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Mantle plumes and Antarctica–New Zealand rifting: evidence from mid-Cretaceous mafic dykes

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Abstract: Ocean floor magnetic anomalies show that New Zealand was the last continental fragment to separate from Antarctica during Gondwana break-up, drifting from Marie Byrd Land, West Antarctica, about 84 Ma ago. Prior to continental drift, a voluminous suite of mafic dykes (dated by Ar–Ar laser stepped heating at 107 ± 5 Ma) and anorogenic silicic rocks, including syenites and peralkaline granitoids (95–102 Ma), were emplaced in Marie Byrd Land during a rifting event. The mafic dyke suite includes both high- and low-Ti basalts. Trace element and Sr and Nd isotope compositions of the mafic dykes may be modelled by mixing between tholeiitic OIB (asthenosphere-derived) and alkaline high- to low-Ti alkaline magmas (lithospheric mantle derived). Pb isotopes indicate that the OIB component had a HIMU composition.

We suggest that the rift-related magmatism was generated in the vicinity of a mantle plume. The plume helped to control the position of continental separation within the very wide region of continental extension that developed when the Pacific–Phoenix spreading ridge approached the subduction zone. Separation of New Zealand from Antarctica occurred when the Pacific–Phoenix spreading centre propagated into the Antarctic continent. Sea floor spreading in the region of the mantle plume may have caused an outburst of volcanism along the spreading ridge generating an oceanic plateau, now represented by the 10–15 km thick Hikurangi Plateau situated alongside the Chatham Rise, New Zealand. The plateau consists of tholeiitic OIB–MORB basalt, regarded as Cretaceous in age, and similar in composition to the putative tholeiitic end-member in the Marie Byrd Land dykes. The mantle plume is proposed to now underlie the western Ross Sea, centred beneath Mount Erebus, where it was largely responsible for the very voluminous, intraplate, alkaline McMurdo Volcanic Group. A second mantle plume beneath Marie Byrd Land formed the Cenozoic alkaline volcanic province.

Keywords: Antarctica, Gondwana, continental drift, geochemistry, magmatism.

The formation of new ocean basins by continental break-up is often accompanied by large amounts of magmatic activity as a result of decompression melting of unusually hot mantle as it rises beneath rifted margins. These 'volcanic margins' are understood to have formed above thermal anomalies, or plumes, in the mantle (White & McKenzie 1989) and contrast with non-volcanic rifted margins formed in regions where normal mantle temperatures occur and consequently the volume of magma produced is relatively low. Volcanic margins are characterised by large volumes of flood basalt, either exposed on the continental edge or interpreted from geophysical techniques to lie offshore, and by igneous rock underplated beneath or intruded as dyke swarms or igneous complexes into continental crust.

The separation of New Zealand from the Antarctic core of Gondwana in Late Cretaceous times was the last of a series of break-up events that occurred during fragmentation of the supercontinent (Storey 1995 for review). Continental separation, in most of these events, was closely associated with extrusion of flood basalt provinces that are generally linked to mantle plumes (White & McKenzie 1989; Hawkesworth *et al.* 1992). New Zealand–Antarctic rifting was associated also with magmatic activity now represented by exposed mafic dyke complexes and anorogenic granitoids along the Marie Byrd Land (Fig. 1) margin of Antarctica (Weaver *et al.* 1994) and some small igneous complexes within New Zealand (Weaver & Pankhurst 1991; Baker *et al.* 1994). Weaver *et al.* (1994) concluded that rifting was caused by changes in plate boundary forces and suggested that mantle plume activity may have begun in mid-Cretaceous times (*c.* 100 Ma) coincident with rifting, triggering melting of the lithosphere and controlling the locus of rifting. Most other authors favour impact of a much younger plume (*c.* 30 Ma) beneath Marie Byrd Land to account only for the extensive Cenozoic alkaline basalt province in Marie Byrd Land (LeMasurier & Rex 1989; Behrendt *et al.* 1991, 1992, 1996; Hole & LeMasurier 1994).

The aim of this paper is to consider all available evidence for the existence of mantle plumes beneath the Marie Byrd Land margin, and the role (active or passive) that a possible plume may have played at the time of New Zealand rifting. We present new geochemical and isotopic data from mafic dyke complexes exposed along the Marie Byrd Land margin (sampled by the South Pacific Rim International Tectonics Expedition, SPRITE; DiVenere *et al.* 1993). We argue that two separate mantle plumes were present beneath Marie Byrd Land; the first became established in mid-Cretaceous times, was responsible for these dyke complexes, and is now present beneath Mount Erebus, a large active volcano on Ross Island (Fig. 2), whereas the second is the Cenozoic plume that became

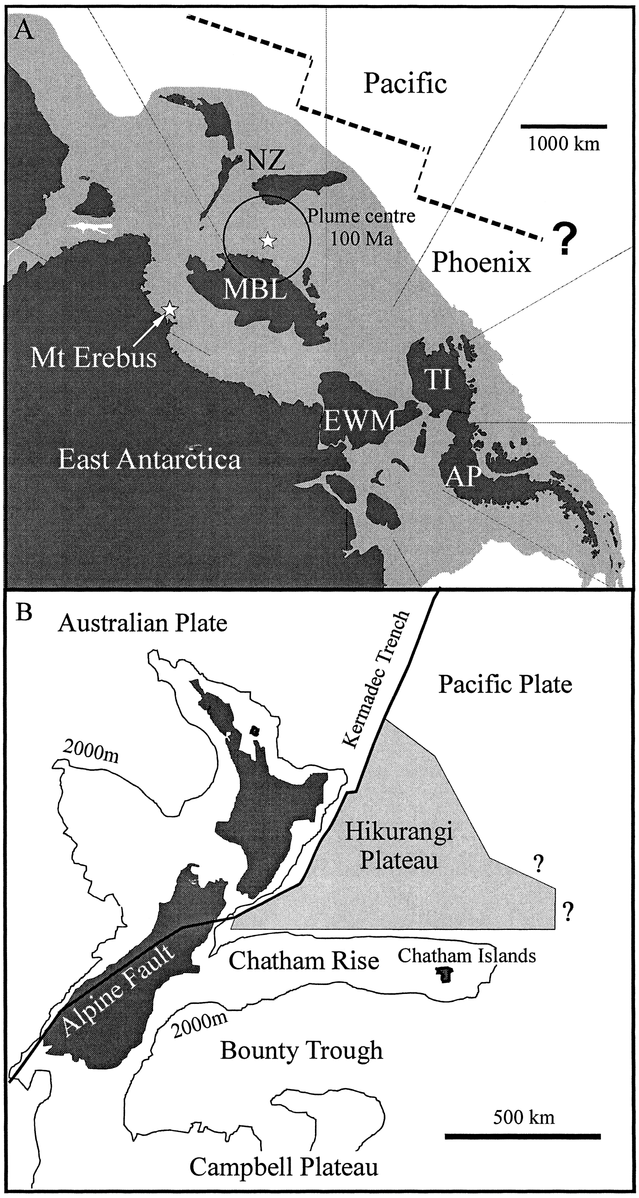


Fig. 1. (**a**) Mid-Cretaceous Gondwana reconstruction showing a New Zealand block (NZ) attached to the Marie Byrd Land (MBL) sector of Antarctica. The position of spreading ridges between the Pacific and Phoenix plates are shown together with the possible position of a mid-Cretaceous mantle plume based on the hot spot reference frame of Mueller *et al.* (1993). We postulate that the plume is now beneath Mount Erebus. AP, Antarctic Peninsula; EWM, Ellsworth–Whitmore mountains; TI, Thurston Island. (**b**) Sketch map of New Zealand showing the position of the Hikurangi Plateau, a large submarine basalt province.

established *c.* 30 Ma and is currently still beneath Marie Byrd Land.

Regional tectonic and magmatic setting

Cretaceous subduction

Prior to New Zealand–Antarctic rifting, the Phoenix plate was being subducted beneath the palaeo-Pacific margin of New Zealand (Fig. 1) with formation of wide accretionary

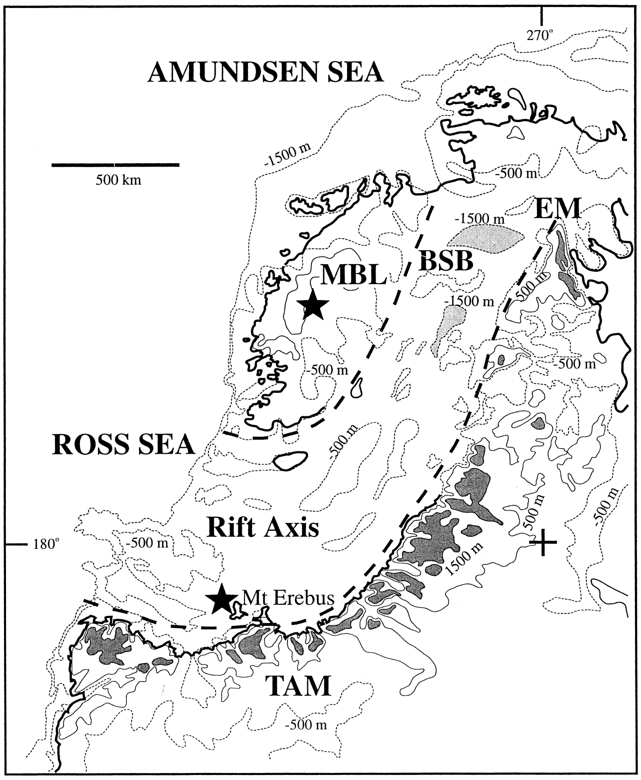


Fig. 2. Sketch map of West Antarctic Rift System showing sub-ice topography and the postulated position (stars) of present day hotspots beneath Mount Erebus and Marie Byrd Land (MBL). The contour interval is 1000 m with land above 1500 m shown with a dark shading and land below 1500 m shown with a lighter shading. BSB, Byrd Subglacial basin; EM, Ellsworth Mountains; TAM, Transantarctic Mountains.

complexes, now exposed in New Zealand, and magmatic arc rocks present in both New Zealand and Marie Byrd Land (Bradshaw 1989; Weaver *et al.* 1994). In Marie Byrd Land a widespread group of Cretaceous (124–95 Ma) calc-alkalic, I-type granodiorite plutons were intruded into basement rocks. In the western sector, the plutons are of Early Cretaceous age (124–108 Ma) whereas they extend to slightly younger ages in eastern Marie Byrd Land. This age pattern suggests that subduction ceased first along the western sector at *c.* 108 Ma and persisted until about 95 Ma in the east.

