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An assessment of the relative roles of crust and mantle in magma genesis: an elemental approach

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The elemental compositions of terrestrial igneous rocks are reviewed with special emphasis on those elements that partition strongly into the liquids in mafic and ultramafic systems. Published data are supplemented by 79 new major- and trace-element analyses. The magmatism of ocean basins is considered in terms of a model that has the following main features: (i) density layering in the sub-lithospheric upper mantle, so that the more fertile source of ocean-island basalts (o.i.b.) underlies the less fertile source of mid-ocean ridge basalts (m.o.r.b.); (ii) the genesis of all mantle-derived magmas restricted to very small degrees of partial fusion; (iii) genesis of m.o.r.b. source mantle as residuum from the loss of a melt fraction (forming o.i.b. magmas and lithospheric veins) from o.i.b.-source mantle; (iv) subduction of o.i.b.-veined lithosphere, with a thin veneer of m.o.r.b. and sediments, to the 670 km seismic discontinuity, followed by re-heating of these components and their buoyant upwelling into the o.i.b.-source reservoir; (v) very little chemical communication across the 670 km discontinuity. All continental anorogenic magmatism (distant from subduction zones in space and time) seems to be related ultimately to the o.i.b.-source mantle reservoir, which therefore must extend beneath the lithospheric roots of continents. The minor sodic-alkalic magmatism of continents is effectively identical in composition to o.i.b. Some continental flood basalts are similar but the majority contain minor contamination (rarely more than 15%) from fusible sialic rocks. Although substantial amounts of sediments appear to be subducted, only a small proportion of them seems to re-appear in the products of island-arc and Cordilleran magmatism. Much larger sediment fractions enter the sparse ultrapotassic magmatism that occurs far behind some subcontinental subduction zones and also characteristically follows the subduction related magmatism of collisional orogenies. The remaining subducted sediments finally pass into the o.i.b.-mantle source reservoir. It is well established that, during and immediately after collisional orogeny, the fusion of sialic crust contributes substantially (or even occasionally exclusively) to batholithic magmatism. Nevertheless, the elemental variation in such magmas implies that the role of fractional crystallization in their genesis has tended to be underestimated in recent years. Mantle-derived mafic to ultramafic magmas appear to be directly or indirectly (as heat sources) involved at deep crustal levels in the parentage of most batholithic intermediate and acid magams. These mantle-derived liquids are subduction-related before continental collisions and then change to o.i.b., several million years after subduction ceases. Enhanced subduction of terrigenous sediments during the final stages of ocean closure leads to the large subducted sialic fractions which re-emerge in the ultrapotassic mafic magmas that characteristically appear immediately after a continental collision.

INTRODUCTION

This paper is a discussion of global magmatism, and focuses on elemental abundances in igneous rocks. We have supplemented published data by analysing approximately 120 samples from worldwide localities, that represent most of the important terrestrial igneous rock groups (as we see them). The analyses of 79 samples are given in table A 1 (see Appendix). Our data for 15 continental flood basalts are published elsewhere (Thompson *et al.* 1983).

Our discussion here will concentrate on the so called incompatible elements in igneous rocks because these are particularly useful when investigating the relative roles of crust and mantle in magma genesis. In most cases these data will be presented diagrammatically, normalized to chondritic abundances (e.g. figure 2) except for Rb, K and P which are normalized to 'primitive terrestrial mantle' values (Sun 1980). Details of the normalization procedure are discussed by Thompson (1982). In the present study we have 'double-normalized' the incompatible-element data for *basic* rocks by setting $\text{Yb}_\text{n} = 10.0$. This manipulation makes the shapes of patterns much easier to compare, by eliminating the crossings of lines caused by variable fractionation of such phases as olivine and pyroxene from the magmas. At a given value of MgO, all the different variants of mafic magmas have Yb contents varying by no more than a factor of about 2. The double-normalization procedure therefore oversimplifies the data somewhat but we think that the benefits of the technique outweigh the disadvantages.

Throughout this paper we use the rather complex normalized multi-element diagrams (e.g. figure 2) in preference to binary or ternary plots (see, for example, Pearce & Cann 1973) because the latter are very prone to ambiguous interpretation, as discussed by Morrison (1978), among others.

EXTENT OF MANTLE FUSION DURING MAGMA GENESIS

The melting of lherzolite upper mantle to basic and ultrabasic magmas has traditionally been envisaged as a cotectic process, in which substantial amounts of compositionally invariant liquid are produced immediately after the solidus temperature is reached. The minimum amount of liquid generated is taken to be a small percentage and the maximum, for mid-ocean-ridge basalts (m.o.r.b.), is set at up to 30% or more. Experiments on natural lherzolites have not fully supported this viewpoint. For instance, the data of Mysen & Kushiro (1977) show rapid growth of the melt fraction with rising temperature above the solidus but with *large progressive changes in the melt composition*, especially its silica-saturation (Thompson 1984).

Ahern & Turcotte (1979) considered the physics of melting and magma separation in upward-convecting mantle within the asthenosphere. They concluded that the generation of large melt fractions within lherzolite *undergoing recrystallization* was impossible. As soon as interconnected porosity is achieved (a figure of 2% melt was suggested), the liquid migrates very rapidly (1 to 10 m a^{-1} at 50 km depth) through the deforming matrix towards the surface. In order to simplify their calculations, Ahern & Turcotte assumed that the upper mantle beneath a mid-ocean ridge melts at a constant temperature and can yield 25% of basaltic magma. In reality the extent of partial fusion is temperature-dependent (Thompson 1984) and the quantity of basic magma that can potentially be extracted from such mantle may amount to not more than a small percentage (Loubet & Allègre 1982). The results of Ahern & Turcotte's calculations therefore suggest the possibility that episodes of mantle fusion may involve the production of only very small fractions (between 'a few' and less than 1%) of melt.

Is there any other evidence on this topic? Except for some discussion of the effects of surface tension on their analysis, Ahern & Turcotte's work has been ignored by almost all petrologists and geochemists, with the notable exception of Kay (1980). Nevertheless, isotope specialists have noted for many years that *elemental* ratios, such as Rb/Sr and Sm/Nd, in basic rocks are usually different from the values that would be required to produce the observed *isotopic* ratios, such as $^{87}Sr/^{86}Sr$ and $^{143}Nd/^{144}Nd$, in the same sample without fractionation of such element pairs as Rb from Sr and Sm from Nd during the genesis of the magmas or relatively shortly before (geologically speaking). The partition coefficients between such strongly-incompatible elements and lherzolite are so low (and similar) that they could only be fractionated during mantle fusion, if the melt fraction was extremely low: an order of magnitude smaller than envisaged by any quasi-cotectic mantle fusion models.

In alkalic basic rocks, the problem of envisaging how light-rare-earth-element (l.r.e.e.) enriched magmas could be generated from mantle sources with time-integrated l.r.e.e.-depletion, relative to chondrites (see, for example, O'Nions *et al.* 1977), has led to widespread postulation of a 'metasomatic event' in the mantle immediately preceding each episode of magma genesis. Cohen *et al.* (1980) pointed out that precisely the same problem faces students of the genesis of *all* oceanic magmas. To quote their discussion of mid-ocean-ridge basalt (m.o.r.b.) glasses: 'The comparatively small variations in Pb-, Nd- and Sr-isotope compositions in the glasses require that the time-integrated (over the duration of Earth history) Rb/Sr, U/Pb and Sm/Nd ratios in the various source regions must have been quite uniform. This is in marked contrast with the order of magnitude variation of Rb/Sr and U/Pb in the glasses .... Thus it must be inferred either that the processes of basalt magma generation and evolution of mid-ocean ridges can effect much greater fractionation of Rb/Sr, U/Pb, and Sm/Nd than is often appreciated or alternatively some mantle process effects large, but transient fractionation before the melting event.'

Recently Allègre & Condomines (1982) have reported Th-isotope disequilibrium data which shed more light on the matter. If their preliminary results are confirmed by future work, they imply that: (1) both m.o.r.b. and ocean-island basalt (o.i.b.) magmas escape rapidly after their genesis in the upper mantle, relative to the half-life of ^{230}Th (75200 a) and that, (2) the melt fractions during both m.o.r.b. and o.i.b. genesis are small enough to permit Th/U to be higher in the liquid than in the residual mantle. If the melt fractions during oceanic basalt genesis are low enough to affect Th/U, they are also certainly low enough to change elemental ratios among Rb, Sr and l.r.e.e. It therefore appears reasonable to consider the possibility that the model of Ahern & Turcotte (1979) is essentially correct. The only alternative we can envisage is to postulate a mysterious *non-magmatic* 'metasomatism' before every mantle magma-genesis episode, including m.o.r.b.!

MANTLE MODEL

If melt fractions within convecting upper mantle are always very small, it is necessary to postulate a global geodynamic model in which mantle recirculation, via subduction zones, is relatively rapid (Kay 1980; Armstrong 1981). Otherwise mantle reservoirs – such as the source of m.o.r.b. – are used up quickly, whatever their size. Figure 1 shows a cartoon of a tentative model (compare Turcotte 1982), which seems to us to be the one best able to explain the elemental and isotopic diversity of terrestrial igneous rocks. We shall discuss and illustrate it in the following sections. But here are some introductory points.

It has been argued (see, for example, O'Nions *et al.* 1980) that the seismic discontinuity at about 670 km depth in the upper mantle is a chemical as well as a physical break because only the mantle above this depth is required, by mass balance arguments, as the reservoir from which the Earth's sialic continents were extracted. Richter & McKenzie (1981) have calculated that, if the chemical break at 670 km is as little as 2% Mg/Fe, a two-layered convective system will be established, with a double thermal boundary layer at the seismic discontinuity. Chemical communication across this boundary will be limited to leakage of primordial rare gases (Kurz *et al.* 1982; Kaneoka 1983) and, possibly, upward migration of magmas if any melting takes place during convection below 670 km.

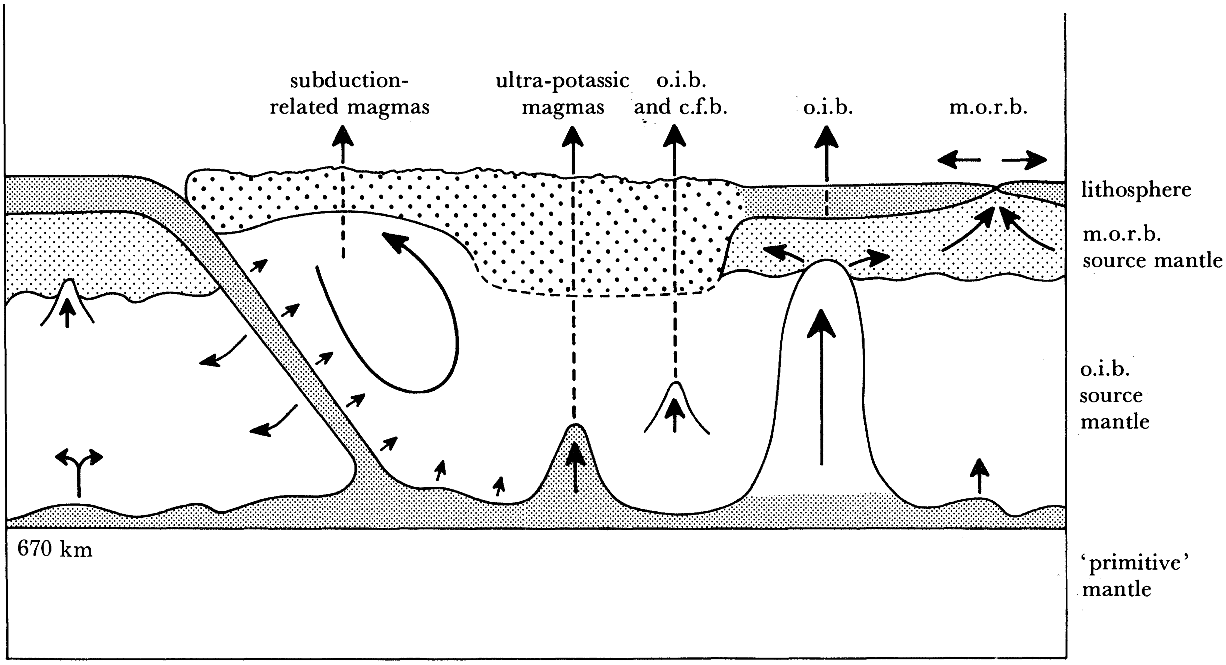

Figure 1. Cartoon of the model for upper-mantle circulation and magma genesis discussed in the text.

Turning to surface magmatism (and excluding subduction-related processes), we shall argue below that almost all terrestrial basic magmatism may be allocated on elemental and isotopic grounds to two mantle sources. In ocean basins these reservoirs feed: (1) the majority of spreading axes with magma (m.o.r.b.) having, for example, ^{87}Sr/^{86}Sr of 0.7021–0.7030 and ^{143}Nd/^{144}Nd of 0.51310–0.51329; (2) ocean islands, such as Hawaii and Iceland, together with adjoining segments of some spreading axes (e.g. Reykjanes Ridge). The latter magmas (o.i.b.) have ^{87}Sr/^{86}Sr of 0.7030–0.7055 and ^{143}Nd/^{144}Nd of 0.51250–0.51309. Although these two groups have considerable internal isotopic variability and grade into each other, it does not seem to us to be an intolerable oversimplification to treat them separately at this stage of our understanding of magmatism. Isotopic studies (see O'Nions *et al.* 1977; Cohen *et al.* 1980) leave no doubt that the m.o.r.b. and o.i.b. mantle sources have been essentially separate for a long time, or that o.i.b.-source mantle is the more 'fertile' of the two, in the sense that it has less of a time-integrated depletion of its basaltic constituents since mantle formation.

We see no logic in attributing major magma-source properties to lithospheric mantle which (by definition) is rigid, sub-solidus and relatively cold. The convection at the base of the lithosphere modelled by Houseman & McKenzie (1982) initially involves *sinking* 'plumes' of cool mantle, scarcely a plausible environment for adiabatic upwelling or radiogenic heat accumulation. We envisage that essentially all basic–ultrabasic magmas originate by decompression-melting of convecting *sub-lithospheric* mantle. We suggest that a convecting upper mantle will tend to develop a density stratification, with m.o.r.b.-source mantle overlying o.i.b.-source (Jordan 1981), although this process will be opposed by the greater internal heat production of the underlying o.i.b.-source layer, tending to lead to convection. In our cartoon (figure 1) we have suggested a 'bouncy' thermal boundary layer at the base of the asthenosphere, between m.o.r.b. and o.i.b. mantle reservoirs; buffeted by convection above and below. This situation seems inevitable, if melts are constantly escaping from the o.i.b.-source layer upwards through the m.o.r.b.-source layer (see below).

If m.o.r.b.-source mantle floats at asthenospheric depths and needs a period measured in Ga during which to develop, its site of generation (by processes discussed below) and principal long-term reservoir must be beneath the Pacific Basin. As our cartoon (figure 1) shows, the lithospheric keel beneath cratons (Jordan 1981) appears to be too deep to allow penetration of m.o.r.b.-source mantle beneath most continents. Similarly, the subducting slabs along the cordilleran plate boundaries around the Pacific should act as effective barriers to the escape of m.o.r.b.-source mantle from that ocean basin, except of course during pauses in subduction or periods of transform movement. How, then, do relatively young oceans, such as the Atlantic, come to erupt m.o.r.b.? We postulate that the m.o.r.b.-source mantle currently underlying the southern and equatorial Atlantic *migrated laterally* from the Pacific into the new ocean basin as it opened. The early stages of this process would have been more complicated than a modern map suggests because of the earlier existence of Tethys (and presumably its underlying m.o.r.b.-source mantle) at equatorial latitudes.

A benefit of the lateral migration hypothesis is that it provides a simple solution to the long-standing problem of the 'anomalous' widespread o.i.b. magmatism in the North Atlantic, in Iceland, the Azores and most of the intervening mid-Atlantic Ridge segment. If the direct routes (avoiding cratons) from the Pacific to the North Atlantic, via the Aleutians or through the Central America–Caribbean region, have been blocked during most of the last 100 Ma by active subduction zones, then the only major supply of m.o.r.b. source has had to make the long journey 'round the Capes'. A mantle volume flowing at 10 cm a^{-1} could theoretically migrate 10000 km in 100 Ma. Seen from this point of view, it is not surprising that the young North Atlantic basin still largely lacks m.o.r.b.-source mantle. We envisage the Reykjanes Ridge to be not the flank of an isolated 'mantle plume' but a low-angle section through the complex gradational boundary layer between overlying m.o.r.b.-source and underlying o.i.b.-source mantle.

Turning to subduction, we accept the model proposed by Ringwood (1982), in which the sinking slabs are not dense enough to penetrate the 670 km discontinuity, but spread out and – after reheating – advect into the upper mantle. Although geologically young sub-oceanic lithosphere, close to spreading axes, must have a bulk composition of m.o.r.b.-source mantle (split into m.o.r.b. and residual mantle) and may therefore be expected to return immediately to the m.o.r.b.-source reservoir, it seems likely that old lithosphere is considerably different. We shall argue below that the lithosphere sinking in most subduction zones contains the feeder and reservoir systems of abundant o.i.b. volcanoes, together with substantial amounts of veining by strongly-alkalic o.i.b. magmas. The re-heating at 670 km and re-circulation of this veined and 'metasomatized' lithosphere into the upper mantle – together with contributions from subducted marine sediments and hydrothermally metamorphosed ocean-crust basalts – makes the main contribution to replenishing the o.i.b.-source mantle in our model (figure 1). We also agree with the reasoning of McKenzie & O'Nions (1983) that the thickened, o.i.b.-veined, sub-continental lithosphere beneath zones of continental collision may become detached, sink to 670 km, re-heat and re-join the o.i.b.-source mantle reservoir (see below).

Ocean basin magmatism

Basic magmas

The chondrite-normalized incompatible-element patterns of a representative selection of basic magmas from ocean basins (excluding island arcs and adjacent areas) are shown in figure 2. In order to avoid the repetition of clumsy nomenclature, this account will follow current usage (Thompson *et al.* 1983) in referring to diagrams such as figure 2 as spider diagrams and the individual patterns as spidergrams. The systematic downward slope from right to left of the m.o.r.b. patterns on figure 2 is an intentional result of the order in which the elements are plotted on this diagram (Sun 1980; Thompson 1982). Samples from transition zones between m.o.r.b. and o.i.b., such as 409-11-3 (Reykjanes Ridge, 63° N) on figure 2, closely resemble m.o.r.b. except for relative enrichments in Nb, Ta and other ultra-incompatible elements.

