1 The Mangaroon Orogeny: Synchronous c. 1.7 Ga magmatism and low-*P*, high-*T* 2 metamorphism in the West Australian Craton

3 Agnieszka M. Piechocka^{1*}, Jian-Wei Zi^{2,3}, Courtney J. Gregory¹, Stephen
4 Sheppard^{1,6}, Fawna J Korhonen⁴, Ian C.W. Fitzsimons¹, T. E. Johnson¹, Birger
5 Rasmussen^{3,5}

6 ¹*School of Earth and Planetary Sciences, Curtin University, Kent Street, Bentley,*
7 *WA 6102, Australia*

8 ²*John de Laeter Centre, Curtin University, Kent Street, Bentley, WA 6102,*
9 *Australia*

10 ³*State Key Laboratory of Geological Processes and Mineral Resources, China*
11 *University of Geosciences, Wuhan, 430074, China*

12 ⁴*Geological Survey of Western Australia, 100 Plain Street, East Perth, WA 6004,*
13 *Australia*

14 ⁵*School of Earth Sciences, The University of Western Australia, Nedlands, WA*
15 *6009, Australia*

16 ⁶*Calidus Resources Limited, Suite 12, 11 Ventnor Avenue, West Perth, WA 6005,*
17 *Australia*

18

19 Abstract

20 The Capricorn Orogen records nearly one billion years of intraplate
21 orogenesis within the West Australian Craton, although the processes
22 responsible for this protracted, punctuated reworking remain unclear. Of the
23 major tectonic events that affected the region, the 1680–1620 Ma Mangaroon
24 Orogeny is one of the least-well understood, mainly due to a lack of direct ages
25 for metamorphism, an absence of pressure (*P*)–temperature (*T*) constraints, and
26 uncertainty regarding the duration of granitic magmatism that is spatially and,
27 possibly, temporally associated with deformation. In this study we define the *P*–
28 *T*–time (*t*) conditions associated with the Mangaroon Orogeny based on *in situ*
29 SHRIMP U–Pb monazite and xenotime geochronology and calculated *P*–*T*
30 pseudosections. Data from a pelitic migmatite constrain the timing of low-*P*,
31 high-*T* metamorphism to 1691 ± 7 Ma at conditions of 665–755 °C and 2.7–4.3

kbar ($\sim 175\text{--}240\text{ }^{\circ}\text{C/kbar}$). Data from a garnet-bearing sillimanite–biotite pelitic gneiss suggests higher pressure during the clockwise prograde history at one locality. Furthermore, the onset of the metamorphism coincides with the oldest granites in the region ($1695 \pm 9\text{ Ma}$), which constrain the onset of the Mangaroon Orogeny. Our results also show that deposition and burial of the precursor sediments occurred, at most, c. 70 million years before the onset of partial melting at c. 1695 Ma. Therefore, there was no long incubation period before the onset of orogenesis. We conclude that, unlike many of the Proterozoic orogenic events in Australia, the c. 1.7 Ga low- P , high- T metamorphism recorded in the upper crust in the Capricorn Orogen cannot be explained by a thermal lid model, but rather was synchronous with granitic magmatism.

Keywords: *intraplate orogeny, U–Pb geochronology, monazite, xenotime, LPHT metamorphism, Proterozoic*

1. Introduction

Proterozoic Australia has been subjected to numerous intraplate orogenic events, typically characterised by low-pressure (low- P) and high-temperature (high- T) metamorphism. Several tectonic models have been proposed to explain the causes and characteristics of low- P , high- T metamorphism, including crustal thickening with coeval magmatism (Loosveld and Etheridge, 1990), crustal extension (Wickham and Oxburgh, 1987; Sandiford and Powell, 1986), and regional contact metamorphism (e.g. Barton and Hanson, 1989). A model commonly invoked for many Australian Proterozoic terrains is that of low- P , high- T metamorphism a product of lithospheric weakening owing to high concentrations of heat-producing elements (McLaren et al., 2005), perhaps

enhanced by thick sedimentary sequences that provided insulation (Sandiford and Hand, 1998).

In the Capricorn Orogen of Western Australia, intraplate metamorphism and deformation associated with the Mesoproterozoic Mutherbukin Tectonic Event was interpreted to reflect elevated heat production beneath a thick thermal blanket, resulting in elevated temperatures that lasted in excess of 100 million years (Korhonen and Johnson, 2015; Korhonen et al., 2017). Although this model has been proposed for other Proterozoic orogens in Australia (e.g. McLaren et al., 2005; Korhonen et al., 2017; Morrissey et al., 2014), it is unclear as to whether all Proterozoic orogenies reflect such a process.

An alternative model for low- P , high- T metamorphism in Proterozoic Australian orogens, is one which invokes intrusion of voluminous granites or mafic rocks to provide the heat for metamorphism (e.g. Clark et al., 2014; Collins and Vernon, 1991; Oliver and Zakowski, 1995; Glasson et al., 2019). Collins and Vernon (1991) proposed that low- P , high- T metamorphism in the upper crust of the Arunta Orogen was caused by granite magmatism, triggered by a mantle perturbation that induced partial melting of the lower crust. Clark et al. (2014) and Glasson et al. (2019) argued that the drivers for low- P , high- T metamorphism associated with Stage I of the Albany–Fraser Orogeny included elevated heat flow and mafic magmatism during lithospheric extension. Furthermore, it was suggested that low- P , high- T metamorphism in the eastern Mt Lofty Ranges, south Australia, was synchronous with granite magmatism in an extensional setting (Oliver and Zakowski, 1995).

The Capricorn Orogen of Western Australia records a prolonged history of repeated intraplate orogenesis, comprising multiple episodes of predominantly

felsic magmatism, metamorphism and deformation. This study focuses on one of these events—the 1680–1620 Ma Mangaroon Orogeny—which has been proposed as an example of low- P , high- T regional metamorphism, based on diagnostic mineral assemblages and a widespread gneissic textures in rocks of the southern Mangaroon Zone, although local occurrences of contact metamorphism have been recorded (Sheppard et al., 2005). However, no detailed work has been done on the timing of metamorphism and the determination of precise P – T conditions. Therefore, the cause and duration of metamorphism associated with the Mangaroon Orogeny remain unknown.

In this contribution, we use SHRIMP U–Pb monazite and xenotime data, and P – T pseudosections to provide the first robust constraints on the duration and P – T conditions of intraplate low- P , high- T metamorphism associated with the Mangaroon Orogeny. Furthermore, we obtained whole-rock geochemistry to determine the peraluminosity of the granites and calculate zircon saturation temperatures. This study indicates that the thermal lid model, which is favoured for other Proterozoic intraplate tectonic events in Australia, cannot easily explain the anomalous thermal gradients in the crust during the Mangaroon Orogeny. The data are then integrated with studies of low-grade metamorphic and hydrothermally altered rocks across the West Australian Craton to show that the Mangaroon Orogeny has a significantly wider footprint than previously thought. Further, we contend that the metamorphism and magmatism recorded in the Capricorn Orogen may have been responsible for widespread circulation of fluids in the upper crust over much of the craton.

2. Regional Geology

106 The study area is in the Capricorn Orogen of Western Australia (Fig. 1).
107 The Capricorn Orogen records the two-stage assembly of the West Australian
108 Craton, separating the Archean Pilbara and Yilgarn cratons to the north and
109 south, respectively. Assembly of the West Australian Craton, which was complete
110 by c. 1950 Ma (Johnson et al., 2011; Occhipinti et al., 2004), was followed by five
111 episodes of intraplate reworking and reactivation spanning nearly 1.5 billion
112 years that shaped the current architecture of the orogen (Korhonen et al., 2017;
113 Piechocka et al., 2018; Piechocka et al., 2019; Sheppard et al., 2005, 2007,
114 2010a). All these tectonic events are recorded in rocks of the Gascoyne Province,
115 a basement complex of magmatic and metamorphic rocks exposed at the
116 western end of the orogen (Fig. 1). The Gascoyne Province is subdivided into
117 several southeast-trending structural and metamorphic zones, each recording a
118 discrete tectonothermal history. Of the five intraplate events, four record
119 reworking (i.e., pervasive deformation accompanied by metamorphism and/or
120 magmatism; Korhonen et al 2017; Piechocka et al., 2019; Sheppard et al., 2005;
121 Sheppard et al., 2007; Sheppard et al., 2010a), whereas the youngest events,
122 including the later stages of the fourth episode of reworking, reflect reactivation
123 along discrete structures, in which the intervening rocks were largely unaffected
124 (Bodorkos and Wingate, 2007; Piechocka et al., 2018).

125 The two Paleoproterozoic intraplate events—the 1820–1775 Ma
126 Capricorn Orogeny (Sheppard et al., 2010a; Piechocka et al., 2019) and the 1680–
127 1620 Ma Mangaroon Orogeny (Sheppard et al., 2005)—produced pervasive
128 metamorphism and deformation accompanied by voluminous, granitic
129 magmatism. The Capricorn Orogeny is characterised by extensive compressional
130 deformation (mostly at low to medium metamorphic grade), magmatism and

some sedimentation (Sheppard et al., 2010a). A recent U–Pb phosphate geochronology study showed that metamorphism during the Capricorn Orogeny was synchronous with magmatism (Piechocka et al., 2019). Deformation and medium-grade metamorphism (<750 °C and <6 kbar) associated with the Mangaroon Orogeny was preceded by the deposition of siliciclastic sedimentary rocks (the Pooranoo Metamorphics; Sheppard et al., 2005).

During the Meso- to Neoproterozoic the Gascoyne Province was subjected to two additional reworking events and two reactivation events. The driver for the older of the reworking events, the 1320–1170 Ma Mutherbukin Tectonic Event in the central Gascoyne Province, was proposed to be thickening of metasedimentary rocks deposited at 1842–1807 Ma which are enriched in high heat-producing elements. According to Korhonen et al. (2017), these Paleoproterozoic rocks formed a thermal lid that generated radiogenic heat for c. 110 million years prior to the onset of peak metamorphism at >650 °C and 4.4–7.0 kbar.

The second reworking event, the 1030–900 Ma Edmundian Orogeny, is characterised by coeval leucogranite magmatism and metamorphism and deformation at 500–550° C and 3–4 kbar (Piechocka et al., 2017; Sheppard et al., 2007). Although metamorphism had ceased by c. 990 Ma, the leucogranite magmatism persisted for another c. 100 million years (Piechocka et al., 2017). Orogen-scale reactivation of pre-existing shear zones at 920–830 Ma involved dextral strike-slip movement in the northern parts of the province and exhumation of the southern portion (Occhipinti & Reddy 2009, Piechocka et al., 2018). This widespread reactivation may be attributed to the Edmundian Orogeny. The effects of the youngest reactivation event, the c. 570 Ma Mulka

Tectonic Event (Bodorkos and Wingate, 2007), are limited to faulting and dextral shearing on discrete structures.

3. Mangaroon Zone

The study area is located in the Mangaroon Zone of the northern Gascoyne Province (Figs 1 and 2) and comprises metasedimentary rocks of the Pooranoo Metamorphics and voluminous granitic intrusions of the Durlacher Supersuite. The Mangaroon Zone is about 40–70 km wide and is overlain by Phanerozoic sedimentary rocks in the west and by Meso- to Neoproterozoic metasedimentary rocks to the east. It is bounded to the north by the Collins Fault and to the south by the Minga Bar Fault. Figure 2 shows the geology of the Mangaroon Zone along with sample locations.

3.1 Pooranoo Metamorphics

The Pooranoo Metamorphics comprise mainly interlayered psammitic schist and feldspathic metasandstone with pelitic rocks and metaconglomerate (Sheppard et al., 2010b). In the Mangaroon Zone, the rocks are commonly schistose, but gneissic and migmatitic metapelitic rocks are known from one locality in the southern Mangaroon Zone. Locally, graded bedding is preserved in the feldspathic metasandstone (Sheppard et al., 2010b). The protoliths to the Pooranoo Metamorphics were deposited as a turbidite sequence in a narrow ~50 km-wide basin, with an unknown thickness. The turbiditic nature of the protoliths suggests a deep-water depositional setting (Sheppard et al., 2005; Sheppard et al., 2010b; Sheppard and Johnson, 2016).

SHRIMP U–Pb zircon geochronology was conducted by the Geological

Survey of Western Australia (GSWA) to constrain the maximum and minimum depositional ages of precursors to the Pooranoo Metamorphics. A sample of the Biddenew Formation of the Mt James Subgroup yielded a youngest detrital zircon population at 1758 ± 18 Ma, which is interpreted as the maximum depositional age of the formation (Sheppard and Johnson, 2016; GSWA 183255: Wingate et al., 2009). A minimum depositional age is provided by the oldest granite to intrude the Pooranoo Metamorphics, which has an igneous crystallization age of 1682 ± 4 Ma (GSWA 208318; Wingate et al., 2013). Therefore, sedimentary precursors to the Pooranoo Metamorphics were deposited between c. 1758 and 1682 Ma.

