Petrogenesis of the volcanic suite of Bouvetøya (Bouvet Island), South Atlantic

TORE PRESTVIK, STEVEN GOLDBERG & GORDON G. GOLES


Whole rock and mineral compositions of volcanic rocks collected during the Norwegian Polarsirkel expedition (1978/79) to the volcanic island of Bouvetøya (close to the Bouvet Triple Junction) are discussed and compared with previously published data from the island. The rock types, hawaiite, benmoreite, and peralkaline trachyte and rhyolite (comendite) are related to each other by crystal fractionation processes. The trace element and radiogenic isotope signatures displayed by the Bouvetøya rocks are those of a moderately enriched oceanic island suite. On several isotope plots Bouvetøya rocks fall on or close to mixing lines between the enriched EM-1 and HIMU mantle components. Mixing between depleted morb mantle (DMM) and enriched components is not likely. Thus, Bouvetøya displays a typical plume signature.

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Introduction

Bouvetøya (official Norwegian spelling) is located 50 km west of the South-West Indian Ridge (SWIR) in the The Bouvet Triple Junction (BTJ) area of the South Atlantic (Fig. 1) and consists entirely of young (<1.4 Ma) volcanic rocks. Bouvetøya is 95% covered by permanent ice and no detailed geological map can be made. Geologic and petrologic features have been described by several authors. Among these are the contributions of Verwoerd et al. (1976) and Imsland et al. (1977), who described the geology and petrology of Prestvik (1982a), who reported on trace element geochemical features, and of le Roex & Erlank (1982), who discussed the petrologic evolution of the volcanic suite of the island. O’Nions & Pankhurst (1974), O’Nions et al. (1977), and Sun (1980) have published some data on Sr, Nd, and Pb isotopes. All these contributions were based on samples collected during short visits to the island (which is of difficult access) from the late 1920s to 1966. Geologic features of the island were described in considerable detail by Prestvik & Winsnes (1981), who participated in the Norwegian Polarsirkel expedition during the Antarctic summer of 1978/79. During this expedition 15 different locations on the island were sampled, resulting in 83 samples with good geological control. According to the IUGS classification based on the non-genetic total alkalis–silica (TAS) diagram, the rocks plot in the fields S1, S2, S3, T and R (Le Bas et al. 1986). The series is sodic (Na2O – 2 ≥ K2O) including a wide variety of hawaiites (grading into mugearite) together with benmoreites and rhyolites. In addition, we have two samples of altered mugearite (B 31 & B 34) and one fresh sample of trachyte (B 16). Trachyte (q <20) has not been reported from Bouvetøya previously. The whole series is oversaturated (q-normative), and the trachyte and rhyolites are peralkaline (ac-normative; see Tables 1, 3). Thus, the rhyolites are comendites, a term that is used in what follows.

The present account is based on material from the Polarsirkel expedition. We present (1) major and trace element composition of the various rock types, (2) microprobe data on the minerals and trace element data on plagioclase separates, and (3) isotope ratios of O, Sr, Nd, and Pb. Furthermore, we (4) discuss the results of

Fig. 1. Schematic configuration of the Bouvet Triple Junction (BTJ), with the location of Bouvetøya. The sketch is based on data from Simonov et al. (1996), Ligi et al. (1997) and Mitchell & Livermore (1998). MAR = Mid-Atlantic Ridge, AAR = American–Antarctic Ridge, SWIR = South-West Indian Ridge. The black arrow indicates that the Spiess Ridge is propagating northwestwards. Also, the location of enriched MAR basalts at 54°S, commonly referred to as the ‘Shona-Ridge anomaly’ (Moreira et al. 1995), is indicated by the stippled pattern of the ridge. The geographic location and plate configuration of the BTJ are shown in the lower, right-hand corner.
hydrothermal alteration, (5) present a revised petrologic model, and (6) discuss the petrogenesis of the suite.

The petrographic features of the major rock types (except trachyte) have been well described in several of the previous studies (op. cit.). The newly discovered trachyte (B 16) mainly consists of a brown glass in which very few microphenocrysts of feldspar (oligoclase to anorthoclase) and almost pure hedenbergite are disseminated together with crystallites of feldspar and Titanomagnetite. Samples of crystalline comendite from a dome at Kapp Valdivia in the north of the island contain alkali feldspar, pure hedenbergite and fayalite, arfvedsonitic amphibole, and ilmenite.

In the study of Imsland et al. (1977) in particular, the rock types are described under different names; 'transitional basalts' I and II, (for hawaiites), and 'transitional icelandite' (for benmoreite), but these authors recognized the peralkaline nature of the rhyolites and used 'comendite'. Le Roex & Erlank (1982) used the terms 'accumulated hawaiite', 'hawaiite', 'mugearite', 'benmoreite' and 'rhyolite'.

Previous petrologic models and associated problems

Both Imsland et al. (1977) and le Roex & Erlank (1982) concluded that the rock types of the Bouvetøya suite were related by crystal fractionation and/or accumulation processes. However, both these studies recognized 'problems' at various model stages; these will be commented on below.