The spidergrams of all magnesian o.i.b. (tholeiitic and alkalic) approximate to smooth concave-downwards curves on figure 2. Apart from analytical inaccuracies, the main causes of irregularities in these curves appear to be: (1) a trough at Sr in less-magnesian samples (e.g. RTH 13, table 1) which have fractionated plagioclase; (2) troughs at K and Rb in some strongly-silica-undersaturated rocks (basanites to melilitites). Sun & Hanson (1975), Clague & Frey (1982) and others have noted that the variation in K and Rb abundances, relative to other ultra-incompatible elements, in well-studied strongly-alkalic sodic suites, such as the Ross Island basanitoids (Antarctica) or the Honolulu Volcanic Series (Hawaii), seems to indicate that phlogopite was present in the mantle residuum during the extraction of these magmas. Finally, note that the o.i.b. spidergrams all peak at Nb and Ta (figure 2) and that the segments of the curves to the left of these peaks have variable slopes. This feature illustrates what we have already emphasized above; the processes during terrestrial evolution whereby ocean-basin magmas are derived from a primordially-chondritic mantle involve *variable changes* in the inter-element ratios of the ultra-incompatible elements, such as Cs, Ba, Rb, U, Th, K, Nb, Ta and l.r.e.e.

In figure 3 we have attempted to model the range of spidergrams for ocean-basin mafic magmas (summarized on figure 2) by applying the equations proposed by Langmuir *et al.* (1977) for incremental melting, with inefficient melt extraction, to a model primordial mantle with chondritic relative abundances of incompatible elements. Our model spidergrams reproduce both the relative and absolute incompatible-element abundances of ocean-basin magmas quite well, except for two discrepancies: (1) we have ignored the troughs at K and Rb in the spidergrams of strongly-alkalic o.i.b. because, although these may easily be modelled by postulating residual phlogopite, their addition to figure 3 would obscure an already-complex diagram; (2) we can see no way of reproducing the Nb and Ta contents of o.i.b., or the ratios of these two elements to other incompatibles, by means of incremental mantle fusion, whatever

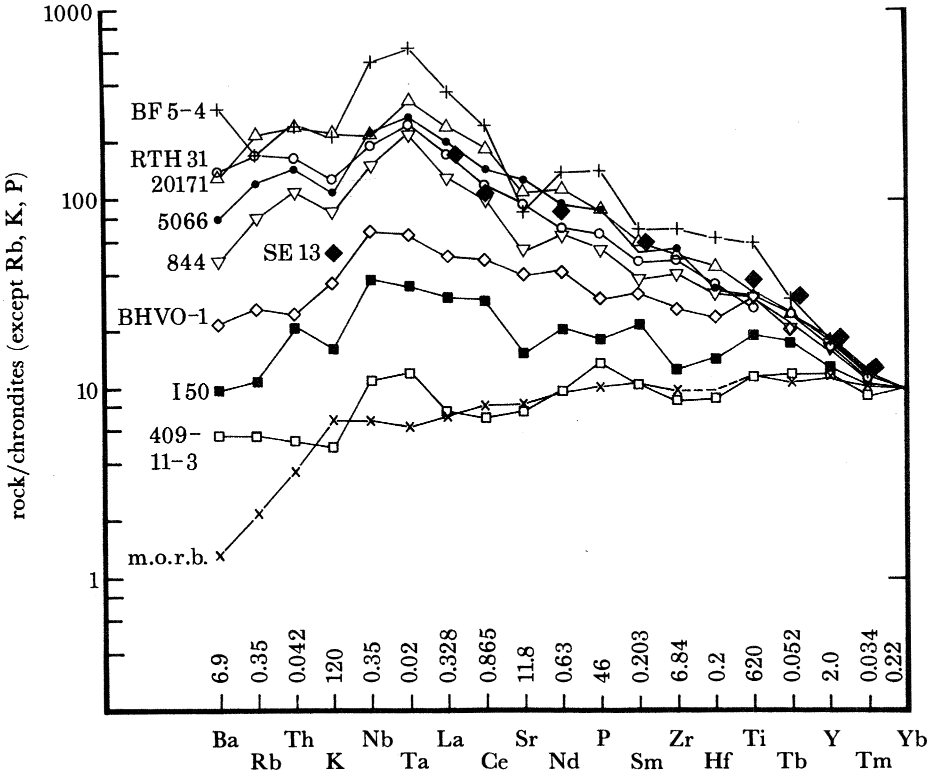

FIGURE 2. Chondrite normalized, trace element abundance diagrams (spidergrams) for representative ocean-basin basic lavas from spreading centres and ocean islands remote from subduction zones. The normalization factors are given on the diagram; see text and Thompson (1982) for details. Tb and Tm interpolated from ICP data (table A 1). In this and subsequent figures, the data for *basic* rocks have been double-normalized to $Yb_n = 10.0$ (see text). Key to samples: BF 5-4, Bermuda; RTH 31, Maui; 20171, Tristan da Cunha; 5066, Tenerife; 844, St Helena (all table A 1); BHVO-1 (international analytical standard), Kilauea, and m.o.r.b. (typical m.o.r.b.) from Sun *et al.* (1979) and Sun (1980); I 50, Eastern Iceland (Wood 1978); 409-11-3 Reykjanes Ridge at 63° N (Wood *et al.* 1979); SE 13 hornblende-mylonite vein cutting dunite-mylonite, St Paul's Rock, S Atlantic (Frey 1970).

values of partition coefficients or melt extracts are chosen. Details of the calculation of individual model spidergrams are given in the legend of figure 3. All the calculated melt fractions are alarmingly low, compared with those invoked in the literature on mantle-derived magmatism during recent years, with the exception of the pioneering study of Kay & Gast (1973) and Kay's subsequent work (for example Kay 1980).

Each model spidergram (figure 3) involves several steps of previous magma extraction from an initially-chondritic mantle. The first few model melting increments are the ones which require exceptionally small liquid fractions. These correspond to the little-understood magmatic processes before 3 Ga. The subsequent melting increments, which lead to 'modern' spidergrams (figure 2), require liquid fractions close to those postulated by Ahern & Turcotte (1979) on the basis of entirely different reasoning. The melting scheme summarized in figure 3 explains the following aspects of isotope geochemistry without recourse to ill-defined metasomatisms: (1) the Th-isotope disequilibrium data (Allègre & Condomines 1982) and implied fractionation of Th from U during mantle fusion; (2) the well known Nd-isotope paradox in many ocean-basin (and other) basic magmas, which have l.r.e.e.-enriched chondrite-normalized r.e.e. patterns (figure 2) but come (with rare exceptions) from sources with time-integrated l.r.e.e.

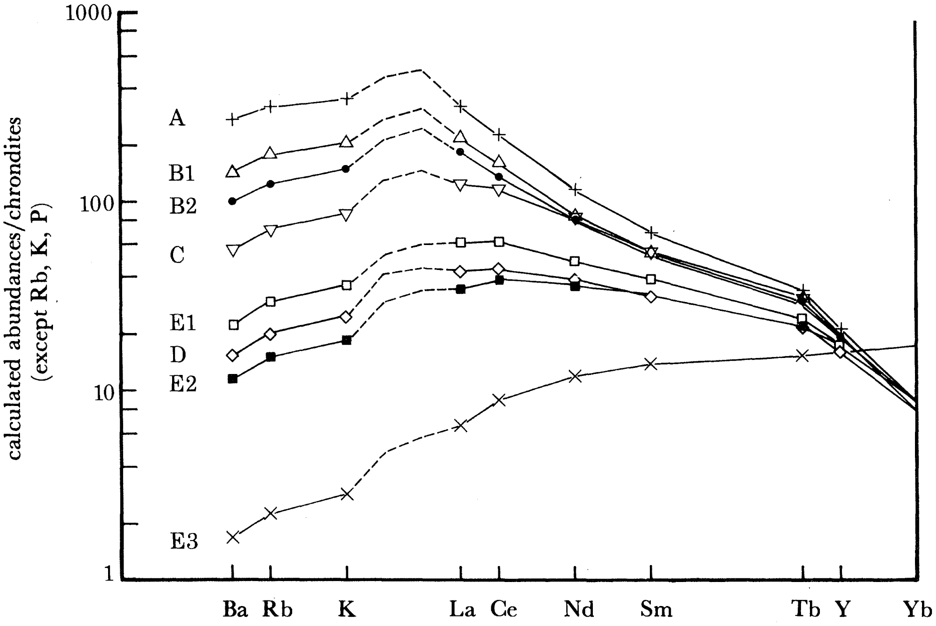

FIGURE 3. Calculated spidergrams for melts derived by dynamic melting (Langmuir *et al.* 1977) of lherzolite with two-times-chondritic incompatible-element abundances. Starting compositions: 60 % olivine, 25 % orthopyroxene, 10 % clinopyroxene, 5 % garnet (model A); 60 % olivine, 20 % orthopyroxene, 15 % clinopyroxene, 5 % garnet (models B to E); melted in proportions 10:10:30:50, respectively. Each of the depicted models was *preceded* by the extraction of two 0.1 % melt fractions, with only 0.05 % melt retained. A: 0.1 % melting, 0.05 % melt retained; B: two steps of 0.1 % melting, 0.2 % melt retained; C: 0.5 % melting, 0.2 % melt retained; D: 1.0 % melting, 2.0 % melt retained; E: two steps of 1.0 % melting, 1.0 % melt retained, residual garnet then converted to clinopyroxene and followed by 5.0 % melting, 2.0 % melt retained; olivine, orthopyroxene and clinopyroxene entering melt in proportions 20:20:60, respectively. Distribution coefficients: r.e.e.: Frey *et al.* (1978) set 3; Ba, Rb, K: Hanson (1977) table A 2, except $K_{D\mathrm{Ba}}^{cpx/l}$ set to 0.0006, $K_{D\mathrm{Ba}}^{gt/l}$ set to 0.0001, $K_{D\mathrm{Rb}}^{ol/l}$ set to 0.00014.

depletion, relative to chondrites (positive ϵ_{Nd}). Before leaving figure 3, we should emphasize that this is an extreme model, in that it assumes unidirectional transport of elements from mantle to crust. We shall argue below for an alternative model, in which re-cycling of elements via subduction zones is important in mantle evolution (see, for example, Armstrong 1981; Patchett *et al.* 1981). In such a model, the postulated melt fractions during incremental mantle fusion can be a little higher than those used to generate figure 3 because the re-cycled elements strongly affect abundances and ratios in the left-hand segments of the spidergrams.

The multi-incremental melting model illustrated in figure 3 implies that m.o.r.b.-source mantle is residual after extraction of o.i.b. from o.i.b.-source mantle. To account for the Sr- and Nd-isotope ratios of m.o.r.b., the main reservoir of m.o.r.b.-source mantle must have been largely isolated for a long period. For this reason, we identify the uppermost mantle beneath the Earth's only geologically ancient ocean, the Pacific, as the terrestrial m.o.r.b.-source mantle factory. We envisage that the more powerful upward-convecting plumes in the underlying o.i.b.-source mantle rise high into the overlying m.o.r.b.-source layer, releasing melt to supply Hawaiian-type oceanic islands, while the mantle residue from this process dissipates laterally into the m.o.r.b.-source reservoir. Batiza (1982) has recently assembled evidence to suggest that we may still grossly underestimate the volume of this type of o.i.b. magmatism and that it may form up to 25 % of the total oceanic crust in regions of the Pacific remote from spreading centres. The heat transported to the lithosphere by magmatism of this order would help to explain the enhanced heat-flow and relative uplift of the older (more than 70 Ma) areas of ocean floors. It would be an additional factor to the various mantle-convection schemes which have been proposed (see, for example, Crough 1979; Yuen *et al.* 1981; Houseman & McKenzie 1982). In addition to this more blatant magmatism, we suggest that there is a much more widespread trickle of o.i.b. magma which is released during minor convective activity at the top of the o.i.b.-source reservoir and leaks upwards through the overlying m.o.r.b.-source mantle layer: much as Frey & Green (1974) supposed. If Ahern & Turcotte's (1979) calculations of magma uprise rates are correct, such o.i.b. liquids could pass *through* a *deforming* overlying m.o.r.b.-source mantle layer much too fast for any significant diffusion-controlled interaction to take place between the two. The compositions and destinations of these deep-seated o.i.b. melts are important to our overall model (figure 1).

Estimation of the major-element composition of the 'trickle' of o.i.b. magmatism, leaking upwards globally from the top of the convecting o.i.b.-source mantle, depends critically upon the choice of gas compositions during mantle fusion. As deep-source o.i.b. magmas are notoriously gas-rich, with CO_2 predominating, it seems reasonable to postulate, on the basis of available experimental data (Wyllie 1984; Thompson 1984), that the 'trickle' melts formed at the top of the o.i.b.-source mantle are picritic members of the basanite-nephelinite-melilitite clan. We agree with Frey & Green (1974) that very little of this low-volume strongly-alkaline magmatism is likely to penetrate the relatively cold, rigid overlying lithosphere and reach the surface. Most of it should freeze in the lower lithosphere (Turcotte 1981), giving rise to veined mantle, together with extensive metasomatism as hydrous CO_2-rich fluids are expelled from the crystallizing magmas (Bergman *et al.* 1981; Stosch 1982). In the older parts of the deeper lithosphere mantle several phases of strong-alkalic magmatism and associated metasomatism may alternate with penetrative deformation and annealing, giving rise to enigmatic 'metasomatized mantle' (see, for example, Frey & Green 1974). As Tarney *et al.* (1980) have stressed, only a small percentage of strongly-alkalic veining is required to re-fertilize barren lithospheric mantle, residual from m.o.r.b. genesis at spreading centres, and create a potential source-mantle for future o.i.b. magmatism. Our overall model (figure 1) differs from that of Tarney *et al.* (1980) in that, rather than supposing *in situ* fusion of veined lithosphere to produce o.i.b. magmas, we suggest that this re-fertilized mantle re-joins the o.i.b.-source reservoir, via subduction and re-heating above the 670 km seismic discontinuity: an extension of the reasoning of Ringwood (1982) and McKenzie & O'Nions (1983).

The incompatible-element compositions of ocean-basin o.i.b. 'trickle' magmatism are illustrated in figure 2, as follows: (1) the basanite spidergrams are closely similar to the available incompatible-element data for the hornblende-mylonite veins (basanitic) within the dunite-mylonite lithospheric mantle which forms St Paul's Rocks (Frey 1970); (2) so-called limburgites (Peckenham *et al.* 1982) form dykes which cut Bermuda seamount and its surroundings about 100 Ma after the latter were originally formed; (3) the 'post-erosional' magmas of the Hawaiian islands post-date both tholeiitic and mildly-alkalic stages of volcano construction by *ca.* 1 Ma (Clague & Frey 1982). This is quite a long period in a system where the drift rate of lithosphere over asthenospheric magma sources is about 9 cm a^{-1}. Noting the subtle isotopic differences between Hawaiian volcano-building and post-erosional magmas (O'Nions *et al.* 1977; Tatsumoto 1978), we speculate that Bermuda magmatism may provide an alternative to the traditional interpretation of Hawaiian magmatism as an evolutionary sequence from tholeiitic,

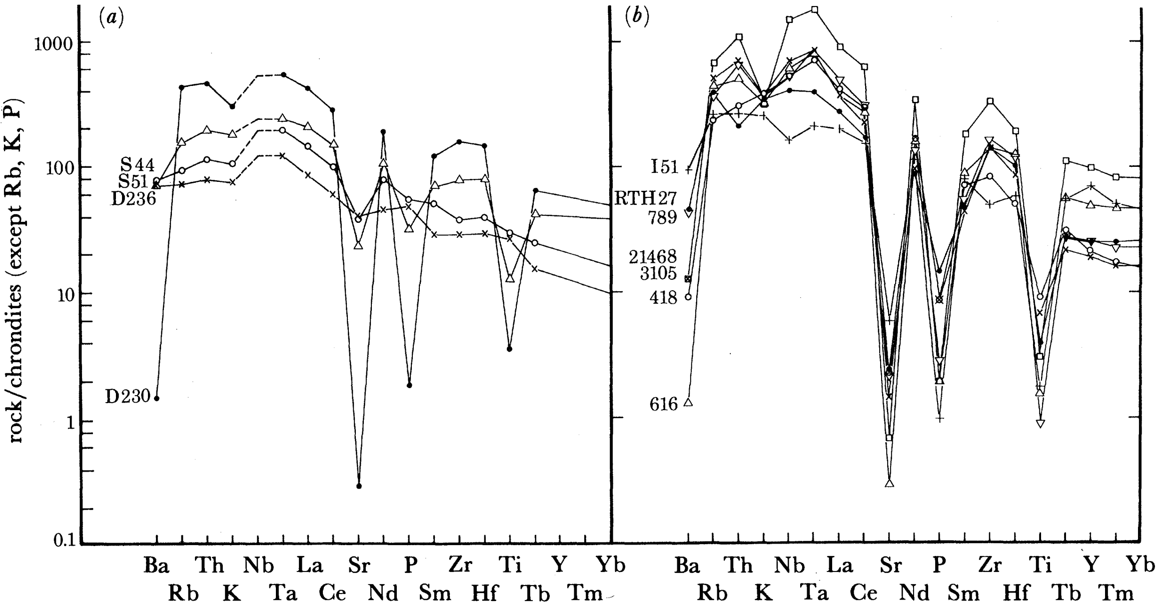

FIGURE 4. Spidergrams for (*a*) Representative lavas from the basalt-to-pantellerite suite of Boina, Ethiopia (Barberi *et al.* 1975). (*b*) Representative leucocratic magmas (lavas and hybabyssal intrusions) from oceanic islands. Key to samples: I 51, Thingmuli, Iceland (Wood 1978); RTH 27, Mauna Kea, Hawaii; 789, St Helena; 21468, Tenerife; 3105, Pantelleria; 418, Grand Canaria; 616 Ascension (all table A 1).

via mildly alkalic, to strongly-alkalic magmatism: all related to a single mantle-source volume. Perhaps the source mantles for the various Hawaiian magmatic suites are different, as is obviously the case in Bermuda.

Acid magmas

We have analysed several examples of extreme leucocratic lavas (silica-saturated and undersaturated) from oceanic islands (see table A 1 in Appendix) in order to compare them with continental leucomagmas (below). In figure 4*a* we illustrate their genesis through fractional crystallization of o.i.b., by plotting the spidergrams of representative members of the basalt–pantellerite suite of Boina, Ethiopia. Although sub-aerial, this volcano is constructed on essentially oceanic crust and Barberi *et al.* (1975) have shown how the ratios of many pairs of incompatible elements do not change significantly throughout the range of erupted magmas (figure 4*a*). In contrast, the spidergrams show vividly the relative depletion of first Sr (tholeiitic andesite), then P and Ti (trachyte) and finally Ba and K (pantellerite), as early crystallizing plagioclase is joined by apatite, Fe–Ti oxides and finally replaced by alkali feldspar in the fractionating crystalline assemblage (Barberi *et al.* 1975). At this point a paradox becomes apparent on figure 4*a*. There are ample published data (see, for example, Pearce & Norry 1979; Leeman & Phelps 1981; Mahood & Hildreth 1983) to demonstrate that the partition coefficients of Rb, Nb, Ta and r.e.e. into feldspar, apatite and Fe–Ti oxides crystallizing from acid magmas are considerably above zero. The apparent total incompatibility of these elements throughout the Boina suite, postulated by Barberi *et al.* (1975), must be an artefact, caused by all of them having approximately the same bulk partition coefficients into the solids. One independent clue that this is the case is provided by the behaviour of Sm/Zr on figure 4*a*. Zircon did not fractionate from this peralkaline suite. Note the changes in Sm/Zr between basalt and pantellerite, caused by m.r.e.e. entering apatite while Zr remained in the liquid. Bailey & Macdonald (1975) have reported similar data for the trachyte–pantellerite suite of Eburru, Kenya, but prefer to interpret them without recourse to an hypothesis of fractional crystallization.