3.2 Durlacher Supersuite in the Mangaroon Zone

In the Mangaroon Zone, the Durlacher Supersuite granites were emplaced as voluminous plutons and dykes that intruded the Pooranoo Metamorphics at c. 1680–1660 Ma (Sheppard et al., 2005) and comprise biotite–muscovite monzogranite, granodiorite, syenogranite and minor muscovite–tourmaline monzogranite (Sheppard et al., 2010b). A deep-crustal seismic reflection survey suggests that granites extend to at least 10 km depth (fig. 10 of Johnson et al., 2013). Durlacher granites are also known from outside the Mangaroon Zone in other parts of the Gascoyne Province: in the Mutherbukin Zone and in the Yarlarweelor Gneiss Complex (Fig. 1). The current upper and lower age limits of Durlacher Supersuite magmatism are constrained by the oldest monzogranite with an igneous crystallization age of 1682 ± 4 Ma, from the Mangaroon Zone, (Sheppard and Johnson 2016; GSWA 208318; Wingate et al., 2013) and a porphyritic biotite monzogranite from the Yarlarweelor Gneiss Complex that

yielded a crystallisation age of 1619 ± 15 Ma (GSWA 142855: Nelson 1998; Sheppard et al., 2005). However, no granites younger than c. 1660 Ma are known from the Mangaroon Zone. Furthermore, no metamorphic assemblages or deformation fabrics associated with the Mangaroon Orogeny are known outside of the Mangaroon Zone. Recent $^{40}\text{Ar}/^{39}\text{Ar}$ mica geochronology, from a monzogranite in the northern Mangaroon Zone (GSWA 195890), yielded a c. 1640 Ma age, interpreted to record cooling and uplift (Piechocka et al., 2018).

3.3 Deformation and metamorphism (D_1/M_1 and D_2/M_2)

In the Mangaroon Zone, of the northern Gascoyne Province, two distinct tectonometamorphic events, D_1/M_1 and D_2/M_2 , have been recognised (Sheppard et al., 2005). D_1/M_1 is characterised by amphibolite facies regionally extensive of the Pooranoo Metamorphics. In the southern Mangaroon Zone, D_1 deformation produced local gneissic fabrics or granofelsic textures, with evidence for partial melting of pelitic and semi-pelitic lithologies at the Star of Mangaroon locality (Sheppard et al., 2005). Also, at the Star of Mangaroon mine local recumbent folds with axial surfaces defined by sillimanite are associated with D_1 (Sheppard et al., 2005). However, some areas preserve a granofelsic texture, which suggests minimal non-coaxial strain. M_1 metamorphic assemblages preserved in pelitic and granofels of biotite–muscovite–quartz–plagioclase–sillimanite, quartz–biotite–cordierite–plagioclase–muscovite(–sillimanite) and plagioclase–biotite–quartz–sillimanite–muscovite–cordierite combined with the general absence of garnet were previously interpreted to record low- P , high- T regional metamorphism (Sheppard et al., 2005). The grade of metamorphism appears to be lower in the northern half of the Mangaroon Zone (Sheppard et al., 2005).

Formerly, the upper age constraint on D_1/M_1 , and the lower age constraint on D_2/M_2 , were based on a diatexitic granite with a zircon crystallisation age of c. 1677 Ma (GSWA 178027: Nelson 2005) that cross-cuts an undated gneissic fabric that was presumed to be associated with D_1 (Sheppard et al., 2005). However, recent zircon ages of c. 1808 Ma obtained from the gneissic host that is cross-cut by the diatexite show that it is part of the older 1820–1775 Ma Moorarie Supersuite (Wingate et al., 2013), and unrelated to the Mangaroon Orogeny. Consequently, the previous upper age constraint on D_1/M_1 and lower age constraint on D_2/M_2 , (i.e., 1677 Ma) may not be valid.

The D_2 deformation event produced a pervasive schistosity, metre- to kilometre-scale upright folds, and retrogression of D_1 metamorphic minerals to greenschist facies assemblages (Sheppard et al., 2005; 2010b). M_2 is characterised by sericite (after sillimanite, cordierite, plagioclase and microcline), and chlorite (after cordierite) with a typical M_2 assemblage consisting of sericite–chlorite–quartz–plagioclase–biotite observed in schists (Sheppard et al., 2005). Many granites of the Durlacher Supersuite with crystallization ages of c. 1675 Ma or older were deformed during D_2 , providing a maximum age constraint for D_2 (Sheppard et al., 2005; 2010b).

4. Sample descriptions and petrography

In total, nine samples of pelitic and granitic rocks were collected from across the Mangaroon Zone for U–Pb geochronology. Monazite was used to determine the crystallisation ages for the granitic rocks (three samples) and monazite and xenotime were selected to date the timing of metamorphism in the pelitic rocks (six samples). Of these pelitic samples, two (GSWA 219742 and

219749) contained a preserved mineral assemblage that was suitable for phase equilibria modelling (GSWA 219742 and GSWA 219749). Locations of all samples used in this study are shown in Figure 2. Descriptions of all the granitic and pelitic samples analysed as part of this study are provided below and summarised in Table 1. Figure 3 shows images of granitic sample sites in outcrop and thin section; Figure 4 consists of outcrop photos of pelitic sample locations; and Figure 5 consists of photomicrographs of pelitic samples.

Granitic rocks (Durlacher Supersuite)

GSWA 195890: muscovite–biotite metamonzogranite

This rock is a medium-grained, porphyritic muscovite–biotite monzogranite (Fig. 3a). The monzogranite shows variation in strain across the outcrop with zones of mylonitisation, quartz veining and brecciation and zones with preserved igneous textures. The vertical metamorphic foliation trends 230°. The typical mineral assemblage consists of K-feldspar, quartz, muscovite and plagioclase with ~25% ferromagnesian minerals including biotite and secondary chlorite, along with accessory Fe-oxides, fluorite, zircon and monazite. In places, zones of pervasive mylonitization resulted in almost complete recrystallisation of the rock during the Neoproterozoic (Piechocka et al., 2018) (Fig. 3c).

P01085: biotite–muscovite syenogranite

This sample is a porphyritic, biotite–muscovite syenogranite consisting of orthoclase, quartz, plagioclase, and muscovite. Mafic minerals make up around 15% of the rock and include biotite and Fe–Ti oxides. Orthoclase predominantly occurs as anhedral crystals and is perthitic. The largest orthoclase crystals occur

as tabular phenocrysts up to 8 mm long by 4 mm wide. Plagioclase shows multiple twinning and quartz shows undulose extinction. Sericite is pervasive throughout the sample.

GSWA 219741: tourmaline–muscovite monzogranite

This sample is a relatively undeformed medium-grained leucocratic monzogranite (Fig. 3b) consisting of orthoclase, quartz, plagioclase and muscovite. Ferromagnesian minerals make up around 10% of the rock and include tourmaline, biotite, chlorite, and Fe-oxides. Pervasive sericite and recrystallised quartz show a minor post magmatic overprint, although the rock looks relatively undeformed in outcrop.

Pelitic rocks (Pooranoo Metamorphics)

GSWA 219749: sillimanite–cordierite pelitic migmatite

The outcrop from which this sample was collected is characterised by stromatic migmatites (Fig. 4a) with leucosomes of feldspar, quartz and biotite and melanosomes of coarse sillimanite and biotite. Seemingly unmigmatised layers of psammite, up to a few meters thick, are cross-cut by a sparse network of leucocratic veins. Folded pelitic and psammitic rocks have axial planar fabrics defined by biotite (D_1) (Martin et al., 2006). At this locality the fold axial surface trends $147^\circ/85^\circ$ (SE) with a fold axis trending $230^\circ/70^\circ$ (SE). In thin section, the foliation (D_1) in the pelitic migmatite is defined by elongate flattened quartz grains that wrap recrystallised quartz, anhedral cordierite porphyroblasts that are altered around their margins (up to 4.5 mm), and magnetite up to 0.5 mm in diameter (Fig. 5a). Fibrolite defines a lineation. Quartz does not show undulose

extinction. Anhedral biotite has brown to orange pleochroism, contains pleochroic halos and is associated with by Fe–Ti oxides at its margins. K-feldspar is anhedral and up to 1 mm across. Secondary muscovite cross-cuts biotite crystals, and sericite alteration suggests post-peak alteration. The interpreted peak metamorphic assemblage (M_1) is sillimanite–cordierite–biotite–magnetite–K-feldspar–quartz–plagioclase–ilmenite–melt.

GSWA 219742: sillimanite–biotite pelitic gneiss

This fine- to coarse-grained rock contains conspicuous porphyroblasts of prismatic sillimanite minor patches of leucosome (Fig. 4b). In outcrop, the rock is not obviously deformed and shows no signs of retrogression. The gneiss is layered (probably reflecting relict bedding) with the sillimanite occurring in the thicker horizons. In thin section this sample contains both fibrolitic and prismatic sillimanite. Fibrolite defines a weak foliation (D_1) that wraps around anhedral biotite porphyroblasts (up to 4 mm) (Fig. 5b). Prismatic sillimanite cross cuts fibrolite and appears to be randomly oriented. Biotite, muscovite and plagioclase crystals have embayed edges. Fine-grained aggregates of pinite are interpreted to be pseudomorphs after cordierite (Fig. 5c). Very rare grains of anhedral garnet occur in association with biotite (Fig. 5d–e). Equigranular quartz crystals do not appear to be recrystallised and rarely display undulose extinction. Ilmenite crystals are distributed more-or-less evenly throughout the sample. Secondary tourmaline overprints biotite. Large plates of randomly oriented muscovite are interpreted to be secondary, probably replacing K-feldspar. The peak metamorphic assemblage (M_1) is sillimanite–biotite–

cordierite–K-feldspar–plagioclase–quartz–ilmenite–melt. Rare anhedral grains of garnet are interpreted to have grown during the prograde evolution.

GSWA 216533: garnet–biotite–epidote pelitic gneiss

In outcrop, the sample is characterised by a gneissic layering and subparallel leucocratic veins (Fig. 4c). The metamorphic foliation trends 241°/30° (SE). In thin section, this sample has a well-developed foliation defined by biotite that wraps garnet porphyroblasts (Fig 5f). The gneissic fabric is defined by alternating layers of biotite and elongate recrystallised quartz. The rock mainly comprises biotite, quartz, muscovite and garnet with minor epidote and ilmenite, and trace plagioclase. with accessory monazite. Garnet porphyroblasts (up to 0.5 mm) are euhedral to subhedral, the latter showing embayed edges. Randomly-oriented inclusions within garnet cores consist of quartz, muscovite and biotite whereas epidote inclusions occur in garnet rims. Late alteration to chlorite post-dates the peak metamorphic assemblage (M₁) of garnet–biotite–quartz–epidote–muscovite–plagioclase–ilmenite.

P01081: quartz–muscovite–tourmaline pelitic schist

This sample is a quartz–muscovite–tourmaline pelitic schist (Fig. 4d). In thin section it consists of quartz and muscovite, with minor tourmaline and plagioclase and accessory Fe-oxides. Rare plagioclase crystals are subhedral to anhedral, range from 1 to 4 mm in length and are partially sericitised. Muscovite is euhedral to anhedral and up to 4 mm. Quartz crystals are typically ~1mm in size and also occur as smaller recrystallised aggregates; both varieties shows undulose extinction. Quartz and muscovite are intergrown, suggesting they

replace K-feldspar. The tourmaline is typically subhedral and is similar in size to the smaller quartz and muscovite crystals.

P16877: quartz-biotite-muscovite pelitic schist

In the field the schist forms a green sub-crop (Fig. 4e) and consists of muscovite, quartz, biotite and sericite with accessory tourmaline. The schist is a protomylonite with a well-developed S-C fabric (S_1 or S_2).

GSWA 216538: biotite-muscovite-tourmaline pelitic gneiss

This sample is a metre-scale inclusion of deformed pelitic gneiss within a leucocratic monzogranite (GSWA 219741) (Fig. 4f). This pelitic gneiss has a well-defined biotite foliation (S_1) alternating with layers of elongate and tabular quartz surrounded by a fine-grained groundmass of sericite. Biotite also occurs as relict porphyroblasts up to 4.5 mm long. Deformed muscovite porphyroblasts (up to 8 mm long) are aligned with the fabric. Prismatic tourmaline crystals (up to 4 mm long) are intergrown with quartz and are also aligned with the foliation. Secondary chlorite and sericite indicate some post-peak retrogression.

