Early Neoproterozoic granulite facies metamorphism of mafic dykes from the Vestfold Block, east Antarctica

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ABSTRACT
Proterozoic mafic dykes from the southwestern Vestfold Block experienced heterogeneous granulite facies metamorphism, characterized by spotted or fractured garnet-bearing aggregates in garnet-absent groundmass. The garnet-absent groundmass typically preserves an ophitic texture composed of lathy plagioclase, intergranular clinopyroxene and Fe–Ti oxides. Garnet-bearing domains consist mainly of a metamorphic assemblage of garnet, clinopyroxene, orthopyroxene, hornblende, biotite, plagioclase, K-feldspar, quartz and Fe–Ti oxides. Chemical compositions and textural relationships suggest that these metamorphic minerals reached local equilibrium in the centre of the garnet-bearing domains. Pseudosection calculations in the model system NCFMASHTO (Na2O–CaO–FeO–MgO–Al2O3–SiO2–H2O–TiO2–Fe2O3) yield P–T estimates of 820–870 °C and 8.4–9.7 kbar. Ion microprobe U–Pb zircon dating reveals that the NW- and N-trending mafic dykes were emplaced at 1764/±25 and 1232/±12 Ma, respectively, whereas their metamorphic ages cluster between 957/±7 and 938/±9 Ma. The identification of granulite facies mineral inclusions in metamorphic zircon domains is also consistent with early Neoproterozoic metamorphism. Therefore, the southwestern margin of the Vestfold Block is inferred to have been buried to depths of ~30–35 km beneath the Rayner orogen during the late stage of the late Mesoproterozoic/early Neoproterozoic collision between the Indian craton and east Antarctica (i.e. the Lambert Terrane or the Ruker craton including the Lambert Terrane). The lack of penetrative deformation and intensive fluid–rock interaction in the rigid Vestfold Block prevented the nucleation and growth of garnet and resulted in the heterogeneous granulite facies metamorphism of the mafic dykes.

Key words: early Neoproterozoic; east Antarctica; granulite facies metamorphism; mafic dykes; P–T conditions; Vestfold Block.

INTRODUCTION
Old cratonic blocks and/or continental marginal basements may experience relatively young orogenic processes. In and near the late Mesoproterozoic/early Neoproterozoic (i.e. Grenvillian) Rayner orogen exposed in Kemp Land, MacRobertson Land and Princess Elizabeth Land of east Antarctica, there exist four Archean/Palaeoproterozoic cratonic blocks that preserve different crustal histories (Harley, 2003; Boger, 2011), including the Lambert Terrane in the southern Prince Charles Mountains, the Napier Complex in Kemp Land and the Rauer Group and Vestfold Block in Prydz Bay (Fig. 1). Given that much of the Rayner orogen was affected by late Neoproterozoic/Cambrian (i.e. Pan-African) high-grade metamorphism and deformation, it remains a matter of debate as to when these cratonic blocks were assembled (Harley et al., 2013; Liu et al., 2013, and references therein). This has led to different formation models of the proposed Rodinia and Gondwana supercontinents (Fitzsimons, 2003; Harley, 2003; Yoshida et al., 2003; Zhao et al., 2003). Therefore, the detailed investigation of the reworking processes and mechanisms of each cratonic block is essential for understanding the nature and role of the Rayner orogeny.

The Vestfold Block is an Archean/Palaeoproterozoic cratonic fragment located 15 km northeast of the Rauer Group. One of the most important features of the Vestfold Block is the occurrence of several mafic dyke swarms of Proterozoic age (Black et al., 1991a; Lanyon et al., 1993). Furthermore, some of these mafic dykes experienced amphibolite facies recrystallization and deformation (Collerson & Sheraton, 1986; Kuehner & Green, 1991; Passchier et al., 1991), although the metamorphic age remains unconstrained. However, careful petrographic observations indicate that mafic dykes from the southwestern Vestfold Block underwent heterogeneous granulite facies metamorphism (Liu et al., 2013). This provides an opportunity to unravel the behaviour of
the Vestfold Block during late Mesoproterozoic to Cambrian orogenies and establish its relationship with other granulite terranes in the Rayner orogen. In this article, we describe the petrological characteristics of metamorphosed mafic dykes from the Vestfold Block, estimate their metamorphic $P-T$ conditions based on pseudosection calculations, and determine the ages for the granulite facies metamorphism using sensitive high-resolution ion microprobe (SHRIMP) U–Pb zircon dating. These data are then used to discuss the tectonic evolution of the Rayner orogen in the context of the assembly of the Indian craton and east Antarctica.

REGIONAL GEOLOGY AND FIELD RELATIONSHIPS

The basement of the Vestfold Block is dominated by granulite facies orthogneiss with subordinate paragneiss (Fig. 2a). The basement rocks can be classified into five units according to composition and relative age with respect to metamorphic and deformational events. These units are the Chelnok Paragneiss, Taynaya Paragneiss, Tryne Mafic Gneiss, Mossel Gneiss and Crooked Lake Gneiss (Collerson et al., 1983; Black et al., 1991b; Harley, 1993; Snape & Harley, 1996; Snape et al., 1997, 2001). The Chelnok Paragneiss consists mainly of garnetiferous pelite and semipelite, with less abundant psammite, quartzite, calc-silicate and banded iron formation. Protoliths were deposited between 2575 and 2520 Ma (Clark et al., 2012). The Taynaya Paragneiss locally crops out as a discontinuous layer or boudin train in the Mossel Gneiss, comprising highly magnesian, silica-undersaturated sapphire-bearing lithologies. The Tryne Mafic Gneiss generally occurs as layered two-pyroxene granulate and also as nodular ultramafic xenoliths within younger units. The Mossel Gneiss consists predominantly of tonalitic orthogneiss, with subordinate trondhjemitic, granodioritic and granitic orthogneiss, which intruded the aforementioned three constituent units between 2526 ± 6 and 2501 ± 4 Ma (Black et al., 1991b). The Crooked Lake Gneiss is the youngest unit and comprises medium- to coarse-grained orthogneiss of mostly dioritic, granodioritic, granitic and less commonly, gabbroic composition. Emplacement ages inferred for the Crooked Lake Gneiss range from 2501 ± 4 to 2484 ± 6 Ma (Black et al., 1991b). Two episodes or a protracted period of high-grade metamorphism and deformation took place from 2520 to 2450 Ma (Collerson et al., 1983; Black et al., 1991b; Snape et al., 1997; Clark et al., 2012).

