Oceanic Island Basalts and Mantle Plumes: The Geochemical Perspective

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isotope and trace element geochemistry

Abstract

Mantle plumes—which are usually, but not always, chemically distinct from
the mid-ocean ridge basalt (MORB)—may be rooted in the core-mantle
boundary and begin with large voluminous heads triggering massive erup-
tions or be headless and arise in the mid-mantle. Geochemistry provides
convincing evidence that mantle plumes are 100–300°C hotter than nor-
mal upper mantle and that upwelling rates within the melting region are
faster than beneath mid-ocean ridges. $^{186}Os/^{188}Os$ hints at the possibil-
ity of material from Earth's core in the Hawaiian plume, but this is not
seen in other oceanic island basalt (OIB) and has not been confirmed by
$^{182}W/^{184}W$ measurements. High $^{3}He/^{4}He$ in plumes does not require a pri-
mordial deep-mantle reservoir. The geochemical signature of mantle plumes
originates primarily through melting in the upper mantle, probably through
creation and subduction of oceanic lithosphere, but the details remain ob-
scure. Plumes are lithologically heterogeneous, consisting of stringers of
mafic material embedded in a more dominant peridotite.

CLOUDS AND VOLCANOES

Mantle plume:
focused, dominantly
vertical, buoyancy-
driven flow of rock
that gives rise to
volcanism or upward
deformation (or both)
at the surface

**Oceanic island basalt
(OIB):** the basaltic
lava typically erupted
by oceanic island and
many seamount
volcanoes

**Mid-ocean ridge
basalt (MORB):** the
basaltic lava typically
erupted on mid-ocean
ridges

Hot spot: persistent
volcanism over
considerable periods of
time at fixed location
despite plate motion;
hot spots are the
surface manifestation
of mantle plumes

Cumulus clouds often form over oceanic island volcanoes because the land heats up, warming the air and initiating atmosphere convection. As the warm air rises, a phase change eventually occurs and some of the water vapor condenses to liquid, forming a cloud. In a similar way, oceanic island volcanoes form over convectively rising hot rock. Within the rising hot rock, a phase change eventually occurs and some of the rock fuses to liquid. The liquid then migrates buoyantly to the surface, forming a volcano. Wilson (1963) recognized that the source of oceanic island volcanoes such as Hawaii was fixed, and therefore must be ultimately deeper than what we now term the lithosphere and viscously coupled asthenosphere. Elaborating on this idea, Morgan (1971) proposed that such volcanoes form over convection currents rising from the deep mantle: mantle plumes. Decompression of rising rock is the most energetically favorable way for melting to occur. Hence, there is a theoretically expected connection between convective rise of rock and melting. Together, these ideas constitute mantle plume theory and provide an elegant and straightforward explanation of most oceanic-island and other within-plate volcanism.

Whereas the mantle plume hypothesis is widely accepted, a small minority argues strenuously against it (e.g., Foulger et al. 2005). In view of the controversy, I begin with a review of the mantle plume hypothesis and how it has evolved in light of new discoveries. The plume hypothesis is essentially a physical one: Hot mantle rises convectively in a plume from the deep mantle and undergoes melting at shallow depth. Campbell (2007) provides an excellent review of the physical aspects of plume theory and how its predictions compare with observations. Chemistry, which is the main focus of this paper, is important in a number of ways. First, thermodynamics-based geothermometry provides a test of an essential prediction of plume theory: that plumes are hotter than surrounding mantle. Consequently, the products of mantle plumes, such as oceanic island basalt (OIB), should be hotter than mid-ocean ridge basalt (MORB) if the plume theory is right. Second, ratios of uranium decay-series isotopes provide an independent estimate of upwelling rates of plumes when they melt: The model predicts that upwelling should be more rapid under oceanic islands than under mid-ocean ridges. Third, as has been known for more than 40 years, OIB and MORB are chemically different, albeit with some overlap. Whereas this difference is consistent with plume theory, plume theory alone does not fully explain why this should be the case. Radiogenic isotope ratios clearly show that the difference is not merely a matter of how the two kinds of magmas are generated (although that plays a role), but that it must arise in part from differences in how the source regions of the two—the shallow mantle and the deep mantle—have evolved over billions of years. Those differences, in turn, are intimately tied to the question of how Earth's deep interior works.

In this paper I focus on OIB to the exclusion of plume-related continental basalts. These hot spots are listed in **Table 1**. Although some of these, such as the Society Islands, appear to be short-lived, the median duration of volcanism at these hot spots is 30 Ma and the mean is 47 Ma. The principal reason for the oceanic focus is that some, although certainly not all, continental basalts are contaminated by the continents through which they rise. Other continental basalts appear to be contaminated by melts from the subcontinental mantle lithosphere. One should bear in mind, however, that mantle plumes induce both oceanic and continental volcanism.

EVOLUTION OF THE PLUME HYPOTHESIS

All good theories, from quantum mechanics to plate tectonics, evolve in light of new observations, and this is true of the plume hypothesis. For example, Morgan specifically suggested that plumes "provide the motive force for continental drift." That idea has been discarded in the modern

Table 1   Vital statistics of mantle plumes' generation of oceanic island chains

Chain	Location	Oceanic crust age	Youngest volcanism	Oldest volcanism	Flux (Mg s^{-1})	Seismic Depth[3] (km)	Large igneous province
Amsterdam-St. Paul	39°S 77°E	3–5 Ma	~0 Ma	>18 Ma[4]			Rajmahal?
Ascension	8°S 14°W	7 Ma	2 Ma	6–7 Ma		~2800	No
Austral-Cook	29°S 140°W	40–90 Ma	0 Ma	30 Ma[5]	3.3[1] 3.9[2]	>1450	No
Azores	38°N 28°W	0 Ma	0 Ma	36 Ma[6]	1.1[1]	~2800	Azores Plateau?
Balleny	65°S 167°E	10–20 Ma[7]	0? Ma[7]	2.6 Ma[7]			Ferrar & Tasmania?[8]
Bouvet	54°S 3°E	<9.6 Ma	0 Ma	2.5 Ma	0.4[1]	≤1450	Karroo?[8]
Canary Is.	28°N 18°W	155 Ma	0 Ma	68 Ma[9]	1[1]	~2800	No
Cape Verde Is.	15°N 24°W	140 Ma	0 Ma	> 20 Ma	1.6[1] 0.5[2]	≥1900	No
Comoros	11°S 43°E	128 Ma	0 Ma	7.7 Ma			Madagascar?[10]
Crozet Is.	46°S 50°E	80 Ma	0 Ma	183? Ma[9]	0.5[1]	~2350	Karoo?[11]
San Felix/San Ambrosio	26°S 80°W	20–33 Ma	0 Ma	3.5 Ma	1.6[1] 2.3[2]		No
Easter Is.	27°S 108°W	4.5 Ma	0.11 Ma	>26 Ma	3.3[1]	~2800	Mid-Pacific mountains[11]
Galapagos	1°S 92°W	5 Ma	0 Ma	~100 Ma[11]	1[1]	≤1000	Caribbean[12]
Tristan-Gough	37°S 12°W	16 Ma	0 Ma	130 Ma	1.7[1] 0.5[2]		Etandeka & Parana
Hawaii	19°N 155°W	93 Ma	0 Ma	>80 Ma	8.7[1] 6.2[2]	≥2350	?
Heard-Kerguelen	49°S 69°E	118 Ma	0 Ma	118 Ma[13]	0.5[1] 0.2[2]	~2350	Kerguelen Plateau-Rajmahal?[12]
Iceland	65°N 10°W	0 Ma	0 Ma	61 Ma	1.4[1]	≤1000	North Atlantic Tertiary
Jan Mayen	80°N 9°W	0 Ma	0 Ma	>0.7 Ma			Siberia?[14]
Juan Fernandez Is.	33°S 80°W	20–33 Ma	1 Ma	4.2 Ma	1.6[1] 1.7[2]	≥1450	No
Madeira	33°N 17°W	129 Ma	1 Ma	70 Ma			No
Marion-Prince Edward	47°S 38°E	0 Ma	0 Ma	88? Ma[15]			Madagascar?[15]
Marquesas	9°S 138°W	63 Ma	0.5 Ma	43? Ma[16]	3.3[1] 4.6[2]		Shatski Rise?[11]
Martin Vas/Trindade	29°S 19′W	90 Ma	1.5 Ma	84 Ma[17]	0.5[1] 0.8[2]		Poxoreu?[17]
Reunion	21°S 56°E	64 Ma	0 Ma	65 Ma	1.9[1] 0.9[2]	≥1900	Deccan
Pitcairn-Gambier	25°S 130°W	23 Ma	0 Ma[18]	8 Ma	3.3[1] 1.7[2]		No
Samoa	15°S 168°W	100 Ma	0 Ma	24 Ma[19]	1.6[1]	~2800	No
Society Is.	18°S 149°W	45–63 Ma	0 Ma	4.3 Ma	3.3[1] 5.8[2]	~2800	No
St. Helena	16°S 6°W	41 Ma	2.5 Ma[20]	>81 Ma[21]	0.5[1] 0.3[2]	~2800	No?