Imsland et al. (1977) divided the basic rocks into two groups on the basis of their K₂O content and concluded that these two groups were not related by simple fractionation, mostly because the model required a more sodic plagioclase than observed in abundant megacrysts. They concluded that the benmoreites (trachytic icelandites) formed by open-system fractionation from evolved hawaiites (transitional basalt II) and that further fractionation resulted in comendite. On the other hand, le Roex & Erlank (1982) suggested that porphyritic hawaiites represented 'regular Bouvetøya hawaiite' with as much as 40% accumulation of plagioclase (34–37%) and clinopyroxene (3–6%). The only problem with this interpretation was an inferred Ba inconsistency (le Roex & Erlank 1982, p. 333). These authors rejected a model in which comendite (rhyolite) was formed from benmoreite, both because it required a fractionating plagioclase of An 47 (benmoreite plagioclase is An 33) and because there was a 'Ba and Sr problem', such that none of the minerals present as phenocrysts in benmoreite could remove enough barium (and strontium) to obtain the lower levels of these elements observed in comendite. Instead, they suggested that comendite was formed directly from mugearite, a very long fractionating step (69%), in which a plagioclase of An 47 was required. However, this is in conflict with the composition of modal plagioclase in mugearite (most calcic plagioclase in mugearite is An 45; le Roex & Erlank (1982), Table III, p. 321). The fractionation model of le Roex & Erlank (1982) can be summarized as follows: (1) hawaiite → mugearite → benmoreite, (2) mugearite → rhyolite, i.e. benmoreite is not an intermediate composition in the formation of rhyolite.

In this study, we address several of the problems connected with the models proposed in the two mentioned studies. Since our radiogenic isotope data (see below) are consistent with a model in which the various rock types at Bouvetøya are closely related, we present a refined crystal fractionation model, based on new mineral and chemical data.

Geochemical features

Analytical methods

Major elements and the trace elements Ba, Rb, Sr, Y, Nb, Cr, Ni, V, Cu, Zn, and Ga were analysed by XRF at Washington State University, Pullman, Washington, USA. Loss on ignition and volatiles on all samples as well as Co (XRF) on samples B 55, B 57, B 58, and B 59 were measured at NTNU, Trondheim, Norway. Scandium, Co, REE, Hf, Ta, and Th on most samples and plagioclase separates were analysed by INAA and Zr by XRF at the University of Oregon, Eugene, Oregon, USA. On samples B 55, B 57, B 58, and B 59, Sc, REE, Hf, Ta, Th, and U were determined by INAA at Imperial College – Reactor Centre, London, UK. Rare earth elements, Sc, Hf, Ta, and Th of sample B 31 were analysed by INAA at the University of Texas at El Paso, USA. Mineral analyses were obtained at NTNU/ SINTEF, Trondheim, Norway using a JEOL microprobe. Oxygen isotopes were measured at the University of South Carolina, Columbia, South Carolina, USA, and some strontium isotopes at University of Rochester, Rochester, New York, USA, while the rest of the Sr isotopes as well as the Nd and Pb isotopes were analysed at the University of North Carolina, Chapel Hill, North Carolina, USA.

Chemical data of fresh whole rocks are presented in Table 1. However, before we discuss the general data we will comment on the fact that in this study, the Na₂O contents of all rock types are consistently higher than those found in the previous studies (Table 2). Given the considerable significance of sodium concentrations in the model calculations, this aspect of our data has been thoroughly studied. Our Na₂O calibration is good for the 0.03–4.38 wt% calibration range (r = 0.99916). However, the intermediate and silicic rocks from Bouvetøya are high in Na₂O. As a check of calibration at higher values, we analysed the international standard STM-1 and an internal standard TMS as unknowns and got 9.15 wt%, which, relatively, is 2.3% higher than the recommended value 8.94 for STM-1 (Govindaraju 1989), and 9.52 wt% for TMS, which, relatively, is 2.0% higher than our preferred value of 9.33 wt%. Even though both of these results are on the high side of recommended or used values, we feel that these are acceptable data.

We have no satisfactory explanation for this interla-
### Table 1: Chemical composition of fresh volcanic rocks from Bouvet Island. Major element data normalized to 100%, volatiles are added.

<table>
<thead>
<tr>
<th></th>
<th>B 51</th>
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<th>B 56</th>
<th>B 57</th>
<th>B 58</th>
<th>B 59</th>
<th>B 62</th>
<th>B 70</th>
<th>B 79</th>
<th>B 28</th>
<th>B 67</th>
<th>B 16</th>
<th>B 44</th>
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<td>50.68</td>
<td>50.98</td>
<td>51.02</td>
<td>50.49</td>
<td>50.40</td>
<td>51.64</td>
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<td>3.52</td>
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<td>0.32</td>
<td>0.23</td>
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<td>4.44</td>
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<td>4.04</td>
<td>4.49</td>
<td>4.64</td>
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<td>8.65</td>
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<td>3.89</td>
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<td>1.18</td>
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<td>0.28</td>
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<td>0.28</td>
<td>0.28</td>
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<td>101.84</td>
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</table>

* = L.O.I.
Table 2. Measured Na2O values for corresponding K2O contents. Before calculating the averages, all data were normalized to 100% volatile free and with total iron as FeO.

<table>
<thead>
<tr>
<th></th>
<th>Hawaiites</th>
<th></th>
<th>Benmoreite</th>
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<th>Comendite</th>
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<td>(3)</td>
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<td>Na2O</td>
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<td>5</td>
<td>3</td>
<td>2</td>
<td>3</td>
<td>5</td>
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</table>

(1) This study; XRF - lithium tetraborate:rock powder = 2:1.
(2) Imsland et al. (1977); atomic absorption spectrometry.
(3) le Roex & Erlank (1982); XRF - data from Verwoerd et al. (1976).

In some laboratories, however, we think it is partly related to alteration (see below) and perhaps to a dilution effect through using strongly porphyritic samples in some studies. Thus, the altered rocks we have studied (volatiles > 2%) are all significantly lower in sodium than their fresh equivalents (Fig. 3e, Table 3). It is therefore possible that

Table 3. Altered basaltic and intermediate rocks. Major element data normalized to 100%, volatiles are added.

<table>
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* = L.O.I.

