

Origin of ocean island basalts: A new perspective from petrology, geochemistry, and mineral physics considerations

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[1] Consideration of petrology, geochemistry, and mineral physics suggests that ancient subducted oceanic crusts cannot be the source materials supplying ocean island basalts (OIB). Melting of oceanic crusts cannot produce high-magnesian OIB lavas. Ancient oceanic crusts (>1 Ga) are isotopically too depleted to meet the required values of most OIB. Subducted oceanic crusts that have passed through subduction zone dehydration are likely to be depleted in water-soluble incompatible elements (e.g., Ba, Rb, Cs, U, K, Sr, Pb) relative to water-insoluble incompatible elements (e.g., Nb, Ta, Zr, Hf, Ti). Melting of residual crusts with such trace element composition cannot produce OIB. Oceanic crusts, if subducted into the lower mantle, will be >2% denser than the ambient mantle at shallow lower mantle depths. This negative buoyancy will impede return of the subducted oceanic crusts into the upper mantle. If subducted oceanic crusts melt at the base of the mantle, the resultant melts are even denser than the ambient peridotitic mantle, perhaps by as much as ~15%. Neither in the solid state nor in melt form can bulk oceanic crusts subducted into the lower mantle return to upper mantle source regions of oceanic basalts. Deep portions of recycled oceanic lithosphere are important geochemical reservoirs hosting volatiles and incompatible elements as a result of metasomatism taking place at the interface between the low-velocity zone and the cooling and thickening oceanic lithosphere. These metasomatized and recycled deep portions of oceanic lithosphere are the most likely candidates for OIB sources in terms of petrology, geochemistry and mineral physics. **INDEX TERMS:** 1025 Geochemistry: Composition of the mantle; 3040 Marine Geology and Geophysics: Plate tectonics (8150, 8155, 8157, 8158); 8121 Tectonophysics: Dynamics, convection currents and mantle plumes; 8124 Tectonophysics: Earth's interior—composition and state (1212); 8125 Tectonophysics: Evolution of the Earth; **KEYWORDS:** OIB sources, mantle plumes, recycled ocean crust, oceanic lithosphere, low-velocity zone metasomatism, mantle convection

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1. Introduction

[2] One of the fundamental advances in modern geochemistry of the solid Earth is the recognition of compositional heterogeneities in the Earth's mantle through studies of oceanic basalts. Ocean island basalts (OIB) are particularly variable in composition such that several isotopically distinct mantle source end-members (e.g., "DM", "EM1", "EM2", "HUMU", "FOZO", "C") are required to explain the variability [e.g., White, 1985; Zindler and Hart, 1986; Hart *et al.*, 1992; Hanan and Graham, 1996]. The isotopic ratio differences among these end-members reflect the differences of the radioactive parent/radiogenic daughter (P/D) ratios (e.g., Rb/Sr, Sm/Nd, U/Pb, and Th/Pb) in their ultimate mantle sources which, with time and the implicit assumption of a two-stage development, evolve to distinct

tive fields in isotope ratio spaces. Significant fractionation of parent/daughter (P/D) ratios in the solid state is unlikely in the deep mantle due to extremely slow diffusion [e.g., Hofmann and Hart, 1978], hence it is logical to suspect that processes known to occur in the upper mantle and crust (e.g., partial melting and magma evolution, dehydration, alteration/metamorphism, differential weathering, transport and sedimentation) are the likely causes of any P/D fractionation, recognizing that partial melting might also be a factor in the transition zone and lower mantle [e.g., Lithgow-Bertelloni and Stixrude, 2000; Hirose *et al.*, 1999; Hirose and Fei, 2002]. These shallow or near-surface P/D-fractionated materials are then introduced into mantle sources of oceanic basalts through subduction zones. Mantle compositional heterogeneity is thus a general consequence of plate tectonics because of crust-mantle recycling.

[3] Among many contributions endeavoring to understand the origin of mantle compositional heterogeneity in the context of plate tectonics is the classic paper by Hofmann and White [1982, p. 421], who proposed that "oceanic crust is returned to the mantle during subduction . . . Eventually, it

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becomes unstable as a consequence of internal heating, and the resulting diapirs become the source plumes of oceanic island basalts (OIB) and hot spot volcanism." They also stressed that this recycled ancient oceanic crust, the plume material, which is the source of OIB is geochemically more "enriched" in K, Rb, U, Th, and light rare earth elements relative to the more "depleted" source of mid-ocean ridge basalts (MORB). While some details are considered conjectural, the principal idea of the model has been widely accepted by the solid Earth community as being, to a first order, correct [e.g., Hofmann, 1988, 1997; Chauvel *et al.*, 1992; Hauri, 1996; Hauri *et al.*, 1996; Rehkämper and Hofmann, 1997; Niu and Batiza, 1997; Lassiter and Hauri, 1998; Sobolev *et al.*, 2000; Cordery *et al.*, 1997; Leitch and Davies, 2001]. In this paper, instead of reviewing numerous works following this line in the literature, we argue that there is no obvious association between ancient subducted oceanic crusts and source materials of OIB. Our arguments are based on well-understood petrological processes, geochemical observations, and recent experimental results on mineral physics under deep mantle conditions [e.g., Agee, 1998; Ohtani and Maeda, 2001].

[4] The arguments in this paper will need to be considered in future models of chemical geodynamics and mantle convection. The concept of mantle metasomatism that has existed in the literature for decades remains valid in explaining many aspects of enriched geochemical signatures in OIB and some enriched MORB (E-MORB) [e.g., Green, 1971; Sun and Hanson, 1975; Frey and Green, 1974, 1978; Wood, 1979; Le Roex *et al.*, 1983; Anderson, 1989, 1994; Sun and McDonough, 1989; McKenzie, 1989; Halliday *et al.*, 1995; Niu *et al.*, 1996, 1999]. Following Niu *et al.* [2002a], we illustrate explicitly that the interface between the low-velocity zone (LVZ) and the cooling and thickening oceanic lithosphere is the most likely location where such metasomatism takes place, and we argue that the subducted metasomatized portions of deep oceanic lithosphere are the best candidates for the ultimate sources feeding hot spot volcanisms.

2. Petrological and Geochemical Arguments

2.1. Melting of Subducted Oceanic Crusts Unable to Produce High-Magnesian OIB Melts

[5] Christensen and Hofmann [1994] explored physical scenarios in which subducted oceanic crust might segregate from the attached lithospheric mantle at the bottom of the lower mantle during mantle convection so as to form deep-rooted plumes to rise and feed hot spot volcanism in the upper mantle. Yasuda and Fujii [1998] made a similar effort for shallower depths. Average bulk oceanic crust is picritic/basaltic in composition and is likely to have no more than ~13 wt % MgO [e.g., Niu, 1997]. Such crust cannot, by melting, produce the high-magnesian picritic lavas seen in many OIB suites. For example, high-magnesian glasses of 15 wt % MgO [Clague *et al.*, 1991] or more [Norman and Garcia, 1999] have been observed in Hawaiian shield picrites. Basaltic melts are derived from more magnesian picritic melts produced by partial melting of mantle peridotites [e.g., O'Hara, 1968a, 1968b; Stolper, 1980; Falloon *et al.*, 1988; Herzberg and O'Hara, 1998, 2002; O'Hara and Herzberg, 2002]. Partial melting of recycled oceanic

crust, which is basaltic/picritic in composition [e.g., Niu, 1997] and quartz-eclogite or eclogite in petrology [e.g., O'Hara and Yoder, 1967; O'Hara and Herzberg, 2002], will not produce basaltic/picritic melts but will produce melts of more silicic composition [e.g., Green and Ringwood, 1968; Wyllie, 1970]. If total melting had occurred (physically unlikely), the melts would be basaltic/picritic in composition, but would still differ in both major and trace element systematics from those of average OIB. It is possible that ancient oceanic crust is more magnesian than the present-day equivalent because of hotter potential mantle temperatures and thus greater extents of melting in the Archean and Proterozoic, but it is, in principle, not possible to partially melt picrites to produce picrites. In fact, primary OIB melts must be more magnesian than most primitive picrite melts (>15% MgO) observed in Hawaii and other OIB suites [Herzberg and O'Hara, 1998, 2002]. Therefore petrologically ancient recycled oceanic crusts cannot become sources feeding hot spot volcanism and OIB. One may argue that OIB are derived from melts of recycled oceanic crusts mixed with predominantly peridotite melts [e.g., Chauvel *et al.*, 1992]. This is possible and will apparently relax the OIB MgO requirement, but then OIB are no longer derived from recycled oceanic crust alone.

2.2. Ancient Oceanic Crusts Isotopically Too Depleted to Yield OIB

[6] It is generally thought that MORB are derived from a mantle source that was depleted in incompatible elements early in Earth's history as a result of extraction of the incompatible-element-enriched continental crust [e.g., Armstrong, 1968; Gast, 1968; O'Nions *et al.*, 1979; Jacobsen and Wasserburg, 1979; DePaolo, 1980; Allègre *et al.*, 1983]. This early differentiation is postulated to have resulted in a relatively homogeneous mantle, from which the depleted mantle (DM) source for MORB and the more enriched source for OIB were then produced over a long period principally by the subduction of oceanic crust [e.g., Hofmann and White, 1982; Hofmann *et al.*, 1986; Hofmann, 1988, 1997]. A simple way to test the hypothesis whether ancient recycled oceanic crusts can serve as enriched sources for OIB is to see what the isotopic signatures of those ancient recycled oceanic crusts would be at present and ask whether they are consistent with those of present-day OIB. This test rejects the hypothesis as illustrated in Figure 1 and detailed below.

[7] To avoid ambiguous conclusions, we use simple Sr-Nd-Hf isotopic systems as opposed to complex Pb or poorly understood Os isotopic systems and the noble gasses. This strategy is logical because if the hypothesis fails for simple systems, interpretations in favor of that hypothesis based on complex systems must also fail. For simplicity, we also assume that ancient subducted oceanic crusts experienced no subduction zone dehydration (the effects of which we consider separately below). It is likely that continental crust has been accreted episodically from ~4.0 Ga to the present [e.g., McCulloch and Bennett, 1998], in which case the Earth's mantle would be depleted progressively during this same long period. The mean age of the continental crust is about 2.0–2.5 Ga [e.g., Jacobsen and Wasserburg, 1979; Taylor and McLennan, 1985]; hence we can assume a similar mean age for the depleted mantle. Using the

depleted MORB melts from the Garrett Transform in the Pacific [Niu and Hékinian, 1997; Wendt *et al.*, 1999] as representing the DM ($\epsilon_{\text{Sr}} = -32$, $\epsilon_{\text{Nd}} = 12$, and estimated $\epsilon_{\text{Hf}} = 24$), we can calculate the evolution curves of the DM which differentiated from the bulk silicate Earth (BSE) or primitive mantle (PM) at ~ 2.5 Ga as a result of continental crust extraction (the choice of this particular composition for the DM is not critical). In the plots of ϵ_{Sr} , ϵ_{Nd} , and ϵ_{Hf} versus time in Figure 1 these curves are labeled 1. Given the known composition of the present-day DM and the decay constants of the radioactive parents (i.e., ^{87}Rb , ^{147}Sm , and ^{176}Lu), the P/D ratios (i.e., $^{87}\text{Rb}/^{87}\text{Sr}$, $^{147}\text{Sm}/^{143}\text{Nd}$, and $^{176}\text{Lu}/^{176}\text{Hf}$) of the DM along these curves are then constrained at any given time in the last 2.5 Gyr.

