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## Mantle plumes from ancient oceanic crust

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We propose the following model for the origin of “hot-spot” volcanism: Oceanic crust is returned to the mantle during subduction. It is separated from the surrounding peridotite, it sinks into the deeper mantle and accumulates at some level of density compensation, possibly the core-mantle boundary. The accumulated layer locally reaches thicknesses exceeding 100 km. Eventually, it becomes unstable as a consequence of internal heating, and the resulting diapirs become the source plumes of ocean island basalts (OIB) and other hot-spot volcanism. This mechanism may trigger upper mantle convection as suggested by Morgan [1]. Our model provides possible explanations for (1) the high trace element concentrations of OIB, (2) the seemingly contradictory isotopic evidence for both enrichment and depletion of magmaphile elements in OIB sources, (3) the phenomenon of apparent mantle isochrons in oceanic basalts, and (4) the apparently episodic nature of continental igneous activity. The model can be tested further, as more knowledge accumulates about the actual bulk composition of the oceanic crust, including its alteration products and sediment cover.

### 1. Introduction

The origin of oceanic islands and aseismic ridges has been a controversial subject throughout the development of plate tectonic theory. Wilson's [2] concept of hot spots was developed into the hypothesis of fixed mantle plumes by Morgan [1]. Schilling ([3] and in a series of subsequent papers) and Hart et al. [4] pioneered geochemical support for the plume idea, but no one has yet furnished conclusive proof for the existence of mantle plumes. In this paper, we adopt Morgan's hypothesis because we find his arguments as persuasive today as they were ten years ago and because, to the best of our knowledge, no viable alternative model has been proposed that would explain the kinematic and geochemical peculiarities of “hot-spot” volcanism (see also Hofmann et al. [5]).

We present the case that the source of plume volcanism is ancient, subducted oceanic crust. A preliminary version of the ideas expressed here was published in the Carnegie Institution of Washington Yearbook [6]. Our argument is essentially that plume material (i.e., the source of ocean island basalts, OIB) is geochemically “enriched” relative to the more “depleted” source of mid-ocean ridge basalts (MORB), but that it is neither primitive nor merely less depleted mantle peridotite. The terms enrichment and depletion refer to the concentrations of incompatible elements such as K, Rb, U, Th and the light rare earth elements.

The evidence for a non-primitive source of OIB is two-fold:

(1) The observed enrichments of incompatible elements in the OIB (both alkali basalts and tholeiites) are so high that conventional melting models using primitive sources indicate extraordinarily small degrees of melting. Such small degrees of melting are in conflict with most of today's con-

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ventional petrological thinking.

(2) The isotopic compositions of OIB also indicate non-primitive sources: Sr, Nd and Pb isotopic compositions usually indicate lower-than-primitive Rb/Sr and Nd/Sm ratios but higher-than-primitive U/Pb, Th/Pb and U/Th ratios.

Because the above evidence for the non-primitive nature of mantle plumes is rather conclusive, it is appropriate to investigate other geologically plausible source compositions. MORB appear to fit the geochemical requirements for OIB sources with a minimum of special pleading: MORB of any age have the approximate trace element concentrations, and ancient MORB may be inferred to have the approximate trace element ratios of modern OIB sources.

To give geological content to these ideas of source identification, we propose a rather specific geological model for the origin of mantle plumes. The model, if correct, does not necessarily invalidate other mantle evolution models, e.g. those linking the geochemical depletion of a portion of the mantle with the development of the continental crust. After a brief summary of previous geochemical earth models, we state the new model and then discuss it from several points of view, including major and trace element chemistry, isotopic chemistry, alteration of the oceanic crust, heat budget, mantle convection, and the possible timing of global magmatic "events".

## 2. Mantle models

### 2.1. Summary of previous geochemical mantle evolution models

This summary is intended to be representative rather than complete and it does not include those models that are primarily based on degassing history only. We hereby apologize to anyone whose model has been omitted.

Early attempts to explain the evolution of Pb isotopes were made by Patterson and Tatsumoto [7] who found from Pb isotope ratios of detrital feldspars that the source U/Pb increased with decreasing age. From this they formulated a mantle differentiation model with an early rapid move-

ment of U, Th, and Pb from the inner to the outer mantle, and a subsequent continued outward movement of U alone. This outer mantle was considered to be a proto-continental layer, from which the actual continents were formed at various times. This model was in apparent conflict with the model of Hurley et al. [8], in which continents are formed throughout geological history from previously undifferentiated mantle. Armstrong [9] and Armstrong and Hein [10] then devised a model involving recycling of continental material into the mantle while keeping the total volume of the continents constant. By using a decreasing rate or recycling with time, they were able to simulate both the Sr and Pb isotopic evolution. Doe and Zartman [11] removed the assumption of constant continental volume. Instead they assumed a constant growth rate and subdivided the crust into an upper, U-enriched, and a lower, U-depleted portion. Hurley's ideas about continued continental accretion were also reinforced by McCulloch and Wasserburg [12] who used Rb-Sr and Sm-Nd data on crustal composites and sedimentary rocks to show that large additions to the continental crust were made 2.6 Gyr. ago and that crust formation from the mantle continued in more recent times.

The slow trickle of crust-mantle evolution models has swelled to a flood since the first Nd isotopic data on oceanic basalts were published [13-15]. It was quickly recognized that Sr and Nd isotopic ratios in oceanic basalts are correlated, and this led to new estimates of the bulk-earth  $^{87}\text{Sr}/^{86}\text{Sr}$  and Rb/Sr ratios. These estimates are based on the assumption that the bulk-earth  $^{143}\text{Nd}/^{144}\text{Nd}$  ratio is chondritic and that the Nd-Sr correlation is produced by a sufficiently simple differentiation process which fractionated both the Rb/Sr and the Sm/Nd ratio in the mantle. The bulk-earth values meant that not only all the mantle sources of MORB but also most of the sources of OIB have had a time-integrated history of depletion of Rb and Nd relative to Sr and Sm. Using Nd and Sr isotopic averages for MORB and continental crust, ignoring the OIB as volumetrically insignificant, and assuming that the enrichment of continental crust had grown at the expense of the MORB-source depletion, O'Nions et al. [16], Allègre et al. [17], Jacobsen and Wasserburg [18],

and DePaolo [19] computed the volume fraction of the depleted mantle and average age of the crust. This simple division into three reservoirs – undepleted mantle, depleted mantle and continental crust – is difficult to reconcile with the evidence for U/Pb enrichment in the mantle. Only if the earth (or the time of core formation) is younger than 4.4 Gyr, could the average MORB source be classified as depleted in U relative to Pb. The most recent models take this “lead paradox” into account by postulating another reservoir, which serves as a storage place for the unwanted lead without affecting the Sr and Nd isotopic systematics. O’Nions and coworkers [16,20,21] postulate lead storage in the lower crust and recycling of upper crustal, high-U/Pb material into the mantle. Duprè and Allègre [22] and Allègre et al. [23] accomplish the necessary U/Pb increase in the mantle by allowing continued Pb loss to the core.

