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Geochemistry and tectonic setting of mafic rocks in western Dronning Maud Land, East Antarctica: implications for the geodynamic evolution of the Proterozoic Maud Belt

EUGENE G. GROSCH[1], AVINASH BISNATH[1], HARTWIG E. FRIMMEL[1,2] & WARWICK S. BOARD[1]
[1]*Department of Geological Sciences, University of Cape Town, Rondebosch 7701, South Africa*
(e-mail: egrosch@geology.uct.ac.za)
[2]*Present address: Institute of Mineralogy, University of Würzburg, Am Hubland, D-97074 Würzburg, Germany*

Abstract: On the basis of new bulk major and trace element (including REE) as well as Sm–Nd and Rb–Sr isotope data, used in conjunction with available geochronological data, a post-tectonic mafic igneous province and four groups of pre- to syntectonic amphibolite are distinguished in the polymetamorphic Maud Belt of western Dronning Maud Land, East Antarctica. Protoliths of the Group 1 amphibolites are interpreted as volcanic arc mafic intrusions with Archaean to Palaeoproterozoic Nd model ages and depletion in Nb and Ta. Isotopic and lithogeochemical characteristics of this earliest group of amphibolite indicate that the Maud Belt was once an active continental volcanic arc. The most likely position of this arc, for which a late Mesoproterozoic age (*c.* 1140 Ma) is indicated by available U–Pb single-zircon age data, was on the southeastern margin of the Kaapvaal–Grunehogna Craton. The protoliths of Group 2 amphibolites are attributed to the 1110 Ma Borgmassivet–Umkondo thermal event on the basis of comparable Nd model ages and trace element distributions. Group 3 amphibolite protoliths are characterized by mid-ocean ridge basalt-type REE patterns and low Th/Yb ratios, and they are related to Neoproterozoic extension. Group 4 amphibolite protoliths are distinguished by high Dy/Yb ratios and are attributed to a phase of syntectonic Pan-African magmatism as indicated by Rb–Sr isotope data.

The high-grade metamorphic Maud Belt of western Dronning Maud Land, East Antarctica, has been interpreted as a juvenile island arc that was tectonically juxtaposed along the margin of an Archaean cratonic block, known as the Grunehogna Craton, at the end of the Mesoproterozoic (Arndt *et al*. 1991; Jacobs *et al*. 1993; Grantham *et al*. 1995; Groenewald *et al*. 1995; Golynsky & Jacobs 2001; Bauer *et al*. 2003*b*; Paulsson & Austrheim 2003; Basson *et al*. 2004). In contrast to the high-grade ortho- and paragneisses of the Maud Belt, a low-grade volcano-sedimentary Mesoproterozoic succession (Ritscherflya Supergroup) overlies Archaean basement on the Grunehogna Craton (Fig. 1). A major geophysical boundary, known as the Pencksökket–Jutulstraumen Discontinuity, separates the Grunehogna Craton from the Maud Belt and has been interpreted as a Mesozoic continental rift with an orientation that was structurally controlled by major 'Grenvillian' (about 1.1 Ga) and/or Pan-African (about 550 Ma) thust shear zones (Ferraccioli *et al*. 2005; Fig. 1). Extensive mafic sills of the 1107 Ma Borgmassivet Suite (Krynauw *et al*. 1991; M. Knoper, unpubl. data reported by Frimmel 2004) intruded the volcano-sedimentary rocks of the Ritscherflya Supergroup, whereas numerous mafic bodies occur as pre-, syn- and post-tectonic amphibolite dykes, sills and boudins in the Maud Belt. Geochemical data for these mafic rocks are scarce in the literature and largely confined to samples from the southwestern-most part of the belt (Bauer *et al*. 2003*a*). To better assess the likely tectonic setting of the various stages of magmatism recorded in the Proterozoic to Cambrian rocks of Dronning Maud Land, new major and trace element (including REE) as well as Sr and Nd isotopic data for a range of metabasites of different age from across the Maud Belt and the adjacent Ritscherflya Basin were collected in this study.

General agreement exists on the Grunehogna Craton having formed part of the Kaapvaal Craton prior to the break-up of Gondwana. Its pre-Gondwana geodynamic evolution remains, however, highly speculative. Three tectonic models have been suggested for the evolution of the southern margin of the Kaapvaal–Grunehogna Craton, including western Dronning Maud Land: (1) closure of a complex 'Tugela Ocean' with one or more intra-oceanic arcs and generally southward-directed subduction zones (Arima *et al*. 2001); (2) a southward-directed subduction, followed by a northward-directed subduction in western Dronning Maud Land (Bauer *et al*. 2003*b*); (3) two southward-directed subduction zones, with the northern one separating the Grunehogna Province from the Kaapvaal Craton (Basson *et al*. 2004). All of these models consider the Maud Belt to represent a continuation of the late Mesoproterozoic Namaqua–Natal Belt of southern Africa into East Antarctica and they involve deposition of at least the lower parts of the Ritscherflya Supergroup in a peripheral forearc cratonic basin in response to accretion and continental collision of an island arc (Maud Belt) to the margin of the Grunehogna Craton (e.g. Groenewald *et al*. 1995; Moyes & Harris 1996; Basson *et al*. 2004). The polarity of subduction is, in most cases, assumed to be directed away from the craton toward allegedly juvenile island arc/s that form the Namaqua–Natal and Maud Belts (e.g. Jacobs *et al*. 1993; Grantham *et al*. 1995; Groenewald *et al*. 1995; Golynsky & Jacobs 2001; Basson *et al*. 2004). Reconstruction of a late Mesoproterozoic *P*–*T*–*t* path for the Maud Belt and direct comparison with that recorded in the Namaqua–Natal Belt was recently shown to be complicated because of major tectonic reworking during Pan-African times at *c*. 540 Ma (Board *et al*. 2005). Consequently, the various mafic bodies that occur as deformed dykes, sills and/or boudins must predate or be syntectonic with the Pan-African orogeny and thus provide

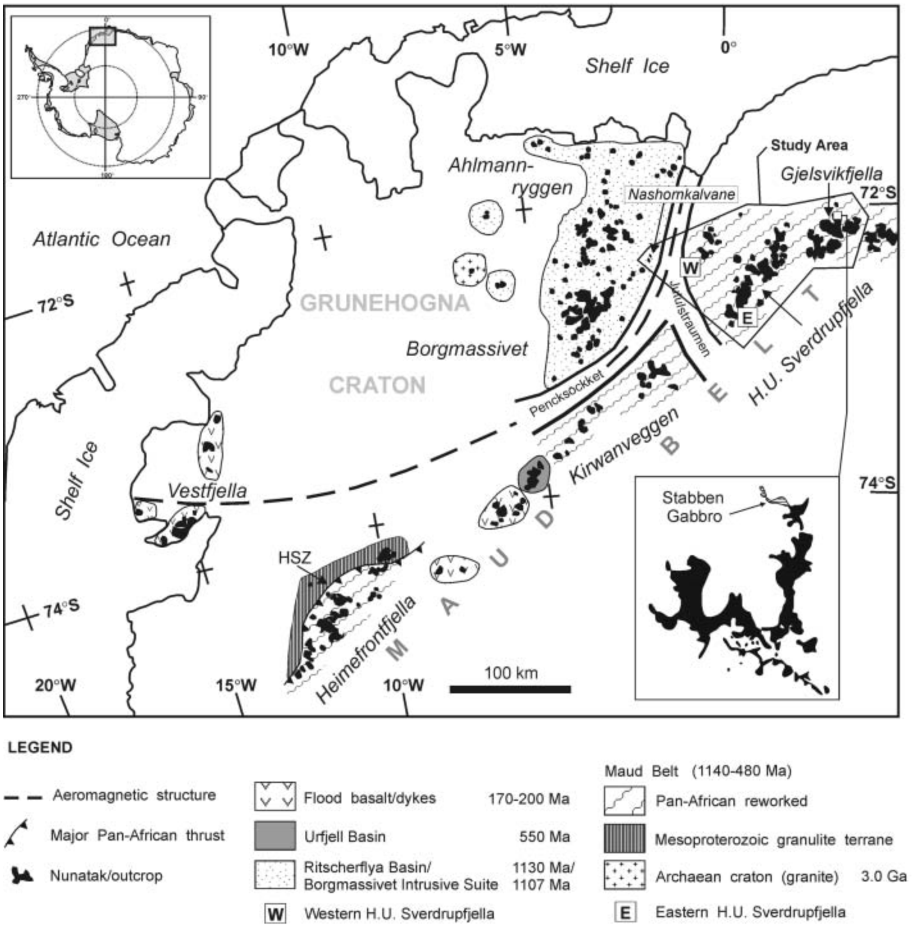

Fig. 1. Tectonic units of Western Dronning Maud Land (after Board *et al.* 2005).

potential information on the late Meso- to Neoproterozoic history of the Maud Belt.

