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Basic Volcanism Associated with Intraplate Linear Features [and Discussion]

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Basic volcanism associated with intraplate linear features

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Intraplate volcanic lineaments include ocean island chains and continental rift systems. Basic lavas erupted in such lineaments form a continuum from tholeiitic basalt in the basements of ocean islands to nephelinites and melilitites in continental rifts and as a capping on ocean islands. All these magma types are enriched in large-ion lithophile elements (l.i.l.e.) compared with mid-ocean ridge basalts (m.o.r.b.), although isotopic data suggest that their mantle sources had been depleted in l.i.l.e. for long periods. In this paper we present a comparison of geochemical data from several suites of basic volcanic rocks ranging from Hawaiian tholeiite to Ugandan melilitite. L.i.l.e. abundance patterns can, in most cases, be explained by variable degrees of melting of a l.i.l.e.-depleted m.o.r.b. mantle source containing l.i.l.e.-rich streaks. Metasomatic enrichment of the mantle source is not a necessary precursor to magmatism.

1. Introduction

Intraplate volcanism is commonly associated with linear features. In the ocean basins these usually take the form of chains of islands and sea mounts thought to be produced by the movement of lithosphere plates over asthenosphere 'hot spots'. Hot-spot trails also occur on the continents but are much less common, probably because continental lithosphere is thicker and therefore less easily penetrable than oceanic lithosphere. The most frequent occurrence of continental intraplate volcanism is in continental rift systems where deep fractures have permitted the escape of magma. Whether the asthenosphere upwelling responsible for rift magmatism is the cause of rifting or simply a consequence of it is still debated (Morgan & Baker 1983). Whatever the cause, asthenosphere upwelling seems to be responsible for both ocean island and continental rift magmatism and we should expect magmas in the two associations to have some geochemical features in common.

Intraplate basaltic rocks, both oceanic and continental, are strikingly enriched in large-ion lithophile elements (l.i.l.e.) by comparison with basalts erupted at mid-ocean ridges (m.o.r.b.). The large degree of asthenosphere melting inferred to take place in a relatively shallow melting zone beneath mid-ocean ridges allows the composition of the m.o.r.b. source to be closely constrained. This source must be depleted in l.i.l.e. (for example, K, Rb, Ba and light rare-earth elements) in comparison with the bulk earth. Isotope ratios of Sr and Nd show that the m.o.r.b. source must have had low Rb/Sr and Nd/Sm ratios (and by implication have been depleted in other l.i.l.e.) for considerable periods of time. Isotope ratios in most intraplate basic volcanic rocks also imply a long-term l.i.l.e.-depleted source, though not quite so depleted as the m.o.r.b. source. This observation has posed a major problem in the understanding of intraplate volcanism. How can such l.i.l.e.-rich magmas be generated from a l.i.l.e.-depleted source?

One solution involves very small degrees of partial melting (less than 1%) to concentrate l.i.l.e. (which are incompatible in mantle phases) into the melt. Such small melt fractions have been considered impossible to extract and simple melting models have been rejected as a consequence. Complex melting processes such as zone refining (Harris 1957) and wall-rock reaction (Green & Ringwood 1967) have been proposed to account for the concentration of K and other incompatible elements into a melt. More recent attempts to account for the high l.i.l.e. concentrations in intraplate volcanic rocks have invoked metasomatic enrichment of the mantle source shortly before melting. The current popularity of mantle metasomatism has led to a proliferation of *ad hoc* hypotheses involving variable degrees of enrichment through the agency of CO_2- and H_2O-rich fluids originating in deeper parts of the mantle. Such hypotheses have been applied to both oceanic (see, for example, Wright 1984) and continental (see, for example, Lloyd & Bailey 1975; Frey *et al.* 1978; Menzies & Murthy 1980, Bailey 1982) intraplate magmas.

That parts of the mantle have been enriched in l.i.l.e. is clearly shown by mantle xenoliths contained in alkali basalts and kimberlites. These xenoliths almost certainly originate in the lithosphere. It is also clear that many of the geochemical characteristics of these enriched mantle fragments can be seen in continental tholeiites (Hawkesworth *et al.* 1984). These tholeiites are the products of extensive melting involving not only the asthenosphere but probably also the mantle part of the continental lithosphere and perhaps even the crust (see, for example, Menzies *et al.* 1984). The continental lithosphere has evolved over hundreds or thousands of millions of years and will bear the imprint of many metamorphic and magmatic events. Melting the continental lithosphere will impart distinctive geochemical signatures to the magmas produced and can account for the wide range of l.i.l.e. concentrations and isotope ratios found in continental flood basalts. However, it is not clear that all intraplate magmas are the products of enriched mantle.

The continental lithosphere is the only part of the Earth where enriched mantle is likely to be preserved for very long periods. The oceanic lithosphere is young and at least as depleted in l.i.l.e. as the asthenosphere, which accounts for the scarcity of metasomatized mantle xenoliths in oceanic volcanic rocks. The asthenosphere will be well stirred by convection and therefore homogeneous on the scale of individual convection cells. If metasomatically enriched mantle plays an important role in the generation of magmas,then we would expect significant geochemical differences between oceanic and continental intraplate basic volcanic rocks. If such differences do not exist then an asthenosphere source is implied, and the case for mantle metasomatism as a necessary precursor to intraplate magmatism will be weakened. A unique area in which to test this hypothesis is provided by a remarkable volcanic lineament known as the Cameroon Line.

