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Mantle Plumes are NOT From Ancient Oceanic Crust

Y. Niu · M. J. O'Hara

7.1 Introduction

Basaltic volcanism mainly occurs in three tectonic settings on the Earth. Volcanism along sea-floor spreading centers produces Mid-Ocean Ridge basalts (MORB) that are depleted in incompatible elements. Volcanism above intra-oceanic subduction zones produces island arc basalts (IAB) that are enriched in water-soluble incompatible elements (e.g., Ba, Rb, Cs, Th, U, K, Pb, Sr), but depleted in water-insoluble incompatible elements (e.g., Nb, Ta, Zr, Hf, Ti). MORB and IAB are products of plate tectonics, and their geochemical differences result from differences in their respective sources and physical mechanisms through which they form. MORB are formed by plate-separation-induced passive mantle upwelling and decompression melting, thus sampling the uppermost mantle that is depleted in incompatible elements. Depletion of the MORB mantle is widely accepted as resulting from the extraction of incompatible element-enriched continental crust during the Earth's early history (Armstrong 1968; Gast 1968; O'Nions and Hamilton 1979; Jacobsen and Wasserburg 1979; DePaolo 1980; Allègre et al. 1983; Hofmann 1998). IAB are widely accepted as resulting from subducting slab-dehydration-induced melting of mantle wedge peridotites, giving rise to the characteristic geochemical signatures of slab "component", which is rich in water and water-soluble elements (e.g., Gill 1981; Tatsumi et al. 1986; McCulloch and Gamble 1991; Stolper and Newman 1994; Hawkins 1995; Pearce and Peate 1995; Davidson 1996).

Basalts are also produced by intraplate volcanism away from plate boundaries. These basalts include flood basalts erupted on land and those erupted/erupting on many oceanic islands. In contrast to MORB and IAB, these basalts are enriched in all incompatible elements and more enriched in the more incompatible elements. If the enriched characteristics in continental flood basalts were caused by continental crust contamination, then the enriched oceanic equivalent, termed ocean-island basalts (OIB), must reflect a mantle source that is enriched in incompatible elements. The observation that the shallow mantle for MORB is depleted in incompatible elements suggests that OIB must be derived from regions deeper than the MORB mantle. As the inferred OIB sources differ from undifferentiated "primitive mantle", it has thus been speculated that OIB source materials must have been previously processed and brought to the upper mantle melting regions by mantle plumes. Among the many contributions endeavoring to understand the origin of mantle plumes and OIB sources in the context of plate tectonics is the classic paper titled "*Mantle plumes from ancient oceanic crust*" by Hofmann and White (1982). These authors proposed "*oceanic crust is returned to the [lower] mantle during subduction … … Eventually, it becomes unstable as a consequence of internal heating, and the resulting diapirs [at the core-mantle boundary] become the source plumes of oceanic island basalts (OIB) and hot-spot volcanism.*"

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In this chapter, we show with evidence that there is no genetic link between ancient subducted oceanic crust and the source materials of OIB. Our arguments are based on well-understood petrology, geochemistry, and experimental data on mineral physics.

7.2 Petrological Arguments

7.2.1 Melting of Oceanic Crust Cannot Produce the High Magnesian Melts Parental to Many OIB Suites

Christensen and Hofmann (1994) explored physical scenarios about how subducted oceanic crust can isolate itself from the attached lithospheric mantle during mantle convection so as to form deep-rooted plumes to rise and feed hotspot volcanism in the upper mantle. It should be understood that the bulk oceanic crust is picritic/basaltic in composition (Niu 1997) and cannot, by melting, produce the high magnesian lavas seen in many OIB suites. It has been well established for many years that basaltic melts are derived from more magnesian picritic melts produced by partial melting of mantle peridotites (O'Hara 1968a,b; 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 compositionally basaltic/picritic (e.g., Niu 1997) and petrologically eclogitic (O'Hara and Yoder 1967; O'Hara and Herzberg 2002), will not produce basaltic/picritic melts, but melts of more silicic composition (Green and Ringwood 1968; Wyllie 1970). If total melting had occurred, the melts would be basaltic/picritic in composition but would still differ in both major and trace element systematics from those of average OIB. In fact, primitive OIB melts are much more magnesian than the most primitive MORB and likely to be more magnesian than bulk oceanic crust (Herzberg and O'Hara 1998, 2002; Clague et al. 1991; Norman and Garcia 1999). Therefore, petrologically ancient recycled oceanic crusts cannot become sources of mantle plumes feeding hotspot volcanisms and OIB.