Subduction ceased as a result either of collision of the Pacific–Phoenix spreading ridge with the subduction zone (Bradshaw 1989) or of abandonment of the spreading ridge just prior to collision followed by transfer of the overriding plate (New Zealand) to the Pacific plate by a process known as subducted slab capture (Luyendyk 1995). In either case, the plate boundary forces changed dramatically, and because the Pacific plate was diverging from the Antarctic plate, a broad extensional province, known as the West Antarctic rift system, developed across West Antarctica (Fig. 2; Behrendt *et al.* 1991, 1996; Lawver & Gahagan 1994). During this initial rifting phase anorogenic magmas were emplaced and Marie Byrd Land was displaced *c.* 300 km away from the Transantarctic Mountains (DiVenere *et al.* 1995) in a dextral transtensional sense (Storey 1991; Wilson 1995).

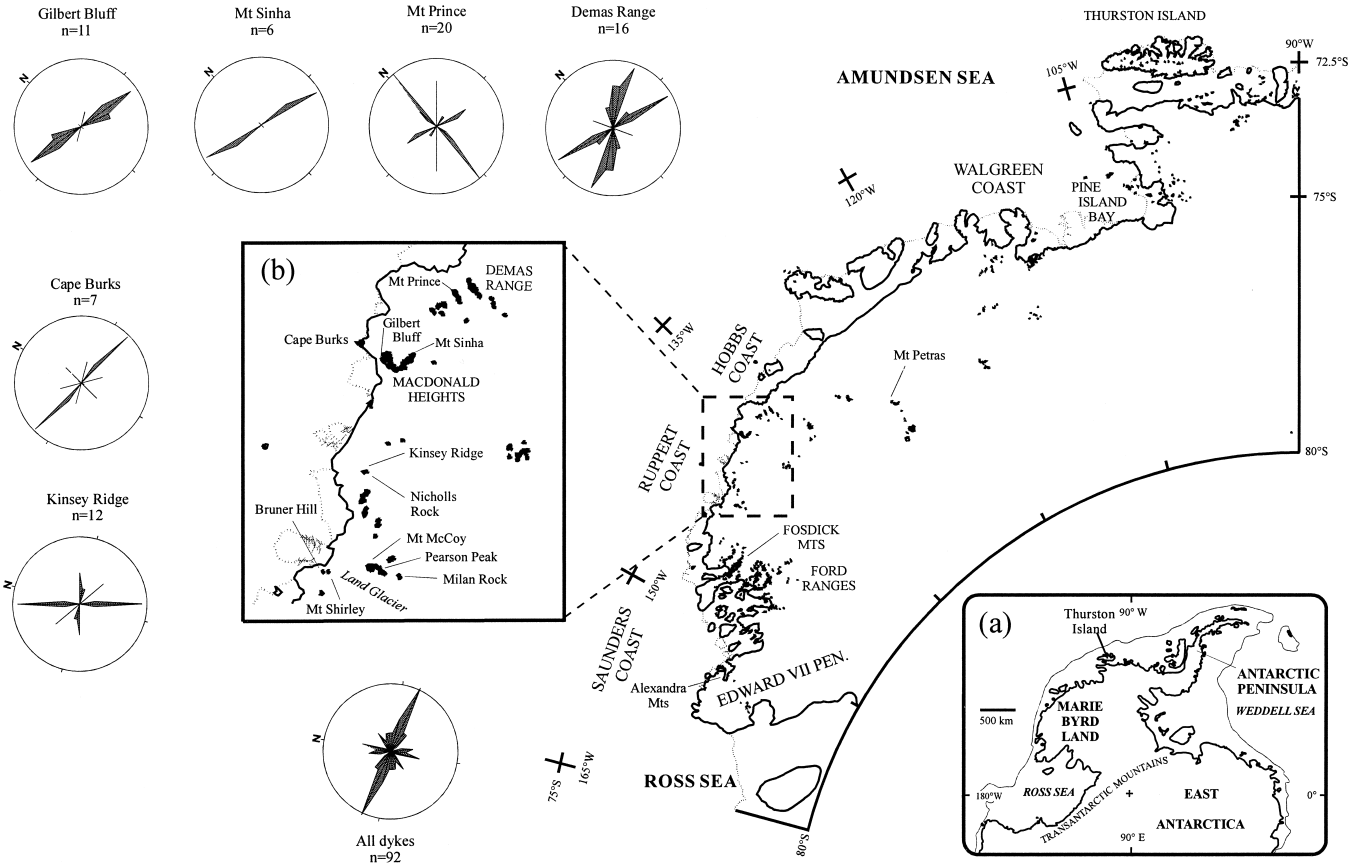


Fig. 3. Sketch maps of Marie Byrd Land (after Pankhurst *et al.* 1997) showing key localities and dyke orientations for different localities mentioned in text. Inset (**a**) shows Marie Byrd Land in relation to the rest of West Antarctica. Inset (**b**) shows a detailed part of Marie Byrd Land.

Cretaceous anorogenic magmatism and metamorphism

In western Marie Byrd Land, a diverse suite of silicic rocks, generally A-type granitoids (emplaced at 95–102 Ma), and including syenites and peralkaline granitoids, and mafic rocks, dominantly dykes, described in detail below, followed the calc-alkaline granodiorites, along a 500 km long sector of the Saunders–Ruppert–Hobbs coasts (Fig. 3). On the Ruppert–Hobbs coasts, the A-type granitoids have initial $^{87}Sr/^{86}Sr$ ratios of 0.7041–0.7045 and εNd_t values of 0 to +4 (Weaver *et al.* 1994), and probably were mantle-derived, whereas on the Edward VII Peninsula, Saunders Coast, A-type syenogranites have initial $^{87}Sr/^{86}Sr$ ratios of 0.7116–0.7206 and εNd_t values of −5 to −6, and probably represent melts of Palaeozoic basement (Weaver *et al.* 1992; Adams *et al.* 1995). Palaeozoic basement rocks together with mid-Cretaceous anorogenic granitoids were metamorphosed to high grades, starting at 105 Ma, and deformed at mid-crustal levels followed by rapid cooling and uplift to form Cretaceous metamorphic complexes (Richard *et al.* 1994). Mineral assemblages record rapid cooling from peak metamorphic conditions of 725–780°C at 4.3–5.6 kbar to below 165°C between 105 and 94 Ma.

A number of small igneous complexes characterized by alkaline gabbro and related rocks are present also in NE South Island, New Zealand. These include alkali gabbro-syenite of the Mandamus complex (Weaver & Pankhurst 1991), the layered ultramafic–mafic Tapuaenuku Igneous Complex (Baker *et al.* 1994), the Blue Mountain complex (Grapes 1975) and numerous smaller mafic intrusions. Alkaline volcanic rocks also overlie deformed sedimentary rocks on the Chatham Islands close to the conjugate Marie Byrd Land margin (Weaver & Smith 1989).

An igneous province that may be of relevance to New Zealand–Marie Byrd Land rifting is the Hikurangi Plateau, an oceanic plateau situated to the north of the Chatham Rise (Fig. 1). It covers some 350 000 km^2, has a crustal thickness of 10–15 km, and is believed to be underlain by basaltic crust (Davey & Wood 1994). The plateau has been sampled at one dredge site, which yielded OIB-MORB-like tholeiitic basalts (Mortimer & Parkinson 1996). Its age is unknown, but it is probably Cretaceous, and is interpreted as a region of thickened oceanic crust related to a mantle plume (Davey & Wood 1994; Mortimer & Parkinson 1996). In view of this interpretation and the potential closeness of the Hikurangi Plateau to the Marie Byrd Land margin during the mid-Cretaceous (Fig. 1), we consider the possibility that the plateau may have formed as a result of excess melt production caused by the influence of a mid-Cretaceous plume at the Marie Byrd Land margin.