2. Cameroon Line, West Africa

The Cameroon Line (figure 1) is a chain of Tertiary to Recent alkaline volcanoes and plutonic complexes extending for 1600 km from the Atlantic Ocean to the interior of Africa. It is unique among intraplate volcanic provinces in comprising both oceanic and continental volcanoes. The Cameroon Line is not a hot-spot trace because there is no evidence for any consistent migration of volcanism with time. Volcanic activity appears to have occurred sporadically along the whole length of the line over its 65 Ma history (Fitton & Dunlop 1985). The siting of the volcanoes shows no evidence of basement control because the line cuts across oceanic transform faults and major fault zones in the continental crust (figure 1). Fitton (1980, 1983) has suggested that the line is related to rifting in the nearby Benue trough.

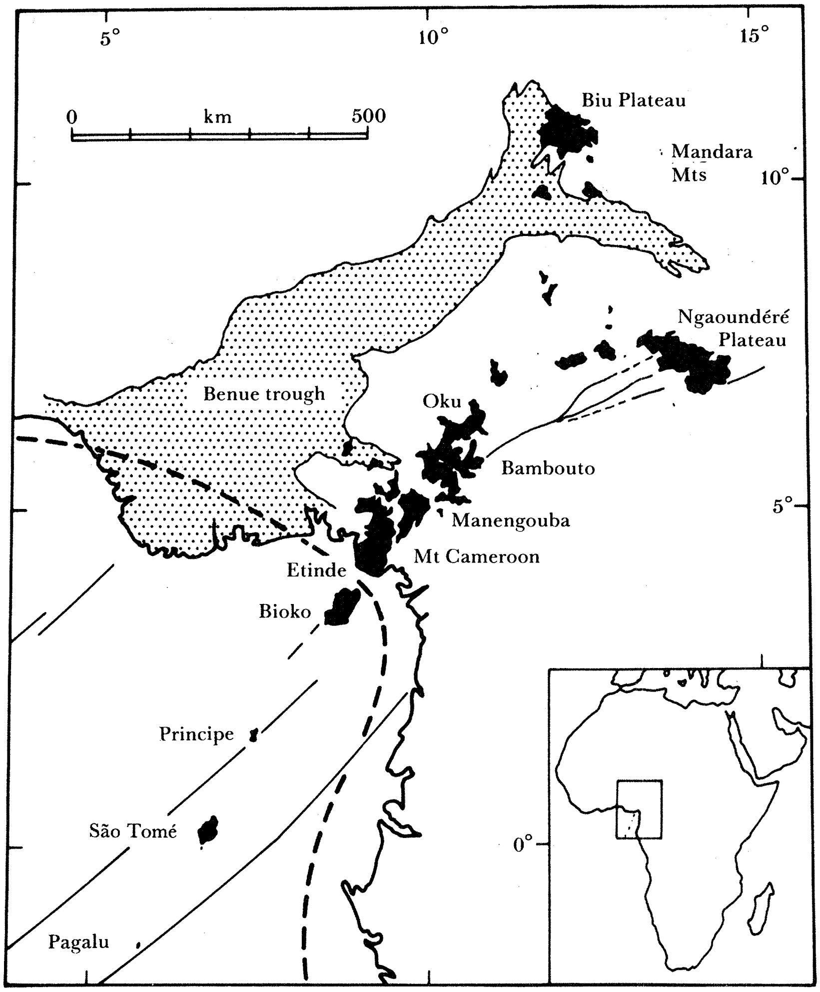

FIGURE 1. Geological map of part of West Africa showing the distribution of the Cameroon Line volcanic rocks (black). The boundary between oceanic and continental crust (heavy broken line) is taken from Emery and Uchupi (1984, chart XI). Oceanic transform faults and the major fault system between Bambouto and Ngaoundéré are from Sibuet & Mascle (1978) and Gazel (1956) respectively.

In an attempt to detect systematic differences between oceanic and continental intraplate volcanism, Fitton & Dunlop (1985) analysed a large number of basic volcanic rocks from the oceanic and continental sectors of the Cameroon Line for major and trace elements and Sr isotope ratios. Surprisingly, the two populations of basalt are chemically and isotopically indistinguishable. This is illustrated in figure 2, where the average concentrations of incompatible elements in basalts from the two sectors are shown normalized to chondrite and undepleted mantle abundances (Sun 1980). The data imply that the oceanic and continental basalts had very similar mantle sources. Because it is most unlikely that the young lithosphere mantle beneath the oceanic sector is chemically and isotopically similar to the ancient lithosphere mantle beneath the continental sector, it follows that lithospheric mantle was not the source of the Cameroon Line basalts. The lack of any consistent migration of volcanism with time rules out a source below the 670 km discontinuity. If the Cameroon Line were the product of a deep mantle plume it would require that lower mantle convection has kept pace exactly, in both

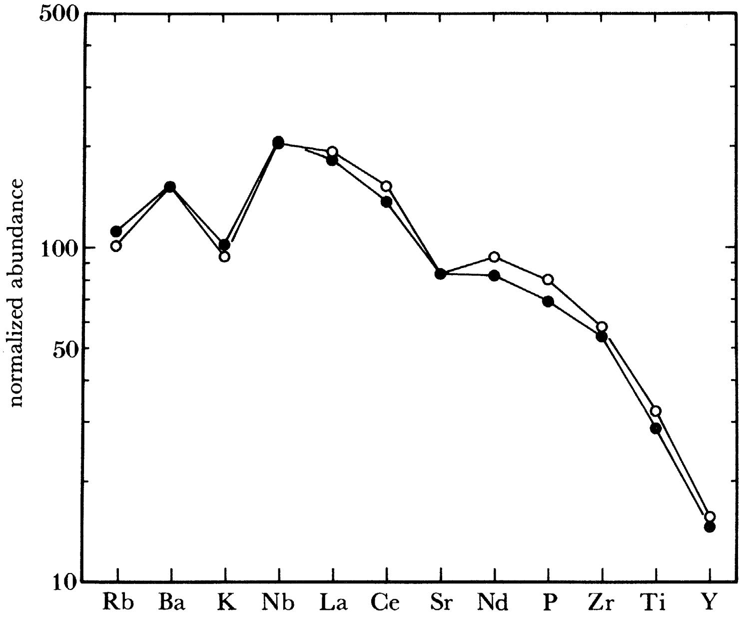

FIGURE 2. Average incompatible-element abundances in basic volcanic rocks (more than 4% MgO) from (○) the oceanic (144 analyses) and (●) continental (134 analyses) sectors of the Cameroon Line. Concentrations have been normalized to chondrite and primitive mantle abundances (Sun 1980).