Ba 165 159 160 208 218 299 371 559 209
Rb 11 6 14 36 3 14 23 12 58 52
Sr 551 510 407 750 500 524 436 550 309 107
Y 42 43 41 34 37 20 53 58 71 55
Zr 285 260 277 251 262 153 366 425 620 476
 Nb 47 46 46 39 41 22.6 61 68 101 73
La 27.4 30.5 25.5 21.5 25.4 13.1 39.5 42.1 58.8 37.0
Ce 63.4 65.6 60.8 49 55.4 32.7 79.0 105 137 90.5
Nd 42 38.8 35.9 31.5 42 26 47.7 64 75 48.5
Sm 8.79 9.2 7.92 7.07 8.1 4.7 11.3 11.4 14.2 9.9
Eu 2.77 3.23 2.68 2.33 2.52 1.6 3.01 3.94 4.34 2.76
Tb 1.38 1.38 1.23 1.06 1.32 0.67 1.74 2.1 2.50 1.67
Yb 3.44 3.75 3.5 2.8 3.16 1.9 4.20 4.68 6.97 5.1
Lu 0.53 0.44 0.43 0.40 0.46 0.25 0.64 0.69 0.98 0.72
Sc 27.7 24.9 25.8 25.2 27.1 21.9 22.6 15.1 12.4 14.7
Cr 27 7 13 31 28 107 18 4 4 6
Co 30.7 38.6 34.5 35.2 35.7 32.1 29.8 14.3 3.52 16.6
Ni 5 0 2 14 15 49 0 0 5 8
V 363 316 322 301 293 195 269 88 13 147
Cu 25 14 38 38 48 53 6 26 2 12
Zn 145 153 123 113 122 70 156 171 211 118
Ga 27 22 20 27 22 21 23 26 30 21
Hf 6.02 5.36 6.13 4.95 5.49 3.21 8.36 9.20 13.6 10.7
Ta 2.45 2.27 2.38 2.26 2.18 1.24 3.49 3.46 5.14 3.60
Th 2.58 2.64 2.77 2.06 2.47 1.38 4.25 4.98 7.93 5.64
Y/Nb 0.89 0.93 0.89 0.87 0.90 0.88 0.87 0.85 0.70 0.75
Zr/Nb 6.06 5.65 6.02 6.44 6.39 6.77 6.00 6.25 6.14 6.52
 δ18O  + +1.9 +2.8 +1.9 +6.0 +6.2 7.2 +7.2 +5.8 +5.1
Q 3.14 7.90 7.15 5.98 1.60 7.20 3.61 9.79 36.61 2.75
Ne
most of the discrepancies might be due to the use of altered and/or porphyritic samples in the previous studies.

**General characteristics**

Both this and the former studies show that the K$_2$O contents of hawaiites vary considerably: from 0.75 to 1.65 wt% (Imsland et al. 1977), 0.72 to 1.31 wt% (le Roex & Erlank 1982) and from 0.97 to 1.54 in this study. The low values are almost always associated with high Al$_2$O$_3$ (>15%) and CaO (>10%) contents, sometimes with ‘high’ MgO (samples B 24 & B 70, this study) and relatively low values of several other elements. These are the effects of plagioclase (and sometimes olivine) accumulation as revealed by the porphyritic nature of many hawaiite samples. The sparsely porphyritic or aphyric Bouvet hawaiites, which are thought to be closer to melt compositions, seem to vary in K$_2$O in the range 1.12–1.54 wt%.

They are high in TiO$_2$ (3.35–3.89 wt%) and total FeO (11.15–12.81 wt%), and low in MgO (3.72–4.78 wt%) and CaO (7.85–9.48 wt%). We have only used such samples in our model calculations.

In Figs. 2 & 3 we have plotted selected major and trace elements in variation diagrams versus Ta, which was apparently incompatible throughout the development of the suite and also rather immobile during alteration. Three different pattern styles are observed: (a) Purely incompatible elements (vs. the observed minerals) such as Nb (Fig. 2a) vary almost linearly with Ta throughout the series. This type of pattern is observed for Zr, Hf, Y, La, Lu, K$_2$O, and Rb, while Ba increases linearly up to the trachyte stage (B 16) after which it drops considerably to and among the comendites (Fig. 2b). (b) FeO (Fig 2c), P$_2$O$_5$ (Fig. 2d), TiO$_2$ (Fig. 2e), MnO (Fig. 2f), and Sc (Fig. 3f) increase from the most primitive rocks and reach a maximum before decreasing in the intermediate and silicic rocks.
Fig. 3. Variation diagram for K2O, Rb, CaO, Sr, Na2O and Sc vs. Ta. Symbols as in Fig. 2.

Note, for example, the pattern of P2O5 contents, which increases linearly to mugearite (B 31 & B 34), indicating that apatite fractionation must have been rather insignificant up to this point, but from then on it decreases in more evolved compositions. TiO2 reaches its maximum somewhat earlier than FeO; we can also see (Fig. 2c, f) that manganese removal is delayed compared to iron, a feature compatible with the increased Mn/Fe ratio of the various ferromagnesian minerals we have analysed in benmoreite and comendite (Table 5). (c) CaO (Fig. 3c), MgO and Sr (Fig. 3d) decrease almost linearly throughout the series.

Alteration

Nine samples are listed in Table 3, all of which are altered by hydrothermal processes. Of these, samples B 10, B 19, B 20, B 22, and B 23 seem to represent 'normal' to evolved Bouvetøya type hawaiite, B 24 represents a hawaiite with accumulated plagioclase, B 31 and B 34 are altered mugearites, and B 49 is altered benmoreite. With the exception of B 23 and B 24, the altered samples were collected from a hydrothermally altered sequence, which is exposed in the cliffs behind the newly formed platform Nyrøysa (<100 m a.s.l.) on the west coast of Bouvetøya. Samples B 23 and B 24 were collected at the site of the active fumaroles in the north of the island (see Prestvik & Winsnes 1981 for details).

