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Basin analysis in polymetamorphic terranes: An example from east Antarctica

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ABSTRACT

U–Pb and Lu–Hf isotopic signatures of detrital and metamorphic zircon in granulite facies paragneiss from Kemp and MacRobertson lands, east Antarctica, distinguish two extensive Precambrian basins: (1) an Archaean basin deposited between *c.* 2570 Ma and *c.* 2500–2430 Ma, which extended across Kemp Land contiguous with equivalent paragneiss in the adjacent Napier Complex in Enderby Land; and (2) a late Mesoproterozoic basin deposited after *c.* 1110–1080 Ma in MacRobertson Land that extends into eastern Kemp Land, which also incorporates rocks currently exposed in the Eastern Ghats Province of India and the Prydz Bay region of east Antarctica. An unconformity that separates the Archaean and Mesoproterozoic sedimentary packages is near the Stillwell Hills. An eastward increase in the proportion of fragmentary detrital zircon in the younger package reflects proximity to late Mesoproterozoic volcanic sources. The Mesoproterozoic basin likely developed along an Andean-type margin to the Indo-Antarctic craton, possibly in a back-arc setting during the assembly of Rodinia. The widespread dissolution of zircon at *c.* 990–940 Ma during Rayner orogenesis in Kemp and MacRobertson lands largely precludes a traditional basin analysis via detrital geochronology, but the Hf isotopic ratio of metamorphic zircon precipitated following anatexis provides a homogenised signature, equivalent to a bulk-rock model age, which can be used as a stratigraphic proxy. This approach captures “hidden” provenance information to enable lithostratigraphic correlation in complex high-grade terranes.

1. Introduction

Zircon detritus can provide a record of the age, composition, pressure, temperature and configuration of eroded protoliths to sediments, and provide information critical to reconstructing pathways between sediment provenance and basin fill. These regional links between sedimentary rocks and their sources are crucial to terrane-scale correlations. High-grade metamorphism drives the anatexis of metasedimentary rocks (*e.g.* Thompson, 1982; Powell and Downes, 1990; Brown, 1994; White et al., 2001, 2003, 2004), and may include the partial to complete dissolution of detrital zircon by partial melt in fertile crust (*e.g.* Watson and Harrison, 1983; Kelsey et al., 2008a; Kelsey and Powell, 2011). The nucleation and growth of metamorphic zircon during cooling and crystallisation of remaining partial melt can be expected to occur both as new grains and overgrowths on relict grains. Any surviving detrital grains (left undissolved once the melt reached Zr-saturation, or protected within unreacted porphyroblastic matrix minerals) are vulnerable to secondary alteration. The results of integrated CL imaging, U–Pb zircon dating and Lu–Hf isotope analysis of zircon domains in metasedimentary rocks can be used to distinguish distinct zircon populations in such rocks, including: (1) relict detrital; (2) *in situ* metamorphic; and (3) isotopically disturbed (Pb-loss), zircon populations. Detrital grains may be highly variable in morphology and internal zonation and yield diverse U–Pb ages and initial ^{176}Hf/^{177}Hf (Hf$_i$) ratios (Griffin et al., 2004; Belousova et al., 2009; Howard et al., 2011). Dissolution and subsequent precipitation of newly grown zircon following anatexis can be expected to homogenise U–Pb ages and Hf$_i$ ratios due to comparatively high diffusivity of Hf in a silicate melt and the similar geochemical behaviour of both Zr and Hf (Gerdes and Zeh, 2009; Zeh et al., 2010b), and result in metamorphic grains or overgrowths crystallised *in situ* that are commonly multi-faceted and sector-zoned (Corfu et al., 2003; Kelly and Harley, 2005). Isotopically disturbed grains are perhaps the most difficult to recognise, zircon of this type either having undergone re-equilibration during pseudomorphic alteration (Geisler et al., 2007), or protracted high-*T*/high-strain volume diffusion (Ashwal et al., 1999; Halpin et al., 2012b). The initial ^{176}Hf/^{177}Hf value incorporated into the zircon lattice during growth, is not modified by these processes that may