7.3 Geochemical Arguments

7.3.1 Melting of Subduction-Zone Dehydrated Residual Oceanic Crusts Cannot Yield the Trace Element Systematics in OIB

The oceanic crust is altered during its accretion at ocean ridges and subsequently pervasively weathered/hydrated on the sea floor. This crust that is atop the subducting slab endures the greatest extent of dehydration in subduction zones. It is widely accepted that the fluids released from this dehydration lowers the solidus of the overlying mantle wedge that melts to produce arc lavas (Gill 1981; Tatsumi et al. 1986; McCulloch and Gamble 1991; Pearce and Peate 1995; Davidson 1996; Tatsumi and Kogiso 1997) as reflected in IAB geochemistry (Fig. 7.1). If this interpretation of IAB genesis

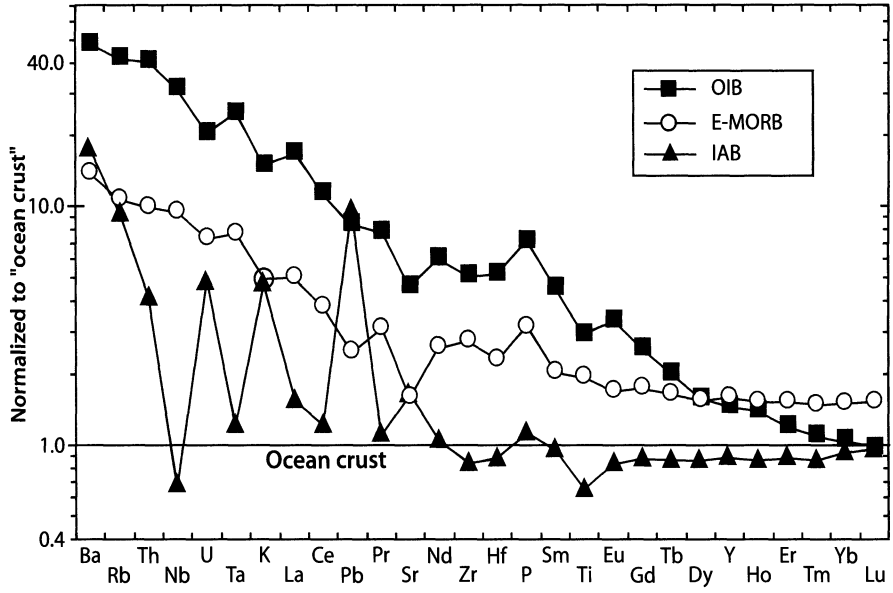

Fig. 7.1. Multi-element "spider-diagrams" of average ocean-island basalts – *OIB* (Sun and McDonough 1989), enriched E-MORB (Niu et al. 2002a), average island arc tholeiites – *IAB* (Ewart et al. 1988; Y. Niu, unpublished data for Tonga and Mariana arc tholeiites) normalized to present-day mean composition of oceanic crust (see Table 7.1 for data)

is indeed correct, then the residual subducted crust that has passed through subduction-zone dehydration reactions will have geochemical signatures that are complementary to the signatures of arc lavas (McDonough 1991; Niu et al. 1999, 2002a). In other words, this residual crust should be relatively enriched in water-insoluble incompatible elements (e.g., Nb, Ta, Zr, Hf and Ti) but highly depleted in water-soluble incompatible elements (e.g., Ba, Rb, Cs, Th, U, K, Sr, Pb etc.) (Fig. 7.2b). It follows logically that if the recycled oceanic crust were geochemically responsible for OIB, then OIB would be highly depleted in these water-soluble incompatible elements. This is not observed. In fact, OIB are enriched in these water-soluble incompatible elements as well as water-insoluble elements (Figs. 7.1 and 7.2a,b) in spite of super-chondritic Nb/Th and Ta/U ratios (Sun and McDonough 1989; Niu and Batiza 1997; Niu et al. 1999). In summary, melting or partially melting residual oceanic crust that has passed through subduction-zone dehydration reactions with the geochemical signatures shown in Fig. 7.2b will neither produce OIB nor any volcanic rocks ever sampled on the Earth's surface. Recycled terrigenous sediments would be enriched in water-soluble elements, but they also dehydrate or even melt in subduction zones, contributing to arc volcanism (Plank and Langmuir 1998; Elliot et al. 1997). If the terrigenous sediments are neither dehydrated nor melted, they still fail to explain the elevated Ce/Pb and Nb/U ratios in most OIB (Hofmann et al. 1986; Niu et al. 1999). Furthermore, terrigenous sediments with detrital zircon crystals will lead to unpredicted Zr-Hf fractionation from REE, thus giving bizarre Hf isotopes (Patchett et al. 1984; White et al. 1986), which is not observed in OIB (see below).