Several mafic dyke swarms cut the high-grade basement of the Vestfold Block. On the basis of geochemistry, isotopic dating, cross-cutting relationships and orientation, at least four generations have been recognized (Collerson & Sheraton, 1986; Black et al., 1991; Harley, 1993; Snape & Harley, 1996; Snape et al., 1997, 2001). The Chelnok Paragneiss consists mainly of garnetiferous pelite and semipelites, with less abundant psammites, quartzites, calc-silicates and banded iron formation. Protoliths were deposited between 2575 and 2520 Ma (Collerson et al., 2012). The Taynaya Paragneiss locally crops out as a discontinuous layer or boudin train in the Mossel Gneiss, comprising highly magnesian, silica-undersaturated sapphire-bearing lithologies. The Tryne Mafic Gneiss generally occurs as layered two-pyroxene granulate and also as nodular ultramafic xenoliths within younger units. The Mossel Gneiss consists predominantly of tonalitic orthogneiss, with subordinate trondhjemitic, granodioritic and granitic orthogneiss, which intruded the aforementioned three constituent units between 2526 ± 6 and 2501 ± 4 Ma (Black et al., 1991b). The Crooked Lake Gneiss is the youngest unit and comprises medium- to coarse-grained orthogneiss of mostly dioritic, granodioritic, granitic and less commonly, gabbroic composition. Emplacement ages inferred for the Crooked Lake Gneiss range from 2501 ± 4 to 2484 ± 6 Ma (Black et al., 1991b). Two episodes or a protracted period of high-grade metamorphism and deformation took place from 2520 to 2450 Ma (Collerson et al., 1983; Black et al., 1991b; Snape et al., 1997; Clark et al., 2012).

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been dated at 1380 ± 7 Ma (Lanyon et al., 1993). The youngest Group III high-Fe mafic dykes are NNE- to N-trending and have been dated at 1248 ± 4 and 1241 ± 5 Ma, respectively (Black et al., 1991a; Lanyon et al., 1993). In addition, alkaline and ultramafic lamprophyric dykes were emplaced before the intrusion of Group II or III mafic dykes. In the northern part of the Vestfold Block, the dykes are un-deformed and display chilled margins and baked contacts, but in the southeastern part of the Vestfold Block, some dykes are metamorphosed and locally deformed (Collerson & Sheraton, 1986; Passchier et al., 1991). Garnet is developed along dyke margins and in fracture zones within and across the dykes, and such features have been interpreted as the result of recrystallization during a Neoproterozoic or younger (1000–500 Ma) amphibolite facies metamorphism (600–660 °C and 5.6–7.5 kbar) (Kuehner, 1986; Kuehner & Green, 1991; Hoek et al., 1992; Snape et al., 2001; Zulbati & Harley, 2007).

Our field investigation focused on the Mule Peninsula of the southwestern Vestfold Block (Fig. 2b). The basement rocks in this area comprise the Chelnock Paragneiss, Mossel Gneiss and Crooked Lake Gneiss (Snape et al., 2001; Clark et al., 2012). These rocks have a regional ENE-trending gneissosity, which dips towards the north at angles of 60–80°. There are two sets of mafic dykes that intrude the basement. One NW-trending set corresponds to the Group I high-Fe mafic dykes. The more common N-trending set (Fig. 3a), with a few dykes showing a NNE trend, cuts both the NW-trending dykes and the gneissosity of the basement gneisses (Fig. 3b), and is probably, part of the Group III high-Fe mafic dykes. All these dykes are 2–8 m in width, and dip at a steep angle of ~80° towards the east for the N-trending dykes and towards the southwest for the NW-trending dykes. Most dykes are massive, and only some show weak deformational features, as evident from oriented fine-grained mineral assemblages (Fig. 3c). The granulite facies metamorphism of most mafic dykes is heterogeneous, characterized by a spotted texture produced by pink garnet-rich aggregates in a grey to dark groundmass (see Fig. 3b,c). Some dykes are only weakly recrystallized, but fine-grained metamorphic orthopyroxene, clinopyroxene and hornblende were observed in all the collected dyke samples. Furthermore, millimetre-sized garnet-bearing zones are also found along the fractures in some of these dykes.

SAMPLES AND ANALYTICAL PROCEDURES

The $P$–$T$ conditions of metamorphism were estimated for nine N- and NW-trending metamorphosed mafic dykes. Of these, six samples were chosen for SHRIMP U–Pb zircon dating to define the metamorphic ages. The sample localities are shown in Fig. 2b, and the coordinates, occurrences, mineral assemblages and age results of the samples are summarized in Table 1. All of the samples have basaltic compositions, with $\text{Mg#}$ values (molar $\text{MgO}/(\text{MgO} + 0.85\text{FeO})$) ranging from 37 to 59 (Table 2).

Whole-rock chemical analyses were carried out using the XRF (3080E) facility at the National
Research Centre for Geoanalysis, Chinese Academy of Geological Sciences. Relative standard deviations of the analytical data are better than 5%. Chemical compositions of minerals were analysed using a JEOL JXA-8230 wavelength-dispersive electron microprobe at the Institute of Mineral Resources, Chinese Academy of Geological Sciences. The operating conditions were as follows: an accelerating voltage of 15 kV, a beam current of $2 \times 10^{-8}$ A and a counting time of 10 s for each peak. The beam diameter was set to 5 μm for most minerals, apart from exsolved pyroxene or inclusions in garnet and zircon (1–2 μm). Natural minerals were used as standards. Ferric iron in garnet, clinopyroxene and orthopyroxene was determined by charge balance. Molar formulae for amphibole were calculated following Holland & Blundy (1994), with modifications following Dale et al. (2000). Ferric iron in biotite was estimated on the basis of the assumption of $\text{Fe}^{3+} / (\text{Fe}^{2+} + \text{Fe}^{3+}) = 0.116$ (Holdaway et al., 1997).