[1]Sleep (1990). [2]Davies (1988). [3]Montelli et al. (2004). [4]Graham et al. (1999). [5]Dickinson (1998). [6]Cannat et al. (1999). [7]Berg et al. (1997). [8]Krishnamurthy (2000). [9]Geldmacher et al. (2005). [10]Melluso & Morra (2000). [11]Courtillot et al. (2003). [12]Duncan & Hargraves (1984). [13]Mahoney et al. (1983). [14]Carlson (1991). [15]Storey et al. (1995). [16]McNutt et al. (1989). [17]Gibson et al. (1997). [18]Stoffers et al. (1990). [19]Hart et al. (2004). [20]O'Connor et al. (1999). [21]O'Connor & le Roex (1992).

version of the theory: Plumes may affect plate motion and even initiate continental breakup, but they do not drive plate motions. Indeed, a relatively common assumption is that plate motion results from convection driven by heat from the mantle, whereas plumes are driven by heat from Earth's outer core (Davies 1988, Sleep 1990).

Early laboratory and numerical experiments revealed that plumes could readily arise at a thermal boundary layer (Whitehead & Luther 1975). As seismic studies revealed that no thermal boundary layer exists at the 660-km seismic discontinuity or elsewhere in the mid-mantle (e.g., Jordan et al. 1993), a thermal boundary layer at the base of the mantle seemed to be the only possible source of mantle plumes. Hence, the idea that plumes are generated at the core-mantle boundary became an integral part of the mantle plume theory.
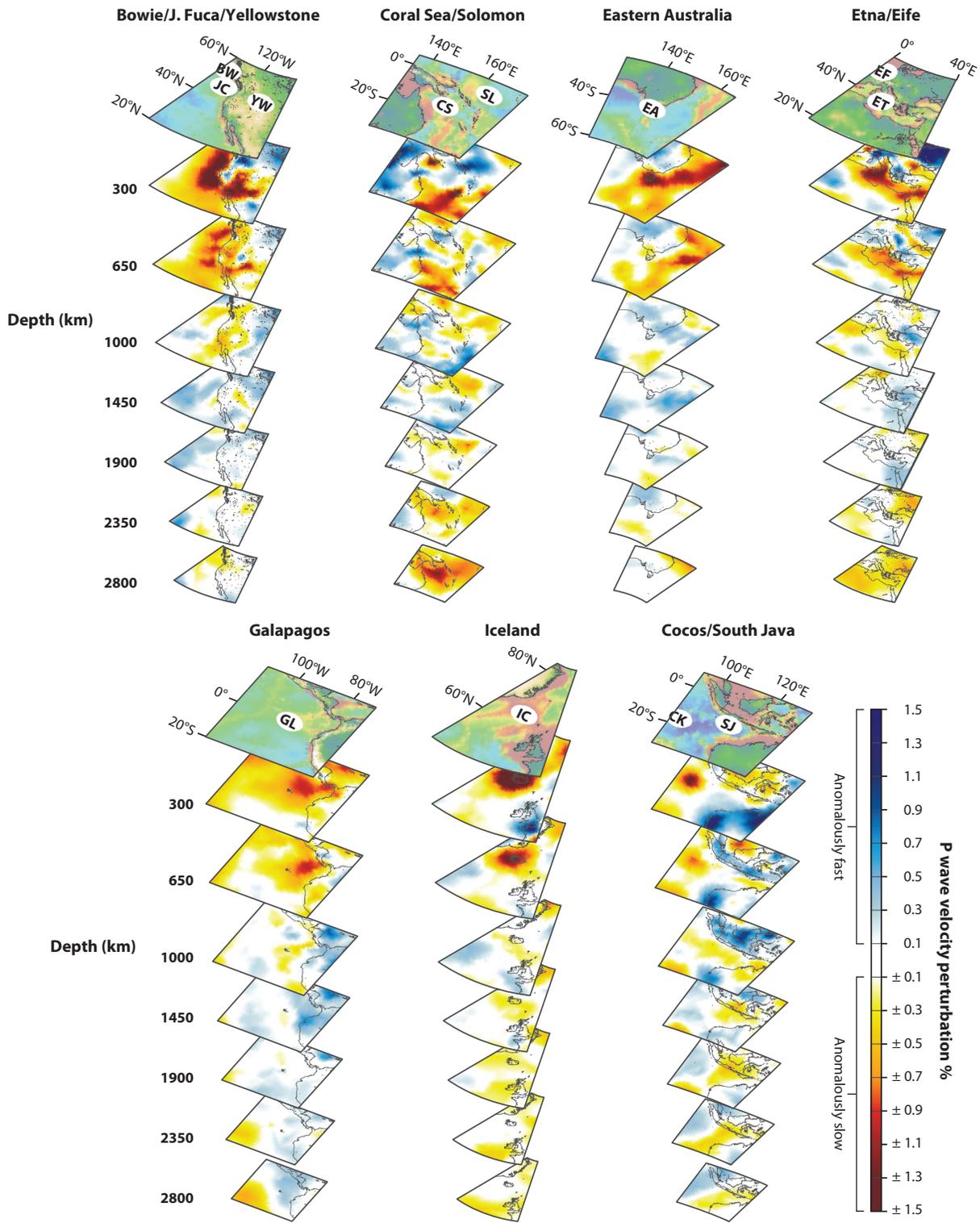
In a remarkable confirmation of the plume theory, Montelli et al. (2004) found negative seismic velocity anomalies beneath 30 hot spots. The slower velocities of seismic waves through these regions imply that they are significantly hotter, as predicted by plume theory. Of these, the anomalies extended through the 660-km seismic discontinuity into the deep mantle in 26 hot spots. In 8 cases, including Ascension/St. Helena, Azores, Canary, Easter, Samoa, and Tahiti (Society Islands), the anomalies extended to the base of the mantle (**Figure 1**). In a several other cases, including Kerguelen and Hawaii, the plumes could be traced to a depth of at least 2350 km. These 8 cases are consistent with the prediction that plumes arise from the thermal boundary layer (perhaps the seismic D″ layer) at the base of the mantle. In the remaining 16 cases, the plumes could not be traced to the core-mantle boundary region, but this could be a problem of resolution. The latter is suggested by the analysis of Burke & Torsvik (2004), who found that most of these remaining 16 cases lie above the margins of low S-wave velocity regions in the lowermost mantle, suggesting a connection between such regions and mantle plumes.

Recent numerical simulations of mantle convection, which take advantage of higher computational power not available in earlier studies and contain more realistic physics and higher resolution, may supply an explanation. First, Farnetani & Samuel (2005) show that thermochemical plumes—that is, plumes that differ in density from surrounding mantle because of both temperature and chemistry—need not be the simple, narrow, vertical conduits found in earlier laboratory and numerical simulations. Instead, they found "a great variety of plume shapes and sizes." Specifically, plumes may neck to the point where they are no longer visibly seismically. This is particularly true as they pass through the 670-km discontinuity. Second, Davies & Bunge (2006) found that cold mantle downwellings plunging into regions of hot mantle can initiate upwelling plumes on the fringes of the hot regions. The deformed hot regions and upwellings look much like the splash of a water droplet, so Davies & Bunge named them splash plumes. The bottom line here is that plumes, particularly minor ones, need not be rooted in a thermal boundary layer.