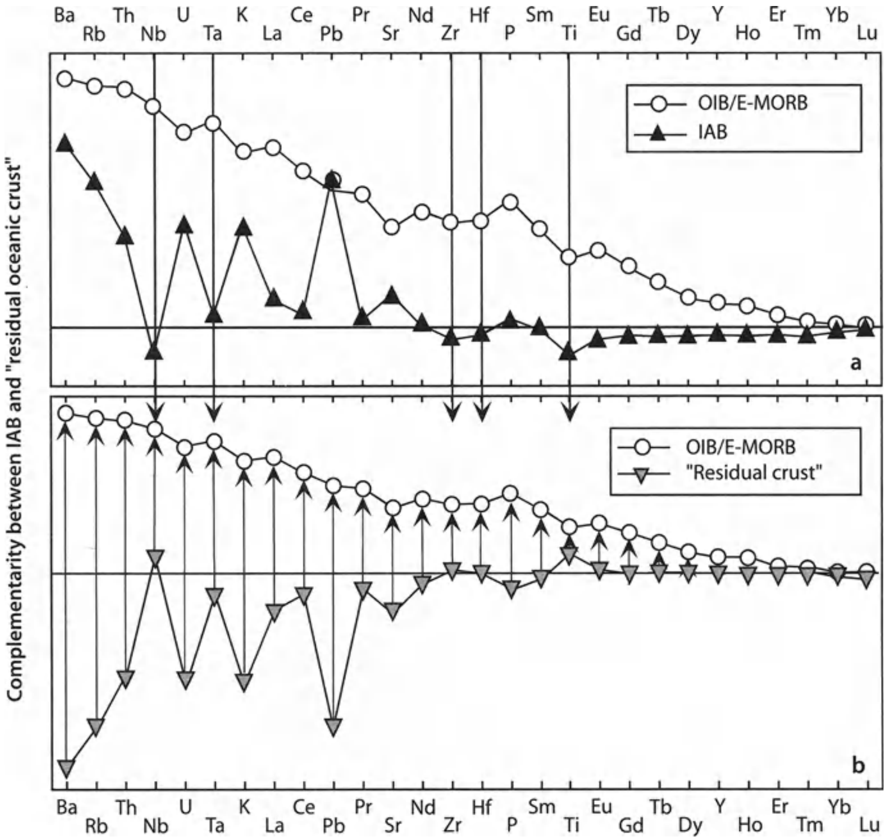

Fig. 7.2. a Schematic representation of Fig. 7.1; **b** schematic illustration of trace element systematics a mean oceanic crust would have after passing through subduction-zone dehydration reactions. The point is that such subduction-zone filtered *residual* oceanic crust is depleted in water-soluble incompatible elements like Ba, Rb, Th, U, K, Pb and Sr while relatively enriched in water-insoluble incompatible elements like Nb, Ta, Zr, Hf, and Ti. It is materially impossible to melt such a *residual* crust to produce magmas with OIB geochemical signatures, unless some form of refertilisation in the deep mantle took place as indicated by the *vertical arrows*. The latter "refertilisation" is entirely *ad hoc* with neither evidence nor physical mechanisms. *The horizontal lines* in *Part a* and *Part b* represent present-day composition of oceanic crust as shown in Fig. 7.1 and Table 7.1)

7.3.2
OIB Sr-Nd-Hf Isotopes Record no Subduction-Zone Dehydration Signatures

If ancient subducted oceanic crusts had indeed played a role in the petrogenesis of OIB, then the isotopic signatures of subduction-zone dehydration, which is non-magmatic, should be preserved in the OIB. This is, however, not observed in Sr-Nd-Hf isotopic systems. We choose these simple isotopic systems instead of complex Pb isotopes or poorly understood systems such as Os and noble gases. Such a choice is logical, because it can avoid ambiguous conclusions. Also, if the hypothesis fails for simple isotopic systems, interpretations based on complex isotopic systems in favor of that hypothesis will collapse accordingly.