$\Upsilon$–$\text{Th}$–$\text{Pb}$ zircon analyses were performed using the SHRIMP II ion microprobe at the Beijing SHRIMP Centre, Chinese Academy of Geological Sciences. Zircon was extracted using conventional techniques, including crushing, sieving, heavy liquid separation and handpicking. Zircon grains were mounted in an epoxy disc along with the TEMORA zircon standard and polished to expose grain centres. Internal structures of zircon grains were revealed by cathodoluminescence (CL) imaging. Instrumental conditions and data acquisition procedures followed Williams (1998). A primary ion beam of 4.5 nA, 10 kV $\text{O}_2^-$ and $~25 \mu m$ in size was used. The measured $^{206}\text{Pb}/^{238}\text{U}$ ratios were calibrated using analyses of the reference zircon TEMORA (416.75 ± 0.24 Ma; Black et al., 2003). Correction for common Pb was made using the measured $^{204}\text{Pb}$. Ages were calculated using the sQUID 1.03 (Ludwig, 2001) and ISOPLOT 3.23 (Ludwig, 2003) programs. The age uncertainties for individual analyses represent one standard deviation (1σ), but the calculated weighted mean $^{207}\text{Pb}/^{206}\text{Pb}$ ages are quoted at the 95% confidence level.

**PETROGRAPHY**

Every metamorphosed mafic dyke contains garnet-bearing and garnet-absent domains. Garnet-absent domains (i.e. groundmass) typically preserve an ophiitic texture composed of lathy plagioclase, intergranular clinopyroxene and Fe–Ti oxides (Fig. 4a). Plagioclase is mostly randomly oriented, with the largest phenocrysts up to 1.2 mm long. All plagioclase laths display optical zoning (Fig. 4b). Their grey cores evident in backscattered electron (BSE) images (Fig. 4c) represent the relics of igneous plagioclase (pl1), whereas the narrow dark rims may reflect recrystallization during metamorphism. Igneous clinopyroxene (cpx1) contains closely spaced exsolution lamellae or patches of orthopyroxene and rare rods of ilmenite (Fig. 4d). Fine-grained orthopyroxene, clinopyroxene (cpx2), hornblende and Fe–Ti oxides are commonly developed around cpx1 and, in many cases, this primary phase has been completely replaced by secondary mineral assemblages. Locally, cpx1 occurs as large phenocrysts that are 1.0–1.2 mm long (Fig. 4e). A medium-sized K-feldspar grain (up to 0.3 mm) was also observed near pl1 in sample VHC74 (Fig. 4f), but it is difficult to determine whether this grain is of igneous or metamorphic origin, as it contains several quartz inclusions. However, in the weakly deformed sample VHC53, cpx1 and Pl1 disappear and hornblende is abundant, forming a typical metamorphic paragenesis (Fig. 4g). In addition, sample VHC73 contains orthopyroxene phenocrysts (opx1) (Fig. 4h) that contain rare clinopyroxene exsolution lamellae. Rims of these orthopyroxene phenocrysts are commonly replaced by an aggregate of fine-grained clinopyroxene (cpx2), orthopyroxene (opx2) and hornblende.

**Fig. 3.** Photographs showing the field occurrences and heterogeneous textures of metamorphosed mafic dykes from the Mule Peninsula of the southwestern Vestfold Block. (a) Distant view of N-trending mafic dykes. The person for scale is 175 cm high. (b) A metamorphosed mafic dyke (sample 08-10) showing an intrusive contact with paragneiss and a spotted texture characterized by pink garnet-rich aggregates in a grey groundmass. The coin for scale is 2.5 cm across. (c) A weakly deformed metamorphosed mafic dyke (sample VHC53) with a spotted texture. The pen for scale is 14 cm long.
Spotted garnet-bearing domains are typically several centimetres in diameter, occupying from ~10% (sample VHC69) to 60% (sample VHC53) of the groundmass by volume. This domain comprises mainly garnet and clinopyroxene (cpx), with minor fine-grained orthopyroxene, hornblende, biotite, plagioclase (pl), K-feldspar, quartz and Fe–Ti oxides (Fig. 5a). Garnet is allotriomorphic to hypidiomorphic and, in some cases, has a tabular shape (Fig. 5b), which is inferred to be pseudomorphs after plagioclase. Garnet in the centre of these domains is relatively coarse-grained (0.1–0.5 mm) and appears to be in textural equilibrium with the other phases (Fig. 5c), although K-feldspar is commonly crystallized around plagioclase or as irregular interstitial grains (Fig. 5d). Towards the margins of the garnet-bearing domains, the garnet grain size gradually decreases. In the transition zone between garnet-bearing and garnet-absent domains, vermicular garnet is commonly observed to have grown on both the primary clinopyroxene (Fig. 5e) and Fe–Ti oxides (Fig. 5f). Garnet contains vermicular inclusions of quartz, plagioclase, K-feldspar, hornblende, clinopyroxene, Fe–Ti oxides, and less commonly orthopyrox-
ene, biotite and pyrite (Fig. 5g). Fractured garnet-bearing domains are observed only in samples VHC73 and VHC74. The characteristics of metamorphic recrystallization in this domain of sample VHC74 are similar to those of spotted garnet-bearing domains in other samples. However, in sample VHC73, garnet is commonly inclusion-poor and is in sharp contact with nearly equigranular clinopyroxene (cpx2), orthopyroxene (opx2), hornblende, biotite, plagioclase (pl2), K-feldspar, quartz and Fe–Ti oxides (Fig. 5h).

MINERAL CHEMISTRY

The end-members of garnet were normalized on the basis of $X_{\text{an}} = \text{Fe}^{3+}/2$, $X_{\text{al}} = \text{Fe}^{2+}/3$, $X_{\text{spss}} = \text{Mn}/3$, $X_{\text{py}} = \text{Mg}/3$ and $X_{\text{gr}} = \text{Ca}/3 - \text{Fe}^{3+/2}/2$. Garnet consists predominantly of almandine ($X_{\text{al}} = 0.51–0.66$), pyrope ($X_{\text{py}} = 0.10–0.25$) and grossular ($X_{\text{gr}} = 0.10–0.21$), with minor spessartine ($X_{\text{spss}} = 0.01–0.04$) and variable andradite ($X_{\text{andr}} = 0.00–0.12$) (Table S1). In general, the garnet compositions depend on the bulk chemistry of the mafic dykes. Garnet has the highest $X_{\text{al}}$ and lowest $X_{\text{py}}$ in sample VHC53 (bulk rock Mg# = 37) and the lowest $X_{\text{al}}$ and highest $X_{\text{py}}$ in sample VHC73 (Mg# = 59), whereas the compositions of garnet from other samples (Mg# = 42–52) are intermediate to these extremes (Fig. 6a). Individual garnet grains are not significantly zoned relative to $X_{\text{al}}$, $X_{\text{py}}$, $X_{\text{gr}}$ and $X_{\text{spss}}$. However, garnet grains commonly show a decrease in $X_{\text{py}}$ and an increase in $X_{\text{al}}$ or $X_{\text{gr}}$ within a short distance (~50 μm) from grain boundaries (Fig. 7). Garnet coronae on clinopyroxene have compositions similar to granular garnet, but coronae on Fe–Ti oxides are slightly enriched in $X_{\text{al}}$ and poor in $X_{\text{py}}$ with respect to the cores of granular garnet.