5. Analytical methods

5.1 In situ SHRIMP U-Pb monazite and xenotime geochronology

For six pelitic samples (P01081, P16877 and GSWA 216533, 216338, 219742 and 219749) and three granitic samples (P01085 and GSWA 195890, 219741) polished thin sections were imaged using a scanning electron microscope (SEM) in back-scattered electron (BSE) mode fitted with an energy dispersive X-ray spectrometer (EDS) to identify suitable monazite and xenotime

crystals for *in situ* SHRIMP geochronology. Monazite and xenotime crystals >10 µm across were drilled out in 3 mm-diameter plugs and cast in a several 25 mm epoxy mounts. In all sessions the reference standards for Pb/U and Pb/Th calibrations and for $^{207}\text{Pb}/^{206}\text{Pb}$ fractionation monitoring were in separate mounts, which were cleaned and Au-coated together with the sample mounts to ensure consistent conductivity.

U–Pb analyses of monazite were conducted using a SHRIMP II ion microprobe in the John de Laeter Centre at Curtin University, Perth. Optical and BSE images were used to guide placement of the primary ion beam during analysis. SHRIMP analytical procedures followed established methods for monazite (Fletcher et al., 2010) and xenotime (Fletcher et al., 2000, 2004). During all analytical sessions, an O_2^- primary beam, with a spot size of 10 to 15 µm, was focussed through a 30 to 50 µm Kohler aperture with a beam intensity of 0.18 to 0.45 nA. The secondary ion beam was focused through a 100 µm collector slit onto an electron multiplier to produce mass peaks with flat tops and a mass resolution (1% peak heights) of >5,200. A post-collector retardation lens was used to reduce background counts produced from stray ions.

Monazite was analysed with a 13-peak run table as defined by Fletcher et al. (2010), which includes mass stations for the estimation of La, Ce and Nd (REEPO_2^+), and Y (YCeO^+). Monazite data were collected in sets of eight scans with a 5:1 unknown to standard ratio. Count times per scan for ^{204}Pb , background position 204.045, ^{206}Pb , ^{207}Pb , and ^{208}Pb were 10, 10, 10, 30, and 10s, respectively. The primary Pb/U and Pb/Th standard used was MAD 1 (known as FRENCH; Foster et al., 2000). Matrix effects on Pb/U data from U and Th were determined using Z2234 standard (Stern and Sanborn, 1998). Standard

Z2908 (provided by Richard Stern) was used to monitor instrumental mass fractionation (IMF) of the instrument.

Some of the analysed monazite grains in this study show distinct alteration corona that indicate fluid-mediated alteration (Finger et al., 1998). However, as the aim of this study was to constrain the timing of peak metamorphism and magmatism rather than the processes of secondary alteration processes, we only analysed primary monazite.

Xenotime was analysed with a nine-peak run table and analytical procedure followed established methods (Fletcher et al., 2000, 2004). The primary Pb/U standard MG-1 (Fletcher et al., 2004) was used for Pb/U, Pb/Th and U abundance calibrations. The secondary standards were z6413 (XENO1: Fletcher et al., 2004; Stern and Rainbird, 2001) used for $^{207}\text{Pb}/^{206}\text{Pb}$ normalisation and used in conjunction with MG-1 for matrix corrections to Pb/U and Pb/Th, and z6410 (XENO2: Fletcher et al., 2004; Stern and Rainbird, 2001) used to monitor matrix effects.

Squid-2.50.11.02.03 software (Ludwig, 2009) was used for initial data reduction, including correction for common Pb. Common Pb corrections were based on individual measured ^{204}Pb abundances, assuming crustal common Pb at the approximate age of the samples modelled by Stacey and Kramers (1975). Corrections for matrix effects in Pb/U and Pb/Th, from U and Th in xenotime (Fletcher et al., 2004) and from U, Th, Pb, and REE in monazite and for instrumental mass fractionation in $^{207}\text{Pb}/^{206}\text{Pb}$ (Fletcher et al., 2010) were applied to the sample data. Weighted mean dates are reported with 95% confidence limits, whereas individual analyses are presented with 1σ uncertainties.

5.3. Whole-rock geochemistry

Whole-rock geochemical analyses of granitic and pelitic samples, including loss on ignition (LOI), were determined by X-ray fluorescence (XRF) spectroscopy at Intertek-Genalysis Australia or at ALS Chemex, Australia by the Geological Survey of Western Australia. All data and additional analytical details are available from the Geological Survey of Western Australia's WACHEM database (<http://geochem.dmp.wa.gov.au/geochem/>). The data were used to determine zircon saturation temperatures and to determine the alumina saturation index (ASI). Pelitic samples were analysed specifically to constrain bulk compositions for use in phase equilibrium modelling.

The granites display a range of SiO₂ compositions (62.9–74.0% wt %), with 94% having >65 wt% SiO₂ (Table 2). The granites are generally peraluminous (i.e., ASI = molar Al₂O₃/(CaO + Na₂O + K₂O) >1 (Supp. Fig. 1); Chappell, 1999; Scaillet et al., 2016; Zen, 1986)--most have ASI values that cluster tightly between 1.12 and 1.20, but two samples have much higher values of 1.55 and 1.61 (Table 2). All but one sample has Zr contents between 70 and 222 ppm--one samples outside this range has 316 ppm Zr (Table 2).

6. Results

6.1 U–Pb monazite and xenotime geochronology

The results are presented as two groups: first granitic rocks and then pelites. The results are discussed in chronological order rather than by the geochronometer used. Figure 6 contains SEM images of analysed monazite and xenotime grains; Figure 7 consists of U–Pb monazite concordia diagrams for the

granitic rocks, Figure 8 consists of U–Pb monazite and xenotime concordia diagrams for the pelitic rocks.

Granitic rocks

GSWA 195890: muscovite-biotite metamonzogranite

The monazite grains are anhedral to subhedral with some crystals showing retrograde corona (Fig. 6a). The crystals range in size from 30 to 80 μm and occur as inclusions in quartz, plagioclase, and muscovite and within the groundmass. Fifty-four analyses were carried out on 29 monazite crystals (Supp. Table 1). Twenty-six analyses are excluded from the final age calculation due to high common ^{206}Pb (>1 wt%) and/or $>5\%$ discordance. A further four analyses are statistical outliers (i.e., >2 standard deviations from the mean) and are not considered further. The remaining twenty-four analytical spots show Th concentrations typically ranging from 50,000 to 93,000 ppm with one analysis at 26,000 ppm, whereas U concentrations cluster into three groups: 300–1300 ppm (fourteen analyses), 2000–3000 ppm (eight analyses) and 5000–6000 ppm (two analyses). The Th/U ratios range from 10 to 198. The 24 analyses yield $^{207}\text{Pb}/^{206}\text{Pb}$ dates between 1732 and 1640 Ma with a weighted mean of 1695 ± 9 Ma (MSWD = 0.62), which is interpreted as the crystallisation age of the monzogranite (Fig. 7a).

P01085: biotite–muscovite syenogranite

The monazite crystals are anhedral to subhedral, and range in size from 30 to 60 μm . Some have secondary apatite and/or retrograde coronas around their margins (Finger et al., 1998; Fig. 6b). The monazites occur both in the

matrix and as inclusions in quartz. Fourteen analyses were conducted on five monazite crystals (Supp. Table 1). Excluded from the age calculations is one analysis >5% discordant that likely represents an overlap between unaltered monazite and the retrograde corona. The remaining 13 analyses have U concentrations of 1100–4600 ppm and tightly-clustered Th contents of 69,000–94,000 ppm, with Th/U ratios from 15 to 72. The 13 analyses yielded $^{207}\text{Pb}/^{206}\text{Pb}$ dates between 1694 Ma and 1676 Ma and a weighted mean of 1680 ± 3 Ma (MSWD = 0.81), which is interpreted as the crystallisation age of the syenogranite (Fig. 7b).

GSWA 219741 tourmaline–muscovite monzogranite

The monazite grains are typically anhedral, and some have retrograde coronas (Fig. 6c). The crystals range in size from 25 to 70 μm and occur in the groundmass and as inclusions in quartz, plagioclase, and muscovite. Thirteen analyses were carried out on seven monazite crystals (Supp. Table 1). Two analyses of a single crystal have high common ^{206}Pb contents (>1%) and were not considered in the final age calculation, nor were a further five analyses >5% discordant. U and Th concentrations of the six remaining analyses are 750–7750 ppm and 13,000–64,500 ppm, respectively, with Th/U ratios of 8–35. The main group of six analyses yielded $^{207}\text{Pb}/^{206}\text{Pb}$ dates between 1671 and 1639 Ma, with a weighted mean of 1659 ± 13 Ma (MSWD = 0.47), which is interpreted as the crystallisation age of the monzogranite (Fig. 7c).

Pelitic rocks

GSWA 219749: cordierite–sillimanite pelitic migmatite

Monazite crystals in this sample are unzoned, anhedral to subhedral, and range in size from 15 to 50 μm . They occur as inclusions in quartz, sillimanite, muscovite, and biotite and are also interstitial to the main silicate phases (Fig. 6d). Twenty analyses were carried out on 11 monazite crystals (Supp. Table 1). Excluded from the age calculations is one discordant analysis (>5%) that overlapped an inclusion. The remaining 19 analyses have U concentrations of 1800 to 4100 ppm and consistently high Th concentrations (35,900–54,000 ppm), with Th/U ratios of 11–24. They yielded $^{207}\text{Pb}/^{206}\text{Pb}$ dates between 1710 Ma and 1652 Ma, and a weighted mean of 1691 ± 7 Ma (MSWD = 0.79) interpreted as the timing of monazite growth during migmatization (Fig. 8a).

GSWA 219742: sillimanite–biotite pelitic gneiss

Subhedral monazite crystals range from 35 to 60 μm and occur as inclusions in quartz, biotite, plagioclase and muscovite. Fe-oxides commonly occur around the margins of the monazite crystals (Fig. 6e). Some grains have a corona of apatite (e.g. Finger et al., 1998), for which a smaller spot size was used to avoid these retrograde areas. Twenty-one analyses were carried out on 11 monazite crystals (Supp. Table 1). Th concentrations range from 39,000 to 60,000 ppm, and U concentrations are mostly 550–5000 ppm (17 analyses) with four analyses having much higher concentrations (8000–10,100 ppm). Th/U ratios show a wide range from 4 to 90. One analysis with >5% discordance was excluded from the final age calculation. The 20 analyses yield $^{207}\text{Pb}/^{206}\text{Pb}$ dates between 1713 and 1624 Ma, and a weighted mean of 1685 ± 9 Ma (MSWD = 1.2) interpreted as the timing of monazite growth during metamorphism (Fig. 8b).

530 *GSWA 216533: garnet–biotite–epidote pelitic gneiss*

531 The monazite crystals range in size from 15 to 40 µm, are anhedral to
532 euhedral, and occur as inclusions in quartz and in garnet cores, and within the
533 matrix (Fig. 6f–h). Monazite inclusions are unzoned and unaltered. In contrast,
534 monazite crystals in the matrix are highly altered to bastnaesite (Fig. 6h) and
535 were unsuitable for SHRIMP analysis due to the small size of surviving monazite
536 (<10 µm). Eight analyses were carried out on five monazite crystals (Supp. Table
537 1). Excluded from the age calculations is one discordant analysis (>5%
538 discordance). The seven remaining analyses have tightly-clustered U and Th
539 concentrations that vary from 1600 to 5000 ppm and 23,000–48,650 ppm,
540 respectively, and Th/U ratios of 8–17. They yield $^{207}\text{Pb}/^{206}\text{Pb}$ dates between
541 1695 and 1673 Ma, and a weighted mean of 1683 ± 6 Ma (MSWD = 0.35)
542 interpreted as the age of monazite growth during metamorphism (Fig. 8c).

543

544 *P01081: quartz–biotite tourmaline schist*

545 The monazites are unzoned, mainly subhedral, and range in size from 20
546 to 70 µm. They occur as inclusions in quartz and muscovite and do not appear to
547 be preferentially aligned with the fabric (Fig. 6i). Some crystals show secondary
548 corona. Eleven analyses were carried out on five monazite crystals (Supp. Table
549 1). Excluded from the age calculations are two analyses that are >5% discordant.
550 The remaining nine analyses show a large spread of U and Th concentrations
551 (700–14,100 ppm and 2400–68,500 ppm, respectively), and Th/U ratios of 1–22.
552 They yield $^{207}\text{Pb}/^{206}\text{Pb}$ dates between 1702 and 1666 Ma and a weighted mean of
553 1681 ± 9 Ma (MSWD = 0.44) interpreted as the timing of monazite growth during
554 metamorphism (Fig. 8d).