[8] The significant coupling between radiogenic isotopes and incompatible element abundances and ratios in many OIB suites, seamount lavas, and MORB [e.g., Saunders *et al.*, 1988; Sun and McDonough, 1989; Niu and Batiza, 1997; Niu *et al.*, 1999, 2002a] suggests that the source materials of these basalts are ancient and have developed their isotopic characteristics over a period in excess of 1.0 Gyr. If recycled ancient oceanic crusts played a role in the source regions of these oceanic basalts, then these ancient oceanic crusts would have to be formed >1.0 Ga. Evolution curves of 2 and 3 in Figure 1 represent these ancient oceanic crusts formed at 2.0 Ga and 1.0 Ga, respectively, from the DM by 20% batch (equilibrium partial) melting using values for the effective bulk distribution coefficients, D , given in Table 1. Such calculations are rather simplistic and effective bulk D can vary substantially [e.g., O'Hara, 2000; O'Hara and Herzberg, 2002], but more complex models do not change the conclusion of the test; 20% melting is higher than a mean value of 15% melting beneath present-day ocean ridges [Niu, 1997] but is reasonable given the hotter mantle in the past. Again, different choices of extent of melting do not affect the conclusion of the test. Figure 1 shows that the isotopic compositions (rectangles on the left vertical axes) of these ancient ocean crusts are quite depleted after 2.0 Ga and 1.0 Ga evolution with $\epsilon_{\text{Sr}} < -24$, $\epsilon_{\text{Nd}} > 8.4$, and $\epsilon_{\text{Hf}} > 19$. They are, on average, much more depleted than most of the present-day OIB [Albarède, 1995; Salters and White, 1998] with $\epsilon_{\text{Sr}} > -20$, $\epsilon_{\text{Nd}} < 6$, and $\epsilon_{\text{Hf}} < 12$ given as histograms along vertical axes to the right. Curves 4 and 5 represent the evolution paths if ocean crust of present-day mean composition (Table 1) had been injected into the Earth's mantle at 2.0 Ga or 1.0 Ga, respectively. This is equivalent to assuming different DM ages and P/D ratios. These paths are obviously more depleted in ϵ_{Sr} because of low Rb/Sr ratio of the present-day mean oceanic crust with respect to old DM, similar in ϵ_{Nd} (similar Sm/Nd ratio) to that of DM, but astonishingly display a more enriched character in ϵ_{Hf} due to the low Lu/Hf ratio relative to DM. The "odd" Hf isotopes reflects the fact that depleted MORB (or bulk ocean crust) often has greater depletion in heavy rare earth elements (REE, e.g., Lu) than middle REE and elements like Zr and Hf (see later). Curves 4 and 5 illustrate the concept, but there are old ophiolitic basalts that are as depleted as present ocean crust [e.g., Bruce *et al.*, 2000]. Changing the mean age of the DM to >2.5 Ga, 3 Ga, or 3.5 Ga will not change the conclusion reached here.

[9] Large databases for MORB and OIB exist [e.g., Lehnert *et al.*, 2000], and in particular, high-quality data

sets on Hf isotopes that have been difficult to collect by thermal ionization mass spectrometer (TIMS) have also begun to appear (although not yet openly available) thanks to multichannel inductively coupled plasma-mass spectrometry (MC-ICP-MS) [e.g., Blichert-Toft and Albarède, 1997; Vervoort and Blichert-Toft, 1999; Vervoort *et al.*, 1999, 2000; Albarède and van der Hilst, 2002]. However, the two data sets we use in Figure 1 (and Figures 4 and 5 in section 2.4) are adequate for the purpose of the paper because these two data sets define the same trends and cover the same spread as the large database seen in the recent literature. For example, the island-averaged data for Sr-Nd isotopes in the compilation of Albarède [1995] define the mantle array and reveal clearly the first-order global systematics. The Nd-Hf isotopic data reported by Salters and White [1998] on many OIB also define the mantle or "terrestrial" array [Vervoort *et al.*, 1999] with global coverage both geographically and in Nd-Hf isotope ratio space.

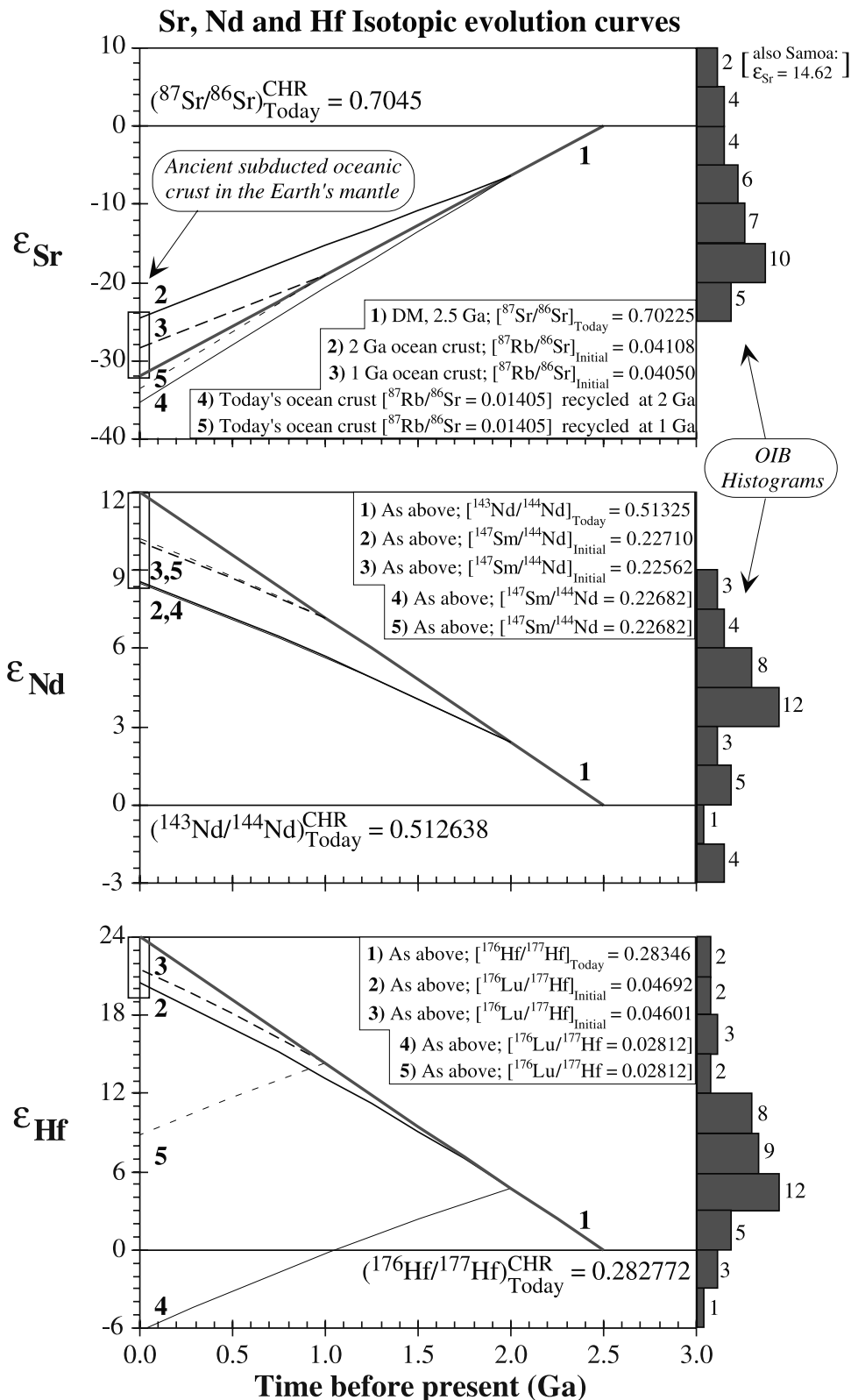
2.3. Recycled Oceanic Crusts After Subduction Zone Dehydration Unsuitable as Sources for OIB

[10] The ocean crust is altered by hydrothermal activity during its accretion at ocean ridges and pervasively weathered/hydrated subsequently on the seafloor. In a widely accepted model, this crust that is on top of the subducting slab endures the greatest extents of dehydration in subduction zones. Fluids released from this dehydration lower the solidus of the overlying mantle wedge, which then melts to produce arc lavas. The incompatible element geochemical signatures of arc lavas largely reflect the signatures of this subducting slab "component" [e.g., Gill, 1981; Tatsumi *et al.*, 1986; McCulloch and Gamble, 1991; Stolper and Newman, 1994; Hawkins, 1995; Pearce and Peate, 1995; Davidson, 1996; Tatsumi and Kogiso, 1997; Ewart *et al.*, 1998]. The slab "component" may include elements of altered ocean crust, and pelagic or terrigenous sediments [e.g., Plank and Langmuir, 1993, 1998; Elliott *et al.*, 1997; Class *et al.*, 2001], but the fluids and the fluid-soluble elements (e.g., Ba, Cs, Rb, U, K, Pb, Sr) released contribute the most to the petrogenesis of arc lavas (see Figures 2 and 3a). If this widely accepted interpretation of IAB genesis is correct, at least to a first order, then the residual subducted crust that has passed through subduction zone dehydration reactions will have geochemical signatures for the fluid-soluble elements that are complementary to the signatures of arc lavas [e.g., McDonough, 1991; Niu *et al.*, 1999, 2002a]. This residual crust should be relatively enriched in water-insoluble or immobile incompatible elements (e.g., Nb, Ta, Zr, Hf, and Ti) and highly depleted in water-soluble or mobile incompatible elements (e.g., Ba, Rb, Cs, Th, U, K, Sr, Pb) (Figure 3b). It follows that if the recycled oceanic crust were the major source of OIB, then these latter basalts should be highly depleted in these water-soluble incompatible elements. However, OIB are enriched not only in these water-soluble incompatible elements but also in water-insoluble incompatible elements relative to mean ocean crust (Figures 2 and 3a).

[11] Figures 2 and 3 illustrate the argument by comparing the average OIB [Sun and McDonough, 1989], average E-MORB [Niu *et al.*, 2002a], and average western Pacific Island arc tholeiites (IAB) with average ocean crust (the first three are normalized to bulk ocean crust; see Table 1 for

details). The OIB and E-MORB are progressively more enriched in the more incompatible elements than average ocean crust, whereas the IAB is depleted in REEs, and significantly more so in water-insoluble incompatible elements, but enriched in water-soluble incompatible elements. Figure 3b shows schematically the geochemical signatures

that average recycled residual oceanic crust should hold after passing through subduction zone dehydration reactions. Assuming heavy REEs are less affected by the subduction zone dehydration, then the contrast in both relative abundances (qualitatively) and patterns of incompatible elements between average OIB/E-MORB and the



recycled residual crust is extraordinary. To melt or partially melt recycled ocean crust with such residual geochemical signatures cannot produce OIB or E-MORB unless this residual crust were refertilized within the mantle by adding significant amounts of Ba, Rb, Cs, Th, U, K, Pb, and Sr and lesser amounts of Nb, Ta, LREEs, Zr, and Hf, i.e., the reverse of subduction zone dehydration. Such postulated refertilization is ad hoc.

[12] Recycled terrigenous sediments would be enriched in these water-soluble elements, but they dehydrate or even melt in subduction zones, contributing to arc volcanism [Plank and Langmuir, 1998; Elliot *et al.*, 1997]. If the subducted terrigenous sediments neither dehydrated nor melted, their presence in the source regions of oceanic basalts would still fail to explain the elevated Ce/Pb and Nb/U ratios in most oceanic basalts [Hofmann *et al.*, 1986; Niu *et al.*, 1999]. Furthermore, terrigenous sediments with detrital zircon crystals will lead to Zr-Hf fractionation from REE and Hf isotopes [Patchett *et al.*, 1984; White *et al.*, 1986; Vervoort *et al.*, 1999], an effect which is not observed in OIB (see below).

