparse, overlap the less-radiogenic portion of the Pilbara zircon field (Fig. 12b). The two Pilbara detrital zircons at c. 3620 Ma with low ε_{Hf} (–2.7) could also have been derived from Yilgarn sources. However, the relative positions of the Pilbara Craton and Narryer Terrane and the remainder of the Yilgarn Craton in the Paleoarchean and Mesoarchean are unknown. Smirnov et al. (2013) used paleomagnetic evidence to conclude that the Pilbara and Yilgarn Cratons did not become closely juxtaposed until 2050–1800 Ma. An alternative source for some of the >c. 3600 Ma Pilbara detrital zircons with relatively low ε_{Hf} could be the Kaapvaal Craton of southern Africa, where zircons with similar age and

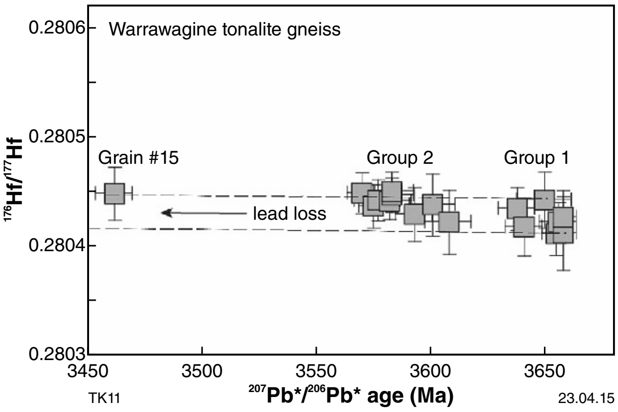


Figure 11. Hf isotope composition of the older zircon grains from the Warrawagine tonalite gneiss showing ancient lead loss vectors from Group 1 zircons

Hf isotope systematics have been identified (Zeh et al., 2011, 2014; Fig. 12b). Continental reconstructions for the Proterozoic have placed the Pilbara and Kaapvaal Cratons much closer together than today, and some paleomagnetic interpretations have concluded that they were contiguous (Zegers et al., 1998; de Kock et al., 2009, 2012) as adjacent parts of the Vaalbara supercontinent (Cheney, 1996).

What then is the age and nature of the substrate to the Pilbara Craton? As discussed above, the existence of >c. 3600 Ma zircon cores in felsic igneous rocks from c. 3460 Ma and younger (Thorpe et al., 1992; this study), presumably entrained from depth, argues in favour of a component that is at least this old. Candidates for relatively intact fragments of this older basement might reasonably include the tonalite gneiss enclaves in the Warrawagine Granitic Complex, and c. 3580 Ma meta-anorthosite pods in the Shaw Granitic Complex (McNaughton et al., 1988). The 3680–3550 Ma detrital zircons in a number of sedimentary packages across the Pilbara Craton is at least consistent with the notion of an older basement.

The oldest of the pre-Pilbara Supergroup zircons (excluding those with $^{207}Pb/^{206}Pb$ ages >c. 3680 Ma) show ε_{Hf} values around zero, whereas the dispersion towards lower ε_{Hf} in slightly younger grains may result from ancient lead loss. The bulk of these old zircons (including those of the Warrawagine tonalite gneiss) therefore originated from source material with an approximately chondritic Hf isotope composition. In considering the earliest growth of the Pilbara Craton, it is necessary to determine the crustal residence prehistory of this source. A common approach is to translate zircon Hf isotope ratios into model crust-formation ages, typically by assuming that the new crust separated from depleted mid-ocean ridge basalt (MORB)-source peridotite (Belousova et al., 2010). For the 3680–3650 Ma pre-Pilbara Supergroup zircons this calculation yields average model crust-formation ages of 4000–3900 Ma.