The spidergrams of representative ocean-basin leucocratic lavas are plotted on figure 4*b*. All but one of these o.i.b. fractionation residua are peralkaline and their spidergrams closely resemble the Boina pantellerite (figure 4*a*). The exception is the Icelandic rhyolite (I50) which fractionated zircon (Wood 1978), causing a distinctive depletion of Zr, relative to Sm and Hf, in its spidergram. If the buffering of K by cotectic equilibria in Petrogeny's Residua System is ignored, the sections between Rb and Ce of the spidergrams of peralkaline oceanic leucomagmas (figure 4*b*) still closely resemble those of their basaltic progenitors (figure 2). The small trough at Th in the pattern of RTH 27 (trachyte, Mauna Kea) looks anomalous at first sight. This relative Th depletion is a local characteristic of many Hawaiian volcano-building basalts (figure 2) and their fractionates (table 1).

We conclude that figure 4 illustrates the widespread development in oceanic islands of leucomagmas that evolved by extreme fractional crystallization of basaltic parents. The long-standing problem of how early-formed crystals and residual liquids became separated from each other appears to be resolved (in outline) by the experiments of McBirney (1980) and Turner & Gustavson (1981), which emphasize crystallization on reservoir walls and the upward migration and accumulation of low-density magmatic residua. We agree with Mittlefehldt & Miller (1983) that the abundances of Ba, Sr, P and Ti are so low in extreme leucomagmas, oceanic and continental (table A1), that their genesis by partial fusion rather than fractional crystallization requires the postulation of either implausible parent rocks or melting conditions. Once due attention has been paid to the minor phases separating from the liquids and to the significant incompatible-element contents of such minerals as clinopyroxene, aenigmatite and Fe–Ti oxides (Hildreth 1979; Michael 1983), the only non-volatile elemental ratio in leucomagmas which seems to defy an explanation in terms of crystal–liquid equilibria is Na/K in strongly-peralkaline suites (Macdonald *et al.* 1970). We suspect that leakage of halogen-rich fluids from the reservoirs that attempt to contain these exotic magmas is the cause of their non-magmatic Na/K variations.

CONTINENTAL ANOROGENIC MAGMATISM

Basic alkalic rocks

Several recent studies have shown that continental basic alkalic magmatic suites from diverse localities are isotopically indistinguishable from o.i.b. (see, for example, Allègre *et al.* 1981; Norry & Fitton 1983). In figure 5*a* we show that the incompatible-element ratios of these two groups are also essentially identical (compare figure 2). As with their oceanic analogues, we suspect that most low-volume batches of strongly-alkalic magmas (basanites, nephelinites, melilitites and kimberlites) fail to penetrate the continental lithosphere and instead become the agents of characteristic veined and metasomatized subcontinental mantle.

Flood basalts

We have recently completed a detailed review of this major terrestrial magma group (Thompson *et al.* 1983) and shall therefore only summarize our conclusions here. (1) Continental flood

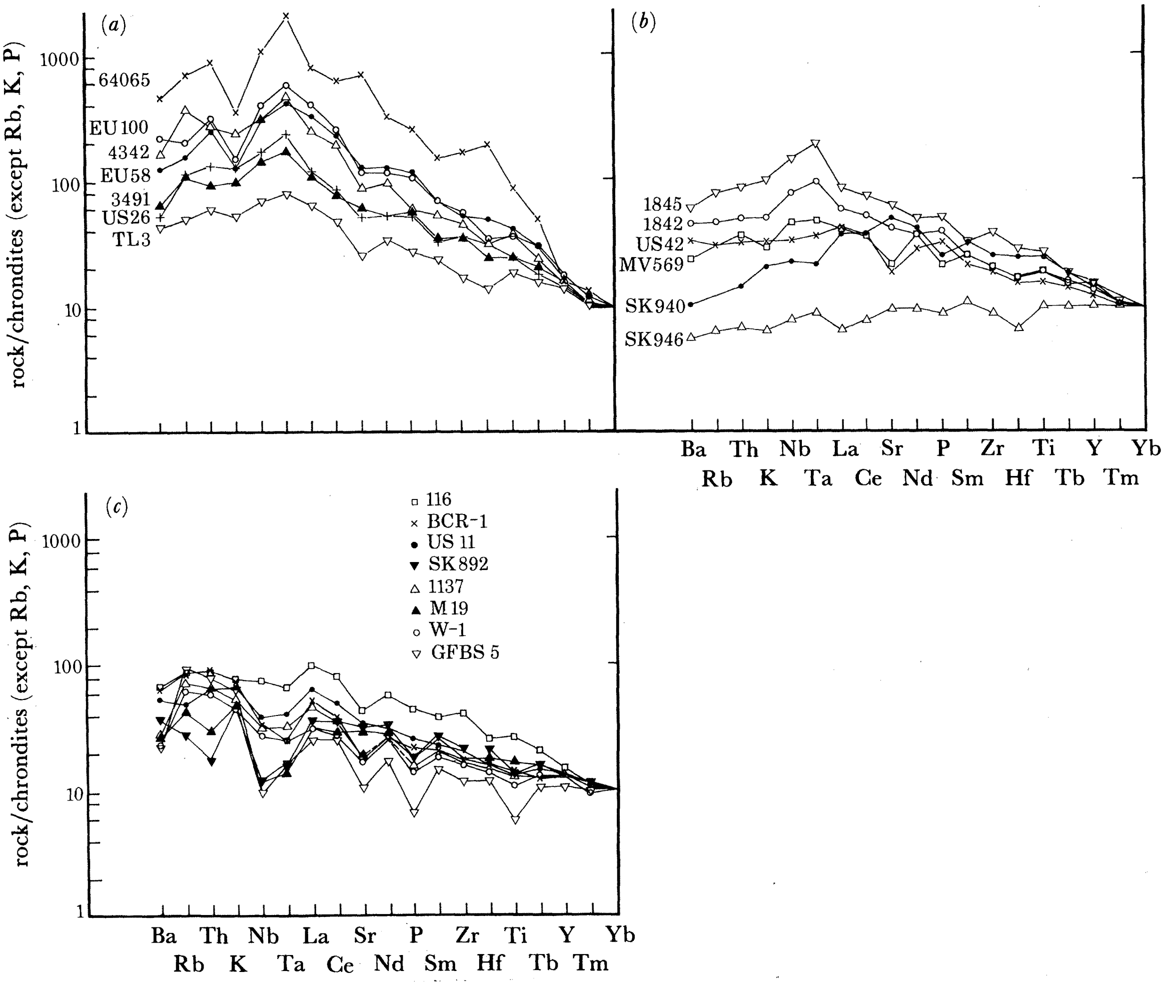

Figure 5. Spidergrams for (*a*) representative continental alkalic basic rocks. Key to samples: 64065, kimberlite, Pipe 200, Lesotho; EU 100, melilitite, Hegau, Germany; 4342, leucitite, Uganda; EU 58, limburgite, Kaiserstuhl, Germany; 3491, alkali-olivine basalt, Calton Hill, Derbyshire; US 26, basanite, Dish Hill, California; TL 3, alkali-olivine basalt, Lake Natron, Tanzania (all table A 1). (*b*) Representative continental flood basalts that resemble o.i.b. in composition. Key to samples: 1842 and 1845, Addis Ababa, Ethiopia; US 42, Snake River Plain, Idaho; MV 569, Midland Valley, Scotland; SK 940 and SK 946, Skye, Scotland. (*c*) Representative continental flood basalts that do not resemble o.i.b. in composition. Key to samples: 116, Parana Basin, Brazil; BCR-1 (international analytical standard) Columbia River Plateau, Washington; US 11, Rio Grande Rift, New Mexico; SK 892, Skye, Scotland; 1137, High Atlas Morocco; M 19, Mull, Scotland; W-1 (international analytical standard), Virginia; GFBS 5 Basement Sill, Antarctica. All the samples in (*b*) and (*c*) are discussed in detail by Thompson *et al.* (1983).

basalts (c.f.b.) range in silica saturation from quartz tholeiite to basanite. Most of the largest provinces are almost exclusively tholeiitic. (2) Only a handful of the samples analysed to date could be primary, unmodified, mantle-derived magmas. The remainder have equilibrated within or at the base of the continental crust before extrusion. (3) Despite their complex pre-eruption histories, some c.f.b. are clearly recognizable, elementally and isotopically, as o.i.b. (figure 5*b*). As implied by our overall model (figure 1), we stress here that we are unaware of any instances of true m.o.r.b. (in terms of both elements and isotopes) erupted through continental crust.

The majority of c.f.b. differ from o.i.b. in both incompatible-element and various isotopic

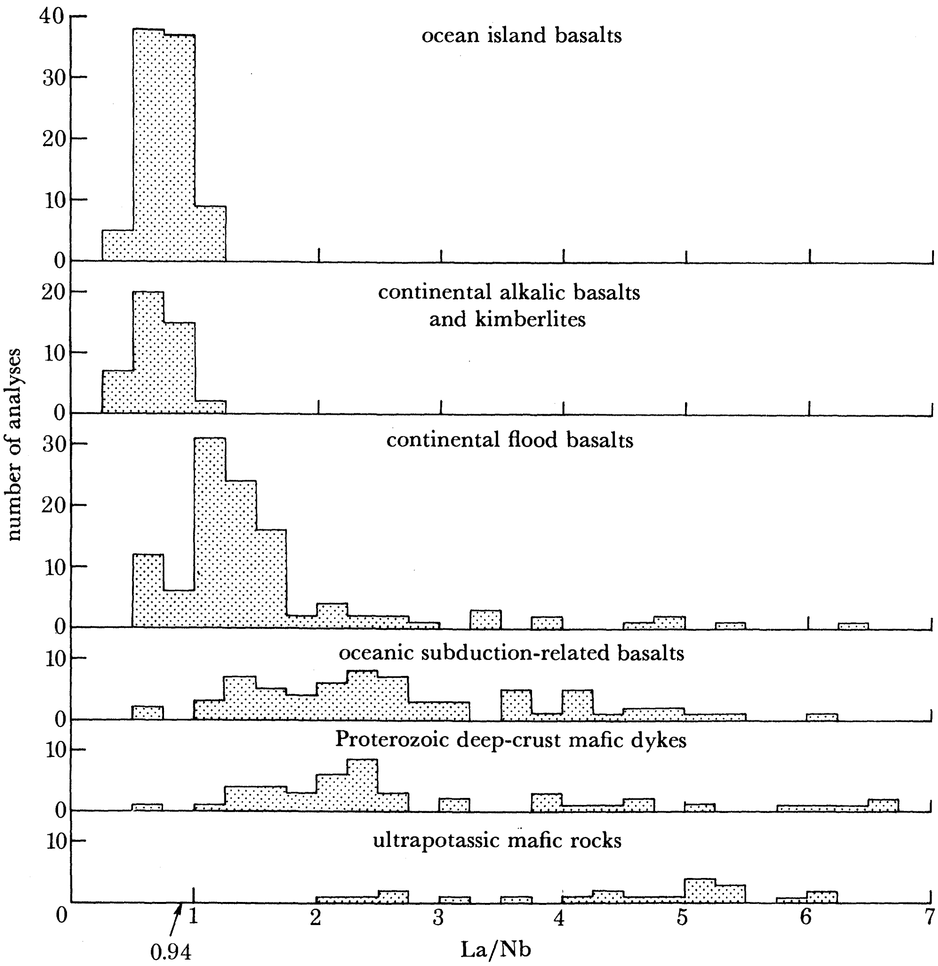

FIGURE 6. Histograms of La/Nb in various igneous rock groups. The chondritic value of La/Nb is 0.94, from the abundances given on figure 2. Thompson *et al.* (1983) list the source references for the oceanic basalts and continental flood basalts. Sources of the other data are table A 1, Baker *et al.* (1977), Dal Piaz *et al.* (1979), Fesq *et al.* (1975), Frey *et al.* (1978), Holm *et al.* (1982), Kesson (1973), Kyle (1981), Nash & Wilkinson (1971), Norry *et al.* (1980), Sheraton & Black (1981), Weaver & Tarney (1981).

ratios (figure 5*c*). We ascribe these differences, in most cases – but not all (Thompson *et al.* 1983) – to contamination of c.f.b. magmas (upwelling from asthenospheric sub-continental o.i.b. mantle sources) during their pre-eruptive equilibration within sialic crust. We have explained elsewhere (Thompson *et al.* 1982, 1983) why we prefer this hypothesis to the alternative that c.f.b. were generated within ancient subcontinental lithosphere.

The spidergrams of fusible crustal rock types (e.g. figures 7, 8 and 10) show that these are usually much richer than most o.i.b. in the elements Ba, Rb, Th, K and l.r.e.e. but have similar or lower contents of Nb, Ta, P, Zr, Hf, Ti, Y and middle-heavy r.e.e. (also U and Th in granulite-facies leucogneisses). This is why crustal contamination leads to the characteristic spidergrams of many c.f.b. (figure 5*c*). It is apparent from figure 5 that Nb/La might be a particularly useful ratio, of a pair of widely-analysed elements, with which to compare o.i.b., c.f.b. and continental basic alkalic rocks. A histogram of Nb/La (figure 6) is consistent with the proposed identification of continental basic sodic alkalic lavas as o.i.b., and of c.f.b. as their variably crustally-contaminated equivalents.

Acid rocks

Because the fusible components of the continental crust are siliceous and mostly peraluminous, it is logical to search among the phonolites and strongly peralkaline rhyolites for leucocratic rocks devoid of sialic contamination. We have analysed several samples specifically to test this point (table A 1, EU 47, K 53T, 40622, 3090). The spidergram of the Ailsa Craig riebeckite microgranite 3090 resembles very closely those of o.i.b. leucocratic rocks (figure 4*b*). Strongly silica-undersaturated phonolites, such as the Fohberg wollastonite phonolite (EU 47), do not show extreme depletion of Sr and Ba because they have fractionated nepheline, rather than plagioclase or alkali feldspar, during the latter stages of their evolution. In addition to nepheline and wollastonite, sample EU47 contains sparse garnet phenocrysts. Fractionation of this phase is presumably the cause of the low h.r.e.e. abundances and high Zr/Hf (106) of this phonolite.

Figure 7*a* shows the spidergrams of several well known examples of anorogenic trachytes, rhyolites and granites which have been studied isotopically and are considered in the published reports of such work to have evolved from mixtures of magmas from both crustal and mantle sources (Barker *et al.* 1975; Thompson *et al.* 1980; Dickin & Exley 1981; Hildreth 1979). Note the resemblance of the left-hand parts of these patterns to those of sial-contaminated continental flood basalts (figure 5*c*). The characteristic trough at Nb and Ta is present in each case, together with a trough at Th in the pattern of trachyte JE 208, which contains a substantial fraction of granulite-facies leucogneiss (Thompson *et al.* 1982).

Figure 7*b* shows the distinctive spidergrams of continental non-peralkaline, ultra-leucocratic acid rocks. The extreme relative depletions of Ba, Sr and P in the patterns suggest that fractional crystallization proceeded even further than is generally possible within ocean-basin volcanoes (compare figure 4*b*). The small troughs at Zr, and values of Zr/Hf (11–24) below the basaltic range (30–60), indicate zircon fractionation from these magmas. The spidergrams of all the ultra-leucocratic rocks show little or no tendency towards troughs at Nb and Ta; the most extreme examples have marked peaks at Ta (compare figures 7*a* and *b*). We can be sure that the changes in La/Nb and La/Ta between figures 7*a* and *b* are related to the processes whereby ultra-leucocratic magmas evolve, because there are two examples of single magmatic suites represented on both diagrams. These are: (1) the Pikes Peak pluton and its associated Redskin cupola (Ludington 1981); (2) the high- and low-temperature parts of the Bishop Tuff (Hildreth 1979). We suggest that fractional crystallization of allanite (chevkinite in some cases) is the mechanism whereby l.r.e.e. are depleted, relative to Nb and Ta, in the ultra-leucocratic magmas. Allanite has been reported in several of the samples plotted in figure 7*b*. We agree with Miller & Mittlefehldt (1982) and Michael (1983) that Hildreth's (1979) objections to fractional crystallization as the mechanism by which these magmas evolve are unnecessary, now that the processes of sidewall crystallization and upward migration of light residual liquids in magma chambers are better understood (McBirney 1980; Sparks *et al.* 1984). Nevertheless, we disagree with Mittlefehldt & Miller (1983) that monazite is usually the phase which largely controls the l.r.e.e. budgets of acid magmas. From their data, it is clear that monazite fractionation leads to massive relative depletions of both l.r.e.e. and Th, giving rise to spidergrams which are quite different from any found in our survey (table 1; figures 7 and 10).

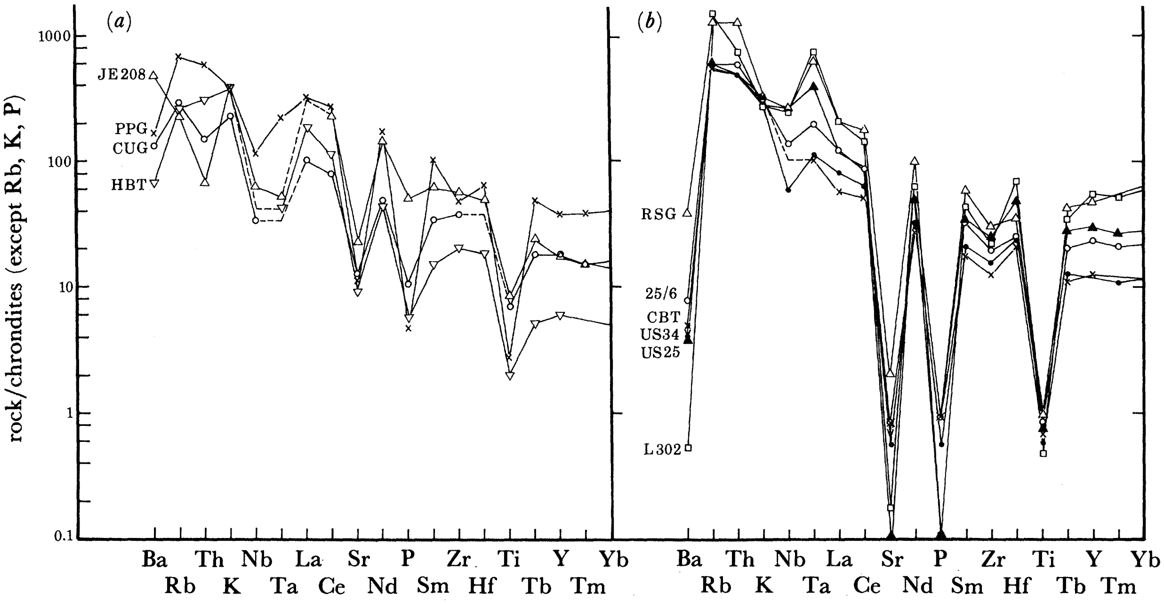

FIGURE 7. Spidergrams for (*a*) representative continental leucocratic magmas that are regarded on isotopic grounds (see text) as having formed by a mixture of liquids from both crustal and mantle sources. Key to samples: JE 208, trachyte flow, Skye, Scotland (Thompson *et al.* (1982); PPG, Pike's Peak Granite, Colorado (Barker *et al.* 1975; Ludington 1981); CUG, Coire Uaigneich Granite, Skye (Dickin & Exley 1981; Thompson 1983); HBT, high-temperature part of Bishop Tuff, California (Hildreth 1979). (*b*) Representative continental ultra-leucocratic magmas. Key to samples: RSG, Redskin Granite, Colorado (Ludington 1981); 25/6, obsidian, Coso Volcanics, California (Bacon *et al.* 1981); CBT, low-temperature part of Bishop Tuff; US 34, obsidian, Mono Craters (table A 1); US 25, obsidian, Jemez Mts (table A 1); L 302, granite, Jere-Sanga, Nigeria (table A 1).