Figure 7.3a plots averages of 40 OIB suites in ε_{Sr}–ε_{Nd} space (data from Albarède 1995). Figure 7.3b plots a number of OIB suites in ε_{Hf}–ε_{Nd} space (data from Salters and White 1998). Except for the so-called EM2 and HIMU OIB suites (Zindler and Hart 1986 for the acronyms) in Fig. 7.3a, all other 36 OIB suites define a scattered, yet statistically significant (at >99.9% confidence levels) inverse linear trend. Except for the HIMU-

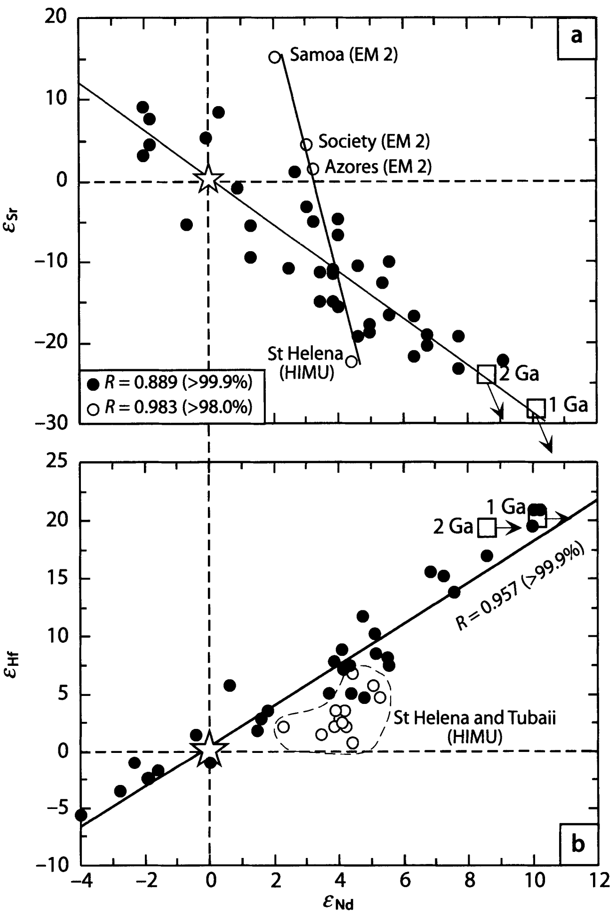

Fig. 7.3. a Plot of OIB in ε_{Sr}–ε_{Nd} space (data from Albarède 1995), where each data point is an average of an ocean island group; **b** plot of OIB in Hf-Nd space (data from Salters and White 1998), where the data represent a number of ocean islands. Note that except for three EM2 OIB suites and one HIMU suite in *Part a* and HIMU suites in *Part b*, all the rest of the OIB data define significant linear trends in these two spaces with recommended chondrite uniform reservoir (CHUR, *open stars*) values lying within the trends. These suggest that the ultimate process/processes that have led to the linear trends must be simple and magmatic because of the very similar effective bulk distribution coefficients of Sr, Nd and Hf (Table 7.1). Note also that the three EM2 and one HIMU OIB suites in *Part a* define a significant linear trend, suggesting that EM2 and HIMU OIB suites may be genetically related. The ε-notations are calculated (use εNd as an example) as $\varepsilon Nd = ({}^{143}Nd/{}^{144}Nd_{[sample]} / {}^{143}Nd/{}^{144}Nd_{(CHUR)} - 1) \times 10\,000$, using the recommended (e.g., Faure 1986; Dickin 1997) present-day CHUR values: ${}^{87}Sr/{}^{86}Sr = 0.70450$; ${}^{143}Nd/{}^{144}Nd = 0.512638$ and ${}^{176}Hf/{}^{177}Hf = 0.282818$. The latter is similar to 0.282772 ± 29 of Blichert-Toft and Albarède (1997). The *open squares* are present-day isotopic compositions the 2 Ga and 1 Ga old oceanic crusts would have. The ancient oceanic crusts were calculated by 20% batch melting from depleted MORB mantle, whose mean age is assumed to be 2.5 Ga. The *arrows* point to the effect of subduction-zone dehydration

Table 7.1a. Compositions of various oceanic rocks, mean crust, and effective bulk distribution coefficients