Two geochemically distinct generations of amphibolite have so far been distinguished in the southernmost part of the Maud Belt and they yielded U–Pb single-zircon ages of 1033 ± 7 Ma and 586 ± 7 Ma (Bauer *et al.* 2003*a*). The geochemistry of these dated late Mesoproterozoic and Neoproterozoic mafic rocks serves as a reference for the variety of pre- to post-tectonic mafic rocks across the Maud Belt for which no radiometric age control is available. A further reference for lithogeochemical and isotopic comparison is given by the voluminous, effectively unmetamorphosed, 1107 Ma mafic sills of the Borgmassivet Suite on the adjacent Grunehogna Craton. Given that the early metamorphic and tectonic evolution of the Maud Belt is largely unknown because of major Pan-African tectonic overprinting, the trace element and isotope geochemistry of the mafic protoliths is used here in an attempt to constrain their tectonic setting. It will be shown that a number of generations of mafic protoliths can be distinguished. After assessing the extent to which the Maud Belt represents juvenile, late Mesoproterozoic crust of an oceanic v. continental volcanic arc, a new geodynamic model for the late Mesoproterozoic evolution of western Dronning Maud Land will be proposed.

Geological setting and history

The Grunehogna Craton is generally interpreted to represent an Archaean fragment of the southeastern Kalahari (amalgamated Kaapvaal–Zimbabwe) Craton that became tectonically disconnected during the break-up of Gondwana (e.g. Ferreira 1988). U–Pb zircon data from a Borgmassivet Suite sill that intrudes the Ritscherflya Supergroup rocks indicates emplacement at *c.* 1107 Ma (M. Knoper, unpubl. data cited by Frimmel 2004). A U–Pb age of 1130 ± 7 Ma, derived from single-zircon grains in tuff beds from the Ahlmannryggen Group representing the lowermost Ritscherflya Supergroup (Frimmel 2004), provided the best constraint so far on the timing of sedimentation. The tectonic setting for these sediments remains problematic. A U–Pb zircon age of 1109.6 ± 0.6 Ma from an Umkondo dolerite sill on the eastern margin of the Archaean Zimbabwe Craton (Hanson *et al.* 1998, 2004) is identical to that of the Borgmassivet Suite. Magmatism at around that time has also been documented about 1200 km SW of the Grunehogna Craton in the southern Coats Land, where zircon grains from rhyolite and granophyre yielded U–Pb ages of 1112 ± 4 Ma and 1106 ± 3 Ma, respectively (Gose *et al.* 1997). The Umkondo and Borgmassivet Suites have been interpreted as products of a large-scale intra-plate magmatic event that took place between 1112 and 1106 Ma that, although short-lived, affected both the Kalahari (including the Grunehogna) and Laurentian Cratons during the assembly of an inferred late Mesoproterozoic supercontinent (Hanson *et al.* 1998, 2004) and might have extended as far as west–central Australia, where the *c.* 1100 Ma Warakurna large igneous province (Wingate *et al.* 2004) may well be an equivalent.

Deciphering the geological history of the Maud Belt has been important for the reconstruction of pre-Gondwana palaeogeography. Traditionally, the Maud Belt has been considered part of a larger Grenville-age mobile belt around the margin of East Antarctica, known as the Circum East Antarctic Mobile Belt (e.g. Tingey 1991; Yoshida 1992; Jacobs *et al*. 1993; Gose *et al*. 1997; Golynsky & Jacobs 2001; Jacobs *et al*. 2003*a*), which continues into the Namaqua–Natal Belt of southern Africa, with early high-grade metamorphism recorded between 1090 and 1030 Ma. The Maud Belt is exposed in a number of areas from SW to NE, namely the Heimefrontfjella, Kirwanveggen, H. U. Sverdrupfjella, Gjelsvikfjella and central Dronning Maud Land (Fig. 1). Available U–Pb zircon protolith ages of the oldest volcanic and plutonic rocks of this proposed island arc of between 1160 and 1130 Ma (Arndt *et al*. 1991; Harris *et al*. 1995; Jackson 1999; Jacobs *et al*. 1999; Paulsson & Austrheim 2003; Board *et al*. 2005) are in good agreement with the U–Pb single-zircon ages of 1130 ± 7 Ma from tuff layers in the Ahlmannryggen Group, pointing to a close spatial relationship between the cratonic depositional basin of the Ritscherflya Supergroup and the Maud Belt volcanic arc.

The tectonothermal evolution of the Maud Belt is complex. The extent of Pan-African tectonic overprinting in the belt has been a contentious issue, with some workers having suggested a mainly thermal overprint (Grantham *et al*. 1995; Groenewald *et al*. 1995; Jackson 1999) and others a tectonothermal overprint (Jacobs *et al*. 1995, 1998; Board *et al*. 2005). In particular, U–Pb ages of *c*. 540 Ma from syntectonic monazite inclusions within metamorphic minerals that define a penetrative top-to-the-NW shear fabric in the H. U. Sverdrupfjella provided strong support for a tectonothermal Pan-African overprint (Board *et al*. 2005). Similar to P–T and U–Pb age data of Board *et al*. (2005), two pulses of high-grade Pan-African tectonism have been reported from the Gjelsvikfjella (Bisnath *et al*. 2006) and central Dronning Maud Land (Jacobs *et al*. 2003*b*) further to the NE between *c*. 580–560 Ma and 530–510 Ma. In central Dronning Maud Land, the second pulse was associated with a phase of syn- to post-tectonic magmatism during orogenic collapse (Jacobs *et al*. 2003*b*; Bisnath & Frimmel 2005). The extent of Pan-African tectonothermal overprinting of Mesoproterozoic reworked volcanic arc crust apparently increases towards the NE (e.g. Jacobs *et al*. 1998; Jacobs & Thomas 2002; Paulsson & Austrheim 2003), which has led to the more widespread occurrence of post-tectonic, 510–480 Ma mafic and felsic intrusions in the northeastern areas of the belt (Harris *et al*. 1995; Mikhalsky *et al*. 1997; Jacobs *et al*. 2003*b*; Paulsson & Austrheim 2003; Bisnath *et al*. 2006).

Results

A total of 80 amphibolitic samples were analysed for major and trace element concentrations by conventional X-ray fluorescence and inductively coupled plasma mass spectrometry techniques at the Department of Geological Sciences, University of Cape Town. All the lithogeochemical data are available online at http://www.geolsoc.org.uk/SUP18261. A hard copy can be obtained from the Society Library. Analytical details have been given by Frimmel *et al*. (2001). Rb–Sr and Sm–Nd isotope ratios (Table 1) were measured on a VG Sector seven-collector mass spectrometer in multi-dynamic mode, following standard chemical separation techniques described by le Roex & Lanyon (1998). The model ages are discussed according to DePaolo (1981). Decay constants for ^{87}Rb and ^{147}Sm of respectively 1.42×10^{-11} and 6.54×10^{-12} were used (Steiger & Jaeger 1977; Lugmair & Marti 1978). Calculation of Nd model ages and ϵ_{Nd} values are based on $^{143}Nd/^{144}Nd = 0.512638$ and $^{147}Sm/^{144}Nd = 0.1967$ for the chondritic uniform reservoir (CHUR, Jacobsen & Wasserburg 1980), and $^{143}Nd/^{144}Nd = 0.513114$ and $^{147}Sm/^{144}Nd = 0.222$ for depleted mantle composition (Michard *et al*. 1985).