2.4. OIB Sr-Nd-Hf Isotopes Lacking Subduction Zone Dehydration Signatures

[13] If ancient subducted oceanic crusts had indeed played an important role in the petrogenesis of OIB, then isotopic signatures of nonmagmatic subduction zone dehydration processes must be preserved in OIB. This is, however, not observed. Figure 4 demonstrates the arguments using island-averaged data compiled by Albarède [1995] for Sr and Nd isotopes and a number of OIB suites with global coverage reported by Salters and White [1998] for Nd and Hf isotopes (also see Figure 1 and discussion above). Figure 4a plots averages of 40 OIB suites in $\epsilon_{\text{Sr}} - \epsilon_{\text{Nd}}$ space, each representing average composition of an ocean island (or islands). Except for three EM2 OIB suites (Samoa, Society, and Azores) and one HIMU OIB suite (St. Helena), all other 36 OIB suites define a scattered, yet

statistically significant (at >99.9% confidence levels) inverse linear trend. Figure 4b plots a number of OIB suites in $\epsilon_{\text{Hf}} - \epsilon_{\text{Nd}}$ space. Except for the two HIMU Islands (St. Helena and Tubaii), the data define a statistically significant (>99.9% confidence levels) positive linear trend. Given the relatively minor occurrences of EM2 and HIMU OIB suites on a global scale, we first focus our discussion on the majority of OIB suites here and discuss the implications of the EM2 and HIMU OIB suites later.

[14] Because of the large differences in relative mobility of $\text{Rb} > \text{Sr} > \text{Nd} > \text{Sm} > \text{Lu} > \text{Hf}$ during subduction dehydration inferred from observations (Table 1) and determined experimentally [Kogiso *et al.*, 1997], the significant correlations in Figures 4a and 4b would not exist or would have been destroyed if sources of these OIB had been involved in, or actually part of, ancient oceanic crusts passing through subduction zone dehydration reactions. The significant correlations thus suggest that (1) elements Sr, Nd, and Hf have behaved similarly in the respective sources of these OIB in the past >1.0 Gyr; (2) the similar behavior would be unlikely if these OIB sources had experienced subduction zone dehydration but is entirely expected if the process or processes these OIB sources had experienced are magmatic because of the similar effective bulk D of these elements (Table 1); and (3) coupled correlations of P/D ratios such as Rb/Sr , Sm/Nd , and Lu/Hf in these OIB sources must also have persisted undisturbed in the last >1.0 Gyr. Figure 5 shows the same plots as in Figure 4 on expanded scales to include a number of additional items. The numbered diamonds are the model results of Figure 1. Ancient oceanic crusts produced 2.0 and 1.0 Gyr ago (labeled 2 and 3, respectively) from the DM plot at the most depleted end of the OIB arrays and thus cannot be sources of OIB as concluded above. The arrows point in the direction of the effect of subduction zone dehydration. The latter makes ancient recycled oceanic crusts even less likely as sources of OIB. The aging or radioactive decay of ocean crust with present-day composition (Figure 1) if injected in

Figure 1. (opposite) Sr, Nd, and Hf isotopic evolution of depleted mantle and ancient oceanic crusts using epsilon notations. The horizontal lines with ϵ_{Sr} , ϵ_{Nd} , and ϵ_{Hf} values of 0 are evolution paths of chondrite uniform reservoir (CHUR), which is often used as representing bulk silicate earth (BSE) or primitive mantle (PM), i.e., mantle plus crust as the Earth's core probably does not take Rb, Sr, Sm, Nd, Lu, and Hf in significant amount. The recommended CHUR values for Sr and Nd isotopes used for calculations are from Faure [1986] and Dickin [1997], who give detailed discussions and ultimate sources: $^{87}\text{Sr}/^{86}\text{Sr}_{\text{today}} = 0.70450$, $^{87}\text{Sr}/^{86}\text{Sr}_{4.55\text{Ga}} = 0.69897$, and $^{87}\text{Rb}/^{86}\text{Sr}_{\text{today}} = 0.0827$; $^{143}\text{Nd}/^{144}\text{Nd}_{\text{today}} = 0.512638$, $^{143}\text{Nd}/^{144}\text{Nd}_{4.55\text{Ga}} = 0.506697$, and $^{147}\text{Sm}/^{143}\text{Nd}_{\text{today}} = 0.1967$. The CHUR values for Hf isotopes are from Blichert-Toft and Albarède [1997]: $^{176}\text{Hf}/^{177}\text{Hf}_{\text{today}} = 0.282772$, $^{176}\text{Hf}/^{177}\text{Hf}_{4.55\text{Ga}} = 0.279742$, and $^{176}\text{Lu}/^{177}\text{Hf}_{\text{today}} = 0.0332$. The relevant decay constants are taken from Dickin [1997]. Epsilon values are calculated using the general formula (use ϵ_{Nd} as an example): $\epsilon_{\text{Nd}}(t) = (^{143}\text{Nd}/^{144}\text{Nd}_{(t,\text{sample})} / ^{143}\text{Nd}/^{144}\text{Nd}_{(t,\text{CHUR})} - 1) \times 10,000$, where t represents time that varies from zero today to the age of the Earth of 4.55 Ga. The depleted mantle (DM) uses the average values of highly depleted lavas from the Garrett Transform in the Pacific [Niu and Hékinian, 1997; Wendt *et al.*, 1999]: $\epsilon_{\text{Sr}} = -32$ ($^{87}\text{Sr}/^{86}\text{Sr} = 0.702245$) $\epsilon_{\text{Nd}} = 12$ ($^{143}\text{Nd}/^{144}\text{Nd} = 0.513254$), and estimated values for $\epsilon_{\text{Hf}} = 24$ ($^{176}\text{Hf}/^{177}\text{Hf} = 0.283451$) from the $\epsilon_{\text{Nd}} - \epsilon_{\text{Hf}}$ relationship in Figure 4. Curve 1 represents evolution of the DM, which differentiated from BSE 2.5 Ga. Curves 2 and 3 are evolution paths of ancient oceanic crusts formed from the DM 2.0 Ga and 1.0 Ga, respectively. Curves 4 and 5 are evolution paths the ocean crust of present composition (Table 1) would have if it had aged/decayed since 2.0 Ga and 1.0 Ga, respectively. Compared with the present-day OIB isotopic compositions given as histograms on vertical axes to the right, it is obvious that ancient subducted oceanic crusts (curves 2 and 3) that evolved to today (rectangles on the left axes) are too depleted to be sources of most present-day OIB. Note that the low ϵ_{Hf} values for curves 4 and 5 result from the low Lu/Hf ratios of the mean ocean crust which has a slightly greater depletion in heavy REEs (including Lu) than middle REEs and similarly incompatible elements like Hf. The present-day Sr-Nd-Hf isotope data for OIB are from Albarède [1995] and Salters and White [1998].

Table 1. Compositions of Various Oceanic Rocks, Mean Crust, and Effective Bulk Distribution Coefficients

	Average OIB ^a	Average N-MORB ^b	Average E-MORB ^c	Average IAB ^d	Average Gabbros ^e	Average Bulk OC ^f	Average Effective D ^g
Ba	350	13.41	99.35	125.6	3.366	7.384	0.0063
Rb	31.0	1.322	7.860	6.800	0.363	0.747	0.0208
Th	4.00	0.201	0.970	0.405	0.032	0.100	0.0547
Nb	48.0	3.333	14.51	1.041	0.396	1.570	0.0651
U	1.02	0.083	0.370	0.245	0.030	0.051	0.0877
Ta	2.70	0.227	0.843	0.135	0.036	0.112	0.0922
K	12000	1163	4016	3811	596.1	822.7	0.1293
La	37.0	4.125	11.14	3.407	1.000	2.250	0.1546
Ce	80.0	12.90	26.69	8.546	3.335	7.161	0.1985
Pb	3.20	0.494	0.938	3.696	0.321	0.390	0.2328
Pr	9.70	2.252	3.829	1.384	0.615	1.270	0.2270
Sr	660	122.9	229.9	237.3	161.9	146.3	0.2500
Nd	38.5	11.32	16.59	6.700	3.316	6.519	0.2533
Zr	280	107.8	149.5	45.24	20.35	55.34	0.2587
Hf	7.80	2.856	3.450	1.333	0.658	1.537	0.2753
P	2700	790.6	1181	432.5	119.0	387.6	0.2737
Sm	10.0	3.752	4.452	2.112	1.236	2.242	0.2814
Ti	17200	10364	11517	3858	3148	6034	0.2932
Eu	3.00	1.361	1.547	0.761	0.635	0.925	0.2938
Gd	7.62	5.043	5.251	2.625	1.744	3.064	0.2925
Tb	1.05	0.868	0.860	0.452	0.315	0.536	0.2966
Dy	5.60	5.783	5.491	3.045	2.161	3.610	0.2987
Y	29.0	33.28	31.61	17.59	11.61	20.28	0.2979
Ho	1.06	1.244	1.161	0.662	0.464	0.776	0.2987
Er	2.62	3.595	3.309	1.938	1.314	2.226	0.2983
Tm	0.35	0.516	0.469	0.276	0.196	0.324	0.2994
Yb	2.16	3.420	3.097	1.908	1.223	2.102	0.2990
Lu	0.30	0.508	0.468	0.295	0.185	0.315	0.2981
Ratios of radioactive parents/radiogenic daughters							
Rb/Sr	0.0470	0.0108	0.0342	0.0287	0.0022	0.0051	
Sm/Nd	0.2597	0.3313	0.2684	0.3152	0.3728	0.3440	
Lu/Hf	0.0385	0.1780	0.1356	0.2213	0.2818	0.2047	
⁸⁷ Rb/ ⁸⁶ Sr	0.1293	0.0296	0.0941	0.0789	0.0062	0.01405	
¹⁴⁷ Sm/ ¹⁴⁴ Nd	0.1713	0.2185	0.1770	0.2078	0.2458	0.22682	
¹⁷⁶ Lu/ ¹⁷⁷ Hf	0.0053	0.0245	0.0186	0.0304	0.0387	0.02812	
	Average OIB ^a	Average N-MORB ^b	Average E-MORB ^c	Average IAB ^d	Average Gabbros ^e	Average Bulk OC ^f	IAB/BOC ^h
Normalized to immobile Nb							
Hf/Nb				1.280		0.979	1.31
Lu/Nb				0.2831		0.2003	1.41
Sm/Nb				2.028		1.428	1.42
Nd/Nb				6.433		4.151	1.55
Sr/Nb				227.9		93.16	2.45
K/Nb				3659.0		523.9	6.98
U/Nb				0.235		0.033	7.19
Rb/Nb				6.529		0.476	13.7
Pb/Nb				3.549		0.248	14.3
Ba/Nb				120.579		4.701	25.6

^aAverage ocean island basalts (OIB) composition of *Sun and McDonough* [1989].^bAverage normal MORB (N-MORB) composition of *Niu et al.* [2002a].^cAverage E-MORB composition of *Niu et al.* [2002a].^dAverage island arc tholeiitic basalts (IAB) composition is derived from *Ewart et al.* [1998] for Tonga arc and Y. Niu (unpublished data) for Tonga and Mariana arcs.^eAverage composition of oceanic lower crustal gabbros (ODP Hole 735B) is derived from 87 whole rock ICP-MS analyses done by Y. Niu (*Bach et al.* [2001]; *Niu et al.* [2002b]; and unpublished data by Y. Niu).^fAverage composition of "bulk ocean crust" is calculated by combining 40% N-MORB (assumed to represent erupted lavas and unerupted feeding dikes of upper crust, equivalent to seismic layer 2a and 2b, respectively) with 60% average gabbros (assumed to be lower crust, equivalent to seismic layer 3) Note that using N-MORB (> 80% of all MORB) as proxy for ocean crust is adequate for the purpose of the paper. Adding lower crust dilutes the abundances of most tabulated elements in N-MORB by, on average, a factor of 3.2445 ("bulk crust" = 3.2445 × N-MORB with $R = 0.995$) without significantly affecting elemental ratios except for Th, Nb, Ta, K, Pb, Sr and Eu. However, the conclusions of the paper are not affected whether or not using lower crust gabbros. Nevertheless, "bulk ocean crust" composition is used throughout the paper.^gEffective bulk distribution coefficients for the tabulated elements are determined by relative variability (defined as $RSD\% = 1\sigma/\text{mean} \times 100$) of MORB data (*Niu and Batiza* [1997]; *Niu et al.* [1999, 2002a]), which is proportional to relative incompatibility, by assuming D for Ba being close to zero while D for heavy rare earth elements being about 0.2997 determined from various published Kd data and polybaric melting reaction of *Niu* [1997]. The latter assumes melting in the spinel lherzolite stability field, which for the majority of N-MORB is reasonable.^hBy normalizing mobile element abundances with respect to the abundances of immobile elements such as Nb, and by comparing IAB with average ocean crust, we can take ratios as reflecting and proportional to the mobility of these elements under subduction zone dehydration reactions.