P16877: quartz–muscovite–biotite schist

The monazites are unzoned, mainly anhedral, range in size from 15 to 40 μm , and are hosted in muscovite and quartz (Fig. 6k). There is no obvious alignment with the enclosing minerals. Twenty analyses were carried out on 13 monazite crystals (Supp. Table 1). One analysis >5% discordant was excluded from the final age calculation. The remaining 19 analyses indicate U concentrations of 800–7700 ppm, bimodal Th concentrations of 15,000–18,000 ppm and 23,000–45,000 ppm, and Th/U ratios of 2–32. They yield $^{207}\text{Pb}/^{206}\text{Pb}$ dates between 1701 and 1667 Ma, and a weighted mean of 1680 ± 7 Ma (MSWD = 0.33) interpreted as the timing of monazite growth during metamorphism (Fig. 8e).

GSWA 216538: biotite–muscovite–tourmaline pelitic gneiss

Both monazite and xenotime from this sample were analysed. Anhedral to subhedral monazite crystals range from 20 to 80 μm and occur as inclusions in quartz and muscovite, with one matrix grain (Fig. 6k). Some crystals show resorption around their margins. Seventeen analyses were carried out on 10 monazite crystals (Supp. Table 1). Th concentrations range from 36,500 to 58,500 ppm and U concentrations between 2700 and 8300 ppm, with Th/U ratios of 6–19. Two analyses with >5% discordance were excluded from the final age calculation, as was one young outlier with a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1656 ± 8 Ma (1602B.1-2). However, this latter outlier may record contact metamorphism associated with the intrusion of the granite host of this gneissic inclusion (sample GSWA 219741). The remaining 14 analyses yield $^{207}\text{Pb}/^{206}\text{Pb}$ dates

between 1692 and 1668 Ma, and a weighted mean of 1680 ± 4 Ma (MSWD = 0.89) interpreted as the timing of monazite growth during metamorphism (Fig. 8f).

Unzoned, unaltered anhedral to subhedral xenotime crystals are 20 to 60 μm in size, and typically occur as inclusions in muscovite and rarely as matrix grains (Fig. 6l). Fourteen analyses were carried out on six xenotime crystals (Supp. Table 1). Uranium concentrations range from 9700 to 17,800 ppm and Th concentrations between 1500 and 2700 ppm, with Th/U ratios of 0.1–0.3. Two analyses showing >5% discordance were excluded from the final age calculation, as was one young outlier with a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 1653 ± 4 Ma (1604G.1-2). The 11 remaining analyses yield $^{207}\text{Pb}/^{206}\text{Pb}$ dates between 1698 and 1670 Ma, and a weighted mean of 1680 ± 4 Ma (MSWD = 2.1) interpreted as the timing of xenotime growth during metamorphism (Fig. 8g).

8. Phase equilibria modelling

Repeated episodes of reworking and reactivation affected rocks of the Mangaroon Zone, making it challenging to find suitable samples for P – T investigations. However, two samples (GSWA 219749 and 219742) from the Pooranoo Metamorphics (Fig. 2) preserve low-variance mineral assemblages best suited to constrain P – T conditions. Based on the presence of leucosomes, samples GSWA 219749 (sillimanite–cordierite pelitic migmatite) and GSWA 219742 (sillimanite–biotite pelitic gneiss) are both interpreted to be former melt-bearing rocks and to have equilibrated above the solidus.

Isochemical phase diagrams (pseudosections) were calculated based on the bulk compositions of the two samples (Table 3). FeO contents were analysed

by titration and Fe_2O_3 was calculated by difference. The H_2O content assumes all LOI as H_2O . As the rocks are interpreted to have been melt-bearing, and likely lost melt at or near the metamorphic peak, their modelled bulk compositions are only valid for constraining the peak and post-peak evolution of these samples. Phase equilibrium modelling used THERMOCALC 3.40i and the internally consistent dataset of Holland and Powell (2011) (tcds62 generated on 6 February 2012), and were calculated in the system $\text{MnO}-\text{Na}_2\text{O}-\text{CaO}-\text{K}_2\text{O}-\text{FeO}-\text{MgO}-\text{Al}_2\text{O}_3-\text{SiO}_2-\text{H}_2\text{O}-\text{TiO}_2-\text{Fe}_2\text{O}_3$ (MnNCKFMASHTO), using the $a-x$ models from White et al. (2014).

8.1 Results

GSWA 219749: sillimanite–cordierite pelitic migmatite

The calculated solidus lies between 670–720°C and 2–8 kbar. The interpreted peak metamorphic assemblage of cordierite–sillimanite–magnetite–biotite–K-feldspar–quartz–plagioclase–ilmenite–melt is stable at 665–755°C and 2.7–4.3 kbar (Fig. 9a). The presence of sillimanite provides a lower temperature constraint, whereas the absence of garnet bounds the inferred peak assemblage field to higher pressure. The presence of biotite and cordierite provide upper and lower temperature limits, respectively.

GSWA 219742: sillimanite–biotite–pelitic gneiss

The calculated solidus lies between 670–700°C and 2–12 kbar. The interpreted metamorphic assemblage of sillimanite–biotite–cordierite–K-feldspar–plagioclase–quartz–ilmenite–melt is stable in a narrow field that extends from around 4 kbar at ~675 °C to around 7.5 kbar at 850 °C (Fig. 9b).

Although the modelled bulk composition is not strictly valid for constraining the prograde evolution due to likely melt loss, the presence of rare grains of anhedral, partially resorbed garnet, suggests higher pressures than those recorded at the peak during the prograde history of this sample (Fig. 9b). This implies a broad clockwise metamorphic P - T path, as illustrated in Fig. 9b.

9. Discussion

Prior to the study of Sheppard et al. (2005), the deformation fabrics and magmatism observed in the Mangaroon Zone were attributed to the 1820–1770 Ma Capricorn Orogeny. However, subsequent field mapping and SHRIMP U–Pb zircon geochronology in the southern part of the Mangaroon Zone, revealed an orogenic event at c. 1680–1620 Ma (Sheppard et al., 2005). Although this study defined the approximate timing of the Mangaroon Orogeny, there remains significant uncertainty on the timing and duration of metamorphism and magmatism, and how these processes relate to each other.

9.1 Duration of magmatism associated with the Mangaroon Orogeny

Before this study, the age range for the intrusion of the Durlacher Supersuite was 1680–1620 Ma (Sheppard et al., 2005). The monazite geochronology demonstrates that magmatism commenced earlier than previously thought, at 1695 ± 9 Ma (GSWA 195890). Furthermore, recent U–Pb geochronology on zircon from granites yields crystallisation ages of 1689 ± 6 Ma (GSWA 208365: Wingate et al., 2017), 1689 ± 4 Ma (GSWA 195888: Wingate et al., 2018) and 1688 ± 4 Ma (GSWA 216532: Wingate et al., 2019a). Combined, these data confirm that magmatism began earlier than 1680 Ma (Fig. 10) and

place a new age constraint on the minimum depositional age of the protoliths of the Pooranoo Metamorphics of >1695 Ma. Because of Neoproterozoic overprinting (fig. 11 of Piechocka et al., 2018) at the locality from which sample GSWA 195890 was collected, it is difficult to assign with any certainty an age to the well-developed fabric formed during the Mangaroon Orogeny. Ar–Ar dating of biotite from the mylonite zone yielded a Neoproterozoic age interpreted as the timing of mylonite formation during fault reactivation (Piechocka et al., 2018).

Our new results combined with previously published zircon data (Table 2) show that magmatism in the Mangaroon Zone may have occurred in two pulses. The first episode likely occurred at 1695–1673 Ma, overlapping with D₁ deformation, with a second episode younger than 1660 ± 9 Ma (constrained by sample GSWA 169059, Table 2) that possibly accompanied D₂ deformation. The only published age for a presumed Durlacher Supersuite granite to lie outside of this 1695–1660 Ma range is the 1620 Ma age for the Discretion Granite (GSWA 142855; Nelson, 1998), located in the southernmost Capricorn Orogen (Fig.1). The isolated nature of this body, which intrudes the Yarlalweelor Gneiss Complex, and its anomalously young age, suggests it may be unrelated to the Mangaroon Orogeny. Therefore, magmatism in the Mangaroon Zone lasted for less than 35 million years (1695–1660 Ma; Fig. 10). However, recent ⁴⁰Ar/³⁹Ar mica geochronology, from a monzogranite yielded a c. 1640 Ma cooling age (Piechocka et al., 2018), and it may be possible that magmatism continued until as late as 1640 Ma.

9.2 P–T–t conditions of early metamorphism

Phase equilibrium modelling indicates low- P , high- T metamorphic conditions (Fig. 9). In particular, the calculated peak field for sample GSWA 219749, at 665–755 °C and 2.7–4.3 kbar is relatively well constrained and indicates that metamorphism occurred at relatively shallow crustal levels (<15 km depth) assuming no significant tectonic overpressure. The interpreted peak metamorphic assemblage of sillimanite–biotite–cordierite–K-feldspar–melt in sample GSWA 219742 suggests higher P – T conditions of ~800 °C, 6–7 kbar. The inferred presence of rare prograde garnet is consistent with some near-peak decompression and a clockwise P – T path (Fig. 9b), although the amount of decompression is unconstrained.

9.3 Timing, duration and causes of metamorphism

The new phosphate geochronology from samples across the Mangaroon Zone show that peak metamorphism occurred at c. 1690 ± 7 Ma and continued until at least 1680 ± 4 Ma (Fig. 10). The absence of younger (<1680 Ma) metamorphic phosphate growth suggests that regional metamorphism may have been relatively short-lived at around 10–20 million years (i.e., 1698–1676 Ma).

The oldest metamorphic monazite age of c. 1691 Ma comes from the cordierite–sillimanite migmatite (GSWA 219749, Fig. 10). This sample defines an apparent thermal gradient of 175 to 240 °C/kbar (Fig. 11), compared to an ambient thermal gradient calculated for the Gascoyne Province of ~46 °C/kbar (calculated at 1320 Ma; fig. 8 of Korhonen and Johnson, 2015). Furthermore, in comparison to typical thermal gradients in the crust (Fig. 1 of Morrissey et al., 2014) the thermal gradient in the Mangaroon Zone can be considered anomalous. High thermal gradients are commonly associated with magmatism

and significant extension (e.g., Smithies et al., 2011; Sandiford and Powell, 1986), or rapid exhumation of deep crustal rocks (e.g., Cooke and O'Brien, 2001). Our geochronology shows that metamorphism was coeval with magmatism at c. 1690 Ma.

Although the geochronology, in this study, may initially point to short duration metamorphism, the absence of post c. 1680 Ma dates does not mean that metamorphism was not more protracted considering that magmatism continued until 1660 Ma. For example, monazite crystals in the matrix in sample GSWA 216533 (Fig. 6h) were so altered such that they could not be dated. Thus, the lack of post-1680 Ma metamorphic ages could be due to pervasive retrogression by fluids released from later magmatism (related to D₂) and/or by fluids flowing through crustal-scale fault systems (e.g., Johnson et al., 2013), possibly during the later stages of D₁ or during D₂. Furthermore, Sheppard et al. (2005) assigned the greenschist-facies schistosity to D₂ deformation, meaning the lower grade mineral assemblages and destruction of monazites in sample 216533 may have been the result of D₂ deformation.

There is some hint of a c. 1660 Ma metamorphic episode recorded in a deformed pelitic gneiss (GSWA 216538) that is intruded by tourmaline–muscovite monzogranite at 1659 ± 13 Ma (GSWA 219741) (Fig. 4f). Sample GSWA 216538 yielded a single monazite analysis with an age of 1656 ± 8 (1σ) Ma and a single xenotime analysis with an age of 1653 ± 4 (1σ) Ma (GSWA 216538, Supp. Table 1), which may reflect contact metamorphism associated with the intrusion of monzogranite (1659 ± 13 Ma; GSWA 219741), possibly during D₂ deformation.

9.4 Temperatures and causes of granite emplacement

Zircon saturation temperatures were used to evaluate the emplacement temperatures of Durlacher Supersuite granites in the Mangaroon Zone. Zirconium contents in the analysed granites range from 70 to 316 ppm, with calculated temperatures ranging from 733 to 863 °C (Supp. Fig. 2), based on the M parameter of Watson and Harrison (1983). These results are also consistent with Miller et al. (2003), who demonstrated that granites with 80–150 ppm Zr record a mean temperature of 766 ± 24 °C and granites with 200–800 ppm Zr record a temperature mean 837 ± 48 °C. Field mapping in the Mangaroon Zone shows that granites account for >50% of the Mangaroon Zone (at surface) (Fig. 2), and extend to at least 10 km depth (Johnson et al., 2013), but it is unclear whether the older or younger granites make up the greater proportion of the crust. These factors suggest that the heat required for metamorphism was in part derived from the granites.