Cretaceous rifting; formation of West Antarctic rift system

Marie Byrd Land forms the northern flank of what has been become widely known as the West Antarctic rift system (Fig. 2; LeMasurier 1990; Behrendt *et al.* 1991, 1996; LeMasurier & Rex 1991; Tessensohn & Wörner 1991), which developed at the same time as separation of New Zealand from Marie Byrd Land. The largely aseismic and asymmetric rift system extends over a 3000 × 750 km, largely ice-covered area from the Ross Sea to the Bellingshausen Sea, comparable in area to the Basin and Range and the East African rift systems. Its southern flank is a spectacular 4 to 5 km high, rift shoulder escarpment extending along the Transantarctic Mountains from northern Victoria Land diverging to the Ellsworth Mountains. Its northern flank, i.e. Marie Byrd Land, is topographically more subdued (LeMasurier 1990). The axial part of the rift is characterized by thin crust (*c.* 20 km) with subglacial bedrock elevations that are commonly 1.0–2.5 km below sea level in the Byrd Subglacial Basin area, and by sedimentary basins with up to 14 km of Late Mesozoic and Cenozoic sediments in the Ross Sea region (Cooper *et al.* 1987, 1991).

The rift system is a composite rift zone (Behrendt *et al.* 1991) with the main period of extension in Late Cretaceous times (Lawver & Gahagan 1994) when sediments filled basins in the Ross Sea region. The uplift history of the southern rift flank, the Transantarctic Mountains, was also episodic. Stump & Fitzgerald (1992) have inferred four phases of uplift for the Transantarctic Mountains from apatite fission-track age profiles at *c.* 110 Ma, 80 Ma, 55 Ma and in Plio-Pleistocene times. Uplift of the Ellsworth Mountains commenced in Early Cretaceous times (Fitzgerald & Stump 1991).

There is a marked lack of correspondence between the age of volcanic provinces within the rift system and extensional and uplift events. Volcanism was greater during the Cenozoic than the Cretaceous in all but the Marie Byrd Land part of the rift system. In contrast, little or no extension occurred during the Cenozoic, except for a period of dextral transtension recognized from fault analysis in the Transantarctic Mountains (Wilson 1995) and continued development of narrow rifts in the Ross Sea (Cooper *et al.* 1987).

Ultimately, *c.* 85 Ma, continental separation occurred, but not as one might have expected in the West Antarctic rift, instead Marie Byrd Land remained attached to West Antarctic and seafloor spreading took place between Marie Byrd Land and New Zealand. The time of separation of Marie Byrd Land from the conjugate Campbell Plateau and Chatham Rise margin of the New Zealand continental block (Fig. 1) is constrained by anomaly Chron 34 (84 Ma; Mayes *et al.* 1990; Royer *et al.* 1990), the oldest identified oceanic magnetic anomaly adjacent to the Campbell Plateau.

Cenozoic magmatism

Cenozoic alkali basalt volcanism associated with the West Antarctic rift system commenced *c.* 30 Ma and continues to the present day. There are two major provinces, the Marie Byrd Land alkaline basalt province (LeMasurier 1990) on the northern flank, and the McMurdo Volcanic Group (Kyle 1990) within the axial rift. Interpretations of aeromagnetic and gravity data indicate an extensive subglacial basalt province along the axis of the rift system (Behrendt *et al.* 1994, 1996). LeMasurier & Rex (1989, 1991) suggested, based on geochemical and structural grounds, that the Marie Byrd Land province is related to a mantle plume with a 600 km diameter beneath Marie Byrd Land, whereas Behrendt *et al.* (1992) proposed a single 3500 km by 2000 km plume head beneath the entire alkaline basalt province.

Table 1. *Summary of Ar–Ar age data for Marie Byrd Land dykes*

Sample	Mineral	Age (Ma)	Error 2σ (Ma)
MB. 154.1S	Plagioclase	107	7
MB. 163.2S	Biotite	107	4
MB. 163.2S	Plagioclase	104	4
MB. 164.1S	Hornblende	107	2
MB. 164.1S	Biotite	108	1
MB. 205.5S	Plagioclase	112	7
MB. 205.9S	Hornblende	108	3

J value for samples was 0.00299 ± 0.000015.

Cretaceous mafic dyke suite from Marie Byrd Land

Field relationships

Mafic dykes and sills crop out along the Saunders–Ruppert–Hobbs coasts of central Marie Byrd Land together with a single layered gabbro intrusion (Fig. 3). In some cases the mafic rocks form up to an estimated 60% of the outcrop (e.g. Mount Prince), and are most abundant within the Cretaceous calc-alkaline granodiorite suite and older basement rocks. Only a few dykes cross-cut the contemporaneous to slightly younger anorogenic granite suite. The dykes vary up to 20 m thick but typically 1–3 m, and vary from trains of disrupted mafic fragments, to linear forms with dyke margins intruded and brecciated by veins of granitic host rock, and dykes with planar margins. Many of these features suggest that the dykes overlap in age with the final stages of the I-type granitoid emplacement. Bimodal dykes are also present indicating magma mingling at lower crustal levels. Dyke orientations are plotted on Fig. 3. They trend predominantly ESE–WNW subparallel to the coastline; the main exceptions are at Mount Prince and at Kinsey Ridge, at both of which there is also a more northerly dyke trend. In the case of Kinsey Ridge, this is due to emplacement into a N–S trending sinistral transtensional shear zone. At some localities, dykes may be grouped according to relative age of emplacement, with a maximum of five separate phases at Mount Prince. The first three phases show brecciated margins and net-vein structures, whereas the remaining two phases have sharp planar boundaries. A single layered gabbro intrusion cut by a small number of E–W trending mafic dykes is exposed at Cape Burks.

Age

Samples were analysed at the Open University, UK by laser stepped heating of small numbers (up to 10) of separated grains using the technique described in Pearson *et al.* (1996). No chemical cleaning was undertaken subsequent to mineral separation. Data were standardized using the GA1550 biotite standard and an age of 97.9 Ma as recommended by McDougall & Harrison (1988).

Dyke emplacement between 110 and 100 Ma is constrained by a U–Pb zircon age of 110 ± 1 Ma from the host granite at Mount Prince and a single U–Pb zircon age of 100.6 ± 0.7 Ma (Mukasa *et al.* 1994) from a felsic dyke at Mount Prince; the felsic dyke cross cuts most but not all of the dykes at Mount Prince and provides a younger age limit on most of the mafic dyke suites. Ar–Ar laser stepped heating age data are presented here (Table 1; full details can be obtained from the Society Library or the British Library Document Supply Centre, Boston Spa, Wetherby, West Yorkshire LS23 7BQ, UK as Supplementary Publication No. SUP 18133 (3 pages) for plagioclase, amphibole and biotite. Plagioclase samples yielded ages of 104 ± 4 Ma (MB. 163.2S), 107 ± 7 Ma (MB.154.1S), 108 ± 3 Ma (MB.205.9S) and 112 ± 7 Ma (MB.205.5S), ages for biotite range from 107 ± 4 Ma (MB.163.2S) to 108 ± 1 Ma (MB.164.1S) and a single amphibole sample yielded an age of 107 ± 2 Ma (MB.164.1S). Although the analytical errors on some of the plagioclase separates are relatively large, they indicate that the dykes were emplaced during a short period centred around the mean age of 107 Ma. The event lasted less than 5 Ma (2 standard deviations on the mean) and is likely to have lasted less than 2 Ma (error on the weighted mean). The age of 107 Ma is consistent with the U–Pb zircon age for the host rock and syn-emplacement structures such as mingling of the mafic and felsic magmas. Some of the mafic magmatism is slightly younger than 107 ± 5 as indicated by the few dykes which crosscut the anorogenic granite suite (102–95 Ma; Weaver *et al.* 1994) and by an Ar–Ar age of 99.9 ± 0.1 Ma from hornblende and biotite separated from a layered gabbro at Cape Burks (Palais *et al.* 1993; Mukasa *et al.* 1994).

Petrography

The mafic dykes are predominantly aphyric ophitic dolerites with varied proportions of augite, green pleochroic hornblende, brown amphibole, biotite, and plagioclase. Where cross-cutting relationships are observed, the oldest dykes are hornblende–biotite dolerites, with hornblende forming single crystals and crystal aggregates replacing original ferromagnesian minerals. The high-Ti dykes, which are the phase 4 dykes at Mount Prince, contain a slightly pink pleochroic augite ± biotite and plagioclase, with the younger phase 5 dykes being low-Ti in composition and containing augite together with a brown pleochroic amphibole. All dykes contain accessory apatite and opaque phases, and some show varied degrees of alteration to secondary actinolite, chlorite, calcite and epidote. Interstitial quartz is present in a few of the more evolved dyke rocks.