Chemically, the alteration is characterized by large, but variable H2O+ and CO2 contents in most samples. The secondary minerals include calcite, celadonite, chlorite, various clay minerals, pyrite, and quartz.

Elements such as Nb (Fig. 2a), Hf, Zr, Y, the REE, and Sc (Fig. 3f) were rather immobile during the alteration (assuming that Ta was also immobile). It is mentioned above that sodium is low in the altered samples (Fig. 3e). With the exception of one sample (B 22), K2O and Rb have
been leached during hydrothermal alteration (Fig. 3a, b). With a few exceptions Ba, P2O5, TiO2, and FeO appear to have been less mobile. CaO, Sr (Fig. 3c, d) and MnO (Fig. 2f) show enrichment in most altered hawaiites. One sample (a hawaiite sill; B 22, the most heavily altered of all samples in terms of total volatiles) is especially notable because it is the only sample considerably enriched in K2O, Rb, Ba, Sr and Ca. In this sample, all phenocrysts and most groundmass plagioclase are completely replaced by sericite ± calcite, and other phases have been transformed into a mixture of clay minerals. Vesicles have been filled with calcite, chlorite, various clay minerals and sometimes quartz.

Oxygen isotope data (Table 7) show low $\delta^{18}O$ values in the heavily altered samples, as low as +1.9‰ in B 22 and B 19. However, in samples B 23 and B 24, $\delta^{18}O$ is +6.0‰ and +6.2‰, respectively, and +7.2‰ in B 34. The significantly higher values of B 23 and B 24 may be related to their location in the fumarole area, while the samples with the lowest $\delta^{18}O$ are from the hyaloclastite section. Apparently, the fumarolic water has other isotopic signatures than the fluids that altered the lower hyaloclastite formation of the island. Average $\delta^{18}O$ for five fresh hawaiites is +5.5‰. Comparing our data on altered rocks with the data on Rb, Sr, Na2O and K2O published by O’Nions & Pankhurst (1974) and O’Nions et al. (1977) on hawaiite and benmoreite from Bouvetøya, we can conclude that some of the samples analysed by these authors represent heavily altered material. However, this alteration does not appear to have affected the $^{87}Sr/^{86}Sr$ and $^{143}Nd/^{144}Nd$ ratios.

Mineral compositions

Several hundred mineral grains have been analysed by electron microprobe. Some olivine, pyroxene and amphibole data are summarized in Table 5 and Fig. 4.

The feldspars show large compositional variation. In the hawaiites they range from An 83 to An 45 with An 60–75 as the most common composition. It is difficult, however, to evaluate the equilibrium plagioclase composition of the hawaiites, but in the sparsely porphyritic samples An 55–62 are common compositions. Plagioclase phenocrysts of the benmoreites are typically around An 30, and microlites grade into the field of anorthoclase. The feldspar of the trachyte (B 16) is anorthoclase, whereas the comendites have a very Ca-poor feldspar; the most potassium-rich varieties are close to the alkali-feldspar minimum melt composition, i.e. Or 35.

Plagioclase separates have been analysed for some trace elements (Table 4). The small concentrations of Fe, Sc, Cr, and Co show that very clean separates were
Fractionation within hawaiites (B 59-B 79)

- Olivine Fo 70
  - 0.71% Si
  - 51.640
  - 51.882

- Clinopyroxene Fs 17
  - 8.84% Ti
  - 3.410
  - 3.431

- Plagioclase An 56
  - 13.42% Al
  - 14.140
  - 14.126

- Ilmenite
  - 1.72% Fe
  - 12.390
  - 12.495

- Titanomagnetite
  - 0.46% Mn
  - 0.210
  - 0.202

Hawaiite B 79
- 74.85% Mg
- 3.720
- 3.741

Fayalite (Fo 0.2, Table 5) is found in several comendite samples.

Rare earth elements (REE) obtained. Especially interesting is the large Ba content of benmoreite plagioclase (397 ppm) compared to 35 and 56 ppm in plagioclase of hawaiites. This contrast shows that the distribution coefficient of Ba (\(D_{Ba}^{\text{plag/liquid}}\)) is increasing from about 0.2 to 0.72 as An contents (and T of equilibrium) decrease. The benmoreite plagioclase (An 33) is considerably richer in the REE than are accumulated bytownites (An 71–75) in hawaiites (see below).

The clinopyroxene data are shown in Fig. 4, and the averages of representative pyroxenes and olivines of the intermediate and silicic rocks are presented in Table 5. Pyroxene of benmoreite is Fe-rich augite, and in trachyte and comendite almost pure hedenbergite is present. Because most hawaiite pyroxenes are irregularly zoned, it is difficult to estimate a reliable 'average' composition. It is, however, important to recognize the compositional variation, because that sheds light on the magma's pre-eruption history.

Olivine varies from Fo 87 to Fo 45 in the hawaiites, but compositions around Fo 70 seem to be most common. These may represent equilibrium olivine, because a range of Fo 75–65 would be expected for the actual whole-rock Mg numbers (Roeder & Emslie 1970). The most magnesium-rich olivine that occurs in the hawaiites is, however, petrogenetically important because it shows the existence of more primitive magmas at deeper levels. The olivine phenocrysts of benmoreite are around Fo 30, whereas pure fayalite (Fo 0.2, Table 5) is found in several comendite samples.

Hedenbergite and fayalite have not been reported from Bouvetøya previously. In both of these minerals the MnO content is relatively large (Table 5). Thus, MnO contents range to 2.59 wt% in olivine of comendite sample B 83.

Imsland et al. (1977) reported rare blue-green alkali amphibole in some comendite samples. We present microprobe data of some amphiboles from a fresh holocry stalline comendite (Table 5). This is arvedsonite according to the classification of Leake (1978).