Metamorphic clinopyroxene (cpx2) has aluminium contents of 0.03–0.12 p.f.u. and $X_{\text{Mg}}$ (Mg/(Mg + Fe²⁺)) of 0.60–0.87 (Table S2). Similar to garnet, clinopyroxene compositions vary with the bulk chemistry of the dykes (Fig. 6b). There is no clear compositional difference for cpx2 from different textural positions, except in sample VHC49 cpx2 from garnet-absent domains has lower $X_{\text{Mg}}$ than that from garnet-bearing domains. Clinopyroxene inclusions in garnet and zircon typically have slightly lower Al contents. In most cases, cpx2 cores are richer in Al and poorer in $X_{\text{Mg}}$ compared with rims, but the opposite trend, particularly for $X_{\text{Mg}}$, is also observed. Igneous clinopyroxene (cpx1) has compositions similar to metamorphic clinopyroxene (cpx2), which suggests that after the exsolution of orthopyroxene, cpx1 approached compositional equilibrium with other metamorphic minerals. If the host clinopyroxene is compositionally reintegrated with the exsolved orthopyroxene, then the original clinopyroxene is augitic.

Metamorphic orthopyroxene has Al of 0.02–0.07 p.f.u. and $X_{\text{Mg}}$ of 0.42–0.64 (Table S3). As for clinopyroxene, orthopyroxene compositions vary with the bulk chemistry of the dykes (Fig. 6c). Within a single sample, orthopyroxene from different textural positions does not show systematic variations, although most orthopyroxene rims and inclusions in zircon tend to have lower Al contents. Igneous orthopyroxene from sample VHC73 contains more Al (0.07–0.08 p.f.u.) than does metamorphic orthopyroxene, but both orthopyroxenes have similar $X_{\text{Mg}}$ values.

Hornblende is pargasite and ferropargasite (Cu₄p > 1.70; Na₂A + Kₐ > 0.60; Ti < 0.30; Si = 5.89–6.46 p.f.u.), with $X_{\text{Mg}}$ varying from 0.41 to 0.64 (Table S4; Fig. 6d). The $X_{\text{Mg}}$ of hornblende cores is similar to that of the rims. However, the compositions of hornblende inclusions in zircon commonly show large variations as compared with matrix hornblende.

Biotite shows highly variable compositions between samples, with Ti of 0.12–0.31 p.f.u. and $X_{\text{Mg}}$ of 0.41–0.75 (Table S5; Fig. 6e). There is little difference in composition between biotite cores and rims. Biotite inclusions in zircon from sample VHC74 have slightly lower $X_{\text{Mg}}$ than does matrix biotite.

The end-members of plagioclase were normalized based on $X_{\text{an}} = \text{Ca}/(\text{Ca} + \text{Na} + \text{K})$, $X_{\text{ab}} = \text{Na}/(\text{Ca} + \text{Na} + \text{K})$ and $X_{\text{or}} = \text{K}/(\text{Ca} + \text{Na} + \text{K})$. Metamorphic plagioclase (pl2) is oligoclase, andesine or labradorite ($X_{\text{an}} = 0.24–0.58$, $X_{\text{ab}} = 0.40–0.73$, $X_{\text{or}} = 0.01–0.03$), and varies in composition among samples (Table S6; Fig. 6f). In most cases, rims contain slightly less $X_{\text{an}}$ than cores. The compositions of plagioclase inclusions in garnet and zircon are indistinguishable from those of matrix plagioclase. Igneous plagioclase (pl1) is mainly labradorite and bytownite with $X_{\text{ab}} = 0.52–0.73$, $X_{\text{ab}} = 0.27–0.47$ and $X_{\text{or}} = 0.01–0.02$. A pl1 lath
from sample VHC69 has the highest $X_{an}$ of 0.80, whereas granular pl$_1$ from samples VHC72 and 08-10 has $X_{an}$ of 0.47–0.49, probably reflecting the effects of recrystallization. All these igneous plagioclase grains have pronounced compositional zoning, with decreasing $X_{an}$ and increasing $X_{ab}$ from core to rim (Fig. 8). The compositions of the pl$_1$ rim are similar to those of pl$_2$, suggesting that the rim compositions completely re-equilibrated during metamorphism.

K-feldspar has $X_{or}$ of 0.88–0.97, $X_{ab}$ of 0.02–0.11 and $X_{an}$ of 0.00–0.01 (Table S7). The cores and rims of K-feldspar, matrix K-feldspar and K-feldspar inclusions in garnet and zircon all have similar compositions. A medium-sized K-feldspar grain from sample VHC74 also shows similar compositions ($X_{or}$ = 0.92–0.93) to other K-feldspars.

**P–T PSEUDOSECTION CALCULATIONS**

A model system NCFMASHTO (Na$_2$O–CaO–FeO–MgO–Al$_2$O$_3$–SiO$_2$–H$_2$O–TiO$_2$–Fe$_2$O$_3$) was chosen for $P$–$T$ pseudosection calculations of the metamorphosed mafic dykes. The effective bulk-rock compositions, normalized into mole proportions in the whole-rock chemical analyses (Table 2). MnO and K$_2$O were ignored in the model system as they occur in only minor amounts. The presence of magnetite indicates a relatively high redox state during metamorphism. It was assumed that 15% of the total iron was ferric, on the basis of previously reported Fe$^{3+}$ contents of similar rocks (Bezos & Humler, 2005; Cottrell & Kelley, 2011) and test modelling. The occurrence of abundant hornblende and minor biotite suggests the presence of relatively large amounts of accessible H$_2$O. Therefore, H$_2$O is considered to be in excess in the calculations. The melt phase is neglected as it would only make up a small amount of the rock in the presence of hornblende (Poli & Schmidt, 2002), and there is no appropriate mixing model available for the melt in mafic rocks. As a result, the high-$T$ fields in the calculated pseudosections are likely to be metastable with respect to the missing melting reactions (Daczko & Halpin, 2009; Pitra et al., 2010).