In all the above models, the distinctive geochemical character of the oceanic islands was largely ignored, because the volume of island basalts is small compared to the volumes of the oceanic and continental crust. The qualitative importance of oceanic islands was clearly recognized by Wilson [2], who proposed that they are formed by hot spots that are fixed in the mantle below the moving lithospheric plates. Morgan [1,24,25] argued that they are formed by deep-mantle convection plumes, which maintain an approximately constant position. Morgan did not attempt to identify the source material, but he did argue that the great age of Pb isotope anomalies in some OIB discovered by Gast et al. [26] indicates a turnover time of about 2 Gyr for mantle convection. Implicit in this suggestion is the recycling of near-surface material into the deep mantle, which is explicitly proposed in the present paper.

Schilling [3] proposed that the plume source is undepleted, i.e., essentially primitive lower mantle material, which contrasts with the depleted asthenosphere. This idea (and the entire plume hypothesis) was vigorously disputed by many geochemists who thought it more likely that the asthenosphere is enriched relative to the more depleted deeper mantle (e.g. [27]). The merits of these arguments were discussed by Hofmann et al. [5].

With the availability of Nd isotopic data it became clear that most OIB are *not* derived from primitive mantle but from a source with (time-integrated) relative light rare earth element (LREE) depletion. At the same time, the OIB themselves showed moderate to strong LREE enrichment as previously noted by Schilling. The resulting dilemma was how to make LREE-enriched basalts from a LREE-depleted mantle source. Wasserburg and DePaolo [28], elaborating on the basic model of Schilling [3] and Hart et al. [4], proposed that the non-primitive Nd isotopic composition could be explained by mixing initially primitive OIB material with upper-mantle-depleted MORB material. As noted above, this model has not been reconciled with the evidence from highly radiogenic Pb isotopes, which indicates a high and non-primitive U/Pb ratio in the OIB sources. Moreover, contamination with MORB material does not solve this dilemma because MORB-type Pb is *less* radiogenic than most OIB-type Pb. The “lead paradox”, first noted by Allègre [29], can be reconciled with an otherwise primitive plume source only by postulating a special removal mechanism for Pb by interaction with the core [23].

Other authors, more concerned with the problem of how to account for the high trace element concentrations of OIB from either primitive or depleted sources postulated metasomatic enrichment of the OIB sources prior to melting (e.g. [30]).

A special role for the subducted oceanic crust has previously been postulated by Ringwood [31] who proposed that this crust sinks to the base of the mantle and is thereby irreversibly lost from the convecting system. Tatsumoto and Knight [32] suggested that subducted oceanic crust could be recycled to form new oceanic crust. Anderson [33–35] has proposed an earth model in which the oceanic crust is stored in the region above 650 km and is then recycled to form new oceanic crust. In his scheme, the OIB source is an enriched peridotite layer in the asthenosphere. Although this model differs from the present one in several important respects, it is similar in the idea that neither MORB nor OIB are derived from primitive mantle.

## 2.2. General statement of the new model

Oceanic ridge volcanism differentiates mantle material into a 5- to 7-km-thick oceanic crust and a more refractory residue. The composition of the crust is initially basaltic and is then modified by hydrothermal alteration, low-temperature alteration, and the deposition of sediments.

Eventually, the crust is subducted into the mantle with further changes of composition through metamorphic processes (dehydration and decarbonation) and possibly partial melting. Despite all these modifications, the bulk composition of the crust thus returned to the mantle is still roughly basaltic. Its density is therefore different from the surrounding mantle material. In the upper mantle, where garnet is stable, the density of the subducted crust is almost certainly higher than mantle densities. In the deeper mantle, where the nature of the phase assemblage is less certain, the density contrast is also uncertain. Assuming that, because of the relatively high Fe, Si and Ca content, the density of the crust is higher than that of ordinary mantle *throughout* the depth of the mantle, we postulate that the crust separates from the associated refractory lithosphere, sinks into the deep mantle, perhaps to its base, and accumulates to form an irregular layer of "degenerate crust".

The layer of stored subducted crust eventually reaches local thicknesses of 100 km or more. Radioactive heat production within this layer and heat transfer from the core [36] cause the temperature to rise until the layer becomes unstable. Diapirs develop and ascend into the upper mantle. The diapirs (or plumes) begin to melt at a depth where upper mantle material is completely solid. The composition of the melt is modified by reaction with the upper mantle. This reaction, for reasons of mass balance, affects the concentrations of the major elements and the compatible trace elements much more strongly than the concentration of the incompatible trace elements. Finally, the melt reaches the surface and forms oceanic islands or continental "hot-spot" volcanoes (Fig. 1).

A tentative time scale for the entire cycle is suggested by apparent mantle isochrons of recent OIB.  $^{207}\text{Pb}/^{204}\text{Pb}$  versus  $^{206}\text{Pb}/^{204}\text{Pb}$  isochrons commonly yield ages of about 1.8 Gyr (see the

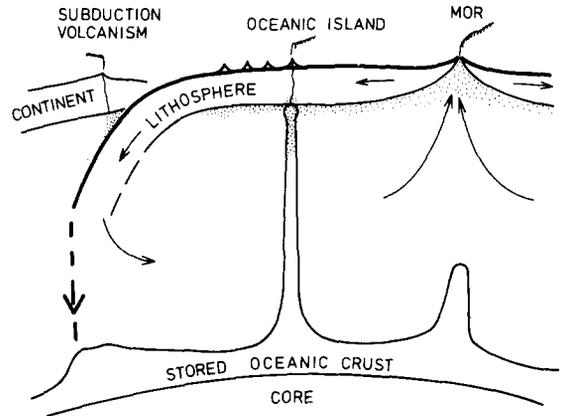


Fig. 1. Cartoon of the model. Oceanic crust is subducted and transformed into relatively dense eclogite and post-eclogite assemblages. It separates, sinks into the deeper mantle, and is accumulated and stored at the core-mantle interface. Eventually it becomes sufficiently hot to form plumes that rise to the surface and produce oceanic island volcanism.