The geochemical data presented in this study include those for a Borgmassivet sill on the eastern margin of the Grunehogna Craton at Nashornkalvane South, and amphibolites from the eastern and western parts of the belt in the H. U. Sverdrupfjella and from the Gjelsvikfjella (Fig. 1). Geochemical analyses of undeformed mafic dykes and the Stabben gabbro in Gjelsvikfjella (Fig. 1), dated at 487 ± 4 Ma (Bisnath *et al*. 2006), are also included. The geochemistry of the amphibolites from the western and eastern H. U. Sverdrupfjella and the Gjelsvikfjella areas is compared with that of the 1107 Ma Borgmassivet sills (this study and Krynauw 1986) and the contemporaneous Umkondo Suite dolerites (Hanson *et al*. 1998, 2004; Munyanyiwa 1999) as well as the two dated groups of amphibolite dykes from the Heimefrontfjella (Bauer *et al*. 2003*a*).

Significance of metasomatism

Before discussing lithogeochemical data for the high-grade metamorphic rocks in terms of protolith composition, a critical assessment of element mobility during metamorphism is required. As large-ion lithophile elements (LILE) are generally considered to be more mobile (non-conservative) elements (e.g. Pearce & Peate 1995), the Cs and Rb concentrations of the sample suite were used to assess potential LILE mobility related to metasomatic alteration. Several samples are distinguished by elevated Cs and Rb concentrations as well as by variable deviation from an otherwise linear relationship between Cs and Rb (labelled samples in Fig. 2a). These samples are likely to have experienced LILE enrichment in the course of their metamorphic history, which is also evident from partial biotitization of these amphibolites after the peak of metamorphism (Bisnath & Frimmel 2005; Board *et al*. 2005) and they are not considered any further in the subsequent discussion on tectonic setting.

The Nb and Zr concentrations of the sample suite were used to assess potential mobility of trace elements typically considered to be less mobile (i.e. conservative elements; e.g. Pearce & Peate 1995) and less sensitive to metamorphic or other post-crystallization alteration. As expected for a suite of unaltered mafic rocks, a good positive correlation between Nb and Zr concentrations is indicated by the majority of the analysed samples (Fig. 2b). Three samples (AB-31A, AB-41A and ABA-63), however, which also correspond to those described above as having highly elevated Cs and Rb concentrations, have high Nb concentrations and plot above the overall trend in Figure 2b, from which a disturbance of the original conservative element ratios is inferred. The same applies to two amphibolite samples (WB-109, WB-34) from the H. U. Sverdrupfjella (Fig. 2b).

The former sample (WB-109) is a highly foliated amphibolite from a high-strain shear zone. It experienced multiple stages of tectonic reactivation and metamorphism and thus may serve as a reference for the open-system behaviour of conservative elements. Not surprisingly, it displays the largest deviation in Nb concentration from the Nb–Zr trend in Figure 2b. Unlike the majority of amphibolite samples in the H. U. Sverdrupfjella, which have moderately enriched light REE (LREE) patterns, this sample is characterized by a LREE-depleted, slightly concave-upward REE pattern (Fig. 3a). By analogy with a number of studies that have illustrated LREE mobilization as a result of external fluid infiltration (e.g. Grauch 1989), it is suspected that

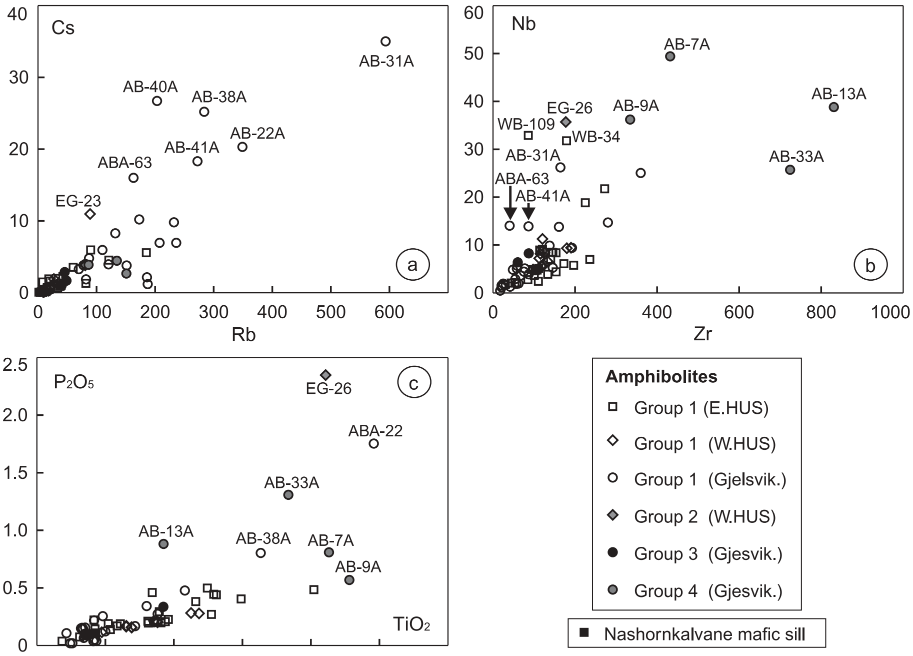

Fig. 2. Binary plots used to illustrate the effects of metasomatism on mafic protolith chemistry as a result of metamorphism or other alteration processes. (**a**) Cs v. Rb. (**b**) Nb v. Zr. (**c**) P_2O_5 v. TiO_2. W.HUS, western H. U. Sverdrupfjella; E.HUS, eastern H. U. Sverdrupfjella; Gjelsvik., Gjelsvikfjella.

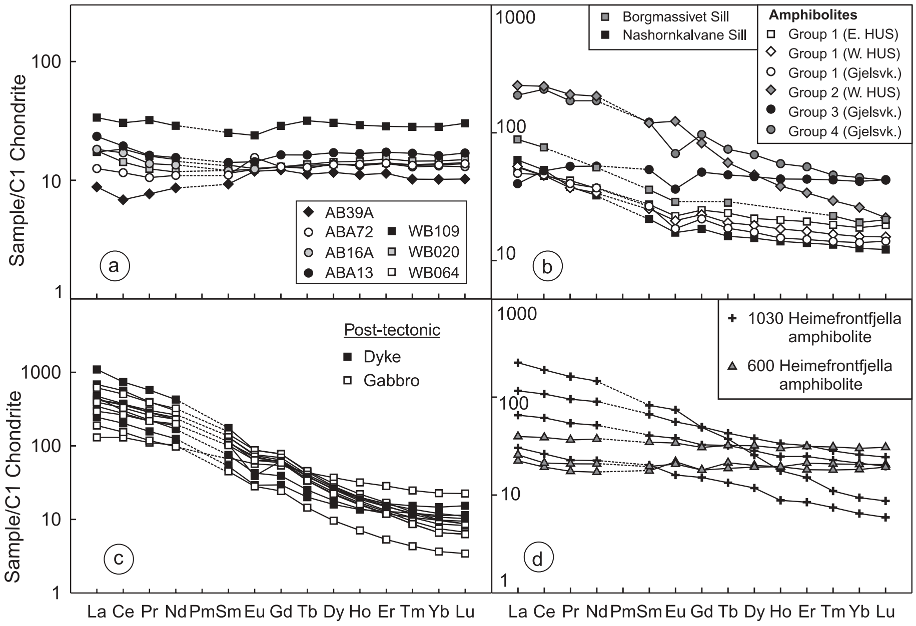

Fig. 3. (**a**) Amphibolites showing concave-upward, LREE-depleted patterns as a result of metamorphic alteration. (**b**) REE patterns of representative amphibolite samples from each amphibolite group identified in the Maud Belt. For abbreviations see Figure 2. Also shown is the REE pattern for the Nashornkalvane mafic sill on the easternmost margin of the craton, which has a REE pattern very similar to that of the 1107 Ma Borgmassivet Suite sills (REE data from Krynauw 1986). (**c**) REE patterns of post-tectonic mafic intrusions. (**d**) Heimefrontfjella 1030 Ma and 600 Ma amphibolite groups (data from Bauer *et al.* 2003a). All samples normalized to the C1 chondrite (Sun & McDonough 1989).