Pseudosection calculations were performed using THERMOCALC 3.33 (Powell & Holland, 1988, 1994; Powell et al., 1998, updated October 2009), employing an updated version (November 2003) of the internally consistent thermodynamic data set (file tcds55.txt; Holland & Powell, 1998). Minerals and activity–composition models used in the calculations are garnet (g; White et al., 2007), clinopyroxene (shown as di; Green et al., 2007), orthopyroxene (opx; White et al., 2002), hornblende (hb; Diener et al., 2007), plagioclase (pl; Holland & Powell, 2003), magnetite (mt; White et al., 2000) and ilmenite (ilm; White et al., 2000). Quartz (q), water (H$_2$O) and rutile (ru) were modelled as pure end-member phases. The $P$–$T$ pseudosections were constructed for the ranges of 650–950 °C and 4–15 kbar for sample VHC46 (Fig. 9a, b) and 750–900 °C and 6–12 kbar for the other eight samples (Fig. 9c–j). All the phase diagrams share similar mineral assemblages. Most equilibria are tri-, quadri- or quinivariant with a divariant equilibrium involving garnet, clinopyroxene, orthopyroxene, hornblende, plagioclase, ilmenite, magnetite and quartz. Garnet-bearing equilibria occur at the high-$P$ region and orthopyroxene only appears at the low-$P$ and high-$T$ region.

The observed assemblage of g + di + opx + hb + pl + mt + ilm + q corresponds to the divariant equilibrium at $P$–$T$ conditions of $\sim$750–890 °C and $\sim$7.0–10.5 kbar in all the calculated pseudosections. Garnet isopleth thermobarometry based on plotting compositional isopleths of garnet as contours on a $P$–$T$ pseudosection (e.g. Zeh & Holness, 2003; Evans, 2004; Kim & Bell, 2005; Gaidies et al., 2006; Le Bayon et al., 2006; Wei & Song, 2008; Liu et al., 2009a; Wang et al., 2014) was applied to estimate peak metamorphic conditions. To eliminate the effect of diffusion during cooling, core compositions of garnet from the centre of garnet-bearing domains were chosen for all the samples. Figure 9b–j shows the contours of compositional isopleths of $X_{Mg}$ (Mg/(Mg + Fe$^{2+}$)) for garnet in relevant fields. Garnet cores from seven samples with intermediate Mg# values (between 42 and 52) have $X_{Mg}$ of 0.23–0.28, which restricts the $P$–$T$ conditions to 820–850 °C and 8.4–9.7 kbar. The stability field of g–hb–di–opx–pl–mt–ilm–(q+ H$_2$O) for the high-Mg# sample VHC73 is extremely narrow, and the plots of garnet core compositions fall slightly beyond the field. This may be due to an inappropriate assumption of Fe$^{3+}$ in this sample. If Fe$^{3+}$ is adjusted to 20% of the total Fe, the window of the stability field enlarges (see Fig. 9h), and the $X_{Mg}$ (30–0.32) of the garnet cores constrains $P$–$T$ conditions to 860–870 °C and 8.8–9.2 kbar, roughly in agreement with the $P$–$T$ data.
Fig. 6. Compositional variations of principal minerals from metamorphosed mafic dykes. (a) Ternary \((X_{\text{alm}} + X_{\text{spss}}) - (X_{\text{gr}} + X_{\text{andr}}) - X_{\text{py}}\) diagram for garnet. (b) Al vs. \(X_{\text{Mg}}\) plot for clinopyroxene. (c) Al vs. \(X_{\text{Mg}}\) plot for orthopyroxene. (d) \(X_{\text{Mg}}\) vs. Si plot for hornblende. (e) Ti vs. \(X_{\text{Mg}}\) plot for biotite. (f) Ternary \(X_{\text{or}} - X_{\text{ab}} - X_{\text{an}}\) diagram for plagioclase. Red colour, igneous mineral; yellow colour, mineral inclusion in garnet and zircon; blue colour, exsolved pyroxene; green colour, core of metamorphic mineral; open symbols, rim of igneous and metamorphic mineral; grey colour, garnet growing on Fe–Ti oxides.

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obtained for other samples. However, the $X_{\text{Mg}}$ of garnet cores from the low-Mg# sample VHC53 ranges only from 0.16 to 0.17, suggesting $P$–$T$ conditions of 770–780 °C and 7.5–8.3 kbar. This discrepancy is unclear; it is suspected that the heterogeneity of granulite facies metamorphism led to a slight mismatch between the bulk-rock compositions used for calculations and the analysed garnet compositions. Overall, we infer that the peak conditions of metamorphism of mafic dykes from the Vestfold Block reached ~820–870 °C and 8.4–9.7 kbar.

**SHRIMP U–Pb ZIRCON AGES**

Zircon from all samples is pale pink and rounded, ovoid or irregular in shape, with grain sizes of 30–100 μm. All zircon grains are homogeneous or show planar, sector and fir-tree zoning in CL images, with or without a weakly oscillatory-zoned or banded core (Fig. 10a–p). Most cores, except some from samples VHC73 and VHC74, are too small to be analysed. Zircon core-rim boundaries are generally embayed,
suggesting the formation of new zircon overgrowths through dissolution–precipitation. Zircon cores from sample VHC53 are commonly homogenized (see Fig. 10c), probably reflecting strong solid-state recrystallization. Numerous hornblende, plagioclase, quartz, and less commonly clinopyroxene, orthopyroxene, diopside, magnetite, ilmenite, rutile.

Fig. 9. $P$–$T$ pseudosections in the NCFMASHTO system for metamorphosed mafic dykes. The dyke compositions were used in molar per cent, which was calculated on the basis of the whole-rock chemical analyses. Contours of $X_{Mg}$ for garnet in the relevant $P$–$T$ range are shown in (b)–(j). di, clinopyroxene; g, garnet; hb, hornblende; ilm, ilmenite; mt, magnetite; opx, orthopyroxene; pl, plagioclase; q, quartz; ru, rutile.