velocity and direction, with the movement of the African plate over the past 65 Ma. The convecting upper mantle is, therefore, the only plausible mantle source.

The convecting upper mantle (of which the asthenosphere is the upper part) is also the source of m.o.r.b. Fitton & Dunlop (1985) proposed that the l.i.l.e.-rich Cameroon Line basalts could be derived from the l.i.l.e.-depleted m.o.r.b. source by small-degree (*ca.* 0.2%) melting, by using bulk partition coefficients consistent with experimentally determined values for mantle silicate phases. They suggested that the convecting upper mantle, although l.i.l.e.-depleted in its bulk composition, contains l.i.l.e.-rich streaks. Partial melts from these streaks will be selectively incorporated into small-degree melts and can account for the small but consistent isotopic differences between m.o.r.b. and intraplate basalts. The extraction of apparently small-degree melts can be accomplished by the migration of larger degree melts along mantle grain boundaries, where they will leach l.i.l.e.-rich grain boundary films. McKenzie (1984, 1985) has recently derived a set of equations describing the compaction of partly molten rock and concluded that melt will begin to flow at very small degrees of melting (less than 0.5%), providing its viscosity is low. The degrees of melting required to produce l.i.l.e.-rich intraplate basalts from the m.o.r.b. source should no longer be regarded as impossibly small.

3. COMPARISON WITH OTHER INTRAPLATE VOLCANIC PROVINCES

If the Cameroon Line basalts can be produced from the m.o.r.b. source, the same must be true for volcanic rocks from other intraplate provinces. To test this hypothesis we have analysed several suites of basic volcanic rocks, chosen to cover a very wide range of compositions and tectonic environments. Ocean islands are represented by tholeiite lavas from historic eruptions of Mauna Loa (Hawaii), transitional basalts from Réunion and nephelinite and melilitite lavas

Table 1. Average compositions of representative intraplate basic volcanic rocks (standard deviations in parentheses)

	Mauna Loa[1]	Réunion[2]	Oahu[3]	Cameroon Line[4]	Jos Plateau[5]	Kisingiri[6]	Toro-Ankole[7]
				percentage by mass			
SiO_2	51.71 (0.62)	47.14 (0.90)	37.94 (1.67)	45.12 (2.33)	46.77 (1.62)	41.17 (2.10)	36.17 (0.35)
Al_2O_3	13.51 (0.39)	13.43 (1.22)	10.77 (0.55)	13.73 (1.79)	14.51 (0.79)	10.91 (2.18)	6.36 (0.28)
Fe_2O_3[8]	12.02 (0.17)	12.41 (1.54)	15.82 (1.58)	12.77 (1.22)	11.54 (0.72)	14.22 (1.14)	12.55 (0.24)
MgO	7.84 (0.78)	9.65 (2.30)	12.35 (1.30)	8.34 (2.67)	9.00 (1.15)	7.09 (2.03)	13.83 (0.12)
CaO	10.38 (0.37)	10.98 (2.28)	12.51 (0.33)	10.17 (1.47)	9.30 (0.52)	14.03 (2.54)	16.86 (0.40)
Na_2O	2.28 (0.04)	2.47 (0.66)	4.50 (0.71)	3.19 (0.76)	3.28 (0.50)	3.21 (1.42)	1.14 (0.20)
K_2O	0.37 (0.03)	0.75 (0.22)	1.52 (0.41)	1.41 (0.47)	1.59 (0.46)	1.72 (0.76)	4.13 (0.45)
TiO_2	2.03 (0.08)	2.49 (0.52)	2.63 (0.17)	3.17 (0.53)	2.33 (0.30)	3.24 (0.43)	4.45 (0.18)
MnO	0.17 (0.00)	0.17 (0.02)	0.23 (0.02)	0.18 (0.02)	0.17 (0.03)	0.23 (0.02)	0.22 (0.01)
P_2O_5	0.23 (0.01)	0.30 (0.09)	1.03 (0.21)	0.79 (0.25)	0.57 (0.16)	0.85 (0.16)	0.99 (0.03)
total	100.54	99.79	99.30	98.87	99.06	96.67	96.70
				p.p.m. (by mass)			
Ni	—	242 (121)	281 (46)	153 (103)	168 (53)	90 (133)	246 (8)
Cr	393 (75)	520 (267)	402(102)	253 (182)	234 (97)	128 (145)	611 (27)
V	—	297 (38)	291 (25)	263 (64)	160 (16)	326 (51)	203 (34)
Sc	29.9 (0.7)	32.6 (5.1)	18.7 (8.1)	22.2 (6.4)	18.5 (2.3)	19.5 (10.1)	29.9 (0.7)
Cu	—	78 (14)	75 (6)	48 (19)	50 (12)	180 (62)	134 (4)
Zn	—	96 (17)	161 (46)	113 (14)	105 (12)	115 (13)	101 (4)
Sr	303 (23)	366 (85)	1400 (315)	919 (260)	753 (183)	1485 (486)	2774 (133)
Rb	5.5 (0.7)	17.4 (6.6)	39.0 (11.0)	37.1 (17.7)	42.6 (14.9)	47.3 (15.0)	120 (2)
Zr	126 (5)	182 (49)	218 (63)	313 (86)	233 (61)	291 (44)	362 (20)
Nb	8.5 (0.5)	22.3 (6.1)	65.5 (22.5)	72 (22)	60 (17)	104 (24)	240 (12)
Ba	76 (4)	172 (49)	914 (122)	580 (179)	623 (169)	1367 (1192)	2426 (38)
La	8.6 (0.6)	18.7 (6.5)	65.0 (9.6)	59 (20)	43 (14)	85 (22)	191 (16)
Ce	22.9 (1.3)	41.8 (15.7)	137.7 (19.5)	118 (39)	84 (27)	170 (43)	384 (41)
Nd	18.1	22.6 (7.1)	63.3 (9.7)	53 (15)	37 (10)	69 (16)	135 (14)
Y	23.0 (0.0)	26.1 (5.0)	26.8 (2.0)	30.3 (4.7)	27.1 (9.5)	29.7 (4.5)	19.0 (1.2)

