Geochronological constraints on nickel metallogeny in the Lake Johnston belt, Southern Cross Domain

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Geochronology and stratigraphic revision of the Lake Johnston greenstone belt and adjacent granitoids and granitic gneiss provide new insight into the age of komatiites in the Southern Cross Domain of the Archean Yilgarn Craton. Roundtop Komatiites are geochemically similar to undated komatiites in the adjacent Ravensthorpe and Southern Cross—Forrestania greenstone belts, and the results can be extrapolated to improve the regional understanding and geodynamic evolution. Consequently, the further refined knowledge of the regional stratigraphy improves the understanding of the evolution and targeting of komatiite-hosted nickel deposits. A minimum age of ca 2773 Ma for the succession of the Lake Johnston greenstone belt is provided by crosscutting granitic rocks, with a maximum age for the underlying Roundtop Komatiite given by a maximum depositional age of ca 2876 Ma for felsic volcanioclastic rocks of the underlying Honman Formation. These new results suggest that komatiites of the Southern Cross Domain are significantly younger than previously assumed, which has implications for Yilgarn-wide geodynamic models regarding ‘plume activity’ and global correlations in the Meso- to Neoarchean.

KEY WORDS: Archean, greenstones, stratigraphy, komatiite, zircon dating, SHRIMP.

INTRODUCTION

The Yilgarn Craton of Western Australia consists mainly of ca 3000–2620 Ma, north- to northwest-trending, deformed greenstone belts, separated by >2620 Ma granitoid and granitic gneiss terranes (Gee et al. 1981; Pidgeon & Wilde 1990; Schiøtte & Campbell 1996; Wang et al. 1996; Pidgeon & Hallberg 2000; Kosticin et al. 2008; Van Kranendonk & Ivanic 2010). The Yilgarn Craton contains world-class nickel deposits (Hoatson et al. 2006) that are associated with major komatiite-hosted nickel-metallogenic events at ca 2.9 and 2.7 Ga (Barnes 2006a, b; and references therein). A major mafic magmatic event occurred at ca 2.8 Ga (Kosticin et al. 2008; Ivanic et al. 2010; Riganti et al. 2010). These magmatic episodes have been suggested to represent mantle-overturn events coeval with plume activity (Campbell & Hill 1988; Barley et al. 1998; Rey et al. 2003; Pirajno 2004; Groves et al. 2005).

The Lake Johnston greenstone belt (Figure 1) is situated in the Youanmi Terrane between the Southern Cross–Forrestania greenstone belts to the west and the Norseman–Wiluna greenstone belt to the east. The geology of the Lake Johnston greenstone belt was first described by Honman (1914) and later mapped by the Geological Survey of Western Australia (Gower & Bunting 1976). Since then, several university and industry-funded research projects have focused on understanding the nickel deposits and crustal architecture of the belt (e.g. Perring et al. 1994; Buck et al. 1998; Heggie 2010; Joly et al. 2010; Heggie et al. 2012a, b).

The Lake Johnston and Southern Cross–Forrestania greenstone belts host significant nickel and gold deposits (Perring et al. 1994, 1995; Barnes et al. 1995; Barnes 2006b). Based on SHRIMP U–Pb zircon geochronology and on stratigraphic correlations with the adjacent Murchison Domain, parts of the greenstone successions have been interpreted to be older than ca 2.9 Ga (Wang et al. 1996; Mueller & McNaughton 2000). Consequently, the komatiites and associated nickel mineralisation in the Lake Johnston and Forrestania greenstone belts have been regarded as older than ca 2.9 Ga, much older than the ca 2.7 to 2.69 Ga Kambalda komatiites in the Eastern Goldfields Superterrane (Swager & Griffin 1990; Swager 1997). However, new field mapping, geochemistry and geochronology challenge this interpretation, and require re-assessment of the age and stratigraphy of the Lake Johnston and Southern Cross–Forrestania greenstone belts.
THE LAKE JOHNSTON GREENSTONE BELT

The Lake Johnston greenstone belt trends northwesterly for about 100 km and is intruded by elongate granitoids. The belt’s present-day maximum width of ~20 km reflects repetition of the stratigraphy caused by folding and shearing. Peak metamorphic conditions have been identified as 596–678°C and 5–7 kbar (Joly et al. 2010). The eastern side of the greenstone belt is represented by a zone, up to 4 km wide, that is strongly deformed and transposed along the major Koolyanobbing Shear Zone (Figure 2; Libby et al. 1991; Libby 1992; Stewart 1992). An older age limit for deformation in the Koolyanobbing Shear Zone is provided by a U–Pb zircon crystallisation age of ca. 2699 Ma for a strongly deformed granodiorite, interpreted as a xenolith in younger granitoids (Fletcher & McNaughton 2002). A younger limit is indicated by the age of ca. 2656 Ma for the post-kinematic Lake Seabrook granite (Qiu et al. 1999). Other major structural features of the belt include the northwest-trending Tay fault and
the Burmiester syncline on the western side of the belt, and the Gordon anticline, which exposes the basal stratigraphy in the central part of the belt (Figure 2; Gower & Bunting 1976). The greenstone succession has historically been divided (from base to top) into the Maggie Hays, Honman and Glasse formations (Figure 3a; Gower & Bunting 1976). More recently, ultramafic units within the Lake Johnston belt stratigraphy have been further divided into the eastern, central and western ultramafic units (Perring 1995; Buck et al. 1998; Heggie 2010, Heggie et al. 2012a, b).