this sheared amphibolite experienced metasomatic modification of its original LREE and other trace element budget during metamorphic alteration. Similarly, the amphibolite sample AB-39A, which displays a pronounced concave-upward, LREE-depleted pattern (Fig. 3a), is suspected to have undergone metasomatic alteration. A few other samples also display this distinctive REE pattern (Fig. 3a) and they are all excluded from the tectonic analyses below.

Amphibolite samples WB-34 and ABA-37 are distinguished by negative Ce and Eu anomalies in their REE patterns. These anomalies could reflect pre-metamorphic basalt–seawater interaction (e.g. Ludden & Thompson 1979).

Major and trace element geochemistry

Notwithstanding some element mobility during metamorphism, and excluding all those samples for which notable metasomatic alteration is suspected (see above), an attempt is made to constrain, as far as possible, the geochemical source characteristics of the mafic rocks using only the relatively immobile elements and isotope data.

The Nashornkalvane sill data plot well within the compositional range of the studied amphibolites (Fig. 2a–c). Four amphibolite samples from the Gjelsvikfjella have high Zr concentrations (sample AB-9A, 334 ppm; AB-7A, 432 ppm; AB-33A, 724 ppm; AB-13A, 831 ppm) and plot off the general trend defined by the other samples (Fig. 2b). These four samples also have elevated concentrations of P_2O_5 (between 0.57 and 1.31 wt%) and Nb (between 25.7 and 49.4 ppm; Fig. 2b and c). The sample (EG-26) with the highest P_2O_5 content (2.35 wt%) also has highly elevated TiO_2 (4.21 wt%) and Nb (35.7 ppm) concentrations despite a Zr concentration that is well within the typical range for most of the amphibolites (Fig. 2c). Samples ABA-2 and ABA-62 are distinguished by high Y concentrations of 69.5 ppm and 82.4 ppm, respectively.

Using the classification criteria of Winchester & Floyd (1977), the majority of the analysed samples have Zr/TiO_2 and Nb/Y ratios that correspond to compositions of typical basalts including subalkaline basalts. In a normal mid-ocean ridge basalt (N-MORB)-normalized multi-element diagram, the Nashornkalvane sill has the chemical signature of a typical Borgmassivet Suite sill, with enrichment in LILE, negative Nb–Ta anomaly and positive Ce anomaly (Fig. 4). The majority of amphibolite samples from the Maud Belt, apart from the high-Zr, high-Y and high P–Ti amphibolites described further below and a few other samples with less prominent Nb–Ta anomalies or variably low Rb, Ba and Th contents, have a very similar MORB-normalized profile to the Borgmassivet Suite sills (Fig. 4).

The chondrite-normalized REE pattern of the Nashornkalvane sill displays a LREE enrichment similar to that of the Borgmassivet Suite mafic sills (Fig. 3b), with $(La/Yb)_N$ ratios between 5.20 and 4.66 and weak negative Eu anomalies ($Eu/Eu^* = 0.81$–0.85). More than one type of REE pattern is identified in the amphibolite sample suite (Fig. 3b). The majority of amphibolites in the Maud Belt have REE patterns similar to those of the Borgmassivet Suite sills, but LREE enrichment is less pronounced. The $(La/Yb)_N$ ratios of the amphibolite samples overlap with those of the Borgmassivet Suite sills, ranging between 1.67 and 4.79 with similar total REE abundances and weak or absent Eu anomalies ($Eu/Eu^* = 0.85$–0.99; Fig. 3b). Using all the REE data for the sample suite, the results for the Nashornkalvane sill and the various amphibolite types are presented in a plot of La/Yb v. Dy/Yb (Fig. 5). Two amphibolite samples (AB-11A and AB-12A) have similar Dy/Yb ratios but noticeably higher La/Yb ratios compared with those of the Nashornkalvane sill. These two samples are characterized by small positive Eu anomalies and probably reflect accumulation of plagioclase in cumulate sills, as indicated by a high modal proportion of plagioclase (70–85 vol%) in these samples. The high P–Ti (EG-26) amphibolite shows the highest enrichment in both LREE and heavy REE (HREE), whereas the two samples with elevated Y concentrations (ABA-2 and ABA-62) have flat REE patterns and are distinguished as having the lowest La/Yb ratios of 1.34 and 1.08, respectively (Fig. 5). The four high-Zr samples show overall REE enrichment and are distinguished by high Dy/Yb ratios ranging between 2.26 and 2.40 (Fig. 5).

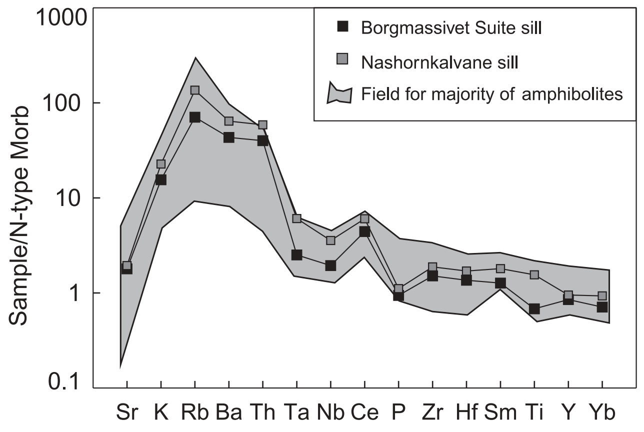

Fig. 4. N-type MORB-normalized multi-element variation diagrams (after Pearce 1983) for the Nashornkalvane sill and the Maud Belt amphibolites (shown as grey field) in comparison with data available for the Borgmassivet Suite mafic sills.

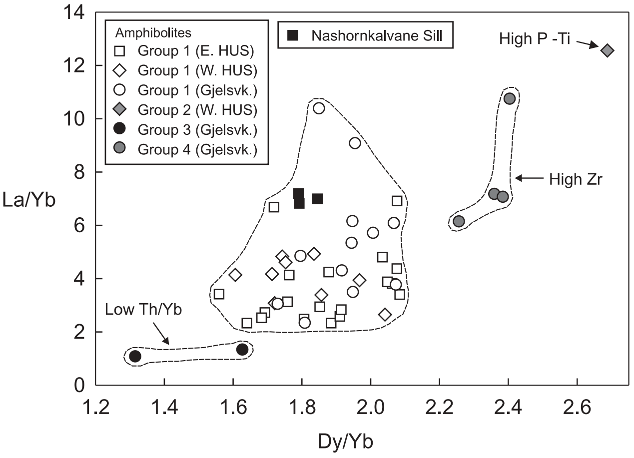

Fig. 5. La/Yb v. Dy/Yb plot showing differences in LREE and HREE chemistry in the Maud Belt amphibolites. Various geochemically distinct groups of amphibolite can be distinguished. Symbols and abbreviations used for amphibolite groups are as in Figure 3.