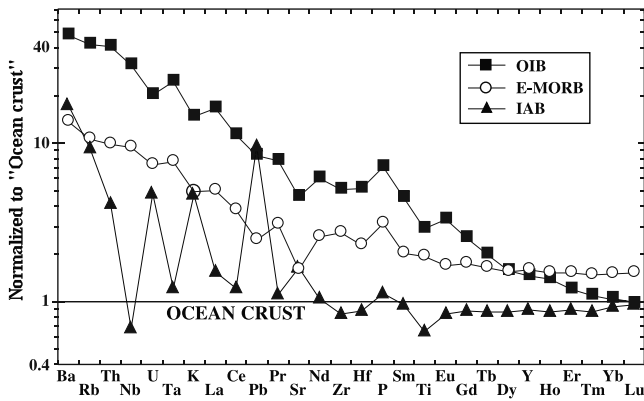


Figure 2. Multielement "spider" grams of average OIB, enriched E-MORB, average island arc tholeiites (IAB) normalized to present-day mean composition of ocean crust (see Table 1 for data).

the mantle at 2.0 and 1.0 Ga (labeled 4 and 5), respectively, would produce odd isotopic ratios that have not been observed anywhere in terrestrial rocks.

[15] The linear trends in Figures 4 and 5 pass through the accepted chondritic or BSE values in the $\epsilon_{\text{Sr}} - \epsilon_{\text{Nd}}$ space but pass above the BSE values in the $\epsilon_{\text{Hf}} - \epsilon_{\text{Nd}}$ space. It is possible that the present-day chondritic value of

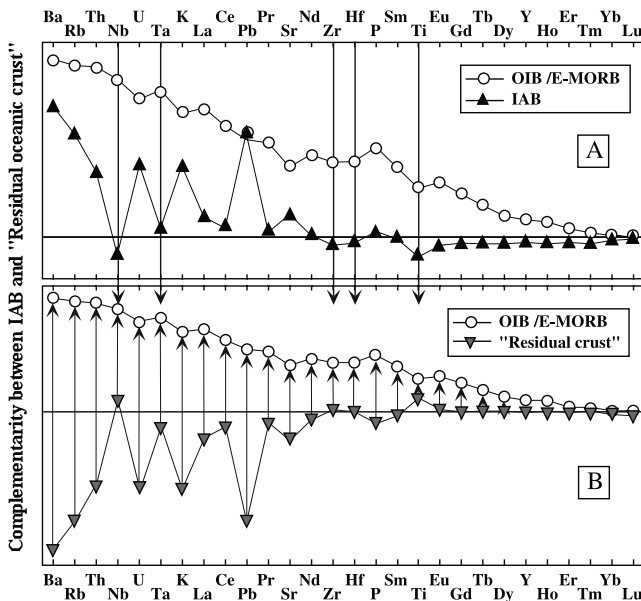


Figure 3. (a) Schematic representation of Figure 2. (b) Schematic presentation of trace element systematics a mean oceanic crust would have after passing through subduction zone dehydration reactions, residual crust. OIB/E-MORB is schematically plotted for comparison. The point is that such subduction zone filtered ocean crust is depleted in water-soluble incompatible elements like Ba, Rb, Th, U, K, Pb, and Sr, while relatively enriched (or less depleted) in water-insoluble incompatible elements like Nb, Ta, Zr, Hf, and Ti. It is not possible to melt such residual crust to produce OIB or E-MORB unless the residual crust had been refertilized in its postsubduction history by some ad hoc processes as indicated by the vertical arrows.

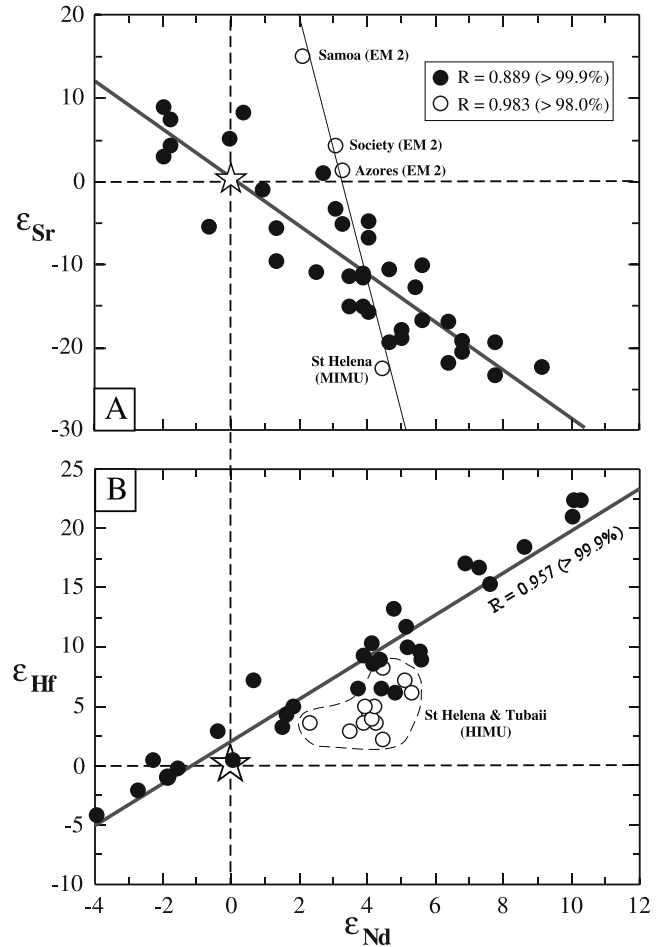


Figure 4. Plots of OIB in (a) $\epsilon_{\text{Sr}} - \epsilon_{\text{Nd}}$ (data from Albarède [1995]) and (b) $\epsilon_{\text{Hf}} - \epsilon_{\text{Nd}}$ (data from Salters and White [1998]) spaces. In Figure 4a, each data point is average representing an ocean island suite, and in Figure 4b, the data are a number of representative ocean islands with global coverage. Note that except for 3 EM2 OIB suites and one HIMU suite in Figure 4a and HIMU suites in Figure 4b, all the rest of the OIB data define statistically significant linear trends coincident with the so-called mantle arrays in these two spaces. The arrays pass through (in Figure 4a) and 1.63 ϵ_{Hf} unit above (in Figure 4b) the recommended CHUR values (Figure 1). The linear trends suggest that the process or processes that have led to the trends may be simple. The EM2 and HIMU OIB suites may be genetically different. Note also that the 3 EM2 and 1 HIMU OIB suites define a significant linear trend in Figure 4a.

$^{176}\text{Hf}/^{177}\text{Hf}_{\text{CH, today}} = 0.282772$ [Blichert-Toft and Albarède, 1997] may be lower than expected although it is not yet known what the ideal value should be [Vervoort et al., 1999]. If the value of $^{176}\text{Hf}/^{177}\text{Hf}_{\text{CH, today}} = 0.282818$ [Patchett, 1983] were used, the ϵ_{Hf} values for all the samples would be 1.63 units lower, and the data trend would perfectly pass through this latter chondritic value. The linear trends defined by the majority of the data in the two spaces are commonly interpreted as mixing trends between DM and an enriched end-member that is more enriched than BSE. We have done such an exercise by choosing two reasonable end-members (see Figure 5 caption for details). We get slightly curved

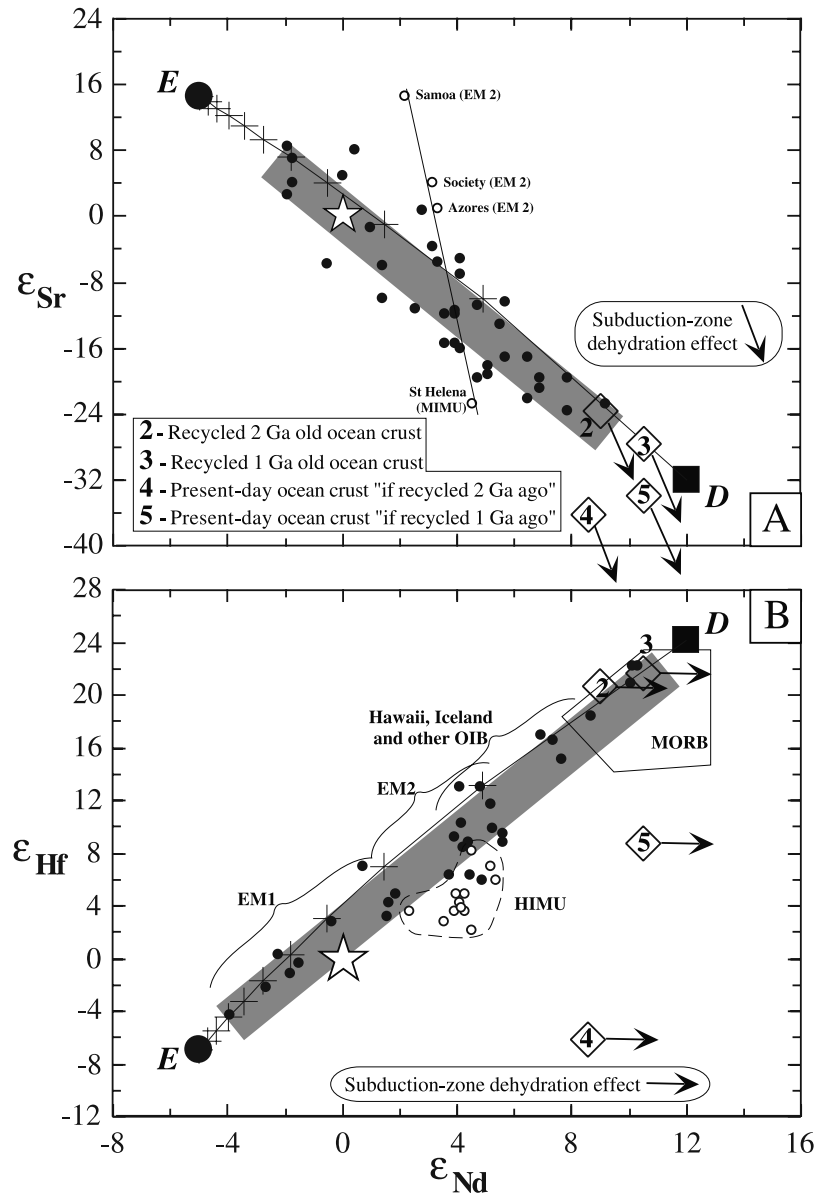


Figure 5. Same plots as in Figure 4 with expanded scales to include some simple model results. The diamonds are model results from Figure 1 as labeled. The arrows with the diamonds pointing to the effect of subduction zone dehydration reactions, which removes more Rb than Sr, more Nd than Sm, and probably have little effect on Lu and Hf, thus leading to decreasing ϵ_{Sr} , increasing ϵ_{Nd} and essentially no change in ϵ_{Hf} during post subduction isotopic evolution. As the linear trends in 2-D spaces are often interpreted as resulting from binary mixing, we have done such an exercise as indicated by the hyperbolic curves with 10% increments labeled with pluses between a depleted (*D*) and an enriched (*E*) end-members. The average of highly depleted MORB melts from the Garrett Transform in the Pacific (see Figure 1) is chosen as *D* with 82 ppm Sr, 5.95 ppm Nd, and 1.58 ppm Hf. *E* is arbitrarily chosen to enclose isotopically most enriched OIB suites along the linear trends in Figure 4, and has 600 ppm Sr, 38.5 ppm Nd, and 7.8 ppm Hf [Sun and McDonough, 1989] with $\epsilon_{\text{Sr}} = 14.52$ ($^{87}\text{Sr}/^{86}\text{Sr} = 0.705523$) $\epsilon_{\text{Nd}} = -5.0$ ($^{143}\text{Nd}/^{144}\text{Nd} = 0.512382$), and estimated values for $\epsilon_{\text{Hf}} = -7.00$ ($^{176}\text{Hf}/^{177}\text{Hf} = 0.282574$). Note that the curvatures and increments are identical in this case if source mixing is invoked. Also, the curvatures remain quite similar if another set of reasonable Sr, Nd, and Hf concentration values are used. The deviation of the mixing curvatures from the linear data trends indicates that mixing is not a logical explanation; all these OIB suites are spatially (globally) unrelated. The MORB field, the ranges of EM1 and EM2 are from *Salters and White* [1998]. Note that the 3 EM2 suites and 1 HIMU suite in Figure 5a define a linear trend, but in Figure 5b, while HIMU OIB suites are below the linear trend, both HIMU and EM2 suites have a similar ϵ_{Nd} and ϵ_{Hf} range in a broad sense, indicating that their sources have similar middle-to-heavy REE patterns despite different Rb/Sr (also U/Pb) ratios.