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roxene, K-feldspar, ilmenite and apatite inclusions are found in the zircon overgrowths. Biotite inclusions are also observed in zircon from sample VHC74; however, garnet was unfortunately never identified as inclusions in zircon.

Twenty-four spot analyses were performed on zircon overgrowth domains from each sample, with the exception of sample 08-10 where only eight spot analyses were carried out on five zircon grains (Table S8). Some zircon analyses for sample VHC53 overlapped with homogenized cores. The majority of the analyses, apart from some older and/or younger data points that may reflect inheritance and post-crystallization Pb loss, respectively, form a discordant array near the concordia for each of the six samples. These data yield upper intercept ages and weighted mean $^{207}\text{Pb} / ^{206}\text{Pb}$ ages, respectively, of 958 ± 12 and 957 ± 7 Ma for sample VHC46, 1016 ± 35 and 990 ± 12 Ma for sample VHC53, 954 ± 14 and 948 ± 8 Ma for sample VHC72, 963 ± 33 and 956 ± 14 Ma for sample VHC73, and 944 ± 14 and 938 ± 9 Ma for sample VHC74 (Fig. 11a–e). In sample 08-10, six analyses produced an upper intercept age of 941 ± 42 Ma (Fig. 11f). The abundances of U and Th are highly variable among the samples, with $U = 241–1193$ ppm, $Th = 7–361$ ppm and $Th/ U = 0.02–0.32$. However, $Th/ U$ ratios of zircon from sample VHC53 are slightly elevated (0.12–0.53) and exhibit a general positive correlation with $^{207}\text{Pb} / ^{206}\text{Pb}$ ages (see Fig. 11b). Note that because most data points plot close to the upper concordia intercepts, their lower intercept ages are imprecise. However, the lower intercept ages of some samples (e.g. VHC53, VHC73 and 08-10) appear to coincide broadly with c. 500 Ma.

Twelve spot analyses were conducted on oscillatory-zoned and banded zircon cores from each of samples VHC73 and VHC74. For sample VHC73, excluding five young data points, the remaining seven analyses plot on or near a discordia with an upper intercept age of 1764 ± 25 Ma (see Fig. 11d), $Th/ U$ ratios for these zircon domains range from 0.10 to 0.98. For sample VHC74, apart from five slightly deviant analyses, the data fall on a discordia with an upper intercept age of 1232 ± 12 Ma (see Fig. 11e). These zircon cores have $Th/ U$ ratios varying from 0.09 to 0.80.

**DISCUSSION**

**Mechanisms and effects of heterogeneous granulite facies metamorphism**

The widespread development of granulite facies mineral assemblages in the mafic dykes indicates that regional high-grade Meso-Neoproterozoic metamorphism took place in the southwestern Vestfold Block. Peak $P-T$ conditions of metamorphism reached 820–870 °C and 8.4–9.7 kbar. However, granulite facies metamorphism is heterogeneous and, in most cases, ophitic textures are preserved. What factors prevented and delayed kinetic reactions amongst minerals during metamorphism of the mafic dykes? Structural studies suggest that only limited compressional deformation characterized by the sharply bound, ductile, <15 cm wide mylonite zones occurred in the Vestfold Block during the late Meso-proterozoic to early Neoproterozoic (Passchier et al., 1991; Hoek et al., 1992; Dirks et al., 1994). Apart from the development of millimetre-sized garnet-bearing zones along a few fractures in some mafic dykes, most of the dykes were unaffected by this deformation event, suggesting that the Vestfold Block was relatively rigid during metamorphism. The lack of penetrative deformation was one of the important controls on the heterogeneous metamorphism of the mafic dykes.

The appearance of hornblende, biotite and interstitial K-feldspar, as well as the growth of garnet along fracture zones, provides evidence of fluid ($H_2O$) activity during metamorphism. However, the mafic dykes were originally dominated by an anhydrous mineral assemblage of clinopyroxene + plagioclase + ilmenite + magnetite ± orthopyroxene, and no signs were observed of alteration having taken place before the metamorphism. On the other hand, the Vestfold Block is an old, dry, granulite terrane, and as a consequence possibly responded only weakly to the early Neoproterozoic high-grade metamorphism. This suggests that fluids had to be introduced from an external source. The criss-crossing of the Vestfold Block by mafic dykes and the development of limited shear zones and brittle faults provided conduits for such an external fluid input. However, $T–M_{H_2O}$ pseudosection modelling at 9 kbar for sample VHC46 suggests that orthopyroxene, clinopyroxene and plagioclase modes decrease and hornblende increases with an increase in $H_2O$ in the granulite facies $H_2O$-unsaturated field, but the garnet mode remains almost constant (Fig. 12). Therefore, although fluid flow is conducive to mass transfer, the nucleation, growth and enrichment of garnet in certain domains of the mafic dykes would have been little affected by the state of saturation or unsaturation of $H_2O$.

Subtle differences in the local compositions of the mafic dykes may have played a critical role in the initial heterogeneous nucleation of garnet. After those heterogeneous nuclei formed, garnet would grow predominantly around the existing nuclei to form large crystals, consequently destroying the ophitic textures. Although the original assemblage and mineral modes may vary in different domains in the mafic dykes, these domains were not completely isolated during metamorphism, particularly when there were fluids sufficient to facilitate element transitions. The growth of garnet may rely on the consumption of the *in situ* reactant minerals. However, the garnet-absent
domains may also have supplied the components for the growth of large garnet crystals in the garnet-bearing domains by fluid transportation and/or diffusion, which may have prevented the further nucleation of garnet in the garnet-absent domains. This may explain the formation of spotted textures in most of the mafic dykes. However, if penetrative deformation took place in the Vestfold Block, nucle-