Recent mapping by the Geological Survey of Western Australia (Figure 4), together with complementary geochemical and geochronological analyses, indicates a more complex stratigraphy, with the intrusive Lake Medcalf Igneous Complex occurring between the Maggie Hays and Honman formations, as well as parts of the Roundtop Komatiite intruding the Honman Formation (Figure 3b). Based on the results of this study, we propose a refined lithostratigraphy, showing the relative times of deposition and intrusion (Figure 3c). The stratigraphic sequence includes (from oldest to youngest): the Maggie Hays Formation, Lake Medcalf Igneous Complex, Honman Formation, Roundtop Komatiite and Glasse Formation. The metamorphic grade in the Lake Johnston belt ranges from greenschist to upper amphibolite facies.

**Maggie Hays Formation**

The Maggie Hays Formation, the lowermost unit exposed in the Lake Johnston greenstone belt, is in sheared contact with internal granitic rocks (i.e. granitoids within the greenstone belt), close to Maggie Hays Hill (Figure 4). The original thickness of the Maggies Hays Formation is difficult to estimate owing to shearing and possible stratigraphic repetition. We estimate the stratigraphic thickness to be ~1600 m, less than the ~2400 m originally estimated by Gower & Bunting (1976).

The Maggie Hays Formation is a variably overprinted submarine volcanic succession. The lowest rocks exposed are amphibolites (after tholeiitic basalt) with preserved pillow structures containing amygdales (Figure 5a). Owing to sporadic outcrop and variable intensity of deformation assessing younging direction is rarely possible. However, weakly deformed pillow-structures in the western limb of the greenstone belt consistently indicate younging to the west.

Fine-grained, massive amphibolites containing variably deformed and recrystallised feldspar porphyroblasts (relict phenocrysts) are commonly associated with the metamorphosed pillow basalts (Figure 5a). Owing to sporadic outcrop and variable intensity of deformation assessing younging direction is rarely possible. However, weakly deformed pillow-structures in the western limb of the greenstone belt consistently indicate younging to the west.

Fine-grained, massive amphibolites containing variably deformed and recrystallised feldspar porphyroblasts (relict phenocrysts) are commonly associated with the metamorphosed pillow basalts (Figure 5b). Above this plagioclase-phryic amphibolite, a package of fine-grained amphibolite is overlain by finely layered, medium-grained amphibolite. These amphibolites are intercalated with pelites and minor metavolcaniclastic horizons (Figure 5c). A prominent horizon of metamorphosed pillow breccias with a hyaloclastite matrix (Figure 5d) is exposed east of Maggie Hays Hill and north of Mount Day. The fragments are heterogeneous and up to 10 cm in diameter. Some are flattened, angular and generally aligned in the finer-grained matrix composed of feldspar, biotite and altered glass. Fine-grained mafic clasts are typically altered. In the upper part of the Maggie Hays Formation, metabasalts are intercalated with thin, metamorphosed iron- and silica-rich interflow
sedimentary rocks, which become more abundant towards the top of the formation.

Lake Medcalf Igneous Complex

A newly conducted gravity survey of the Lake Johnston area indicates that mafic-ultramafic intrusive rocks of the Lake Medcalf Igneous Complex are distributed more extensively than previously realised (Figure 2b). The Complex forms concordant, sill-like intrusions into the upper levels of the Maggie Hays Formation. The protoliths either show a gradual fractionation from medium- to coarse-grained metapyroxenite to metaleucogabbro and meta-anorthosite, or form medium- to coarse-grained metagabbro. Chilled margins mark the contact with the Maggie Hays Formation (Figure 4). In the central part of the western limb, fractionated metagabbros show younging to the west, consistent with the above-mentioned pillow structure in metabasalts.

The intrusive rocks are typically highly strained (Figure 5e), form mega-boudins within the Maggie Hays Formation, and are aligned parallel to the trend of the greenstone belt. Less-deformed portions of the Lake Medcalf Igneous Complex metagabbros preserve a primary porphyritic texture. Metamorphic conditions reached upper amphibolite facies and resulted in development of coarse-grained amphibolite after gabbro, and actinolite-chlorite schist after pyroxenite. In thin-section, the rocks show evidence of dynamic recrystallisation of plagioclase and growth of green metamorphic amphibole.

Honman Formation

The Honman Formation (Gower & Bunting 1976) is mainly exposed on the western side of the greenstone belt with only few remnants in the east (Figure 4). A detailed, mine-scale description was recently published by Heggie et al. (2012a, b). The succession grades from pale grey, metadacitic to metarhyolitic, metavolcanic and volcaniclastic rocks to clastic metasedimentary rocks. The uppermost package consists of quartzite, metachert, and meta-BIF which form Honman Ridge and most of Roundtop Hill. Heggie et al. (2012a) described a ‘Transition Zone’ above the felsic volcanic rocks including garnetite, amphibolite, metarhyolite and clastic metasedimentary rocks. These are overlain by the meta-BIF. The top of the Honman Formation is formed by garnetite and quartz-arenite. The fine-grained felsic metavolcanic rocks are poorly exposed and the lower contact of the Honman Formation is typically obscured by deformed pegmatite veins. At Roundtop Hill, the metavolcanic rocks are tuffaceous (Figure 5f) and conformably overlain by folded meta-BIF (Figure 5g). Additionally, granitic gneiss disrupts the stratigraphy at the bottom of Roundtop Hill. The mine-scale study of Joly et al. (2010) described the occurrence of garnet-bearing pegmatite in a similar stratigraphic position.