The Durlacher Supersuite magmatism is characterised by melting and recycling of older crust (i.e., the 1820–1775 Ma Moorarie Supersuite) with no evidence for a significant mantle component (fig. 4 of Johnson et al., 2017). The peraluminous nature of the Durlacher granites in the Mangaroon Zone suggests they also have a metasedimentary source (Table 2) and Johnson et al., (2017) suggested the Durlacher granites are likely to have been derived from partial melting of the Pooranoo Metamorphics, consistent with the age of xenocrystic zircon (Supp. Fig. 3).

Heat production values of crustal rocks in the Mangaroon Zone show that the source components are quite radiogenic, with average values of 3.5 and 4 $\mu\text{W}/\text{m}^3$ (at 1.7 Ga) for the Pooranoo Metamorphics and the Moorarie Supersuite,

respectively (F. J. Korhonen, 2018, personal communication). Thermal modelling of the Mesoproterozoic Gascoyne Province crust showed that it would take c. 140 million years to reach temperatures in excess of 650 °C at depths greater than 15 km (Fig. 8 of Korhonen and Johnson, 2015). However, it is unclear whether the conditions during the Mesoproterozoic are applicable to the Mangaroon Orogeny. Nevertheless, in the Mangaroon Zone, at c. 1.7 Ga, anatexis of the Pooranoo Metamorphics (protoliths were deposited, buried and melted between c. 1758 and 1695 Ma) to generate the Durlacher Supersuite occurred at depths of less than about 15 km. This time frame (c. 63 million years) suggests there was probably insufficient time for a thermal lid, overlying the Pooranoo Metamorphics, to incubate and heat the crust and cause melting. Therefore, it appears unlikely that radiogenic heating alone was the heat source for crustal melting.

9.5 New age constraints on D_1/D_2

Constraining D_1 deformation has been challenging due to a lack of consistent field relationships and structures across the Mangaroon Zone. The previous upper age constraint on D_1/M_1 , from the southern Mangaroon Zone, was c. 1675 Ma (Sheppard et al., 2005), but recent zircon geochronology showed that a deformation fabric previously assigned to the Mangaroon Orogeny formed during an older orogenic event (Wingate et al., 2013). Furthermore, there was a lack of direct constraints on the age of metamorphism and uncertainty about the timing of granite magmatism across the Mangaroon Zone. The geochronology from the northern, central and southern Mangaroon Zone, demonstrate that metamorphism and deformation commenced at 1691 ± 7 Ma,

coeval with magmatism at 1695 ± 9 Ma. Therefore, the new lower constraint on the onset of deformation and metamorphism, and coeval magmatism, D_1/M_1 during the Mangaroon Orogeny is c. 1695 Ma. In contrast to the metamorphic assemblages reported by Sheppard et al. (2005), the new interpretation of M_1 assemblage includes the presence of K-feldspar (in GSWA 219749 and 219742) and early garnet in GSWA 219742.

Although the timing of peak metamorphism in sample GSWA 216533 from the northern Mangaroon Zone was constrained at 1683 ± 6 Ma based on data from monazite inclusions in garnet cores and monazite within matrix quartz, the timing of foliation development and the growth of garnet rims could not be dated due to severe retrogression and small size of the matrix monazite. However, recent zircon geochronology from granites in the northern Mangaroon Zone, only ~ 700 m to the west of GSWA 216533, may be used to interpret the timing of foliation development in sample GSWA 216533. There, an undeformed muscovite–tourmaline monzogranite with an age of 1673 ± 4 Ma (GSWA 216531: Wingate et al., 2019b) intrudes a foliated biotite–metagranodiorite with an age of 1688 ± 4 Ma (GSWA 216532: Wingate et al., 2019a). This relationship indicates that the foliation (D_1) in the garnet–biotite–epidote pelitic gneiss (GSWA 216533) may have developed between 1688 ± 4 and 1673 ± 4 Ma, which overlaps with the timing of D_1 in the central (GSWA 219742) and southern Mangaroon Zone (GSWA 219749) at 1685 ± 9 Ma and 1691 ± 7 Ma, respectively. Although, the timing of D_1 deformation is constrained the lack of megascopic structures across the Mangaroon Zone, typical of compressional or transpressional settings, suggests that D_1 may have occurred in an extensional setting.

Compared to the limited structures associated with D_1 deformation, D_2 deformation produced a pervasive schistosity, metre- to kilometre-scale upright folds, and retrogression of D_1 metamorphic minerals to greenschist facies assemblages (Sheppard et al., 2005; 2010b). The age constraints for granite magmatism in the Mangaroon Zone show that there may have been two pulses of magmatism associated with the Durlacher Supersuite. The first pulse at 1695–1673 Ma, coincides with D_1 deformation and peak metamorphism. The second pulse of magmatism at c. 1660 Ma maybe be linked with D_2 deformation. The existing constraints for D_2 come from deformed granites older than c. 1675 Ma that suggests regional deformation occurred after 1675 Ma. The interpreted crystallisation age of 1659 ± 13 Ma (GSWA 219741), from the undeformed tourmaline–muscovite monzogranite, provides a new minimum age for D_2 , which is constrained to the interval c 1675–1660 Ma.

9. 6 Geodynamic setting of the Mangaroon Zone

Many models have been invoked for low- P , high- T metamorphism, but the most common one for Proterozoic Australian orogens is the thermal lid model. In this section we will discuss how the commonly applied thermal lid model does not apply to the metamorphism associated with the Mangaroon Orogeny. Any model for the Mangaroon Orogeny must explain the following:

- burial and melting of the protoliths to the Pooranoo Metamorphics took place within c. 63 million years of them being deposited;
- granitic magmatism at c. 1695 Ma was coeval with peak metamorphism at c. 1691 Ma during D_1 deformation (Fig. 10);

- the Mangaroon Orogeny was associated with an anomalous thermal gradient 175 to 240 °C/kbar (Fig. 11) compared to an ambient thermal gradient of ~46 °C/kbar for the Gascoyne Province;
- the presence of rare grains of garnet at one locality, suggests higher pressures than those recorded at the peak during the prograde history, implying a broad clockwise metamorphic P - T path (Fig. 9b);

The thermal lid model requires high concentrations of heat-producing elements in the crust combined with an incubation period typically lasting >100 million years (Korhonen et al., 2015; 2017; Morrissey et al., 2014). Although the crust in the Mangaroon Zone was radiogenic at c. 1.7 Ga, deposition, burial and melting of the Pooranoo Metamorphics (source rocks for the Durlacher Supersuite) occurred within c. 63 million years. Therefore, it is unlikely that there was sufficient time for radiogenic heat build up to cause the upper crustal metamorphism during the Mangaroon Orogeny. If radiogenic heating alone was unlikely the cause of the crustal melting, another heat source is required. For instance, the elevated Moho below the Mangaroon Zone (Johnson et al., 2013) and the anomalous thermal gradient during peak metamorphism, which is commonly associated with magmatism and significant extension (Smithies et al., 2011), suggests lithospheric thinning during the Mangaroon Orogeny.

Mantle upwelling during lithospheric thinning is a common driver proposed for low- P , high- T metamorphism (Collins, 1994; Vilà et al., 2007). In the Mangaroon Zone there is a lack of evidence for mafic magmatism; therefore, it is difficult to say with certainty that there was a mantle contribution.

9.7 Mangaroon Orogeny: high-grade core, low-grade halo

Although medium- to high-grade rocks formed during the Mangaroon Orogeny are restricted to the Gascoyne Province, disparate studies in recent years suggest that the effects of the orogeny may be present across much of the West Australian Craton (Fig. 12). For instance, in the Ashburton Basin at the Paulsens Gold Mine, SHRIMP U–Pb monazite dating yielded an age of 1680 ± 9 Ma and at the nearby Belvedere orebody xenotime yielded an age of c. 1680 Ma, both of which are interpreted as the timing of hydrothermal gold mineralisation as a result of fault reactivation during the Mangaroon Orogeny (Fielding et al., 2017; 2018). Further north in the Pilbara Craton, hydrothermal monazite growth occurred at c. 1650 Ma (Rasmussen et al., 2007a), and in the southern Pilbara at Mt Tom Price, hydrothermal xenotime growth occurred at c. 1670 Ma (Rasmussen et al., 2007b). Both these SHRIMP U–Pb studies attributed phosphate growth to reactivation of pre-existing crustal-scale structures during the Mangaroon Orogeny. Furthermore, shear zones within the Sylvania Inlier in the Pilbara Craton yielded $^{40}\text{Ar}/^{39}\text{Ar}$ mica ages of c. 1650 Ma (Sheppard et al., 2006). Similarly, a c. 1650 Ma $^{40}\text{Ar}/^{39}\text{Ar}$ mica age, from quartz-mica schist, was identified in the Earraheedy Basin, and was interpreted as the timing of deformation within the Stanley Fold Belt (Pirajno et al., 2009). These scattered low-grade events highlight that what initially appeared to be a very localised high-grade event in the Mangaroon Zone had a significant regional impact across the West Australian Craton.

Tectonic activity occurring synchronously with the Mangaroon Orogeny is also known more widely in western and central Australia. For instance, in the adjacent Albany–Fraser Orogen, two tectonic episodes are known: the 1710–

1650 Ma Biranup Orogeny (Spaggiari et al., 2014) and, the c. 1680 Ma high-grade metamorphic Zanthus Event (Kirkland et al., 2011). In the Gawler Craton of the South Australia Craton (Fig. 1), ultra-high temperature metamorphism at 1690–1650 Ma was the result of initial crustal thinning followed by crustal thickening (Cutts et al., 2013). In the Arunta Inlier of the North Australian Craton (Fig. 1) the Argilke Event at 1680–1660 Ma involved magmatism, migmatisation and metamorphism (Collins and Shaw, 1995; Black and Shaw, 1995) and high temperature metamorphism associated with the 1640–1635 Ma Liebig Orogeny was the result of the accretion of the Warumpi Province onto the North Australian Craton (Scrimgeour et al., 2005). These (presently) separated occurrences of c. 1.7 Ga tectonism in the western, northern and southern parts of Australia suggests that the Mangaroon Orogeny was perhaps part of a continent-scale tectonic event.

Conclusions

The combined U–Pb geochronology and *P–T* results suggest that upper crustal metamorphism at 1691 ± 7 Ma (665–755 °C and 2.7–4.3 kbar) coincided with the oldest phases of the Durlacher Supersuite coeval at 1695 ± 9 Ma, and continued until 1680 ± 4 Ma, with continued magmatism until c. 1660 Ma, in the Mangaroon Zone. The monazite geochronology shows that magmatism commenced at least 15 million years earlier than previously thought. Although the geochronology suggests a short duration of metamorphism (~22 million years), associated with D₁, the magmatism continued until c. 1660 Ma. The absence of robust metamorphic dates younger than c. 1680 Ma does not imply that metamorphism ceased. Rather, the absence of dates is likely due to the

severe destruction of matrix monazite, possibly related to the continued magmatism. A low-grade schistosity fabric was previously assigned to D₂ deformation; therefore, there is indication of continued metamorphism.

Our new results show that, in the Mangaroon Zone, the Pooranoo Metamorphics were deposited, buried and melted in c. 63 million years (<1758 and >1695 Ma). The new constraints on the ages of sedimentation and magmatism, combined with the elevated thermal gradient immediately following sediment deposition, suggests that upper-crustal granite emplacement likely occurred in an extensional tectonic setting. Unlike other intraplate metamorphic events in Proterozoic Australia, the thermal lid model is unlikely to explain the metamorphism in the Mangaroon Zone during the Mangaroon Orogeny. Rather, our results show that magmatism was synchronous with upper crustal metamorphism. The low-grade halo of the Mangaroon Orogeny can be traced well outside of the Capricorn Orogen to the northern parts of the Pilbara Craton and the eastern Earaheedy Basin.

Acknowledgements

The authors thank Natasha Wodicka and an anonymous reviewer for their constructive and insightful comments resulting in a more focused manuscript. Simon Johnson is thanked for his assistance with sample collection and field work. This PhD project was funded through an Australian Research Council (ARC) grant (LP130100922) to BR and the Western Australian government Exploration Incentive Scheme. U–Pb measurements were conducted using the SHRIMP ion microprobes at the John de Laeter Centre at Curtin University. FJK

publishes with permission of the Executive Director of the Geological Survey of Western Australia.

References

Barton, M. D., & Hanson, R. B. (1989). Magmatism and the development of low-pressure metamorphic belts: Implications from the western United States and thermal modeling. *GSA Bulletin* 101, 1051–1065.

Black, L. P., & Shaw, R. D. (1995). An assessment, based on U–Pb zircon data, of Rb/Sr dating in the Arunta Inlier, central Australia. *Precambrian Research* 71, 3–15.

Bodorkos, S., and Wingate, M. T. D. (2007). The contribution of geochronology to GSWA's mapping programs: current perspectives and future directions: *Geological Survey of Western Australia, Record* 2007/2, p. 10–11.