The best examples of post-orogenic mafic rocks come from Gjelsvikfjella, where undeformed mafic dykes are abundant, and they include what has been mapped as the Stabben ‘gabbro’ (Fig. 1). The lithogeochemistry of the latter (see the Supplementary Publication) does not conform to a typical basaltic melt composition, thus precluding a meaningful comparison with the other mafic rocks studied. A very similar LREE-enriched pattern (high average La/Yb ratio of around 70) to that of the post-tectonic mafic dykes is noted, however (Fig. 3c).

Sr and Nd isotope data

A total of 17 samples with the most primitive composition, representative of each group of mafic rocks, were selected for isotope analyses (Table 1). The age-corrected initial $^{87}Sr/^{86}Sr$ ratio obtained for the 1107 Ma Borgmassivet sill is 0.7117 (using an average of three samples). Assuming a similar age for the mafic protolith emplacement of the amphibolites in the western H. U. Sverdrupfjella, initial $^{87}Sr/^{86}Sr$ ratios between 0.70923 and 0.71603 are calculated for these rocks. For a number of amphibolite samples from the Gjelsvikfjella, including those with high Dy/Yb and low La/Yb, initial $^{87}Sr/^{86}Sr$ ratios are unreasonably low at an assumed age of 1110 Ma (i.e. less than chondritic values). Consequently, their emplacement ages must be younger. The maximum age required to obtain realistic initial $^{87}Sr/^{86}Sr$ ratios for the amphibolite samples with high Dy/Yb as well as for those that have trace element characteristics similar to the Nashornkalvane sill (samples AB-15A and AB-43A) is about 530 Ma. This age corresponds to the final stage of synorogenic high-grade metamorphism in the Gjelsvikfjella and to a phase of syntectonic magmatism further east (Jacobs *et al.* 2003*b*). The corresponding initial $^{87}Sr/^{86}Sr$ ratios ($\geqslant 0.7031$) are similar to or

Table 1. *Sm–Nd and Rb–Sr isotopic data for mafic rocks in western Dronning Maud Land*

Sample	Sm (ppm)	Nd (ppm)	$^{143}Nd/^{144}Nd$	Error (2σ)	$^{147}Sm/^{144}Nd$	$^{143}Nd/^{144}Nd$ (i)	ϵNd (t)	T_{DM} (Ma)	Rb (ppm)	Sr (ppm)	$^{87}Sr/^{86}Sr$	Error (2σ)	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$ (i)
*Group 1 (1110 Ma)**														
EG-032	3.52	14.34	0.51172	0.0009	0.14825	0.51064	−11.1	2866	21.26	203.55	0.71969	0.0017	0.30268	0.71488
EG-030	3.71	15.75	0.51173	0.0021	0.14212	0.51069	−10.1	2634	21.24	159.18	0.71538	0.0016	0.38661	0.70923
EG-021b	5.74	21.28	0.51208	0.0009	0.16291	0.51090	−6.0	2643	35.38	129.78	0.72859	0.0017	0.79072	0.71603
EG-022	3.92	16.46	0.51150	0.0010	0.14381	0.51045	−14.7	3124	27.18	175.36	0.72303	0.0019	0.44938	0.71589
*Group 2 (1110 Ma)**														
EG-N1	3.35	15.86	0.51182	0.0009	0.12774	0.51089	−6.2	2090	39.56	165.40	0.72265	0.0029	0.69346	0.71163
EG-N2	3.24	15.09	0.51181	0.0008	0.12967	0.51087	−6.6	2144	37.74	157.16	0.72315	0.0017	0.69627	0.71209
EG-N3	3.10	14.48	0.51182	0.0009	0.12929	0.51088	−6.4	2119	38.66	157.39	0.72271	0.0025	0.71212	0.71140
EG-026	18.24	90.28	0.51175	0.0009	0.12202	0.51087	−6.6	2066	80.68	753.63	0.71755	0.0020	0.31022	0.71262
*Group 3 (600 Ma)**														
ABA/2	7.88	25.60	0.51260	0.0009	0.18593	0.51187	0.0	2177	48.60	60.30	0.73740	0.0019	2.33995	0.71738
ABA/62	7.21	23.00	0.51255	0.0011	0.18935	0.51180	−1.2	2631	45.50	82.40	0.71902	0.0019	1.60027	0.70533
ABA/63	3.53	14.00	0.51240	0.0025	0.15229	0.51240	−1.2	1556						
*Group 4 (530 Ma)**														
AB43A	6.81	29.10	0.51241	0.0009	0.14135	0.51192	−0.6	1323	232.03	246.38	0.72857	0.0016	2.73182	0.70793
AB-15A	3.37	13.50	0.51234	0.0011	0.15077	0.51182	−2.6	1643	187.31	47.49	0.80796	0.0018	11.52990	0.72086
AB-33A	18.50	83.10	0.51226	0.0010	0.13446	0.51179	−3.2	1491	86.31	404.41	0.71045	0.0017	0.61799	0.70578
AB-7A	13.00	58.90	0.51253	0.0010	0.13331	0.51206	2.1	1010	134.58	294.00	0.71373	0.0019	1.32592	0.70371
AB-9A	10.60	47.60	0.51253	0.0009	0.13451	0.51206	2.0	1023	21.29	347.63	0.70446	0.0017	0.17723	0.70312
AB-22A	16.20	60.80	0.51240	0.0022	0.16093	0.51184	−2.2	1773	349.00	94.40	0.78783	0.0019	10.78630	0.70634
AB-38A	10.40	44.80	0.51235	0.0009	0.14021	0.51187	−1.8	1418	283.75	225.31	0.73291	0.0019	3.65471	0.70530
*Post-tectonic (490 Ma)**														
ABA-65	15.60	92.80	0.51173	0.0010	0.10152	0.51141	−11.7	1746	13.00	2252	0.70818	0.0018	0.01671	0.70807
ABA-58	17.20	113.00	0.51164	0.0011	0.09192	0.51135	−12.9	1723	64.00	2607	0.70849	0.0017	0.07107	0.70799

*Assumed age.

lower than those for the post-tectonic Stabben 'gabbro' (0.7080), but are within the range of continental mafic rocks.

The initial $^{143}Nd/^{144}Nd$ ratios for the various mafic rocks were used in the calculation of ϵ_{Nd} values and Nd model ages (Table 1) to distinguish between possibly different mantle source regions and/or different generations. As both the Nashornkalvane sill and the majority of amphibolites in the Maud Belt have similar trace element characteristics, their Sm–Nd model ages are compared to assess their source characteristics and whether they were derived from similar mantle sources.

Nd model ages (T_{DM}) and ϵ_{Nd} for the mafic rocks in Western Dronning Maud Land show a wide spread (Fig. 6; Table 1), which argues against melt derivation from the same source. Strongly negative ϵ_{Nd} values between −14.7 and −6.0 were obtained for the Nashornkalvane sill and the western H. U. Sverdrupfjella amphibolites, using an initial $^{143}Nd/^{144}Nd$ ratio at 1110 and 1130 Ma, which approximates to the time of emplacement of the Borgmassivet mafic sills and the formation of the volcanic arc. The Nd model ages for three samples from the Nashornkalvane sill on the Grunehogna Craton are between 2090 and 2144 Ma. With the exception of the high Ti–P amphibolite (EG-26) with a similar model age of 2066 Ma, the remaining amphibolites from the western H. U. Sverdrupfjella, which have trace element signatures representative of the majority of amphibolite in the Maud Belt, have distinctly older Nd model ages of between 2634 and 2134 Ma.