	Average Bulk Ocean Crust[a]	Average OIB[b]	Average IAB[c]	Average Effective D[d]
Ba	7.384	350	125.6	0.0063
Rb	0.747	31.0	6.800	0.0208
Th	0.100	4.00	0.405	0.0547
Nb	1.570	48.0	1.041	0.0651
U	0.051	1.02	0.245	0.0877
Ta	0.112	2.70	0.135	0.0922
K	822.7	12000	3811	0.1293
La	2.250	37.0	3.407	0.1546
Ce	7.161	80.0	8.546	0.1985
Pb	0.390	3.20	3.696	0.2328
Pr	1.270	9.70	1.384	0.2270
Sr	146.3	660	237.3	0.2500
Nd	6.519	38.5	6.700	0.2533
Zr	55.34	280	45.24	0.2587
Hf	1.537	7.80	1.333	0.2753
P	387.6	2700	432.5	0.2737
Sm	2.242	10.0	2.112	0.2814
Ti	6034	17200	3858	0.2932
Eu	0.925	3.00	0.761	0.2938
Gd	3.064	7.62	2.625	0.2925
Tb	0.536	1.05	0.452	0.2966
Dy	3.610	5.60	3.045	0.2987
Y	20.28	29.0	17.59	0.2979
Ho	0.776	1.06	0.662	0.2987
Er	2.226	2.62	1.938	0.2983
Tm	0.324	0.35	0.276	0.2994
Yb	2.102	2.16	1.908	0.2990
Lu	0.315	0.30	0.295	0.2981

[a] Average composition of "bulk ocean crust" (BOC) is calculated by combining 40% N-MORB (average of 132 N-MORB glass samples analyzed using ICP-MS by Y. Niu, see Niu et al. (2002a); assumed to represent erupted lavas and unerupted feeding dikes of upper ocean crust, equivalent to seismic layers 2a and 2b, respectively) with 60% oceanic gabbros (average of 87 whole-rock gabbroic samples of ODP Hole 735B analyzed using ICP-MS by Y. Niu, see Niu et al. (2002b); assumed to be lower ocean crust, equivalent seismic layer 3).

[b] Average ocean island basalts (OIB) composition of Sun and McDonough (1989).

[c] Average island arc tholeiitic basalts (IAB) composition derived from Ewart et al. (1998) for Tonga arc, and unpublished data by Y. Niu for Tonga and Mariana arcs.

[d] Effective bulk distribution coefficients for the tabulated elements are determined by relative variability (defined as RDS% = 1σ/mean × 100) of MORB data (Niu and Batiza 1997; Niu et al. 1999, 2002a), which is proportional to relative incompatibility, by assuming bulk D for Ba being close to zero while bulk D for heavy rare earth elements being about 0.2997 determined from various published Kd data and polybaric melting relation of Niu (1997).

Table 7.1b. Normalized to immobile Nb

	Average Bulk Ocean Crust	Average IAB	IAB/BOC[a]
Hf/Nb	0.979	1.280	1.31
Lu/Nb	0.2003	0.2831	1.41
Sm/Nb	1.428	2.028	1.42
Nd/Nb	4.151	6.433	1.55
Sr/Nb	93.16	227.9	2.45
K/Nb	523.9	3659	6.98
U/Nb	0.033	0.235	7.19
Rb/Nb	0.476	6.529	13.7
Pb/Nb	0.248	3.549	14.3
Ba/Nb	4.701	120.579	25.6

[a] By normalizing mobile element abundances with respect to the abundances of immobile elements such as Nb, and by comparing IAB with average bulk ocean crust (BOC), we can take the ratios as reflecting and proportional to the mobility of these elements during subduction-zone dehydration reactions.

like OIB, the data define a statistically significant (>99.9% confidence levels) positive linear trend (Fig. 7.3b). Given the relatively minor occurrences of EM2 and HIMU OIB suites on a global scale, we will first focus our discussion on the majority of OIB suites here and discuss the implications of EM2 and HIMU OIB suites later.

Because of the large differences in relative mobility of Rb > Sr > Nd > Sm > Lu > Hf during subduction dehydration inferred from observations (Table 7.1) and determined experimentally (Kogiso et al. 1997), the significant correlations in Fig. 7.3a,b 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 linear correlations thus suggest that (1) the elements Sr, Nd and Hf have behaved similarly in the respective sources of these OIB suites in the past >1 Ga; (2) the similar behavior would be unlikely if these OIB sources had experienced subduction-zone dehydration, but is to be expected if the process or processes these OIB sources had experienced were magmatic because of the similar effective bulk *D*'s of these elements (Table 7.1); and (3) coupled correlations of a radioactive parent over radiogenic daughter (*P*/*D*) for ratios such as Rb/Sr, Sm/Nd and Lu/Hf in these OIB sources must also have existed without having been disturbed in the last >1 Ga.