SUBDUCTION-RELATED MAGMATISM

In figure 8*a* we show the spidergrams for representative subduction-related basic lavas from localities not underlain by continental crust. The dramatic departures from smooth curves shown by all these patterns tends to obscure the fact that their *overall* slopes are in the same range as m.o.r.b. and o.i.b. (i.e. from l.r.e.e.-depleted to strongly l.r.e.e.-enriched). We suggest that the reason for this is that subduction-related magmas are simply m.o.r.b.–o.i.b. (generated within convecting mantle wedges behind subduction zones) with complexities introduced by added components brought into the system through subduction (see, for example, Kay 1980; Armstrong 1981). The only added component about which everyone seems to be agreed is water (Delaney *et al.* 1978). It is clear from the available experimental data (Basaltic Volcanism Study Project 1981; Wyllie 1984; Thompson 1984) that water addition systematically raises the silica-saturation of mantle melts, provided that other *PTX* constraints remain unchanged. Thus, we envisage that low-K, island-arc tholeiites (*Q* normative) may be regarded as the hydrous subduction-related analogues of m.o.r.b. (mostly *ol*+*hy* normative), while calc-alkaline basalts (either *ol*+*hy* or *Q* normative) are similarly related to o.i.b. tholeiites and mildly-alkalic magmas (either *ne* or *ol*+*hy* normative), and that shoshonites (mildly *ne* normative) are the subduction-related equivalents of basanites, nephelinites and melilitites. Pearce (1982) suggests a similar relation between the suites of mafic ocean-basin and subduction-related magmas but explains it in terms of a complex array of non-magmatic metasomatisms,

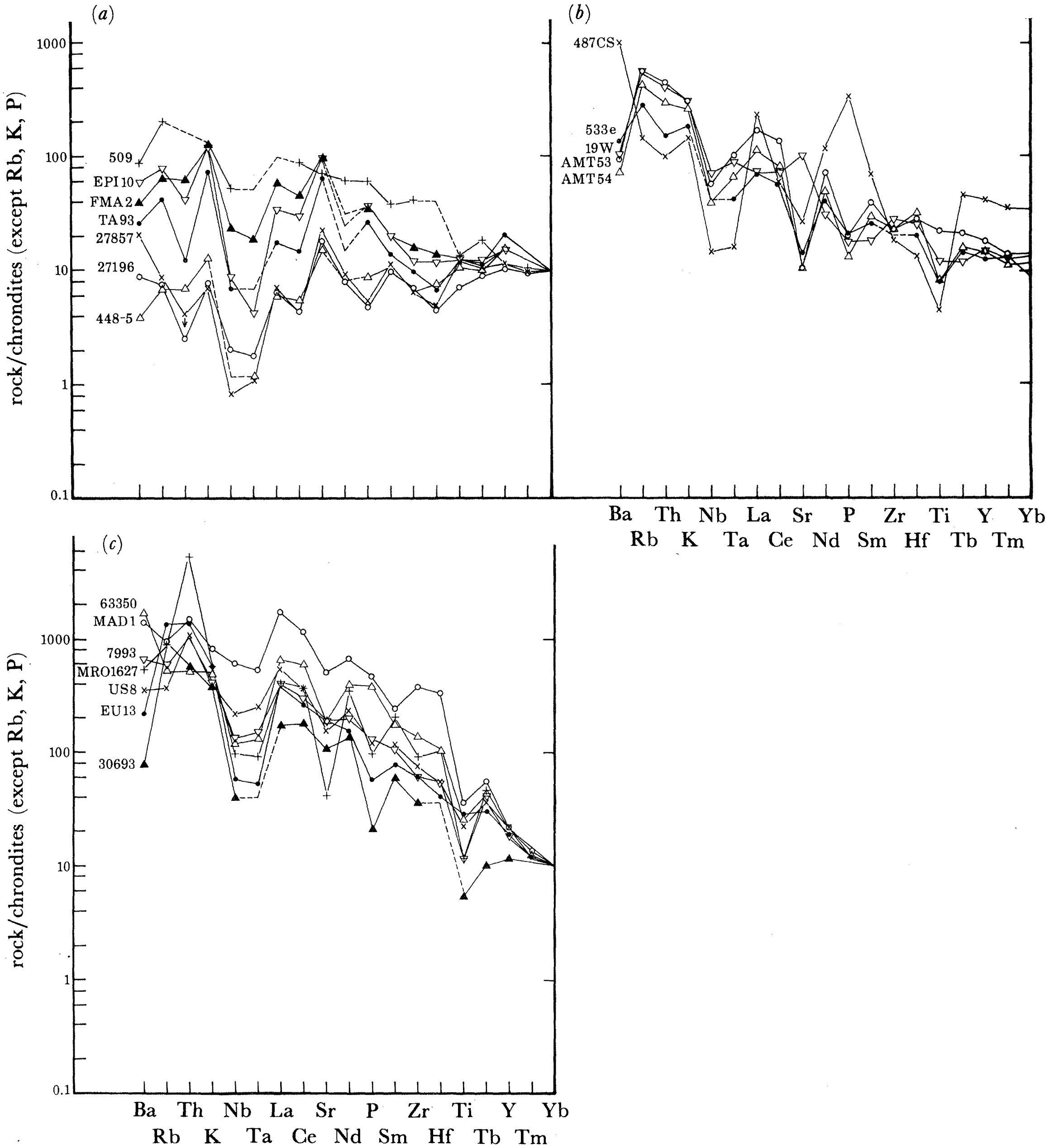

FIGURE 8. Spidergrams for (*a*) representative oceanic subduction-related basalts. Key to samples: 509, Grenada, Lesser Antilles (Shimizu & Arculus 1975); EPI 10, FMA 2 and TA 93, New Hebrides (Dupuy *et al.* 1982); 27857, Izu Peninsula, Japan (table A 1); 27196, Warner flow, California (table A 1); 448-5, DSDP Leg 59, Palau-Kyushu Ridge (Wood *et al.* 1981). The Izu and Warner basalts are generally supposed to be underlain by oceanic crust. (*b*) Marine sediments. Key to samples: 487 CS, pelagic mud overlying basalt, DSDP Leg 66, Middle America Trench (table A 1); 533e, Cretaceous argillite, Alaska (Hill *et al.* 1981); 19W, slate, Skiddaw, Cumbria (table A 1); AMT 53 slate, Blaenau Ffestiniog, Wales (table A 1); AMT 54 slate, Shap, Cumbria (table A 1). (*c*) Representative ultrapotassic mafic magmas. Key to samples: 63350, minette, Summer Isles, Scotland (table A 1); MAD 1, madupite, Leucite Hills, Wyoming (table A 1); 7993, minette; Sedberg, Cumbria (table A 1); MRO 1627, minette, Champoluc, Western Alps (Dal Piaz *et al.* 1979). US8, minette, Agathla Neck, Arizona (table A 1); EU 13, augite leucitite, Lago di Albano, Italy (table A 1); 30693, olivine leucite melilitite, San Venanzo, Italy (Holm *et al.* 1982).

rather than by the approach based on silicate crystal–liquid equilibria which we employ throughout this paper.

In the foregoing classification we are concerned only with mafic–ultramafic magmas and geochemical features supposed to be inherited from processes within the mantle. We agree with Green (1982) and Kay *et al.* (1982) that fractional crystallization of such magmas at depth, with separation of amphibole and titanomagnetite, is frequently a dominant influence in the development of the characteristic chemical features of calc-alkaline andesites and dacites. An excellent small-scale example of an analogous process within a Hawaiian shield volcano is provided by the rhyodacite at Kauaopuu Ridge, Waianai volcano, Oahu (table A 1, RTH 7), which contains amphibole, biotite and Fe–Ti oxide phenocrysts. The chemical relation between this lava and typical Hawaiian shield-building tholeiites (e.g. table A 1, RTH 13) is a classic 'calc-alkaline trend'.

Returning to mafic–ultramafic magmas, the widespread tendency for low-K tholeiites to occur above the shallower parts of subduction zones, while shoshonites occur above the deepest parts, is a confirmation of our general argument that, beneath ocean basins, m.o.r.b.-source mantle overlies o.i.b.-source mantle (figure 1). Exceptions to this rule (see, for example, Arculus & Johnson 1978) are hardly surprising, if convection is active in the mantle wedge and disrupts the tendency towards density-fertility layering in the mantle. Our hypothesis that m.o.r.b.-source mantle has migrated laterally out of the sub-Pacific region into young ocean basins, such as the Atlantic, explains the curious southward change from low-K tholeiites to basanites along the Lesser Antilles Arc at an approximately constant depth from the volcanoes to the Benioff zone (Arculus & Johnson 1978). Because it is cut off on all sides, except northwards, from m.o.r.b.-source mantle by Benioff zones, one would expect the southern part of the Lesser Antilles mantle wedge eventually to become starved of m.o.r.b. source mantle.

Several studies have pointed out that there is an *isotopic* gradation between subduction-related basalts and m.o.r.b.–o.i.b. (see, for example, Meijer 1976; Kay *et al.* 1978; DePaolo & Johnson 1979). Using spidergrams we find that the same intergradation occurs with incompatible *elemental* patterns. Examples in the published literature include many of the Grenada basanites (see, for example, Shimizu & Arculus 1975), basalts from New Britain and the Marianas (DePaolo & Johnson 1979; Meijer & Reagan 1983), and shoshonites from many localities (Kay 1980). We have also analysed two basalts from the part of Mexico beneath which the Cocos Plate is currently subducting (table A 1, 26550 and 64064). All these samples have spidergrams that are unmistakably o.i.b.-like overall, although differing in detail in such features as La/Nb (see below).

Published discussions of the chemical differences between subduction-related and other oceanic basic lavas (see, for example, Kay 1980; Pearce 1982 and references cited therein) generally postulate a combination of added components (solutions or melts?) from the subducting slab to the overlying mantle wedge, together with stabilization of various residual solid phases in the magma sources (slab or wedge?), to account for the marked relative depletions of some elements. Our new analyses and survey of the published data lead us to the following conclusions on these points.

(i) The most persistent chemical feature of subduction-related basalts is their marked relative depletion in Nb and Ta, which leads to high values of La/Nb (figure 6) and troughs on their spidergrams (figure 8*a*). We agree with other authors (see, for example, Saunders *et al.*

1980) that the most likely explanation for this phenomenon is that Nb and Ta are retained in the residuum during magma generation in a hydrous environment. The solid phase is generally supposed to be a Ti-rich mineral. Our data (table A 1) and plots (figure 8) suggest that this is not so, because troughs at Ti on the spidergrams by no means always accompany those at Nb and Ta (compare figure 8*c*, EU13 and 30693).

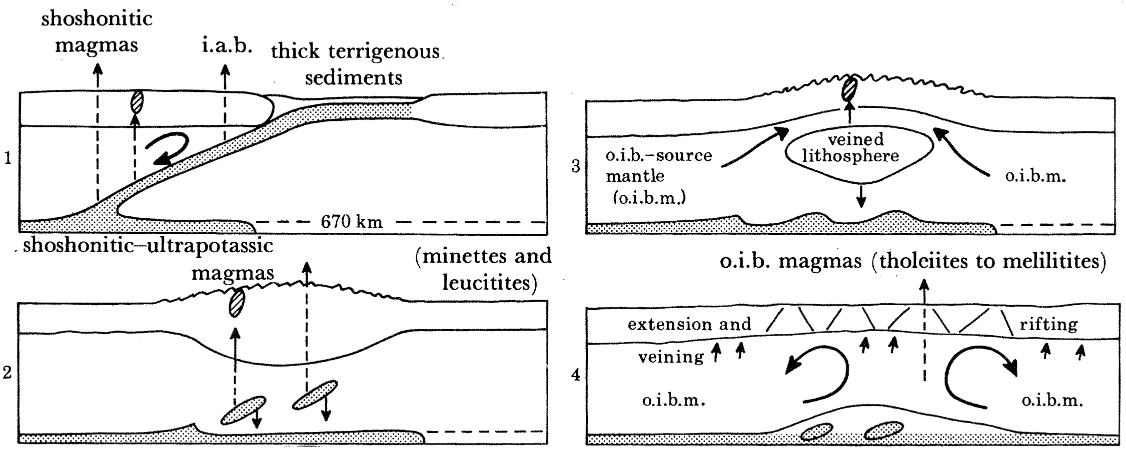

FIGURE 9. Cartoon of the model for upper-mantle circulation and magma genesis, during and after continental collision, discussed in the text.

(ii) Troughs at Zr and Hf appear in the spidergrams of many of the more silica-saturated subduction-related basalts (figure 8*a*). In such cases there is usually also a marked trough at Th and a relative depletion of U (table 1). We agree with Saunders *et al.* (1980) that the logical explanation for these phenomena, at first sight, is the stability of zircon in the residuum during genesis of these magmas. Nevertheless, we are very concerned that the supposed occurrence of residual zircon during the genesis of a major part of the magmas that contributes to continental growth does not appear to have left any discernible imprint on the evolution of terrestrial Hf isotopes, despite the immense Hf/Lu fractionation in zircon (Patchett *et al.* 1981).

(iii) The spidergrams (figure 8*a*) and data in table 1 show clearly why it is generally supposed that oceanic subduction-related basalts are enriched in Ba, Rb, K and Sr – together with U, Th and Pb in some cases – relative to m.o.r.b. and o.i.b. We do not feel that either elemental or isotopic studies have yet shown unequivocally whether the agent of this enrichment at the melting sites within the mantle wedge is silicate melt (from oceanic sediment and basaltic sources) or hydrous fluid released during dehydration of the subducting slab. Nevertheless, detailed geochemical investigations of boninites (Hickey & Frey 1982) and recent Nd-isotope studies (Cohen & O'Nions 1982) both strongly support the idea that l.r.e.e. are also among the enriched elements and that the agent of transport is therefore probably a silicate melt. The recent Be-isotope data of Brown *et al.* (1982) provide further evidence for this point of view (Kay 1984).

We have analysed three terrigenous slates and one pelagic mud (table 1; figure 8*b*) as a contribution to understanding chemically the types of material which may be subducted in various circumstances. Except during ocean closure (see below) and near to river deltas, the majority of sediment liable to be subducted is pelagic (Karig & Kay 1981; Wilson† 1984).

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† Formerly Powell.

The brown pelagic mud (table A 1, 487CS) is a composite of the material immediately overlying ocean-crust basalt adjacent to the Middle America Trench (DSDP leg 66, hole 487). Our analysis shows high concentrations of Fe, Mn, Ba and P, which may be related to hydrothermal activity during deposition of the sediment near a spreading centre. Kay (1984) gives a detailed discussion of the sediment input to subduction zones.

The foregoing account has concentrated on island arcs within ocean basins. As soon as subduction-related basalts penetrate and fractionate within continental crust, it becomes excruciatingly difficult to disentangle (by means of elements, isotopes or both) the sialic contributions from two separate sources, via subduction or continental crustal contamination. The recent breakthrough in the use of Be isotopes (Brown *et al.* 1982) confirms that *some* subducted ocean-floor sediment is entering island-arc volcanics erupted on both oceanic and continental crust. To that extent, it seems possible to rule out the extreme crustal-contamination model proposed by Arculus & Johnson (1981). Nevertheless, we suspect that the majority of continental subduction-related magmatic suites are variably contaminated by the sial through which they pass, as may be diagnosed in favourable cases (see, for example, Thorpe *et al.* 1984).

Despite these contamination problems, the continents are interesting places for the study of subduction-related magmatism because – behind the main volcanic arc – it is often possible to obtain samples of contemporaneous 'back-arc' magmatism. Similar situations exist in the ocean basins (see, for example, Kay *et al.* 1978) but do not survive for long within the geological record because they are eliminated by ocean-ridge spreading and subsequent subduction. Thorpe *et al.* (1984) notes 'normal' o.i.b. magmatism to the east of the Andean subduction-related volcanoes. In the northwestern U.S.A. the Columbia River Plateau, Snake River Plain and Yellowstone magmas have erupted behind the Cascades arc during the last 16 Ma. Recent accounts have debated whether crustal, sub-crustal lithospheric, subducted or other sources are responsible for the distinctive compositions of these magmas (Thompson 1977*a*; DePaolo 1983; Thompson *et al.* 1983; Menzies *et al.* 1984).

During the Tertiary there were several cases where extremely potassic basic magmas were erupted or emplaced far to the east of the U.S.A. Pacific coast and its associated subduction. Published chemical data on these suites show that, far from resembling the sodic o.i.b. magmas which – together with m.o.r.b. – characterize oceanic back-arc basins (Tarney *et al.* 1981), these ultra-potassic magmas seem to be extreme variants of the shoshonitic suites of subduction-related magmas (Edgar 1980). We have analysed samples from two classic localities, Leucite Hills, Wyoming, (table A 1, MAD 1, 2) and the Navajo Province, Arizona, (US 8) and confirm that elementally (figures 6 and 8*c*) they are extreme but typical members of the subduction-related clan. The Agathla minette (US 8) is particularly important because it is virtually identical in composition (Rogers *et al.* 1982; Ehrenberg 1982) to the nearby contemporary minette of the Thumb, which contains garnet-lherzolite xenoliths and appears to have traversed the continental crust and underlying lithospheric mantle very fast from a depth of origin of at least 110 km (Ehrenberg 1982). Although low-angle subduction beneath the western U.S.A. in Tertiary times has been postulated (Lipman *et al.* 1972), it is difficult to see why – once they are dense enough to plunge into the asthenosphere – the sinking slabs should extend sub-horizontally for enormous distances beneath a continent, within asthenospheric mantle but not sinking further into it. Our preferred model (figure 1) is that the slabs subducting at the U.S.A. Pacific coast during the Tertiary sank to the 670 km seismic discontinuity (Ringwood 1982) and were then over-ridden by westward drift of the American Plate (necessitated to permit the opening of the North Atlantic). We therefore suggest that the Tertiary ultrapotassic magmas were fed ultimately by continuing fusion of slab remnants on the 670 km 'floor'; the fusion products entering upward-convecting plumes in overlying o.i.b.-source mantle and releasing ultramafic ultrapotassic wet magmas at depths of 110 km or more (Nb and Ta remaining largely locked in the crystalline residuum).