hyperbolic mixing lines on both plots. Mixing between “melts” or “solid sources” will give the same mixing lines in the designated scenarios. The apparent deviation of the mixing lines from the data trends would conventionally suggest a missing component that is unsampled by oceanic basalts, and that must be located somewhere lower left and lower right relative to the data trends in Figures 5a and 5b, respectively. Such a hidden component may indeed exist [McDonough, 1991; Rudnick *et al.*, 2000; Blichert-Toft and Albarède, 1997; Albarède and van der Hilst, 2002] as will be discussed later. Vervoort *et al.* [1999] argued that the mantle array will pass through the BSE values in the $\epsilon_{\text{Hf}} - \epsilon_{\text{Nd}}$ space if a higher $^{176}\text{Hf}/^{177}\text{Hf}_{\text{CH, today}}$ value is used, in which case a hidden component would not be required. However, the mixing lines are necessarily curved if proper end-member compositions are chosen [Albarède and van der Hilst, 2002]. Therefore the choice of $^{176}\text{Hf}/^{177}\text{Hf}_{\text{CH, today}}$ value is not a valid argument for or against the presence of a hidden component.

[16] Given the similar effective bulk D of Sr, Nd, and Hf in crystal-liquid processes, we suggest that this simple process is magmatic. Extraction of the bulk continental crust that depleted parts of the mantle, for example, is mostly a magmatic process. Small amounts of low mass fraction (low- F) partial melt may form at the core-mantle boundary [e.g., Williams and Garnero, 1996] or in the low-velocity zone (LVZ [e.g., Niu *et al.*, 2002a]). This low- F melt with high Rb/Sr, and low Sm/Nd and Lu/Hf ratios will be an important metasomatic agent that ascends and refertilizes, to variable extents, the DM, and less depleted or perhaps even undifferentiated mantle. This would lead to different mean Sr, Nd, and Hf isotopic signatures in the source regions of different OIB suites and to the linear correlations in Figures 4 and 5. We suggest that the linear trends result from such magmatic low-degree melt metasomatism, but the scatter about the linear trends is a function of the age of the refertilization and also possibly the small differences in the composition of both premetasomatized mantle and low-degree melts.

[17] Although their rocks are relatively minor in abundance, the deviation of EM2 and HIMU OIB suites from the main linear OIB trends in Figures 4 and 5 needs some attention. In particular, the three EM2 OIB suites and the St Helena HIMU OIB suite define a simple linear trend (Figures 4a and 5a) that is statistically significant (>98% confidence level). This suggests that their origin may be somehow related. It is generally thought that EM2 OIB reflect melting of recycled terrigenous sediments [e.g., Weaver, 1991; Hofmann, 1997], whereas HIMU OIB result from melting of recycled oceanic crust [e.g., Hofmann, 1997]. These interpretations are favored by interpretations of oxygen isotope studies [Eiler *et al.*, 1997]. It is important to note that neither EM2 nor HIMU OIB suites show trace element patterns [Weaver, 1991] that are consistent with going through subduction zone dehydration reactions (see Figure 3b). Contamination by altered oceanic crust and sediment within the superstructure of the volcanoes may be a factor [O'Hara, 1998], but the only physical scenario in which terrigenous sediments and “ocean crust” may be introduced into the mantle source region without possibly experiencing subduction zone dehydration is where a subduction zone begins to initiate along passive continental

margins [Niu *et al.*, 2001a, 2003]. Many parts of passive margins are characterized by “seaward dipping reflectors,” the thick sequences of volcanics and intrusives associated with continental breakup [Eldholm and Coffin, 2000]. Parts of the sequences overlie the oldest oceanic lithosphere (denser than continental lithosphere and oceanic plateau roots [Niu *et al.*, 2001a, 2003]). Such magmatic constructions are unlikely to have been as heavily altered and hydrated as normal oceanic crust, and thus should not have experienced significant dehydration when subducted as metamorphosed dense eclogites, along with the terrigenous sediments, into the mantle.

3. Mineral Physics Arguments

[18] One of the great achievements in mantle tomographic studies in recent years is the consensus that subducting oceanic lithosphere can indeed penetrate the 660 km seismic discontinuity (660-D) into the lower mantle [van der Hilst *et al.*, 1997; Grand *et al.*, 1997]. This supports the whole mantle convection model and the proposal that mantle plumes originate from the core-mantle boundary [e.g., Griffiths and Campbell, 1990; Davies and Richards, 1992]. This also lends support to the model by Hofmann and White [1982] that ancient oceanic crusts may be heated and segregated from the ambient mantle at the core-mantle boundary [Christensen and Hofmann, 1994], feeding mantle plumes and hot spot volcanism. To balance the downward flow of subduction, upward mass transfer from the lower mantle to the upper mantle is required. It is possible that plume flux from the deep mantle is the most important upward flow that feeds the upper mantle and also possibly the sources of oceanic basalts [Phipps Morgan *et al.*, 1995; Niu *et al.*, 1999], but it is also viable that the upward mass transfer takes place in the form of regional “swell” across the 660-D. In either scenario, the fundamental question is whether or not the ancient oceanic crust, if subducted into the lower mantle, can return to the upper mantle source regions of oceanic basalts. Recent mineral physics studies suggest that this is improbable.

3.1. Subducted Oceanic Crusts Too Dense in the Lower Mantle Conditions to Rise to the Upper Mantle

[19] Ono *et al.* [2001] have shown that subducted basaltic oceanic crust turns into an assemblage of stishovite (~24 vol %), Mg-perovskite (~33%), Ca-perovskite (~23%), and Ca-ferrite (~20%) at pressures equivalent to 800–950 km depths, which is in the upper portion of the lower mantle. This assemblage has a bulk density significantly greater than the ambient peridotitic mantle. Figure 6 summarizes their results by comparing the bulk density of oceanic crust with that of mantle peridotite under shallow lower mantle conditions as a function of depth. Assuming a whole mantle convection scenario, and considering a depth of 780 km, the temperature of ~2000 K at this depth is reasonable. In this case, the subducted oceanic crust is >2.3% denser than the ambient peridotite mantle. Such huge negative buoyancy will impede the ascent of the subducted oceanic crust to the upper mantle. If the crustal portion of the subducted lithosphere were segregated at greater depths as proposed [Christensen and Hofmann, 1994], then this crust would only rise to the level of neutral buoyancy, which is at depths of about 1600

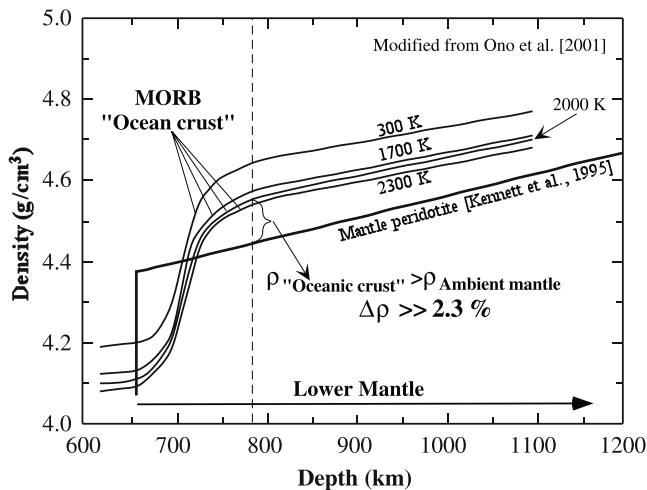


Figure 6. Modified from *Ono et al.* [2001] (copyright 2001, with permission from Elsevier Science) to show that oceanic crust subducted into the lower mantle will be transformed to a high-pressure mineral assemblage whose bulk rock density is significantly greater than the ambient peridotite mantle [Kennett et al., 1995]. For a whole mantle convection scenario, the mantle temperature would be ~ 2000 K at ~ 780 km. In this case, the oceanic crust, if subducted into the lower mantle, would be $>2.3\%$ denser than the ambient mantle. Such huge negative buoyancy impedes the rise of subducted crust into the upper mantle source regions of oceanic basalts.

km [Kesson et al., 1998; Ono et al., 2001]. If the observed seismological heterogeneity at this and deeper depths [e.g., Kaneshima and Helffrich, 1999; Kellogg et al., 1999; van der Hilst and Kárason, 1999; Albarède and van der Hilst, 2002] is indeed controlled by the level of neutral buoyancy, then it is unlikely that solid oceanic crust subducted to the core-mantle boundary will ever return as solid materials to the upper mantle source regions of oceanic basalts.

[20] In this context, subducted oceanic crust under lower mantle conditions should have greater seismic velocities than the ambient peridotitic mantle. This is because the subducted crustal lithologies have abundant high-velocity minerals such as Al-rich Mg-perovskite, Al-rich Ca-ferrite, and importantly stishovite [e.g., Ono et al., 2001; Anderson, 1989], whereas the ambient peridotite mantle lacks these phases and has abundant low-velocity minerals like magnesiowüstite [e.g., Anderson, 1989]. This suggests that the widespread small (<10 km) bodies detected using seismic scatter waves in the lower mantle [e.g., Hedlin et al., 1997] are not fragments of subducted oceanic crust contrary to some interpretations [Kaneshima and Helffrich, 1999; Helffrich and Wood, 2001] because these bodies have seismic velocities up to 4% slower, not faster, than their surroundings [Helffrich and Wood, 2001].

3.2. Basaltic Melts in Lower Mantle Conditions Probably Denser Than Ambient Solid Peridotites

[21] As rocks of basaltic compositions have lower melting temperatures than peridotitic materials in lower mantle conditions [e.g., Hirose et al., 1999; Hirose and Fei, 2002], subducted oceanic crust may melt preferentially, as

detected seismically in the seismic D'' region near the core-mantle boundary [Williams and Garnero, 1996]. It may thus be argued that subducted oceanic crusts may partially or totally melt in the deep lower mantle and the melt might then rise to the upper mantle source regions of oceanic basalts. This argument, however, is unsupported by presently available knowledge.

[22] Agee [1998, p. 107] noted the following:

Equilibrium liquidus crystals will float in variety of terrestrial and lunar magmas within a wide range of mantle depths. Density inversions exist for ultrabasic mantle melts and olivine at depths between 245 and 500 km in the Earth. Olivine flotation occurs in high-Ti basalts at ~ 400 km in the lunar mantle. Basaltic and komatiitic magmas have density inversions with garnet in the upper mantle transition zone and perovskite in the lower mantle. Peridotitic magma, because of its comparatively large bulk modulus (K_T) and pressure derivative of the bulk modulus (K'), is likely to be less dense than liquidus garnet and perovskite at all depths within the Earth.