Rare earth elements (REE). – In the petrologic models of Imsland et al. (1977) and le Roex & Erlank (1982), as well as in our own models (Table 6), plagioclase fractionation is important. Extensive plagioclase fractionation is usually thought to deplete the residual melt in Eu because this element readily replaces Ca (and Sr) in the feldspar lattice.

Plagioclase separates of hawaiite and benmoreite all show a significant positive Eu anomaly (Fig. 5) where Eu is enriched about 10 times compared to the neighbour elements in hawaiite and about 30 times in benmoreite. As can be seen from Fig. 5, benmoreite does not display negative Eu anomalies despite the approximately 24%
Table 7. Isotope data for Bouvetøya rocks.

<table>
<thead>
<tr>
<th>Sample</th>
<th>$\delta^{18}$O**</th>
<th>$^{87}$Sr/$^{86}$Sr</th>
<th>$^{143}$Nd/$^{144}$Nd</th>
<th>$^{206}$Pb/$^{204}$Pb</th>
<th>$^{207}$Pb/$^{204}$Pb</th>
<th>$^{208}$Pb/$^{204}$Pb</th>
</tr>
</thead>
<tbody>
<tr>
<td>Hawaiite B 56</td>
<td>+5.8</td>
<td>0.703626</td>
<td>0.512895</td>
<td>ND</td>
<td>ND</td>
<td>ND</td>
</tr>
<tr>
<td>Hawaiite B 59</td>
<td>ND</td>
<td>0.703695</td>
<td>0.512961</td>
<td>ND</td>
<td>ND</td>
<td>ND</td>
</tr>
<tr>
<td>Hawaiite B 62</td>
<td>+5.4</td>
<td>0.703629</td>
<td>0.512919</td>
<td>ND</td>
<td>ND</td>
<td>ND</td>
</tr>
<tr>
<td>Hawaiite B 79</td>
<td>+6.0</td>
<td>0.703575</td>
<td>0.512891</td>
<td>19.105</td>
<td>15.671</td>
<td>38.836</td>
</tr>
<tr>
<td>Benmor. B 28</td>
<td>+6.5</td>
<td>0.703609</td>
<td>0.512887</td>
<td>ND</td>
<td>ND</td>
<td>ND</td>
</tr>
<tr>
<td>Hawaiite B 2</td>
<td>+5.9</td>
<td>ND</td>
<td>0.703695</td>
<td>ND</td>
<td>ND</td>
<td>ND</td>
</tr>
<tr>
<td>Hawaiite B 6</td>
<td>+5.4</td>
<td>0.703629</td>
<td>0.512919</td>
<td>ND</td>
<td>ND</td>
<td>ND</td>
</tr>
<tr>
<td>Hawaiite B 7</td>
<td>+6.0</td>
<td>0.703575</td>
<td>0.512891</td>
<td>19.105</td>
<td>15.671</td>
<td>38.836</td>
</tr>
<tr>
<td>Benmor. B 2</td>
<td>+6.5</td>
<td>0.703609</td>
<td>0.512887</td>
<td>ND</td>
<td>ND</td>
<td>ND</td>
</tr>
<tr>
<td>Hawaiite B 1</td>
<td>+4.2</td>
<td>0.703675</td>
<td>0.512883</td>
<td>ND</td>
<td>ND</td>
<td>ND</td>
</tr>
<tr>
<td>Hawaiite B 2</td>
<td>+1.9</td>
<td>0.703739</td>
<td>0.512850</td>
<td>19.420</td>
<td>15.630</td>
<td>39.053</td>
</tr>
<tr>
<td>Comendite B 39</td>
<td>+7.7</td>
<td>0.703754</td>
<td>0.512852</td>
<td>19.606</td>
<td>15.636</td>
<td>39.165</td>
</tr>
<tr>
<td>Comendite B 44</td>
<td>+7.4</td>
<td>0.703615</td>
<td>0.512900</td>
<td>ND</td>
<td>ND</td>
<td>ND</td>
</tr>
<tr>
<td>*Hawaiite B 10</td>
<td>+4.2</td>
<td>0.703675</td>
<td>0.512883</td>
<td>ND</td>
<td>ND</td>
<td>ND</td>
</tr>
<tr>
<td>*Hawaiite B 2</td>
<td>+1.9</td>
<td>0.703739</td>
<td>0.512850</td>
<td>19.420</td>
<td>15.630</td>
<td>39.053</td>
</tr>
<tr>
<td>*Mugear. B 34</td>
<td>+7.2</td>
<td>0.703524</td>
<td>0.512888</td>
<td>ND</td>
<td>ND</td>
<td>ND</td>
</tr>
<tr>
<td>*Benmor. B 49</td>
<td>+5.8</td>
<td>0.703665</td>
<td>0.512845</td>
<td>19.687</td>
<td>15.723</td>
<td>39.434</td>
</tr>
</tbody>
</table>

* Altered samples.
** Additional $\delta^{18}$O determinations on two fresh hawaiites are +5.0 and +5.5, and from +1.9 to +6.2 in 5 altered hawaiites (see also Table 3).
ND = not determined.

plagioclase fractionation that is indicated from the petrologic models (Table 6; model II). Intuitively, and as was stated by Dickey et al. (1977), one should expect a considerable effect on the Eu content of the derived melt when Eu-enriched plagioclase is removed. For the Bouvetøya suite, therefore, two questions arise: (i) Is it possible to fractionate a feldspar with a pronounced positive Eu anomaly without developing a negative anomaly in the derived melt, and/or (ii) Is the petrologic model principally wrong, perhaps because plagioclase fractionation was not that important? The answers appear to be very simple: the plots in Fig. 5 also show another important feature; the bytownites of hawaiite have very small concentrations of REE compared to those of the whole rock compositions. Even the apparently large Eu values are significantly lower than in the host hawaiites, i.e. the plot is a visual way of showing that Eu has a small aggregate distribution coefficient (for Eu$^{3+}$ and Eu$^{2+}$ combined) between basic plagioclase and a mafic melt, a fact that has been known for many years. A similar relationship exists between the plagioclase (An 33) and the host benmoreite (Fig. 5) except that in this case the Eu content is larger than in the melt the plagioclase crystallized from; i.e. $D_{Eu}$ is greater than unity (about 1.3), and therefore further fractionation may be very efficient in developing negative Eu anomalies in the more silicic melts. It is thus logical that trachyte (B 16), which is somewhat less silicic than the comendite, displays an REE pattern which is parallel with that of comendite, but has smaller REE abundances and a less pronounced negative Eu anomaly.