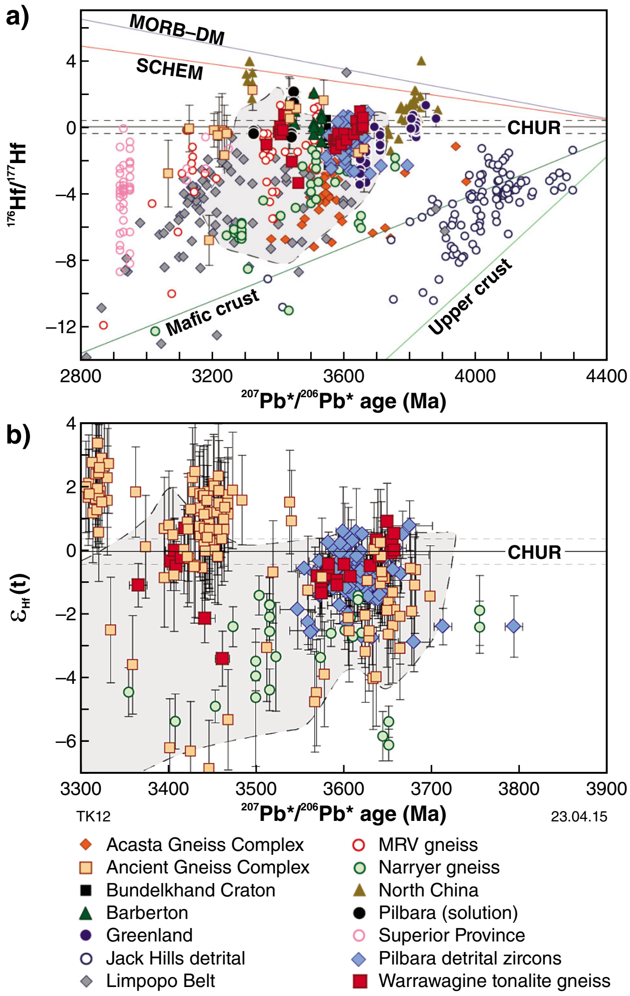


Figure 12. a) Compilation of the Hf isotope compositions of zircons from meta-igneous rocks of various Archean cratons, also showing one dataset from the Jack Hills detrital zircons (Kemp et al., 2010), and data from this study. Data sources are from Kemp and Hawkesworth (2013), and include Kaur et al. (2014) (Bundelkhand Craton), Satkoski et al. (2013) (Minnesota River Valley (MRV) gneisses), Wu et al. (2008) (North China Craton), and Zeh et al. (2011) (Ancient Gneiss Complex). The shaded field encompasses data from the Limpopo Belt zircons of Zeh et al. (2014). SCHEM, super-chondritic Earth model (Caro and Bourdon, 2010). The trajectories of putative Hadean 'mafic crust' and 'upper crust' were calculated assuming extraction from chondritic mantle at 4.5 Ga and evolution with $^{176}Lu/^{177}Hf$ values of 0.022 and 0.009, respectively; b) direct Hf isotope comparison between the pre-Pilbara Supergroup zircons of this study with zircons from the Narryer Gneisses of the Yilgarn Craton (data from Kemp et al., 2010), zircons of tonalite–trondhjemite–granodiorite (TTG) from the Ancient Gneiss Complex of Swaziland (Zeh et al., 2011), and detrital zircons from c. 3.1 Ga quartzite of the Central Zone of the Limpopo Belt, South Africa (grey shaded field; Zeh et al., 2014). Error bars are the 2σ uncertainties of individual analyses.

The conventional interpretation would dictate that the magmas from which these zircons crystallized formed from source rocks that were up to 400 Ma older. Model crust-formation ages are, however, underpinned by many assumptions that limit their utility in crustal evolutionary studies (Arndt and Goldstein, 1987; see Kemp and Hawkesworth, 2013). In particular, there is little justification for extraction of early Archean crust from strongly depleted mantle extrapolated from modern MORB. Indeed, zircons of the oldest (>3500 Ma) felsic rocks in several Archean cratons, such as the North Atlantic Craton (Greenland: Hiess et al., 2009; Kemp et al., 2009; Naeraa et al., 2012), the Superior Province (Satkoski et al., 2013), and the Kalahari Craton (Amelin et al., 2000; Zeh et al., 2009) also show approximately chondritic Hf (Fig. 12). There is currently little evidence for derivation of continental rocks older than 3600 Ma from long-term, relatively depleted (superchondritic Lu/Hf value) mantle sources (see also the compilation in Guitreau et al., 2012).

We therefore suggest that most 3680–3650 Ma Pilbara zircons crystallized from juvenile magmas that were extracted from mantle of broadly chondritic Hf isotope composition at this time. In this interpretation, these zircons record the incipient stages of felsic crust formation in the craton. It is not possible to make robust inferences about the volume and overall composition of this proto-cratonic crust. Part of this depends on whether the Warrawagine tonalite gneiss enclaves are genuinely representative of the composition of a c. 3650 Ma tonalite (as opposed to a c. 3410 Ma intrusive rock with a large degree of inheritance — 65% of grains >3570 Ma). Zircon typically crystallizes from silicic melts, and the intricate oscillatory zoning of many studied grains resembles that of zircons found in granites (e.g. Kemp et al., 2005). The broad oscillatory banding and/or sector zoning of other zircons are more akin to formation from residual silicic melts in relatively mafic bulk rock compositions, such as diorites and gabbros (e.g. Hoskin, 2000). It is therefore probable that felsic material was a component of the earliest crust in the Pilbara Craton, regardless of the overall composition of this crust. Further isotopic work is required to gauge the extent to which the oldest Pilbara crust was reworked by the voluminous younger magmas in the craton, which could provide additional clues on the volume and extent of this ancient nucleus. Notably, existing solution U–Pb and Hf isotope data on zircons from Pilbara felsic volcanic rocks (3450–3320 Ma) provide no evidence for substantially older crustal source components (Amelin et al., 2000).