Both laboratory and numerical simulations (e.g., Whitehead & Luther 1975, Campbell & Griffiths 1990, Griffiths & Campbell 1990, Olsen 1990) suggested that plumes arising from a thermal boundary layer necessarily start with a large voluminous head. As Morgan (1972) noted, a number of oceanic island and seamount chains extrapolate to large igneous provinces (LIPs). For example, the Reunion plume, currently beneath volcanically active Reunion Island, can be connected through the Mauritius, the Mascarene Plateau, and the Chagos-Laccadive Ridge to the Deccan flood basalt province of India, which erupted 65 Mya. These observations led to the hypothesis that LIPs are the products of plume heads and oceanic islands are the products of

Large igneous provinces (LIPs): massive (>10^6 km³) eruptions of basaltic lava over relatively geologically short (a million years) time spans

Figure 1

Seismic tomographic sections of mantle plumes. Maps are 40° by 40°, appropriately scaled with depth. Yellows and reds indicate anomalously slow seismic velocities; blues indicate anomalously fast velocities. From Montelli et al. (2004). Reprinted with permission.

plume tails (Richards et al. 1989). This idea became an essential component of the plume theory. A difficulty with this, as often pointed out by plume theory opponents, is that many oceanic island chains apparently do not begin with LIPs. Of the oceanic island chains listed in **Table 1**, only five can be confidently associated with LIPs; another eight or so chains have been tentatively associated with LIPs. Again, recent computer geodynamic simulations may provide an answer. Farnetani & Samuel (2005) found that, as opposed to purely thermal plumes, the surface expression of thermochemical plumes "may be a headless, age-progressive volcanic chain." Similarly, the splash plumes that Davies & Bunge (2006) found in their simulations are not associated with the voluminous heads typical of Rayleigh-Taylor instabilities. Thus in the current version of the plume theory, plumes may or may not begin with voluminous heads.

An essential aspect of Morgan's theory was that plumes are "fixed" relative to one another. Subsequent work showed, however, that hot spots do move relative to one another (Molnar & Stock 1987), although the velocities are much lower than those of lithospheric plates. In retrospect, expecting hot spots to remain stationary was always overly simplistic. One should no more expect this than expect smoke from a chimney to always rise vertically or for a thunderhead to remain stationary. In particular, even if the base of plumes remains fixed (and numerical and laboratory simulations suggest they need not be), their surface manifestations will move because of distortion by convective motion of the mantle (mantle wind) and intraplate deformation. The former results from the fact that mass balance requires a horizontal component of flow at depth in the mantle to balance plate motions. Additional mantle flow can occur as a result of density differences, just as in the atmosphere and oceans. Taking these factors into account, Steinberger et al. (2004) were able to explain nearly all the apparent relative motion between hot spots. Indeed, when one thinks about it, the surprising thing is how little, not how much, hot spots move relative to one another. This is because $3/4$ of a plume's journey to the surface takes place through the lower mantle, whose viscosity is probably one to two orders of magnitude greater than that of the upper mantle (the latter is approximately 10^{21} to 10^{22} Pa s). Thus whereas most of the mantle flows, it does so sluggishly.

Morgan (1971) noted that his model was "compatible with the observation that there is a difference between oceanic island and ocean ridge basalts." Classic studies by Schilling (1973) and Hart et al. (1973) demonstrated that a progressive change in the chemistry of basalts erupted along the Reykjanes Ridge as Iceland was approached; in particular, changes in ratios such as La/Sm and $^{87}Sr/^{86}Sr$ required a change in the chemistry of the mantle beneath the region. Furthermore, the chemical and isotopic variations were consistent with plume material flowing southward away from Iceland in a manner predicted by Morgan. Thus the idea that plumes are chemically distinct from the upper-mantle source of MORB became an integral part of the mantle plume hypothesis. Nevertheless, at least one hot spot chain, the Cobb-Eichelberg in the Northeast Pacific, does not have a geochemical signature distinct from MORB (Chadwick et al. 2005). In addition, there are numerous regions of the mid-ocean ridge system that erupt basalts with OIB-like geochemistry (e.g., the 45°N area of the Mid-Atlantic Ridge), where there is no evidence of a mantle plume.

To summarize, in the modern version of the plume theory, plumes are buoyant upwellings of hot mantle that are usually chemically distinct from the upper-mantle source of MORB, but need not be. Some plumes may be rooted in the thermal boundary layer at the core-mantle boundary and appear to have initiated with large voluminous heads that triggered massive eruption of basalt. Other plumes, however, may be headless or may have small heads that produce small plateaus (e.g., the Azores), and may arise through still poorly understood geodynamics in the mid-mantle. Plumes are not completely stationary, but they do move relative to one another much less than lithospheric plates do.

PLUME TEMPERATURES AND UPWELLING RATES

It is possible in principle that plumes could buoyantly rise solely by virtue of their composition (Anderson 1975). However, the trace element and isotopic compositions of OIB indicate that their sources are less depleted in incompatible elements—such as the alkalis, alkaline earths, and light rare earths—than is the source of MORB. Because depletion of these incompatible elements has almost certainly been achieved through loss of a partial melt, plumes should be richer in basaltic components than the MORB source, and should be more (not less) dense than the MORB source. Consequently, if plumes rise buoyantly, it can only be because they are hot.

Petrologic data suggest that they are. The first such paper to report evidence of this was probably Dick et al. (1984), who found that abyssal peridotites in the vicinity of hot spots had mineral abundances and compositions that suggested they had melted to a greater degree than had peridotites away from hot spots. They noted that the peridotite mineralogy was correlated regionally with basalt chemistry and concluded that hot spots were indeed hotter and hence likely products of mantle plumes. Michael & Bonatti (1985) reached similar conclusions from the study of peridotite compositions. Klein & Langmuir (1987) concluded from a comprehensive study of MORB chemistry that mantle temperatures beneath the mid-ocean ridges varied by as much as 250°C. The highest temperatures were invariably found near hot spots such as Iceland. Wyllie (1988) estimated that the Hawaiian plume was perhaps 300°C hotter than mantle beneath mid-ocean ridges. A decade later, Herzberg & O'Hara (1998) estimated that the Hawaiian plume was "200–250°C hotter than present day ridges" whereas Iceland was only 100–150°C hotter than the mantle beneath ridges.

None of these studies were truly quantitative in the sense of relying on thermodynamics-based geothermometry. Because most basalts contain olivine, the Roeder & Emslie (1970) olivine geothermometer, which is based on the temperature dependency of the Mg-Fe exchange between olivine and melt, would seem the most logical choice for addressing the question of melting temperatures. The original calibration of Roeder & Emslie (1970) has been refined and extended over the years, culminating in the studies of Putirka (2005), Putirka et al. (2007), and Herzberg et al. (2007), who did careful reexaminations and recalibrations of the geothermometer while taking these factors into account. Putirka et al. (2007) then used their calculated melt temperatures to derive mantle potential temperature. They found that the mantle potential temperature beneath mid-ocean ridges is 1454 ± 81°C, that the potential temperatures of the Hawaiian and Samoan plumes are identical at 1722°C, and that the Iceland plume has a potential temperature of 1616°C. These translate to excess temperatures of 268°C and 162°C, respectively, in overall good agreement with earlier studies. Herzberg et al. (2007) found lower potential temperatures for ambient mantle—in the range of 1280–1400°C—but nevertheless conclude that plumes are typically 200–300°C hotter. Both these estimates are also in good agreement with studies that use entirely different approaches, such as the width of the geochemical anomaly along ridges (Schilling 1991) and excess topography and geoid height (Watson & McKenzie 1991). Finally, Courtier et al. (2007) found a correlation between petrologically determined temperatures in OIB and MORB and the seismically determined thickness of the underlying mantle transition zone. Beneath hot spots, the transition zone is thinner—precisely the expected result if hot spots result from hot upwellings passing through the transition zone.