Southwest of Honman Ridge, only the upper part of the Honman Formation, consisting of recumbently folded quartzite overlain by metachert with finely laminated meta-BIF at the top, is exposed. At the lowest exposed level, the metamorphic overprint has caused mobilisation of quartz in the quartzite producing in situ formation of quartz veins.

Roundtop Komatiite

Previous studies by Buck et al. (1998), Fiorentini et al. (2011) and Heggie et al. (2012a, b) identified three major metakomatiite units within the newly defined Roundtop Komatiite. These are interpreted to rest conformably on the Honman Formation and intrude the Maggie Hays and lower levels of the Honman formations.

The western ultramafic unit (Buck et al. 1998) is up to 400 m thick and consists of a series of fractionated, thin overprinted komatiite showing well-preserved, green-grey olivine-spinifex-textured flow tops and yellow-brown olivine orthocumulate to mesocumulate flow.

|                   | Honman Formation | Glasse Formation | Lake Medcalf Igneous Complex |
|-------------------|------------------|------------------|------------------------------|
| 2903 Ma           | komatite         | basalt           | gabbro                       |
| 2921 Ma           | komatite         | basalt           | gabbro                       |

Figure 3 Stratigraphic succession of the Lake Johnston greenstone belt: (a) previous stratigraphic interpretation (Gower & Bunting 1976; Wang et al. 1996); (b) revised tectonostratigraphy, representing an E-W profile, based on younging-to-W indicators; and (c) revised stratigraphy. ¹Wang et al. 1996; ²this study.
bases. Within costeans southeast of the Maggie Hays deposit, the size of olivine plates from the western ultramafic unit increases westward, indicating younging to the west. The metamorphic and alteration products are talc–carbonate schists, which weather to a distinct pale-yellow soil. The metamorphic assemblage in spinifex-textured komatiites is tremolite–serpentinite–chlorite and opaque phases. Southeast of the Maggie Hays deposit, a small exposure of talc–chlorite schist exhibits kink bands.

The eastern ultramafic unit, interpreted to be located within the upper Maggie Hays Formation, forms a discontinuous series of lenses <80 m thick. Sparse outcrop and drill chips suggest a series of thin, differentiated metakomatiite flows with a range of magnesium contents (Buck et al. 1998). Heggie et al. (2012a) describe the unit as an igneous body. However, recent mapping shows that this unit represents a structural duplication (e.g. by folding) of the western ultramafic unit, associated with slivers of the Honman Formation along strike. Nevertheless, presently mineralisation has not been identified within the eastern ultramafic unit.

Undifferentiated metakomatiites of the central ultramafic unit (Heggie 2010) occur as stratigraphically conformable sills below the meta-BIF unit of the Honman Formation in the central part of the Lake Johnston belt. The sequence hosts the Maggie Hays nickel deposit (Buck et al. 1998). Detailed research by Perring et al. (1994) and Perring (1995) shows that olivine mesocumulates and acumulates fractionate upwards into
pyroxene ± olivine cumulates and pyroxene-plagioclase cumulates. Heggie et al. (2012b) describes the ore as a subhorizontal tube-like body up to 300-400 m thick and >40 m wide.

**Glasse Formation**

The Glasse Formation is the uppermost formation in the Lake Johnston belt (Gower & Bunting 1976). The formation is up to 1200 m thick and consists of deformed and massive, fine-grained green-grey metabasalt. The formation contains rare horizons with amygdales up to 0.5 cm across and relict feldspar phenocrysts. The weathering style of the Glasse Formation is blockier and more angular than the mafic extrusive rocks of the Maggie Hays Formation. In thin-section, the fine-grained metabasalts consist of a matrix of small, dynamically recrystallised plagioclase, aligned green actinolite and epidote, and larger, spotted, twinned green amphibole grains. The rock is crosscut by fine, deformed quartz veins with subgrain rotation of quartz, which is a typical quartz deformation feature at temperatures of 400-500°C (e.g. Stipp et al. 2002).

Ultramafic rocks previously described as part of the Glasse Formation by Gower & Bunting (1976) are located to the east and below the mafic extrusive rocks of the Glasse Formation, and are separated from them by a thin zone of metasedimentary rocks. Thus, they may be
part of the Roundtop Komatiite associated with the Honman Formation rather than an intrusiv unit of the Glasse Formation.

**Isolated greenstone belt remnants**

Greenstone belt remnants within migmatisites, and granitic gneisses between the Forrestania and Lake Johnston greenstone belts, contain strongly deformed metasedimentary rocks, including meta-BIF, amphibolite and undifferentiated metakomatiite. The greenstone remnants primarily preserve metasedimentary rocks of the Honman Formation, minor amphibolites of the Maggie Hays Formation, and the metamorphosed mafic–ultramafic intrusive rocks of the Lake Medcalf Igneous Complex. Gower & Bunting (1976) described the occurrence of pumpellyite in the meta-BIF and, therefore interpreted a prehnite–pumpellyite facies metamorphic overprint. However, recent studies have shown that pumpellyite can be stable to up to 1050°C (Liu 1989; Fockenberg 1998). Moreover, the occurrence of amphibolite indicates that the greenstone remnants underwent at least amphibolite facies conditions.