Geochemistry

Samples of Cretaceous mafic dykes for Marie Byrd Land were analysed by standard XRF methods at the University of Keele, UK, and by standard INAA procedures at the Open University, UK. Sr, Nd and Pb isotopes were measured at the NERC Isotope Geosciences Laboratory, Keyworth, UK. Static multi-collector $^{87}Sr/^{86}Sr$ ratios, normalized to $^{86}Sr/^{88}Sr=0.1194$ are accurate to 0.005% (1-sigma) and are reported relative to a value of 0.710225 for NBS 987 (standard rock BHVO-1 gave 0.70343). Rb/Sr ratios were derived from geochronological-grade X-ray fluorescence data obtained at British Geological Survey. εSr_{100} values are relative to a bulk earth $^{87}Sr/^{86}Sr$ of 0.7045. Dynamic multi-collector $^{143}Nd/^{144}Nd$ ratios, normalized to $^{146}Nd/^{144}Nd=0.7219$, are accurate to *q* 0.005% (1-sigma); BHVO-1 gave a measured value of 0.512990. Sm and Nd concentrations were measured by isotope dilution. Initial Sr and Nd compositions were calculated for 107 Ma. Pb isotopic ratios were normalized to preferred values for NBS 981 (Todt *et al.* 1984). The Pb data are not corrected for decay since the Cretaceous, but this correction is small, up to 0.1, 0.01 and 0.08 in the values of the ratios of ^{206}Pb, ^{207}Pb and ^{208}Pb to ^{204}Pb, respectively.

The samples have between 46.6 and 55.9 wt% SiO_2, and are classified as basalts (29), shoshonites (7), mugearites (5), hawaiites (4), potassic trachybasalts (4) and basaltic andesites (2) according to the recommended alkali-silica scheme (Le Maitre 1989); they range from sub-alkaline to weakly alkaline (Fig. 4a). Most of the rocks are sodic, although two are potassic ($K_2O>Na_2O$), and many of the samples are K-enriched relative to oceanic suites. MgO ranges from 8.9 wt% (close to values in melts in equilibrium with mantle peridotite) to 3.6 wt% (indicating significant fractionation from such melts). In common with other basalt provinces associated with Gondwana break-up (Brewer *et al.* 1992; Peate *et al.* 1992; Sweeney *et al.* 1994; Gibson *et al.* 1995), the Marie Byrd Land sample set includes high-Ti and low-Ti representatives. A threshold of 2.5 wt% TiO_2 at 46 wt% SiO_2 (Fig. 4b) can be used to separate a high-Ti Group A from a low-Ti Group B. Group A has higher Fe_2O_3 (>*c.* 13 wt% at 46 wt% SiO_2) and Zr (>*c.* 200 ppm) abundances than Group B (Fig. 4c, d). A subgroup of six samples straddling the threshold between high-Ti and low-Ti groups in the TiO_2 v. SiO_2 plot has low Fe characteristics of the low-Ti group, but high Zr abundances of the high-Ti group. These samples are treated as a separate Group C.

The three groups have distinctive elemental abundances and ratios (Figs 5 & 6). Group A is characterized by high P_2O_5 (0.4–1.44 wt%), Y (28–59 ppm), Zr (191–361 ppm), and relatively low Zr/Y (4.6–6.9). A few of the samples have low La/Ta ratios, close to values for ocean island basalts (OIB). Group B has low P_2O_5 (0.2–0.7), Y (15–34 ppm), Zr (98–243 ppm), and a range in Zr/Y (4.3–9.4) (Fig. 5b). They generally have higher La/Ta ratios than Group A. Group C (Fig. 5c) have the highest Zr (266–388 ppm), P_2O_5 (0.8–1.5wt%), Sr (781–1101 ppm) and La (23–64 ppm) abundances, and Zr/Y ratios (9.2–11.0) and are moderately potassic ($Na_2O/K_2O=0.54$–1.1). Two samples in this group (MB.219.4 and 219.2W) are unusual in that chondrite-normalized values for Ba, Rb, Th and K are less than those for La. Rare earth element (REE) abundances are similar for groups A and B (Fig. 6; La_N/Yb_N ranges 4.1–7.5 and 5.1–11.2 respectively). Group C has the steepest patterns (Fig. 6a; $La_N/Yb_N=14.7$). All the patterns are appropriate for continental basalts generated within the garnet stability zone of the mantle. Eu*/Eu anomalies are absent or positive and small, indicating that the magmas experienced no significant plagioclase fractionation or assimilation of Eu-depleted crust. There are striking geochemical resemblances between groups A, B and C and the high-Fe, low Ti-Zr and high Ti-Zr groups (respectively) described from the Karoo lavas by Sweeney *et al.* (1994).

The Marie Byrd Land dykes have restricted ranges in radiogenic isotopes (Table 2), all samples plotting in the OIB-array on Sr–Nd plots (Fig. 7a). Group A have the least radiogenic Sr ($\varepsilon Sr_{100}=0.3$–9.1) and most radiogenic Nd ($\varepsilon Nd_{100}=3.4$–4.0). Group C sample MB.219.4S has the most radiogenic Sr ($\varepsilon Sr_{100}=8.7$) and the least radiogenic Nd ($\varepsilon Nd_{100}=-0.3$). Group B are intermediate between these values. All groups have similar $^{207}Pb/^{204}Pb$ (15.61–15.63). Group A samples have relatively high $^{206}Pb/^{204}Pb$ and $^{208}Pb/^{204}Pb$ ratios (Fig. 7b, c; 18.88–19.02 and 38.68–38.78 respectively) compared with Group B (Fig. 7b, c; 18.74–18.82 and 38.53–38.61 respectively). Group C has Pb isotopes identical to those of Group B.

Petrogenesis

In this section, we compare the compositions of the Marie Byrd Land mafic dykes with those of Hikurangi oceanic

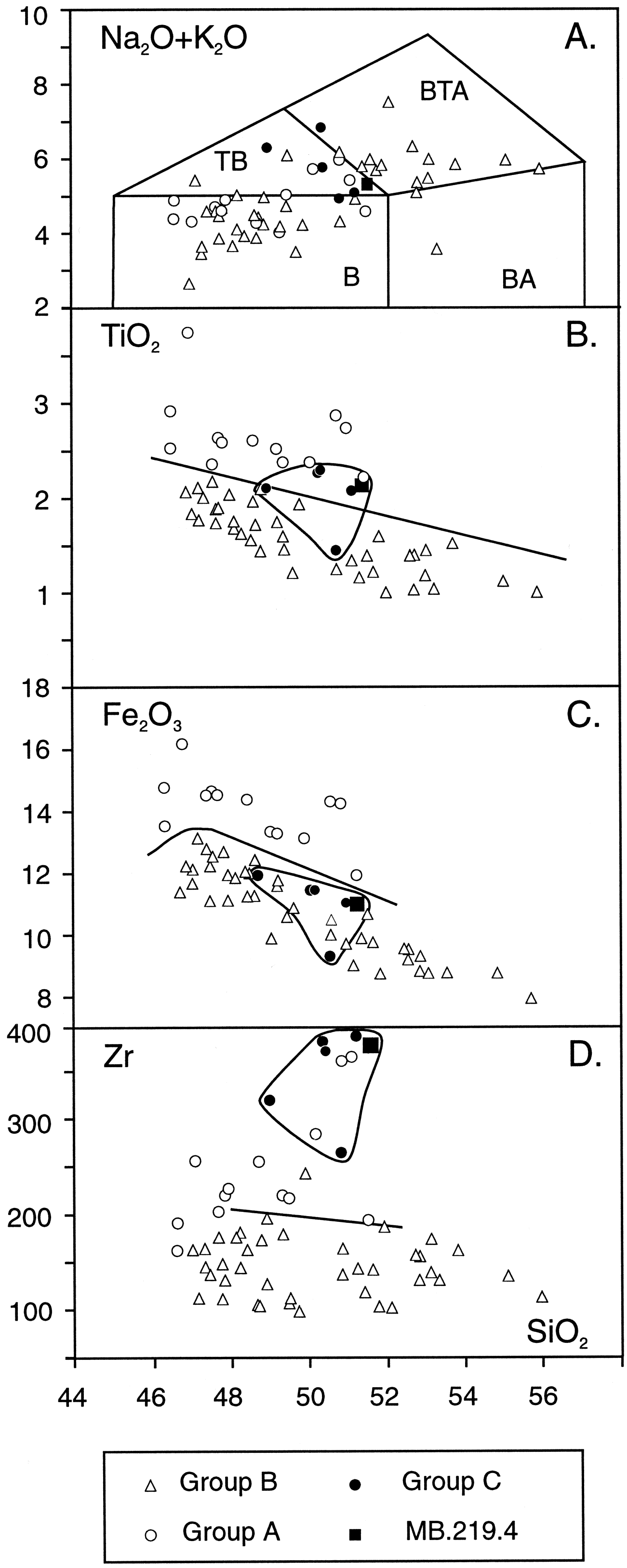


Fig. 4. Variation diagrams for mid-Cretaceous dykes, Marie Byrd Land. Rock types in (**a**) are: B, basalt; BA, basaltic andesite; TB, trachybasalt; BTA, basaltic trachyandesite (LeMaitre 1989). Lines in (**b–d**) separate high-Ti from low-Ti varieties. The different dyke groupings are explained in the text.