The two amphibolite samples with low La/Yb ratios have very similar flat, MORB-type REE patterns and low Th/Yb ratios (see Figs 5 and 7). Thus they strongly resemble a group of amphibolites in the Heimefrontfjella for which an age of *c.* 600 Ma has been determined (Bauer *et al.* 2003*a*; see Fig. 3b and d). A calculated ϵ_{Nd} (600 Ma) value for such an amphibolite with MORB characteristics (ABA-2) is 0.0. Its T_{DM} (600 Ma) is with 2127 Ma comparable to the range obtained for Group 2. The other amphibolite with an analogous REE pattern has a similar ϵ_{Nd} (600 Ma) of −1.2, but an even older T_{DM} (600 Ma)

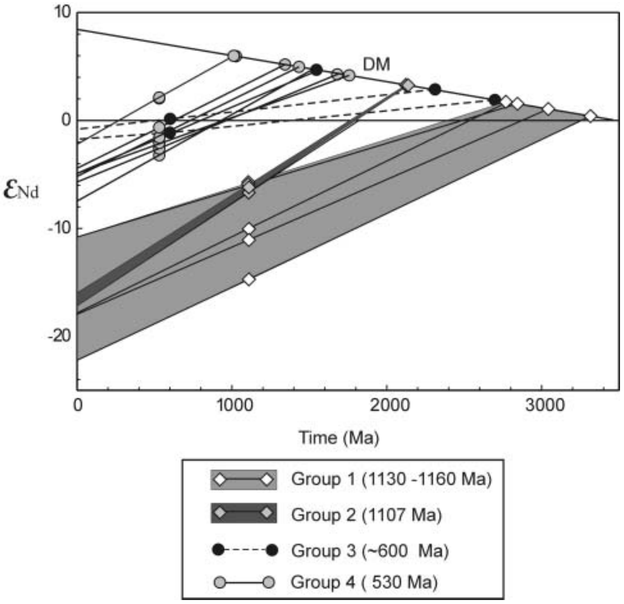

Fig. 6. Sm–Nd evolution diagram for pre- to syntectonic mafic rocks in Western Dronning Maud Land. (Group 2 is samples EG-N1 and EG-26.) Symbols and abbreviations used for amphibolite groups are as in Figure 5. DM, depleted mantle.

age of 2631 Ma comparable to Group 1. Reasonable initial ^{87}Sr/^{86}Sr ratios ($\geqslant$0.70533) are also obtained when an age of 600 Ma is assumed for these two amphibolite samples. A further group of syn-tectonic amphibolites is distinguished by lower T_{DM} ages between 1010 and 1773 Ma, with ϵ_{Nd} values between -3.2 and 2.0. (Fig. 6). The post-tectonic Stabben 'gabbro' has a T_{DM} (490 Ma) of around 1740 Ma but much lower ϵ_{Nd} (-12.0).

Four generations of amphibolite

Based on the trace element and isotope data presented here, at least four geochemically distinct varieties of pre-Pan-African amphibolites can be distinguished in the Maud Belt. Each records a stage of mafic magmatism that occurred between the late Mesoproterozoic and the Cambrian. The main distinguishing chemical characteristics of the amphibolite groups are outlined below.

Group 1

Group 1 represents the majority of amphibolites in the Maud Belt. It is characterized by moderate LREE enrichment and a MORB-normalized multi-element pattern indicating LILE enrichment, a negative Nb–Ta anomaly and a positive Ce anomaly. Four samples from the western H. U. Sverdrupfjella within this group were found to have Palaeoproterozoic to Archaean Nd model ages (2634–3124 Ma) and a wide spread in overall negative ϵ_{Nd} (1110 Ma) values (-6.0 to -14.7). This spread points to a heterogeneous sublithospheric mantle source, as can be found below Archaean cratons that record a complex growth and stabilization history (e.g. Pearson 1999), rather than depleted mantle source regions below juvenile Mesoproterozoic island arc crust. This postulated Archaean craton is probably the Kaapvaal–Grunehogna Craton.

The interpretation of the Maud Belt as an ancient continental arc is in agreement with the trace element and isotopic geochemical signature of the Group 1 amphibolites. Their MORB-normalized diagrams show a negative Nb–Ta anomaly and a positive Ce anomaly typical of subduction-related volcanic arc rocks. In the Zr/Y v. Zr diagram of Pearce & Norry (1979), a within-plate signature is indicated for most of these amphibolites, although a trend towards the MORB field is noticeable (Fig. 8). It should be noted that basalts of continental arcs were subsequently shown to overlap with the within-plate basalt field in this diagram (Pearce 1983). Their chemical distribution in this diagram is very similar to that of the Borgmassivet and Umkondo Suite mafic sills, which also plot in the within-plate field (Fig. 8). In the plot of Th (non-conservative element) v. Nb (conservative element) normalized to Yb (Pearce & Peate 1995), these amphibolites are displaced above the composition of the mantle array, with almost all samples plotting in the continental arc field (Fig. 7). An active continental volcanic arc origin for these mafic rocks is also indicated by the Ti–Zr–Y diagram of Pearce & Cann (1973) and the Nb–Zr–Y tectonic diagram of Meschede (1986), in which most samples plot within the volcanic arc + within-plate and the calc-alkaline fields, respectively, with trends towards the more depleted MORB fields (not shown). The Group 1 amphibolites are therefore interpreted to represent the oldest group of amphibolite with their protoliths emplaced during mafic magmatism associated with continental volcanic arc formation, now represented by reworked crust of the Maud Belt.

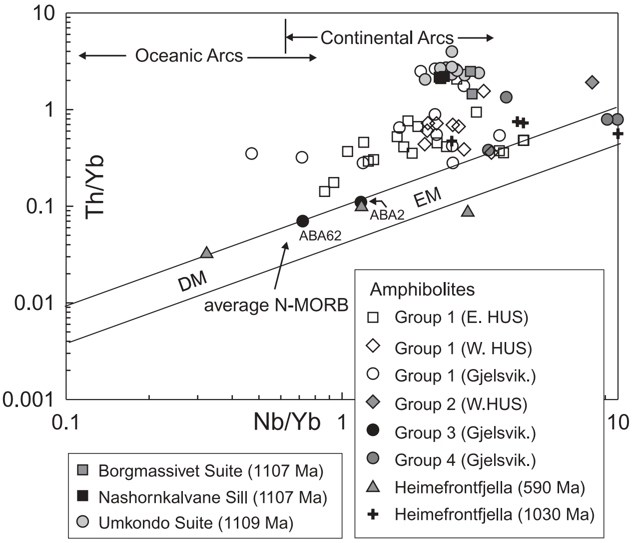

Fig. 7. Th/Yb v. Nb/Yb plot for the mafic rocks in Western Dronning Maud Land. EM, enriched mantle; DM, depleted mantle (after Pearce 1983; Pearce & Peate 1995). Symbols and abbreviations used for amphibolite groups are as in Figure 3.

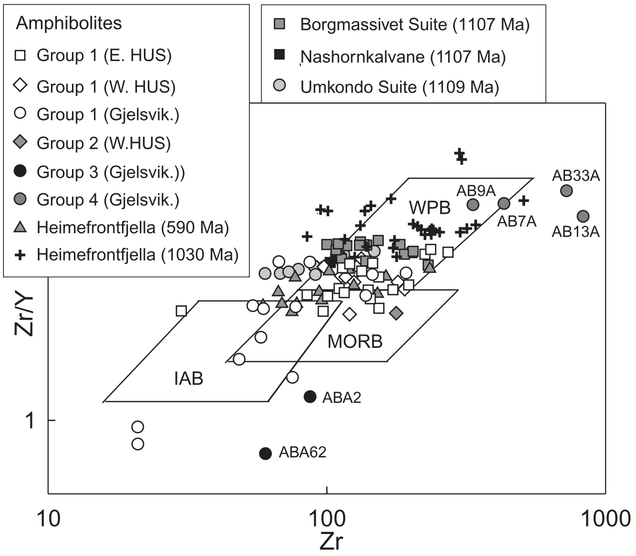

Fig. 8. Zr/Y v. Zr diagram (after Pearce & Norry 1979) for the mafic rocks from Western Dronning Maud Land. Geochemical and geochronological data for the Heimefrontfjella 1030 Ma and 600 Ma amphibolite groups from Bauer *et al*. (2003*a*); geochemical and geochronological data for Umkondo mafic sills from Munyanyiwa (1999) and Hanson *et al*. (2004), respectively. Geochemical data for the Borgmassivet Suite sills after Krynauw (1986); age data for the Borgmassivet Suite were reported by Frimmel (2004). All other geochemical data for various amphibolite (western H. U. Sverdrupfjella, eastern Sverdrupfjella, Gjelsvikfjella) and the Nashornkalvane sill are from this study. Symbols and abbreviations used for amphibolite groups are as in Figure 3.