recent compilation by Chase [37] and Rb/Sr apparent isochrons yield about 1.6 Gyr [38]. Alternatively, one may correlate periods of intense continental magmatism at 3.6, 2.7, 1.8, and 1.0 Gyr ago with times of plume-triggered mantle convection, and the intervening "quiet" periods with times of mostly stagnant conditions in the mantle, when the newly accumulated subducted crust was gradually heating up but had not reached the point of instability. Perhaps the two suggested

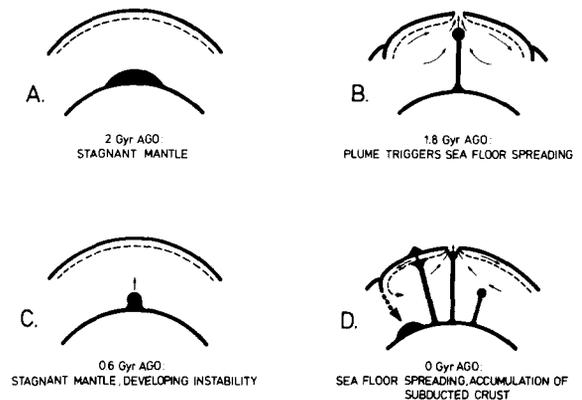


Fig. 2. Possible sequence of mantle events. Periods of plume-triggered mantle-wide convection alternate with stagnant periods, during which the stored oceanic crust gradually heats up.

time scales of 0.9 and 1.8 Gyr are not in conflict, but the regions where 1-Gyr-old accumulated crust dominates have not yet reached the temperature of instability. A possible sequence of intermittent mantle convection alternating with stagnant conditions is sketched in Fig. 2.

### 3. Elaboration and discussion

We believe the model outlined above offers possible solutions to several geochemical and geophysical puzzles. Among these are: (1) How can plume convection occur in an internally heated medium; (2) how can the abnormally high concentrations of magmaphile trace elements in OIB be reconciled with "reasonable" degrees of melting; (3) why is continental igneous activity episodic or perhaps periodic; (4) why are Rb/Sr and Nd/Sm ratios low, while the apparent U/Pb ratio is high, in most OIB sources relative to bulk-earth values. Perhaps the most simple-minded question is this: If any significant portions of the subducted oceanic crust are ever returned to the surface, what would be the geological consequences? It appears inevitable that the recycled oceanic crust would melt and that the melts would be distinguished by high concentrations of incompatible trace elements and a relatively high degree of chemical and isotopic heterogeneity. The group of basalts that fit these criteria best are clearly OIB (as well as some continental basalts).

In the sections that follow, we elaborate on several geochemical and geophysical aspects of the model and its consequence for the evolution of the earth.

#### 3.1. Subduction, separation and depth of accumulation

The effects of the subduction process on the bulk composition of the crust are difficult to assess quantitatively. The sediment cover may be scraped off or it may be subducted. Later, the crust will be metamorphosed. Dehydration and partial melting may remove significant but unknown amounts of incompatible elements from the crust. We assume that these processes are not thorough and the

remaining crust is still enriched in incompatible elements, especially K, U, Th. Some justification for this assumption may be derived from the mass balance of crust subduction and magma production in island arcs. Karig and Kay [39] have summarized intrusion, extrusion and subduction rates for several island arcs. Of these, the Mariana arc has the highest estimated total magma production rate of 20 km<sup>3</sup>/Myr and a crustal subduction rate of 600 km<sup>3</sup>/Myr per km arc length. Thus, if all the magma produced were derived from the subducted (basaltic) crust, this would amount to only 3% of the basaltic portion of the crust. Using the figures of Karig and Kay [39], we find that all the K in Mariana arc volcanism could be accounted for by 31% of the K in the subducted basaltic crust. If, as Karig and Kay suggest, only 60% of the total K is derived from the basaltic crust (the remainder being derived from sediment and the overlying mantle wedge), less than 20% of the K in the basaltic crust is extracted by subduction zone volcanism. This suggests that most of the subducted potassium is not extracted from the crust but remains in that crust on its way into the mantle.

At some point during subduction, the crust is converted to eclogite. This material is substantially denser than either undepleted peridotite or depleted lithospheric material. Because of this and because the crust is cooler than the lithosphere below it, we postulate that the crust will physically separate from the lithosphere and sink into the deep mantle. This separation need not be, and is unlikely to be, complete. Thus some depleted peridotite of the lithosphere remains with the crust and some "enriched eclogite" remains with the lithosphere. The lithosphere is returned to the MORB source reservoir. Anderson [34,35] proposed that this eclogite sinks no deeper than 670 km and that it is recycled to form new oceanic crust (rather than oceanic islands). In view of the uncertainties of the phase relations at very high pressures, it seems equally plausible that subducted crust accumulates elsewhere, for example at the base of the mantle. In fact, the method used by Ruff and Anderson [40] to calculate relative densities of materials at the base of the mantle indicates that basaltic material would be about 0.1

$\text{g cm}^{-3}$  denser than pyrolite at that pressure. Although a particular depth of accumulation need not be specified for the purpose of our model, we have, for simplicity assumed that it is at the base of the mantle (e.g. Fig. 1).

### 3.2. Accumulation and heating

The present-day average rate of production of new oceanic crust is about  $3 \text{ km}^2 \text{ yr}^{-1}$  [41]. If this crust were evenly deposited at the base of the mantle the accumulation of one crustal thickness would require  $5 \times 10^7$  years, and a thickness of 100 km (= 15–20 layers) would require  $8 \times 10^8$  years. This is necessarily a very rough estimate because the spreading rate may have decreased and/or oscillated with time. More important, accumulation is probably highly uneven in space. For example, most of the subduction of the present cycle of sea-floor spreading is strongly concentrated in the Pacific. Consequently, thicknesses exceeding 100 km will probably accumulate locally within about  $10^8$  years.

The oceanic crust contains approximately  $1/F$  times the incompatible element concentration of the respective source material. For a range of melt fractions between  $F=0.3$  and  $0.1$  the concentration increase is therefore 3 to 10 fold, and the accumulated crust will have significantly higher concentrations of K, U and Th than the surrounding mantle. The high concentrations of heat-producing elements may cause a temperature increase sufficient to make the layer unstable, even if there is no additional heating from below. To obtain rough estimates of the temperature increase in the source layer as a result of internal heating alone, the following calculations are made assuming no heat production in the overlying mantle and no heat transfer from the core.