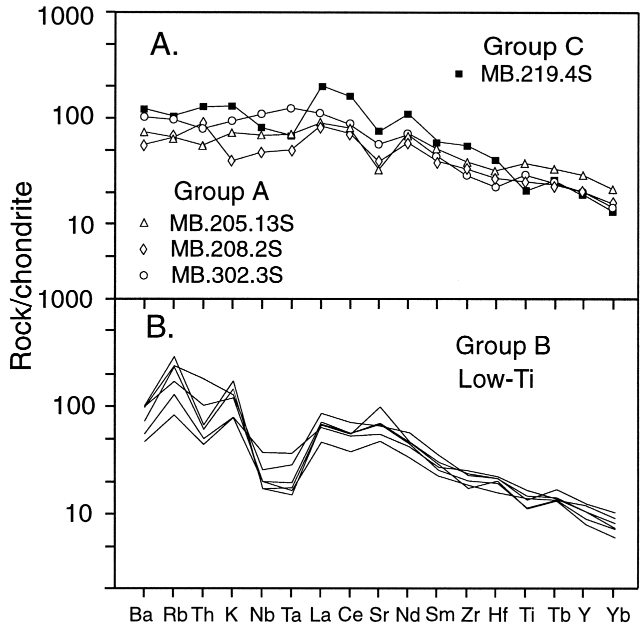


Fig. 5. Chondrite-normalized (Thompson *et al.* 1984) multi-element diagram showing comparative enrichments in incompatible elements for Marie Byrd Land mid-Cretaceous mafic dyke groups (see text for explanation of dyke groups).

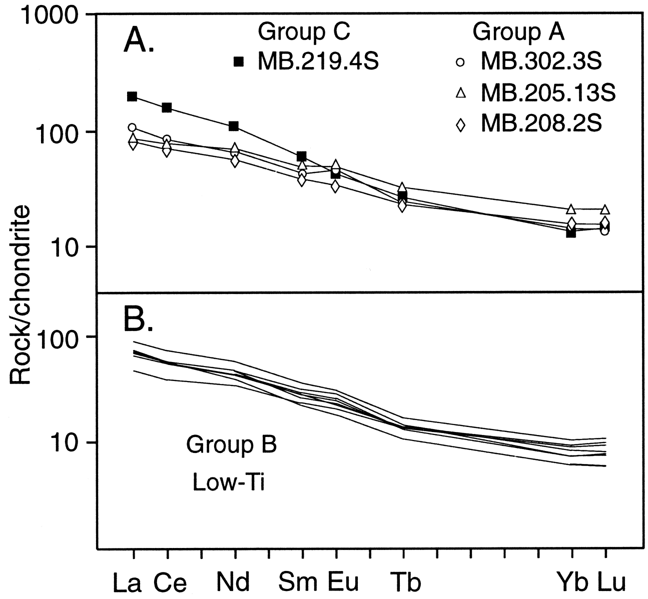


Fig. 6. Chondrite-normalized REE plots for Marie Byrd Land mid-Cretaceous mafic dyke groups.

tholeiites and West Antarctic Cenozoic basalts—the McMurdo Volcanic Group, which are significantly fractionationated from mantle-derived compositions (Wörner *et al.* 1989; Rocholl *et al.* 1995), and the chemically very similar MBL alkaline basalt province (Hole & LeMasurier 1994). We shall argue that: (i) crustal contamination of mantle-derived magmas could not have produced the observed compositions of the MBL mafic dykes; (ii) many features of the dykes, especially of Group C, were inherited from sources in underlying subcontinental lithospheric mantle; (iii) Groups A and B were produced by mixing between lithospheric melts and

Table 2. *Representative geochemical and isotopic analyses of mid-Cretaceous mafic dykes, Marie Byrd Land*

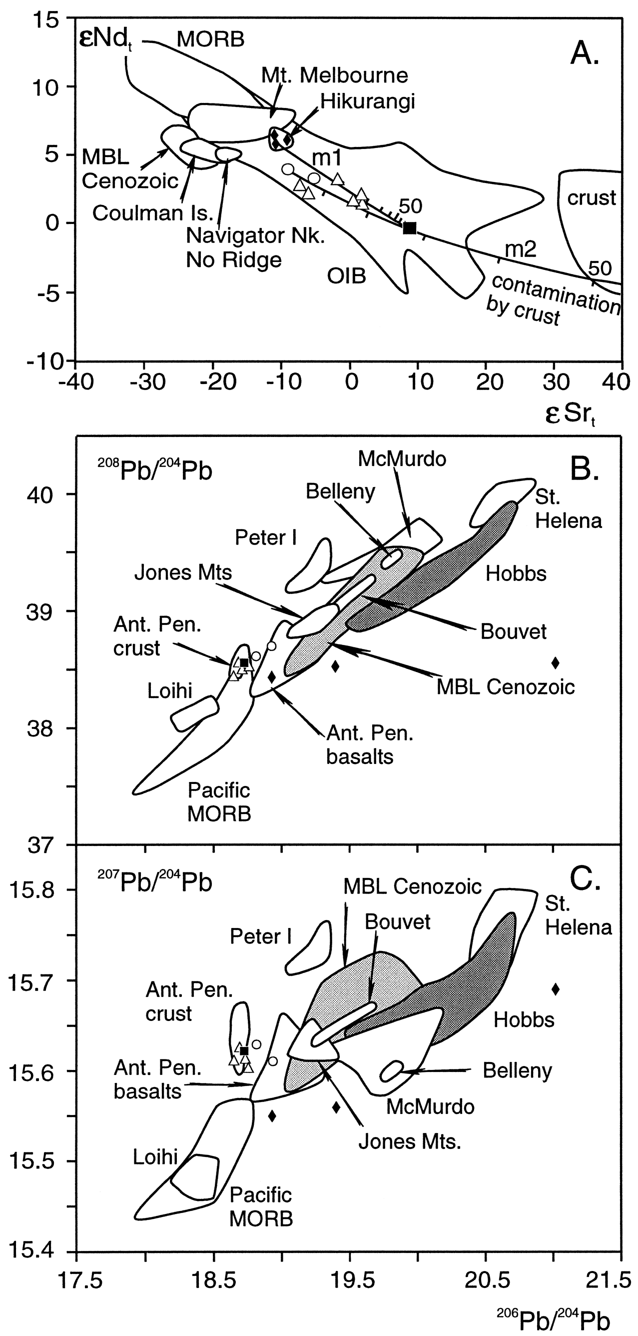
Sample Group	MB.202.1S B	MB.153.1S B	MB.205.4S B	MB.301.5S B	MB.205.1S B	MB.205.7S B	MB.205.11S B	MB.163.3S B	MB.302.3S A	MB.205.13S A	MB.208.2S A	MB.219.4S C
SiO_2	48.26	48.63	50.98	51.91	52.63	53.06	54.77	55.57	45.34	46.59	48.19	50.93
TiO_2	1.72	1.44	1.16	1.52	1.40	1.18	1.12	1.00	2.92	3.75	2.52	2.14
Al_2O_3	15.61	16.58	16.82	16.62	16.10	17.71	18.16	16.32	15.23	13.26	15.43	15.46
Fe_2O_3(T)	12.01	11.27	9.01	8.77	9.53	8.82	8.75	7.93	13.53	16.17	13.35	10.96
MnO	0.16	0.17	0.13	0.13	0.14	0.13	0.13	0.12	0.20	0.25	0.18	0.17
MgO	8.08	8.10	7.15	3.86	6.44	4.72	3.33	6.03	4.39	4.63	5.57	4.94
CaO	8.21	8.75	7.76	7.41	7.55	8.43	6.84	6.37	9.45	9.15	7.93	7.72
Na_2O	3.24	3.08	3.66	3.15	3.50	3.75	3.86	3.65	3.47	3.27	3.38	3.38
K_2O	1.14	1.13	2.09	2.49	1.85	1.73	2.07	2.03	1.28	1.00	0.55	1.84
P_2O_5	0.53	0.30	0.39	0.61	0.48	0.37	0.39	0.29	1.44	0.91	0.65	1.31
H_2O+	1.52	0.60	0.64	3.69	0.70	0.48	0.66	0.72	2.99	0.98	2.48	1.04
Total	100.48	100.05	99.79	100.16	100.32	100.38	100.08	100.03	100.24	99.96	100.23	99.89
Trace elements by XRF												
Cr	284	224	295	13	203	0	4	246	64	42	109	82
Cu	68	49	42	64	78	38	63	44	50	55	45	37
Ga	21	19	21	14	20	22	22	19	22	25	23	22
Nb	13	7	6	6	9	7	6	5	36	23	16	28
Ni	196	140	135	13	78	26	8	134	60	31	74	37
Pb	12	12	18	8	13	17	18	18	12	12	14	16
Rb	45	29	82	102	84	60	67	92	32	22	22	35
Sr	651	559	824	1169	774	802	774	684	639	373	440	884
V	235	208	219	239	223	221	231	171	192	386	298	206
Y	21	24	16	21	25	18	22	15	39	56	39	39
Zn	107	86	93	63	85	50	103	82	100	127	137	116
Zr	173	127	118	162	156	139	135	113	191	256	220	374
Ba	381	323	499	694	675	686	608	611	668	485	372	827
Trace elements by INAA												
La	20.8	15.2	22.3	21.9	28.2	23.4	23.0	22.6	34.4	27.9	26.3	64.1
Ce	45.8	32.7	48.6	47.7	61.1	48.5	46.9	47.6	71.6	66.3	59.4	135
Nd	26.5	21.0	29.3	29.5	35.9	28.2	26.7	23.8	41.9	43.4	35.4	68.1
Sm	5.52	4.57	5.84	6.15	7.13	5.17	5.55	4.39	8.55	9.97	7.71	12.0
Eu	1.85	1.57	1.94	2.14	2.33	1.75	1.69	1.35	3.49	3.72	2.57	3.28
Tb	0.72	0.70	0.69	0.74	0.88	0.70	0.72	0.54	1.23	1.67	1.19	1.35
Yb	1.80	2.00	1.33	1.62	2.26	1.59	2.02	1.35	3.09	4.56	3.35	2.93
Lu	0.27	0.31	0.20	0.25	0.36	0.26	0.33	0.20	0.47	0.70	0.52	0.48
Th	2.10	1.85	2.56	2.80	7.59	4.28	6.11	6.50	3.17	2.23	3.57	5.23
U	0.76	0.69	0.82	1.11	1.41	1.50	1.57	1.87	1.03	0.70	1.10	1.75
Ta	0.73	0.33	0.30	0.35	0.57	0.39	0.35	0.41	2.35	1.34	0.96	1.35
Hf	4.46	3.16	4.04	4.26	4.29	3.86	3.68	3.32	4.34	6.18	5.25	8.04
$^{87}Rb/^{86}Sr$	0.200	0.150	0.288	0.252	0.314	0.216	0.250	0.389	0.145	0.171	0.145	0.115
$^{87}Sr/^{86}Sr_{(m)}$	0.704244	0.704071	0.704831	0.704622	0.704872	0.705032	0.704668	0.705052	0.703949	0.704252	0.704610	0.705157
$^{87}Sr/^{86}Sr_{(i)}$	0.703960	0.703858	0.704422	0.704263	0.704426	0.704724	0.704312	0.704499	0.703743	0.704009	0.704404	0.704994
$\varepsilon Sr_{(i)}$	−6.1	−7.5	0.4	−1.9	0.4	4.7	−1.2	1.3	−9.1	−5.3	0.3	8.7
Sm	5.698	4.829	5.708	6.575	7.088			4.354	8.654	10.011		12.333
Nd	26.973	21.044	27.140	30.149	33.908			23.001	40.952	41.453		67.242
$^{143}Nd/^{144}Nd_{(m)}$	0.512710	0.512751	0.512709	0.512766	0.512680			0.512673	0.512796	0.512778		0.512564
$^{147}Sm/^{144}Nd$	0.1277	0.1387	0.1272	0.1318	0.1264			0.1144	0.1278	0.1460		0.1109
$^{143}Nd/^{144}Nd_{(i)}$	0.512626	0.512660	0.512626	0.512680	0.512597			0.512598	0.512712	0.512682		0.512491
$\varepsilon Nd_{(i)}$	2.3	3.0	2.3	3.3	1.7			1.8	4.0	3.4		−0.3
$^{206}Pb/^{204}Pb$	18.820	18.781	18.744			18.737			19.024	18.876		18.829
$^{207}Pb/^{204}Pb$	15.607	15.617	15.628			15.617			15.616	15.632		15.628
$^{208}Pb/^{204}Pb$	38.600	38.583	38.610			38.529			38.785	38.680		38.668