**GEOCHRONOLOGY**

A major difficulty in lithostratigraphic correlation of Archean greenstone belts composed of mainly mafic successions is the rarity of dateable minerals. Zircon and baddeleyite are the minerals of choice for U–Pb dating of mafic rocks. Zircon is relatively rare in mafic rocks and is commonly xenocrystic, whereas baddeleyite (ZrO₂), although rarely occurring as xenocrysts, breaks down to form zircon during metamorphism (Davidson & van Bree 1998; Heaman & LeCheminant 1993). Typically, the best way to determine the age of a mafic-ultramafic succession is by its cross-cutting relations with dateable felsic or intermediate rocks. Until now, the age of the Lake Johnston greenstone belt has been poorly constrained, and some of the existing zircon U–Pb geochronological data (e.g. Wang et al. 1996) are difficult to interpret.

**Methodology**

Appropriate samples were selected in the field using a NITON field-portable X-ray fluorescence (XRF) analyser to ensure sufficient zirconium content. After removing weathered surfaces and inclusions, each 2-3 kg sample was processed at Minsep Laboratories in Denmark, Western Australia (samples LJD-0017-222, LJD-0011-770, ARC 23), or at the Geological Survey of Western Australia Carlisle Laboratory (samples GSWA 182308 and 199046). Each sample was crushed using a jaw crusher, powdered using a ring mill, and sieved. Zircons were then concentrated using conventional density and magnetic methods, and isolated by hand-picking. Zircons were mounted in 25 mm diameter epoxy-resin disks with chips U/Pb zircon standard BR266 (559 Ma, 903 ppm U; Stern 2001), crystals of zircon 238U/206Pb standard OGC1 (3465 Ma; OGI of Stern et al. 2009), and NBS610 glass. Cathodoluminescence (CL) and backscattered electron (BSE) imaging on gold-coated mounts were performed using a Philips XL30 scanning electron microscope (SEM) at the Department of Imaging and Applied Physics at Curtin University in Perth, Western Australia, and a JEOL 6400 SEM at the Centre for Microscopy and Microanalysis at the University of Western Australia.

U–Pb analyses were performed using the SHRIMP II ion microprobes at Curtin University, Western Australia, using standard operating procedures similar to those described by Compston et al. (1992) and Wingate & Kirkland (2001). The diameter of the primary ion probe used during all sessions was about 20 μm. Six data collection cycles (scans) were performed per analysis, and count times (per scan) were 10 s for the 204Pb, 206Pb, and 208Pb mass peaks and background, and 30 s for the 238U mass peak. Analyses of unknown zircons were referenced to multiple analyses of the BR266 standard for U/Pb calibration. Calibration and reproducibility uncertainties are included in the errors of 238U/206Pb ratios and dates listed in supplementary papers. Data reduction was carried out using the software SQUID v2.2 and 2.5 and ISOPLOT v3.0 (MS Excel add-ins by Ludwig 2003, 2009). Corrections for common Pb were based on measured 206Pb, assuming an average crustal composition (Stacey & Kramers 1975) appropriate to the age of the mineral. Mean ages are quoted below with 95% uncertainties (τMSWD; Ludwig 2003). Data are listed in Supplementary Papers (Table A1) and illustrated in concordia diagrams.

**GREENSTONE SUCCESSION**

Two drill core samples, LJD-0017-222 and LJD-0011-770, were collected at the Maggie Hays deposit from a meta-rhyolite and a felsic metavolcaniclastic rock located at the base and top of the Honman Formation, respectively. Additionally a metadacite sample (GSWA 199046) was collected at Roundtop Hill.

**Table 1 Major element oxide geochemistry measured by XRF**

| Sample ID | SiO₂ | Al₂O₃ | Fe₂O₃ (tot) | FeO | MgO | CaO | Na₂O | K₂O | TiO₂ | P₂O₅ | SO₃ | Cr₂O₃ | V₂O₅ | LOI | Total |
|-----------|------|------|------------|-----|-----|-----|------|-----|------|------|-----|-------|------|-----|-------|
| LJD00011-770 | 63.14 | 9.72 | 15.39 | 14.03 | 1.71 | 2.51 | 0.37 | 3.02 | 0.25 | 0.03 | 3.67 | 0.01 | 0.004 | 2.4 | 103.7 |
| LJD0017-222 | 68.54 | 18.12 | 0.78 | 0.88 | 1.57 | 8.43 | 1.29 | 0.07 | 0.25 | 0.09 | 0.006 | 0.7 | 100.2 |

*Fe is presented in four forms: new data from this study analysed Fe as Fe₂O₃ (tot), which was converted to FeO via the canonical method.

*Uncorrected analytical totals.
Sample LJD-0017-222, Metarhyolite (Lat: -32.25120; Long: 120.5146)

This sample is a fine-grained (0.2–0.5 mm), equigranular rhyolite containing subhedral platy biotite (1–2 vol%), anhedral chlorite (2–5 vol%), and anhedral feldspar (15–20 vol%), within an anhedral–granular matrix of quartz. Zircon extracted from this sample are subhedral to euhedral, equant to slightly elongate, typically 150–100 μm, and colourless to dark brown. The CL images of sub-facetted and elongated grains show texturally discordant cores with broad sector zoning or cloudy textures overgrown by concentric oscillatory-zoned rims (Figure 6a).