These suggestions are supported by other studies [Suzuki et al., 1998; Ohtani and Maeda, 2001]. Figure 7, which is

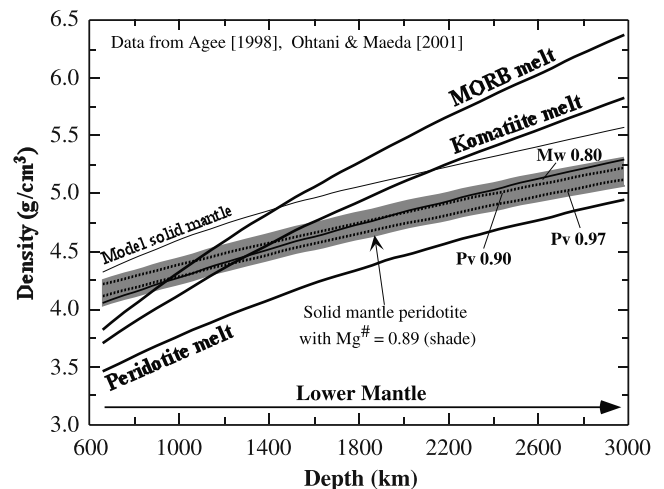


Figure 7. Modified from Agee [1998] reported at 3000°C and Ohtani and Maeda [2001] reported at 4000 K (3727°C) to show that if oceanic crust subducted into the lower mantle melted, this melt of basaltic composition would be denser and progressively more so with depth than solid minerals of proper compositions (Mw, magnesiowüstite, Pv, perovskite, where the numbers refer to $\text{Mg} \# = \text{Mg}/[\text{Mg} + \text{Fe}^{2+}]$) and the bulk mantle peridotites at depths in excess of 1400 km. At depths approaching the core-mantle boundary where the melt may actually form, the basaltic melt would be $\sim 15\%$ denser than the bulk solid mantle. This would suggest that subducted oceanic crust in the lower mantle cannot return to the upper mantle in the melt form. It must be stressed, however, that uncertainties associated with both experiments and models are probably very large, and the inferences drawn from these results thus cannot yet be validated without further experimental demonstrations. The model solid mantle is from Kennett et al. [1995], which is the same as the PREM model of Dziewonski and Anderson [1981]. The shaded band is for mantle peridotite with $\text{Mg} \# = 0.89$ [Ohtani and Maeda, 2001]. The latter is significantly less dense than the model lower mantle, probably reflecting uncertainties.

Table 2. Peridotite (KLB-1) and Basalt (MORB) Compositions^a

P, GPa	T, deg	SiO ₂	TiO ₂	Al ₂ O ₃	FeO [†]	MnO	MgO	CaO	Na ₂ O	K ₂ O	Cr ₂ O ₃	FeO/MgO
<i>KLB-1</i>												
25	2500	47.14	0.11	4.86	5.62	0.09	39.95	2.54	0.17	0	0.13	0.14
27	2450	47.91	0.1	4.28	8.18	0.14	37.84	4.35	0.26	0.01	0.32	0.22
<i>MORB</i>												
27	2430	45.8	2.33	12.86	17.23	0.23	10.37	6.89	3.93	0.33	0.05	1.66
27.5	2430	44.81	2.32	14.56	16.16	0.26	12.38	6.42	4.17	0.17	0.1	1.31

^aFeO[†] is total Fe.

modified after Agee [1998] and Ohtani and Maeda [2001], illustrates the argument. A MORB melt (e.g., from total melting of subducted oceanic crust) becomes denser, and progressively more so as pressure increases, than solid mantle minerals (perovskite and magnesiowüstite) and bulk mantle peridotite compositions at depths in excess of 1400 km. At depths close to the core-mantle boundary (or D'' region), the MORB melt would be ~15% denser than the bulk peridotitic solid mantle. The negative buoyancy of the MORB melt is so large that it is physically difficult for it to rise at all, let alone to arrive in the source regions of oceanic basalts in the upper mantle even without considering the hazard of their freezing on the way up. Starting from the popular geochemical interpretation and widely accepted notion of "mantle plumes from ancient oceanic crust" [Hofmann and White, 1982], Ohtani and Maeda [2001] had to invoke slab-derived water to lower the MORB melt density so that it could ascend as mantle plumes, but to overcome a 15% negative buoyancy is difficult. Furthermore, the subducted crust, which is on top of the subducting lithosphere, experiences the greatest extents of dehydration in subduction zones, and subsequently is water poor. On the other hand, highly serpentinized peridotites, which are likely the major lithology atop the lithospheric mantle beneath the crust [Dick, 1989; Niu and Hékinian, 1997], can contain up to 13% water. Serpentine can be stable up to 7 GPa [Ulmer and Trommsdorff, 1995], and then transformed to dense hydrous phases to carry the water deep into the lower mantle [Kuroda and Irifune, 1998]. Much of the water introduced into the deep mantle will be physically and genetically associated with recycled mantle peridotites, not recycled crust. It is possible that a water-rich fluid could be a free phase able to move around at such deep levels, but why subducted crusts should be the sinks for water released in this way is not immediately obvious.

[23] Figure 7 also shows that peridotite melts are less dense than MORB melts and komatiitic melts, and, significantly, less dense than the solid mantle peridotites under the lower mantle conditions. Formation of peridotite melts at deep mantle conditions may indeed be promoted by the subducted water as discussed above. The peridotite melts are physically the best candidates to flow into the upper mantle, feeding hot spot volcanism, a process which would also explain many petrological and geochemical observations of OIB. For example, primitive OIB are quite magnesian, a feature which could well be derived/evolved from peridotite magmas formed in the lower mantle.

[24] Melting experiments under uppermost lower mantle conditions (~25–27.5 GPa or ~700–755 km) by Hirose and Fei [2002] provide some revealing perspectives on possible melt generations and ascent at these depths. They

noted that the temperature of the mantle at these depths is lower than the anhydrous melting (solidus) temperature of both peridotite and MORB compositions. This suggests that there would be no melt produced from either types of source material at these depths without either the presence of water (including excess alkalis and other volatiles) or local thermal anomalies. If there were indeed melts produced at these depths, these melts would then be less dense than the ambient mantle (see Figure 7) [Suzuki *et al.*, 1998; Agee, 1998; Ohtani and Maeda, 2001], making it possible for them to rise. Studying the melting phase relations at these depths, Hirose and Fei produced partial melts for both peridotite (KLB-1) and basalt (MORB), with the compositions given in Table 2.

[25] While DENSICAL [Niu and Batiza, 1991] suitable for MORB and OIB is unlikely to be valid for calculating densities of melts at these great pressures, we can readily see that the melts produced by partial melting of MORB composition should be significantly denser than the melts by partial melting of peridotite. As MgO and FeO have the largest bulk moduli (87.8 and 73.6 GPa, respectively) [e.g., Herzberg, 1987] and also very large abundance differences between the two types of melts among all the major element components, the bulk melt density depends, thus, largely on relative abundances of FeO (heavy) and MgO (light) or FeO/MgO ratios. This is indeed consistent with available knowledge (Figure 7) [Suzuki *et al.*, 1998; Agee, 1998; Ohtani and Maeda, 2001]. We thus can infer that if there were melts produced at the top of the lower mantle, ultramafic melts will likely ascend readily, but melts produced by recycled crust (MORB compositions), if any, may not ascend at all or ascend with difficulties.

[26] How probable is melt production at these depths given the very high solidus temperatures (>2350°C) of both peridotite (e.g., KLB-1) and MORB composition [e.g., Herzberg and Zhang, 1996; Fei and Bertka, 1999; Hirose *et al.*, 1999; Irifune and Ringwood, 1993; Yasuda *et al.*, 1994] relative to the likely geotherm of ~1800–1900°C estimated by Stacey [1992] to be most consistent with the PREM model of Dziewonski and Anderson [1981]? An excess temperature of >400°C is clearly required for any melt to be produced if the mantle material at these depths is dry. Presence of water (accompanied by alkalis and other volatiles) will relax this requirement, but much water will still be needed to produce significant amounts of melts. Recycled oceanic crust which has passed through subduction zone dehydration reactions (eclogites) will be dry, but the underlying serpentinized lithospheric mantle may carry significant amount of water down to the lower mantle [e.g., Ulmer and Trommsdorff, 1995; Kuroda and Irifune, 1998]. As we discuss below, metasomatized deep portions of

recycled oceanic lithosphere, which is peridotitic in bulk composition, should be highly enriched in incompatible elements, alkalis as well as water and other volatiles. In summary, volatile-bearing peridotites are better candidates than recycled oceanic crust for the generation of melts atop the lower mantle.

4. Discussion

[27] The foregoing demonstrates petrologically, geochemically, and in terms of mineral physics that recycled oceanic crust cannot be the source of solid mantle plumes or partial melts feeding the bulk of hot spot volcanism and OIB. However, we cannot rule out entirely some material contributions to OIB volcanism from recycled oceanic crust, terrigenous sediments, pelagic sediments, and other shallow and near-surface processed materials in which P/D ratios and oxygen isotopes may have been fractionated. For example, EM2 OIB sources may indeed have a component of subducted terrigenous sediments [Eiler *et al.*, 1997]. As trace element patterns of EM2 lava [Weaver, 1991] show no subduction zone dehydration effects (Figure 3b), these sediments may not have been introduced into mantle through “normal” subduction zones where dehydration reactions are pervasive, and sediments may melt [Elliot *et al.*, 1997; Plank and Langmuir, 1998]. Instead, they may have been subducted along with thick basaltic piles, as opposed to hydrothermally altered and weathered normal oceanic crust, at passive continental margins during subduction initiation [Niu *et al.*, 2001a, 2003], in which case significant subduction zone dehydration may not take place. A few HIMU OIB suites have been shown to have low $\delta^{18}\text{O}$ [Eiler *et al.*, 1997], which requires a recycled gabbroic cumulate assemblage in their mantle sources if crustal level assimilation can be ruled out in OIB petrogenesis [Eiler *et al.*, 1996; O'Hara, 1998]. However, the lack of subduction zone dehydration signatures (Figure 3b) in HIMU OIB requires that the recycled crust undergo no subduction dehydration. It is possible that some HIMU OIB may have a source component derived from cumulate piles at passive margins as interpreted above for possible sediment sources of EM2 OIB. In this case, however, (1) the “basaltic crust” subducted is not normal oceanic crust, and (2) this subducted “crust” must not go down to the lower mantle otherwise it will not come back to the source regions of oceanic basalts in the upper mantle (see above). The significance of sediment and crustal material subduction during subduction initiation at passive margins requires further evaluation.

[28] Elevated $\delta^{18}\text{O}$ in olivines of some Hawaii Scientific Drilling Project (HSDP) lavas was interpreted to arise from contributions of recycled terrigenous sediments [Eiler *et al.*, 1996]. Hauri *et al.* [1996] and Lassiter and Hauri [1998] used Os isotopes to postulate that some of these lavas have a component derived from recycled oceanic crust and an associated minor amount of pelagic sediment, whereas other lavas may in fact come from recycled oceanic lithosphere. Hofmann and Jochum [1996] interpreted these Hawaiian lavas as resulting from recycled gabbros of the oceanic lower crust. Sobolev *et al.* [2000] further argued that the composition of olivine melt inclusions in some Hawaiian lavas recorded “ghost” plagioclase of ancient recycled

gabbros. While all these interpretations are interesting, their validity requires careful testing for the effects of crustal assimilation in magma chambers [Eiler *et al.*, 1996; O'Hara, 1998]. In particular, melt inclusions are often assumed to be melt trapped by liquidus minerals which are in equilibrium with that melt. This is often not the case [e.g., Gaetani and Watson, 2000]. Also, it is not clear that melt inclusions necessarily record melting processes and primary mantle melt compositions, given that the host minerals are crystallizing from the cooling and evolving melts in magma chambers. Except for a few exciting findings by Eiler *et al.* [1996, 1997], the constancy of oxygen isotopic compositions in MORB and most OIB suites sets perhaps a maximum limit on the amount of recycled oceanic crust involvement in the OIB genesis.