Group 2

Group 2 is represented by a high P–Ti amphibolite in the westernmost part of the Maud Belt that has a very similar Sm–Nd model age (*c*. 2066 Ma) and ϵ_{Nd} (1110 Ma) value (-6.6) to that of the Borgmassivet Suite Nashornkalvane mafic sill on the eastern margin of the craton.

The similarity in ϵ_{Nd} values for the high P–Ti amphibolite (−6.6) and the Nashornkalvane Borgmassivet Suite sill (−6.2 to −6.6) and very similar Nd model ages point to a possible genetic relationship to the Borgmassivet–Umkondo thermal event. The strongly negative epsilon values overlap only with the uppermost values in the range for the Group 1 amphibolites and also indicate the involvement of an ancient source, such as the subcontinental lithospheric mantle below the Kaapvaal–Grunehogna Craton.

Considering the possible connection of the Borgmassivet–Umkondo thermal event with a *c*. 1107 Ma mantle plume (Frimmel 2004; Hanson *et al.* 2004), this group of mafic rocks may have been the result of impingement of such a plume on the base of a sublithospheric mantle that had been previously metasomatized by subduction-related processes a few tens of million years prior to melting. The presence of the high P–Ti amphibolite with a possibly similar mantle source to the Borgmassivet Suite sill could indicate that many of the amphibolite protoliths in the Maud Belt with subduction-zone geochemical signatures may have been the result of that mantle plume. The subduction-related trace element enrichment of the Nashornkalvane sill in this group as well as many of these plume-related amphibolites is, however, not distinguishable from the composition of Group 1.

Group 3

Group 3 amphibolites from the Gjelsvikfjella are distinguished by having flat, MORB-type REE patterns with the lowest La/Yb ratios. They also have the lowest Th/Yb ratios and the highest Y concentrations. An ϵ_{Nd} (600 Ma) value of 0.0 supports extraction of the protolith mafic melt from a MORB mantle source with CHUR-type characteristics at around that time.

The geochemistry of this group of amphibolites is similar to that of dated Neoproterozoic amphibolites in the Heimefrontfjella in the southernmost part of the Maud Belt (Fig. 3d). By analogy with the dated example from Heimefrontfjella (Bauer *et al.* 2003*a*), Group 3 amphibolites probably represent oceanic basalts and mafic dykes that were emplaced into the Maud Belt during opening of a Neoproterozoic basin, whereby the mafic melts tapped a heterogeneous mantle during thinning of the crust and subcontinental lithospheric mantle.

Group 4

Group 4 amphibolites from the Gjelsvikfjella are distinguished by the highest Zr/Y and Dy/Yb ratios. These amphibolites are also characterized by elevated Nb, P_2O_5 and TiO_2 concentrations. Reasonable initial $^{87}Sr/^{86}Sr$ ratios for these amphibolites are obtained only when a maximum age of 530 Ma is assumed. Their Nd model ages (Table 1) support reworking of heterogeneous Mesoproterozoic crust.

An age of 530 Ma indicated by the Rb–Sr isotope data as the protolith emplacement age for this group of amphibolites corresponds to a time of high-grade Pan-African tectonism in the Gjelsvikfjella and to a phase of syntectonic mafic magmatism further east in the Maud Belt (Jacobs *et al.* 2003*b*). In the Zr/Y v. Zr diagram of Pearce & Norry (1979) their Zr/Y ratios are similar to those for a heterogeneous group of 1030 Ma amphibolites in the Heimefrontfjella (Fig. 8), which have been explained by post-orogenic collapse (Bauer *et al.* 2003*b*). Their high Zr, Nb and P_2O_5 concentrations point toward the mafic protoliths probably being derived from highly evolved magmas that were sourced from within the sublithospheric mantle.

The 490 Ma post-tectonic Stabben ‘gabbro’ has a similar but somewhat higher Sm–Nd model age of *c*. 1740 Ma in comparison with the high-Zr amphibolite group. Following the tectonometamorphic studies by Paulsson & Austrheim (2003), Engvik & Elvevold (2004) and Bisnath & Frimmel (2005), the Group 4 amphibolites and the post-tectonic gabbro and mafic dykes probably represent a late- to post-tectonic mafic igneous province that is related to post-orogenic collapse caused by the intrusion of asthenospheric material following the delamination of lithospheric mantle.

Geodynamic evolution

In the current tectonic models suggested for the Maud Belt (Arndt *et al*. 1991; Jacobs *et al*. 1993; Grantham *et al*. 1995; Groenewald *et al*. 1995; Golynsky & Jacobs 2001; Bauer *et al*. 2003*a*; Paulsson & Austrheim 2003; Basson *et al*. 2004) a rotation in the principal compressional tectonic stress by more than 90° around the south and southeastern margin of the Kaapvaal Craton is required during the accretion of the Namaqua–Natal–Maud Belt island arc following the closure of the proposed Tugela Ocean. Contiguity of a volcanic arc stretching around the southern Kaapvaal–Grunehogna Craton from the Namaqua sector to the Maud Belt is, however, not supported by more recent, precise U–Pb zircon age data from the Namaqua–Natal and Maud Belts (Frimmel 2004). These data indicate that arc formation in the Maud Belt postdated that in the Namaqua–Natal Belt by several tens to 100 Ma.

Basson *et al*. (2004) have pointed out that the southern Mfongosi Valley sediments in the Natal Belt have an active continental margin geochemical signature similar to the sediments of the Ahlmannryggen Group but distinct from the northern Mfongosi sediments for which an oceanic island arc source is indicated. The age of these sediments in the Natal Belt is, however, not well constrained.

Limited Sm–Nd isotope data on felsic rocks from the Gjelsvikfjella, including a homogeneous granodioritic, arc-related migmatite with a protolith age of *c*. 1160 Ma and aplitic dykes dated at *c*. 500 Ma, yielded Proterozoic Nd model ages of 1390–1770 Ma and 1800–2220, respectively, which are older than volcanic arc formation (Paulsson & Austrheim 2003). The preferred model presented by Paulsson & Austrheim (2003) follows that of the above workers; that is, interpreting the Maud Belt as a continuation of the Natal island arc. Paulsson & Austrheim have argued that the old (2220 Ma) Nd model ages probably reflect mixing of melts derived from Proterozoic island arc crust with another source from an unknown craton. Sm–Nd isotope data for various felsic, mafic and metasedimentary samples from the Heimefrontfjella and Mannefallknausane in the southern parts of the belt were found to scatter widely about a 1.1 Ga reference line and the corresponding ϵ_{Nd} values (assuming an age of 1.1 Ga) range between +4.4 and −3.7, indicating juvenile late Mesoproterozoic protoliths (Arndt *et al*. 1991). The ϵ_{Nd} values were found to decrease from north to south in the Heimefrontfjella and were interpreted to indicate late Mesoproterozoic mixing between mantle-derived mafic magmas with older crustal material from an inferred older continent to the south of the Heimefrontfjella (Arndt *et al*. 1991).

Results obtained in this study show that the northern Maud volcanic arc was not oceanic but formed in close proximity to an Archaean Craton, probably the Kalahari–Grunehogna Craton. A close spatial relation between the on-craton, arc-derived, volcano-sedimentary Ritscherflya Supergroup on the Grunehogna Craton and an active volcanic arc is supported by the good agreement in U–Pb ages between the Ahlmannryggen Group and the oldest plutonic, arc-related rocks identified in the Maud Belt. Such a relationship would imply westward-directed subduction below the craton margin, in which case the basin for the Ritscherflya Supergroup would take a back-arc position.