In the view presented here, the presence of a considerable fraction of subducted crustal material in the western U.S.A. deep-source ultrapotassic magmas is crucial evidence that both sediments and water are carried far deeper and further in subduction zones than might be supposed from the study of island-arc magmas alone. To this extent we support the view of Armstrong (1981) and Patchett *et al.* (1981), although we agree with the latter in not accepting the whole of Armstrong's model. Although the available experimental phase equilibria show clearly that sediments could not survive unmolten below about 100 km, perhaps they continue to be carried down as phlogopite-rich reaction complexes (Wyllie & Sekine 1982), physically protected from dispersion into convecting overlying mantle by a carapace of relatively rigid peridotite 'frozen' on to the cold slab during its subduction.

In the light of our overall model, it is logical to expect that an isotopically identifiable crustal component should be re-gurgitated from the mantle in the magmatism of back-arc-basin spreading centres (Cohen & O'Nions 1982). We suggest (figure 1) that the subducted slabs of veined sub-ocean-basin lithospheric mantle, together with whatever remains of their igneous and sedimentary ocean-crust cover, are re-heated above the 670 km seismic discontinuity until they become buoyant again, homogenizing by advection as they rise (Richter & Ribe 1979). In some instances the rising plumes may feed o.i.b. magmatism directly (McKenzie & O'Nions 1983) but we suspect that in most cases the bulk density of the upwelling material is such that it joins the worldwide o.i.b. source-mantle reservoir, completing the geochemical cycle we have summarized in figure 1. The Nb and Ta retained in the solid residuum of subduction-related magma genesis also appear to find their way into the o.i.b.-source mantle reservoir. We agree with Tarney *et al.* (1980) that this is the logical explanation of the high Nb and Ta abundances, relative to other incompatible elements, in o.i.b. magmas (figures 2 and 5*a*). As we emphasized above, it does not seem possible to model the Nb and Ta in o.i.b. by calculations of incremental mantle melting alone (figure 3). The behaviour of these two elements in oceanic magmatism appears to be strong evidence for the type of mantle re-cycling model we have summarized in figure 1. Finally, we note that a model along these lines is considered by both Patchett *et al.* (1981) and DePaolo (personal communication, 1983) to be essential in order to explain the evolution of the terrestrial Nd- and Hf-isotope systems during the last 3 Ga.

COLLISIONAL AND POST-COLLISIONAL MAGMATISM

Model

Figure 9 is a cartoon summarizing our views on the relation between magmatism and events in the underlying mantle before, during and shortly after collision between two continental lithospheric plates. Certain aspects are discussed in more detail below. At this stage we note the following points.

(i) Chemical zonation of subduction-related magmatism perpendicular to the continental margin during ocean closure is unlikely to be detected unambiguously (when it occurs) unless abundant Mg-rich basic magmas are extruded or intruded, so that the effects of interaction between mantle-derived magmas and sialic crust are minimized (see, for example, Thirlwall 1982). Our sample MS 143 (table A 1) – kentallenite from the type locality in western Scotland – illustrates a rare instance where a volcanic neck of subduction-related shoshonitic picrite has reached a crustal level dominated by batholithic granites, by using the vertical plane of weakness through the lithosphere provided by the major transcurrent Great Glen Fault.

(ii) During the final stages of their closure, ocean basins receive large supplies of terrigenous sediment (figure 8*b*). Substantial amounts of this may be subducted, if the floor of the contracting ocean becomes imbricated during its final closure (Thompson 1977*b*).

(iii) We recognize shoshonites and their more chemically extreme ultra-potassic associates from two distinct sites in place and time during continental collision. First, they sometimes occur in a 'back-arc' situation, over or beyond the deepest parts of Benioff zones, during active subduction. Second, they occur in a 'post-orogenic' setting, immediately after plate collision, on both sides of the suture (see below).

(iv) The concept of gravitational subduction or sloughing-off of the thickened sub-continental lithospheric mantle root of a collisional orogen (Houseman *et al.* 1981) carries with it an implication for post-collisional magmatism, because sub-continental asthenosphere must flow upwards into the space left by the sinking 'root' and, in doing so, may be expected to melt and give rise to copious magmas. We suggest that this is the ultimate cause of the burst of 'post-orogenic' magmatism, straddling the suture, which characterizes the British Isles segment of the Caledonides and many other continental collision zones. The flip from subduction-related to o.i.b. magmas at mantle depths may not become apparent immediately in the overlying batholithic suites, if the mafic magmas are entering hot crust which contributes substantially to the liquids. Eventually, as the system cools, there is frequently a phase of 'late-orogenic' or 'early-post-orogenic' *peralkaline* granites (see, for example, Harris & Marriner 1980) of unmistakable o.i.b. parentage.

(v) We illustrate our concept of the magmatic sequence throughout a continental collision even by means of the spidergrams (figure 10*a*) of *basic* rocks between Lower Devonian and Permian age from NW Scotland. This region was several hundred km behind the subduction zone along the NW margin of Iapetus; hence the shoshonitic affinities of the subduction-related magmas (Thirlwall 1982). Transcurrent movement on the Great Glen Fault complicates a simple interpretation of figure 10*a* but we cannot resolve this matter, so long as the activity of this fault remains controversial (Van de Voo & Scotese 1982). The large time gap between the minettes (*ca.* 400 Ma) and subsequent o.i.b. nephelinites (*ca.* 280 Ma) in NW Scotland could be reduced (if we had the relevant analyses) by considering an area nearer to the Midland Valley, where rift-related o.i.b. magmatism began at about 365 Ma (Pankhurst 1982), in Uppermost Devonian (Harland *et al.* 1982).

Granites

Ideas about the genesis of the batholithic tonalite–granite suite have changed substantially during the last few years. Following the realization that island-arc and Andean magmatism was linked to subduction, there was a period when cordilleran volcanic magmas were generally held to be predominantly mantle-derived, while the batholithic plutons associated with them (and possessing almost identical elemental and isotopic ranges) were usually ascribed to crustal fusion. At present, cordilleran batholiths are generally considered as ultimately mantle-derived, with sial added variably to the system (see, for example, Atherton *et al.* 1979; DePaolo 1981*b*).

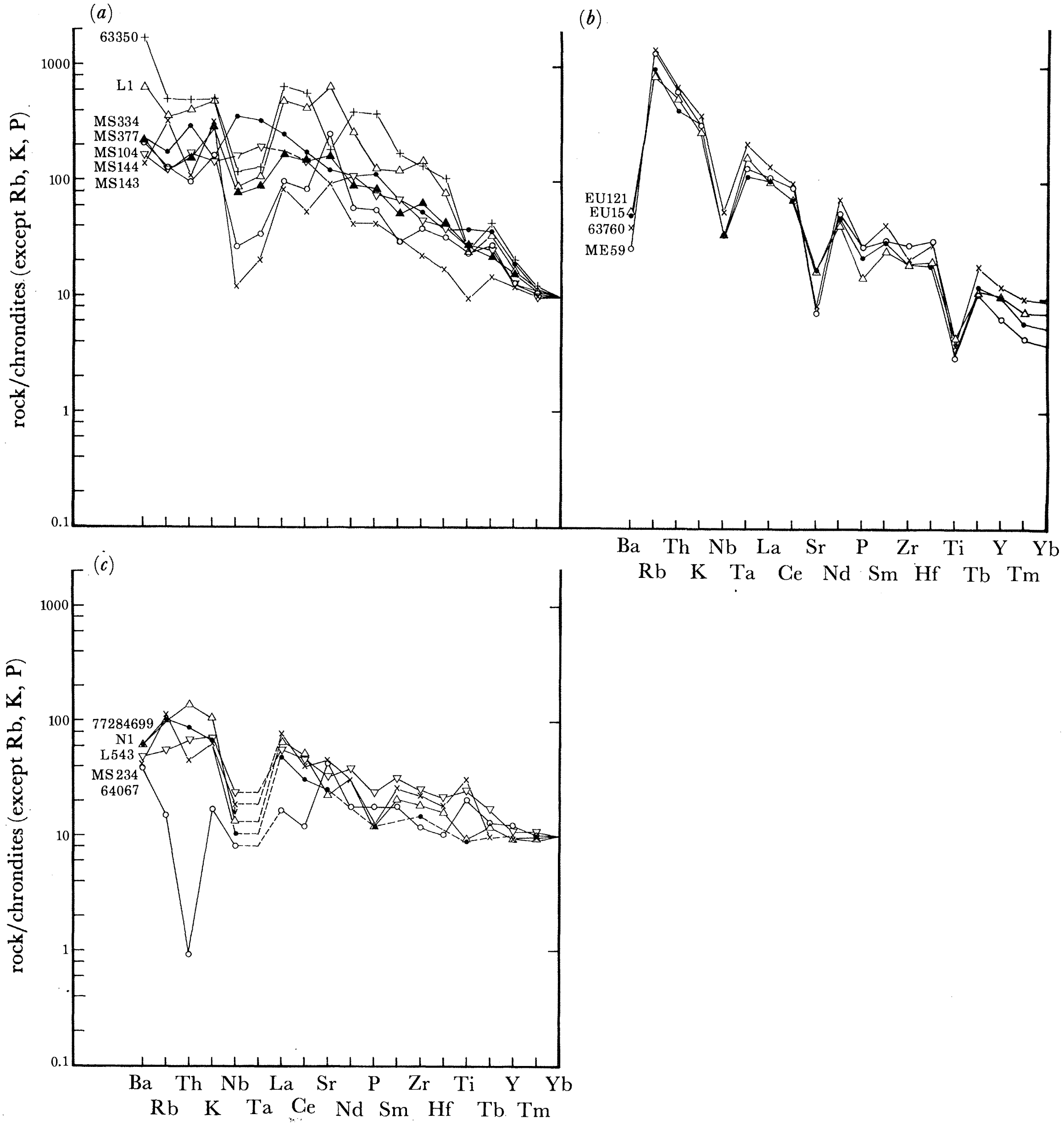

FIGURE 10. Spidergrams for (*a*) Caledonian and post-Caledonian (Devonian–Permian) basic magmatism in western Scotland. Key to samples (all in table A 1) and their age relations (Pankhurst 1982): MS 143, Kentallenite (shoshonitic picrite). 405–410 Ma. Cut by Ballachulish tonalite; MS 377, Basalt. 400 Ma. Lowest flow of Lorne Plateau lavas on Kerrera; MS 104, Gabbro/diorite (MgO = 5.52 %). Glencoe Fault Intrusion. Cuts Lorne lavas and is cut by Etive Complex, emplaced at 400 Ma (Harmon *et al.* 1983); L1, Minette. Dyke of swarm which cuts Glenelg-Ratagain complex (emplaced at 400 Ma); 63350, Minette. Unknown age but closely resembles L1 and post-granite minettes in SW Mull dated at 406 Ma by Beckinsale & Obradovich (1973); MS 144, Camptonite (nephelinite) *ca.* 280 Ma. Dyke of swarm which cuts L1; MS 334, Monchiquite (nephelinite) *ca.* 280 Ma. Dyke of same swarm as MS 144. (*b*) Examples of 'S-type' magmas. Key to samples (all in table A 1): EU 121, Mt Capanne pluton, Elba; EU 15, vitreous rhyodacite, San Vincenzo, Tuscany; 63760, Lands End pluton, Porthcurno, Cornwall; ME 59, Bodmin pluton, Trebartha, Cornwall. (*c*) Basic and ultrabasic intrusions which are thought to have been subduction-related magmas, emplaced deep in the crust beneath contemporary tonalite-granite batholiths and cordilleran volcanoes (see text). Key to samples: 77284699, norite dyke, Enderby Land, Antarctica (Sheraton & Black 1981); N1 norite and L543 olivine gabbro, Scourie dykes, NW Scotland (Weaver & Tarney 1981); MS 234, bytownite two-pyroxene cumulate, Currywongaun layered pluton, Connemara (table A 1); 64067, fine-grained olivine gabbro, Insch pluton, Aberdeenshire (table A 1).

In contrast, most accounts still favour genesis by crustal fusion for collisional and post-collisional granites (tonalite-granite suites), such as those of the British Isles Caledonides and Hercynides (see, for example, Leake 1978; Pitcher 1979), despite the thermal problems emphasized by Wyllie (1984) and the occurrence of small amounts of diorite, gabbro and their hydrated equivalents (appinites) in many such complexes.

Current isotopic studies appear to give unequivocal evidence for simple crustal melting as the source of granites in only a small proportion of instances (see, for example, Taylor *et al.* 1984; DePaolo 1984). The elemental case for crustal melts is strongest for the so-called S-type magmas (Chappell 1984). We have analysed four widely-recognized examples of S-type acid magmas (three plutonic, one volcanic), in order to see how their incompatible-element compositions relate to those of sediments. At first sight, the spidergrams of the S-type magmas (figure 10*b*) are very similar to those of shales (figure 8*b*). Nevertheless, there are subtle differences. Perhaps the most interesting (and unexpected) are the low values of Nb/Ta in the S-type magmas, relative to those of shales or worldwide basic magmas. During our analytical study we have found that low Nb/Ta is characteristic of many acid rocks (table 1) but is confined to those samples which also show lower Zr/Hf than in basalts. We therefore suspect that some of the Nb in acid magmas is trivalent. The ionic radius of Nb^{3+} is very close to that of Hf^{4+} (Shannon & Prewitt 1969) and Nb^{3+} might then be expected to enter zircon readily. Our own prejudice is that zircon separated from the magmas plotted in figure 10*b* during their fractional crystallization from more mafic parents. We note that all of them are considerably more mafic than the leucogranitic 'minimum-melting' compositions that S-type magmas are generally supposed to be. Chappell (1984) argues that the reason for this is that the magma was emplaced as a mush of restite solid phases in a minimum-melt liquid. In the coarse-grained granodiorites of SW England and Elba (figure 10*b*) we can find no petrographic evidence for or against this view. In the acid lava (figure 10*b*, EU 15) of San Vincenzo, however, the matrix is fresh glass and the crystals are euhedral phenocrysts of quartz, sanidine, plagioclase, biotite, cordierite and magnetite. We agree with Barberi *et al.* (1967) that these crystals are clearly an equilibrium assemblage, formed during cooling of an original all-liquid magma. The melting behaviour of both EU 15 and the Elba andalusite-granodiorite (EU 121) has been studied experimentally, under dry conditions and at 1 $kbar_{H_2O}$ (J. Nolan, personal communication)†. Both Tuscan S-type magmas have liquidi approximately 200 °C higher than the minimum-melting composition in the synthetic system $NaAlSi_3O_8$–$KAlSi_3O_8$–SiO_2–H_2O under the same conditions.

Turning from the peraluminous, ^{87}Sr- and ^{18}O-rich minority of granites to the majority of batholithic plutons, we agree entirely with the geochemical models developed by DePaolo (1981 *a*, *b*), Tindle & Pearce (1981) and others, that envisage fractional crystallization of mantle-derived hydrous basic magmas, trapped in hot thick crust, together with progressive assimilation of sialic country rocks. The experience of the Coire Uaigneich granite, in the British Tertiary Volcanic Province, should be remembered when interpreting inherited crustal zircons in these plutons (Pidgeon & Aftalion 1978). As Dickin & Exley (1981) have shown in this case, acid magmas full of inherited zircons are not necessarily devoid of a mantle-derived fraction; only a comprehensive isotopic study can resolve the matter.

We have analysed a number of batholithic tonalite–granite samples (table 1 and unpublished). Because many of them are from the British Isles segment of the Caledonides and because they show very complex elemental behaviour, we shall postpone a detailed discussion of them to a subsequent report on Caledonian magmatism. In cases where the crustal input into the frac-

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† 1 bar = 10^5 Pa.

tionating system is unimportant, the residuum from silica-saturated mafic liquids will be granitic (see, for example, Atherton *et al.* 1979), while shoshonitic parental magmas (normative *ne* or $ol+hy$) will evolve into syenites. These differ in many chemical details from o.i.b.-related syenites; notably their spidergrams have deep troughs at Nb and Ta. They are particularly well represented within the Caledonian orogenic belt and on the Lewisian foreland of NW Scotland (table A 1, 63758 and unpublished analyses).

It is not easy to deduce the magmatic plumbing beneath batholiths, because currently-exposed examples of them appear from geophysical studies to extend for several kilometres into the upper crust, so that their 'roots' are deeply buried. Lynn *et al.* (1981) have pointed out that all the available high-quality reflection-seismic profiles through batholithic complexes in the U.S.A. show sub-horizontal layered reflections beneath the granites at 6–10 km depth. We agree with these authors that the reflectors are most probably layered gabbros. Very extensive layered mafic–ultramafic complexes, frequently deformed and metamorphosed, are characteristic of the deep-crustal levels of orogenic belts. Where igneous assemblages are preserved in such plutons, these frequently include the phases orthopyroxene, amphibole or very calcic plagioclase: key minerals fractionating during the evolution of island-arc magmas from mafic mantle-derived parents when the calc-alkaline trend is followed (Perfit *et al.* 1980; Kay *et al.* 1982). Excellent examples of such mafic plutons are provided by the Tremadocian gabbros, norites and peridotites of Aberdeenshire and Connemara in the segment of the Caledonian orogenic belt within the British Isles. These plutons are well known for having melted and incorporated the country rock (high-grade metasediments) around them on a large-scale (Gribble & O'Hara 1967; Leake 1970).

We have analysed a very fresh, fine-grained (doleritic), olivine-rich facies of the Insch Pluton, Aberdeenshire (Clarke & Wadsworth 1970). This rock has a spidergram (figure 10*c*, 64067) closely resembling those of island-arc tholeiites (figure 8*a*). We have also analysed a bytownite-augite-hypersthene cumulate from the Currywongaun layered pluton, Connemara (Leake 1970). The incompatible elements in this rock are largely held in its sparse intercumulus phases and therefore have such low abundances that Nb and Ta are below our detection limits. Nevertheless, it is clear (figure 10*c*, MS 234) that the spidergram of this rock is very similar to those of calc-alkaline lavas from oceanic island arcs (figure 8*a*). Our data therefore support the view of Yardley *et al.* (1982) that these Tremadocian mafic plutons are the roots of a Caledonian subduction-related volcanic arc; we would go further and interpose batholithic tonalite–granite plutons within the (eroded) upper crust. Detritus from the erosion of Connemara during the Ordovician forms the sedimentary infill of the South Mayo Trough, immediately to the north. Among these sediments are Llanvirnian submarine-canyon conglomerates, packed with large boulders of varied granites derived from the 'missing' batholiths (Archer 1977).