Set against the results from these geothermometry studies are the results of Green et al. (2001) and Falloon et al. (2007). Also using the olivine geothermometer, they found no difference in potential temperatures between the sub-ridge and the sub-Hawaiian mantle. In an appendix, Putirka et al. (2007) discuss the possible reasons for this difference. We note here only that the results of Green et al. (2001) and Falloon et al. (2007) appear to be at odds not only with other

Incompatible elements: chemical elements that because of either ionic size or charge are not easily accommodated in mantle minerals
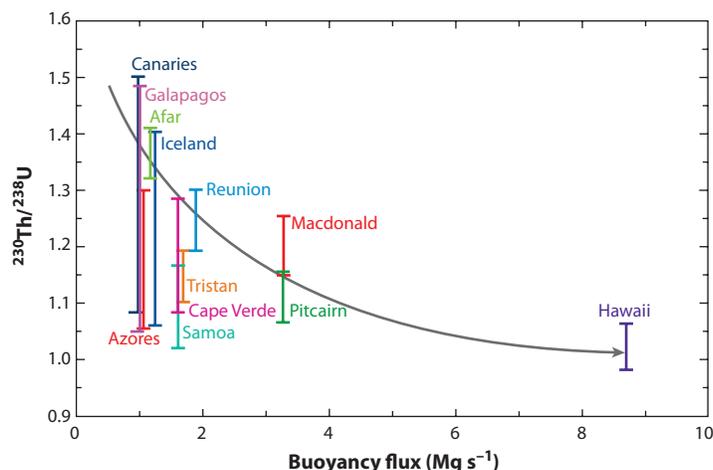
Figure 2

Plot of (^{230}Th/^{238}U) versus buoyancy flux of Sleep (1990). Includes unpublished data for the Galapagos of A. Saal (personal communication, 2009). Updated from Chabaux & Allègre (1994).

geothermometry but also with petrological and geophysical studies that have found that plumes are hot.

Melting occurs in plumes because mantle is rising. The reason is that although ascending mantle cools as a result of adiabatic decompression, its temperature drops less rapidly than does the solidus. Eventually the solidus is crossed and melting begins. The melt percolates upward, eventually ponding in crustal-magma chambers. The melt production rate is consequently related to the upwelling rate. The extent of radioactive disequilibrium between ^{230}Th and its parent, ^{238}U, should depend on these rates (e.g., Spiegelman & Elliot 1993). Consequently, the degree to which the ^{230}Th/^{238}U activity ratio (activity ratios are hereafter designated by parentheses) exceeds 1 should be a measure of the rate of mantle upwelling within the melting region—depths of 150 km and less. Indeed, Chabaux & Allègre (1994) demonstrated that (^{230}Th/^{238}U) ratios in OIB are inversely correlated with buoyancy flux as estimated by Sleep (1990). **Figure 2** shows their plot, updated to take into account subsequently published data. Bourdon et al. (2006) take this even further. With simple assumptions about melt porosity and partition coefficients and the relationship between melt production and upwelling rate, they demonstrate that the relationship observed by Chabaux & Allègre is precisely what is expected. They further show that (^{230}Th/^{238}U) and (^{231}Pa/^{235}U) ratios vary with distance from the assumed center of plumes in a manner consistent with buoyancy fluxes and excess temperatures in the range of 100–300°C. Thus OIB geochemistry provides an indirect confirmation of buoyancy fluxes estimated from geophysical observations.

A DEEP MANTLE OR CORE SIGNATURE IN PLUMES? W, OS, AND NOBLE-GAS ISOTOPE RATIOS

Geochemistry clearly requires that OIB and MORB be derived from different mantle reservoirs, but for the most part offers little about where these reservoirs might be located—even whether or not they occupy separate regions of space. The noble gases and isotopes of W and Os are potentially an exception. Noble gases such as He are thought to readily degas from the mantle when it ascends to shallow depth and melts. Hence it is commonly argued that the least degassed reservoir should be located deep in the mantle. W and Os are siderophile elements and as such

Siderophile elements: elements that partition preferentially into an iron liquid from a silicate one; lithophile elements do the opposite

are concentrated in Earth's core, Os more so than W. There is reason to believe that the isotopic compositions of these two elements should differ between core and mantle. If plumes could be shown to have Os and W with the isotopic compositions expected of the core, such a result would provide strong evidence that plumes do indeed come from the core-mantle boundary.

The isotopic composition of Os varies because ^{187}Re decays to ^{187}Os with a half-life of 42 billion years (Ga) and ^{190}Pt decays to ^{186}Os with a half-life of 450 Ga (Shirey & Walker 1998). The rarity of ^{190}Pt and its long half-life mean that variations in ^{186}Os/^{188}Os are quite small and difficult to detect. Nevertheless, Walker et al. (1997) and Brandon et al. (1998) found correlated variations in ^{186}Os/^{188}Os and ^{187}Os/^{188}Os in ores associated with the Siberia Traps flood basalts and Hawaiian basalts that they argued could only come from Earth's core. The argument is based on observed fractionations of Re/Os and Pt/Os in iron meteorites, which are fragments of asteroidal cores. As the iron of the core crystallizes, Os partitions preferentially into the crystalline phase, the inner core in the case of Earth, so that Re/Os and Pt/Os ratios increase in the remaining liquid, the outer core (Walker et al. 1995). Subsequently, similar correlated variations were found in Gorgona Island komatiites, thought to be associated with the initial stage of the Galapagos plume (Brandon et al. 2003). However, none of the other OIB and flood basalts analyzed to date have elevated ^{186}Os/^{188}Os ratios.

Tungsten isotopic composition varies in nature because ^{182}Hf decays to ^{182}W with a half-life of 9 million years (Ma). With this short half-life, essentially all isotopic variability in W in the solar system was locked in within the first 60 Ma. This decay is interesting because whereas W is siderophile and concentrated in the core, Hf is lithophile and concentrated in the mantle and crust. If Earth's core separated from the mantle before ^{182}Hf had completely decayed away, tungsten remaining in the mantle should have ended up with high ^{182}W/^{184}W. This seems to have been the case because samples from the mantle and crust do have uniformly higher ^{182}W/^{184}W than chondritic meteorites, which never experienced silicate-metal separation (Yin et al. 2002). One consequently expects that the core should have low ^{182}W/^{184}W, as do iron meteorites. If elevated ^{186}Os/^{188}Os in Hawaiian basalts results from incorporation of core material at the core-mantle boundary by the Hawaiian plume, the plume should have low ^{182}W/^{184}W. However, Scherstén et al. (2004) found that Hawaiian basalts do not have low ^{182}W/^{184}W. In addition, ^{182}W/^{184}W lower than in chondritic meteorites has not been found in any terrestrial materials, which suggests that material from the core has never reached the surface.

Because Os is far more concentrated in the core relative to the mantle than is W, the absence of a core signature in W isotopes cannot be used to entirely rule out the possibility of a core component in the Hawaiian plume. It does, however, make one wonder whether there might be some other explanation, in particular whether there are ancient crustal or mantle materials that could have evolved high ^{186}Os/^{188}Os. Fe-Mn nodules have high Pt/Os, but as Brandon et al. (1998) point out, they constitute an infinitesimal fraction of marine sediment, so it seems unlikely that they could be the source of the high ^{186}Os/^{188}Os. High Pt/Os ratios have also been observed in mantle clinopyroxenite veins (Büchl et al. 2002) and are less easily eliminated as a source of the high ^{186}Os/^{188}Os. Thus the debate about whether plumes contain a core component continues (Brandon & Walker 2005, Hawkesworth & Scherstén 2007).

Whereas ^{4}He is continually created by alpha decay, Earth's inventory of ^{3}He steadily diminishes as it degasses from Earth's interior and is then lost to space. Precisely because He is so ethereal, many geochemists have reasoned that a high ^{3}He/^{4}He reservoir should be located deep in Earth and therefore that high ^{3}He/^{4}He ratios are indicative of a less degassed, or perhaps even primitive deep-mantle reservoir. As **Figure 3** shows, OIB typically have higher ^{3}He/^{4}He than do MORB. The mean and standard deviation of the OIB data are 12.2 $\pm$ 5 R/R_A compared with 9.2 $\pm$ 3 R/R_A for MORB (R/R_A indicates that the ratio is relative to the atmospheric ratio, which is
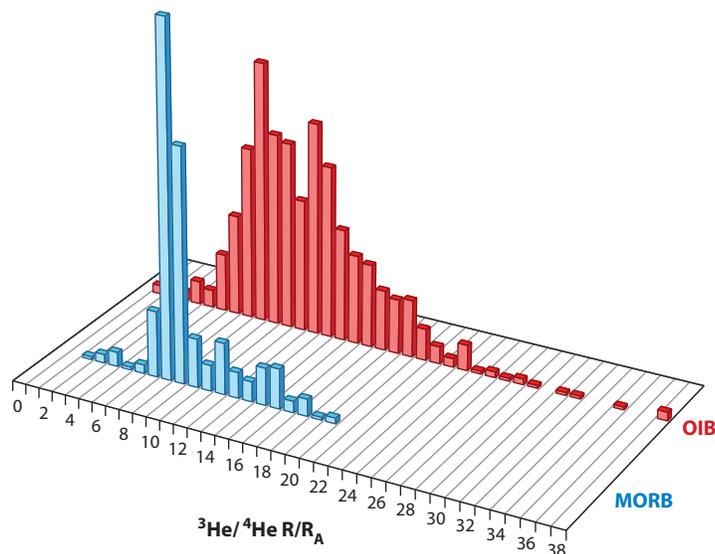
Figure 3

Histogram of ^{3}He/^{4}He in MORB and OIB. Data drawn from the EarthChem databases
(**http://www.earthchem.org**). The MORB mean is 9.2 and the OIB mean is 12.2.