The abundances in present-day MORB of the heat producing elements are of the order  $U=0.1$ ,  $\text{Th}=0.4$ , and  $\text{K}=1000$  ppm. However, estimates of the concentrations of these elements in stored, recycled oceanic crust are much more difficult to make because of the uncertainties of the effects of: (1) additions from seawater and sediment, (2) "stripping" of these elements by dehydration and/or partial melting in subduction zones, and

(3) the admixture of depleted mantle. In addition, we must recall that we are dealing with Proterozoic, not modern, ocean crust as the source of modern mantle plumes. For the present purpose, we simply assume the above estimates. Also we use values of heat production corresponding to 2 Gyr ago. A heat-generating layer of 100 km thickness, initially isothermal with its surrounding, will heat up depending on the boundary conditions, which may be approximately isothermal if the surrounding medium convects. On the other hand, if the surrounding medium is stagnant it will be heated by conduction. These two simplest cases are shown in Figs. 3 and 4. The temperature rise is dramatic if the layer and its environment do not convect (Fig. 3). For layer thicknesses between 100 and 200 km (between 50 and 100 km, if no heat is lost into the core), the central temperature increases by  $400\text{--}1000^\circ\text{C}$  within 1–2 Gyr. This is similar to the temperature increase in the lower boundary layer of the mantle, estimated to be about  $1000^\circ\text{C}$  by Elsasser et al. [36] on the basis of heat transfer from the core. It is likely that such a temperature rise will decrease the density of the layer and make it unstable. For  $\Delta T=1000^\circ\text{C}$ ,  $\rho=5 \text{ g cm}^{-3}$  and a coefficient of expansion  $\alpha=10^{-5} \text{ deg}^{-1}$ , the density changes by  $\Delta\rho=0.05 \text{ g/cm}^{-3}$ . This is of the

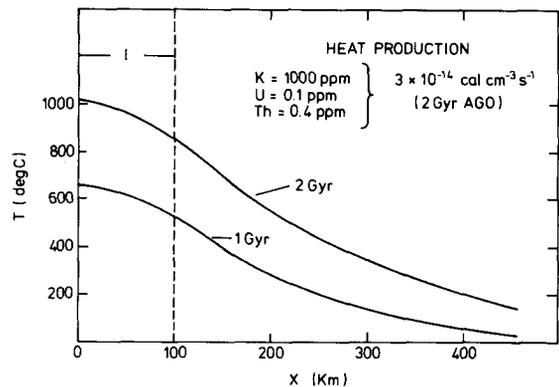


Fig. 3. Temperature profiles in an infinite medium containing a heat-producing layer of  $l=100$  km half-thickness with zero initial temperature and no convection. The profiles also apply to a layer of 100 km total thickness, if no heat is conducted through the base,  $X=0$ . The heat production is assumed constant at  $3 \times 10^{-14} \text{ cal cm}^{-3} \text{ s}^{-1}$ , and the heat diffusivity is assumed to be  $1 \times 10^{-2} \text{ cm}^2 \text{ s}^{-1}$ . The heat production is appropriate for typical oceanic crust at a time 2 Gyr ago.

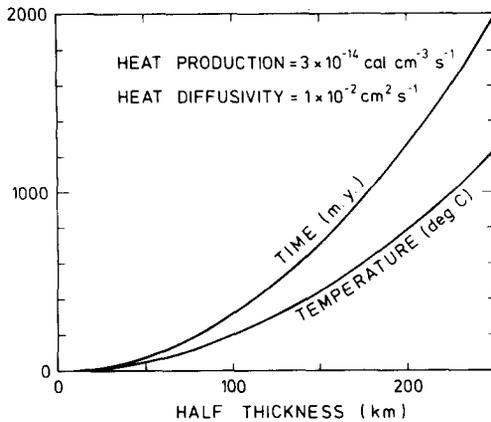


Fig. 4. Times (in Myr) required to reach 91% of the maximum (= steady state) temperature and the values of this temperature as a function of the half-thickness of a layer of constant heat production and zero initial and surface temperature. This also corresponds to the case where efficient convection outside the layer maintains a constant temperature at the surface of the layer. A layer of half-thickness 100 km (corresponding to the case shown in Fig. 3) reaches a temperature of 200°C in about 300 Myr at its center. Heat production =  $3 \times 10^{-14}$  cal cm $^{-3}$  s $^{-1}$ ; heat diffusivity =  $1 \times 10^{-2}$  cm $^2$  s $^{-1}$ .

same order of magnitude as the isothermal density difference of  $\Delta\rho = 0.1 \pm 0.1$  g cm $^{-3}$  between a basaltic composition and normal mantle, e.g. pyrolyte, calculated from shock wave data and using the method described by Ruff and Anderson [40]. The temperature may be even higher, if the layer is heated both internally and from below.

The other extreme is shown in fig. 4. Here, the enriched layer is assumed to be embedded in an isothermal medium. This corresponds to the case where the heat conducted to the boundaries of the layer is carried away efficiently by convection. The maximum temperature, reached at the center of the layer is much lower than in the previous case and reaches a steady-state value, which depends on the thickness of the layer. The figure shows the times required to reach approximately 90% of the maximum (= steady state) temperature and the values of this temperature as a function of half-thickness of the layer. To achieve a temperature increase of 1000°C, the layer must have a half-thickness of about 225 km and the time required is about 1.5 Gyr. The accumulation of this much

subducted oceanic crust seems unlikely. For this reason, and others to be discussed later, the case of a temporarily stagnant (or quasi-stagnant) lower mantle is preferred.

These rough calculations show that large vertical (and possibly horizontal) temperature differences may result from the proposed accumulation of oceanic crust at the base of the mantle. Elsasser et al. [36] discussed other arguments including seismic evidence favoring a hot boundary layer at the base of the mantle. Our purpose is only to point out that internal heating can also be significant if the boundary layer is chemically distinct and has a high heat production as proposed in our model.

### 3.3. Diapir (plume) formation and ascent

Once the temperature rise has reduced density sufficiently for the stored oceanic crust to become buoyant and to overcome viscous forces, Rayleigh-Taylor instabilities develop and the material begins to rise as diapirs. Whitehead and Luther [42] and Marsh [43] conducted laboratory experiments under conditions that may be relevant to our proposed model. A thin, low-density boundary layer is overlain by a higher-density fluid of differing viscosity. Narrow, ascending features develop in all cases, with the specific shape depending on the sign and the magnitude of the viscosity contrast. These experiments show that plume convection is possible, given the appropriate material properties and boundary conditions.