Twenty-four analyses were obtained from 24 zircons (Figure 7a). Five analyses >5% discordant, and one analysis >4% reversely discordant, were not included in the age calculation. The remaining 18 analyses are concordant to slightly discordant and yield a weighted mean 207Pb/206Pb date of 2920 ± 4 Ma (MSWD = 2.9). The somewhat elevated MSWD value may indicate minor degrees of ancient radiogenic-Pb loss that affect the result. However, the exclusion of data points furthest from the mean does not significantly change the age or uncertainty.

Sample GSWA 199046, Metadacite (Lat: -32.1642; Long: 120.4486)

The sample is a foliated and folded, fine- to medium-grained metadacite, with partly recrystallised relict feldspar phenocrysts (Figure 5f). The rocks are overlain by a thick package of bleached meta-BIF (Figure 5g). Zircon from this sample are colourless to dark brown and subhedral to euhedral. In CL images, concentric zoning is ubiquitous, and some crystals appear to contain older cores (Figure 6c). Three analyses are >5% discordant, and two analyses indicate high within-run variation of isotope ratios; these five analyses are not considered further. Twelve analyses yield a concordia age of 2922 ± 4 Ma (MSWD = 1.3), interpreted as the magmatic crystallisation age. One analysis of a zircon core yields a 207Pb/206Pb date of 2943 ± 8 Ma (1σ), interpreted as the age of an inherited component (Figure 8).

Sample LJD-0011-770, Volcaniclastic Metasedimentary Rock (Lat: -32.2380; Long: 120.5077)

This sample is a fine-grained (0.2–0.5 mm), foliated, equigranular metavolcaniclastic tuffaceous sandstone containing subhedral, platy biotite (5–10 vol%) and anhedral plagioclase (2–5 vol%) within an anhedral–granular matrix of quartz. Zircon from this sample are anhedral to subhedral, equant to slightly elongate, typically 100–80 μm, and colourless. CL images show rounded crystals with broad concentric oscillatory and/or sector zoning. Some zircons contain mineral inclusions (Figure 6b).

Eighteen analyses were obtained from 18 zircons (Figure 7b). All analyses are <6% discordant. Seventeen analyses yield a weighted mean 207Pb/206Pb date of 2876 ± 3 Ma (MSWD = 1.7). The single excluded analysis is significantly younger than the mean, and yields a 207Pb/206Pb date of 2856 ± 4 Ma (1σ).
GRANITIC ROCKS

Sample GSWA 182308, Foliated Metamonzogranite (Lat: –32.3730; Long: 120.7521)
This sample is a pale-grey to pinkish, foliated, medium-grained metamonzogranite that intruded in the central part of the greenstone belt. Zircons from the sample are subhedral to euhedral, equant to slightly elongate, and colourless to dark brown. Oscillatory zoning is common, and many crystals contain areas or zones that are metamict. The CL image shows concentric oscillatory zoned short prismatic to sub-facetted grains, partly containing discordant cores. Some grains developed an unzoned rim (Figure 6d).

Twenty-three analyses were obtained from 23 zircons (Figure 9a). All data, which include 19 results between 5 and 90% discordant, define a discordia with an upper intercept date of $2720 \pm 9$ Ma (MSWD = 2.9) and a zero-age lower intercept. The four least discordant analyses ($<5\%$ discordant) yield a weighted mean $^{207}$Pb/$^{206}$Pb date of $2718 \pm 6$ Ma (MSWD = 0.23).

Sample ARC 23, Equigranular Metamonzogranite–Quartz Monzonite (Lat: –32.2892; Long: 120.5322)
This sample is a pinkish equigranular medium grained monzogranite to quartz monzonite. The larger granitoid platform consists of equigranular monzogranite to quartz monzonite and flow-foliated monzogranite with

**Figure 7** U–Pb analytical data for zircons from supracrustal rocks of the Honman Formation: (a) metarhyolite (LJD-0017-222), and (b) metavolcaniclastic rock (LJD-0011-770). Crossed squares indicate data not included in the age calculation (see text). Error bars are 1 sigma. Mean ages are quoted with 95% uncertainties. Pb*, radiogenic Pb.

**Figure 8** U–Pb analytical data for zircons from supracrustal rocks of the Honman Formation; sample GSWA 199046. Crossed squares indicate data not included in the age calculation (see text). Error bars are 1 sigma. Mean ages are quoted with 95% uncertainties. Pb*, radiogenic Pb.

**Figure 9** U–Pb analytical data for zircons from granitoid samples. (a) GSWA 182308; data are shown in a Wetherill concordia diagram to illustrate the good fit to a zero-age discordia; the inset shows four data points <5% discordant on which the age of the sample is based. Crossed squares indicate discordant data (see text). (b) ARC 23; seven data >35% discordant are not shown. Error bars are 1 sigma. Mean ages are quoted with 95% uncertainties. Pb*, radiogenic Pb.
aligned plagioclase and mafic schlieren. Zircons are 100–200 μm long, mainly euhedral, and locally rounded. Oscillatory zoning is common, but some crystals contain metamict areas. CL images (Figure 10) show concentric oscillatory zoning. Some zircons contain texturally discordant cores, and some are mantled by unzoned rims. Owing to high uranium concentrations, the rims could not be analysed successfully.