A, B, C are dyke groups refered to in the text.

asthenosphere-derived melts similar to those of Hikurangi Plateau.

Mixing between high-Ti basalt and local Swanson Formation metasedimentary crust could generate the range of Sr and Nd isotopic compositions of the MBL mafic dykes with only 30% contamination of basalt by crust (Fig. 7a). Nevertheless, certain trace element compositions of the dykes cannot be generated by contamination of any observed mafic magma composition by any plausible crustal rocks. For example, although local crust has the low Ti/Zr ratios required to generate the low Ti/Zr MBL dyke magmas from Hikurangi tholeiite, the amount of assimilation required, up to 80%, would result in magmas much more silicic than those observed (Fig. 8a). Moreover, crustal assimilation of any likely crust cannot produce the high Zr/Y, Group C magmas from either tholeiitic or alkaline parents (Fig. 8a).

The MBL dykes form a trend between high Ti/Zr, low Zr/Y basalt, similar to that of Hikurangi, and low Ti/Zr, high Zr/Y alkaline magmas of Group C. There is also a positive relationship between εNd_t and Ti/Zr (Fig. 8b). Since enrichment in incompatible elements in the low Ti/Zr magmas renders them more resistant to the effects of crustal contamination, as measured by Sr and Nd isotopes, it is highly unlikely that the low εNd_t composition of Group C is a result of crustal contamination. It is much more probable that relatively alkaline, high Zr/Y, low εNd_t Group C represents magmas derived from lithospheric mantle sources. Conversely, the high εNd_t and Ti/Zr component in Fig. 8b has $\varepsilon Nd_t \geq 4$, close to asthenosphere-derived Cenozoic alkali basalts (Hole & LeMasurier 1994), and is likely also to have been asthenosphere-derived. Highest Ti/Zr ratios (asthenospheric end-member) in Groups A and B are 94 and 114 respectively, close to values for N-MORB and Hikurangi tholeiite (103 and 110 respectively; Sun & McDonough 1989; Mortimer & Parkinson 1996). Similar relationships occur for Ce/Y ratios in that MBL mafic dykes lie between values for Group C and a low Ce/Y end-member. Lowest Ce/Y ratios in Groups A and B are 1.3 and 0.8 respectively, slightly higher than values for E-MORB (0.7; Sun & McDonough 1989). It is not possible to compare these ratios with those of Hikurangi Plateau tholeiite, whose LREE abundances are poorly known, owing to effects of alteration and analytical uncertainty (Mortimer & Parkinson 1996), but whose Ce/Y ratios are clearly low (≤ 1). These relationships suggest that the dominant asthenospheric end-member for the MBL mafic dykes was tholeiitic in composition, and similar to Hikurangi tholeiite and E-MORB.



In Fig. 8c, Group A forms a mixing array between Group C and a tholeiitic end-member, T, with Ce/Nb increasing modestly from *c.* 2.2 in T to 5.5 in Group C. The increase in Ce/Nb with increasing Ce/Y suggests an inherited subduction component in the lithospheric source. Most Group B samples have higher Ce/Nb than this trend, having Ce/Nb ratios up to 13.5. An implication is that Group B samples contain a component of a different end-member. This third end-member (C2 in Fig. 8c) has lower Zr/Y and Ce/Y than Group C. Its εNd_t values of *c.* +1 to +2 indicate that it represents partial melts of lithospheric mantle having higher time-integrated Sm/Nd ratios than the source of Group C. We propose that Groups A and B represent magma mixing between asthenosphere-derived tholeiite and two different partial melts of lithospheric mantle, although the asthenosphere component of group A might have had higher Ce/Y, and hence represent smaller degree partial melts than the asthenosphere component of Group B. This may explain the distinctive high Ti compositions of Group A in Fig. 8d which plot between Group C and Cenozoic basalts. In the same diagram, Group B samples appear to plot on a mixing trend between tholeiite T and C2 (trend b1), and modified by variable amounts of fractional crystallization (trend b2). The implication of the relatively even spread of Group A and B compositions between the several end-member compositions (Fig. 8d) suggests that many samples contain components of all three end-members.

The Cenozoic volcanic rocks of West Antarctica form a dominantly HIMU OIB province, having low εNd_t, low εSr_t—mostly plotting below the MORB-OIB array (Fig. 7a)—and high $^{206}Pb/^{204}Pb$, $^{207}Pb/^{204}Pb$ and $^{208}Pb/^{204}Pb$ ratios relative to Pacific MORB and Hawaii OIB (Fig. 7b, c). Rocholl *et al.* (1995) argued that three mantle sources were tapped by the McMurdo Volcanic Group: a HIMU plume source, a depleted mantle plume source and an enriched lithospheric mantle source. They were unable to determine whether the HIMU source represents a contemporary plume or old plume material fossilized in the lithosphere. Group B and C samples from the MBL mafic dykes have Pb isotopic

Fig. 7. Sr, Nd and Pb isotopic relationships of Mesozoic–Cenozoic rocks in West Antarctica. Symbols as in Fig. 4, except filled diamonds are Hikurangi Plateau basalts (Mortimer & Parkinson 1996). (**a**) εNd_t versus εSr_t showing Marie Bryd Land mid-Cretaceous mafic dykes relative to fields for MORB, OIB (Staudigel *et al.* 1984), Marie Byrd Land Cenozoic mafic volcanic rocks (Hole & LeMasurier 1994), McMurdo Volcanic Group (Mt Melbourne, Wörner *et al.* 1989; Coulman Island, Navigator Nunatak and No Ridge, Rochall *et al.* 1995), Carboniferous–Triassic igneous and metasedimentary Marie Byrd Land crust (Pankhurst *et al.* 1998) and Hikurangi Plateau tholeiites. Mixing curve m1 is between Hikurangi tholeiite and sample MB.219.4S; mixing curve m2 models contamination of Group A sample MB.302.3S by metasediment sample MB.211.1W (Pankhurst *et al.* 1998). (**b, c**) Comparison of mid-Cretaceous mafic dykes with the HIMU Antarctic Province. End member compositions are St Helena (Chaffey *et al.* 1989), Pacific MORB (Ito *et al.* 1987) and Loihi (Staudigel *et al.* 1984). Other fields: Hobbs Lineament and other Marie Byrd Land Cenozoic basalts (Hart *et al.* 1997), Bouvet (Sun 1980), McMurdo Volcanic Group (Rochall *et al.* 1995), Balleny Island (Hart 1988), Peter I Island (Prestvik *et al.* 1990, Hart *et al.* 1995), Jones Mountains (Hart *et al.* 1995), Antarctic Peninsula Cenozoic alkaline mafic volcanic rocks (Hole *et al.* 1993) and Antarctic Peninsula galenas which record local crustal Pb isotope ratios (Willan & Swainbank 1995).