The spacing of plumes is expected to be irregular because of the irregular thickness of the accumulated source layer. The actual spacing of Morgan's plumes varies from several hundred to several thousand kilometers. The theory of Rayleigh-Taylor instabilities [42,43] indicates a distance of:

$$L = \frac{2\pi h}{2.15} (\mu_1/\mu_2)^{1/3}$$

between plumes where  $h$  is the thickness of the initially uniform low-viscosity, low-density layer, and  $\mu_1$  and  $\mu_2$  are the viscosities of the upper and lower layers, respectively.

In good agreement with this theory, the experiments (42,43) yielded spacings of  $L/h = 5$  and 10 for viscosity ratios of 8 and 44, respectively. This would indicate a spacing of 500 and 1000 km in the mantle, if the layer were uniformly 100 km thick. This result is useful in that it confirms the orders of magnitude of spacing and thickness.

Do plumes actually “drive” sea-floor spreading in the manner suggested by Morgan [1,24,25], namely by breaking up plates and causing passive upwelling of upper mantle material? We do not know of any quantitative assessment of this mechanism. We adopt this mechanism for the reasons given by Morgan, namely the high elevation of plume loci, the high degree of symmetry of the spreading plates, and the large-scale geochemical gradients observed in the vicinity of plumes [4,44] which suggest radial horizontal flow of plume material and mixing with asthenospheric material. An additional argument in favour of Morgan’s mechanism is the fact that MORB source material is so depleted in heat-producing elements that it must be either passively heated from deeper heat sources [45], or it must be mechanically driven to well up and fill in the accreting plate margins.

### 3.4. *Melting and interaction with the upper mantle*

Because the plume is initially hotter than the surrounding mantle, we expect that it would begin to melt at greater depths than, say, ascending upper mantle material beneath “normal” mid-ocean ridge segments. Depending on the excess temperature and on the amount of volatiles present within the diapir, melting may start at depths considerably greater than 100 km, i.e., below the base of the oceanic lithosphere [46]. The melt fraction may then react with the asthenospheric “wall rock” (which may also be partially molten). This interaction will have two important effects on the plume, namely cooling and chemical buffering. A detailed treatment of the magnitude of these effects is beyond the scope of this paper. We believe the cooling effect will prevent the plume from melting completely. Chemical buffering will be most important for the major elements and the compatible trace elements. If the composition of the plume melt is initially controlled primarily by

clinopyroxene and garnet (i.e., the eclogite assemblage), the buffering capacity of upper mantle olivine and orthopyroxene will change the melt composition toward that of a basalt in equilibrium with all four solid phases. These effects will be discussed in the section on geochemistry.

If the plume is long lived (i.e., on the order of 100 Myr), the buffering and cooling capacity of the adjacent upper mantle might become exhausted. On the other hand, viscous drag may cause fresh upper mantle material to advect toward the plume, so that a steady state may be the result. Until the physics of melting during diapiric ascent has been investigated in more detail, these questions cannot be answered definitively.

## 4. **Geochemistry of oceanic island basalts**

### 4.1. *Trace element concentrations*

The high concentrations of magmaphile trace elements in OIB have puzzled geochemists for years. Gast [47] explained this by relatively low degrees of melting at the OIB sources relative to MORB sources. If this were the sole explanation, then the on-ridge islands like Iceland would need a much larger flux of source mantle to be advected to the melting site than the surrounding normal mid-ocean ridge. A very rough estimate of the excess flux can be obtained from the enrichment factor of the magmaphile elements, let us say 3, and the excess production of basaltic crust (per unit time) on the island, also roughly a factor of 3. This estimate of the excess production of basalt on ocean islands is derived from a comparison of crustal thickness on Iceland of about 20 km [48] and normal oceanic crust of 5–7 km. This means an excess influx of source material below Iceland that is greater by a factor of 9 than the ordinary advective mantle flux to the adjacent regions of the mid-ocean ridge. In this excess material is brought in from greater depths in the mantle, and if its rise is roughly adiabatic, we would expect it to be melted to at least the same degree as the normal MORB source.

Another way of stating the same dilemma is that it seems improbable that those source regions

with high Rb/Sr, Nd/Sm, U/Pb, and Th/Pb ratios (namely the OIB source regions) should be melted consistently to a lesser extent than those regions that have low Rb/Sr, etc., ratios (namely the MORB source regions). It might be argued that the degree of melting could be lowered by the buildup of relatively thick oceanic crust beneath the islands. In that case, the degree of melting should be correlated with some parameter related to the thickness of the crust (or lithosphere). At present, we are not aware of such a correlation on a global scale.

Similarly, quantitative attempts to model source compositions *without* assuming the degree of melting also yield enriched source regions (relative to a primitive mantle) for island basalts (e.g. [49]). The apparent enrichment of OIB sources in magmaphile trace element concentrations must be reconciled with evidence for *relative* depletion of Rb (with respect to Sr) and Nd (with respect to Sm), which is indicated by the isotopic compositions of Sr and Nd shown in Fig. 7. One currently popular explanation for this paradox is that the OIB may have undergone metasomatic trace element enrichment prior to eruption. Unfortunately, the only direct evidence for mantle metasomatism is found in mantle xenoliths, and these xenoliths are normally *not* related to the source of the basalt in which they are incorporated (e.g. [50]). Consequently, we regard metasomatism as one possible, though awkward, escape from this geochemical dilemma.

Subducted oceanic crust is our preferred choice for the source material of OIB. Its composition meets the above-listed geochemical requirements. Its concentrations of magmaphile trace elements are elevated over those of primitive mantle; for example, rare earth concentrations are typically about ten times the chondritic values, yet its Rb/Sr and Nd/Sm ratios are variable and generally less than those of primitive mantle material. Such a source may produce OIB with "normal" degrees of melting, i.e., 10–30% for tholeiites. This is illustrated in Fig. 5 for the REE. This example shows that the REE patterns of a typical ocean island tholeiite (Kohala volcano, Hawaii) can be produced by about 30% melting of a source material with a typical MORB-type pattern. The

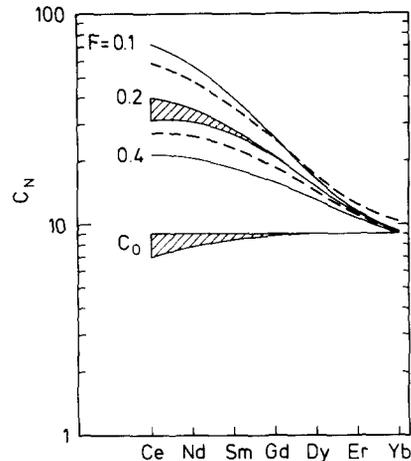


Fig. 5. Calculated rare earth concentrations, normalized to chondrites, using a simple batch-melting model from a garnet pyroxenite source with an initial REE pattern typical of ocean floor basalts. The two assumed source patterns are (1) flat, nine-times chondritic values, and (2) LREE-depleted to a value of seven times chondritic for cerium. The bulk partition coefficients from Ce to Yb are 0.03, 0.06, 0.14, 0.28, 0.49, 0.79 and 1.00. These coefficients correspond to a source containing about 40% opx, 40% cpx and 20% gnt, depending on the specific choice of mineral-melt coefficients. The heavy dashed lines are two REE patterns from Kohala volcano, Hawaii [72], for a typical tholeiite and a transitional (toward alkalic composition) basalt. For the flat source pattern, these Kohala basalts would be formed by 30% and 13% modal melting of the garnet pyroxenite.

source must contain both garnet and clinopyroxene for the melting models to be successful [51,52].