1.39×10^{-6}). Much of the variation of the latter results from higher ratios found in MORB near
hot spots, such as Iceland. If these are excluded, the MORB average is 8.75 ± 2 (Graham 2002).
Thus the higher ^{3}He/^{4}He observed in OIB has formed the basis of one argument that plumes
come from a less degassed and perhaps primitive deep-mantle reservoir (e.g., Allègre 1982, Allègre
et al. 1987). Some oceanic islands, however, such as St. Helena, the Austral-Cook chain, and the
Comoros, have lower ^{3}He/^{4}He than MORB have (see **Figure 3**). This suggests that those plumes
sample a more—not less—degassed mantle. Such low ^{3}He/^{4}He plumes seem, however, to be the
exception.

The interpretation of He isotope ratios remains a matter of debate. For one thing, although
examples of local correlations exist (e.g., Kurz et al. 1987), ^{3}He/^{4}He correlates weakly at best with
other isotope ratios in OIB and MORB overall (e.g., Graham 2002). Strong correlations among
these isotope ratios would be expected if plumes were mixtures of primitive mantle and other
components. For another thing, OIB with the highest ^{3}He/^{4}He have Sr, Nd, and Pb isotopic
signatures indicative of incompatible-element-depleted sources. Although this point has been
made by others (e.g., Farley et al. 1992, Hanan & Graham 1996), it was perhaps best made by
Class & Goldstein (2005). As **Figure 4** shows, most oceanic islands with very high ^{3}He/^{4}He—
notably Iceland, the Galapagos, and Hawaii—have Nd and Pb (and Sr) isotopic compositions
closer to MORB than most OIB. Class & Goldstein (2005) present model calculations showing
that high ^{3}He/^{4}He ratios could be preserved in the mantle despite extensive melting, volcanism,
and degassing. A key assumption is that noble gases "are not extracted with 100% efficiency, but
rather behave like highly incompatible elements during melting."

This assumption is supported by new experimental data on the solubility and partitioning of
He between melt and crystals. Indeed, these results suggest that He might actually be extracted
from mantle by melting less efficiently than either U or Th (Parman et al. 2005, Heber et al. 2007),
meaning that depleted mantle—mantle that had previously experienced melt extraction—could
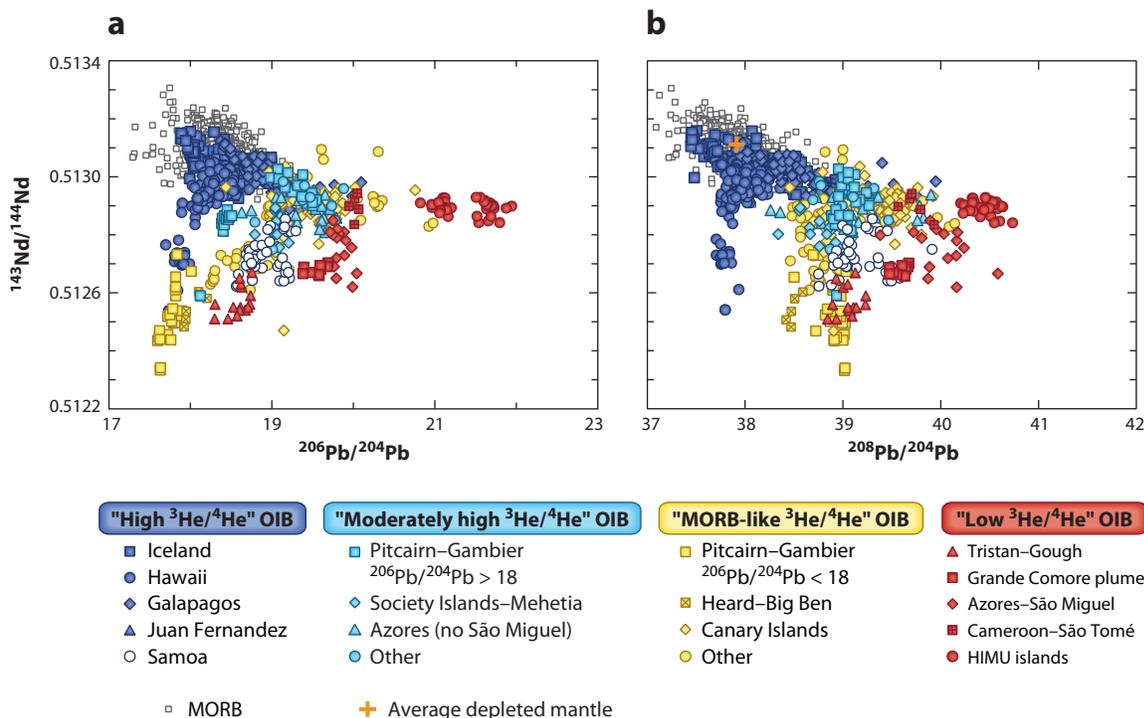be a high ^{3}He/^{4}He reservoir.

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Figure 4

Nd–Pb–He isotopes of OIB and MORB. Symbol colors correspond to the helium isotope groups of global plumes. Orange plus sign indicates estimate of average depleted (MORB) mantle. Other moderately high ^{3}He/^{4}He OIB include Cape Verde northern Islands and Reunion Island. Other MORB-like ^{3}He/^{4}He OIB include Austral-Cook Islands Rarotonga and Rurutu younger series, Cape Verde southern Islands, and Society seamounts Tehetia and Rocard. From Class & Goldstein (2005). Reprinted with permission.

Gonnermann & Mukhopadhyay (2009) argue that even if melting does extensively degas He, an isolated, high ^{3}He/^{4}He reservoir in the deep mantle is unnecessary to explain the observed data. They suggest that subduction of degassed oceanic lithosphere into the deep mantle dilutes the nobles gases to the point where "even when the mass flux across the 660-km seismic discontinuity is equivalent to approximately one lower-mantle mass over the Earth's history, high ^{3}He contents, high ^{3}He/^{4}He ratios and ^{40}Ar concentrations high enough to satisfy the ^{40}Ar mass balance of the Earth can be preserved in the lower mantle." Essentially, they find that as long as the lower mantle is "processed"—that is, melted—at a slower rate than the upper mantle, differences in noble-gas isotope ratios similar to those between OIB and MORB will develop.

In another interesting development, Jackson et al. (2008) found that ^{3}He/^{4}He is correlated with Ti enrichment in OIB such that the highest ^{3}He/^{4}He ratios always occur in Ti-rich basalts. They also found a relationship between ^{3}He/^{4}He and Nb enrichment in OIB. Finally, they found that OIB with the highest ^{3}He/^{4}He have intermediate ^{187}Os/^{188}Os, as Brandon et al. (2007) also found. Jackson et al. (2008) suggest a model in which (*a*) the high ^{3}He/^{4}He signal is contained within the peridotitic lower part of the oceanic lithosphere, and (*b*) the Ti and Nb signal is carried by the basaltic upper part of the oceanic lithosphere, which is processed into refractory eclogite during subduction. In their model, these two parts of the oceanic lithosphere remain "intimately associated" as a "package in space and time within a subducted plate." After sinking to the deep mantle, this package is entrained by mantle plumes.