We regard the Tremadocian plutons of the Caledonides as some of the most shallow-emplaced examples of sub-batholithic mafic complexes. Other typical representatives of such plutons, intruded at deeper crustal levels, would include the Baltimore Complex, U.S.A., and various suites of hypersthenic charnockites in ancient terrains. Going deeper still into the crust, we need to consider regions of granulite-facies Archaean and Proterozoic gneisses, in order to try to see what sort of magmatic conduit system lay beneath the sub-batholithic mafic complexes. The granulites of these regions are frequently cross-cut by swarms of mafic dykes. Examples of these which have been studied in detail, such as the Early Proterozoic dykes of NW Scotland (Weaver & Tarney 1981) and Enderby Land, Antarctica (Sheraton & Black 1981) show a range of compositions from silica-saturated basalts to ultramafites rich in olivine, orthopyroxene or both. One of the compositions listed by Sheraton & Black contains 55.1 % SiO_2 and 10.6 % MgO (*op. cit.* table 3, B). This is typical of the lavas known as 'high-magnesian andesites' in modern SW Pacific island arcs (Tatsumi & Ishizaka 1982). The incompatible-element spidergrams (figure 10*c*) and La/Nb range (figure 6) of these Proterozoic dykes are extremely similar to those of modern subduction-related magmas. Weaver & Tarney (1981) note petrographic evidence that the Scourie dykes were originally hydrous liquids and comment on their chemical similarity to the magmas of present-day island arcs. We conclude that dykes such as these were originally the deep-crust roots of batholiths.

Post-collisional minettes and leucitites

The emplacement of swarms of minette dykes, cutting both the folded and metamorphosed country rocks and the batholithic granites emplaced in them is a well known feature of collisional orogenic belts. Good examples of this phenomenon are seen in the Caledonides and the Alps. Although the majority of minettes usually post-date the regional deformation, there is frequently overlap between the latest folding and earliest ultrapotassic magmatism. Leucitites, the erupted equivalents of minettes, are rare products of current subduction-related volcanism in Indonesia (Foden & Varne 1980). But the widespread post-collisional minettes seem to be a different and distinctive magmatic phenomenon.

We have analysed minette dykes from both sides of the Caledonian suture in Britain, from northern England (7993) and NW Scotland (63350). The spidergrams of these are plotted on figure 8*c*, together with an Oligocene minette (MRO 1627) from the Alps (Dal Piaz *et al.* 1979). All of them are clearly very similar extreme members of the subduction-related magma clan, which resemble closely in composition the deep-source Agathla minette, discussed above. On figure 8*c* we have also plotted two representative leucitites from the Roman Magmatic Province of Italy. Of these, it has been established experimentally by Thompson (1977*b*) that sample EU 13 had reached cotectic equilibrium within the upper mantle immediately before its final uprise and eruption.

As indicated in figure 9, we envisage that all these ultrapotassic magmas originate from deep within the upper mantle, with major contributions from the melting of stagnant subducted slabs, up to several million years after continental collisions. Enhanced subduction of terrigenous shale during ocean-basin closure seems to be the most likely reason for the association of these distinctive magmas with collisional orogeny. In the case of Italy, the collision event involved the Corsica–Sardinia microplate (Alvarez 1972). Apart from their well-known isotopic features (see, for example, Kay & Gast 1973; Turi & Taylor 1976; Hawkesworth & Vollmer 1979), there is elemental evidence that these mafic ultrapotassic magmas contain large subducted-sediment fractions. Their spidergrams show extremely deep Nb and Ta troughs, and some also have negative Eu-anomalies in their r.e.e. patterns (table A 1, EU 13), even when their bulk compositions are much too silica-undersaturated for feldspar to precipitate. As McLennan & Taylor (1981) have emphasized, this chemical feature is to be expected in subduction-related magmas containing substantial fractions of subducted sediment.

Extensive fractional crystallization of the ultrapotassic magmas of the Roman Province has taken place, and has lead to trachytic and phonolitic residua. We have analysed a trachyte from Ischia (table A 1), 1093) which appears, on isotopic grounds, to have evolved from mafic parentage without appreciable concomitant crustal contamination (Turi & Taylor 1976).

The Biella syenite stock occurs in the same part of the Alps as the swarm of minette and more-evolved related dykes studied by Dal Piaz *et al.* (1979). Both are of Oligocene age. Our analysis of a sample of this substantial syenite pluton (table 1, 2403) shows the same shoshonitic affinities as the Scottish Caledonian syenites, discussed above, and its exceptionally high abundances of U and Th are features also shown by the minettes. We therefore support the tentative conclusion of Dal Piaz *et al.* (1979) that the Biella syenite is a fractionation residuum of ultrapotassic post-collisional magmatism.

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APPENDIX. TABLE A 1

(For key to samples, see table A 2)

	RTH 13	RTH 7	RTH 24	27001	RTH 27	RTH 31	877	5066	21468	418	21101	616	844	789	20171
SiO_2	50.36	68.07	46.44	49.28	62.62	45.23	46.60	45.06	59.15	62.46	43.21	70.50	45.93	62.23	46.18
TiO_2	3.62	0.60	2.34	2.66	0.36	2.25	1.99	3.37	0.69	0.92	2.86	0.16	2.85	0.09	3.26
Al_2O_3	13.41	14.94	11.94	17.23	17.46	13.33	10.27	14.72	18.98	17.19	12.61	13.36	14.59	17.93	16.51
$Fe_2O_3^*$	13.63	3.18	12.60	11.47	4.66	13.31	10.05	12.01	3.87	4.14	13.89	3.37	13.13	5.21	11.74
MnO	0.18	0.07	0.17	0.21	0.35	0.19	0.17	0.19	0.20	0.29	0.19	0.15	0.18	0.14	0.19
MgO	5.52	2.05	12.38	3.99	0.39	10.37	13.46	7.76	0.36	0.90	11.94	0.00	8.45	0.00	4.60
CaO	9.60	2.90	11.41	7.29	1.07	9.69	13.34	11.11	1.02	0.99	11.47	0.45	10.69	1.27	9.59
Na_2O	2.80	4.74	2.16	5.14	9.05	4.19	1.65	3.61	9.99	7.29	2.98	6.62	3.15	7.99	3.80
K_2O	0.77	3.62	0.57	2.00	4.93	1.52	1.27	1.71	5.39	5.47	0.98	4.52	1.10	5.14	3.16
P_2O_5	0.42	0.18	0.28	1.02	0.15	0.56	0.30	0.99	0.09	0.09	0.67	0.02	0.55	0.03	0.94
Ba	191	247	194	630	305	791	496	582	87	61	307	9	295	298	870
Be	1.1	8.6	0.9	2.5	6.8	1.8	0.6	2.1	7.9	3.7	1.4	8.0	1.3	7.0	1.6
Cr	81	27	616	6	< 3	405	714	213	< 3	< 3	542	< 3	234	< 3	10
Cu	98	14	62	24	4	61	79	60	11	9	68	8	59	12	33
Ga	22	38	17	26	26	23	11	21	29	26	22	31	21	31	23
Hf	5.95	6.52	3.62	10.46	20.13	5.76	2.85	7.22	16.9	9.88	6.40	24.7	5.84	23.7	8.84
Nb	21.5	46	16.8	56	140	55	41	85	240	183	54	210	49	179	75
Ni	78	40	337	<2	2	270	241	120	< 2	< 2	276	< 2	118	2	8
Pb	2	15	<2	5	9	6	< 2	9	21	11	10	17	5	19	—
Rb	15.4	110	10.6	41	133	49	31	46	174	80	21	151	25	159	76
Sr	395	317	446	1251	28	899	429	1589	17.0	27	735	3.4	589	25	1261
Ta	1.50	3.28	1.17	3.56	7.81	4.06	2.84	5.75	16.6	13.7	4.25	15.6	4.05	17.0	6.49
Th	1.64	3.7	1.3	3.1	8.8	5.61	2.81	6.51	28.7	12.5	4.54	20.5	4.29	27.4	10.1
U	1.0	2.4	< 0.4	1.4	2.6	1.43	0.74	1.53	7.23	1.94	1.20	6.02	1.26	3.97	2.24
W	< 2	< 2	< 2	< 2	4	< 2	< 2	< 2	5	< 2	< 2	7	< 2	3	< 2
Y	42	46	24	48	49	27	18	37	38	41	27	98	29	52	35
Zn	119	116	87	106	188	122	60	103	126	137	103	178	93	182	108
Zr	227	171	162	506	935	266	142	400	961	566	279	959	255	1115	349
La	24	24	16.5	52	78	46	29	70	118	134	41	120	40	161	78
Ce	53	49	43	129	150	83	55	132	191	248	82	234	78	265	159
Pr	6.9	5.2	5.1	14.5	13.5	8.3	6.5	14.83	15.1	23.7	9.63	22.5	8.6	19.7	16.93
Nd	35.1	25.3	24.1	70	57	36	27.6	63	60	104	43	98	37	82	70.2
Sm	8.9	8.0	5.73	15.1	9.5	7.8	5.05	11.38	8.75	14.5	8.63	18.1	7.36	11.0	11.91
Eu	2.98	2.35	1.88	4.94	2.23	2.63	1.53	3.54	1.94	3.73	2.71	0.99	2.37	1.99	3.56
Gd	9.1	9.7	5.5	13.4	7.5	7.0	4.25	9.35	7.0	10,3	7.25	16.2	6.55	8.8	9.05
Dy	7.58	9.45	4.57	9.79	8.78	5.21	3.39	6.73	6.17	8.64	5.19	17.87	5.15	8.30	6.24
Ho	1.41	1.45	0.83	1.74	1.71	0.91	0.62	1.20	1.21	1.59	0.88	3.60	0.94	1.68	1.11
Er	3.64	3.03	2.01	4.53	5.28	2.48	1.68	3.05	—	—	2.30	10.59	2.47	—	2.91
Yb	3.04	1.78	1.60	3.48	5.70	1.77	1.44	2.33	3.63	3.43	1.54	10.33	2.01	5.12	2.15
Lu	0.42	0.21	0.23	0.52	0.87	0.28	0.18	0.29	0.56	0.50	0.18	1.52	0.28	0.79	0.29

	3113	2649	3105	BF5-4	BD5-4	27857	64064	26650	27196	EU 92	2211	MS 377	MS 334	US 26	EU 142
SiO_2	46.82	65.38	68.55	40.69	38.91	48.77	51.47	50.90	48.17	52.85	74.80	51.96	38.54	44.66	51.09
TiO_2	2.95	0.57	0.31	5.30	5.46	0.86	1.69	1.83	0.73	1.41	0.08	1.54	2.76	2.92	2.02
Al_2O_3	15.15	12.79	8.22	12.32	10.83	16.86	16.65	16.02	18.04	16.17	12.87	16.97	12.03	15.45	15.58
$Fe_2O_3^*$	13.11	7.80	9.20	14.08	15.33	12.64	9.48	9.65	9.67	10.83	1.82	8.10	12.04	12.09	10.56
MnO	0.19	0.28	0.34	0.20	0.19	0.21	0.14	0.15	0.16	0.13	0.08	0.12	0.25	0.19	0.13
MgO	7.20	0.17	0.00	9.37	13.76	7.81	7.58	8.62	10.10	6.40	0.00	7.47	8.40	8.86	7.37
CaO	10.62	1.49	0.57	10.96	10.43	12.00	8.01	8.18	11.19	7.96	0.74	7.13	12.92	9.59	7.94
Na_2O	3.05	6.71	6.89	2.29	1.89	1.28	4.09	3.76	2.55	3.97	4.31	4.37	2.63	4.04	3.87
K_2O	0.93	4.84	4.36	2.69	1.32	0.07	1.10	1.17	0.12	0.59	4.70	2.19	1.64	2.06	1.84
P_2O_5	0.74	0.05	0.02	1.29	1.29	0.04	0.38	0.53	0.06	0.18	0.003	0.47	0.84	0.64	0.47
Ba	380	200	86	1780	1457	98	249	295	59	223	17	837	1062	402	724
Be	0.8	5.6	13.7	2.4	2.2	0.2	1.2	1.3	0.1	0.6	7.2	1.2	2.6	1.6	1.2
Cr	94	< 3	< 2	20	231	181	220	289	154	186	< 3	250	384	181	186
Cu	58	16	17	111	83	93	36	35	77	42	12	18	—	55	41
Ga	17	36	33	20	17	14	18	19	15	22	18	21	—	20	21
Hf	3.58	18.8	37.8	10.8	10.9	0.72	4.18	4.07	0.88	3.09	5.9	4.61	5.26	5.8	4.07
Nb	32	182	506	162	162	0.2	16.1	18.6	0.7	9.8	34	14.5	88	68	36
Ni	62	< 2	< 2	23	68	35	154	172	176	129	< 2	174	224	153	155
Pb	—	6	20	6	7	9	6	4	< 2	5	33	6	10	4	7
Rb	13.6	106	230	53	36	1.9	16.1	19.1	2.6	15.6	323	26	43	45	39
Sr	490	9.4	8.0	872	518	183	523	541	212	346	17.9	1049	1014	689	744
Ta	2.34	14.9	35.1	10.6	10.7	0.015	1.24	1.28	0.035	0.62	3.49	0.95	4.63	5.34	2.56
Th	1.99	15.7	44.5	9.33	11.2	< 0.12	1.92	2.15	0.20	1.74	59.0	3.53	8.76	6.28	3.90
U	0.48	5.10	13.1	2.96	2.88	0.053	0.64	0.74	0.072	0.42	16.9	0.73	—	1.48	0.82
W	< 2	6	9	38	4	< 2	< 2	< 2	< 2	< 2	< 2	< 2	—	< 2	< 2
Y	28	83	191	32	31	16	30	33	20	18	51	17	28	31	20
Zn	86	230	502	117	96	81	72	77	48	98	56	74	—	85	103
Zr	149	757	2240	410	379	30	204	243	47	103	174	232	263	281	208
La	28	118	290	106	112	1.61	20	23	2.06	13	71	30	57	44	33
Ce	60	219	541	184	196	2.63	42	55	3.79	24	130	70	103	84	65
Pr	6.64	21.6	49.3	19.3	22.1	0.60	4.72	5.8	0.80	2.8	11.4	7.1	—	9.0	5.9
Nd	32	91	189	75.5	87.2	4.10	22	26.3	5.0	13.4	47	31.4	48	38	26.8
Sm	6.73	17.1	36.2	12.13	13.5	1.60	5.10	5.9	1.90	3.7	8.0	5.7	9.63	7.44	5.6
Eu	2.65	3.08	5.08	3.57	3.95	0.61	1.63	1.84	0.78	1.27	0.18	1.72	2.92	2.35	1.8
Gd	6.35	15.6	33.7	9.0	9.9	2.0	4.85	5.5	2.4	3.7	6.4	4.4	7.23	6.75	4.7
Dy	4.97	14.9	33.0	6.25	6.60	2.45	4.58	4.99	3.03	3.21	6.35	3.11	—	5.42	3.52
Ho	0.92	2.97	6.66	1.10	1.14	0.51	0.89	0.98	0.66	0.59	1.28	0.57	—	1.02	0.62
Er	—	8.26	18.71	2.71	2.83	1.53	2.55	2.55	2.04	1.35	3.86	1.43	—	2.75	1.43
Yb	1.89	7.99	17.89	1.89	1.78	1.51	2.43	2.51	2.14	1.27	4.23	1.17	1.51	2.42	1.13
Lu	0.26	1.21	2.61	0.24	0.21	0.21	0.35	0.38	0.34	0.18	0.64	0.20	—	0.33	0.15

Appendix table A 1 (*cont.*)

(For key to samples, see table A 2)

	EU 100	EU 58	EU 47	K 38T	K 53T	TL 2	TL 3	TL 4	TL 5	4342	EU 90	3491	PB 98	PB 48	DT 770
SiO_2	39.10	43.88	53.41	41.70	51.74	46.62	47.48	47.54	47.33	43.15	52.53	45.89	51.15	62.61	73.33
TiO_2	2.86	2.91	0.37	3.60	1.16	2.69	2.14	2.64	2.64	3.71	1.84	2.32	1.62	0.39	0.33
Al_2O_3	10.74	11.72	19.66	12.02	19.28	14.56	14.94	11.56	11.53	12.24	15.12	13.83	15.73	17.62	12.30
$Fe_2O_3^*$	11.48	12.54	4.47	15.50	6.45	13.89	13.23	14.14	14.29	13.08	10.60	12.84	12.30	5.25	4.05
MnO	0.21	0.18	0.31	0.21	0.23	0.22	0.20	0.18	0.19	0.20	0.13	0.19	0.15	0.14	0.08
MgO	14.64	11.02	0.61	10.44	1.48	5.92	6.51	9.78	10.01	8.54	7.76	10.77	6.21	0.39	0.14
CaO	14.57	12.88	9.61	11.60	3.79	10.94	11.27	10.34	10.32	11.83	7.97	10.03	9.54	2.91	0.79
Na_2O	2.95	3.16	5.68	2.05	9.66	3.06	3.60	2.69	2.56	2.03	3.65	2.57	2.77	5.91	4.90
K_2O	1.69	1.23	5.60	1.63	4.77	1.18	0.88	1.21	1.05	3.40	1.44	1.29	0.78	5.00	4.85
P_2O_5	0.85	0.83	0.15	0.92	0.94	0.52	0.33	0.33	0.31	0.62	0.32	0.53	0.17	0.10	0.01
Ba	1167	563	2587	596	915	489	330	422	411	1119	554	401	135	2864	135
Be	1.3	1.6	8.2	0.9	10.4	1.1	0.9	1.5	1.6	1.5	0.8	1.2	0.5	1.8	4.8
Cr	617	278	< 3	403	6	76	85	413	448	369	238	355	119	< 3	8
Cu	65	72	24	69	17	124	107	97	98	43	46	57	49	18	10
Ga	16	19	30	16	25	20	19	21	19	20	21	21	19	28	42
Hf	7.7	6.53	5.66	4.81	22.0	4.07	3.02	4.40	4.29	10.4	3.71	4.37	2.93	12.1	21.1
Nb	109	72	433	51	360	38	28	45	45	108	25.5	46	6.6	33	90
Ni	275	206	2	148	< 2	57	70	207	225	74	184	267	105	< 2	< 2
Pb	7	< 2	41	< 2	10	5	< 2	5	6	10	3	5	8	9	15
Rb	55	36	215	48	169	22	19.5	31	27	127	35	34	27	53	101
Sr	1068	991	3213	1172	992	421	358	544	521	1005	592	652	229	218	22
Ta	9.02	5.53	20.6	3.51	23.8	2.51	1.80	2.98	2.83	9.12	1.82	3.15	0.52	2.87	6.35
Th	10.3	7.00	51.9	4.11	35.0	4.26	2.84	6.54	6.26	11.3	2.94	3.48	3.13	3.33	9.16
U	3.32	1.47	18.5	0.81	4.13	—	—	—	—	2.25	0.70	0.97	0.31	1.20	2.00
W	97	< 2	<2	< 2	< 2	< 2	< 2	< 2	< 2	130	< 2	< 2	< 2	< 2	4
Y	27	22	15	33	43	37	31	26	25	29	21	28	20	32	82
Zn	79	91	236	103	122	89	79	95	96	97	102	92	109	78	187
Zr	295	242	599	209	1270	172	131	179	179	306	158	218	94	639	750
La	90	72	240	56	93	35	24	43	43	81	24	31	15	34	154
Ce	153	134	364	109	155	68	47	78	77	164	47	61	31	67	285
Pr	15.4	13.0	31.6	11.5	15.2	7.2	5.0	7.9	7.8	14.9	5.0	6.7	3.23	6.0	31.7
Nd	66	53.3	97	51	56	33.7	24.1	35.4	35.0	64	23.3	30.3	15.0	28.6	129
Sm	10.8	9.1	9.13	9.56	9.4	7.15	5.25	6.85	6.85	10.7	5.2	6.46	3.81	6.3	22.9
Eu	3.07	2.75	2.15	3.11	2.82	2.51	1.92	2.24	2.24	2.97	1.71	2.11	1.16	5.70	2.14
Gd	8.0	7.0	5.3	8.15	7.3	7.4	5.6	6.3	6.4	8.1	4.7	5.95	3.75	5.7	18.6
Dy	5.65	4.5	3.18	5.95	6.27	6.07	4.92	4.58	4.58	5.51	3.56	4.96	3.62	4.78	16.69
Ho	0.98	0.79	0.61	1.07	1.29	1.17	0.95	0.80	0.81	1.01	0.66	0.90	0.67	0.93	3.19
Er	—	2.01	—	2.70	3.85	—	—	—	—	—	1.46	2.38	1.80	—	—
Yb	1.65	1.42	1.18	2.12	4.77	3.00	2.41	1.69	1.67	2.11	1.23	1.93	1.53	2.55	8.12
Lu	0.23	0.20	0.17	0.26	0.80	0.44	0.37	0.24	0.25	0.34	0.18	0.25	0.21	0.47	1.24