Concentrations of compatible trace elements such as Ni and Cr might be used to constrain the choice of OIB sources, if the partitioning of these elements were sufficiently well understood. Because of the high partition coefficients of Ni in olivine/melt and Cr in clinopyroxene/melt and spinel/melt, the melt is expected to have much *lower* Ni and Cr concentrations than the source. If this argument is applied to our model, Cr and Ni abundances should be much lower in OIB than in MORB (G. Brey, personal communication, 1981). Taken at face value, this is a serious weakness of our hypothesis. Nevertheless, we do not believe the weakness is fatal, because these compatible elements are easily buffered by olivine and pyroxene external to the source material *sensu stricto*. This may occur in at least two ways:

(1) We relax the conditions of the source accumulation process to the point of allowing some incorporation, say 10–20% of mantle peridotite together with the oceanic crust. This would cause only modest changes in density, incompatible element concentration, and heat production, but it would add enough olivine and pyroxene with “normal” (= mantle) concentrations of Ni and Cr, so that these elements would be buffered at “normal” melt concentrations.

(2) Similar buffering may also occur by reaction between the ascending plume and upper-mantle peridotite as postulated in section 3.4.

The chemical effects of either kind of buffering reaction are illustrated in Fig. 6. Three initial melts with Ni concentrations of 50, 100 and 200 ppm are equilibrated with a peridotite (referred to as “wall rock”,  $W$ ) of 2000 ppm Ni with a bulk partition coefficient of  $D_{Ni} = 3$ . It can be seen that for wall rock to melt ratios above about 0.5, the Ni concentration of the melt is greater than 400 ppm regardless of the initial Ni concentration in the melt. In contrast, the concentration of Sr, a moderately incompatible element, remains virtually un-

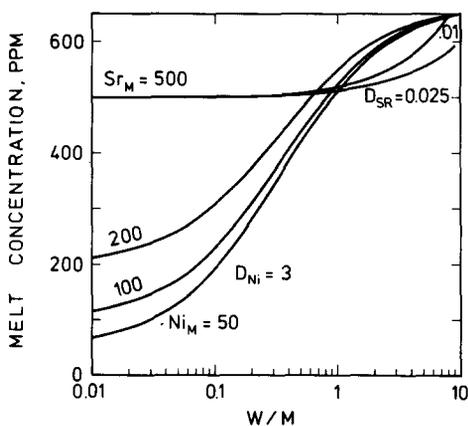


Fig. 6. Contamination of basaltic melt due to equilibration with a peridotitic “wall rock” for an incompatible element (Sr) and a compatible element (Ni). The mass ratio of wall rock to melt is labeled  $W/M$ ; the initial concentrations for Sr are assumed to be  $Sr_M = 500$  ppm in the melt and  $Sr_W = 25$  ppm in the wall rock; for Ni they are  $Ni_M = 50, 100,$  and  $200$  ppm and  $Ni_W = 2000$  ppm. The bulk partition coefficients are  $D_{Sr} = 0.025$  and  $0.01$ , and  $D_{Ni} = 3$ . Ni concentrations in the melt increase significantly while Sr concentrations remain virtually unchanged.

changed until the wall rock to melt ratio increases above a value of about 5. Consequently, equilibration of the eclogite-derived plume melt with a modest amount of peridotite, i.e.,  $0.5 < W/M < 5$ , causes the compatible element concentrations to be controlled by the peridotite, the incompatible element concentrations by the eclogite.

#### 4.2. Major element composition

The idea that OIB could be melting products of a source that has essentially a basaltic composition itself is not dear to petrologists. The prevailing opinion is that basalts are produced by partial fusion of peridotite (e.g. [53, chapter 15]). However, Yoder [54, chapters 2 and 8] points out that partial fusion of a source rock of basaltic composition and eclogitic phase assemblage could indeed yield a basaltic melt. Green and Ringwood [55] conducted melting experiments on eclogites and found the resulting melts to be saturated with  $SiO_2$ . In contrast, most OIB are olivine saturated. Nevertheless, we do not regard this to be a strong constraint because our suggested source is pure eclogite only in an idealized model. A more likely “real” source material would be eclogite intermixed or “contaminated” with some peridotite. In this case, the phase relationships during melting are probably indistinguishable from those of an ordinary “fertile” (i.e., garnet- and clinopyroxene-bearing) peridotite.

In general, it can be said that the major element concentrations (as well as the “compatible” trace element concentrations) will tend to be buffered by the presence of the four major mantle minerals olivine, orthopyroxene, clinopyroxene and garnet. The proportions of these phases in the melting region have a comparatively minor effect on the major element composition of the melt, as long as none of the minerals is consumed completely. Whether this buffering occurs in a well-mixed source assemblage or during some type of wall rock reaction between the melt and upper mantle peridotite is of little consequence in this regard. The behaviour of the major elements is analogous to that of the compatible trace elements discussed above.

In particular, the Fe/Mg ratio would be much

higher in a melt derived from pure oceanic crust than in an ordinary OIB, but buffering by upper mantle olivine or by peridotitic “impurities” in the source plume should bring the Fe/Mg ratio back to values normal for basalt.

In summary, it can be said that trace and major element abundances are generally consistent with the model but do not provide strong constraints. Nevertheless, the most striking chemical difference between MORB tholeiites and OIB tholeiites, namely the high concentrations of incompatible elements in OIB, is more easily explained by partial melting of eclogitic oceanic crust than by melting of ordinary mantle peridotite because of the small degrees of melting required by the peridotite model.