It seems unlikely that we have heard the final word on the significance of He and other noble-gas isotope ratios in OIB. It does seem quite clear that high ^{3}He/^{4}He ratios in OIB do not require existence of a primordial reservoir. The recent advances in our understanding have arisen in considerable part from improvements in our understanding of the fundamental chemistry and physics of noble gas behavior. This will likely be true in the future.

ORIGIN OF THE PLUME RESERVOIR

One of the most vexing problems concerning mantle plumes and OIB is the question of what, ultimately, is responsible for the chemical differences between OIB and MORB and the chemical variability within OIB. Morgan (1971) referred to plumes as "relatively primordial." Schilling (1973) also used the word "primordial" in describing the Iceland plume. Emerging Nd isotope data seemed to support the idea of a primordial or primitive lower mantle (DePaolo & Wasserburg 1976), as did He isotope data. Mass-balance calculations (e.g., DePaolo & Wasserburg 1979, O'Nions et al. 1979) also supported the idea because the continental crust appeared to have been extracted from only a third to half of the mantle. However, as more isotope data on OIB were published, it became apparent that OIB could not be explained simply as mixtures of two components (White & Hofmann 1982). This led to a new generation of models to explain mantle chemistry and evolution.

Before considering how mantle plume reservoirs might have arisen, we should first ask this: What is the point of departure for mantle chemical evolution? In other words, what is the bulk composition of the silicate Earth? Geochemists place much faith in a modified chondritic model of Earth's composition (e.g., McDonough & Sun 1995). The model postulates that refractory elements—including the rare earths, Sr, Hf, U, and Th, but not K, Rb, or Pb—are present in Earth in the same relative abundances as they are in chondritic meteorites. This assumption underlies virtually all models of chemical evolution of Earth. One of the most significant developments in geochemistry in the past decade has been the discovery that Earth, or at least the accessible part of it, does not have a chondritic Sm/Nd ratio (Boyet & Carlson 2005). That might seem trivial, but it places the whole chondritic model in jeopardy, including the assumption that the bulk Earth has chondritic ^{143}Nd/^{144}Nd (i.e., $\varepsilon_{Nd} = 0$; ε_{Nd} is the deviation in parts per 10,000 from the chondritic isotope ratio).

Boyet & Carlson (2005, 2006) demonstrated that all terrestrial materials, with the exception of rare early Archean rocks, have ^{142}Nd/^{144}Nd ratios that are approximately 20 ppm (0.2 epsilon units) higher than those in chondritic meteorites. ^{142}Nd is the α-decay product of ^{146}Sm, whose half-life is 106 Ma and is therefore now extinct. All variation in the ^{142}Nd/^{144}Nd ratio in the solar system, including Earth, must have been established within the first 500 Ma. The high ^{142}Nd/^{144}Nd ratio implies that the Sm/Nd ratio of the accessible Earth is approximately 6% higher than the average of chondritic meteorites. This implies that ε_{Nd} for the accessible Earth is approximately +8—close to the mean value for MORB. There are only two possible explanations for these observations. The first is Boyet & Carlson's (2005, 2006) hypothesis that Earth underwent an early differentiation that produced an incompatible-element-enriched reservoir with a lower than chondritic Sm/Nd ratio and a complementary incompatible-element-depleted reservoir. They proposed that this early enriched reservoir (EER) formed as an early protocrust and was subsequently subducted into the deep mantle where it has remained ever since. In this model, the complementary early depleted reservoir, after being initially homogenized to eliminate variation in ^{142}Nd/^{144}Nd, subsequently differentiated into the continental crust and the various mantle reservoirs sampled by volcanism. In other words, all of the accessible Earth derives from the early depleted reservoir. The alternative is that Earth is nonchondritic, even for refractory lithophile elements such as the rare earths. Caro

et al. (2008) argue that the Moon, Mars, and possibly all the terrestrial planets have nonchondritic rare-earth abundances. Neither alternative seems attractive. It is unlikely that nebular processes could fractionate Sm and Nd, whose 50% condensation temperatures are 1560 K and 1563 K, respectively. On the other hand, the existence of a mantle reservoir that has remained completely isolated and unsampled for more than 4 billion years also seems problematic.

Both of these possibilities have significant implications for mantle evolution and the origin of mantle plume chemical reservoirs. To begin with, an ε_{Nd} value of +8 for Earth or the accessible Earth would mean that most mantle plume reservoirs have experienced net incompatible-element enrichment since the early differentiation. And as Boyet & Carlson (2006) explain, this would also mean that earlier mass-balance calculations are invalid and that much, perhaps 80% or more, of the mantle has been depleted in incompatible elements by continent extraction. Thus most of the mantle apparently consists of the kind of depleted mantle that gives rise to MORB, and the mantle plume reservoir must be volumetrically small. Finally, if the Boyet & Carlson (2006) model of early differentiation is correct, the most likely location for the unsampled early enriched layer would be the bottom of the mantle. That implies that plumes do not come from the core-mantle boundary.

Two other key observations constrain models of the origin of mantle plume reservoirs. First, the age of mantle heterogeneity revealed by oceanic basalts is much younger than the age of Earth. **Figure 5** is a plot of $^{207}Pb/^{204}Pb$-$^{206}Pb/^{204}Pb$ in OIB. In this plot, slopes correspond to ages (because of the different rates of decay of ^{235}U and ^{238}U). The slope of the data shown here corresponds to an age of 1.68 Ga. In this case, the age almost certainly does not correspond to a discrete event (particularly because there is not a perfect correlation between $^{207}Pb/^{204}Pb$ and $^{206}Pb/^{204}Pb$), but it likely does reflect the average age of mantle heterogeneity. It rules out, for example, the possibility that this heterogeneity arose during an early Earth differentiation event such as that proposed by Boyet & Carlson (2005). On the other hand, mantle heterogeneity must be much older than the timescale for creation and destruction of oceanic lithospheric plates.

The second key observation is that melting within the upper mantle is the primary way that mantle heterogeneity is created. Many OIB sources are enriched, and the MORB source is depleted, in incompatible elements—precisely the elements excluded by upper-mantle minerals and concentrated in magma during melting. This is illustrated in **Figure 6**, which is a spider diagram for representative OIB and MORB. In this diagram, elements are arranged from left to right in order of increasing compatibility. Melts will be enriched, and residual solids depleted, in elements on the left. The spider diagrams suggest complex histories of enrichment and depletion, but with some exceptions (most notably Pb), an element's enrichment is related to its compatibility. Melting in the lower mantle, because it would involve minerals with very different structures, would create patterns of enrichment and depletion quite different from those in **Figure 6** (e.g., Kato et al. 1988, Corgne et al. 2005). Thus whereas plumes may come from the deep mantle, they carry chemical signatures created in the upper mantle.