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**Fig. 10.** Cathodoluminescence (CL) images of zircon from metamorphosed mafic dykes. (a) A homogeneous grain with a dark luminescent relict core from sample VHC46. (b) A homogeneous grain containing an apatite inclusion from sample VHC46. (c) Sample VHC53 showing a homogenized core and a sector-zoned rim. Hornblende occurs as inclusions in the sector-zoned rim. (d) Sample VHC53 showing sector zoning. (e) Sample VHC72 showing sector zoning and containing a K-feldspar inclusion. (f) Sample VHC72 showing fir-tree zoning and containing a K-feldspar inclusion. (g) Sample VHC73 showing sector zoning. (h) Sample VHC73 showing sector zoning containing inclusions of orthopyroxene, clinopyroxene, hornblende and plagioclase in planar/sector-zoned domains. (i) A homogeneous grain with a bright luminescent relict core from sample 08-10. (j) A homogeneous grain from sample 08-10. Circles with numbers are analytical spots with their identification numbers. Ages are given with errors of 1σ. Scale bars are 20 μm.
Fig. 11. SHRIMP zircon U–Pb concordia diagrams of metamorphosed mafic dykes. (a) Sample VHC46. (b) Sample VHC53. Inset shows the relationships between the $^{207}\text{Pb}/^{206}\text{Pb}$ apparent ages and Th/U ratios for the same analysed spots. (c) Sample VHC72. (d) Sample VHC73. (e) Sample VHC74. (f) Sample 08-10. Grey ellipses, oscillatory-zoned or banded cores of zircon; open ellipses, overgrowth domains of zircon; dashed ellipses, data points excluded from the calculations of the weighted mean and/or intercept ages.
ation would have been promoted by strain, and fluid flow would also have accelerated, resulting in intensive fluid–rock interaction. In this case, a more pervasive distribution of garnet nuclei might be expected, and equilibrium is more likely to have been attained between minerals, leading to the formation of relatively homogeneous metamorphic textures in the mafic dykes.

**Ages of protoliths and metamorphism**

The cores of zircon from the NW-trending (sample VHC73) and N-trending (sample VHC74) dykes have weak oscillatory zoning and/or banded structures, as observed in zircon from some gabbrics (Rubatto & Gebauer, 2000; Corfu et al., 2003). Therefore, the ages of 1764 ± 25 and 1232 ± 12 Ma obtained for these cores should represent the emplacement ages of the two dykes. These results coincide with the emplacement ages of Group I and III high-Fe mafic dykes (c. 1750 and 1240 Ma respectively) reported from other parts of the Vestfold Block (Black et al., 1991a; Lanyon et al., 1993).

Homogeneous and planar-, sector- or fir-tree-zoned zircon grains and overgrowths from most samples, except for VHC53, yielded weighted mean 207Pb/206Pb ages between 957±7 and 938±9 Ma. These zircon domains have Th/U ratios mostly <0.30, similar to those of granulite facies metamorphic zircon (Vavra et al., 1999; Möller et al., 2003). Coupled with the preservation of granulite facies mineral inclusions in these zircon domains, we interpret the age of c. 960–940 Ma as the best estimate for the timing of metamorphism of the mafic dykes. This age is also consistent with a lower intercept age of 966 ± 41 Ma obtained for zircon from a paragneiss in the same area (Clark et al., 2012). Some zircon analyses from sample VHC53 were performed on the inherited cores and have slightly higher Th/U ratios (mostly >0.30). An older mean age of 990±12 Ma obtained for this sample is consequently inferred to have been affected by the partial recrystallization of igneous zircon cores during metamorphism, and may have no geological significance.

The imprecise U–Pb lower intercept ages of c. 500 Ma may reflect the influence of a late Neoproterozoic/Cambrian metamorphic event that occurred widely in Prydy Bay. Recently, we obtained Sm–Nd mineral–whole-rock ‘isochron’ ages (reset ages?) of c. 670–590 Ma as well as 40Ar/39Ar hornblende and biotite ages of c. 530–510 Ma from the metamorphosed mafic dykes and associated mafic granulites (Liu et al., 2013). These data are in agreement with Rb–Sr mineral–whole-rock isochron ages of c. 620–500 Ma previously reported by Collerson & Sheraton (1986). This may suggest that at least the southwestern Vestfold Block did not escape the reworking of the late Neoproterozoic/Cambrian orogeny.
Implications for the evolution of the Rayner orogen

Different metamorphic conditions and P–T paths have been identified for various rocks from the Rayner orogen (Fig. 13). Metamorphism of much of the Rayner Complex, comprising late Mesoproterozoic rocks in MacRobertson Land, attained medium- to low-pressure granulite facies conditions (800–900 °C and 5–7 kbar), followed by a near-isobaric cooling (Clarke et al., 1989; Fitzsimons & Harley, 1992, 1994; Thost & Hensen, 1992; Hand et al., 1994; Scrimgeour & Hand, 1997; Stephenson & Cook, 1997; Boger & White, 2003; Halpin et al., 2007a). In contrast, metamorphism of the reworked marginal basement of the Archaean Napier Complex in Kemp Land reached higher P–T conditions (850–990 °C and 8–10 kbar), followed by near-isothermal decompression or decompressive cooling (Eills, 1983; Kelly et al., 2000; Kelly & Harley, 2004; Halpin et al., 2007b). The peak metamorphic conditions of mafic dykes in the Vestfold Block are estimated to be 820–870 °C and 8.4–9.7 kbar, which are consistent with P–T data for rocks from Kemp Land. However, metamorphic ages of 960–940 Ma in the Vestfold Block appear to be intermediate between those of MacRobertson Land (mainly c. 1000–970 Ma) and Kemp Land (c. 940–900 Ma) (Grew et al., 1988; Kinny et al., 1997; Dunkley et al., 2002; Kelly et al., 2002; Halpin et al., 2007a,b, 2012). In addition, metamorphism in the Rauer Group, 15 km southwest of the Vestfold Block, reached ultrahigh-temperature (UHT) conditions of 950–1050 °C and 9.5–12.0 kbar, followed by decompression to >7 kbar at >800–850 °C (Harley & Fitzsimons, 1991; Harley, 1998; Kelsey et al., 2003a; Tong & Wilson, 2006). However, although U(Th)–Pb zircon or monazite ages of c. 1030–820 Ma have been reported from some metapelites (Kinny, 1998; Kelsey et al., 2007; Wang et al., 2007), the results of microstructure-controlled in situ monazite chemical dating favour a late Neoproterozoic/Cambrian UHT event (Kelsey et al., 2003b). Therefore, the P–T evolution of the late Mesoproterozoic/early Neoproterozoic metamorphism in the Rauer Group needs to be investigated further. In any case, the Napier Complex, Rauer Group and Vestfold Block at the northern margin of the Rayner Complex were entirely or partially affected by the Rayner orogeny.