#### 4.3. Radiogenic isotope chemistry

Recent studies of Nd isotopes of mantle-derived basalts indicate that the Nd/Sm ratio in large portions of the mantle decreased with decreasing age (for a recent compilation of data, see Hart and Brooks [56]). Another way of saying this is, the LREE in the mantle became progressively depleted with time. If this is correct, then Proterozoic MORB had a less “depleted” (i.e., flatter) REE pattern than modern MORB. Also, one would expect that some Proterozoic MORB (just as some modern MORB) actually have REE patterns with slight relative enrichments of the LREE. Consequently, if Proterozoic MORB material is the source of modern OIB, one would expect variable but generally lower  $^{143}\text{Nd}/^{144}\text{Nd}$  ratios (and correspondingly higher  $^{87}\text{Sr}/^{86}\text{Sr}$  ratios) in OIB than in modern MORB. Also, on the Sr-Nd isotope correlation diagram most, but not all, of the OIB would still lie on the “depleted” side of the bulk-earth point (Fig. 7a). This is exactly what is observed in modern OIB. Consequently, the linear correlation of Sr and Nd isotopic data from oceanic islands may be interpreted as the result of magmatic Sm-Nd and Rb-Sr fractionation, which produced the ancient oceanic crust from the oceanic mantle at that time. Allègre et al. [17] have shown that the slope of the correlation line is compatible with extraction of liquid from the mantle and remixing at various times. By includ-

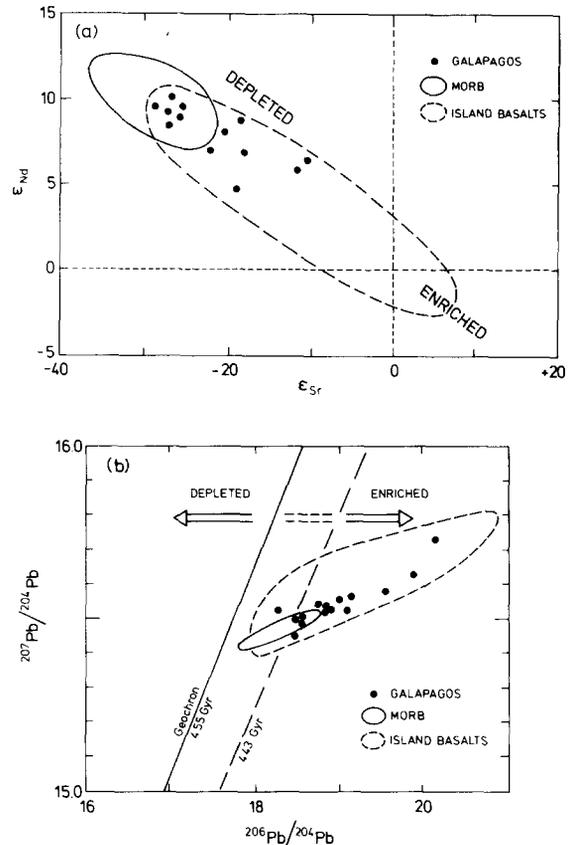


Fig. 7. Outlines of isotopic composition of oceanic basalts. The solid lines represent the fields of mid-ocean ridge basalts (MORB), the dashed lines show the oceanic island basalts. The solid dots are shown as examples from the Galapagos Islands [73,74]. The fields marked “depleted” and “enriched” indicate source regions with low and high ratios, respectively, of Rb/Sr, Nd/Sm, and U/Pb in relation to bulk-earth values (inferred from meteorite data). The inferred bulk-earth values are model dependent especially in the U-Pb system where the value depends strongly on the assumed age of the Earth. Nevertheless, even if the Earth is as young as 4.43 Gyr, most of the oceanic Pb data fall in the region of U/Pb enrichment, whereas most of the Nd and Sr data indicate Rb/Sr and Nd/Sm depletion of the source.

ing the oceanic crust as part of the array, the mantle array may extend into the “enriched” quadrant of the diagram, a condition actually observed in some OIB. This would not be possible if all OIB were merely mixtures of primitive and depleted sources.

Lead isotope systematics are not as easy to

explain. OIB have more variable and generally higher  $^{206}\text{Pb}/^{204}\text{Pb}$ ,  $^{207}\text{Pb}/^{204}\text{Pb}$  and  $^{208}\text{Pb}/^{204}\text{Pb}$  ratios than MORB (Fig. 7b). This relationship can be explained in a similar way as the Nd-Sr systematics: Proterozoic MORB had variably fractionated but generally higher U/Pb and Th/Pb ratios than their source mantle, and this will produce the variable but correlated enrichments in the radiogenic Pb isotopes  $^{206}\text{Pb}$ ,  $^{207}\text{Pb}$  and  $^{208}\text{Pb}$  of modern OIB. The fly in the ointment is that virtually *all* modern oceanic basalts show radiogenic lead enrichment relative to a single-stage U-Th-Pb isochron. In other words, all the basalt sources, even modern MORB sources, experienced an *increase* in their U/Pb ratios at some time significantly *later* than the formation of the earth. This observation has been explained by Vidal and Dosso [57] and by Dupré and Allègre [22] as the result of late Pb uptake by the earth's core, and by O'Nions et al. [16] as the result of Pb storage in the lower crust. Both mechanisms are compatible with our model and may explain the enrichment in radiogenic lead. However, the Pb isotopes also point rather clearly to an increase in the U/Th ratio in the mantle [58] and this cannot be explained by Pb storage alone. The U/Th ratio in the mantle might be increased by magmatic fractionation or by preferential recycling into mantle of uranium from the upper continental crust. This might occur through subduction of U-bearing sediments or through absorption of uranium into the basaltic oceanic crust by hydrothermal circulation at ridge crests or low-temperature alteration on the sea floor. This possibility is discussed in more detail in the following section.

#### 4.4. Alteration of the oceanic crust

The composition of the oceanic crust may be changed by three kinds of processes: high-temperature (hydrothermal) alteration, low-temperature alteration, and addition of sediments. A thorough evaluation of these processes is beyond the scope of this paper but will ultimately be needed to test our model because the subducted, altered crust is the postulated source of OIB. In this section, we discuss briefly some of the possible isotopic effects of sea water alteration.

The anomalously low  $^{18}\text{O}$  values found in some Icelandic basalts [59] are an indication that the source of these basalts was affected by hydrothermal alteration. Unfortunately, we do not at present know if this alteration process took place recently within the Icelandic crust, or whether it affected the Iceland source material during its creation as ancient oceanic crust.

Deviations from the Nd-Sr "mantle array" may be caused by the addition of radiogenic strontium from seawater or from the continental crust. Such deviations have been reported from island arc volcanics [60,61] and they are easily explained in this context. However, similar deviations have also been reported for ocean islands that are not related to subduction, namely the Azores [62], and Samoa [63]. The fact that these OIB deviate from the mantle array suggests that their sources contain a crustal component enriched in  $^{87}\text{Sr}$ . Whether this radiogenic strontium was located primarily in the altered basalt or in the sedimentary cover is not immediately obvious. Most likely, both components are important, since the basalt crust is the vehicle for the sediments on their way back into mantle.