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disturb or reset the U–Pb isotopic system (Patchett, 1983; Kinny et al., 1991; Kinny and Maas, 2003; Lenting et al., 2010). Therefore, the decoupling of U–Pb and Lu–Hf isotopic systems during zircon alteration/modification is a valuable tool to distinguish these grains (Bomparola et al., 2007; Whitehouse and Kemp, 2010; Halpin et al., 2012b). The compounding effects of multiple metamorphic events increases complexity in the decoding of provenance in ancient terranes, and can require the discrimination of several *in situ* metamorphic populations from relict detrital grains multiply and variably disturbed. The zircon population in common metapelitic gneiss could be expected to totally dissolve at high-degrees of partial melting and strip the rock of its detrital isotopic record. In contrast, detrital zircon is more likely to survive high-grade metamorphism in low-fertility rocks, such as quartzite (Grew et al., 2012). Provenance studies in granulite terranes therefore require the evaluation of zircon from diverse rock compositions.

This study uses U–Pb and Lu–Hf isotopic zircon data to interrogate the age, provenance and tectonothermal history of nine metasedimentary rocks from a composite high-grade terrain in east Antarctica. The data define the extent of Archaean and late Mesoproterozoic basins between MacRobertson and Kemp lands (Fig. 1), a *c.* 300 km stretch of coastline intensely deformed and metamorphosed at *c.* 990–910 Ma (*i.e.* the Rayner Complex; Kamenev, 1972; Black et al., 1987; Sheraton et al., 1987). Isotopic data are used to establish provenance and lithostratigraphic correlations both within east Antarctica and across conjugate margins in India. These data reflect the juxtaposition of SE India and east Antarctica from at least the Palaeoproterozoic to Gondwana dispersal, refining our understanding of this continental fragment in supercontinent assembly.

2. Geological framework

Rocks of the MacRobertson and Kemp Land coast (Fig. 1) are divided into four major units: the Mawson Charnockite, Painted Gneiss, Colbeck Gneiss and Stillwell Gneiss (Crohn, 1959; McCarthy and Trail, 1964; Trail et al., 1967; Trail, 1970). Islands and nunataks in the vicinity of Mawson Station and mountains to the south (Fig. 1b) are mainly formed of Mawson Charnockite, originally defined (*sensu stricto*) as being emplaced at *c.* 970 Ma (Young and Black, 1991). The body is now known to be a composite pluton formed from several pulses of magmatism between *c.* 1150 and 960 Ma with two-stage (crustal) Hf model ages ($T_{\mathrm{DM}}^{\mathrm{Hf(c)}}$) *c.* 2.3–1.9 Ga (where Lu/Hf = 0.01; Halpin et al., 2012b). The charnockite intrudes, and contains abundant inclusions of, layered felsic gneiss, calc-silicate, metapelite and minor mafic gneiss collectively termed the Painted Gneiss. The age of this unit is not well constrained. Rb–Sr whole-rock ages between at *c.* 1250 and 1150 Ma for felsic gneiss from the Framnes Mountains (Fig. 1b) most probably reflect the partial to complete recrystallisation of protoliths age(s) during *c.* 990–910 Ma metamorphism (Young and Black, 1991). Zircon from orthogneiss xenoliths in Mawson Charnockite at Mawson Station yield a spectrum of apparent ^{207}Pb/^{206}Pb ages between *c.* 1000 and *c.* 2450 Ma. Most analyses cluster between 1700 Ma and 2000 Ma, including a chord reflecting *c.* 1000 Ma Pb-loss, and a concordant population at *c.* 1300–1400 Ma (Young and Black, 1991).

Metasedimentary and granitic gneiss termed the Colbeck Gneiss forms most coastal exposures between Chapman Ridge and the Scoble Glacier (Fig. 1b). These rocks resemble the Painted Gneiss (Trail, 1970) and include felsic paragneiss, metapelite and calc-silicate gneiss (White and Clarke, 1993; Dunkley et al., 2002). Both units were intruded by the Mawson Charnockite and are likely to be conformable and/or partially equivalent. U–Pb zircon ages from exposures at Cape Bruce reflect extensive *c.* 990–910 Ma felsic magmatism and anatexis (Dunkley et al., 2002).