Two basic hypotheses, both of which posit creation of incompatible-element-enriched reservoirs in the upper mantle, have been proposed to explain the chemistry of the mantle plume reservoir. Hofmann & White (1982) suggested that mantle plumes consist of oceanic crust that had been subducted into the deep mantle, possibly the core-mantle boundary. The oceanic crust, created by melting, is enriched in incompatible elements compared with mantle peridotite. Thus they argued that the oceanic crust would eventually evolve more enriched isotopic signatures than depleted mantle. McKenzie & O'Nions (1983) suggested that the subcontinental lithosphere rather than the oceanic lithosphere was recycled into plumes. They noted that many peridotite xenoliths from the subcontinental mantle lithosphere carry incompatible-element-enriched isotopic signatures. This enrichment results when the peridotite reacts with melts or hydrous fluids
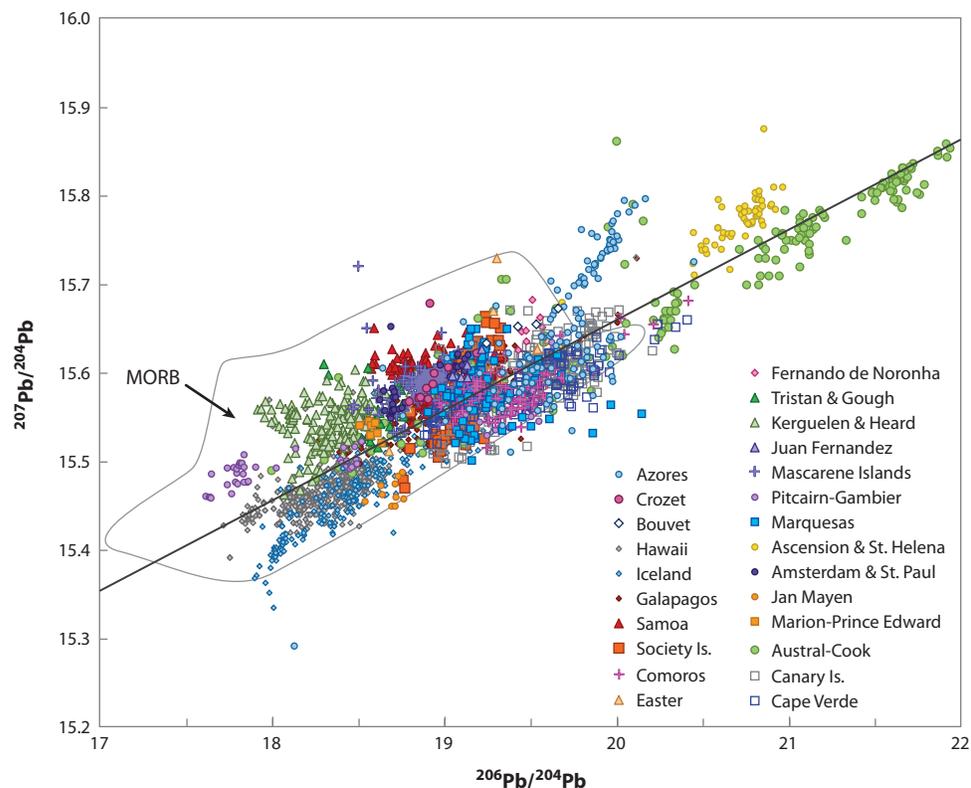
Figure 5

$^{207}Pb/^{204}Pb$-$^{206}Pb/^{204}Pb$ in OIB and MORB. A regression line based on all data is shown. The slope of the line corresponds to an age of 1.68 Ga. Data are drawn from the Geochemistry of Rocks of the Oceans and Continents (GEOROC) database (**http://georoc.mpch-mainz.gwdg.de/georoc/**). Field for MORB is shown based on data from the PetDB database (**http://www.petdb.org**).

passing through them. The melts and fluids in turn might arise from incipient melting in the asthenosphere or perhaps dehydration in subduction zones. They suggested that the subcontinental mantle lithosphere could detach from overlying crust in orogenic zones such as the Himalayas. Once detached, the cold lithosphere would sink to the base of the mantle. In a manner analogous to that described by Hofmann & White (1982), this material would eventually heat up and rise buoyantly as mantle plumes.

Most ideas proposed subsequently have been a variation on these themes (e.g., Chauvel et al. 1992, Pilet et al. 2005, Geldmacher et al. 2008). Nevertheless, neither the oceanic crust–recycling hypothesis nor the subcontinental lithosphere–recycling hypothesis is entirely satisfactory. **Figure 7** addresses the latter: It shows a plot of Os and Nd isotope ratios in OIB and in xenoliths from the subcontinental lithosphere. Much of the subcontinental lithosphere has a negative γ_{Os} signature, implying long-term depletion of Re relative to Os (Carlson & Irving 1994, Pearson et al. 1995, Pearson et al. 2004). In contrast, γ_{Os} values in OIB are almost exclusively positive. The difference between subcontinental lithosphere and OIB is even more striking when Nd isotopes are considered. OIB have predominantly positive ε_{Nd} values, although a few basalts (such as those from the Pitcairn Islands, shown in **Figure 7**) can have modestly negative ε_{Nd} values, as low as −5.
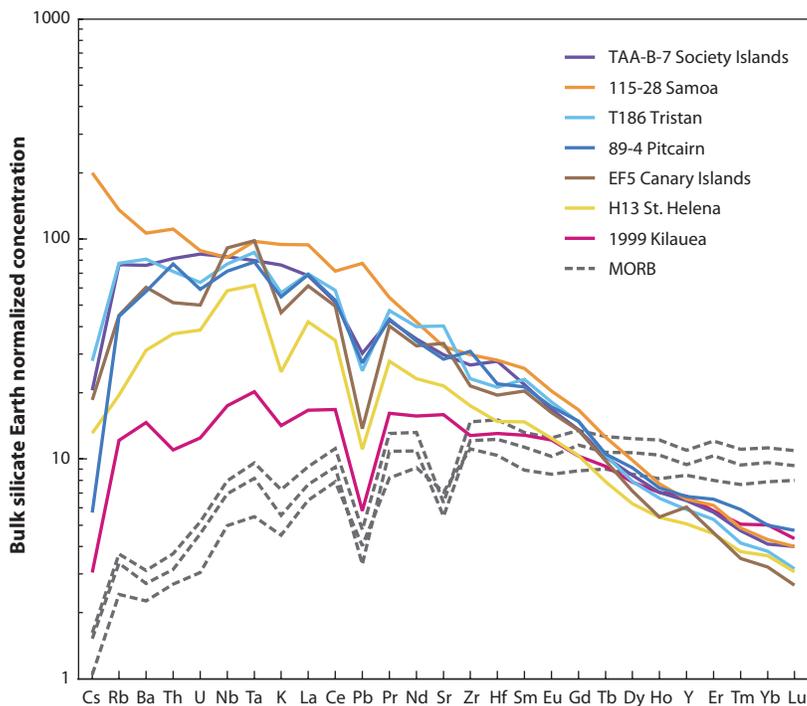
Figure 6

Spider diagram displaying the bulk earth (McDonough & Sun 1995). Normalized concentrations of trace elements are ordered by increasing compatibility. The plot shows representative patterns for MORB, Hawaii (Tanaka & Nakamura 2005), the Canary Islands (Lundstrom et al. 2003), St. Helena (Willbold & Stracke 2006) (an example of a HIMU—high μ or $^{238}U/^{204}Pb$—OIB), Tristan da Cunha (Willbold & Stracke 2006), Pitcairn (Willbold & Stracke 2006) (examples of EM I—enriched mantle 1—OIB), Samoa (Jackson et al. 2007), and the Society Islands (Hauri & Hart 1997) (examples of EM II—enriched mantle 2—OIB).

In contrast, the subcontinental lithosphere often has highly negative ε_{Nd} values, implying long-term enrichment of Nd relative to Sm. This combination of depleted-Os and enriched-Nd isotopic signatures does not occur in oceanic basalts and appears unique to the subcontinental lithospheric mantle. Apparently, it results from fluid- or melt-metasomatism that enriches the subcontinental lithosphere in Nd relative to Sm, but not Re relative to Os (e.g., Pearson et al. 1995, 2004). Given the absence of this signature within OIB, it seems unlikely that subcontinental lithosphere contributes significantly to the plume reservoir (Eisele et al. 2002, Escrig et al. 2004).

The oceanic crust–recycling hypothesis of Hofmann & White (1982) was attacked in detail by Niu & O'Hara (2003). Niu & O'Hara argue that (*a*) oceanic crust is too dense to rise as plumes, (*b*) partial melting of basalt cannot give rise to basalt, (*c*) oceanic crust should be depleted in water-soluble elements after passing through subduction zones, and (*d*) oceanic crust is too incompatible-element-depleted to give rise to plume reservoirs.