Implications for early continental volumes

The Pilbara Craton is one of the largest and best preserved tracts of Paleoarchean to Mesoarchean crust on Earth. Hf isotope data from our study of some of the oldest zircons of this craton provide no compelling evidence for the existence of stable Hadean to early Eoarchean continental crust. This suggests either the reworking of older rocks, as would be revealed by distinctly subchondritic $^{176}Hf/^{177}Hf$ values, or the sampling of high Lu/Hf mantle depleted by prior extraction of large crustal volumes, as would be registered by strongly superchondritic zircon $^{176}Hf/^{177}Hf$ values. These conclusions are tempered by uncertainties as to whether CHUR is an appropriate Hf isotope proxy for the bulk Earth. In a superchondritic Earth model (SCHEM) (Caro and Bourdon, 2010), the Pilbara Hf isotope analyses would plot well below the bulk Earth baseline (Fig. 12). At present, however, the applicability of this alternative model is unresolved.

Conclusions

Hf isotope data are reported for a selection of detrital and inherited zircons from the Pilbara Craton that have $^{207}Pb^*/^{206}Pb^*$ ages between c. 3800 and 3550 Ma, substantially older than the most ancient exposed stratigraphy of this area, the Pilbara Supergroup (<c. 3530 Ma). Most of these zircons are interpreted to have been derived from ancient meta-igneous basement intrinsic to the craton, although the origins of three >3680 Ma grains with relatively unradiogenic Hf are unresolved. The pre-Pilbara Supergroup zircons show a narrow range of Hf isotope compositions that are consistent with derivation from igneous source rocks, possibly no longer preserved at the surface, of approximately chondritic composition. These source rocks are considered juvenile and extracted from chondritic mantle between c. 3700 and 3600 Ma, thereby marking an early stage in the growth of the craton. The Hf isotope data preclude the incorporation of Hadean components into magmas from which the >c. 3600 Ma zircons crystallized. The present data suggest either that the Pilbara Craton developed remote from the isotopic influence of the putative Hadean continental masses, or that the stabilized volumes of the earliest continents have been overestimated.

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Appendix 1

Tectonic settings, environments, and zircon characteristics of metasedimentary rocks analysed for Hf isotopes, with evidence for provenance

GSWA 169199: lithic quartz metasandstone, Mosquito Creek Formation

Depositional age of formation: Between c. 2980 and 2930 Ma

Oldest detrital zircon: 3648 ± 8 Ma

Tectonic setting: Central section of a deep-water rift basin (Mosquito Creek Basin) between the East Pilbara and Kurrana Terranes (Bagas et al., 2005)

Lithology and mineralogy: Lithic quartz metasandstone composed of poorly sorted grains of vein quartz, chert, and sericite-rich clasts in matrix of quartz–sericite schist. Original clay content may have been deposited from distal turbidity currents.

Unconformity proximity: Remote from the basal unconformity of the Mosquito Creek Basin

Zircon morphology: Equant and angular to elongate with rounded terminations. Many fractured and metamict grains. Grain surfaces are pitted, consistent with detrital transport.

Post-3530 Ma zircon age components: Age components are consistent with derivation from the Pilbara Craton: 3495 ± 18 Ma, within 3525–3490 Ma range of the Coonterunah Subgroup;[*†] 3428 ± 18 Ma, within 3450–3420 Ma range of the Tambina Supersuite; and 3290 ± 11 Ma, within 3325–3290 Ma range of the Emu Pool Supersuite.

Suggested provenance: Zircon rounding and the silica-rich composition of the rock indicate a relatively mature sedimentary protolith. Sedimentary differentiation probably resulted from longitudinal transport along the rift basin (Bagas et al., 2005). However, all grains could have been derived from known igneous units of the Pilbara Craton.