	L 302	40622	3090	63758	6982	2403	64067	MS 234	MS 143	MS 104	MS186	MS 392	63760	ME 59	EU 121
SiO_2	77.71	73.62	71.33	60.65	54.97	65.33	47.66	51.57	48.55	52.98	60.49	76.55	70.96	72.74	68.30
TiO_2	0.05	0.18	0.08	0.59	0.73	0.48	1.16	0.25	0.55	1.27	0.98	0.13	0.40	0.32	0.47
Al_2O_3	11.77	7.35	14.07	17.63	21.63	14.96	16.97	6.86	10.46	18.18	16.81	13.34	14.94	14.78	15.83
$Fe_2O_3^*$	1.39	8.72	2.38	4.10	3.90	4.20	10.25	12.94	9.22	8.54	5.19	1.18	2.89	2.22	3.14
MnO	0.02	0.08	0.06	0.07	0.18	0.08	0.16	0.20	0.15	0.11	0.07	0.04	0.03	0.05	0.05
MgO	0.00	0.04	0.00	0.99	0.82	2.29	9.60	21.33	18.14	5.52	3.91	0.28	1.21	0.66	1.66
CaO	0.19	0.31	0.17	2.92	4.72	3.79	12.45	8.04	7.80	8.25	6.07	0.67	0.85	1.13	2.88
Na_2O	4.49	5.55	6.98	5.34	7.04	3.35	2.00	0.14	2.37	3.66	5.22	3.93	3.26	3.59	3.54
K_2O	4.02	3.79	4.06	5.87	5.07	5.40	0.13	0.07	2.65	1.23	2.15	4.50	5.47	4.57	4.02
P_2O_5	0.00	0.14	0.00	0.24	0.13	0.44	0.10	0.01	0.25	0.31	0.36	0.04	0.29	0.28	0.16
Ba	3.8	355	22	4441	2061	1840	140	23	523	753	1159	488	283	189	374
Be	5.6	11.5	14.1	1.6	3.7	13.6	0.2	0.2	1.4	1.1	1.9	8.6	5.4	5.9	9.8
Cr	< 3	4	< 3	6	9	20	273	961	813	84	67	< 3	13	9	31
Cu	6	17	5	67	16	29	53	75	52	38	13	< 3	10	5	10
Ga	47	49	44	25	25	19	15	—	12	23	—	—	23	25	23
Hf	14.2	38.4	29.8	1.78	8.64	9.80	1.06	0.28	1.91	3.32	4.75	4.43	5.77	4.31	4.26
Nb	89	110	114	10.3	185	33	1.5	< 0.5	2.4	4.8	10.7	30	19.6	12.4	11.9
Ni	< 2	5	3	< 2	< 2	5	86	306	640	48	47	< 2	4	8	11
Pb	18	—	47	47	8	68	4	< 2	15	3	15	21	27	31	58
Rb	533	146	174	100	152	370	2.8	3.1	65	23.4	43	408	497	466	291
Sr	2.1	18.8	3.1	5583	1980	774	257	42	628	1561	1349	142	101	87	207
Ta	15.1	6.60	8.39	2.05	9.52	6.10	—	—	0.24	0.35	1.07	7.20	4.31	2.70	3.28
Th	31.1	10.1	19.9	5.56	23.7	160	0.02	0.15	2.59	2.06	1.93	25.6	28.4	26.1	22.4
U	7.15	3.38	4.81	1.09	7.14	54.0	< 0.04	< 0.05	0.91	0.62	1.15	6.38	16.7	6.60	9.64
W	7	4	7	< 2	< 2	13	< 2	< 2	< 2	< 2	< 2	3	5	2	< 2
Y	129	113	204	24	18	22	13	1.5	14	13	14	40	25	13	21
Zn	125	268	245	58	75	41	59	62	71	82	65	30	45	57	54
Zr	153	1412	708	357	473	264	42	12	87	135	234	95	149	131	139
La	69	80	51	67	105	57	2.90	2.0	15.3	16	36	27.5	46	37	33
Ce	126	157	124	128	142	104	5.60	2.0	26	37	56	53	84	79	62
Pr	12.0	19.0	17.7	14.2	12.3	10.2	0.93	0.30	2.90	4.3	6.0	5.1	10.7	8.8	6.8
Nd	40	84	84	61	37.2	43	5.9	1.5	15.1	19.1	23	20.7	45	34.7	27.1
Sm	9.0	17.9	24.6	9.7	4.0	7.95	1.91	0.4	3.6	3.10	4.03	4.55	9.0	6.43	5.33
Eu	0.13	3.51	1.76	3.02	1.40	1.72	0.99	0.11	1.03	1.34	1.25	0.37	0.86	0.67	0.94
Gd	8.4	17.6	27.0	6.5	3.1	5.7	2.05	0.20	3.0	3.0	3.0	4.3	6.5	4.1	4.0
Dy	13.38	21.51	32.94	4.03	2.86	3.67	2.16	0.27	2.29	2.41	2.17	4.85	4.78	2.41	3.23
Ho	2.97	4.51	6.63	0.73	0.66	0.68	0.42	0.05	0.44	0.47	0.42	0.97	0.85	0.43	0.61
Er	10.07	13.36	19.59	2.45	—	—	—	—	—	1.18	1.12	—	2.26	1.02	1.67
Yb	12.92	14.0	18.40	1.49	1.98	2.00	1.16	0.17	1.22	1.11	1.08	3.92	2.08	0.87	1.63
Lu	1.80	2.12	2.52	0.28	0.32	0.32	0.17	0.04	0.22	0.17	0.17	0.59	0.30	0.12	0.25

APPENDIX TABLE A 1 (*cont.*)

(For key to samples, see table A 2)

	EU 15	US 30	US 34	US 25	L 1	63350	7993	US 8	MAD 1	MAD 2	22650	63672	64065	EU 13	1093
SiO_2	70.61	68.82	76.45	77.20	58.10	43.06	50.72	50.51	44.65	45.43	36.55	29.41	35.48	46.85	62.22
TiO_2	0.35	0.50	0.06	0.08	0.90	2.08	1.05	1.86	2.25	2.30	5.58	3.32	2.92	3.25	0.69
Al_2O_3	15.08	14.50	12.47	12.09	15.46	9.95	11.30	10.45	8.16	8.38	5.34	5.86	2.09	15.30	17.56
$Fe_2O_3^*$	2.42	3.11	1.22	1.17	5.16	9.17	7.50	8.90	6.38	6.44	7.05	14.39	12.51	8.31	3.20
MnO	0.04	0.06	0.05	0.08	0.08	0.14	0.16	0.13	0.12	0.12	0.14	0.26	0.19	0.17	0.27
MgO	1.01	1.26	0.01	0.02	5.69	15.37	11.92	10.57	12.30	12.61	7.73	25.33	36.14	3.42	0.48
CaO	1.87	2.47	0.54	0.28	5.26	10.08	9.76	9.72	14.06	12.67	21.51	16.88	8.68	9.23	1.20
Na_2O	3.08	3.77	4.48	4.55	4.81	0.74	0.87	2.65	0.93	1.01	0.52	0.00	0.00	2.97	7.90
K_2O	4.65	4.56	4.33	4.16	4.32	5.52	5.71	4.84	7.18	7.90	7.74	2.59	1.64	9.26	5.88
P_2O_5	0.23	0.13	0.01	0.00	0.80	3.10	1.23	1.06	3.04	1.88	2.21	1.66	0.87	0.65	0.46
Ba	352	1162	29	27	2650	9024	4120	1995	5844	7404	3.4 %	1539	1008	1644	3.6
Be	8.6	2.0	3.8	6.1	3.1	3.7	6.1	3.5	7.3	7.8	3.8	1.8	1.9	14.2	13.6
Cr	23	11	4	5	153	551	563	399	488	523	56	625	1048	4	< 3
Cu	8	13	5	< 3	29	42	45	51	52	59	97	62	63	79	9
Ga	23	17	17	25	21	15	14	20	19	22	16	9	5	18	27
Hf	3.84	7.53	4.79	9.77	9.33	16.1	8.18	8.93	40.8	39.3	48.2	9.49	12.3	8.80	19.8
Nb	11.5	11.4	20.9	96	18.1	32	11.2	63	131	128	127	213	123	22	107
Ni	10	< 2	< 2	< 2	106	445	405	296	173	184	88	492	1514	20	< 2
Pb	51	19	29	44	19	12	48	30	54	58	23	11	4	95	55
Rb	341	217	195	208	75	135	186	106	201	228	454	110	79	512	498
Sr	203	331	6.8	1.2	4631	1687	2063	1500	3663	5433	2041	1321	737	2457	6.9
Ta	2.29	2.52	2.28	8.09	1.31	1.98	0.90	4.15	6.52	8.92	3.59	17.3	13.1	1.15	5.77
Th	17.8	29.3	20.7	21.2	10.6	16.4	39.6	37.8	37.8	49.8	6.06	24.5	12.0	63.8	76.0
U	11.9	6.48	6.00	8.35	2.75	3.97	6.99	8.81	8.43	11.1	2.27	6.34	3.08	14.5	15.2
W	3	< 2	5	11	< 2	< 2	< 2	< 2	< 2	< 2	< 2	< 2	< 2	136	< 2
Y	20	20	23	59	21	32	32	36	27	26	17	24	10	42	76
Zn	69	43	42	86	79	121	72	103	98	103	62	82	70	83	96
Zr	135	186	107	168	622	709	379	414	1587	1734	1771	363	370	450	814
La	33	39	27	39	99	165	120	148	347	338	31	164	84	136	211
Ce	61	54	56	78	224	391	235	256	605	630	50	278	174	248	262
Pr	7.2	6.7	5.1	7.2	24.2	44.3	25.1	29.4	69.0	69.9	6.4	30.7	16.9	25.4	36.5
Nd	31.0	26.8	20.5	30	100	191	112	121	258	265	23.4	119	67	105	130
Sm	6.3	4.9	4.34	7.4	15.1	27.2	19.05	19.71	30.25	30.6	2.9	17.04	9.8	17.45	20.6
Eu	0.87	0.92	0.17	0.12	4.01	6.43	5.42	4.90	6.78	6.82	0.96	4.55	2.69	3.82	1.33
Gd	4.4	3.6	3.8	7.8	9.4	15.7	13.4	12.95	16.05	16.0	2.0	11.15	6.6	12.6	15.5
Dy	3.28	2.90	3.92	9.25	4.53	6.75	6.24	6.63	6.48	6.69	2.41	5.81	3.59	7.05	13.56
Ho	0.55	0.57	0.81	1.89	0.79	1.14	1.05	1.12	1.15	1.10	0.43	0.95	0.59	1.21	2.79
Er	—	1.63	2.29	5.62	1.96	2.74	—	2.87	—	—	1.14	2.59	1.48	—	7.97
Yb	1.22	1.66	2.53	6.10	1.31	1.66	1.97	1.80	1.33	1.28	1.15	1.15	0.69	2.39	7.41
Lu	0.18	0.27	0.38	0.89	0.20	0.24	0.29	0.25	0.19	0.20	0.19	0.15	0.09	0.34	1.13

	AMT 53	AMT 54	19 W	487 CS
SiO_2	57.69	60.50	59.08	45.81
TiO_2	1.14	0.85	1.23	0.46
Al_2O_3	24.93	16.80	23.16	10.99
$Fe_2O_3^*$	8.34	7.51	9.03	22.69
MnO	0.22	0.11	0.37	6.86
MgO	2.04	4.66	2.43	3.51
CaO	0.54	5.28	0.24	1.56
Na_2O	0.73	1.63	0.83	3.98
K_2O	4.34	3.76	4.37	2.11
P_2O_5	0.21	0.14	0.19	3.44
Ba	648	498	703	6875
Be	3.5	2.5	3.0	2.1
Cr	105	127	114	60
Cu	18	25	21	301
Ga	29	21	27	11
Hf	5.5	6.3	5.06	2.63
Nb	20.1	13.6	20.9	5.1
Ni	34	93	45	324
Pb	16	13	133	70
Rb	195	146	186	51
Sr	122	127	85	319
Ta	2.01	1.30	1.78	0.32
Th	18.6	12.2	16.8	4.06
U	2.36	3.23	2.62	2.37
W	< 2	< 2	< 2	< 2
Y	36	29	30	84
Zn	84	84	92	285
Zr	154	156	189	124
La	54	37	24	76
Ce	117	71	63	50
Pr	9.6	7.0	4.7	12.5
Nd	44	30	19.7	58
Sm	7.95	5.85	3.75	12.05
Eu	1.78	1.37	0.80	3.13
Gd	6.8	5.3	3.20	14.0
Dy	5.49	4.32	3.87	13.45
Ho	1.04	0.84	0.84	2.88
Er	—	—	—	8.10
Yb	3.05	2.56	2.97	7.47
Lu	0.47	0.41	0.46	1.17

Notes

Oxides in percentage mass, trace elements in micrograms per gram.

All Fe reported as Fe_2O_3.

Major elements on a 'dry' basis, except for MS 334 and EU 13 (Thompson 1977*b*). Analyses on ignited powders of RTH 13, RTH 24, 27001, 844, 26650, 27196, EU 92, EU 100 and PB 98.

All samples crushed in agate, except for BF 5-4, BD 5-4, EU 100, 4342 and EU 13 (tungsten carbide).

Major elements analysed by X.r.f.; Be, Pr, Sm and middle-heavy r.e.e. by ICP; U, Th, Hf and Ta by INAA; other trace elements by X.r.f.

X.r.f. analyses made with a Philips PW 1450 spectrometer. Major elements determined on 40 mm fusion beads (1:5 Spectroflux 100B). Trace elements on 40 mm powder briquettes. For details see Hendry (1982, 1983). Approximate Cl determined in the following samples (X.r.f.): RTH 27, 0.1 %; 2649, 0.2 %; 3105, 0.6 %; 2211, 0.3 %; US 25, 0.1 %.

ICP analyses made at Imperial College (Be) and Kings College, London (r.e.e.).

INAA analyses were made at the Reactor Centre, Imperial College at Silwood Park, with epithermal neutron activation under a cadmium filter. U, Th and Ta were measured after 20 h irradiation as ^{239}Np, ^{233}Pa and ^{182}Ta, respectively. Hf was measured after a 30 s irradiation as $^{179m}Hf\{T_{\frac{1}{2}}: 18.7\ s\}$ (Parry 1982).