Uranium from the upper continental crust may be transported back into the mantle by dissolution in seawater and reprecipitation in fresh oceanic crust by low-temperature alteration and hydrothermal interaction. The importance of this mechanism is suggested by inspection of Table 1, which gives estimates of the maximum possible concentration increases of the oceanic crust due to uptake of U, Th, Rb, Sr, and Nd dissolved in seawater. These estimates are calculated from the total river inflow, the production rate of oceanic crust ( $\approx 20 \text{ km}^3/\text{yr}$ ) and the assumptions that this crust is the only sink of these elements and that steady state prevails. The actual uptake by the oceanic crust will be much smaller for most of these elements because there are more efficient sinks (such as shallow-water sediments), but no such sink appears to be available for uranium. Uranium may be removed from seawater by low-temperature alteration of sea-floor basalt [64,65] and by hydrothermal interaction with the oceanic crust at mid-ocean ridges [66]. The amount of uranium recycling estimated in this way is larger

TABLE 1  
Dissolved element flux through ocean water \*

	U	Th	Pb	Rb	Sr	Nd
River influx (in $10^{10}$ g yr <sup>-1</sup> )	(0.13) 1.9	0.03		3.2	59	
Pelagic precipitation ( $10^{10}$ g yr <sup>-1</sup> )			1.1			0.2-1.4
<i>Assuming the entire flux is added to the oceanic crust:</i>						
Net concentration increase of crust (in ppm)	(0.02) 0.33	0.005	0.2	0.5	10	<0.24
Percent concentration increase of crust (in % over fresh MORB)	(20) 300	1	20	50	10	<2

\* The upper two rows represent input estimates either from measurements of element concentrations in river waters or from estimates of prehistoric pelagic precipitation rates at the ocean bottom. The lower two columns are estimates of the bulk concentration increase of the oceanic crust if the entire influx is added to the oceanic crust of basaltic composition, produced at a rate of 18 km<sup>3</sup> per year. These are maximum estimates for most of the elements, especially Pb which is rapidly removed from solution by shallow-water organisms, and Sr which is coprecipitated with shallow-water carbonates. On the other hand, uranium is added to the oceanic crust by hydrothermal activity [66]. Note, however, the U figures in parentheses, which are based on the U concentration of Amazon river water only (see compilation by Bloch [65]). Other influx data are from Edmond et al. [67], Goldberg et al. [68], Goldberg [69], Schaule and Patterson [70] and Turekian and Chan [71].

than the uranium increase inferred from the Pb isotopes. The reason for this may be because the river influx estimates are biased due to runoff from phosphate fertilizers. If U influx from the Amazon river (given in parentheses on Table 1) were representative of uncontaminated, world-wide riverine U influx, the increase in U/Pb and U/Th in the subducted crust would be much more modest, though still significant.

#### 4.5. Mantle isochrons

If the source of OIB is indeed an ancient oceanic crust that was formed by magmatic processes, and if these processes fractionated U/Pb, Rb/Sr, and Sm/Nd ratios, it may be that the roughly linear correlations on the various isochron diagrams can be interpreted as "mantle isochrons" as proposed by Brooks et al. [38].

Model ages calculated from <sup>207</sup>Pb/<sup>204</sup>Pb vs. <sup>206</sup>Pb/<sup>204</sup>Pb ratios of several island groups and MORB have recently been reviewed by Chase [37]. The dominant model age of about 1.8 Gyr is striking as is its approximate concordance with the Rb-Sr isochron age for oceanic tholeiites reported by Brooks et al. [38]. It is therefore possible that

this is the age of the subduction event that created the source material for today's OIB. It is also possible, however, that the linear correlations represent mixing lines between this same source material and an upper mantle that underwent subsequent changes in chemistry. In either case, apparent or real mantle isochrons would be a natural consequence of our model.

Allègre et al. [23] have shown how a continuously differentiating mantle can produce mantle isochrons in Rb-Sr and U-Pb decay systems. They found that by letting the differentiation and mixing parameters vary with time, the calculated apparent Pb-Pb isochrons can match the observed ones. However, these authors also point out that the problem is poorly constrained and that it would be desirable to formulate the problem as a stochastic rather than a deterministic one. Thus, while mantle isochrons are a *possible* consequence of continuous mantle processes, they have so far not been shown to be an *inevitable* consequence. In view of this we reiterate that the simplest explanation of an apparent isochron may be that it is a real one.

Although our model is not inherently and necessarily dependent on an episodic mantle evolu-

tion, it is certainly compatible with it. Because of this, we have pointed out some of the possible consequences of episodic mantle convection (episodic continental igneous activity, apparent mantle isochrons) as well as its possible cause (delayed instability of plume sources because of slow heating of newly subducted oceanic crust). We note, however, that continuous or semicontinuous versions of the model are also possible and may be more plausible to those who are strongly inclined toward uniformitarianism.

### 5. Concluding remarks

The ideas expressed in this paper are largely qualitative at present. The model we have proposed is an extreme one, so that it will present a broad target for criticism. Other, more "modest" versions are easy to think up, if some of the details do not fit future observations. For example, the "mantle isochrons" may not give the correct age for the recycling process after all, or the mechanism for storage and heating may need modification. The fundamental aspect of the model, which can be tested more immediately, rests on the inference that the source of ocean-island-magmatism is not primitive (or slightly depleted) mantle, but contains significant amounts of recycled oceanic crust. The first stage of the recycling process is the subduction of the oceanic crust, which consists of ocean floor basalt (and its gabbroic equivalent), its alteration products, and a sediment cover of varying thickness.

It should be possible in the near future to test the model through better estimates of the actual bulk composition of the subducted crust, a more complete characterization of OIB trace element and isotope chemistry and its relationship to primitive and evolved sources, improved geochemical modeling of OIB petrogenesis, and a geophysical identification of the source depth and ascent paths of mantle plumes.

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Yearbook [6]. Since that time Chase [37] has independently published a paper containing very similar ideas, based on the Pb isotopic composition of oceanic basalts alone. Among the many people who have helped us with constructive criticism of our ideas are C.J. Allègre, D.L. Anderson, G. Brey, R.C. Fletcher, S.R. Hart, D.E. James, E. Jagoutz, W.J. Morgan, I. Nicholls, R. Stern, and G.W. Wetherill.

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