U–Pb geochronology reference: Nelson (2004a)

GSWA 169200: lithic quartz metasandstone, Mosquito Creek Formation

Depositional age of formation: Between c. 2980 and 2930 Ma

Oldest detrital zircon: 3681 ± 9 Ma

Tectonic setting: Central section of a deep-water rift basin (Mosquito Creek Basin) between the East Pilbara and Kurrana Terranes (Bagas et al., 2005)

Sample lithology and mineralogy: Quartz metasandstone containing up to 20% lithic grains, principally chert

Unconformity proximity: Remote from the basal unconformity of the Mosquito Creek Basin

Zircon morphology: Equant and angular to elongate with rounded terminations. Grain surfaces are pitted, consistent with detrital transport.

Post-3530 Ma zircon age components: A zircon age component dated at 3366 ± 9 Ma does not match any known magmatic event in the northern Pilbara Craton. Other components are consistent with derivation from the craton, although with one possible exception not from the adjacent East Pilbara and Kurrana Terranes: 3410 ± 15 Ma, close to minimum age of the 3450–3420 Ma Tambina Supersuite; 2997 ± 40 Ma, within the 3000–2980 Ma range of the Maitland River Supersuite and the 3010–2990 Ma Whim Creek Group;[†] and 2948 ± 24 Ma, within the 2955–2920 Ma range of the Sisters Supersuite.

Suggested provenance: Zircon age components indicate distal sources containing rocks of similar age to those in the west Pilbara Craton. The infrequency of zircon ages typical of the East Pilbara and Kurrana Terranes suggests long-distance current transport along the centre of the basin (Bagas et al., 2005). The wide dispersion of ages for concordant zircons, between c. 3681 and 2955 Ma, indicates multiple sources and extensive erosion of source regions prior to deposition of the sandstone.

Stratigraphic reference: Van Kranendonk et al. (2002)

U–Pb geochronology: Nelson (2004b)

*Previously published age ranges of stratigraphic units within the Pilbara Craton are from Van Kranendonk et al. (2006) unless referenced otherwise: †Hickman (2012), ‡Van Kranendonk et al. (2007), §Hickman and Van Kranendonk (2012), and #Hickman et al. (2010).

GSWA 178010: lithic quartz sandstone, Mosquito Creek Formation

Depositional age of formation: Between c. 2980 and 2930 Ma

Oldest detrital zircon: 3714 ± 5 Ma

Tectonic setting: Northern margin of the Mosquito Creek Basin adjacent to the East Pilbara Terrane. Deposition occurred in submarine fans during rifting of the basin margin (Bagas et al., 2005).

Lithology and mineralogy: Coarse-grained lithic quartz sandstone, very weakly metamorphosed, composed of poorly sorted grains of vein quartz, chert, and carbonaceous clasts

Unconformity proximity: Close to the basal unconformity of the Mosquito Creek Basin

Zircon morphology: Equant to slightly elongate and subhedral grains. Grain surfaces are pitted consistent with detrital transport.

Post-3530 Ma zircon age components: All zircons are Paleoarchean with age components consistent with derivation from the East Pilbara Terrane of the Pilbara Craton: 3461 ± 7 Ma, within the 3490–3460 Ma range of the Callina Supersuite; 3424 ± 12 Ma, within the 3450–3420 Ma range of the Tambina Supersuite; and 3294 ± 14 Ma, within the age range of the 3325–3290 Ma Emu Pool Supersuite.

Suggested provenance: All Paleoarchean zircons were derived from the East Pilbara Terrane. Zircons older than c. 3530 Ma were therefore derived either directly from older crust beneath the East Pilbara Terrane or by recycling of inherited zircons within the Paleoarchean sedimentary or igneous rocks of the terrane.

U–Pb geochronology: Nelson (2005a)

GSWA 142942: quartzite, Constantine Sandstone

Depositional age of formation: Between c. 3015 and 2948 Ma

Oldest detrital zircon: 3795 ± 5 Ma

Tectonic setting: Northern part of the Mallina Basin adjacent to the Whim Creek greenstone belt. The Mallina Basin is a rift basin similar to the Mosquito Creek Basin (Van Kranendonk et al., 2002).

Lithology and mineralogy: Quartzite derived from metamorphism of quartz sandstone. Sericite and minor albite make up about 20% of the rock, occurring interstitially between original quartz grains. The high quartz content indicates sedimentary differentiation resulting in mature quartz sand.

Unconformity proximity: Remote from the basal unconformity of the basin

Zircon morphology: Subrounded and broken grains, commonly zoned

Post-3530 Ma zircon age components: All zircon age components are consistent with derivation from the Pilbara Craton: 3425 ± 24 Ma, within the age range of the 3450–3420 Ma Tambina Supersuite; 3200 ± 17 Ma, similar to the c. 3200 Ma age of the Regal Formation[‡] and the 3200–3165 Ma Soanesville Group;[§] 3155 ± 16 Ma, between the ages of the 3200–3165 Ma Mount Billroth Supersuite and the 3130–3110 Ma Railway Supersuite; and 2994 ± 4 Ma, the main component, within the age ranges of the 3010–2990 Ma Whim Creek Group[†] and the 3000–2980 Ma Maitland River Supersuite.