It was mentioned above that Eu$^{2+}$ is expected to have geochemical behaviour similar to Sr. However, Eu occurs...
in both the $2^+$ and the $3^+$ states. Thus, the aggregate distribution coefficient for Eu depends on the prevailing oxidation states. Distribution coefficients are also temperature dependent (Blundy & Wood 1991); they increase with decreasing temperature. If we apply distribution coefficients between plagioclase and melt for Eu$^{2+}$ and Eu$^{3+}$ similar to those for Sr$^{2+}$ and REE$^{3+}$, respectively, we can estimate the proportion of the two Eu ions in the melt. If we distribute the observed aggregate $D_{\text{Eu}}$ of 0.2 between bytownite plagioclase and hawaiite with $D_{\text{Eu}^{2+}} = 1$ and $D_{\text{Eu}^{3+}} = 0.03$, we find that the Eu$^{3+}$/Eu$^{2+}$ ratio in the melt is ca. 4.7; in other words, Eu is rather oxidized. Choosing a higher coefficient for Eu$^{2+}$ ($D_{\text{plagioclase/melt}}$ range between 1 and 3) would correspond to an even more oxidized melt. Making use of the same procedure for andesine/benmoreite, using aggregate $D_{\text{Eu}} = 1.3$ (as observed), $D_{\text{Eu}^{3+}} = 0.05$, and $D_{\text{Eu}^{2+}} = 2.55$ gives a Eu$^{3+}$/Eu$^{2+}$ ratio of 0.85, i.e. considerably more reduced than in the basic melts. These results are in general agreement with the conclusion drawn by Prestvik (1982a) on a qualitative basis. If we compare with the experimental data of Weill & Drake (1973), we find that $f_{\text{O}_2}$ for bytownite/hawaiite at 1100°C is about $10^{-3}$ Pa and about $10^{-5.5}$ Pa for andesine/benmoreite at 1000°C. For comparison, le Roex & Erlank (1982) estimated $f_{\text{O}_2}$ for hawaiite at 1000°C to be about $10^{-6.0}$ Pa on the basis of coexisting Ti-magnetite and ilmenite.

New petrologic model

We used a common least squares mass balance mixing technique (Bryan et al. 1969) to evaluate the possible relationships between the observed rock and mineral compositions. Before modelling, all data were normalized to 100% volatile-free, and total iron was expressed as FeO. In models involving silicic rocks (models IV and V), Mg was omitted in model IV, because the poor analytical accuracy of whole rock analyses with MgO $\leq 0.10\%$ made error estimates meaningless. Mn was omitted from models IV and V. Our models are presented in Table 6.

We made the calculations in several steps. As described above, there is considerable chemical variation within the hawaiite group, and we wanted to include this variation in our modelling. We chose sample B 59 ($K_2O = 1.18$) as representative of the least evolved hawaiites and B 79 ($K_2O = 1.54$ and transitional to mugearite) as the most evolved. Both these groups are almost non-porphryitic. Models IA and IB (Table 6) represent variation within the group of hawaiites. Models II and III show the evolution from evolved hawaiite (B 79) to benmoreite (B 28; $K_2O = 2.97$) and furthermore to trachyte (B 16; $K_2O = 4.50$). The evolution of comendite (B 39; $K_2O = 4.93$) is presented in two models, model IV from trachyte and model V directly from benmoreite. With one exception (IB), we present models with low residuals ($\Sigma^R < 0.5$), but we also rejected models with low residuals if the relative discrepancy for Na and K was $>1.5\%$.

Our preferred sequence of evolution of the Bouvet suite is as follows: hawaiite $\rightarrow$ evolved hawaiite $\rightarrow$ benmoreite $\rightarrow$ trachyte $\rightarrow$ comendite (models IA, II, III & IV), see discussion below.
Discussion

Former models

We agree with Imsland et al. (1977) that the irregularly zoned and commonly corroded basic plagioclase mega­
crysts in the hawaiites are not real phenocrysts. On the basis of mixing calculations, le Roex & Erlank (1982) suggested that such phryic hawaiites represented regular Bouvetøya hawaiite with as much as 40% accumulation of plagioclase (34–37%) and clinopyroxene (3–6%). As mentioned above, the only problem with this interpretation was an inferred Ba inconsistency (le Roex & Erlank 1982, p. 333). However, the very small Ba concentrations obtained by us on hawaiite plagioclase separates (Table 4) explain and eliminate this problem.