Outcrops in Kemp Land westward from the Scoble Glacier to Edward VIII Gulf (Fig. 1b) are dominated by distinctive, layered orthogneiss called the Stillwell Gneiss. It predominantly comprises alternating pyroxene-plagioclase-rich and quartz-feldspar-rich layers and includes multiple protoliths of distinct age. The oldest recognised components are formally named the Stillwell Orthogneiss in the vicinity of the Stillwell Hills (Halpin et al., 2012a) and give *c.* 3650–3460 Ma U–Pb zircon ages (minimum emplacement) with $T_{\mathrm{DM}}^{\mathrm{Hf(c)}}$ average values between 3.9 and 3.6 Ga (Kelly et al., 2004; Halpin et al., 2005). These are similar to those of rocks from the adjacent Achaean Napier Complex (Halpin et al., 2005). Younger cross-cutting felsic orthogneiss identified in the Oygarden Group has a minimum emplacement age of *c.* 2800 Ma (Kelly et al., 2004), and a *c.* 1600 Ma intrusive charnockite has been identified in the Stillwell Hills (Scoresby Charnockite; Halpin et al., 2005, 2012a). Subordinate supracrustal rocks including quartzite, and pelitic, psammitic and calc-silicate gneiss are intercalated and interfolded with the Stillwell Gneiss (Clarke, 1988). Some of these supracrustal layers constitute mappable units in the Stillwell Hills, of which the particularly distinctive Ives Paragneiss is tentatively correlated with the Colbeck and Painted gneisses exposed to the east (McLeod et al., 1966). Subordinate metapelitic and calc-silicate gneiss occurs as discontinuous lenses or pods within Archaean orthogneiss in the Oygarden Group (Kelly, 2000). In most cases, the effects of intense deformation mask the nature of contacts between components of the Stillwell Gneiss. Metamorphosed basic dykes, which locally cut Archaean-Palaeoproterozoic foliations in the Stillwell Gneiss are interpreted as equivalent to the *c.* 1190 Ma Amundsen dyke suite in the Napier Complex (Sheraton et al., 1980; Sheraton and Black, 1981). Neoproterozoic magmatism in Kemp Land appears to be restricted to pegmatitic intrusions (Clarke, 1988; Grew et al., 1988; James et al., 1991; Kelly et al., 2002).

The zircon U–Pb and Hf isotopic data distinguish at least two terranes exposed along this east Antarctic coastline (Fig. 1): (i) an older craton of Archaean to Palaeoproterozoic orthogneiss and subordinate supracrustal material tectonically re-worked during the Proterozoic (Kemp Land); and (ii) an adjacent block dominated by comparatively juvenile Proterozoic orthogneiss and paragneiss (MacRobertson Land). Much of the Kemp Land terrane represents part of the Napier Complex tectonically reworked during the Rayner Orogen. The MacRobertson Land terrane represents Proterozoic crust accreted to a complexly deformed Kemp-Napier Archaean craton after *c.* 1600 Ma. Any incorporation of juvenile material during orogenesis was minor, indicating that this part of the Rayner Complex consists mainly of ensialic crust.

The Rayner Orogeny caused widespread deformation and recrystallisation at granulite facies conditions along the Kemp and MacRobertson land coast. Metamorphic assemblages in Kemp Land record a clockwise P–T–t path with peak conditions at $T \geq 800$–$990\,^{\circ}$C and $P = 7.4$–10 kbar, with pressure increasing westward towards the Napier Complex (Kelly and Harley, 2004; Halpin et al., 2007b). Electron microprobe-derived (Th + U)–Pb monazite ages indicate that the major mineral textures in these rocks developed at *c.* 940 Ma and the high-T recrystallisation occurred over *c.* 25–40 m.y. (Halpin et al., 2007b; Kelly et al., 2012). In contrast, metapelitic gneiss from the MacRobertson Land coast records an anticlockwise P–T–t path involving heating with an increase in pressure to peak conditions of $P = 5.6$–6.2 kbar and $T = 850\,^{\circ}$C (Cape Bruce; Fig. 1b) and $P = 5.4$–6.2 kbar at $T = 900\,^{\circ}$C (Forbes Glacier; Fig. 1b), prior to the emplacement of regionally extensive charnockite and granitoid. Chemical monazite ages suggest that peak P–T conditions occurred at *c.* 990–970 Ma; elevated temperatures are interpreted to have been sustained by successive magmatic intrusions, and subsequent cooling occurred over *c.* 80 m.y. without any appreciable change in pressure (Halpin et al., 2007a). The apparently conflicting P–T–t paths within these contemporary parts of