Twenty-five analyses were obtained from 24 zircons (Figure 9b). Ten analyses ≥10% discordant and/or containing >1% common ²⁰⁶Pb are not considered further. The remaining 15 analyses can be divided into four groups. Two analyses (Group 1) yield identical ²⁰⁷Pb/²⁰⁶Pb dates and a weighted mean ²⁰⁷Pb/²⁰⁶Pb date of 2719 ± 16 Ma. Nine analyses (Group 2) yield a weighted mean ²⁰⁷Pb/²⁰⁶Pb date of 2773 ± 8 Ma (MSWD = 1.5). Three analyses (Group 3) yield a weighted mean ²⁰⁷Pb/²⁰⁶Pb date of 2901 ± 11 Ma (MSWD = 0.32). The single remaining analysis (Group 4) indicates a ²⁰⁷Pb/²⁰⁶Pb date of 2985 ± 14 Ma (1σ).

DISCUSSION

U-Pb zircon ages

The weighted mean ²⁰⁷Pb/²⁰⁶Pb dates of 2920 ± 4 Ma for the metarhyolite (LJD-0017-222) and 2718 ± 6 Ma for foliated monzogranite (GSWA 182308) appear to represent single age components (with variable disturbance), and are interpreted as the ages of igneous crystallisation.
AGE OF THE GREENSTONE SUCCESSION

Our new SHRIMP zircon ages for the Honman Formation constrain the timing of sediment deposition and magmatism in the Lake Johnston greenstone belt. Previous U–Pb dates of 2921 ± 4 and 2903 ± 5 Ma (Wang et al. 1996) for intermediate rocks of the Honman Formation are in good agreement with the age of 2920 ± 4 Ma obtained for the metarhyolite (LJD-0017-222) and metabasite (GSWA 199046). Two zircons in one of the samples studied by Wang et al. (1996) yielded younger dates of ca 2856 Ma, which were interpreted by them to be metamorphic. Although Wang et al. (1996) interpreted these phryic rocks as extrusive, Mueller & McNaughton (2000) re-interpreted them as metamorphosed sills that intruded after deposition of the Honman Formation and the komatiites.

Although none of these interpretations is conclusive, the maximum depositional age of 2876 ± 4 Ma determined in this study for the volcaniclastic rock (sample LJD-0017-770) is clear evidence that at least parts of the Honman Formation and overlying supracrustal rocks are younger than ca 2850 Ma. A younger limit for the greenstone formation is provided by the 2773 ± 8 Ma age (sample ARC 23) of granitoid intrusions that crosscut the stratigraphy. The upper part of the succession of the Lake Johnston greenstone belt, including the Roundtop Komatiites, was deposited between 2879 and 2765 or 2720 Ma.

Heggie et al. (2012a) discussed the possibility of the Transition Zone being a stratigraphic unit, or a structural zone, e.g. hydrothermally altered shear zone. The felsic volcaniclastic rock occurs within the Transition Zone, and its younger age supports a younger stratigraphic unit with slightly differing components, rather than a ‘structural’ zone. Nevertheless deformation and alteration has most likely occurred within the Transition Zone. Mechanisms for resetting and recrystallising zircon are discussed below, and indicate that the younger age is unlikely to be the sole result of resetting zircons and metamorphic growth.

Furthermore, the geochemical analyses indicate different sources for the felsic volcanic and volcaniclastic rocks (Table 1). The primitive mantle-normalised, multi-element and chondrite-normalised REE-patterns for LJD-0011-770 and LJD-0017-222 are very similar; although subtle differences suggest a different source for each. Both samples show a slight positive Th anomaly, and negative Ta, Nb and Ti anomalies. However, the volcaniclastic sample shows a small negative Sr anomaly and greater enrichment in all elements relative to the volcanic sample. The volcaniclastic sample is slightly more enriched in the LREE and has a negative Eu anomaly, whereas the volcanic sample shows no Eu anomaly. These variations in geochemistry are consistent with a plagioclase-rich source for the 2875 Ma felsic volcaniclastic rock and a garnet-rich source for the 2919 Ma felsic volcanic rock.

Our new results for the metarhyolite (LJD-0017-222) do not include any ca 2856 Ma zircons with high Th/U ratios, such as those interpreted as a possible metamorphic component by Wang et al. (1996). Although zircons with Th/U ratios above 0.2 are typically magmatic in origin (Hoskin & Black 2000), this criterion is not conclusive. However, it is possible that owing to a lack of zirconium, zircon components in samples described by Wang et al. (1996) and this study are xenocrystic and the age of eruption for all samples is ca 2856 Ma or younger.

Since the late 1990s intensive investigations into fluid flow in zircons, zircon annealing, radiation damage and element ratios in zircons, have been undertaken (e.g. Cherniak et al. 1998; Meldrum et al. 1998; Nasdala et al. 1999, 2001, 2002; Vavra et al. 1999; Hoskin & Black 2000; Geisler et al. 2002, 2003, 2007), and the understanding of zircon formation has improved significantly. Work by Watson & Harrison (1983), Mezger & Krogstad (1997), and Hoskin & Schaltegger (2003) have shown that zircon does not recrystallise below ~1000°C.

Geisler et al. (2007) proposed that U–Pb ages from re-equilibrated domains in zircons are more likely to correspond to the time of fluid influx or melt production than to the age of metamorphism. Ayers et al. (2003) suggested that growth of metamorphic rims around primary zircons by Ostwald ripening is strongly enhanced by the presence of an aqueous fluid or melts. These zircon domains, which evolved during dissolution–recrystallisation processes, should contain lower minor and trace-element concentrations compared with the parent zircon. Metamict zircons retain the parent isotopic characteristics and a high concentration of non-formula elements (Geisler et al. 2003, 2004; Rayner et al. 2005).