[1] Average of 6 analyses of tholeiite lavas from historic eruptions of Mauna Loa, Hawaii (Basaltic Volcanism Study Project 1981; pp. 166 and 173, HAW-5 omitted). Nd abundance has been estimated from the Mauna Loa R.E.E. data of Budahn & Schmitt (1985).

[2] Average of 20 analyses of transitional basalt lavas from Réunion, Indian Ocean.

[3] Average of 6 analyses of nephelinite and melilitite lavas from the Honolulu Series of Oahu.

[4] Average of 278 analyses of basic volcanic rocks (more than 4% MgO) from the Cameroon Line.

[5] Average of 13 analyses of basaltic lavas from the Jos Plateau, Nigeria.

[6] Average of 32 analyses of nephelinite and melilitite lavas (more than 4% MgO) from Kisingiri, Kenya.

[7] Average of 3 analyses of olivine melilitite (katungite) from Katunga, S.W. Uganda. Samples C4078, C4409 and C4414 collected by A. D. Combe.

[8] Total Fe as Fe_2O_3.

from the Honolulu Volcanic Series (H.V.S.) of Oahu. A suite of Miocene nephelinite and melilitite lavas from Kisingiri in the Kavirondo branch of the Kenya Rift and three samples of Pliocene olivine melilitite (katungite) from Katunga (Uganda) in the Western Rift were taken to represent extreme compositions from continental rift valleys. The Jos Plateau (Nigeria) suite comprises Pliocene and Pleistocene basalts from an area of uplift thought to be related to continental hot-spot activity (Bowden *et al.* 1976). All the rock samples, with the exception of the Mauna Loa tholeiites, were analysed in Edinburgh by X-ray fluorescence techniques (Fitton & Dunlop 1985).

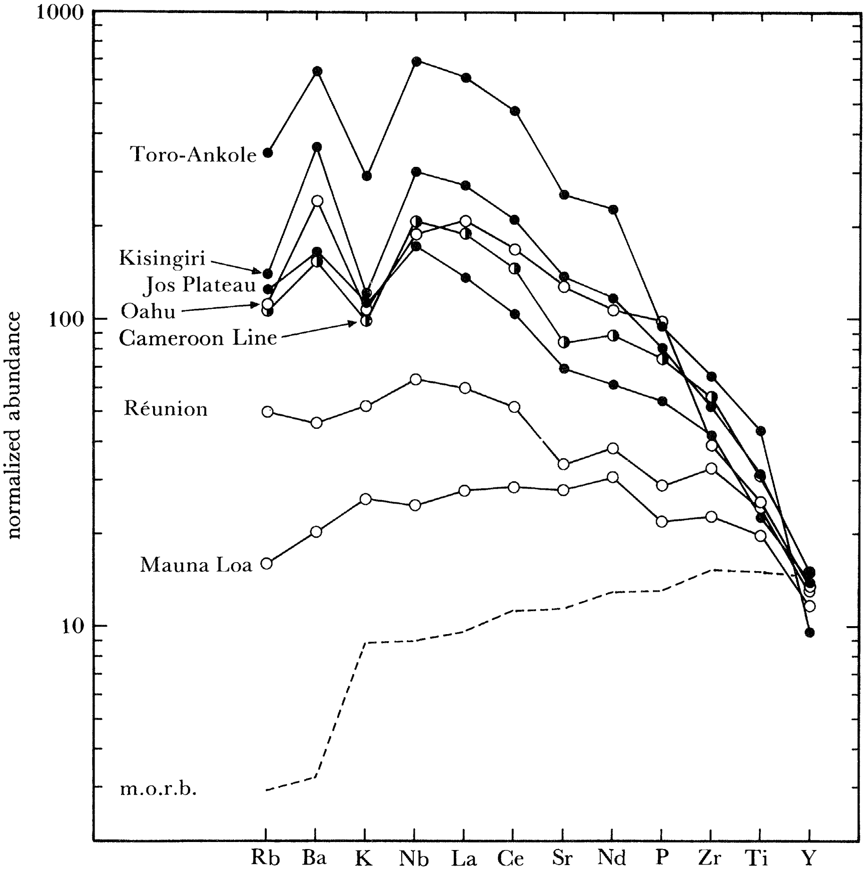

Figure 3. A comparison of normalized incompatible-element abundances in several suites of intraplate basic volcanic rocks (table 1). Open and solid symbols represent oceanic and continental suites respectively. The Cameroon Line data include analyses from both the oceanic and continental sectors. Typical m.o.r.b. abundances (Sun 1980) are shown for comparison.