Discrepancies in the available tectonic models can be resolved if the Maud Belt is regarded as a former continental volcanic arc that formed in an independent and younger tectonic regime compared with island arc magmatism in the Natal Belt. By the time inboard subduction beneath the eastern flank of the Kaapvaal–Grunehogna Craton took place, the island arc(s) of the Natal Belt would have already been obducted to the southern margin of the craton. In our preferred model, subduction beneath the craton margin induced back-arc spreading, resulting in the Ritscherflya Supergroup being deposited in a cratonic back-arc basin with material derived from the craton and the Maud volcanic arc (Fig. 9). Superimposed on the later stages of back-arc basin development was extension during to the Borgmassivet–Umkondo thermal event at *c*. 1107 Ma and emplacement of the Borgmassivet Suite sills into the Ritscherflya basin and also into the adjacent volcanic arc. Early U–Pb metamorphic zircon overgrowth ages at 1104 ± 5 Ma (Arndt *et al*. 1991) and 1112 ± 29 Ma (Bauer *et al*. 2003*b*) in the Heimefrontfjella, 1104 ± 8.3 Ma in the Gjelsvikfjella (Bisnath *et al*. 2006), and leucosome development dated at 1098 ± 5 Ma in the Kirwanveggen, probably record this event as high-grade thermal metamorphism throughout the Maud Belt.

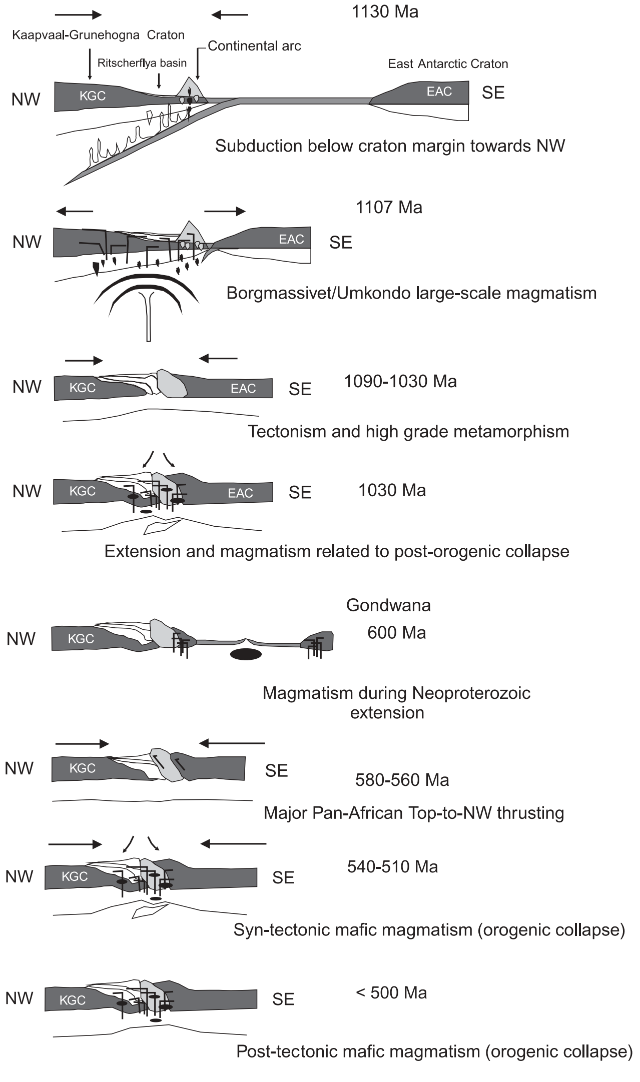

Fig. 9. Schematic geodynamic evolution of Western Dronning Maud Land.

Following arc formation, thermal metamorphism and back-arc basin extension, continental collision of the Kaapvaal Craton with an unconstrained ('East Antarctic') craton is recorded in zircon overgrowths and zircon in anatectic leucosome throughout the Maud Belt with U–Pb ages of between 1030 and 1060 Ma. Post-orogenic collapse following crustal thickening during this continental collision was accompanied by intrusion of mafic dykes at *c*. 1030 Ma, so far recorded only from the Heimefrontfjella (Bauer *et al*. 2003*a*).

Mafic dykes, *c*. 600 Ma in age, were emplaced in the southwestern area of the belt (Bauer *et al*. 2003*a*). Comparable dykes are now also indicated for the northeastern area of the belt. The timing and MORB-type geochemical character of this magmatic event suggests that it was probably related to the formation of oceanic basalts and mafic dykes during opening of a Neoproterozoic basin, possibly the extension of the Mozambique Ocean in the north and/or the oceanic basin that is recorded in the Shackleton Range in the south (Talarico *et al*. 1999). A period of major Pan-African tectonism followed in the form of two tectonic pulses, between *c*. 570–560 Ma and 540–530 Ma during the amalgamation of Gondwana (e.g. Board *et al*. 2005). The second phase of tectonism was associated with syntectonic mafic magmatism, whereas further post-tectonic mafic magmatism is indicated for an orogenic collapse stage between 500 and 480 Ma.

The new isotope data do not support the existence of a major late Mesoproterozoic crustal suture or boundary between a former Maud Belt island arc and the Grunehogna Craton. Instead, that boundary could simply represent the rear end of a Mesoproterozoic continental volcanic arc, and the major Maud–Grunehogna crustal discontinuity indicated in various aerogeophysical studies (e.g. Golynsky *et al*. 1997; Ferraccioli *et al*. 2005) is rather regarded as a major thrust or fault that formed during Pan-African orogeny. It cannot be ruled out, however, that Pan-African collisional to transpressional tectonic forces reactivated an older continental back-arc fault there.

Conclusions

A combination of trace element distribution and isotope data for metabasites from across the Maud Belt and adjacent Grunehogna Craton has revealed a series of protolith generations, which provide important information regarding the geodynamic evolution of western Dronning Maud Land from as early as the late Mesoproterozoic. The variety of mafic rocks identified in this study include four groups of pre- to syntectonic amphibolite. The petrogenesis of mafic protoliths to the Group 1 amphibolite is related to arc formation between 1160 and 1130 Ma. The Group 2 amphibolite records the Umkondo thermal event, which was responsible not only for intrusion of the Borgmassivet Suite mafic sills, but also for widespread and short-lived mafic magmatism in and on the Kalahari–Grunehogna Craton (possible stretching as far as southeastern Laurentia and west–central Australia) at around 1110 Ma. Amphibolite of Group 3 reflects oceanic basalts and/or dolerite dykes related to Neoproterozoic extension at around 600 Ma. Further mafic magmatism is indicated for a post-orogenic collapse phase following Pan-

African continental collision, possibly as a result of the amalgamation of East and West Gondwana.

Sm–Nd isotope data for the earliest mafic group reveal a strong involvement of Archaean continental crust in the formation of a late Mesoproterozoic volcanic arc, which provided the bulk of the protoliths in the Maud Belt, and indicate a melt source in a heterogeneous subcontinental mantle lithosphere. Although previous models all indicate outboard subduction with one or more subduction zones directed away from the craton margin, subduction below the craton towards the west or NW leading to growth, stabilization and metasomatism of the sublithospheric mantle is favoured in this study, with the sediments of the Ritscherflya Supergroup having been deposited in a back-arc basin.

This study was carried out on samples collected during three field campaigns to Western Dronning Maud Land within the South African National Antarctic Programme (SANAP) between 1998 and 2002. D. Reid, A. Späth and S. Govender assisted with XRF, ICPMS, and Sr and Nd isotope analyses, respectively. We thank W. Bauer and an anonymous reviewer for constructive comments on an earlier version of the manuscript. Financial support by SANAP and the South African National Research Foundation (grant to HEF) is gratefully acknowledged.

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