Transitional basalts I of Imsland et al. (1977) correspond well to ‘accumulated’ hawaiite of le Roex & Erlank (1982). This rock type is represented by samples such as B 24 and B 56 (the hawaiite samples from which we separated plagioclase). However, the strongly phryic hawaiites not only have An-rich plagioclase, but also tend to have the most primitive clinopyroxene and olivine, i.e. a mineral assemblage representing more primitive melts. Our data show that Mg numbers (Mg#s) of the hawaiites from Bouvetøya range from 36 to 55 (not tabulated). The largest Mg#s are found in the phryic hawaiites. If these rocks were formed by accumulation of pheno/xenocrysts, as suggested by le Roex & Erlank (1982), it would require that as much as 15% cpx + ol was added to hawaiite melt (B 59) in addition to plagioclase. Unless significant resorption occurred, this seems to be unlikely, based on observed hawaiite modal composition. We favour an interpretation in which more primitive hawaiitic magmas (than those represented by the aphyric hawaiites) were present and fractionated at depth, and that only porphyritic varieties are represented in our sample collection. Sometimes, also more evolved hawaiite melts accumulated phenocrysts (especially plagioclase) from their basic precursors and brought these to the surface.

There is a tendency, however, for a gradual composi­
tional change also within the group of sparsely phryic or non-phryic hawaiites. Low Mg#s and generally small concentrations of compatible elements such as Ni, Co, Sc, and Cr in these hawaiites (Table 1) strongly indicate a history in which olivine and clinopyroxene (+ spinel?) fractionated from the more primitive magmas. Pre­
fractionation of plagioclase is not documented by negative Eu anomalies in REE patterns of the basic rocks, because plagioclase has such small Eu concentrations (Table 4). However, the strong, negative anomaly of strontium in the spidergram of the Bouvetøya hawaiite (Fig. 6) indicates that parental magmas at depth also crystallized plagioclase, because distribution coefficients of Sr are large enough (1–3; Henderson 1982) effectively to remove this element during plagioclase fractionation.

New models

We present two fractionation models (Table 6, IA & IB) to explain compositional variation in the hawaiites. Both suggest ca. 25% fractionation of ol + cpx + plag + ilm + mt, all of which occur in hawaiite. The difference in the two models is the plagioclase composi­
tion (An 56 in IA; An 65 in IB). However, the unacceptably high residual in model IB makes that model
unrealistic. Even though modal plagioclase in many hawaiites falls in the range An 60 – 75, we did not consider models using plagioclase more calcic than An 65. As was discussed above, the true equilibrium plagioclase composition may well be lower; and we think the plagioclase (An 56) used in model IA is a realistic choice. The composition of olivine and clinopyroxene is as observed in hawaiite. Thus, IA is our preferred model for this stage. The amount of fractionation is consistent with the observed incompatible element variation.

In model II, benmoreite is derived from evolved hawaiite by slightly more than 50% fractionation. This is about 10–15% less than reported by both Imsland et al. (1977) and le Roex & Erlank (1982). However, the parent hawaiites used in the models of these authors were less evolved than those we have used. The model uses minerals and compositions that are typical in evolved hawaiite, except that the plagioclase is slightly more sodic than is common in hawaiite. Since this step is so large, we find it reasonable that the average fractionating plagioclase is intermediate between parent and daughter compositions. Olivine and clinopyroxene compositions are both fairly magnesium-rich, but that can be explained by postulating that these minerals fractionated early in the sequence. This model has very low residuals. Model III shows how trachyte can be formed by ca. 38% fractionation from benmoreite. Only benmoreite mineral compositions were used in this model, which has a somewhat high residual. This is mainly due to fits in the Al and Fe contents (both ca. 4.5% relative). We still believe this is an acceptable model for the derivation of trachyte, not least because trachyte is intimately related to benmoreite in the field. The trachyte is relatively enriched in incompatible elements such as Ba, Rb, Y, Zr and Nb compared to benmoreite, but strontium is considerably lower (89 vs. ca. 315 ppm). This requires bulk D\text{Sr} and D\text{Ba} of ca. 3.6 and 0.5, respectively. With the actual proportion of plagioclase in the fractionating assemblage, this gives D\text{Ba}^{\text{ plag}} of 5.4, which is reasonable for plagioclase An 31 (Nash & Creecraft 1985), and D\text{Ba}^{\text{ plag}} = 0.72, which is identical to calculated values from the analysed plagioclase separate from benmoreite (Table 4).

In models IV and V (Table 6), we have sought to determine how further fractionation from trachyte (IV) or benmoreite (V) can yield comendite. In model IV, there is a considerable deficiency in Na content (4.49% relative) of the derived comendite. In this model, we used mineral compositions from the trachyte, including alkali feldspar. In order to avoid the Na deficiency, a much less sodic and unrealistic feldspar composition was attempted, which in turn created problems with other elements. In model V, mineral compositions of phenocrysts in benmoreite were used, and the fit is almost perfect for all elements. However, le Roex & Erlank (1982) rejected their model in which comendite could be formed from benmoreite, because the model required a fractionating plagioclase of An 47 (An 31 in our model V), and because there was a 'Ba and Sr problem'. Instead, they suggested that comendite was formed directly from mugearite, a very long fractionating step (69%) in which a plagioclase of An 47 was required.