Average compositions of these basic rock suites are presented in table 1 along with an average of the Cameroon Line data. Normalized incompatible-element abundance patterns for all the suites are shown on figure 3. A typical m.o.r.b. pattern (Sun 1980) is shown for comparison. It is clear that the patterns form a compositional continuum ranging from Hawaiian tholeiite to Ugandan melilitite. In §4 we shall discuss the possibility that these compositions can be derived from a common m.o.r.b. source.

4. Partial melting models

It was noted in §3 that the incompatible element concentrations in the Cameroon Line magmas can be modelled by small-degree melting of a m.o.r.b. source with the use of reasonable bulk partition coefficients (D). If this conclusion is applicable to intraplate magmas in general, then it should be possible to reproduce the patterns in figure 3 by varying the fraction of melting (F) in the equilibrium partial-melting equation

$$C_1/C_0 = 1/(D+F-DF),$$

where C_1 is the concentration of an element in the liquid, C_0 its concentration in the source and D the bulk partition coefficient between the residual solid phases and the liquid. Strictly speaking, this equation is only applicable when the solid phases enter the melt in their original proportions (i.e. modal melting) but it can also be used for non-modal melting providing D is small (Shaw 1979). In the course of melting, D will vary as phases are consumed, but for small degrees of melting it can be assumed to be constant.

The average compositions given in table 1 represent magmas which have undergone variable degrees of crystal fractionation. It is therefore necessary to estimate the composition of their respective primary magmas before attempting to model the partial melting process. To do this, one must assume that the magmas have crystallized only olivine since segregating from the mantle. This is a reasonable assumption for the Mauna Loa tholeiite (Wright 1984) but is less certain for the other compositions. Consequently, only the most magnesian compositions (Oahu and Toro-Ankole) from table 1 were used in the calculations. An average of 86 Cameroon Line basalt analyses with more than 10% MgO was also used.

To estimate the primary magma composition, Fe_2O_3/FeO is first normalized to 0.15 and the equilibrium olivine calculated by using a K_D value of 0.3 (Roeder & Emslie 1970). A small amount (0.1%) of this olivine is added and a new magma composition calculated. The process is repeated until the magma would be in equilibrium with mantle olivine (assumed to be Fo_{92}). Finally the trace-element abundance in the primary magma is calculated by 'diluting' the analytical data with the total amount of olivine added. The resulting primary magma compositions are given in table 2. These are, of course, model dependent, but they have all been normalized to be in equilibrium with the same mantle olivine and errors in the assumptions made (mantle olivine composition and initial oxidation state) will tend to cancel out during the subsequent modelling calculations.

The next stage is to calculate melting parameters for one set of data and then apply these to the other sets. The Cameroon data have been used because there is good evidence that the lavas had an asthenosphere source. Published D values cannot be used to model the partial melting process because most are not known to better than an order of magnitude. At the degrees of melting involved, order-of-magnitude variations in calculated compositions would be produced. Instead we have assumed a small-degree partial melting model and used the equilibrium melting equation to calculate D values required to derive the Cameroon Line primary magma from a m.o.r.b. source (Fitton & Dunlop 1985) over a range of F. Positive values of D for all elements are obtained only for $0 < F < 0.0044$ (0–0.44% melting). Because values of D and F close to zero are unrealistic, we have used a set of D values for $F = 0.003$ in subsequent calculations. These are consistent with the range of measured D values for likely

Table 2. Calculated primary magma compositions

	Cameroon Line	Mauna Loa	Oahu	Toro-Ankole
		percentage by mass		
SiO_2	42.95	48.62	38.77	36.84
Al_2O_3	9.14	9.73	7.63	5.44
Fe_2O_3	1.22	1.03	1.33	1.28
FeO	10.39	9.91	11.88	9.82
MgO	20.13	19.19	23.01	19.03
CaO	8.51	7.48	8.86	14.43
Na_2O	2.01	1.64	3.19	0.98
K_2O	0.88	0.27	1.07	3.54
TiO_2	2.39	1.46	1.86	3.81
MnO	0.14	0.12	0.16	0.19
P_2O_5	0.57	0.17	0.73	0.85
total	98.33	99.62	98.49	96.21
		p.p.m. (by mass)		
Sr	639	218	992	2375
Rb	24.4	4.0	27.6	102
Zr	202	91	155	310
Nb	48	6	46	206
Ba	409	55	648	2077
La	40	6.2	46	163
Ce	78	16.5	98	328
Nd	36	13.0	45	116
Y	21.3	16.6	19.0	16.3
olivine added (percentage by mass)	29.9	38.8	41.1	16.8

Primary magma compositions have been calculated by adding 0.1 % increments of equilibrium olivine (using $K_D = 0.3$; Roeder & Emslie 1970) until the equilibrium olivine reaches Fo_{92}. Fe_2O_3:FeO was normalized to 0.15 before the addition of olivine.

The calculations were based on the data in table 1 except for the Cameroon Line primary magma, for which an average of 86 analyses of basalt with more than 10 % MgO was used. This average has SiO_2, 43.58; Al_2O_3, 11.87; total Fe as Fe_2O_3, 13.28; MgO, 11.43; CaO, 11.05; Na_2O, 2.61; K_2O, 1.14; TiO_2, 3.10; MnO, 0.18; P_2O_5, 0.74; all as percentages by mass.

mantle phases and are appropriate for a garnet–lherzolite mantle source containing a small amount of a potassic phase (see Fitton & Dunlop 1985, Fig. 11).