Given these arguments, only an isotopic and trace element study of the different zones and patches within the
zircon could shed light into their genesis, because metamorphic zones should show a different pattern. Nevertheless, a study by Kirkland et al. (2009) showed that even with high fluid flow age information may be preserved in zircons within shear zones.

**REGIONAL CONTEXT**

The apparent absence of ca 2876 Ma and younger zircons in the metarhyolite (LJD-0017-222) and metadacite (GSWA 199046) can be explained by several processes and scenarios, which are discussed below:

1. The stratigraphic scheme of Heggie et al. (2012a), assuming that the metarhyolites and metadacites are part of the felsic volcanic rock association, but the volcaniclastic metasedimentary rock (LJD-0017-770) is part of the Transition Zone, can account for a 70 Ma hiatus between the deposition of these two stratigraphic levels.

2. On the other hand, if the there were no new zircon growth in the felsic volcanic rocks, owing to insufficient zirconium content, the eruption age would most likely be very similar to the maximum age for deposition of the volcaniclastic rocks (ca 2876 Ma).

3. If the felsic rock were an intrusive sill, as proposed by Mueller & McNaughton (2000), then the samples dated by Wang et al. (1996) and this study may be as young as ca 2720 Ma, depending on the age of deformation of the greenstone belt, because the felsic rock is sub-concordant to the greenstone belt stratigraphy and, therefore pre-dates deformation. The magma may have intruded along the contact between mafic extrusive rocks and felsic volcaniclastic rocks.

4. If the ca 2930 Ma age of Wang et al. (1996) were the age of the sill, then the Maggie Hays Formation would be older. This scenario would imply a hiatus of ca 70 Ma between felsic volcanic and volcaniclastic rocks, with intrusion of the sill prior to deposition of the felsic volcaniclastic rocks. Therefore, the sill would have intruded into, and postdate, the Maggie Hays Formation.

In the case of the lack of new zircon growth in the older samples, a similar observation has been made for the Southern Cross greenstone belt, where younger cross-cutting porphyries either do not contain young zircons or only a few younger discordant zircons (e.g. 2834 ± 7 Ma, Southern Star; Mueller & McNaughton 2000). Moreover, these samples do not show evidence of zircon growth during high-grade metamorphism or later skarn formation (Mueller & McNaughton 2000). A 2722 ± 5 Ma zircon age for the Copperhead porphyry in the Southern Cross greenstone belt is interpreted to reflect recrystallization, owing to similarity with ca 2775 Ma ages obtained for granitoid in the nearby Ghooli dome (Mueller & McNaughton 2000). Backscattered electron (BSE) and CL images provided by Mueller & McNaughton (2000) suggest that the younger dates are from areas within the zircons that exhibit rather featureless zoning and patchy reaction zones, whereas rounded zircons older than ca 2930 Ma preserve oscillatory zoning typical of magmatic growth (e.g. Mezger & Kroghstad 1997; Vavra et al. 1999).

The age of the Maggie Hays Formation remains unknown but, because it stratigraphically underlies the Honman Formation, it must be older than ca 2873 or 2930 Ma. Because the Lake Medcalf Igneous Complex intrudes the uppermost Maggie Hays Formation, the Lake Medcalf Igneous Complex is most likely younger than the Honman Formation. Intrusion of the Lake Medcalf Igneous Complex would have required the lithostatic overburden provided by the overlying Honman Formation and perhaps the Glasse Formation. Furthermore, the consistent intrusion level suggests an intact (i.e. undeformed) stratigraphy during intrusion.

In the Murchison Domain of the Youanmi Terrane, extensive and voluminous igneous complexes were emplaced at ca 2800 Ma (e.g. Windimurra and Narndee; Ivanic et al. 2010 and references therein), and less voluminous mafic intrusions ranging from ca 2810 to ca 2740 Ma are scattered across the Murchison and Southern Cross domains (e.g. Grass Flat Gabbro, Wingate et al. 2011a; Kathleen Valley Gabbro; Liu et al. 2002), and the Burtville (e.g. Swincer Dolerite and Mapa Igneous Complex GSWA 185968 and 185976; Wingate et al. 2011b, c) and Kurnalpi terranes (Kositcin et al. 2009) of the Eastern Goldfields Superterrane.

Because the greenstone succession of the Lake Johnston greenstone belt was deposited prior to intrusion of the ca 2774 and ca 2720 Ma granitoids, the Lake Medcalf intrusive rocks, which are concordant with the greenstone belt stratigraphy, must be older and possibly associated with the other ca 2810 Ma igneous complexes known from the Murchison Domain (Ivanic et al. 2010) and the northern Southern Cross Domain (Riganti et al. 2010).

Although the age and geochemical relationships between the Lake Medcalf Igneous Complex and the Roundtop Komatiites are still unknown, the komatiites have to be deposited within a short time span. Buck et al. (1996) assumed that the deposition of the Western Ultramafic Unit was coeval with deposition of the Honman Formation. The chemical sedimentation, resulting in the formation of BIF, and concomitant felsic volcanism, took place during quiescent periods in the Western Ultramafic Unit volcanic event. Heggie et al. (2012a) proposed an intrusive level of komatiites (central ultramafic unit) and a cogenetic extrusive level (western ultramafic unit), which require the komatiites postdate the Honman Formation. Either way, the ca 2874 Ma felsic volcaniclastic rocks (LJD-0011-770) imply that the Roundtop Komatiite is younger than ca 2830 Ma. Hence, the ‘Barberton’-type komatiite (Heggie et al. 2012b) of the Lake Johnston greenstone belt were deposited between ca 2786 and ca 2720 Ma, and are older than the ca 2710–2700 Ma ‘Munro’-type komatiites of the Eastern Goldfields Superterrane (Kositcin et al. 2008 and references therein).