Suggested provenance: Multiple sources of detritus are indicated although the Paleoarchean to Mesoarchean age components are consistent with derivation from the Pilbara Craton. The main zircon age component (2994 Ma) indicates significant input of detritus from the Whim Creek Group and Maitland River Supersuite directly northwest from the basin margin. The c. 3425 Ma zircons were most likely derived either from recycling of inherited zircons in the Karratha Terrane or Soanesville Group, or from a more distal source in the East Pilbara Terrane. The latter possibility would require longitudinal current transport along the rift basin.

U–Pb geochronology: Nelson (2000a)

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GSWA 142943: quartzite, Constantine Sandstone

Depositional age of formation: Between c. 3015 and 2948 Ma

Oldest detrital zircon: 3640 ± 6 Ma

Tectonic setting: Central part of the Mallina Basin. Deposition was from deep-water submarine fans, most probably from the southeast (Eriksson, 1982; Smithies et al., 1999, 2001).

Lithology and mineralogy: Quartzite derived from metamorphism of quartz sandstone. Quartz makes up at least 90% of the rock, and the original grain diameter was up to 3 mm.

Unconformity proximity: Within about 1 km stratigraphically above the basal unconformity of the Mallina Basin

Zircon morphology: Euhedral to subrounded grains, mainly zoned, and commonly metamict

Post-3530 Ma zircon age components: Almost all zircons are Paleoarchean with age components consistent with derivation from the Pilbara Craton: 3427 ± 5 Ma, within the 3450–3420 Ma range of the Tambina Supersuite; 3279 ± 21 Ma, within error of the published minimum age of c. 3290 Ma for the Emu Pool Supersuite; and 3201 ± 17 Ma, consistent with the ages of the 3200–3165 Ma Soanesville Group§ and the c. 3200 Ma Regal Formation.‡

Suggested provenance: If detritus was derived from the southeast, as previously suggested, the Soanesville Group of the Pilbara Well greenstone belt was the most probable source, with Paleoarchean grains recycled from older sources in the East Pilbara Terrane. The absence of Mesoarchean grains indicates little or no influx of detritus from the northwest side of the basin (e.g. Whim Creek greenstone belt) although available geochronology# does not preclude derivation of the 3279 ± 21 Ma component from the 3280–3260 Ma Roebourne Group and Karratha Granodiorite in the Karratha Terrane.

U–Pb geochronology: Nelson (2000b)

GSWA 142188: subarkosic metasandstone, Mallina Formation

Depositional age of formation: Between c. 3015 and 2931 Ma

Oldest detrital zircon: 3668 ± 8 Ma

Tectonic setting: Southeast part of the Mallina Basin (rift basin). Deposition occurred in submarine fans during crustal extension and rifting of the basin (Smithies et al., 2001).

Lithology and mineralogy: Poorly sorted, medium-grained subarkose unit within a turbidite succession. Grains of quartz (dominant), feldspar, and fragments of rock are angular to rounded.

Unconformity proximity: The sampling site was several kilometres stratigraphically above the basal unconformity of the Mallina Basin.

Zircon morphology: Large rounded and subrounded grains, commonly elongate

Post-3530 Ma zircon age components: Paleoarchean and Mesoarchean ages consistent with derivation from the East Pilbara Terrane and the Mesoarchean granites that intruded it. Some detritus may also have been derived from the 2955–2920 Ma Sisters Supersuite that intruded the Mallina Basin during clastic deposition. Age components: 3422 ± 17 Ma, within the age range of the 3450–3420 Ma Tambina Supersuite; 3288 ± 11 Ma, within error of the published 3325–3290 Ma range of Emu Pool Supersuite; 3066 ± 7 Ma, consistent with the 3068–3066 Ma range of Elizabeth Hill Supersuite; and 2941 ± 9 Ma, within the 2955–2920 Ma range of the Sisters Supersuite.

Suggested provenance: Post-3530 Ma zircon ages are consistent with igneous sources in the adjacent East Pilbara Terrane. The northwest part of this terrane was extensively intruded by Mesoarchean granites that most likely provided sources for the c. 3066 and 2941 Ma zircons. Syn-depositional erosion of c. 2941 Ma granites within the Mallina Basin may account for some of the youngest grains.