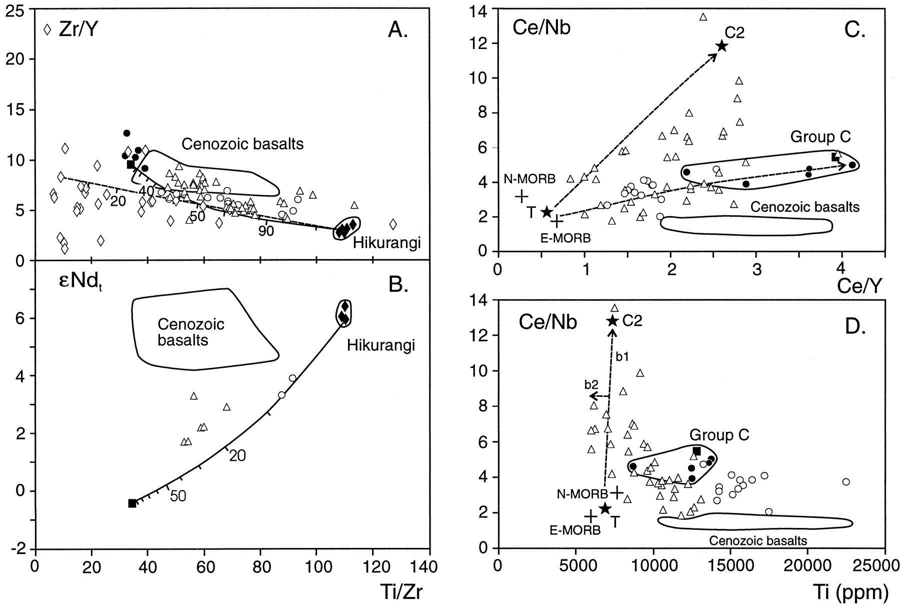


Fig. 8. Incompatible element and Nd isotopic variations in Marie Byrd Land mid-Cretaceous mafic dykes. Symbols as in Fig. 4 except open diamonds are Marie Byrd Land pre-mid-Cretaceous igneous crust (Pankhurst *et al.* 1998), filled diamonds are Hikurangi Plateau basalts (Mortimer & Parkinson 1996). Field for Marie Byrd Land Cenozoic basalts is from Hole & LeMasurier (1994), and MORB average compositions Sun & McDonough (1989). Mixing curve models contamination of Hikurangi tholeiite with sample MB.219.4S, and Triassic monzogranite sample MB.212.3P (Pankhurst *et al.* 1998). In C and D, stars T and C2 represent putative end member compositions: dashed lines b1 and b2 are idealized petrogenetic trends explained in the text.

compositions close to Pacific MORB and Antarctic Peninsula crust and, by implication, most Marie Byrd Land and Antarctic Peninsula lithosphere was formed by active margin processes from Pacific ocean plate, consistent with current models of West Antarctic lithosphere evolution (Pankhurst *et al.* 1997). Group A samples plot to higher $^{206}Pb/^{204}Pb$ and $^{208}Pb/^{204}Pb$ ratios than Groups B and C (Fig. 7b, c). Inasmuch as these Group A samples probably contain large components from an asthenospheric source (Fig. 7b, c), we suggest that this source was HIMU, similar to that of Cenozoic rocks of West Antarctica (Fig. 7b, c). These Group A Cretaceous samples represent the earliest record of HIMU mantle compositions in West Antarctica and their source can be specified as dominantly asthenospheric. Subsequent Cenozoic magmatism may have incorporated this HIMU signature fossilized as lithospheric mantle into the alkaline basalts (Rocholl *et al.* 1995; Hart *et al.* 1997). Alternatively, HIMU mantle convected by mantle plumes may have contaminated large areas of the sub-West Antarctic asthenosphere, so all Cenozoic basalts from the region from the Ross Sea (Rocholl *et al.* 1995) to the Antarctic Peninsula (Hole *et al.* 1993), including oceanic representatives (Peter I Island and Bellany Island, Hart 1988; Hart *et al.* 1995) contain elements of the HIMU signature.

Hikurangi Plateau tholeiites have similar εNd to, but higher εSr than, all Cenozoic volcanic rocks in West Antarctica except certain Mount Melbourne (McMurdo Volcanic Group) samples interpreted as containing components from lithospheric sources (Rocholl *et al.* 1995). Pb isotopic compositions of Hikurangi Plateau tholeiites are scattered, implying mobilization of Pb (Mortimer & Parkinson 1996) but, nevertheless, two of the samples have $^{206}Pb/^{204}Pb$, $^{207}Pb/^{204}Pb$ and $^{208}Pb/^{204}Pb$ between Pacific MORB and HIMU, suggesting that both these sources contributed to the oceanic plateau magmas. The Hikurangi Plateau has εSr_{100} values of -10, instead of *c.* -20, as would be expected if it had a HIMU-MORB source. This may be a result of seawater alteration (although the tight clustering of the three values makes this unlikely). Alternative explanations are that their mantle source was contaminated by subducted continental sediment or that assimilation of continental lithosphere by the magmas took place. Either process could have occurred if, as proposed, the plateau formed adjacent to the continental margin.

Evidence for mantle plumes beneath Marie Byrd Land

Cenozoic Marie Byrd Land plume

There is little doubt that a mantle plume has existed beneath the Marie Byrd Land Cenozoic alkaline basalt field since *c.* 30 Ma ago. The main evidence is as follows.

(1) There is a positive geoid anomaly over western Marie Byrd Land. It has an amplitude of 5–7 m and a diameter of some 800 km (SCAR 1990) similar in dimensions to those of sub-oceanic mantle plumes.

(2) Marie Byrd Land was regionally and differentially uplifted by 600–2700 m over an area some 1000 km across (LeMasurier & Rex 1983) during the Cenozoic. The uplift most probably reflects dynamic or thermal support by a mantle plume and/or underplating by mafic magma.

(3) Magma compositions are consistent with a plume derived asthenospheric source (Hole & LeMasurier 1994; Panter *et al.* 1994; Hart *et al.* 1997).

(4) Measured heat flow values of 1.7–1.9 HFU recorded in ice cores from Marie Byrd Land (Alley & Bently 1988) are some 0.2–0.4 HFU above normal values for continental crust and may be due to high mantle-derived heat flow associated with a mantle plume.

(5) Silicic volcanism shows a systematic age progression from the centre of the plume uplift to its peripheries, interpreted to be the result of deformation associated with crustal doming over the head of a mantle plume (LeMasurier & Rex 1989).

Cretaceous plume

The main evidence for a mantle plume beneath the mid-Cretaceous anorogenic and mafic dykes suites along the Marie Byrd Land margin is a follows.

(1) The volume of anorogenic silicic and mafic magma along the Marie Byrd Land margin is typical of volcanic margins formed above a mantle plume.

(2) The mafic dyke rocks contained an asthenospheric end-member component that was tholeiitic in composition and similar to Hikurangi tholeiite and E-MORB. The tholeiitic component indicates that the degree of mantle melting was greater than that associated with normal mantle temperatures, assuming that lithospheric thicknesses were not unusually thin.

(3) Isotopic data indicate that the asthenospheric component was HIMU, typical of plume-derived sources in the South Pacific region.

(4) Cretaceous metamorphic complexes along the Marie Byrd Land margin indicate rapid uplift following peak metamorphic conditions between 105 and 94 Ma. This is in marked contrast to the conjugate New Zealand margin which subsided as New Zealand moved away from Marie Byrd Land to form deep sedimentary basins.

Mount Erebus plume

Allowing for motion of the Antarctic plate relative to the hot-spot reference frame (Mueller *et al.* 1993), a stationary hot-spot situated below the Ruppert–Hobbs coast during mid-Cretaceous times (Weaver *et al.* 1994) would now be situated beneath Mount Erebus (Fig. 1). This leads us to conclude that there are currently two mantle plumes beneath West Antarctica; the Marie Byrd Land plume was initiated *c.* 30 Ma ago, and the Mount Erebus plume (Fig. 1) in mid-Cretaceous times beneath the Ruppert–Hobbs coasts. Most previous authors have considered a single large Cenozoic plume head beneath this region (e.g. Behrendt *et al.* 1996) but we suggest two separate plume heads. This interpretation is supported by the fact that there is a separate *c.* 5 m positive geoid anomaly having a diameter of 600 km, centred 100 km west of Mount Erebus, by the fact that the eastern Ross Sea is an area of high heat flow, possibly the highest in Antarctica (Blackman *et al.* 1987), having heat flow values of 66–114 m Wm^{-2} (1.7–2.9 HFU, Blackman *et al.* 1987), averaging well above the continental average of *c.* 60 m Wm^{-2}, and by the high estimates of magma production rates for Mount Erebus which suggest that the source of Mount Erebus is hot mantle convected in a mantle plume (Kyle *et al.* 1992). The uplift history and fault pattern of the Transantarctic Mountains, the rift shoulder, are also more compatible with a Cretaceous plume than a Cenozoic Marie Byrd Land plume, as the main period of uplift of the Transantarctic Mountains and extension within the West Antarctic rift system occured prior to inception of the Marie Byrd Land plume. The rift margin in the Ross embayment strongly cross-cuts the major structure within and between the Palaeozoic Robertson Bay, Bowers and Wilson terranes (Stump 1995). This is contrary to the major faults of most continental rifts, which tend to follow pre-existing crustal structure and consistent with the interpretation that the position of the rift margin along the 500 km radius curvilinear Ross embayment was controlled by progressive eastward collapse of the Palaeozoic crust of the Transantarctic Mountains into the developing rift as the lithosphere migrated over the plume head. Such collapse could have resulted from lithospheric weakening owing to heating by the mantle plume.