The open squares in Fig. 7.3a,b represent the present-day isotopic compositions that ancient oceanic crusts would have if they were produced 2 Ga and 1 Ga ago, respectively, from the depleted MORB mantle (DMM). The calculation assumes a mean age of 2.5 Ga for the DMM corresponding to the mean age of continental crust (Jacobsen and Wasserburg 1979; Taylor and McLennan 1985) because of the agreement that DMM resulted from continental crust extraction in the Earth's early history (Armstrong 1968; O'Nions and Hamilton 1979; Jacobsen and Wasserburg 1979; DePaolo 1980; Allègre et al. 1983). Regardless of model details, we cannot avoid the conclusion that the ancient subducted oceanic crusts are isotopically too depleted (too unradiogenic Sr, and too radiogenic Nd and Hf) to meet the required isotopic values of present-day OIB. Therefore, ancient recycled oceanic crusts cannot be mantle plume sources feeding intraplate volcanism and OIB. The arrows next to the open squares point to the effect of subduction-zone dehydration. Obviously, ancient oceanic crusts passing through subduction-zone dehydration would be isotopically even more depleted, therefore even more unlikely to be sources of OIB.

Although relatively minor in occurrence, the deviation of EM2 and HIMU OIB suites from the main linear OIB trends needs attention (Fig. 7.3a,b). In particular, the three EM2 OIB suites (Samoa, Society and Azores) and the HIMU OIB suite (St. Helena) define a simple but statistically significant (at >98% confidence level) linear trend (Fig. 7.3a). This suggests that their origin may be somehow related. It is generally thought that EM2 OIB reflect a source component of recycled terrigenous sediments (Weaver 1991; Hofmann 1997), whereas HIMU OIB reflect a source component of recycled oceanic crust (Hofmann 1997). Surface or near-surface processes are likely to have caused P/D (e.g., Rb/Sr, Sm/Nd, Lu/Hf, U/Pb, Th/Pb etc.) fractionation. Therefore, these near-surface, processed materials with fractionated P/D ratios, with time and when returned to mantle source regions of oceanic basalts, would produce peculiar isotopic signatures such as EM2, HIMU etc. in some OIB. However, it is imperative to note that neither EM2 nor HIMU OIB suites show trace element systematics (Weaver 1991) that are consistent with having experienced subduction-zone dehydration reactions (see Fig. 7.2b). The only physical scenario in which terrigenous sediments and "oceanic crust" could be introduced into the mantle without possibly experiencing subduction-zone dehydration is where the subduction zone begins to initiate along passive continental margins (Niu et al. 2003). Many parts of passive margins are characterized by thick sequences of volcanics and intrusives of mantle plume origin during continental break-up (Eldholm and Coffin 2000). Such mantle plume generated magmatic constructions are unlikely to have been heavily altered and hydrated (vs. normal oceanic crust) and thus should not have experienced significant dehydration when subducted as metamorphosed dense eclogites, along with the loaded terrigenous sediments, into the mantle. In this case, however, (1) the "basaltic crust" subducted is not normal oceanic crust, and (2) this subducted "crust" does not go down to the lower mantle; otherwise, it will never come back to the source regions of oceanic basalts in the upper mantle (see below). The significance of sediment and crustal subduction during subduction initiation at passive margins as a speculative hypothesis (Niu et al. 2003) requires further evaluation.

7.4 Mineral Physics Arguments

Recent mantle tomographic studies have reached the consensus that subducting oceanic lithosphere can penetrate the 660 km seismic discontinuity (660-D) into the lower mantle (van der Hist 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 (Griffiths and Campbell 1990; Davies and Richards 1992). This also lends support to the model by Hofmann and White (1982) that ancient oceanic crusts could be heated and segregated from the ambient mantle at the core-mantle boundary (Christensen and Hofmann 1994), feeding mantle plumes and hotspot volcanism. To balance the downward flow of subduction, upward mass transfer from the lower mantle to the upper mantle is required. Plume flux from the deep mantle may be the most important upward flow that feeds the upper mantle (Phipps Morgan et al. 1995; Niu et al. 1999), but it is also feasible that the upward mass transfer takes place in the form of a regional "swell" across the 660-D. In either scenario, the fundamental question is whether or not the subducted crust can return to the upper mantle source regions of oceanic basalts. Recent mineral physics studies indicate that this is physically unlikely.