4.1. Origin of OIB Sources: Our Perspective

[29] Following the detailed analysis by Niu *et al.* [2002a] in the context of the petrogenesis of near-ridge seamounts on the flanks of the East Pacific Rise, we emphasize that mantle metasomatism is the most important process that produces geochemically enriched domains as dikes or veins dispersed in the predominantly depleted peridotite mantle. The concept of mantle metasomatism has been in the literature for three decades [e.g., Green, 1971; Sun and Hanson, 1975; Frey and Green, 1974, 1978; Wood, 1979; Le Roex *et al.*, 1983; Anderson, 1989, 1994; Sun and McDonough, 1989; McKenzie, 1989; Halliday *et al.*, 1995; Niu *et al.*, 1996, 1999], but we illustrate when, where, how and under what conditions this metasomatism takes place, and how this metasomatized mantle material (1) affects the geochemistry of OIB and MORB on short timescales and (2) becomes a source feeding hot spot volcanism and OIB in the long term in the context of plate tectonics and mantle convection. This model is illustrated in Figure 8, which is significantly modified after Niu *et al.* [2002a]. The reasoning that leads to this view is detailed in the following sections.

4.1.1. A Two-Component Mantle Source (Composite Lithologies) for Oceanic Basalts

[30] Many studies suggest that fertile sources of oceanic island basalts are very heterogeneous on variably small scales and also vary with time beneath a given ocean island, a given seamount and a given ridge segment. A common view is that such a heterogeneous source essentially has two components, a volumetrically small and easily melted component enriched in volatiles, alkalis and other incompatible elements dispersed in ambient depleted and refractory peridotitic matrix [Hanson, 1977; Wood, 1979; Le Roex *et al.*, 1983; Prinzhofer *et al.*, 1989; Hirschmann and Stolper, 1996; Niu *et al.*, 1996, 1999, 2001b, 2002a; Niu and Batiza, 1997; Phipps Morgan and Morgan, 1999]. The two components themselves can each be compositionally heterogeneous [Niu *et al.*, 1999, 2002a].

4.1.2. Both Components Ancient With Different Isotopic Histories

[31] In most cases, clear “mixing” trends are found in erupted lavas between an enriched component and a depleted end-member with elevated abundances of incompatible elements associated with radiogenic Sr and unradiogenic Nd isotopes and vice versa. Mixing of this kind in geochemical spaces is in fact the consequence of melting

such a two-component mantle, which may be termed “melting-induced mixing” [e.g., *Niu et al.*, 1996, 1999, 2001b, 2002a]. Melting to variable extents of such a two-component mantle or melting to similar extents of a mantle containing variable proportions of the two components results in similar mixing relationships. Such melting induced mixing implies strong coupling between incompatible trace elements and radiogenic isotopes, which further suggests that both enriched and depleted lithologies must develop their isotopic characteristic independently over a period in excess of >1.0 Gyr [e.g., *Niu et al.*, 1999, 2002a]. The melting-induced mixing has been quantitatively described previously [O'Hara, 1985, 1995] by considering the size and shape of the melting regime, and a power factor, n . Integrated melting regimes described by large values of n (lots of high-F central melting, little low-F partial melt from the periphery) would tend to produce relatively homogeneous liquids representative of the average bulk source. Those described by small and variable values of this power factor (large contributions from low-F melts near the periphery) would show the apparent effects of mixing between the lowest solidus temperature component of the mantle (the metasomatic veins) and the more abundant but less readily melted component contributed by partial melting in the central region. Melt regimes described by low values of n will also be more vulnerable to potential complications arising from the persistence of scarce mineral phases which are rich in some of the normally incompatible trace elements [O'Hara *et al.*, 2001].

4.1.3. Low-F Magmatic Origin of the Enriched Component

[32] In many sample suites with melting-induced mixing trends, the enriched samples are progressively more enriched in the more incompatible elements than in the less incompatible elements and such enrichment correlates positively with radiogenic Sr (proportional to Rb/Sr) and unradiogenic Nd (proportional to Sm/Nd). This relative incompatibility-dependent enrichment indicates that the enriched component is of magmatic origin [Niu *et al.*, 2002a]. Furthermore, this enrichment correlates positively with ratios of incompatible elements of very similar bulk distribution coefficients such as Nb/Ta, Zr/Hf, Rb/Cs, Nb/U, Ce/Pb. This may indicate that this magmatic process leading to formation of the enriched component is low-F melting because low-F melts host the highest abundances of incompatible elements, and because low-F melting can effectively fractionate elements with only subtle differences in incompatibility. Note that complex partial crystallization processes could also cause fractionation of these elements [O'Hara and Fry, 1996; O'Hara and Herzberg, 2002], but such processes will disturb the coupled correlations observed with radiogenic isotope ratios.

4.1.4. Low-F Melt Metasomatism at the Interface Between the LVZ and the Growing/Thickening Oceanic Lithosphere

[33] Low-F metasomatism is a wide spread phenomenon in the earth's mantle and perhaps throughout much of Earth's history [Green, 1991; Sun and McDonough, 1989] as evidenced from mantle xenoliths [e.g., Frey and Green, 1974, 1978; O'Reilly and Griffin, 1988], massif peridotites on land [Frey *et al.*, 1985; Takazawa *et al.*, 2000], and volcanics and xenoliths from ocean islands [e.g., Frey,

1980; Halliday *et al.*, 1995]. Delamination of metasomatized continental lithosphere has also been proposed in the literature as a mechanism to introduce enriched materials in the source regions for oceanic basalts [e.g., McKenzie and O'Nions, 1995]. However, as old (>1.0 Ga) subcontinental lithosphere is compositionally refractory and physically buoyant, its recycling into the deep mantle is difficult although possible [Griffin *et al.*, 1999; O'Reilly *et al.*, 2001]. Subduction of oceanic lithosphere into deep mantle is known to occur, and therefore, metasomatized oceanic lithospheric mantle is the better candidate (Figure 8).

[34] Volumetrically small low-F melts enriched in volatiles and incompatible elements may develop in the LVZ and migrate upward and metasomatize the cooling and growing lithosphere as fine dikes and veins [e.g., Niu *et al.*, 1999, 2002a]. The interface between the LVZ and the thickening oceanic lithosphere is the effective solidus of the multicomponent natural system where low-F melt coexists with the thickening lithosphere (Figure 8). The LVZ has the lowest seismic velocity and the lowest viscosity in the mantle. It is thus the most likely region in the asthenosphere where fluid phases or small amounts of melt may exist. These fluid phases or low-F melts are enriched in whatever volatiles and incompatible elements present, both immobile and mobile incompatible elements. As oceanic lithosphere thickens due to cooling with time through basal accretion, these enriched low-F melts are readily frozen and incorporated as “metasomatic” dikes or veins in the thickening lithosphere at the LVZ-lithosphere interface. Consequently, although the topmost oceanic lithosphere may be highly depleted (MORB residues represented by abyssal peridotites), by the time of subduction its thickened deeper portions are reservoirs of volatiles and incompatible elements hosted in metasomatized dikes or veins. We envisage that this is where “two-component” mantle develops enriched dikes and veins (phlogopite- and amphibole-bearing pyroxenites/garnet pyroxenites) dispersed in depleted peridotitic matrix.

[35] It is less likely that these deep (>60 km) portions of the lithosphere would undergo significant dehydration in subduction zones because of the thermal structure of the subducting slab. Reinjection of this lithosphere into the deep mantle with a subduction-processed oceanic crust on the top, with MORB melting residues (serpentinized harzburgites) in the middle and with a metasomatized peridotite section at the base will contribute to small-scale and large-amplitude compositional heterogeneities in the Earth's mantle [Niu *et al.*, 1999, 2002a].

4.1.5. Recycling of the Metasomatized Oceanic Lithosphere as Sources Feeding Hot Spot Volcanisms and OIB

[36] In this discussion, we assume that subducting oceanic lithosphere can penetrate the 660-D into the lower mantle and probably can approach the core-mantle boundary [e.g., Grand *et al.*, 1997], where this subducted lithosphere will be heated with time. The heating raises its bulk buoyancy. Its crust and mantle components separate during ascent by buoyancy contrast. The crust tends to sink (Figure 6), but the thermally buoyant peridotitic mantle rises. The presence of volatiles and alkalis in the metasomatized lithosphere may facilitate partial or localized total melting, which (1) enhances the buoyancy because peridotite melt is least

dense among all likely materials at all these depths (Figure 7) and (2) reduces the bulk viscosity, physically easing the ascent. The top portion of the subducted lithosphere (i.e., "serpentinized MORB residues") may still hold significant amounts of water inherited from incomplete dehydration of serpentine [Kuroda and Irifune, 1998], which may further facilitate melting. Given the elevated viscosity in the ambient lower mantle and the expectation that any bulk material sufficiently fertile to account for the local volume of OIB eruptives will also be significantly denser than the ambient upper mantle, we believe the presence of melt is important for both the initiation and ascent of the mantle plumes [O'Hara, 1975]. This differs from the traditional view that plumes are primarily hot and dry during their initiation and ascent [e.g., Griffiths and Campbell, 1990; Davies and Richards, 1992].

[37] We believe the principal constituent of what ascends in "mantle plumes" is peridotitic in composition. This contrasts with the popular view of "mantle plumes from ancient oceanic crust" [Hofmann and White, 1982; Christensen and Hofmann, 1994], which implies that the source is basaltic, and would be eclogitic in most of the upper mantle and probably far too dense to ascend. Mantle plumes of predominantly peridotitic composition with the presence of peridotitic melt are physically buoyant (Figure 7) and will rise. Chemically, the peridotitic composition of plumes explains the high-magnesian composition of primitive OIB. In contrast, plumes of basaltic composition meet neither the physical nor the chemical requirements.

4.1.6. Two Timescales of "Mantle Source Enrichments"

[38] The geophysical observation that oceanic lithosphere reaches full thickness in the time frame of ~ 70 – 80 Ma suggests that in a vertical section of a mature oceanic lithosphere, the age differences of the metasomatic dikes or veins can be as large as ~ 80 Myr. This time span is in fact quite short on geological timescale and in terms of

radiogenic isotope evolution. Intraplate hot spot volcanism, while having most of its melts derived from asthenospheric mantle, will assimilate with, remobilize or even melt the recently metasomatized lithosphere [O'Hara, 1998; Niu *et al.*, 1999, 2002a], leading to extremely enriched lavas such as alkali basalts, nephelinite, basanite on many ocean islands and some intraplate seamounts [Batiza and Vanko, 1984; Zindler *et al.*, 1984]. In this case, incompatible trace elements and radiogenic isotopes are often decoupled because of significant P/D fractionation by the recent metasomatism, yet without having enough time to produce radiogenic isotopes [e.g., Halliday *et al.*, 1995]. Some of these metasomatized or metasomatizing deep portions of the lithosphere may be perturbed by rideward asthenospheric flow, leading to MORB melts with clear trace element-isotope decoupling [Mahoney *et al.*, 1994; Halliday *et al.*, 1995; Niu *et al.*, 1996].

[39] Many OIB suites, including some Hawaiian lava suites [Hauri, 1996; Lassiter and Hauri, 1998], seamount lavas, and MORB show excellent coupling between incompatible trace elements and radiogenic isotopes and even major elements [Castillo *et al.*, 1998; Niu *et al.*, 1999, 2002a; Regelous *et al.*, 1999; Wendt *et al.*, 1999]. Such coupling requires a time interval between formation of metasomatized mantle and the volcanism in excess of 1.0 Gyr, which may represent a minimum period from the time of the subduction of metasomatized oceanic lithosphere into deep mantle to the return of these materials feeding hot spot volcanism and OIB in the upper mantle.