Chappell, B. W. (1999). Aluminium saturation in I- and S-type granites and the characterization of fractionated haplogranites. *Lithos* 46, 535–551.

Clark, C., Kirkland, C. L., Spaggiari, C. V., Oorschot, C., Wingate, M. T. D., & Taylor, R. J. (2014). Proterozoic granulite formation driven by mafic magmatism: An example from the Fraser Range Metamorphics, Western Australia. *Precambrian Research* 240, 1–21.

Collins, W. J. (1994). Upper- and middle-crustal response to delamination: An example from the Lachlan fold belt, eastern Australia. *Geology* 22, 143–146.

Collins, W. J., & Shaw, R. D. (1995). Geochronological constraints on orogenic events in the Arunta Inlier: a review. *Precambrian Research* 71, 315–346.

Collins, W. J., & Vernon, R. H. (1991). Orogeny associated with anticlockwise P–T–t

951 paths: Evidence from low-P, high-T metamorphic terranes in the Arunta
 952 inlier, central Australia. *Geology* 19, 835–838.

953 Cooke, R. A., & O'Brien, P. J. (2001). Resolving the relationship between high P–T
 954 rocks and gneisses in collisional terranes: an example from the Gföhl
 955 gneiss–granulite association in the Moldanubian Zone, Austria. *Lithos* 58,
 956 33–54.

957 Cutts, K. A., Kelsey, D. E., & Hand, M. (2013). Evidence for late Paleoproterozoic
 958 (ca 1690–1665Ma) high- to ultrahigh-temperature metamorphism in
 959 southern Australia: Implications for Proterozoic supercontinent models.
 960 *Gondwana Research* 23, 617–640.

961 Fielding, I. O. H., Johnson, S. P., Zi, J. W., Rasmussen, B., Muhling, J. R., Dunkley, D.
 962 J., Sheppard, S., Wingate, M. T. D., & Rogers, J. R. (2017). Using In Situ
 963 SHRIMP U-Pb Monazite and Xenotime Geochronology to Determine the
 964 Age of Orogenic Gold Mineralization: An Example from the Paulsens Mine,
 965 Southern Pilbara Craton. *Economic Geology* 112, 1205–1230.

966 Fielding, I. O. H., Johnson, S. P., Zi, J.-W., Sheppard, S., & Rasmussen, B. (2018).
 967 Neighbouring orogenic gold deposits may be the products of unrelated
 968 mineralizing events. *Ore Geology Reviews* 95, 593–603.

969 Finger, F., Broska, I., Roberts, M. P., & Schermaier, A. (1998). Replacement of
 970 primary monazite by apatite-allanite-epidote coronas in an amphibolite
 971 facies granite gneiss from the eastern Alps. *American Mineralogist* 83,
 972 248–258.

973 Fletcher, I. R., McNaughton, N. J., Aleinikoff, J. A., Rasmussen, B., & Kamo, S.L.
 974 (2004). Improved calibration procedures and new standards for U-Pb and

975 Th-Pb dating of Phanerozoic xenotime by ion microprobe. Chemical
 976 Geology 209, 295–314.
 977 Fletcher, I. R., McNaughton, N. J., Davis, W. J., & Rasmussen, B. (2010). Matrix
 978 effects and calibration limitations in ion probe U–Pb and Th–Pb dating of
 979 monazite. Chemical Geology 270, 31–44.
 980 Fletcher, I.R., McNaughton, N.J., & Rasmussen, B. (2000). SHRIMP U–Pb
 981 geochronology of authigenic xenotime and its potential for dating
 982 sedimentary basins. Australian Journal of Earth Sciences 47, 845–859.
 983 Foster, G., Kinny, P., Vance, D., Prince, C., and Harris, N. (2000). The significance
 984 of monazite U–Th–Pb age data in metamorphic assemblages; a combined
 985 study of monazite and garnet chronometry. Earth and Planetary Science
 986 Letters, 181, 327–340.
 987 Glasson, K.J., Johnson, T.E., Kirkland, C.L., Gardiner, N.J., Clark, C., Blereau, E.,
 988 Hartnady, M.I., Spaggiari, C. and Smithies, H. (2019). A window into an
 989 ancient backarc? The magmatic and metamorphic history of the Fraser
 990 Zone, Western Australia. Precambrian Research, 323, 55–69.
 991 GSWA. (2016a). 1:500 000 State interpreted bedrock geology of Western
 992 Australia, 2016: Geological Survey of Western Australia, digital data layer.
 993 www.dmp.wa.gov.au/geoview
 994 GSWA. (2016b). Compilation of geochronology information, 2016 update, digital
 995 data product. www.dmp.wa.gov.au/geoview
 996 Holland T. J, B., & Powell, R. (2011). An improved and extended internally
 997 consistent thermodynamic dataset for phases of petrological interest,
 998 involving a new equation of state for solids. Journal of Metamorphic
 999 Geology 29, 333–383.

1000 Johnson, S. P., Sheppard, S., Rasmussen, B., Wingate, M. T. D., Kirkland, C. L.,
 1001 Muhling, J. R., Fletcher, I. R., & Belousova, E. A. (2011). Two collisions, two
 1002 sutures: Punctuated pre-1950Ma assembly of the West Australian Craton
 1003 during the Ophtharmian and Glenburgh Orogenies. *Precambrian Research*
 1004 189, 239–262.

1005 Johnson, S. P., Korhonen, F. J., Kirkland, C. L., Cliff, J. B., Belousova, E. A., &
 1006 Sheppard, S. (2017). An isotopic perspective on growth and
 1007 differentiation of Proterozoic orogenic crust: From subduction
 1008 magmatism to cratonization. *Lithos* 268, 76–86.

1009 Johnson, S. P., Thorne, A. M., Tyler, I. M., Korsch, R. J., Kennett, B. L. N., Cutten, H.
 1010 N., Goodwin, J., Blay, O., Blewett, R. S., Joly, A., Dentith, M. C., Aitken, A. R.
 1011 A., Holzschuh, J., Salmon, M., Reading, A., Heinson, G., Boren, G., Ross, J.,
 1012 Costelloe, R. D., & Fomin, T. (2013). Crustal architecture of the Capricorn
 1013 Orogen, Western Australia and associated metallogeny. *Australian Journal*
 1014 *of Earth Sciences* 60, 681–705.

1015 Kirkland, C. L., Spaggiari, C. V., Pawley, M. J., Wingate, M. T. D., Smithies, R. H.,
 1016 Howard, H. M., Tyler, I. M., Belousova, E. A., & Poujol, M. (2011). On the
 1017 edge: U–Pb, Lu–Hf, and Sm–Nd data suggests reworking of the Yilgarn
 1018 craton margin during formation of the Albany-Fraser Orogen.
 1019 *Precambrian Research* 187, 223–247.

1020 Korhonen, F. J., & Johnson, S. P. (2015). The role of radiogenic heat in prolonged
 1021 intraplate reworking: The Capricorn Orogen explained?. *Earth and*
 1022 *Planetary Science Letters* 428, 22–32.

1023 Korhonen, F. J., Johnson, S. P., Wingate, M. T. D., Kirkland, C. L., Fletcher, I. R.,

1024 Dunkley, D. J., Roberts, M. P., Sheppard, S., Muhling, J. R., & Rasmussen, B.
 1025 (2017). Radiogenic heating and craton-margin plate stresses as drivers
 1026 for intraplate orogeny. *Journal of Metamorphic Geology* 35, 631–661.
 1027 Loosveld, R. J. H., & Etheridge, M. A. (1990). A model for low-pressure facies
 1028 metamorphism during crustal thickening. *Journal of Metamorphic*
 1029 *Geology* 8, 257–267.
 1030 Ludwig, K.R. (2009). *Squid 2.50, A User's Manual*. Berkeley Geochronology
 1031 Centre Special Publication, pp. 95.
 1032 Martin, D. M., Sheppard, S., Thorne, A., Farrell, T. R., and Groenewald, P. B. (2006).
 1033 Proterozoic Geology of the Western Capricorn Orogen - a field guide:
 1034 Geological Survey of Western Australia, Record 2006/18, p.43.
 1035 McLaren, S., Sandiford, M., & Powell, R. (2005). Contrasting styles of Proterozoic
 1036 crustal evolution: A hot-plate tectonic model for Australian terranes.
 1037 *Geology* 33, 673–676.
 1038 Miller, C. F., Meschter McDowell, S., & Mapes, R. W., 2003, Hot and cold granites?
 1039 Implications of zircon saturation temperatures and preservation of
 1040 inheritance. *Geology* 31, 529–532.
 1041 Morrissey, L. J., Hand, M., Raimondo, T., and Kelsey, D. E., 2014, Long-lived high-T,
 1042 low-P granulite facies metamorphism in the Arunta Region, central
 1043 Australia: *Journal of Metamorphic Geology*, v. 32, no. 1, p. 25-47.
 1044 Nelson, D. R. (1998). 142855: porphyritic monzogranite, Anderson well; in
 1045 Compilation of SHRIMP U–Pb zircon geochronology data, 1997: Western
 1046 Australia Geological Survey, Record 1998/2, 194–197.
 1047 Nelson, D. R. (2002). 169053: biotite muscovite monzogranite, Fraser Well; in

1048 Compilation of geochronology data, 2001: Western Australia Geological
1049 Survey, Record 2002/2, p. 35–38.

1050 Nelson, D. R. (2005). 178027: biotite-muscovite granodiorite, Mangaroon
1051 Homestead; Geochronology dataset 536; in Compilation of geochronology
1052 data, June 2006 update; Western Australia Geological Survey.

1053 Occhipinti, S. A., Sheppard, S., Passchier, C., Tyler, I. M., & Nelson, D. R. (2004).
1054 Palaeoproterozoic crustal accretion and collision in the southern
1055 Capricorn Orogen: the Glenburgh Orogeny. *Precambrian Research* 128,
1056 237–255.

1057 Occhipinti, S. A., & Reddy, S. M. (2009). Neoproterozoic reworking of the
1058 Palaeoproterozoic Capricorn Orogen of Western Australia and
1059 implications for the amalgamation of Rodinia. Geological Society, London,
1060 Special Publications 327, 445–456.

1061 Oliver, N. H. S., & Zakowski, S. (1995). Timing and geometry of deformation, low-
1062 pressure metamorphism and anatexis in the eastern Mt Lofty Ranges: The
1063 possible role of extension. *Australian Journal of Earth Sciences* 42, 501–
1064 507.

1065 Pearson, J. M. (1996). Alkaline rocks of the Gifford Creek Complex, Gascoyne
1066 Province, Western Australia — their petrogenetic and tectonic
1067 significance: University of Western Australia, Perth, PhD thesis
1068 (unpublished).

1069 Piechocka, A. M., Gregory, C. J., Zi, J.-W., Sheppard, S., Wingate, M. T. D., &
1070 Rasmussen, B. (2017). Monazite trumps zircon: applying SHRIMP U–Pb
1071 geochronology to systematically evaluate emplacement ages of

1072 leucocratic, low-temperature granites in a complex Precambrian orogen.
 1073 Contributions to Mineralogy and Petrology 172, 1–17.
 1074 Piechocka, A. M., Sheppard, S., Fitzsimons, I. C. W., Johnson, S. P., Rasmussen, B., &
 1075 Jourdan, F. (2018). Neoproterozoic $^{40}\text{Ar}/^{39}\text{Ar}$ mica ages mark the
 1076 termination of a billion years of intraplate reworking in the Capricorn
 1077 Orogen, Western Australia. Precambrian Research 310, 391–406.
 1078 Piechocka, A. M., Zi, J. W., Gregory, C. J., Sheppard, S., and Rasmussen, B. (2019).
 1079 SHRIMP U–Pb phosphate dating shows metamorphism was synchronous
 1080 with magmatism during the Paleoproterozoic Capricorn Orogeny:
 1081 Australian Journal of Earth Sciences 66, 973–990.
 1082 Pirajno, F., Hocking, R. M., Reddy, S. M., & Jones, A. J. (2009). A review of the
 1083 geology and geodynamic evolution of the Palaeoproterozoic Earaheedy
 1084 Basin, Western Australia. Earth-Science Reviews 94, 39–77.
 1085 Rasmussen, B., Fletcher, I. R., & Muhling, J. R. (2007a). In situ U–Pb dating and
 1086 element mapping of three generations of monazite: Unravelling cryptic
 1087 tectonothermal events in low-grade terranes. Geochimica et
 1088 Cosmochimica Acta 71, 670–690.
 1089 Rasmussen, B., Fletcher, I. R., Muhling, J. R., Thorne, W. S., & Broadbent, G. C.
 1090 (2007b). Prolonged history of episodic fluid flow in giant hematite ore
 1091 bodies: Evidence from in situ U–Pb geochronology of hydrothermal
 1092 xenotime. Earth and Planetary Science Letters 258, 249–259.
 1093 Sandiford, M., & Hand, M. (1998). Controls on the locus of intraplate deformation
 1094 in central Australia. Earth and Planetary Science Letters 162, 97–110.
 1095 Sandiford, M., & Powell, R. (1986). Deep crustal metamorphism during

1096 continental extension: modern and ancient examples. *Earth and Planetary*
1097 *Science Letters* 79, 151–158.