The occurrence of several mafic dyke swarms with ages between 2240 and 1230 Ma indicates an extensional environment for the Vestfold Block since the early Palaeoproterozoic. Pressure estimates from the igneous mineral paragenesis of the 2240 Ma mafic dykes and metamorphic mineral assemblages of sapphire-bearing granulate xenoliths trapped in the dykes suggest that the Vestfold Block basement was exhumed to within 8–11 km of the surface during the early Palaeoproterozoic (Seitz, 1994; Harley & Christy, 1995). Subsequently, the Vestfold Block resided at shallow crustal levels (10–13 km) until the middle Mesoproterozoic, as evidenced by the development of shear zones, brittle faults and pseudotachylites during emplacement of the mafic dyke swarms (Passchier et al., 1991; Hoek et al., 1992; Dirks et al., 1994). This indicates that at least the southwestern margin of the Vestfold Block underwent tectonic burial of ~20 km to lower crustal levels (~30–35 km) during the Rayner orogeny. On the other hand, although mafic dykes from the northeastern Vestfold Block appear to be undeformed and unmetamorphosed, some ductile mylonites transecting major mafic dykes have amphibolite facies minerals in their fine-grained matrix (Kuehner, 1986; Passchier et al., 1991). These metamorphic conditions have been estimated to be 550–600 °C and 6 kbar (Kuehner, 1986). If the reset zircon U–Pb age of 1025 ± 56 Ma obtained for a mafic dyke from this area (Black et al., 1991a) represents the timing of the amphibolite facies metamorphism, then progressive metamorphism from the northeast to southwest would be inferred. According to the orientation and sense of movement of the amphibolite facies mylonites, some researchers have proposed that the Rauer Group was thrust over the Vestfold Block towards the northeast during the Rayner orogeny (Passchier et al., 1991; Hoek et al., 1992). Although it remains to be confirmed whether the overthrust terrane is the Rauer Group or the Rayner Complex, this hypothesis could explain not only the burial mechanism of

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**Fig. 13.** P–T diagram showing metamorphic conditions of mafic dykes from the southwestern Vestfold Block and P–T paths estimated for different localities in the Rayner orogen (modified after Liu et al., 2013).
the relatively rigid Vestfold Block with limited internal deformation but also the progressive metamorphism towards the southwest.

The formation of the Rayner orogen is generally considered to have resulted from the collision between the Indian craton and part of east Antarctica (the Lambert Terrane or the Ruker craton including the Lambert Terrane). As is the case for the Napier Complex, the Vestfold Block is thought to be part of the Indian craton, although it preserves a different crustal history before the Neoproterozoic (Fitzsimons, 2003; Stein et al., 2004; Boger, 2011; Clark et al., 2012; Flowerdew et al., 2013). Geochronological and geochemical data indicate that much of the Rayner Complex represents a continental arc, which underwent long-lived magmatic accretion between 1380 and 1020 Ma (Sheraton et al., 1996; Stephenson, 2000; Mikhailsky et al., 2001; Liu et al., 2007, 2009b, 2014; Wang et al., 2008; Grew et al., 2012, 2013). The recent discovery of 1.33 Ga ophiolitic mélange in the Eastern Ghats Belt of India (Dharma Rao et al., 2011) suggests that this continental arc was separated from the Indian craton by an ocean basin. If this was the case, the Rauer Group would be the relict basement of the continental arc or a cratonic block involved in the Rayner orogen. Two tectonic models, involving a protracted continuous collision from 1000 to 900 Ma (Boger et al., 2000; Carson et al., 2000) or a two-stage collision involving an initial arc-continent collision during 1000–970 Ma followed by a final continent-continent collision at 940–900 Ma (Kelly et al., 2002; Halpin et al., 2007b; Liu et al., 2013, 2014), have been proposed to explain the tectonic evolution of the Rayner orogen between the Indian craton and east Antarctica. In any case, the margin of the Indian craton represented by part of the Napier Complex and Vestfold Block was buried to depths of ~30–35 km beneath the Rayner orogeny, and the depth of burial of the Vestfold Block decreases from southwest to northeast. However, the metamorphism of the Vestfold Block post-dates that of the southeastern margin of the Napier Complex by at least c. 20 Ma, which may indicate a diachronous collision between the Rayner continental arc and the Indian craton.

CONCLUSIONS

Mafic dykes from the Vestfold Block, east Antarctica, experienced heterogeneous granulite facies metamorphism characterized by spotted or fractured garnet-bearing aggregates in a garnet-absent groundmass. Peak $P$–$T$ conditions reached 820–870 °C and 8.4–9.7 kbar and are comparable with those of the southeastern margin of the Napier Complex. The heterogeneity of metamorphism was related to the lack of penetrative deformation and intensive fluid–rock interaction, as a consequence of the residence of the rigid Vestfold Block at deep crustal levels.

SHRIMP U–Pb zircon dating reveals the emplacement ages of NW- and N-trending mafic dykes to be $1764 \pm 25$ and $1232 \pm 12$ Ma, respectively, consistent with the ages of mafic dykes with the same trends from the northeastern Vestfold Block. These mafic dykes subsequently experienced high-grade metamorphism at c. 960–940 Ma, which is slightly later than the metamorphic age (mainly c. 1000–970 Ma) of the Rayner Complex, but earlier than that (c. 940–900 Ma) of the southeastern margin of the Napier Complex.

The Vestfold Block was affected by the Rayner orogeny during the late Mesoproterozoic/early Neoproterozoic. Like the southeastern margin of the Napier Complex, the southwestern margin of the Vestfold Block was buried to depths of ~30–35 km beneath the Rayner orogen during the late stage of collision between the Indian craton and east Antarctica. The different metamorphic ages of the margins of the two cratonic blocks may indicate a diachronous collision during the late continental convergence.

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SUPPORTING INFORMATION

Additional Supporting Information may be found in the online version of this article at the publisher’s web site:

**Table S1.** Representative microprobe analyses of garnet.

**Table S2.** Representative microprobe analyses of clinopyroxene.

**Table S3.** Representative microprobe analyses of orthopyroxene.

**Table S4.** Representative microprobe analyses of hornblende.

**Table S5.** Representative microprobe analyses of biotite.

**Table S6.** Representative microprobe analyses of plagioclase.

**Table S7.** Representative microprobe analyses of K-feldspar.

**Table S8.** SHRIMP U–Pb zircon analyses for metamorphosed mafic dykes.

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