The *D* values so obtained were then used to calculate incompatible-element abundances for liquids derived by a range of degrees of melting ($F = 0.001$–0.05). Normalized abundance patterns for these liquids are shown in figure 4 compared with patterns for the Mauna Loa, Oahu (H.V.S.) and Toro-Ankole primary magmas. The Cameroon Line primary magma data are not plotted on this diagram because they were used in its construction. It is clear from figure 4 that, to a first approximation, the Mauna Loa, Oahu (H.V.S.) and Cameroon Line primary magmas could all have been derived by the partial melting of similar m.o.r.b. sources. There is no need to postulate source enrichment before melting. The degrees of melting implied by these calculations (*ca.* 0.1 % for Oahu nephelinites to *ca.* 7 % for Mauna Loa tholeiites) are not absolute values but depend upon the degree of melting (0.3 %) assumed for the Cameroon Line data. Varying this figure within the permitted range (0–0.44 %) will not greatly affect these degrees of melting.

Several factors could be responsible for the degree of misfit in figure 4. First, the upper mantle beneath the Hawaiian islands will have been convecting independently of that beneath the

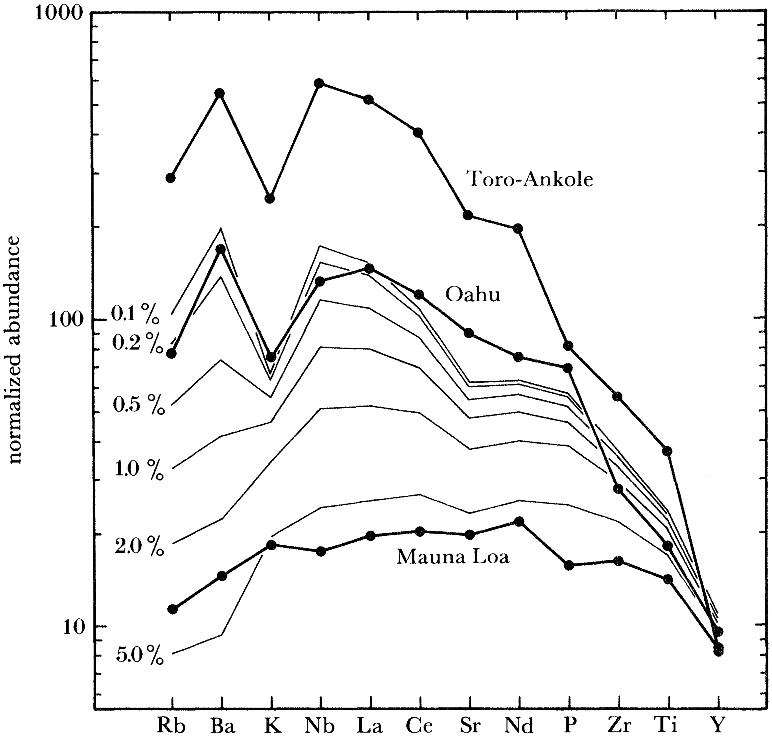

FIGURE 4. Normalized incompatible-element abundances in calculated primary magmas (table 2) compared with normalized abundance patterns generated by various degrees of melting of a m.o.r.b. source. The *D* values used in the melting calculations were obtained by assuming that the Cameroon Line primary magma was derived by 0.3 % equilibrium partial melting of the m.o.r.b. source: Rb, 0.00316; Ba, 0.00141; K, 0.0189; Nb, 0.00696; La, 0.00881; Ce, 0.0149; Sr, 0.0275; Nd, 0.0308; P, 0.0353; Zr, 0.0640; Ti, 0.111; Y, 0.237. These are comparable with published *D* values for likely mantle phases (Fitton & Dunlop 1985).

Cameroon Line for very long periods, and may have evolved to a slightly different composition. Global variations in m.o.r.b. and ocean island basalt compositions have been well documented by Dupré & Allègre (1983) and Hart (1984). Secondly, bulk *D* values will vary with the modal mineralogy of the mantle, which is likely to change with depth and degree of melting. Leeman *et al.* (1980), for example, have suggested that the small-scale compositional variation shown by Hawaiian tholeiites may be a result of variations in the ratio of garnet to clinopyroxene in the source. Finally, the assumptions made when calculating primary magma compositions are all subject to uncertainty. When all these factors are taken into account, the degree of fit shown by figure 4 is remarkable.

Because the patterns in figure 3 appear to be a related series it is likely that the Réunion, Jos and Kisingiri primary magmas would also fit the calculated patterns on figure 4. This is not true for the Toro-Ankole data, however, because the l.i.l.e. content of the calculated primary magma is much too high to be explained by melting of a m.o.r.b. source. One solution to this problem could be that the magmas have not undergone a small amount of olivine fractionation (table 2) but are instead the product of very large amounts of pyroxene fractionation. There is experimental evidence that clinopyroxene is on the liquidus of katungite magma at high pressures (Arima & Edgar 1983) but the high Cr and Sc contents of the lavas (table 1) makes the possibility of large amounts of clinopyroxene fractionation unlikely. We must conclude, therefore, that the Toro-Ankole lavas had a l.i.l.e.-enriched mantle source. There is isotopic evidence for enriched sources for many of the ultrapotassic lavas of the Western Rift (Bell & Powell 1969; Vollmer & Norry 1983). The Western Rift, unlike the Eastern Rift, cuts Archaean crust and is therefore probably underlain by anomalously thick and very ancient lithosphere.