1098 Scaillet, B., Holtz, F., & Pichavant, M. (2016). Experimental Constraints on the
1099 Formation of Silicic Magmas. *Elements* 12, 109–114.

1100 Scrimgeour, I. R., Kinny, P. D., Close, D. F., & Edgoose, C. J. (2005). High-T
1101 granulites and polymetamorphism in the southern Arunta Region, central
1102 Australia. Evidence for a 1.64 Ga accretional event. *Precambrian Research*
1103 142, 1–27.

1104 Sheppard, S., Bodorkos, S., Johnson, S. P., Wingate, M. T. D., & Kirkland, C. L.
1105 (2010a). The Paleoproterozoic Capricorn Orogeny: Intracontinental
1106 Reworking Not Continent-Continent Collision: Geological Survey of
1107 Western Australia, Report 108, p. 33.

1108 Sheppard, S., Farrell, T. R., Bodorkos, S., Hollingsworth, D., Tyler I., M., & Pirajno,
1109 F. (2006). Late Paleoproterozoic (1680–1620) sedimentation,
1110 magmatism, and tectonism in the Capricorn Orogen, *in* GSWA 2006
1111 extended abstracts: promoting the prospectivity of Western Australia:
1112 Geological Survey of Western Australia, Record 2006/3, p. 11–12.

1113 Sheppard, S., & Johnson, S. P. (2016). Pooranoo Metamorphics (P₁-PO-md):
1114 Geological Survey of Western Australia, WA Geology Online, Explanatory
1115 Notes extract, viewed 20 November 2017, www.dmp.wa.gov.au/ens.

1116 Sheppard, S., Johnson, S. P., Wingate, M. T. D., Kirkland, C. L., & Pirajno, F.
1117 (2010b). Explanatory notes for the Gascoyne Province: Geological Survey
1118 of Western Australia, 1:100 000 Explanatory Notes, p. 336.

1119 Sheppard, S., Occhipinti, S. A., & Nelson, D. R. (2005). Intracontinental reworking

1120 in the Capricorn Orogen, Western Australia: the 1680–1620 Ma
 1121 Mangaroon Orogeny. *Australian Journal of Earth Sciences* 52, 443–460.
 1122 Sheppard, S., Rasmussen, B., Muhling, J. R., Farrell, T. R., and Fletcher, I. R. (2007).
 1123 Grenvillian-aged orogenesis in the Palaeoproterozoic Gascoyne Complex,
 1124 Western Australia: 1030–950 Ma reworking of the Proterozoic Capricorn
 1125 Orogen. *Journal of Metamorphic Geology* 25, 477–494.
 1126 Smithies, R. H., Howard, H. M., Evins, P. M., Kirkland, C. L., Kelsey, D. E., Hand, M.,
 1127 Wingate, M. T. D., Collins, A. S., & Belousova, E. (2011). High-Temperature
 1128 Granite Magmatism, Crust–Mantle Interaction and the Mesoproterozoic
 1129 Intracontinental Evolution of the Musgrave Province, Central Australia.
 1130 *Journal of Petrology* 52, 931–958.
 1131 Spaggiari, C. V., Kirkland, C. L., Smithies, R. H., & Wingate, M. T. D. (2014).
 1132 Tectonic links between Proterozoic sedimentary cycles, basin formation
 1133 and magmatism in the Albany-Fraser Orogen: Geological Survey of
 1134 Western Australia, Report 133.
 1135 Stacey, J.S., & Kramers, J.D. (1975). Approximation of terrestrial lead isotope
 1136 evolution by a two-stage model. *Earth and Planetary Science Letters* 26,
 1137 207–221.
 1138 Stern, R. A., & Rainbird, R. H. (2001). Advancements in xenotime U–Pb
 1139 geochronology by ion microprobe, Eleventh V.I. Goldschmidt Conference
 1140 Lunar and Planetary Science Institute, Houston.
 1141 Stern, T. A., & Sanborn, N. (1998). Monazite U–Pb and Th–Pb geochronology by
 1142 high-resolution secondary ion mass spectrometry. *Radiogenic Age and*
 1143 *Isotope Studies: Report 11*, Geological Survey of Canada, Current
 1144 Research 1998-F, 1–18.

- 1145 VilÀ, M., Pin, C., Liesa, M., & Enrique, P. (2007). LPHT metamorphism in a late
1146 orogenic transpressional setting, Albera Massif, NE Iberia: implications
1147 for the geodynamic evolution of the Variscan Pyrenees. *Journal of*
1148 *Metamorphic Geology* 25, 321–347.
- 1149 Watson, E. B., & Harrison, T. M. (1983). Zircon saturation revisited: temperature
1150 and composition effects in a variety of crustal magma types. *Earth and*
1151 *Planetary Science Letters* 64, 295–304.
- 1152 White, R. W., Powell, R., & Johnson, T. E. (2014). The effect of Mn on mineral
1153 stability in pelites revisited: new a–x relations for manganese-bearing
1154 minerals. *Journal of Metamorphic Geology* 32, 809–828.
- 1155 Wickham, S. M., & Oxburgh, F. R. S. (1987). Low-pressure regional
1156 metamorphism in the Pyrenees and its implications for the thermal
1157 evolution of rifted continental crust. *Philosophical Transactions of the*
1158 *Royal Society of London. Series A, Mathematical and Physical Sciences*
1159 321, 219–242.
- 1160 Wingate, M. T. D., Kirkland, C. L., & Korhonen F. J. (2013). 208318: metagranite,
1161 Brown Well; *Geochronology Record* 1174: Geological Survey of Western
1162 Australia, 4p.
- 1163 Wingate, M. T. D., Kirkland, C. L., Bodorkos, S., S, Sheppard, & Farrell, T. R.
1164 (2009). 183255: metasandstone, Mount Samuel; *Geochronology Record*
1165 772: Geological Survey of Western Australia, 5p.
- 1166 Wingate, M. T. D., Lu, Y., Korhonen, F. J., & Johnson, S. P. (2019a). 216532:
1167 granodiorite gneiss, Mountain Creek; *Geochronology Record* 1559:
1168 Geological Survey of Western Australia, 4p.
- 1169 Wingate, M. T. D., Lu, Y., Korhonen, F. J., & Johnson, S. P. (2019b). 216531:

1170 muscovite monzogranite, Mountain Creek; Geochronology Record 1558:
 1171 Geological Survey of Western Australia, 4p.

1172 Wingate, M. T. D., Lu, Y., Korhonen, F. J., & Johnson, S. P. (2018). 195888:
 1173 metamonzogranite, Brown Well; Geochronology Record 1552: Geological
 1174 Survey of Western Australia, 6p.

1175 Wingate, M. T. D., Lu, Y., Korhonen, F. J., & Johnson, S. P. (2017). 208365:
 1176 biotite metagranodiorite, Kimbers well; Geochronology Record 1356:
 1177 Geological Survey of Western Australia, 4p.

1178 Zen, E. (1986). Aluminum Enrichment in Silicate Melts by Fractional
 1179 Crystallization: Some Mineralogic and Petrographic Constraints. Journal
 1180 of Petrology 27, 1095–1117.

1181

1182 **Figure and Table Captions**

1183 **Fig. 1:** Regional geological setting of the Capricorn Orogen in relation to Australia
 1184 (inset map) and simplified regional geology of the Gascoyne Province showing
 1185 discrete fault- bounded NW-SE trending domains. PC = Pilbara Craton, YC =
 1186 Yilgarn Craton, WAC = West Australian Craton, AFO = Albany- Fraser Orogen, GC
 1187 = Gawler Craton, KC = Kimberley Craton, NAC = North Australian Craton, SAC =
 1188 South Australian Craton, YGC = Yarlalweelor Gneiss Complex (adapted from
 1189 Johnson et al. (2017).

1190

1191 **Fig. 2:** Geology of the Mangaroon Zone showing location of samples used in this
 1192 study and locations of previously dated granitic samples. The Minga Bar Fault,
 1193 the Collins Fault and the Edmund Basin bound the Mangaroon Zone. Coordinate
 1194 system is GDA 1994 MGA Zone 50. Geology polygons and linear features were

1195 extracted from the 1:500 000 Geological Map of Western Australia (GSWA
1196 2016a).
1197
1198 **Fig. 3:** Field photos of granites and representative photomicrograph images of
1199 granitic sample 195890 in cross-polarised light. analysed in this study. A.
1200 Variably deformed porphyritic muscovite–biotite monzogranite characterised by
1201 discrete cm-scale mylonite zones (GSWA 195890). B. Leucocratic undeformed
1202 tourmaline–muscovite monzogranite (GSWA 219741). C. Fine-grained biotite
1203 matrix with recrystallised quartz and feldspar porphyroblasts showing sigma
1204 tails (GSWA 195890). D. A less deformed part of the rock (GSWA 195890)
1205 showing feldspar and muscovite porphyroclasts. Abbreviations: fsp = feldspar;
1206 ms = muscovite; qtz = quartz.

1207
1208 **Fig. 4:** Field photos of pelitic samples analysed in this study. A. Stromatic
1209 cordierite–sillimanite pelitic migmatite (GSWA 219749). B. Sillimanite–biotite
1210 pelitic gneiss that in outcrop appears bedded (GSWA 219742). Local melting
1211 observed at several locations across the outcrop. C. Meter-wide biotite-rich
1212 zones with leucosomes (GSWA 216533). D. Weathered outcrop of quartz–
1213 biotite–tourmaline schist (P01081). E. Sub-crop of deformed muscovite–biotite
1214 schist (P16877). F. Inclusion of deformed biotite–muscovite–tourmaline pelitic
1215 gneiss (GSWA 216538) enclosed within a relatively undeformed monzogranite
1216 (GSWA 219741).

1217
1218 **Fig. 5:** Representative photomicrograph images of analysed pelitic samples in
1219 plane-polarised light (PPL) and cross-polarised light (XPL). A. Fabric defined by

1220 fibrous sillimanite, biotite, quartz and ilmenite in the cordierite–sillimanite
 1221 pelitic migmatite (GSWA 219749; PPL). B. Sillimanite–biotite pelitic gneiss with
 1222 tabular and fibrous sillimanite (GSWA 219742; PPL). C. Possible ex-cordierite
 1223 crystal (GSWA 219742; PPL). D. Anhedral garnet associated with biotite (GSWA
 1224 219742; PPL). E. As per D but in XPL. F. Garnet porphyroblast with pressure
 1225 shadow and biotite–quartz foliation in the garnet–biotite–epidote pelitic gneiss.
 1226 Inclusions are restricted to the garnet cores and are randomly oriented (GSWA
 1227 216533; PPL). Abbreviations: bt = biotite; crd = cordierite; grt = garnet; ilm =
 1228 ilmenite; pl = plagioclase; qtz = quartz; sil = sillimanite.

1229

1230 **Fig. 6:** Back-scattered electron (BSE) images of representative monazites and
 1231 xenotime from samples analysed *in situ*. A. GSWA 195890 (muscovite–biotite
 1232 monzogranite) monazite grain 1612A. B. P01085 (biotite–muscovite
 1233 syenogranite) monazite crystal C. C. GSWA 219741 (tourmaline–muscovite
 1234 monzogranite) monazite crystal C. D. GSWA 219749 (cordierite–sillimanite
 1235 pelitic migmatite) crystals H. E. GSWA 219742 (sillimanite–biotite pelitic gneiss)
 1236 monazite crystal D. F. GSWA 216533 (garnet–biotite–epidote pelitic gneiss),
 1237 garnet crystal. G. GSWA 216533 monazite inclusion within garnet crystal, 1503D,
 1238 showing an alteration pathway to the left. H. GSWA 216533 example of altered
 1239 monazite surrounded by apatite corona and bastnaesite (bas) rim. I. P01081
 1240 (quartz–muscovite–tourmaline schist) monazite crystal G. J. P16877 (quartz–
 1241 biotite–muscovite schist) monazite crystal A. J. GSWA 216538 (biotite–
 1242 muscovite–tourmaline pelitic gneiss) monazite crystal D. K. GSWA 216538
 1243 xenotime crystal A. Red circles denote SHRIMP spot locations. Ages are indicated

in black with errors indicated at 95% confidence level. Analytical data are presented in Supplementary Table 1.

Fig. 7: U–Pb concordia diagrams for monazite from granitic samples A–C. Data plotted with 68.3% error ellipses; red ellipses indicate analyses used to calculate the age of the sample, and unfilled ellipses indicate data excluded due to discordance >5%, high common Pb, or as statistical outliers. Weighted mean dates are based on $^{207}\text{Pb}/^{206}\text{Pb}$ ratios and quoted with 95% confidence intervals.