U–Pb geochronology: Nelson (1999)

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GSWA 142951: sandstone, Lalla Rookh Sandstone

Depositional age of formation: Between c. 2988 and 2931 Ma

Oldest detrital zircon: 3668 ± 8 Ma

Tectonic setting: Deposition in a small intracratonic basin (Krapez, 1984)

Lithology and mineralogy: Feldspathic sandstone containing approximately 60% quartz grains within a matrix of muscovite, sericite, and plagioclase. Very few quartz grains show significant rounding, indicating limited transport and sorting.

Unconformity proximity: The Lalla Rookh Sandstone is up to 5 km thick and unconformably overlies the Gorge Creek Basin and the East Pilbara Terrane. The stratigraphic level sampled is interpreted as more than 1 km above the basal unconformity although the sampling site is less than 3 km from both the northwest and southeast faulted basin margins.

Zircon morphology: Euhedral to anhedral grains, partly rounded and some metamict

Post-3530 Ma zircon age components: Three Paleoarchean zircon age components were distinguished by Nelson (2000): 3414 ± 7 Ma, within error of the c. 3420 Ma minimum age of the Tambina Supersuite; and 3329 ± 31 and 3304 ± 6 Ma, both consistent with derivation from the 3325–3290 Ma Emu Pool Supersuite.

Suggested provenance: Post-3530 Ma zircon age components are consistent with derivation from the East Pilbara Terrane. This terrane outcropped on both sides of the depositional basin.

U–Pb geochronology: Nelson (2000c)

GSWA 178045: sandstone, Corboy Formation

Depositional age of formation: Between c. 3228 and 3190 Ma

Oldest detrital zircon: 3656 ± 21 Ma

Tectonic setting: Passive margin succession on the western margin of the recently rifted East Pilbara Terrane (Van Kranendonk et al., 2010; Hickman, 2012)

Lithology and mineralogy: Quartz sandstone containing approximately 10% feldspar, principally in thin layers up to 2 mm thick

Unconformity proximity: Less than 20 m above an erosional unconformity on undated banded granitic gneiss of the Yule Granitic Complex

Zircon morphology: Most grains are elongate and euhedral although some have pitted surfaces consistent with detrital transport. Most grains are zoned and some have fractured unzoned cores.

Post-3530 Ma zircon age components: Most grains analysed are older than c. 3530 Ma. Of the remainder, the dominant age component is 3273 ± 14 Ma, within range of the 3275–3225 Ma Cleland Supersuite. Three analyses of two zircons indicated a second age component at 3192 ± 74 Ma, within range of the 3200–3165 Ma Mount Billroth Supersuite and also consistent with the 3200–3165 Ma depositional age of the Soanesville Group.§

Suggested provenance: The feldspathic content of the sandstone, low degree of abrasion of the zircon grains from detrital transport, and close proximity to the underlying erosional unconformity are consistent with relatively proximal sources.

U–Pb geochronology: Nelson (2005b)

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GSWA 169013: metasandstone, ?Corboy Formation

Depositional age of formation: Correlation of this metasandstone with the 3228–3190 Ma Corboy Formation is very tentative. An alternative interpretation is that it was deposited as part of the same succession preserved in the Pilbara Well greenstone belt about 30 km to the north. The Pilbara Well succession includes sandstone that might belong to the Coonterunah Subgroup (see GSWA 180095 on this page).

Oldest detrital zircon: 3666 ± 8 Ma

Tectonic setting: The Corboy Formation belongs to a passive margin succession (Soanesville Group) deposited on the western margin of the East Pilbara Terrane (Van Kranendonk et al., 2010; Hickman, 2012).

Lithology and mineralogy: Poorly sorted, medium-grained metasandstone containing approximately 50% quartz, 35% clinopyroxene, and 15% microcline. The sandstone was metamorphosed at amphibolite facies.

Unconformity proximity: The metasandstone outcrops close to granitic rocks and amphibolites within the Yule Granitic Complex but as a consequence of late Mesoarchean granitic intrusion no unconformable contacts are preserved. It is interpreted that the sandstone was deposited directly on Paleoarchean granites and greenstones of the Yule Granitic Complex, and in this interpretation the basal unconformity was probably within 1 km.

Zircon morphology: Euhedral to subrounded grains, commonly metamict

Post-3530 Ma zircon age components: Apart from one grain dated at 3331 ± 5 Ma (close to the age range of the 3325–3190 Ma Emu Pool Supersuite) other post-3530 Ma grains belong to a group dated at 3524 ± 5 Ma (the same age range as zircons within the Coonterunah Subgroup). The c. 3331 Ma grain may define the maximum depositional age of the sandstone; however, zircons of this age might have been introduced when the sandstone was intruded by granite and metamorphosed.