Break-up model

If it is accepted that two mantle plumes were present below West Antarctic during Cretaceous–Cenozoic times, the break-up and post-break-up evolution of Marie Byrd Land can be readily explained. The reconstructed track of the Erebus mantle plume, shows that it was below the site of New Zealand–Marie Byrd Land rifting during the mid-Cretaceous break-up event (Fig. 1). We propose that it was this plume (Weaver *et al.* 1994) that was responsible for the large volume of magmatism emplaced in Marie Byrd Land at that time, and for the mid-Cretaceous uplift. We further propose that the large-volume Hikurangi Plateau was generated from mantle convected in this mantle plume, while the Phoenix–Pacific spreading centre was migrating along the outboard coast of the New Zealand block, but within the influence of the Mount Erebus mantle plume. Marie Byrd Land subsequently drifted away from the influence of this plume, as the Antarctic plate moved slowly west and as Marie Byrd Land rifted away from the Transantarctic Mountains to form Cretaceous basins in the Ross Sea embayment region. The initial uplift phase of the Transantarctic Mountains was most probably related to this event. Marie Byrd Land experienced tectonic and magmatic quiescence from mid-Cretaceous to Late Cenozoic times, when it was distant from the Mount Erebus plume. At about 30 Ma, Marie Byrd Land started to be influenced by the Marie Byrd Land plume, generating a second phase of uplift and renewed magmatism. This may represent the impact of a newly formed plume on the base of the lithosphere, or westward drift of the continental edge over an established plume. If our interpretation is correct, the Hikurangi Plateau, the mid-Cretaceous dykes of Marie Byrd Land, and the McMurdo volcanic Group were all ultimately generated from the Mount Erebus plume. The continued presence of the Mount Erebus plume within the West Antarctic rift system has provided the dynamic and thermal support to maintain the Transantarctic Mountains as a rift shoulder and produce Cenozoic basalts within the rift zone (Behrendt *et al.* 1996).

The role of the Cretaceous Erebus plume in New Zealand–Marie Byrd Land rifting is by no means certain. The broad extensional province that developed within the West Antarctic rift system, including Marie Byrd Land, in mid-Cretaceous times is most probably related to changes in plate boundary forces along the New Zealand margin due to cessation of subduction at the time of, or just prior to, collision of an oceanic spreading centre with the trench (Fig. 1). Continued divergence of the Pacific and Antarctic plate resulted in formation of the broad extensional province. We conclude, in agreement with Weaver *et al.* (1994), that the eventual locus of seafloor spreading between Marie Byrd Land and New Zealand was controlled by the position of the mantle plume. If it were not for the mantle plume, continental separation may have occurred between the Transantarctic Mountains and Marie Byrd Land.

A mid-Cretaceous plume is by no means without precedent in the South Pacific region. In fact it fits in well with the idea of Larson (1991) who postulated a mid-Cretaceous superplume event in the South Pacific region. The plume may consequently be one of series of plumes in this region that produced some of the largest oceanic plateaus in the world and was responsible for the final demise of Gondwana and the separation of New Zealand from Antarctica. Component plumes of the superplume event may have been responsible for underplating parts of Gondwana (e.g. Marie Byrd Land and

New Zealand) with a HIMU lithospheric mantle component to provide, upon remobilization, a source for inherited HIMU signatures in younger Cenozoic basalts in the South Pacific region (Lanyon *et al.* 1993; Weaver *et al.* 1994; Hart *et al.* 1997).

Low- and high-Ti magmas

Classification of basalts associated with Gondwana basalts into high- and low-Ti types is well entrenched in the literature: nevertheless criteria for distinguishing between them are matters of debate. All except one of both high- and low-Ti types of the Marie Byrd Land mafic dykes have Ti/Y ratios > *c.* 300, and so would be classified as high-Ti types, according to this discriminant (Peate *et al.* 1992). The Marie Byrd Land mafic dykes have neither the very low Ti/Y ratios of extreme low-Ti basalts from the Ferrar and Paraná provinces, nor the very high Ti abundances of parts of the Karoo province (Brewer *et al.* 1992). The striking similarity of the Marie Byrd Land groups and high-and low-Ti groups in the Karoo lavas (Sweeney *et al.* 1994) nevertheless suggests that the Marie Byrd Land dykes are real representatives of the Gondwana-wide groupings. Our interpretation that the high-Ti and low-Ti magmas in the Marie Byrd Land mafic dykes represent mixtures between asthenosphere-derived magma and two other compositional end-members, both of which represent partial melts of lithospheric mantle, is in agreement with the interpretation of the origin of Gondwana-wide basalt provinces proposed by Gibson *et al.* (1995) in so far as mixing of magmas from lithospheric and asthenospheric end-members is seen as a key process. Gibson *et al.* (1995) further proposed that, across the Gondwana interior, high-Ti lithospheric mantle partial melts were generated from younger lithosphere (Proterozoic mobile belts) than low-Ti partial melts (cratonic areas). In Marie Byrd Land, crustal Nd model ages span much of the Proterozoic. The time when the lithospheric mantle source of Group C magmas became stabilized might have been Proterozoic also. The high εNd_t values of the C2 end-member would imply that its source is younger than that of Group C. The relatively young age and strongly arc-like character of the C2 end member suggests that it developed later, perhaps during Mesozoic arc magmatism in MBL. An implication is that such low-Ti end-members are not restricted to cratonic areas in Gondwana, as proposed by Gibson *et al.* (1995), but they also developed much later at active plate margins.

Conclusions

(1) A voluminous suite of mafic dykes, dated at 107 Ma by Ar–Ar laser stepped heating, was emplaced in mid-Cretaceous times in Marie Byrd Land. Their emplacement predated separation of Marie Byrd Land from New Zealand by seafloor spreading and may indicate rifting in the presence of a mantle plume.

(2) Magmatism in the West Antarctic Rift System was related to two mantle plumes. The Marie Byrd Land mantle plume, now situated below Marie Byrd Land, was responsible for the Cenozoic alkali magmatism and is still active today. The Mount Erebus plume, now below the eastern Ross Sea, was situated below Marie Byrd Land in the mid Cretaceous, and was responsible for magmatism associated with Marie Byrd Land–New Zealand rifting.

(3) The mid-Cretaceous mafic dykes in Marie Byrd Land form co-existing high and low Ti suites. The origin of the suites can be explained by mixing between at least three end members, with minimal involvement of crust. The end members were, a plume derived (HIMU) OIB component, a high K/Nb, La/Ta, low Ti lithosphere-derived component, and a low K/Ta, moderate La/Ta, high Ti, high Zr, K, lithosphere-derived component.

(4) The high-Ti and low-Ti dyke suites were produced by mixing of athenosphere-derived magmas with partial melts of two different lithospheric mantle sources, consistent with the model of Gibson *et al.* (1995). Moreover, the asthenosphere-derived component in the high-Ti group may have been produced by smaller degrees of partial melting in the low-Ti group.

(5) The HIMU component in the Marie Byrd Land dykes is typical of Cenozoic rocks in the area, but represents the first known Cretaceous occurrence of the end member in West Antarctica.

(6) The progression from lithosphere-OIB mixing (mid-Cretaceous) to OIB only (Cenozoic) magmatism in Marie Byrd Land is similar to the progression in the Ross Sea embayment during Tertiary times. The feature in both areas is consistent with models of lithospheric–asthenospheric interaction below rifts, but the west-younging trends are also consistent with westerly relative motion of underlying plumes.

(7) A mid-Cretaceous mantle plume beneath Marie Byrd Land controlled the position of break-up between New Zealand and Marie Byrd Land within the West Antarctic rift system. The cause of the Cretaceous extension within the West Antarctic Rift System was mostly due to cessation of subduction along the New Zealand margin and transfer of Marie Byrd Land from the Antarctic plate to the Pacific plate.

(8) A phase of uplift of the Transantarctic Mountains in Cretaceous times was most probably due to impingement of the Cretaceous plume at the base of the lithosphere. The continued presence of this plume head has provided the thermal and dynamic support for this mountain range throughout the Cenozoic with potential variations in heat flux controlling episodic uplift events during the Cenozoic.

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