Fig. 8: U–Pb concordia diagrams for monazite (A–F) and xenotime (H) from pelitic samples. Data plotted with 68.3% error ellipses; red ellipses indicate analyses used to calculate the age of the sample; unfilled ellipses indicate data excluded due to discordance >5% or as statistical outliers. Weighted mean dates are based on $^{207}\text{Pb}/^{206}\text{Pb}$ ratios and quoted with 95% confidence intervals.

Fig. 9: *P–T* pseudosections for (a) GSWA 219749 and (b) GSWA 219742. Peak metamorphism field is outlined by thick lines; solidus shown by thick dashed lines; H₂O field shown in (b) by thick dotted line. Abbreviations used on pseudosections: and, andalusite; bi, biotite; cd, cordierite; g, garnet; H₂O, H₂O; ilm, ilmenite; ksp, K-feldspar; ky, kyanite; mt, magnetite; liq, melt; mu, muscovite; pl, plagioclase; q, quartz; sill, sillimanite.

Fig. 10: Compilation diagram of Mangaroon-age geochronology from the Mangaroon Zone. Includes results from this study and published data (GSWA 2016b). The data show that magmatism was initiated at 1695 Ma followed by

1269 upper crustal metamorphism at 1691 Ma. Magmatism continued for an
1270 additional 20 million years after peak metamorphism. $^{40}\text{Ar}/^{39}\text{Ar}$ mica ages of c.
1271 1640 Ma are interpreted to indicate the timing of cooling and exhumation of the
1272 Mangaroon Zone crust. Sample numbers in bold indicate this study. Asterisks
1273 indicates samples used for P - T modelling. Blue dashed lines indicate the
1274 geochronology samples used to constrain magmatism associated with D_1
1275 deformation.

1276

1277 **Fig. 11:** Summary diagram showing P - T pseudosections from GSWA 219742 and
1278 219749 plotted in P - T space. Sample (b) indicates a higher prograde
1279 temperature than sample (a).

1280

1281 **Fig 12:** The Mangaroon orogeny high-grade core, low-grade halo in the West
1282 Australian Craton. The red oval represents the high-grade 1691–1680 Ma
1283 metamorphism recorded in the Mangaroon Zone. The stars represent low-grade
1284 occurrences of 1680–1650 Ma ages from outside of the Gascoyne Province. The
1285 grey dashed lines represent the northern and southern limits of the Capricorn
1286 Orogen. These data show that the footprint of the Mangaroon Orogeny extents
1287 far beyond the high-grade core.

1288

1289 **Table 1** Summary of pelitic and granitic samples from the Mangaroon Zone
1290 analysed in this study.

1291

1292 **Table 2:** Whole-rock geochemistry data for Mangaroon Zone Durlacher
1293 Supersuite granite samples, from this study and published data (GSWA 2016b).

1294

1295 **Table 3:** Whole-rock chemistry used to construct P – T pseudosections for

1296 samples GSWA 219742 and 219749.

1297

1298 Supplementary data:

1299

1300 **Supplementary Fig. 1:** Plot of ASI values for Durlacher granites from the

1301 Mangaroon Zone. * A/CNK = molar $Al_2O_3/(CaO+Na_2O+K_2O)$.

1302

1303 **Supplementary Fig. 2:** Calculated zircon saturation temperatures ($^{\circ}C$) (M

1304 parameter from Watson and Harrison 1993) plotted against igneous

1305 crystallisation ages in Ma show that Durlacher Supersuite granites were typically

1306 $>750^{\circ}C$. Geochemistry data is available from Table 3 and/or the Geological

1307 Survey of Western Australia WACHEM database

1308 (<http://geochem.dmp.wa.gov.au/geochem/>).

1309

1310 **Supplementary Fig. 3:** Probability density diagrams and histograms of (a)

1311 xenocrystic zircons from the Durlacher Supersuite and (b) detrital zircon from

1312 the Pooranoo Metamorphics, from the Mangaroon Zone. Includes data ($<5\%$

1313 discordant) from published data (GSWA 2016b).

1314

1315 **Supplementary Table:** U–Pb SHRIMP monazite and xenotime data tables.

1316

GSWA Sample ^a	Easting	Northing	Sample description	Structure	<i>P–T</i> constraints	Mineral	U–Pb Age (Ma)
<i>Granite samples^b</i>							
195890	338457	7397374	ms–bt metamonzogranite	230°/90° metamorphic foliation		monazite	1695 ± 9
P01085 ^c	436219	7343766	bt–ms syenogranite			monazite	1682 ± 3
219741	339577	7382285	tur–ms monzogranite			monazite	1659 ± 13
<i>Supporting samples^d</i>							
208365	326186	7440703	bt metagranodiorite	228°/70° SE metamorphic foliation; D ₁		zircon	1689 ± 6
195888	331433	7390220	bt monzogranite			zircon	1689 ± 4
216532	331336	7424614	bt metagranodiorite	260/85° SE gneissic layering; D ₁		zircon	1688 ± 4
195887	341014	7379817	tur–ms monzogranite			zircon	1682 ± 9
216531	331336	7424614	ms–tur monzogranite			zircon	1673 ± 4
<i>Pelitic samples^e</i>							
219749	372340	7359950	cd–sill pelitic migmatite	147°/85° SE fold axial surface; 230°/70° SE fold axis; D ₁	665–755°C; 2.7–4.3 kbar	monazite	1691 ± 7
						675–850°C; 4–7.5 kbar	
219742	353086	7402982	sill–bt pelitic gneiss	241°/30° SE metamorphic foliation; D ₁		monazite	1685 ± 9
216533	330613	7424503	gt–bt–ep pelitic gneiss			monazite	1683 ± 6
P01081 ^c	419044	7355430	qtz–ms–tour schist			monazite	1681 ± 9
P16877 ^c	422009	7355110	qtz–bt–ms schist			monazite	1680 ± 7
216538	339577	7382285	bt–ms–tur pelitic gneiss			monazite	1680 ± 4
216538	339577	7382285	bt–ms–tur pelitic gneiss			xenotime	1680 ± 4

Sample locations quoted in coordinate system MGA 94 Zone 50.

Planar structures are quoted as strike/dip and dip direction.

^a Sample prefix is GSWA unless otherwise stated.

^b Durlacher Supersuite.

^c Samples collected by J. Pearson (Pearson 1996).

^d Published GSWA data. References within text.

^e Pooranoo Metamorphics.

Sample	195890 ^b	195888 ^c	208365 ^c	216532 ^c	88410 ^c	195887 ^c	169092 ^c	178030
	<i>Ms–bt</i> <i>metamonzogranite</i>	<i>Bt porphyritic</i> <i>monzogranite</i>	<i>Bt metagranodiorite</i>	<i>Bt porphyritic</i> <i>monzogranite</i>	<i>Ms–bt</i> <i>monnzograni</i> <i>te</i>	<i>Tur</i> <i>monzogranite</i>	<i>Bt–ms</i> <i>monzogranite</i>	<i>Bt</i> <i>granodiorite</i>
Rock Type								
wt%								
SiO ₂	-	67.40	65.09	67.43	73.99	72.90	73.85	62.93
TiO ₂	-	0.58	0.64	0.51	0.14	0.19	0.19	0.81
Al ₂ O ₃	-	16.00	15.26	15.16	13.57	14.85	13.86	16.00
Fe ₂ O ₃ ^T	-	5.72	5.29	4.67	1.60	1.69	1.593	7.232
FeO	-	5.14	3.08	4.20	1.17	1.52	1.08	3.8
MgO	-	1.79	1.74	1.36	0.33	0.39	0.39	2.12
MnO	-	0.07	0.11	0.11	0.06	0.05	0.06	0.17
CaO	-	1.36	3.01	2.31	1.00	0.91	0.81	3.39
Na ₂ O	-	1.79	2.60	2.76	2.60	2.98	2.86	2.35
K ₂ O	-	4.19	3.64	3.97	5.46	5.41	5.21	3.91
P ₂ O ₅	-	0.13	0.22	0.20	0.10	0.20	0.24	0.24
LOI	-	1.50	1.87	1.02	1.03	0.89	0.85	1.03
A/CNK ^a		1.61	1.12	1.16	1.13	1.20	1.17	1.12
ppm								
Zr	-	159.0	181.0	161.0	93.0	95.0	70.0	191.0
Age (Ma)	1695 ± 9 [^]	1689 ± 4 [*]	1689 ± 6 [*]	1688 ± 4 [*]	1688 ± 7 [*]	1682 ± 9 [*]	1681 ± 10 [*]	1678 ± 6 [*]

Sample	178027 ^c	169053 ^c	178029 ^c	178028 ^c	169062 ^c	216531 ^c	169060 ^c	169059 ^c	169055 ^c	219741 ^b
	<i>Bt-ms granodiorite</i>	<i>Bt-ms monzogranite</i>	<i>Bt monzogranite</i>	<i>Bt-ms syenogranite dyke</i>	<i>Porphyritic syenogranite</i>	<i>Ms-tur monzogranite</i>	<i>Porphyritic syenogranite</i>	<i>Ms-bt monzogranite</i>	<i>Bt-ms monzogranite</i>	<i>Tur-ms monzogranite</i>
Rock type										
wt%										
SiO ₂	65.38	68.44	70.50	70.89	70.08	72.92	72.18	71.85	72.15	-
TiO ₂	0.69	0.46	0.25	0.38	0.49	0.19	0.37	0.32	0.30	-
Al ₂ O ₃	15.99	15.00	14.44	14.31	14.71	14.29	14.18	14.69	14.13	-
Fe ₂ O ₃ ^T	5.12	4.97	2.123	2.728	2.149	1.67	2.80	1.82	2.15	-
FeO	2.79	2.25	1.43	1.72	1.13	1.50	1.76	0.80	1.26	-
MgO	1.21	1.39	0.48	0.75	0.61	0.39	0.68	0.40	0.41	-
MnO	0.10	0.09	0.07	0.07	0.05	0.05	0.06	0.04	0.06	-
CaO	2.70	1.06	0.80	0.79	1.34	0.87	1.92	0.96	0.94	-
Na ₂ O	2.89	1.84	2.45	1.92	2.35	3.10	3.19	2.55	2.47	-
K ₂ O	3.67	4.36	6.23	6.73	6.25	5.26	3.66	6.08	5.66	-
P ₂ O ₅	0.11	0.14	0.21	0.25	0.30	0.09	0.13	0.26	0.21	-
LOI	2.30	2.36	2.49	1.20	1.02	0.71	0.91	1.00	1.53	-
A/CNK ^a	1.17	1.55	1.18	1.20	1.13	1.15	1.12	1.17	1.19	-
ppm										
Zr	112.0	131.0	147.0	222.0	316.0	126.0	167.0	174.0	140.0	-
Age (Ma)	1677 ± 5 [*]	1677 ± 6 ^{#*}	1675 ± 11 [*]	1674 ± 8 [*]	1674 ± 8 [*]	1673 ± 4 [*]	1673 ± 15 [*]	1660 ± 9 [*]	1659 ± 10 [*]	1659 ± 13 [^]

^a A/CNK = molar Al₂O₃/(CaO+Na₂O+K₂O); ^b this study; ^c = published data (GSWA 2016b); - = not analysed for whole-rock geochemistry; * = zircon age; ^ = monazite age; # age recalculated from Nelson 2002;

Ms = muscovite; bt = biotite, tur = tourmaline;

Sample	GSWA 219742	GSWA 219749
<i>XRF whole rock compositions (wt%)</i>		
SiO ₂	57.24	58.56
TiO ₂	0.78	0.83
Al ₂ O ₃	21.37	21.81
Fe ₂ O ₃ ^(a)	4.78	1.93
FeO ^(a)	4.78	6.24
MnO	0.23	0.17
MgO	2.72	2.93
CaO	0.61	0.34
Na ₂ O	0.92	0.62
K ₂ O	3.7	4.48
LOI	1.85	0.89
Total	98.98	98.80
<i>Normalised molar proportions used for phase equilibria modelling^(b)</i>		
SiO ₂	63.22	65.59
TiO ₂	0.65	0.70
Al ₂ O ₃	13.91	14.39
O	1.99	0.81
FeO	4.42	5.84
MnO	0.22	0.16
MgO	4.48	4.89
CaO	0.72	0.41
Na ₂ O	0.99	0.67
K ₂ O	2.61	3.20
H ₂ O	6.81	3.32
Total	100.00	100.00

^(a) FeO analysed by Fe²⁺ titration; Fe₂O₃ calculated by difference

^(b) final compositions used for P-T pseudosection, based on adjusted FeO:Fe₂O₃