Suggested provenance: The source of grains in this sample may have been similar to that for grains in GSWA 180095.

U–Pb geochronology: Nelson (2005c)

GSWA 180095: metasandstone, unit unassigned

Depositional age of formation: Unknown. Based on geochronology on overlying formations the sandstone has been provisionally mapped as part of the Soanesville Group. However, an unconformity may exist between this sandstone and the Soanesville Group because the youngest zircon of all 54 grains analysed from GSWA 180095 was dated at 3514 ± 2 Ma. The geochronology is therefore consistent with an alternative interpretation that the sandstone is part of the Coonterunah Subgroup of the Warrawoona Group (Pilbara Supergroup). Geochronology in another greenstone belt of the East Pilbara Terrane indicated that a felsic volcanic formation in the central part of the Coonterunah Subgroup has an isotopic date range of c. 3515 to 3490 Ma. Another possibility is that the metasandstone might be laterally equivalent to the c. 3480 Ma Dresser Formation. At North Pole the Dresser Formation is dominated by detrital zircons dated between c. 3534 and 3507 Ma (GSWA 180070, Wingate et al., 2009).

Oldest detrital zircon: 3647 ± 4 Ma

Tectonic setting: Basal section of the Pilbara Well greenstone belt. If the sandstone is part of the Coonterunah Subgroup it is positioned near the stratigraphic base of the Pilbara Supergroup. The Pilbara Supergroup is interpreted to have been deposited directly on older continental crust of granite–greenstone composition (Hickman, 1981; Green et al., 2000; Van Kranendonk et al., 2007, 2015; Hickman and Van Kranendonk, 2012). If the Pilbara Well metasandstone is correlated with the c. 3480 Ma Dresser Formation it was most likely deposited above a thick basaltic succession older than 3490 Ma. If the metasandstone is part of the Soanesville Group it was deposited on a subsiding passive margin, although in this scenario the reason for the absence of any detrital zircon components dated between c. 3514 and 3200 Ma is unknown.

Lithology and mineralogy: Quartz metasandstone containing rounded grains of quartz and less-abundant altered rock fragments in a siliceous matrix including microcrystalline muscovite or sericite. The protolith was sandstone containing small grains of igneous rock, either volcanic or intrusive.

Unconformity proximity: If the sandstone is part of the Coonterunah Subgroup it closely overlies the basal unconformity of the Pilbara Supergroup. However, no basal stratigraphic contact has been preserved because of folding, faulting, and strong shearing against c. 3420 Ma granitic rocks of the Tambina Supersuite. If the sandstone is part of the Soanesville Group it contains detritus shed from the rifted margin of the East Pilbara Terrane.

Zircon morphology: Equant anhedral to euhedral grains, variably rounded consistent with detrital transport. Euhedral growth zoning is common and many of the grains are partly metamict.

Post-3530 Ma zircon age components: A single early Paleoarchean age component dated at 3518 ± 4 Ma, within the previously interpreted 3525–3490 Ma age range of the Coonterunah Subgroup†.

Suggested provenance: Post-3530 Ma zircon ages are tightly constrained within the age range of the Coonterunah Subgroup, suggesting no younger sources were involved. If the sandstone is part of the Coonterunah Subgroup it was derived partly from erosion of felsic igneous rocks of the same age and partly from older crust. A lack of sedimentological data on the unit prevents any conclusions on the proximity of the source of the detritus.

U–Pb geochronology: Wingate et al. (2009)

GSWA 168992: volcaniclastic metasandstone, Wyman Formation

Depositional age of formation: Between c. 3325 and 3315 Ma

Oldest detrital zircon: 3676 ± 13 Ma

Tectonic setting: Deposition during the final stages of the Kelly volcanic cycle on the Paleoarchean volcanic plateau of the East Pilbara Terrane (Hickman, 2011, 2012)

Lithology and mineralogy: Volcaniclastic metasandstone composed of altered lithic and vitric fragments and rare quartz grains within a siliceous matrix

Unconformity proximity: The basal unconformity of the Kelly Group is several hundred metres beneath the Wyman Formation.

Zircon morphology: Most grains are euhedral although a minority show rounding consistent with detrital transport.

Post-3530 Ma zircon age components: Detrital zircon grains have various Paleoarchean ages consistent with derivation from different underlying units of the East Pilbara Terrane: two zircons have ages consistent with derivation from the 3490–3460 Ma Callina Supersuite and two zircons are similar in age to the 3450–3420 Ma Tambina Supersuite. The main age component is dated at 3324 ± 4 Ma, which is within the 3325–3290 Ma range of the Emu Pool Supersuite.