5. Volume considerations

If we are correct in concluding that most intraplate primary magmas are derived by small-degree (0.1–7 %) partial melting of the asthenosphere and that source enrichment is not necessary, then very large volumes of mantle will need to be processed to generate the volumes of magma observed in some of the larger intraplate volcanoes. It is significant that the most l.i.l.e.-rich magmas (nephelinites and melilitites) account for only a minute proportion of intraplate volcanism and therefore do not present a volume problem. On the other hand, ocean island tholeiite and continental alkali basalt often form very large volcanoes. In this section we shall estimate the volumes of mantle that must have undergone partial melting to produce the Hawaiian-Emperor chain and Mt Cameroon, a large continental alkali basalt volcano.

The Hawaiian–Emperor chain is generally believed to be the product of melting within a rising mantle plume. The exposed part of the chain is composed largely of tholeiite. Alkali basalt and more undersaturated lavas form only a thin capping. By using the volumes calculated by Shaw *et al.* 1980) from bathymetry, Crisp (1984) has calculated an average rate of magma production of 1.5×10^{-2} km^3 per year. However, Shaw *et al.* (1980) point out that their volumes are underestimated by a factor of 1.3–2.2 because the mass of volcanic material has depressed the Moho beneath each volcano. This implies a true production rate of around 3×10^{-2} km^3 per year. There is evidence for the existence of large permanent magma reservoirs within the volcanic structure of Hawaii (Ryan *et al.* 1981) where crystal fractionation could take place, though some fractionation will almost certainly also occur at mantle depths (Wright 1971). The rate of primary magma production will therefore be between 3 and 4×10^{-2} km^3 per year. If melt is extracted from a cylindrical melt zone of radius r, then the minimum velocity (v) with which upwelling mantle must enter the melt zone will be given by

$$v = P/F\pi r^2,$$

where P is the rate of primary magma production and F the degree of melting. Density differences between mantle peridotite and solid basalt are neglected. Shaw *et al.* (1980) estimated that the melt zone is up to several hundred kilometres wide. Taking a conservative estimate of 100 km for r, a value of 0.07 for F (from figure 4), and primary magma production rates of $3–4 \times 10^{-2}$ km^3 per year gives upwelling velocities of about 1.4–1.8 cm per year. These are about half the velocity estimated by McKenzie (1984) for rising jets within a convecting mantle.

Mt Cameroon is one of the world's largest alkali basalt volcanoes with a height of 4095 m and an estimated volume of 1500 km^3. Its lavas are very similar in composition to the average Cameroon Line basalt (Fitton *et al.* 1983, table 1). The Mt Cameroon primary magma would, in our model, be produced by about 0.3 % partial melting of the asthenosphere. After correcting for olivine fractionation (assumed to take place in sub-volcanic magma reservoirs) the volume of primary magma involved would be about 2000 km^3, which means that 6.7×10^5 km^3 of mantle would be involved in magma generation. This is equivalent to a cylinder with a radius of 50 km (about twice the radius of the volcano at its base) and a height of 85 km. Such a volume would fit comfortably in the asthenosphere.

6. Nature of the mantle source

Our data suggest that a wide range of intraplate magmas share a common source within the convecting upper mantle. The possibility of a deeper mantle source (below the 670 km discontinuity) can be discounted for the Cameroon Line basalts because of the lack of migration of volcanism with time. Deep mantle plumes remain a possibility for other areas, however, although they are unnecessary in our model. The only evidence for any transport of material across the 670 km discontinuity comes from high $^3He/^4He$ ratios in some (but by no means all) ocean island basalts compared with m.o.r.b. (Kurz *et al.* 1982).

Because the convecting upper mantle must also be the m.o.r.b. source we have to account for the small but systematic isotopic differences between m.o.r.b. and intraplate basic volcanic rocks. These imply a heterogeneous source but provide no information on the scale of heterogeneity. Convection will ensure homogeneity on a large scale, but small-scale (metre to kilometre) heterogeneities may persist for long periods (Olsen *et al.* 1984). These may be generated in the course of intraplate magmatism or by the return of ocean crust (including ocean islands) to the mantle during subduction, and will be l.i.l.e.-rich in comparison with the bulk mantle. They will be drawn out into streaks by convection.

Over long periods these streaks will evolve to higher $^{87}Sr/^{86}Sr$ and lower $^{143}Nd/^{144}Nd$ ratios than will the bulk mantle. Because they will contain a higher proportion of basaltic components than the surrounding mantle, they will have lower solidus temperatures and will therefore be selectively incorporated into small-degree melts. Large-degree melting at mid-ocean ridges will effectively homogenize the mantle and produce m.o.r.b. with isotope ratios similar to the bulk mantle. The isotopic evolution of a streaky mantle and the effects that melting such a mantle will have on isotope ratios in the melts have been discussed by Fitton & Dunlop (1985).

Evidence for the existence of enriched streaks in a generally depleted mantle comes from global variations in isotope ratios in m.o.r.b. Batiza (1984) and Allègre *et al.* (1984) have commented upon the inverse correlation between diversity in $^{87}Sr/^{86}Sr$ and spreading rate. Both explain this as the result of variable rates of homogenization of a mantle source containing small-scale heterogeneities. A similar conclusion was reached by Zindler *et al.* (1984) from a study of chemical and isotopic variation among basalt samples dredged from young Pacific Ocean sea mounts.