TABLE A 2. KEY TO SAMPLES IN TABLE A 1

sample	rock type	location
RTH 13	basalt	1955 flow. Roadside on Rte. 130, Kilauea, Hawaii
RTH 7	rhyodacite	Kauaopuu Ridge, Waianae, Oahu
RTH 24	ankaramite	Roadside on Rte. 190, Mauna Kea, Hawaii
27001	hawaiite	Popoo Gulch, Mauna Kea, Hawaii
RTH 27	trachyte (obsidian)	Puu Waawaa, Haulalai, Hawaii
RTH 31	basanite	Post-erosion flow. Puu Kilea, Maui
877	ankaramite	Jameson Bay, Jan Mayen
5066	basanite	Tenerife
21468	phonolite	Tenerife
418	trachyte	N of Guinamar, Grand Canaria
21101	basanite	Funchal, Madeira
616	rhyolite	Ascension
844	ankaramite	St Helena (Baker 1969)
789	trachyte	St Helena (Baker 1969)
20171	trachybasalt	Tristan da Cunha
3113	basalt	Pantelleria
2649	trachyte	Pantelleria
3105	pantellerite	Pantelleria
BF 5-4	nephelinite	Bermuda Rise
BD 5-4	nephelinite	Bermuda Rise
27857	olivine basalt	W of Atami, Izu Peninsula, Japan
64064	olivine basalt	S of Tlalpan, Mexico
26550	olivine basalt	Xitli, Cuicuilco, Mexico
27196	olivine basalt	Warner flow, Medicine Lake Highlands, California
EU 92	basalt	Roadside at junction of N 442 and N 131, Uras, Sardinia
2211	rhyolite (obsidian)	Lipari
MS 377	basalt	Basal flow, SW end of Kerrera, Oban, Argyllshire (NM789276)
MS 334	monchiquite (nephelinite)	Dyke. S face of Streap, Argyllshire (NM 951858)
US 26	basanite	Dish Hill, California (Lherzolite xenoliths)
EU 142	basalt	Mt Cuiaru, NE of Bonorva, Sardinia
EU 100	melilitite	Plug. Hövenegg, Hegau, Germany
EU 58	limburgite	Type locality (Limburg), Kaiserstuhl, Germany
EU 47	wollastonite phonolite	Stock. Fohberg, Kaiserstuhl, Germany
K 38T	ankaramite	Dyke, Hobley Valley, Mt Kenya, Kenya
K 53T	phonolite	Barrow Ridge, Mt Kenya, Kenya
TL 2	basalt	Humbu lavas, Lake Natron, Tanzania
TL 3	basalt	Humbu lavas, Lake Natron, Tanzania
TL 4	basalt	Sambu lavas, Lake Natron, Tanzania
TL 5	basalt	1.85 Ma flow, Olduvai Gorge, Tanzania
4342	olivine augite leucitite	Bufumbira, Uganda
EU 90	basalt	Dyke. Genna Spina, Mt Arci, Sardinia
3491	basalt	Calton Hill, Derbyshire (Lherzolite xenoliths)
PB 98	gabbro	Nigeria
PB 48	syenite	Pankshin, Sara-Fier, Nigeria
DT 770	fayalite-ferrohedenbergite porphyry	Kila, Nigeria
L 302	hypersolvus biotite granite	Jere-Sanga, Nigeria
40622	aegirine granite	Rockall
3090	riebeckite microgranite	Ailsa Craig, Firth of Clyde, Scotland
63758	syenite	Glen Dessary, Argyllshire (NM 956924)
6982	syenite	Rouzac, Bagneres de Bigorre, Pyrenees, France
2403	syenite	Biella, Western Alps, Italy
64067	gabbro	Pitscurry Quarry, Insch Pluton, Aberdeenshire (NJ 728267)
MS 234	gabbro (pyroxene-rich cumulate)	Currywongaun, Connemara, Eire
MS 143	kentallenite (shoshonitic picrite)	Type locality (Kentallen), Argyllshire (NN 008573)
MS 104	diorite	Fault Intrusion. Cam Ghleann, Glencoe, Argyllshire (NN 250523)
MS 186	tonalite	Roadside on A861, Strontian, Argyllshire (NM 846610)
MS 392	adamellite	Foot of nursery drag lift, Cairngorm, Grampian

sample	rock type	location
63760	adamellite	Porthcurno, Cornwall
ME 59	adamellite	Hawk's Tor, Bodmin Moor, Cornwall (SX 255764)
EU 121	granodiorite	Mt Capanne pluton, Mt Perone, Elba, Italy
EU 15	cordierite rhyodacite	Quarry on San Carlo road, San Vincenzo, Tuscany, Italy
US 30	adamellite	June Lake Village, Sierra Nevada, California
US 34	rhyolite (obsidian)	North Crater, Mono Craters, California
US 25	rhyolite (obsidian)	Ridge between Alamos and Frijoles canyons, Jemez Mts, New Mexico
L 1	minette	Dyke. Avernish Point, Lochalsh, NW Scotland (NG 835257)
63350	minette	Dyke, Summer Isles, NW Scotland
7993	minette	Dyke. Helmside Gill, Sedberg, Cumbria
US 8	minette	Dyke. Agathla Neck, Monument Valley, Arizona
MAD 1	madupite	Leucite Hills, Wyoming
MAD 2	madupite	Leucite Hills, Wyoming
22650	wolgidite	Wolgidee Hills, West Kimberley, Australia
63672	kimberlite	Lipelenang, Lesotho
64065	kimberlite	Pipe 200, Lesotho
EU 13	augite leucitite	Lago di Albano, Italy (Thompson 1977*b*)
1093	trachyte	Scarrupata, Ischia, Italy
AMT 53	slate	Blaenau Ffestiniog, Gwynedd, Wales
AMT 54	slate	Cutting on A6 near Shap, Cumbria
19 W	slate	Skiddaw, Cumbria
487 CS	composite	subequal parts of samples 16-1-35/9, 18-4-97/103 and 18-5-89/93, hole 487, DSDP leg 66

Discussion

P. R. SIMPSON (*Institute of Geological Sciences*, 64/78 *Grays Inn Road, London WC*1*X* 8*NG, U.K.*). Has Dr Thompson considered the possible role of fluorine and carbon species to account for the partitioning of uranium and thorium that he attributes to partial melting? The high partial pressures of fluorine, carbon species and other volatiles involved in mantle metasomatism mitigate the requirement for the very low degrees of partial melting or extreme fractionation in the case he describes.

Rb, Sr and r.e.e. such as Nd would also be particularly mobile in such a régime and thus contribute to mantle zone-refining to form the source rocks enriched in l.i.l. and other incompatible elements which were partially melted to form the late discordant granites of Aberdeenshire such as Cairngorm to which he refers.

R. N. THOMPSON. We explain in our text why we are reluctant to accept non-magmatic metasomatic processes as the source of all terrestrial mafic magmatism. The evidence for Th/U fractionation during basic magma genesis reported by Allègre & Condomines (1982) includes samples of m.o.r.b. Viewing the evolution of geochemical theory during the last few years, we confidently predict that a metasomatic origin for m.o.r.b. will shortly become widely accepted. Nevertheless, we do not wish to be the first people to suggest it!

In table A 1 we have included analyses of the Cairngorm pluton and the tonalitic member of the Strontian complex. But we have not discussed the Caledonian granite suite of the British Isles in any detail because it would lengthen our paper considerably. Looking at batholithic granites (*s.l.*) in general, we identify two non-metasomatic processes which we suspect may be important in the genesis of so-called 'enriched' granites: (1) extensive fractional crystallization. Some of the published reports of 'enriched' granites discuss minor elements alone and it is not apparent to the reader that the major-element compositions of the plutons are those of strongly-fractionated magmas. The Cairngorm group of Aberdeenshire Caledonian plutons are much more acid than many of the other intrusions in the province with which they have been compared geochemically in some published accounts. (2) Alkalic magmas (shoshonitic or o.i.b.) as the basic mantle-derived input at the bottom of the column in which fractionation, crustal fusion and magma-mixing takes place. Our data (table A 1) show differences of between one and almost two orders of magnitude, at a constant value of MgO, in the abundances of strongly-incompatible elements (including Be) between various types of basic magmas.

R. Hutchison (*Mineralogy Department*, *British Museum* (*Natural History*), *Cromwell Road*, *London SW*7 5*BD*, *U.K.*). It is difficult to envisage the extraction of very small (below a few per cent?) proportions of magma from the upper mantle. Spinel lherzolite xenoliths from two French localities about 20 km apart have almost identical major and minor element compositions consistent with 'undepleted' upper mantle. Those from one locality are light r.e.e. enriched, with normal K and U abundances, whereas the others are light r.e.e. depleted, with very low K and U contents. One set of xenoliths seems complementary to the other, but because Ti, Al, Ca and Na are equally abundant, a magmatic means of transport for light r.e.e., K and U between the suites seems ruled out. Metasomatic mantle enrichment followed by large volume melting appears to be a more probable mechanism for the production of highly fractionated magmas than small volume melting.

R. N. Thompson. Dr Hutchison concentrates his attention on the rigid sub-solidus lithosphere' whereas we are concerned with magma genesis from the convecting asthenosphere and below. Obviously, very small amounts of melt are unlikely to escape from a rigid matrix. If you put a wet sponge on a table, it remains wet. The way to extract the water is to squeeze the sponge. Recent analyses of the physics of magma escape from *internally-deforming* mantle (Walker *et al.* 1978; Ahern & Turcotte 1979) all emphasize the impossibility of preventing liquid fractions greater than a few percent from migrating upwards. A possible exception may occur at sub-asthenospheric depths, if the densities of the liquids become equal to or greater than those of the surrounding mantle (Stolper *et al.* 1981). As we have explained in our text, we accept the sort of evidence that Dr Hutchison has summarized concerning the compositions of various types of lithospheric mantle. Where we differ from him is that he appears to have no difficulty in envisaging the sub-solidus metasomatized lithospheric mantle xenoliths which he describes as becoming transformed by an unspecified process into a major, largely-molten, near-surface magma source. We suggest alternatively that the logical way to fuse lithospheric mantle is to subduct it, either beneath an island arc or following continental collision, and thus to re-introduce it to the convecting sub-lithospheric part of the upper mantle. Our model carries the added advantage of permitting subducted sediment to reach the source of alkalic basic magmas, as required by the findings of several recent isotopic studies.

M. J. O'Hara, F.R.S. (*Department of Geology*, *University College of Wales*, *Aberystwyth*, *U.K.*). The authors have argued that their data base, and in particular their data for uranium and thorium, can only be interpreted in terms of very small to small percentages of fractional partial melting of source regions containing variable amounts of trapped liquid.

The operation of an r.t.f. magma chamber (see comments on paper by Dickin *et al.* in this volume) can readily cause trace element variations in erupted basalts to simulate those expected during the progress of equilibrium partial melting of an apparently homogeneous but 'fertile' source region (which would be identical with the parental magma plus contaminants entering the r.t.f. chamber). It is easy to envisage situations in which the resulting trace element chemistry would simulate that arising during the first 0.1% to 1% of partial melting. It is *also* possible for the products of an r.t.f. chamber to simulate those of perfect fractional partial melting of a similar source (so far as highly incompatible elements are concerned) between 0 and 0.16% melt extraction.

In less favourable circumstances, the products of an r.t.f. magma chamber, even with constant composition inputs, can simulate those expected to result from variable but in some circumstances small degrees of partial melting of heterogeneous source regions (see O'Hara & Mathews 1981). Contamination, and mineral facies variations in the source regions add still further to the complexities.

The persistent occurrence of several phenocryst species in the majority of erupted magmas indicates the possibility of fractional crystallization at one or more depths. If the plumbing systems involved do operate r.t.f. chambers the interpretation of geochemical data offered by the authors may be non-unique (see also discussion of paper by Sparks *et al.* this volume).

The authors do not appear to have attempted to evaluate their data along these lines, and I would welcome a statement of the criteria which they have used in determining that such evaluation would be superfluous for their data set.

The field evidence from ophiolite complexes, other peridotite masses and ultramafic nodule assemblages in kimberlite is unambiguous. Substantial melt fractions, i.e. *ca.* 30% by mass, have been removed from more fertile peridotites to leave the residual peridotites actually observed. The incompatible trace element concentrations in liquids and residues will be fundamentally different if this melt fraction is extracted by fractional or equilibrium partial melting. It appears to be the authors' contention that to fit their data it is necessary to use a continuous melting model in which the required total mass of melt is extracted as a series of very small aliquots from a system which retains a significant proportion of partial melt throughout the process.

It is not easy to visualize physical models of partial melting processes that lead to the output of significant successive volumes of magma whose compositions are controlled by the equations describing *perfect fractional partial melting*. One such model is that there are 'sultanas' of fertile mantle, which do undergo partial melting, carried upwards within a convecting 'pudding' of infertile or residual mantle which undergoes no significant partial melting. Requirements of the model are that the dimensions of the 'sultanas' in the direction of flow are small (*ca.* 1%) relative to the total path described through the partial melting zone, and that 'sultanas' are separated from each other spatially by distances of the order of size of the partial melting zone itself. How plausible this model is, particularly with respect to the survival of the required short dimensions of the 'sultanas' during large scale mantle convection, is debatable.

As soon as it is conceded that several 'sultanas' in different stages of the fractional melting process are contributing to the total emerging magma, or that 'sultanas' are sufficiently elongated to produce a similar effect, then the compositions of the extracted liquids will move towards those of *aggregated perfect fractional partial melts* which are described by relations that predict compositions differing relatively little in their properties from those obtained by equilibrium partial melting, which in turn cannot easily be distinguished from liquids emerging from an r.t.f. magma chamber.

Introduction of a trapped liquid fraction into the considerations in a *continuous partial melting process* has a marked effect on the equations in so far as they describe the behaviour of highly incompatible elements.

The conspicuous feature of perfect fractional partial melting without trapped melt which distinguishes it from equilibrium partial melting is that it causes the very rapid elimination of (highly incompatible) elements of very low distribution coefficient and the progressive elimination of elements of successively higher distribution coefficient as melting proceeds. The presence of trapped partial melt during a fractional (but not perfect fractional) partial melting event is to produce three régimes.

Where the mass fraction of trapped melt exceeds the total mass fraction which has been fractionally removed, the liquid products differ very little from those produced in equilibrium partial melting.

Where the mass fraction of extracted melt significantly exceeds that of trapped melt, the liquid products differ little from those produced in perfect fractional partial melting, but 'elimination' of highly incompatible elements is delayed to higher total percentages of melt extraction.

Where the mass fractions of trapped liquid and total melt extracted are more comparable the liquid products are transitional between those produced in equilibrium and perfect fractional partial melting (the continuous melting régime). It is desired to produce liquid in which very highly incompatible elements are depleted with respect to highly incompatible elements relative to their source region *and at the same time* generate concentrations of those elements in the liquids which are significantly higher than in the source region.

If this is to be achieved *in one operation* then there is apparently no alternative to *selecting* a narrow range of very low mass fraction near perfect fractional partial melts (rejecting the initial smallest fractions *which must not be allowed to aggregate* with the selected fractions) produced under conditions where there is very little trapped melt present.

If, however, it is conceded that most upper mantle source regions have been subjected to several previous cycles of partial melt extraction and that the extracted liquids are often subject to r.t.f. magma chamber evolution before eruption, preliminary studies suggest that the required effects can be produced in events in which significant (i.e. *ca.* 5 % total) aggregated fractional partial melts are removed from (depleted) source regions which contained substantial (i.e. *ca.* 10–20 %) trapped melt throughout the extraction process. This model predicts what should (in principle) be a detectable long term secular change in highly incompatible element behaviour in the Archaean at least.

However, before expending a lot of effort on attempting to demonstrate that this prediction is or is not accurate it is important to establish

(*a*) that the effect noted by the authors is general for all very low distribution coefficient elements,

(*b*) that it is restricted to very low distribution coefficient elements (as it should be if melting and fractionation processes dominate), and

(*c*) that there are no effects involving either formation of complex ions or diffusion or both which complicate the simple models considered thus far.

R. N. Thompson. We are grateful to Professor O'Hara for his extensive and thoughtful comments on our paper. Turning first to his later comments, we are gratified to see that he has reached conclusions extremely similar to our own, as to the style of magma genesis and geochemical evolution of the mantle necessitated by the array of abundances and ratios among variously incompatible elements, illustrated by our spidergrams for non-subduction-related basic rocks (see also Hofmann & White 1983). The second half of his discussion should be read in conjunction with our figure 3 and its legend, which it supplements and clarifies. Both of us have independently noted that the model implies slightly different secular geochemical evolution of the mantle in Archaean and post-Archaean times. We both find that the melt fractions necessary to generate the observed incompatible element patterns during simple incremental fusion of an originally-chondritic mantle are 'alarmingly low', to use our own words. We both consider the possibility that other processes may permit the model to operate at somewhat larger degrees of mantle fusion during the incremental steps. To achieve this result, O'Hara inevitably postulates r.t.f. chambers within all magmatic systems, while we prefer to appeal to limited recirculation of sialic crust (as oceanic sediments), via subduction zones into the upper mantle, for reasons that we discuss in our text (see also White & Hofmann 1982).

O'Hara asserts that the peridotites in ophiolite complexes, Alpine-type masses and ultramafic nodule suites are residua from more fertile primordial mantle after the extraction of about 30% of melt. This comes as a surprise to us because we have previously understood his position to be strongly opposed to any petrogenetic scheme involving the more-fertile variants of mantle periodotite (O'Hara *et al.* 1975). The view he expresses here is the same as our own, except that we are uncertain as to the fraction of extracted melt that relates fertile to residual mantle peridotite; Loubet & Allègre (1982) suggest a figure of only about 5%.

Turning to r.t.f. magma chambers, we have not discussed them in our text because their possibilities were explored exhaustively by O'Hara & Mathews (1981) and we had to be selective in the topics we covered in our text, for reasons of length. In addition, we felt confident that O'Hara himself would raise the matter, as he has indeed done! We have already published several discussions as to whether or not r.t.f. magma chambers occur beneath continental flood-basalt provinces (Thompson 1982; Thompson *et al.* 1982, 1983). Elsewhere in this volume we present the results of a detailed chemical study of a suite of Skye lavas which was undertaken specifically to test whether they might be products of an r.t.f. magma chamber.

O'Hara & Mathews (1981, p. 250) calculated that pairs of ultra-incompatible elements, such as U and Th, could become separated in an r.t.f. magma chamber after a very large number of cycles of magma replenishment and crystallization, but *only* during periods when *little or no* residual liquid was leaking out of the reservoir. Although some senescent long-lived volcanoes might fulfil these criteria, they do not seem appropriate for such examples as Etna or Kilauea, where frequent substantial effusions of magma occur. Nevertheless, both these volcanoes show Th-isotope evidence for U/Th fractionation (Allègre & Condomines 1982).

The universality of the r.t.f. process advocated by O'Hara depends, of course, on whether such chambers actually exist within every magmatic system. No less widespread a process would be a viable explanation for the chemical relations discussed in our paper because we are concerned with the whole of terrestrial magmatism. A classic r.t.f. chamber operated in the Early Tertiary beneath Rhum (Brown 1956; Palacz 1983) and presumably also beneath the *central intrusive complexes* in at least some of the other centres in the British Tertiary Province. But it seems unlikely that similar chambers are widespread beneath other volcanoes because the immense positive gravity anomalies with which they are associated are uncommon.

There are three other main points which dissuade us, at present, from joining O'Hara in his enthusiasm for r.t.f. processes as explanation for much of the incompatible-element variation in our data set.

(i) Throughout his writings O'Hara has stressed repeatedly his conviction that *high-level* magma chambers are common and that most (all?) mantle-derived magmas fractionate extensively in them on the way to the surface (see O'Hara & Mathews 1981, p. 275). If this is true, the cotectic precipitation of plagioclase-bearing assemblages from basic magmas would leave an unmistakable imprint on the spidergrams of the basalts finally erupted. The relatively high partition coefficient of Sr between plagioclase and basic magma means that the liquid in a cotectic r.t.f. magma chamber must progressively develop a large deficiency of Sr, *relative* to other incompatible elements; causing a trough at Sr in the chondrite-normalized spidergram. It is clear from our analyses (table A 1; figures 2 and 5) and published data that this is not the case for most magnesian basalts.

(ii) Mindful that the question of low-pressure cotectic crystallization would probably arise during the discussion of our work, we analysed several samples which either contained mantle peridotite xenoliths themselves (spinel or garnet facies) or came from igneous provinces where compositionally-similar rocks from adjacent localities contained such xenoliths. The numbers of these samples (table A 1) are: RTH 31, MS 334, US 26, EU 142, EU 100, 3491, US 8, 63672 and 64065. Obviously one could postulate that they have passed through r.t.f. magma chambers deep in the lithospheric upper mantle. Nevertheless, the Sr-in-plagioclase problem would remain, unless this phase was replaced by garnet, which would cause strong fractionation of Y and the heavy r.e.e., *relative* to Ti, P, Zr and Hf (Pearce & Norry 1979). The spidergrams of the samples listed above show no signs of such effects.

(iii) Iyer's survey (this volume) shows that enough geophysical data are now available to be reasonably sure that *large* crustal magma chambers (liquids plus semi-consolidated cumulates) occur beneath only a small proportion of currently-active volcanic centres. To produce fractionations of such element pairs as U and Th, the mechanisms described by O'Hara & Mathews (1981) require the development of *large and long-lived* r.t.f. magma chambers.

Finally, we agree with O'Hara's concluding remarks; especially about the need for more data to restrain the various theoretical possibilities. Table A 1 is our present contribution to this end.

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