7.4.1
Subducted Oceanic Crusts are too Dense to Rise to the Upper Mantle

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 shallow upper mantle conditions. This assemblage is significantly denser than the ambient peridotitic mantle. Figure 7.4 compares the bulk density of oceanic crust with that of ambient 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 2 000 K at this depth is reasonable. There, the subducted oceanic crust is >2.3% denser than the ambient peridotite mantle. Such huge negative buoyancy will impede the rise of the subducted oceanic crust into the upper mantle. If the crustal portion of the subducted lithosphere was 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 1 600 km (Kesson et al. 1998; Ono et al. 2002). If the observed seismological heterogeneity at this and deeper depths (Kaneshima and Helffrich 1999; Kellogg et al. 1999; van der Hilst and Kárason 1999) is controlled by the level of neutral buoyancy, then it is unlikely that oceanic crust subducted to the deep mantle at the core-mantle boundary will rise to the upper mantle source regions of oceanic basalts.

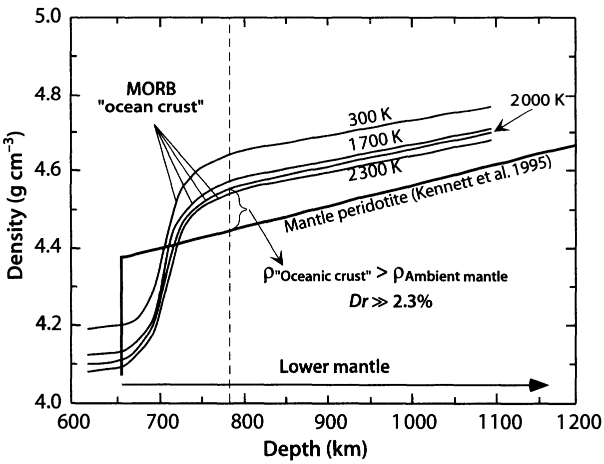

Fig. 7.4. Modified from Ono et al. (2001) to show that oceanic crust subducted into the lower mantle will be transformed to a high-pressure mineral assemblage whose bulk-rock density is significantly greater than that of the ambient peridotite mantle (Kennett et al. 1995). For a whole-mantle convection scenario, the mantle temperature would be about ~2 000 K at ~780 km. In this case, the subducted oceanic crust 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

7.4.2
Basaltic Melts in the Lower Mantle Conditions are Denser than Ambient Solid Peridotites

Melts may exist in the seismic D" region near the core-mantle boundary (Williams and Garnero 1996). As oceanic crusts likely have lower solidus temperatures relative to the ambient mantle, it is possible that subducted oceanic crusts may have contributed to the partial or total melting. However, basaltic melts are again too dense in comparison to the ambient mantle to rise (Suzuki et al. 1998; Ohtani and Maeda 2001; Agee 2001). Figure 7.5 demonstrates that a basaltic melt (compositionally equivalent to oceanic crust) becomes denser and progressively more so than solid mantle minerals (perovskite and magnesiowuestite) and bulk mantle peridotites at depths of >1 400 km. At depths close to the core-mantle boundary (or D" region), this basaltic melt is >~15% denser than the bulk peridotitic solid mantle. The negative buoyancy of the basaltic melt is so large that it is physically difficult to rise at all, let alone arrive in the source regions of oceanic basalts in the upper mantle. Because of the classic geochemical interpretation of "mantle plumes from ancient oceanic crust" (Hofmann and White 1982), Ohtani and Maeda (2001) had to invoke slab-derived water to lower the basaltic melt density in order to be consistent with that geochemical interpretation. It is practically and physically impossible to overcome the 15% negative buoyancy by adding finite water in the basaltic melt. Furthermore, the subducted crust, which is atop the subducting lithosphere, experiences the greatest extent of dehydration in subduction zones, and is thus water poor relative to the serpentinized peridotites atop the lithospheric mantle beneath the crust (Dick 1989; Niu and Hekinian 1997) and the metasomatized deep portions of oceanic lithosphere (Niu et al. 2002a; Niu and O'Hara, 2003).