The selective incorporation of streak-derived melt at low degrees of melting will result in magmas whose compositions are controlled more by streak mineralogy and composition than by the composition of the bulk mantle. However, the rate at which the melt moves through the mantle will be controlled by the proportion of melt (McKenzie 1985). Very small degree (and therefore l.i.l.e.-rich) melts will move slowly and tend to equilibrate with the bulk mantle. At the other end of the magmatic spectrum, large-degree melts (m.o.r.b.) will also have isotopic ratios close to those in the bulk mantle. In the middle, however, there may be degrees of melting small enough to incorporate streaks selectively but large enough for the melt to move rapidly in comparison with diffusion rates. Such magmas would have the highest $^{87}Sr/^{86}Sr$ and lowest

^{143}Nd/^{144}Nd ratios. This effect could explain why ocean island tholeiites often have higher ^{87}Sr/^{86}Sr ratios than do alkali basalts and nephelinites. Chen & Frey (1983) explain the inverse correlation between l.i.l.e. enrichment and ^{87}Sr/^{86}Sr in Hawaiian lavas by mixing small-degree melts from the m.o.r.b. source with larger-degree melts from an enriched mantle source. This explanation is consistent with our streaky mantle hypothesis.

If ocean island tholeiites preserve some of the isotopic characteristics of the streaks then the 1.5 Ga 'isochron' obtained from such rocks by Brooks *et al.* (1976) could represent the average age of the streaks. It is significant that oceanic alkali basalts do not show similar isochrons. It may also be significant that the greatest degree of misfit shown by the Mauna Loa tholeiite data on figure 4 is for Rb, Ba and K. These are some of the most incompatible elements and therefore those in which the streaks will be most enriched.

The model discussed here requires a common mantle source for m.o.r.b. and more enriched basalts such as those erupted in ocean islands and continental rifts. In this respect it differs from the commonly held view that m.o.r.b. and the other basalt types are derived from separate regions in the mantle (see, for example, Allègre 1982). In such models, m.o.r.b. are derived from a l.i.l.e.-depleted upper mantle and the other magmas from undepleted mantle below the 670 km discontinuity. However, they fail to explain why magmas with the unique composition of m.o.r.b. are confined entirely to mid-ocean ridges. One would, for example, expect m.o.r.b.-like magmas to be erupted occasionally in continental areas. Thompson *et al.* (1984) have proposed a model in which the m.o.r.b. source floats on less depleted asthenosphere and is restricted to the ocean basins by thicker lithosphere beneath the continents. The m.o.r.b. source migrates slowly into new oceanic areas (such as the Atlantic) from its principal reservoir beneath the Pacific Ocean. The model implies that the whole of the Atlantic Ocean floor north of Iceland and, presumably, the older parts of the ocean floor further south are not made of m.o.r.b.! One would also expect to find significant compositional differences between basic volcanic rocks erupted in the oceanic and continental sectors of the Cameroon Line (see figure 2).

Models involving complete separation of mantle sources do not provide an adequate explanation for the occurrence and composition of intraplate magmas and m.o.r.b. The model we propose can account for both the complete restriction of m.o.r.b.-like magmas to mid-ocean ridges, and the composition of ocean island and continental rift magmas. Mid-ocean ridges are the only places on earth where asthenosphere can undergo sufficiently large-degree, shallow melting to produce magmas whose chemical and isotopic characteristics are controlled by the *bulk* mantle composition. Elsewhere, small-degree melting with the selective incorporation of melt from enriched streaks will produce l.i.l.e.-rich magmas chemically and isotopically different from m.o.r.b.

7. Conclusions

We have shown that the incompatible-element geochemistry and volumetric abundance of basic volcanic rocks ranging from ocean island tholeiite to continental rift nephelinite and melilitite can be explained by variable degrees of melting of a common source in the convecting upper mantle. This extends from the base of the lithosphere to the 670 km discontinuity and must also, therefore, be the source for m.o.r.b. The systematic isotopic differences between m.o.r.b. and intraplate basic volcanic rocks requires the convecting upper mantle to be heterogeneous. The isotopic characteristics of ocean island basalt are consistent with their derivation from a l.i.l.e.-depleted mantle containing l.i.l.e.-rich streaks.

There is ample evidence for the existence of ancient enriched mantle in the continental lithosphere. This will contribute to intraplate magmas during extensive, high-level melting events such as the generation of continental flood basalts. It may also be involved during less extensive melting, beneath regions where the lithosphere is very old and therefore thick, as in the Western Rift of East Africa. The ultrapotassic, highly undersaturated magmas in this area require an enriched source. This is probably also true for extremely l.i.l.e.-rich magmas elsewhere. However, metasomatic enrichment of the mantle is not necessary to account for the geochemistry of ocean island and most continental rift basic volcanic rocks.

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Discussion

A. M. Quennell (*Department of Geology, University of Bristol*). This paper adds usefully to the concept of a lineament crossing an intraplate boundary between the oceanic and continental parts of a plate. The identity of specific types of volcanism with active or passive continental rifting is believed to be accepted. It is understood that the rifting of the region discussed is passive in nature but the volcanics described appear to conform more with those belonging to active rifting. East Africa has been mentioned and it is here that the distinction can be seen between volcanism of active rift zones, for example, the Kenya domal uplift and the Virunga volcanic field, and that of the passive rift zones, for example, south Gregory zone and Fort Portal. In the latter fields the nepheline–carbonatite volcanic association is distinctive, with its explosive vents and violent eruptions from central volcanoes.

J. G. Fitton. Dr Quennell's point about the distinction between volcanism in active and passive rift zones is an interesting one. However, we do not share his confidence in being able to recognize the two sorts of rift in the field. Şengör & Burke (1978) have pointed out that uplift, rifting and volcanism will occur in both but the sequence will be different in active and passive rifts. The relative timing of events in East Africa has not yet been unequivocally resolved and so one cannot confidently assign an active or passive origin to the rift system as a whole or to its component parts.

The character of rift volcanism and the nature of the products will be largely controlled by degree of melting in the mantle. Nephelinites and carbonatites will be the products of much smaller degrees of melting than will alkali basalts. We can see no compelling reason why the degree of mantle melting should be influenced by the type of rifting involved.

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