The barium (and strontium) problem

Prestvik (1982a) reported Ba contents of three comendites (183–216–245 ppm) from Bouvetøya. Similarly, le Roex & Erlank (1982) reported 180 and 298 ppm Ba in two comendites. In this work, we obtained 312 ppm Ba in comendite sample B 39 (obsidian). However, our crystalline comendites are much more depleted in Ba (and Sr), with only 80 ppm Ba. Apparently, the 'barium problem' would have been even larger if these samples had been used in the models. From the whole rock data we have calculated the value of D\text{Ba} and D\text{Sr} for the fractionating feldspar of models IV and V (the comendite models). In model IV (formation of comendite from trachyte) we get D\text{Ba} = 3.8 and D\text{Sr} = 5.7. The fractionating feldspar in this model is about 2/3 alkali feldspar and 1/3 anorthoclase (An 10). The estimated D’s are well within the possible ranges for both elements (Henderson 1982; Blundy & Wood 1991). For model V (formation of comendite directly from benmoreite) the values of D\text{Ba} and D\text{Sr} for fractionating feldspar are 2.8 and 4.2, respectively. For plagioclase (An 31) this is a plausible value, but the D\text{Ba} value is much larger than that found by us for this plagioclase when we analysed the mineral separate (D\text{Ba} = 0.72). We thus agree with le Roex & Erlank (1982) that comendite did not form by fractionation directly from benmoreite. We disagree, however, that the comendite was formed by fractionation directly from mugearite. Despite the Na discrepancy of the comendite derived by fractionation of trachyte, we think that model IV provides a plausible explanation for the way in which Bouvetøya comendites are derived. The very high Ba content of trachyte – and much lower content in comendite – explains how the benmoreite magmas fractionate through a trachyte stage and shows that there is a shift from plagioclase to alkali feldspar as a fractionating phase around the trachyte stage. Furthermore, the ferromagnesian minerals are much more evolved in trachyte than benmoreite. Fractionation of these minerals not only results in comenditic composition, but also explains the trace element data, such as the Ba and Sr concentration. We have no valid explanation for the Na discrepancy seen in model IV, but the model is, of course, ‘vulnerable’, owing to the fact that we have only one sample of trachyte.

We have also tried to model (not tabulated) the relationship between comendite B 39 (glass) and crystalline comendite. This is possible by approximately 7% fractionation of alkali feldspar and minor amounts (ca. 2%) of ol + cpx + mt if D\text{feldspar} for both Sr and Ba is about 19. This is within the published range for Sr, but higher than the value we have seen published for Ba (up to 12.9; Henderson (1982)).
Mantle signature

On the basis of Y – Ti – Zr ratios Prestvik (1982b) characterized the basic rocks from Bouvetøya as ‘within plate’ type (i.e. influenced by a mantle plume in more modern terms). Based on trace element ratios such as Y/Nb, Zr/Nb, (La/Yb)N, le Roex (1987) characterized them as identical to (P)lume-type MORB of the American–Antarctic Ridge (AAR) and South-West Indian Ridge, which he suggested had formed by partial melting of depleted asthenosphere that had been veined by enriched melt (2–4%) in a previous melting event. Our new data show that the geochemical signatures displayed by the Bouvetøya suite (Fig. 6) are those of a ‘mildly enriched’ ocean island type (Thompson et al. 1984), with typical high normalized abundances of the HFS elements Nb and Ta. The average Y/Nb and Zr/Nb ratios of Bouvetøya hawaiites, which are 0.86 and 6.06, respectively (Table 1), also attest to their enriched nature.

Although there is a considerable amount of Sr and Nd isotope data for the ridge axes of the BTJ area (e.g. Dickey et al. 1977; le Roex et al. 1983; 1985; 1987), very few such data are published from Bouvetøya (O’Nions et al. 1977), and the only known Pb isotope data from Bouvetøya are three determinations of Sun 1974 (1980). Here we report new Sr and Nd (as well as O) isotopic data on 11 samples from Bouvetøya and Pb isotope data on 5 samples (Table 7), all collected during the Polarsirkel expedition. In Fig. 7 we have plotted some of the new data together with the previously published Bouvetøya data (O’Nions & Pankhurst 1974; O’Nions et al. 1977; Sun 1980). The Bouvetøya rocks fall on, or very close to, mixing lines between EM-1 and HIMU basalts in both 207Pb/204Pb vs. 206Pb/204Pb, 208Pb/204Pb vs. 206Pb/204Pb and 87Sr/86Sr vs. 206Pb/204Pb plots (Fig. 7) and very close to such mixing trends in 143Nd/144Nd vs. 206Pb/204Pb and 143Nd/144Nd vs. 87Sr/86Sr diagrams (not shown). However, the combined isotope data (e.g. the 87Sr/86Sr vs. 206Pb/204Pb plot) are not compatible with mixing between a depleted mantle source (DMM in Fig. 7) and enriched sources such as EM-1 (and EM-2) or HIMU sources. This implies that the model of le Roex (1987) has to be modified.

Enrichment was probably integral to the plume. HIMU signatures are primarily seen from elevated U/Pb ratios because of loss of Pb (Vidal 1992). Unfortunately, we lack elemental Pb data on the Bouvetøya rocks, but typical HIMU signatures are apparent also from the Ce/Rb and Ba/Nb ratios of hawaiites (average 3.13 and 4.78, respectively, of 9 samples).

Dredged samples from the South-West Indian Ridge immediately east of Bouvet (Fig. 1) are isotopically almost identical to the Bouvetøya rocks and apparently related to the Bouvet plume as well. However, the regional influence of the Bouvet plume is not well understood. The possible presence of more than one plume in the BTJ area has been discussed by several authors (i.e. le Roex 1987). Thus, Moreira et al. (1995), Douglas et al. (1995) and Simonov et al. (1996) indicated that the anomalously enriched segment of the southern MAR east of the Shona seamount (the ‘Shona ridge-anomaly’, Fig. 1) – only about 600 km from Bouvetøya – might represent a separate Shona plume.

Conclusions

Our study supports previously published models which concluded that the main rock types of Bouvetøya are interrelated by fractional crystallization. However, we find that the best evolutionary path of the suite was as follows: hawaiite → evolved hawaiite → benmoreite → trachyte → comendite.

Primitive megacrysts of olivine, clinopyroxene and plagioclase in hawaiite represent phenocrysts of more primitive (parental) magmas that crystallized at depth, indicating the presence of an open-system magma chamber(s) in which variously evolved melts and crystal assemblages were able to mix.

Similarity in trace element and isotope ratios indicate that the Bouvetøya suite, and the much more primitive basalts dredged from the Bouvet Ridge east of the island, originated from a common, enriched mantle source. The radiogenic isotope data show that this source is complex, consisting of several mantle components.

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