Even though the older age constraint may be within the range of the ‘Munro’-type komatiites of the Eastern Goldfields Superterrane, the Lake Johnston greenstone belt was already deformed prior to granitoid intrusion. Moreover, the Glasse Formation was deposited above the Roundtop Komatiite, indicating that the Roundtop Komatiite is not the youngest part of the greenstone stratigraphy. The assumption that the older age constraint may be ca 2720 Ma raises the possibility...
that the Glasse Formation may be related to younger (ca 2805 or 2770 Ma) mafic events recognised elsewhere in Yilgarn (Liu 2007; Ivanic et al. 2010; Pawley et al. 2012).

Heggie (2010) and Fiorentini et al. (2011) have shown that komatiites in the Lake Johnston greenstone belt are comparable, and may be related, to komatiites in the Forrestania greenstone belt. If this were correct, the Forrestania komatiites would likely to be the same age as the Roundtop Komatiite and imply that both the Forrestania and Lake Johnston greenstone belts may not preserve komatiites older than ca 2870 Ma.

The Southern Cross greenstone belt (Figure 1) includes high-MgO komatiites, which display transitional character between Munro and Barberton types. These komatiites are comparable with those in the upper younger portions of the Forrestania greenstone belt (Thébaud & Barnes 2012). So far, based on geochronological evidence, it is interpreted that the Forrestania and Southern Cross greenstone belt sequences were deposited at the same time, and the difference in komatiite geochemistry is the result of lateral variation in depth of melting (Thébaud & Barnes 2012).

The Ravensthorpe greenstone belt, the southernmost greenstone belt in the Southern Cross Domain, has a slightly different stratigraphic, with the sedimentary Chester Formation underlying olivine spinifex-textured komatiites. Parts of this stratigraphy, consisting of felsic volcanic rocks, sedimentary rocks and komatiites, may be equivalent to the Honman Formation of the Lake Johnston greenstone belt. Above the Maydon Basalt, the topmost Hatfield Formation is marked by felsic volcanic and sedimentary rocks (Witt 1997). As yet, the only available geochronology from the Ravensthorpe belt includes one tonalite (2966 ± 12 Ma; Savage et al. 1996), a felsic volcanic rock from the upper silicic volcanic and sedimentary succession (2958 ± 4 Ma; Nelson 1995) and the Annabelle Volcanics (2989 ± 11 Ma; Witt 1999). The intimate spatial association of tonalite, tonalite porphyry dykes, and calc-alkaline volcanic rocks suggests a comagmatic relationship (Witt 1997).

Older felsic components have been reported in intrusive and extrusive felsic rocks of the ca 2935 Ma Mount Gibson (Yeats et al. 1996) and ca 2950 Ma Golden Grove (Wang et al. 1998) areas in the Murchison Domain and the Penneshaw Formation in the Norseman (Murchison Domain). Like the Lake Johnston rocks, the Hornet felsic schist of the Mount Gibson area contains a young (2827 ± 13 Ma), high Th/U zircon generation. Yeats et al. (1996) interpreted these zircons as hydrothermal and associated with gold formation because a nearby post-tectonic granitite porphyry is slightly younger (2623 ± 7 Ma). However, in the case of the Lake Johnston belt, the ca 2856 Ma (Wang et al. 1996) and ca 2876 Ma (this study) ages are too old to be related to a metamorphic or hydrothermal event. The only age constraint on metamorphism and hydrothermal overprint in the Lake Johnston belt is a study by Joly et al. (2010), who reported monazite growth at ca 2630 Ma. This reported age coincides with the age of late-stage gold mineralisation in adjacent greenstone belts (Mueller et al. 2004).

Given that the internal granitoids are younger than the greenstone sequences in the Lake Johnston belt, they could have triggered zircon growth. The foliated metagranite (GSWA 182308) yields a single 2718 ± 6 Ma zircon age component. This may indicate new heat influx at the time, with magma generation and new zircon growth. However, no zircons have been identified within rocks of the greenstone sequences at this time.

CONCLUSIONS

The Lake Johnston greenstone belt consists of five successions, based on lithology and geochemistry: the lowermost Maggie Hays Formation, Honman Formation, Roundtop Komatiite, Glassie Formation and Lake Medcall Igneous Complex, respectively. The ca 2874 Ma felsic volcaniclastic rocks of the upper Honman Formation and the ca 2774 to 2720 Ma crosscutting granitoid provide new younger age constraints on the greenstone succession of the Lake Johnston greenstone belt, including the ‘Barberton-type’ Roundtop Komatiites. Moreover, our new results question the existence of ca 2.9 Ga komatiites in the Southern Cross Domain of the Yilgarn Craton. Nevertheless, the age of the Maggie Hays Formation is still unknown, and the Ravensthorpe greenstone belt does appear to contain an older greenstone succession.

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SUPPLEMENTARY PAPERS

Table A1 Ion microprobe analytical results for zircons from samples from the Lake Johnston greenstone belt.