Suggested provenance: Post-3530 Ma zircon ages are consistent with sources in the adjacent East Pilbara Terrane.

U–Pb geochronology: Nelson (2004d)

142836: volcaniclastic metasandstone, Strelley Pool Formation

Depositional age of formation: Between c. 3430 and 3350 Ma

Oldest detrital zircon: Two zircons were dated at 3622 ± 6 and 3622 ± 12 Ma.

Tectonic setting: Deposition occurred on a volcanic plateau during a c. 75 million year break in volcanism between the Salgash and Kelly volcanic cycles (Hickman, 2008, 2011, 2012; Van Kranendonk et al., 2015).

Lithology and mineralogy: Quartz sandstone containing angular to rounded quartz grains and lithic fragments

Unconformity proximity: Directly overlying a regional erosional unconformity developed on the Panorama Formation of the Warrawoona Group (Hickman, 2008)

Zircon morphology: Most zircons are rounded, suggesting detrital transport.

Post-3530 Ma zircon age components: A single age component was dated at 3426 ± 10 Ma, within the 3450–3420 Ma range of the Tambina Supersuite.

Suggested provenance: Post-3530 Ma zircon ages are consistent with derivation from either the Tambina Supersuite or the contemporaneous Panorama Formation. This suggests that older formations of the Warrawoona Group did not contribute detritus to the Strelley Pool Formation in the area sampled, probably as a consequence of local concealment by the Panorama Formation. By contrast, zircon U–Pb geochronology on the Strelley Pool Formation in other greenstone belts indicates erosion of almost all formations of the Warrawoona Group or granitic rocks of similar age (Hickman, 2008).

U–Pb geochronology: Nelson (1998)

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Nelson, DR 2004c, 169013: metasandstone, Punduna Pool; Geochronology Record 138; Geological Survey of Western Australia, 5p.

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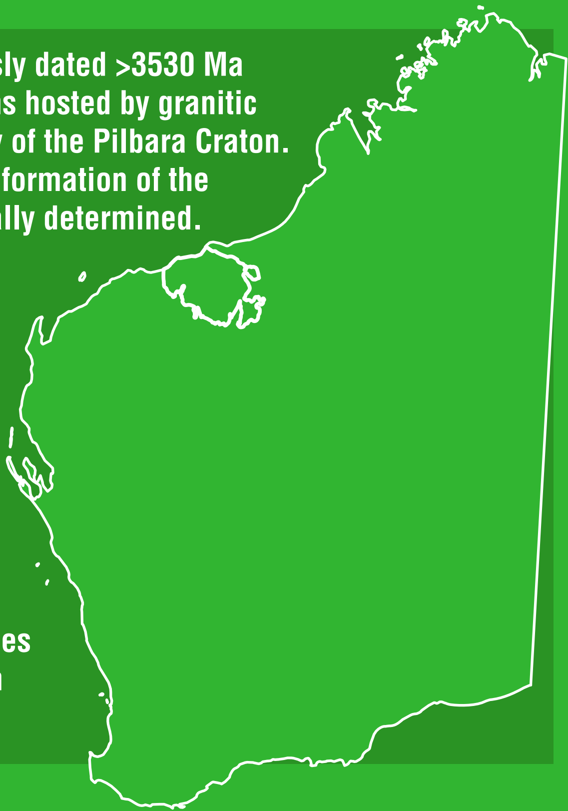
New Lu–Hf isotope data are presented for previously dated >3530 Ma detrital zircons in sandstones, and inherited zircons hosted by granitic gneisses, to investigate the earliest growth history of the Pilbara Craton. The source of the detrital zircons that pre-date the formation of the exposed Pilbara Supergroup cannot be unequivocally determined. However, the U–Pb and Hf isotope systematics are consistent with the 3680–3600 Ma detrital grains being eroded from meta-igneous rocks similar to the oldest gneisses of the East Pilbara Terrane. Hf isotope compositions of the <3680 Ma zircons suggests that the early crustal components of the Pilbara Craton separated from the mantle between 3700 Ma and 3600 Ma. Results from our study suggest that if significantly older crust was ever present in the Pilbara Craton data it was very minor. One implication is that the stabilized volumes of the earliest continents (pre-3700 Ma) have been overestimated.



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EARLY EVOLUTION OF THE PILBARA CRATON FROM HAFNIUM ISOTOPES IN DETRITAL AND INHERITED ZIRCONS

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