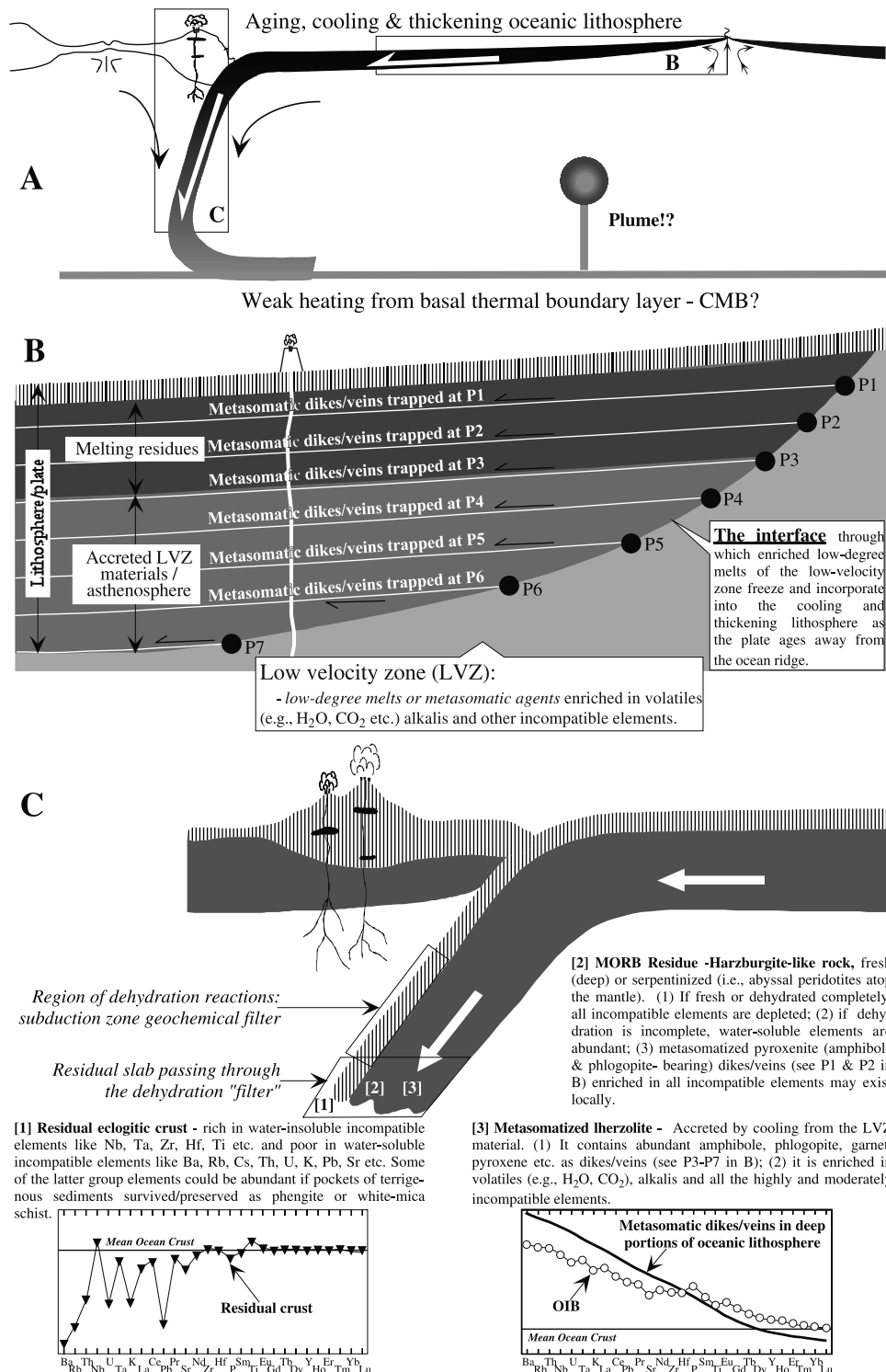
4.2. Implications for the Chemical Structure of the Earth's Mantle

[40] As discussed above and illustrated in Figures 6 and 7, oceanic crust subducted into the lower mantle will not return in bulk to the upper mantle because of the negative buoyancy in both solid and liquid states. Transfer of basaltic crust to the lower mantle would be an irreversible

Figure 8. (opposite) Schematic representations of our preferred interpretation for the origin of enriched geochemical characteristics of OIB in the context of plate tectonics and mantle convection. (a) A general sketch of the origin of oceanic lithosphere at an ocean ridge, its thickening/growth due to heat loss with time, and its ultimate return into the deep mantle. (b) A snapshot of the growing lithosphere. The oceanic lithosphere thickens with time through basal accretion due solely to heat loss to the surface. As the material to be accreted to the growing lithosphere with time comes from the LVZ, the interface between the lithosphere and the LVZ as a natural solidus representing the freezing front through which low-F melts enriched in volatiles and incompatible elements originating in the LVZ is incorporated into the thickening lithosphere in the forms of trapped metasomatic melts or fine dikes or veins. The points labeled P1–P7 indicate the low-F melts incorporated into the thickening lithosphere at different times and at different depths. While enriched materials incorporated at P1 would be the oldest, metasomatism or melt freezing can take place from P1 through P7 all at the same time. Therefore the ages of the enriched heterogeneities in the lithosphere on a vertical section cannot be resolved in the time frame (~ 70 – 80 Ma) before the lithosphere reaches its full thickness. These enriched heterogeneities would be similarly old when the recycled lithosphere returns to the source regions of oceanic basalts after >1 Ga. (c) A stratigraphic sketch of the residual oceanic lithosphere after passing through subduction zone dehydration reactions: (1) The residual crust is depleted in water-soluble incompatible elements (see inset spidergram), and is thus unsuitable as sources for mantle plumes and OIB; (2) the MORB melting residues atop the lithospheric mantle may still retain significant amount of water stored in dense hydrous phases [Kuroda and Irifune, 1998] inherited from serpentines [Ulmer and Trommsdorff, 1995] thus should also retain geochemical signatures of both MORB melting residues and serpentinization; (3) the deep portions of the lithosphere contains abundant metasomatized dikes or veins enriched in volatiles and incompatible elements (see inset spidergram). These metasomatized dikes/veins could be remobilized upward or assimilated/melted during intraplate eruptions, giving rise to some extremely enriched lavas on many ocean islands. On the other hand, when subducted into lower mantle, the metasomatized deep portions of the lithosphere will become sources of mantle plumes feeding hot spot volcanisms and OIB. Significantly modified after Niu *et al.* [2002a] (copyright 2002, with permission from Elsevier Science).

process. This supports the argument for a hidden component deep in the lower mantle that has not been sampled by known volcanism [McDonough, 1991; Blichert-Toft and Albarède, 1997; Rudnick et al., 2000; Albarède and van der Hilst, 2002] and would also lead to chemical stratification of the mantle with the mean composition of the lower mantle becoming progressively enriched in residual ocean crust lithologies (i.e., compositionally lower in Ca/Al, and higher in Fe/Mg, Si/Mg, Al, and water-

insoluble incompatible elements such as Ti, Nb, Ta, Zr, and Hf). If subduction of oceanic crusts into the lower mantle has continued for some time, then a large compositional contrast in terms of these elements must exist between the upper and lower mantle. The magnitude of such compositional contrast depends on many factors to be explored. One of the most obvious factors is when the subducting oceanic lithosphere began to penetrate the 660-D. This question is relevant because the Earth was likely



to be hotter in its early history and the oceanic lithosphere then would be thinner, reach thermal equilibrium more rapidly with the ambient mantle and thus may have lost its negative thermal buoyancy before reaching 660-D. Another important question is whether oceanic crusts can indeed descend into the lower mantle through 660-D along with the subducting lithospheric mantle. It is not improbable that the subducting crust may be stripped, at least partly, off the slab and remain in the transition zone [Ringwood and Irifune, 1988] given the lower density of the oceanic crust than the peridotite mantle around 660 km (Figure 6) [Irifune and Ringwood, 1993; Hirose et al., 1999].

5. Summary

[41] 1. There is no obvious association between ancient recycled oceanic crusts and OIB sources (mantle plumes) in terms of petrology, geochemistry, and mineral physics.

[42] 2. Melting of oceanic crusts with basaltic/picritic compositions cannot produce high-magnesian lavas parental to most OIB. Primitive OIB melts (>15% MgO) are likely more magnesian than bulk ocean crust (<13% MgO).

[43] 3. Oceanic crusts passing through subduction zone dehydration reactions should be depleted in water-soluble incompatible elements such as Ba, Rb, Cs, Th, U, K, Sr, Pb relative to water-insoluble incompatible elements such as Nb, Ta, Zr, Hf, Ti, etc. Residual crusts with such trace element systematics are unsuitable as fertile sources for OIB.

[44] 4. Ancient oceanic crusts are isotopically too depleted to produce the isotopic signatures of most OIB. OIB Sr-Nd-Hf isotopes preserve no signature that indicates previous subduction zone dehydration histories.

[45] 5. Subducted oceanic crusts at shallow lower mantle conditions form mineral assemblages that are much (>2.3%) denser than the ambient peridotitic mantle. If the subducted crusts melt in the deep lower mantle, this melt, depending upon its composition, may have still greater (up to 15%) density than solid peridotitic mantle. Therefore ancient oceanic crusts that have subducted into the deep lower mantle will not return in bulk to the upper mantle in either the solid or molten states.

[46] 6. Small fragmented components of subducted oceanic crusts could be returned to the upper mantle source regions of oceanic basalts provided they were carried along with streams of ascending buoyant material, but there is no convincing evidence for the presence of bulk subducted crust in the source regions of oceanic basalts, noting points 3 and 4 above.

[47] 7. Models invoking recycled oceanic crust to explain the geochemistry of OIB must be able to demonstrate how such crust can, by melting, produce high-magnesian lavas in many OIB suites and must also explain the lack of subduction zone dehydration signatures in OIB. Models that require ancient subducted crusts as plume sources reactivated from the lower mantle also require physical mechanisms to overcome the huge negative buoyancy of those subducted crusts in both lower and upper mantle.

[48] 8. Recycled deep portions of oceanic peridotitic lithosphere are the best candidates for the sources feeding hot spot volcanism and OIB. These deep portions of oceanic

lithosphere are likely to have been enriched in water-soluble incompatible elements like Ba, Rb, Cs, Th, U, K, Sr, Pb as well as all other incompatible elements as a result of low-F melt metasomatism at the interface between the LVZ and the cooling and thickening oceanic lithosphere. These metasomatized lithospheric materials are peridotitic in bulk composition. They can, by partial or locally total melting, produce high-magnesian melts for primitive OIB. Such materials may develop positive thermal buoyancy upon heating in the deep mantle, especially in the presence of a peridotitic melt phase, making it possible for the bulk material to ascend as plumes.

[49] 9. The solidus temperatures of both dry peridotites and MORB (proxy for recycled oceanic crust) are too high for melting to occur commonly at uppermost lower mantle conditions. However, if such melts are formed, the best candidates for a source rock would be recycled deep portions of oceanic lithosphere with abundant volatiles and alkalis (see point 8), which will lower the solidus temperature significantly. Recycled incompletely dehydrated serpentinized MORB residues atop the lithospheric mantle ("abyssal peridotites") with abundant water will also be a favored source material for melt generation.

[50] 10. If oceanic crusts subducted into the lower mantle cannot return to the upper mantle because of their negative buoyancy in both the solid and liquid states, then a "hidden" component must exist in the deep mantle that has not been sampled by known volcanism. As a result, a compositional contrast between the upper and lower mantle must have developed. The bulk lower mantle would have become progressively higher in Si/Mg, Fe/Mg etc. than the upper mantle. Given the geodynamic significance of the inference, further quantitative evaluation is needed.

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