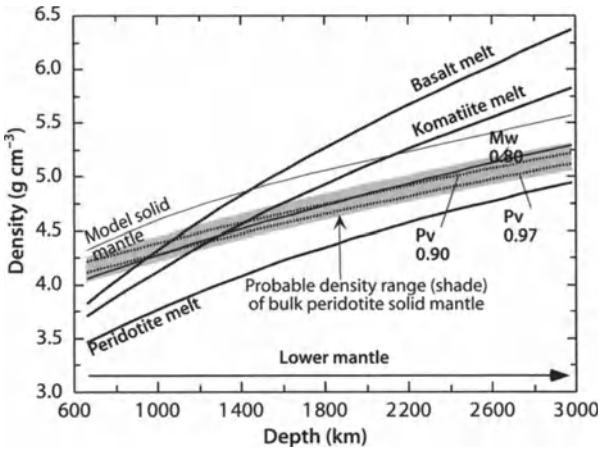

Fig. 7.5. Modified from Agee (1998) and Ohtani and Maeda (2001) to show that if oceanic crust subducted into the lower mantle melts, this melt of basaltic composition will be denser and progressively more so with depth than solid minerals of proper compositions (*Mw:* magnesiowuestite, *Pv:* perovskite) and the bulk mantle peridotites at depths in excess of 1 400 km. At depths approaching the core-mantle boundary, the basaltic melt is >15% denser than the bulk solid mantle. Therefore, subducted oceanic crust in the lower mantle cannot return to the upper mantle in the form of melt

Figure 7.5 also shows that peridotite melts are less dense than basaltic melts, komatiitic melts, and most importantly, than the solid mantle peridotites in the lower mantle conditions. We emphasize that peridotites and peridotite melts are better candidates for mantle plumes feeding intraplate volcanism and OIB. Physically, peridotites and peridotite melts are the least dense in the lower mantle conditions. Chemically, peridotites and peridotite melts can produce high magnesian basaltic/picritic melts as required by primitive OIB, which is impossible by melting recycled oceanic crusts of basaltic composition.

7.5 Summary

We have shown in terms of straightforward petrology, geochemistry and mineral physics that ancient subducted oceanic crusts cannot be source materials of mantle plumes feeding intraplate volcanism and OIB. Melting of oceanic crusts cannot produce high magnesian OIB lavas. Oceanic crusts produced from depleted mantle >1 Ga ago are isotopically too depleted to meet the required values of most OIB. Subducted oceanic crusts that have passed through subduction-zone dehydration must be depleted in water-soluble incompatible elements such as Ba, Rb, Cs, Th, U, K, Sr, Pb but relatively enriched in water-insoluble incompatible elements such as Nb, Ta, Zr, Hf, Ti. Melting of residual crusts with such trace element composition cannot produce OIB or any volcanic rocks sampled on the Earth. Oceanic crusts subducted into the lower mantle will be >2% denser than the ambient mantle at shallow lower-mantle depths. Such huge negative buoyancy will impede the subducted oceanic crusts from rising into the upper mantle. If subducted oceanic crusts melt at depths near the core-mantle boundary, such melts are even denser, up to ~15%, than the ambient peridotitic mantle. Therefore, subducted bulk oceanic crusts can neither in the solid state nor in the melt form rise into the upper mantle source regions of oceanic basalts. However, we cannot rule out the possibility that minor elements of subducted oceanic crusts might be carried into the upper mantle along with buoyant ascending plumes of peridotite compositions. Models invoking recycled oceanic crust to explain the geochemistry of OIB must be able to demonstrate how such crust can, by melting, produce the high magnesian lavas observed in many OIB suites. They must also be able to explain the lack of subduction-zone dehydration signatures in OIB. Models that require ancient subducted oceanic crusts become plume sources derived from the lower mantle must also explain the physical mechanisms needed to overcome the huge negative buoyancy of the subducted crusts in both solid state and melt form.

We suggest, following Niu et al. (2002a), that deep portions of recycled oceanic lithosphere are the best candidates for mantle plume sources. These deep portions of oceanic lithosphere are filled with dykes or veins enriched in volatiles, alkalis, and all incompatible elements as a result of low-degree melt metasomatism at the interface between the low velocity zone and the cooling and thickening oceanic lithosphere. These metasomatized lithospheric materials are peridotitic in bulk composition and can, by partial or locally total melting, produce the high magnesian melts required for primitive OIB. Such peridotite-dominated material will necessarily develop positive thermal buoyancy upon heating in the deep mantle, with or without the presence of a melt phase, making it possible for the material to ascend as plumes.

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