The first criticism may arise from the misconception that Hofmann & White (1982) suggested that plumes consist exclusively of oceanic crust. In fact, we always imagined that plumes consisted of a mix of recycled oceanic crust and peridotite, in part because of the impossibility of a complete physical separation of the two. Niu & O'Hara argue that oceanic crust should be 2.3% denser than normal peridotite. Simple calculations assuming a coefficient of thermal expansion of 5×10^{-5}
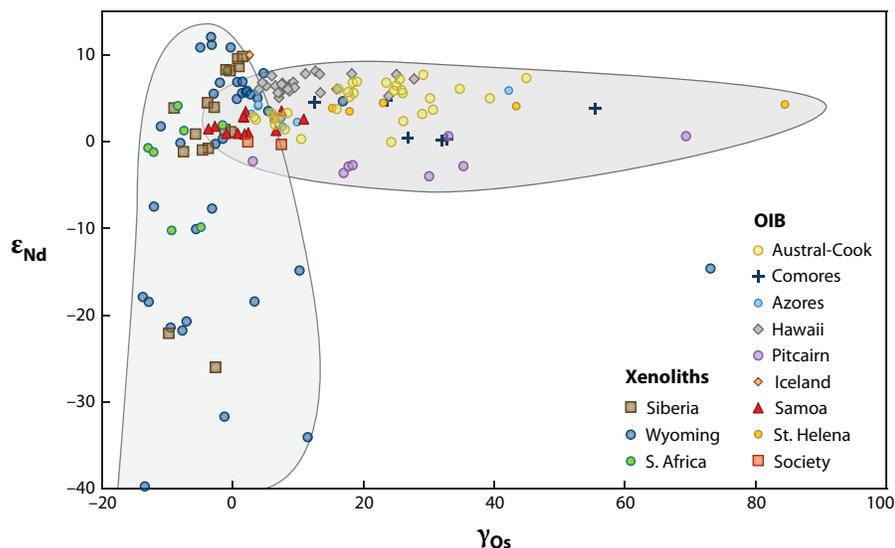
Figure 7

ε_{Nd} versus γ_{Os} in OIB and peridotite xenoliths from the subcontinental lithosphere. γ_{Os} is defined as the relative deviation of $^{187}Os/^{188}Os$ from the chondritic value (0.127) in percent.

show that roughly half of this density excess is overcome if plumes have an excess temperature of 250°C. Consequently, any hot plume containing recycled crust will indeed rise provided the fraction of crust is less than 50%. The second point is strictly true, but again, it is germane only if plumes consist exclusively of oceanic crust. The consequences of melting mixed eclogite-peridotite lithologies are discussed in greater detail in the following section.

The third criticism has been addressed in a new study by Porter & White (in press), who used a mass-balance approach to assess the effects of subduction-zone processing on subducting oceanic crust and sediment. They found that soluble elements, such as K, Rb, and Sr, are indeed extracted from the subducting slab in preference to insoluble ones such as Nb and Zr. However, the effect on the composition of the slab is small, as 75% or more of the inventories of soluble elements survive into the deep mantle. Indeed, a greater problem is that there is insufficient loss of soluble relative elements and that the subducting slab retains a crustal "flavor" with respect to some elements.

Pb/Ce and Nb/U ratios are very different between mantle and crust; furthermore, the mantle appears to have relatively uniform values of these ratios, with little difference between OIB and MORB (Hofmann et al. 1986, Newsom et al. 1986). For example, the mean Pb/Ce in OIB is 0.036 and in MORB is 0.038, whereas this ratio in the continental crust is 0.26 (e.g., Rudnick & Gao 2003). Similarly, Nb/U ratios of OIB and MORB are nearly identical—at 49 and 45, respectively—which are quite different from the crustal value of 6.2 (Rudnick & Gao 2003). Subduction of oceanic crust is invariably accompanied by subduction of marine sediment, which is primarily derived from the continents and carries a continental signature with respect to these ratios. For example, the global average of subducting sediment of Plank & Langmuir (1998) has Pb/Ce of 0.347 and Nb/U of 5.32. Thus subducted oceanic crust should have, in these and several other respects, something of a continental-crustal flavor. Porter & White (in press) found that whereas Pb and U are extracted in preference to Ce and Nb, the change in Pb/Ce and Nb/U ratios in the subducting slab is limited. The average Pb/Ce of subducting slabs decreases from

0.113 to 0.095 and Nb/U increases from 23 to 27, but these values remain well above and below, respectively, the values found in most OIB. There are, however, exceptions, as discussed below.

Niu & O'Hara's (2003) final criticism, although flawed, has some validity. They assume that subducting oceanic crust consists of only normal depleted MORB (N-MORB), when in fact it consists of both N-MORB and enriched MORB influenced by mantle plumes. Furthermore, they do not consider the sediment that is invariably subducted along with oceanic crust. Porter & White (in press) nevertheless found that modern subducting slabs, even with their sedimentary veneers, are not sufficiently light-rare-earth-enriched to generate the negative ε_{Nd} observed in some OIB. The mean Sm/Nd of subducting slabs in eight modern subduction zones they studied is 0.309, which increases to 0.314 as a result of subduction-zone processing. This corresponds to a $^{147}Sm/^{144}Nd$ ratio of 0.19, only slightly lower than the chondritic value (but much lower than the ratio of 0.226 assumed by Niu & O'Hara). Porter & White calculate that 1.8 Ga–old subducted oceanic crust and sediment would have a minimum ε_{Nd} of approximately +2.5, whereas ε_{Nd} in some OIB is as low as −6 (**Figure 8**).

On the other hand, one can make a strong case for crustal recycling in the Samoan plume. Jackson et al. (2007) report $^{87}Sr/^{86}Sr$ as high as 0.7204 and ε_{Nd} as low as −6 in Samoan lavas
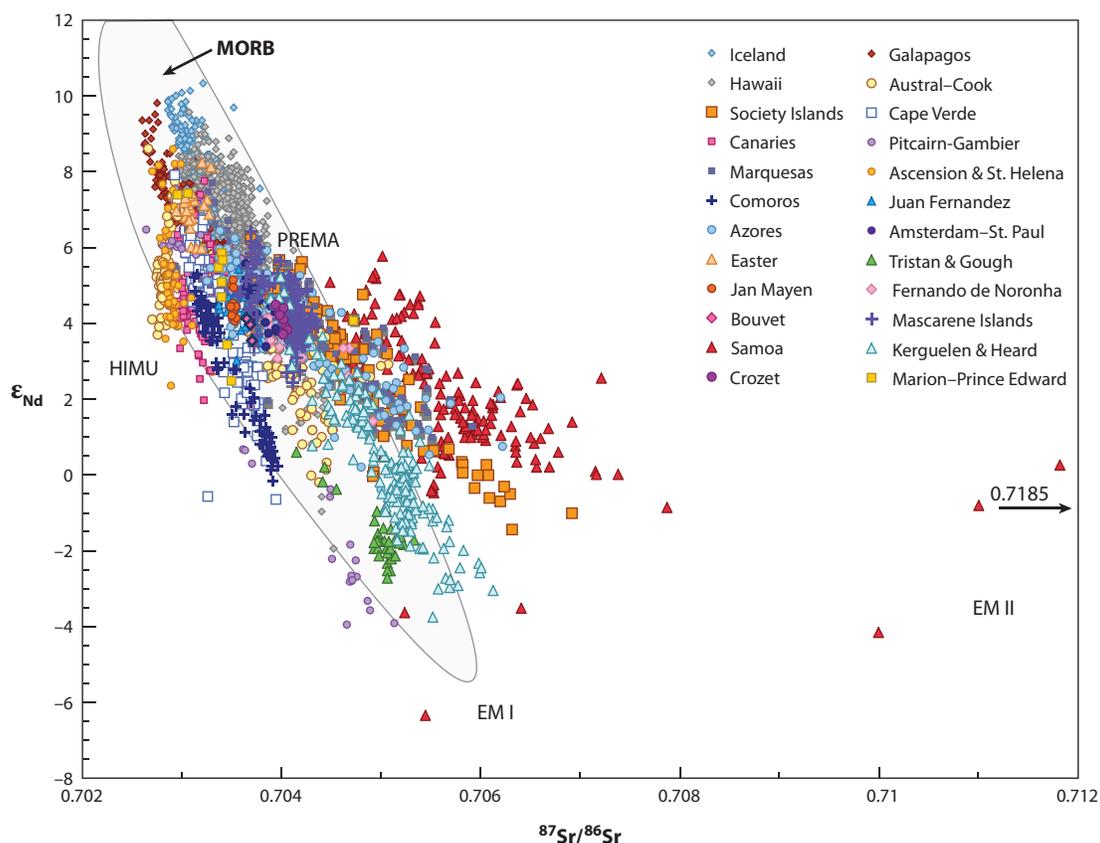

Figure 8

ε_{Nd} versus $^{87}Sr/^{86}Sr$ for OIB and MORB. Data are drawn from the GEOROC database (**http://georoc.mpch-mainz.gwdg.de/georoc/**). Field for MORB is based on data from the PetDB database (**http://www.petdb.org**). PREMA is Zindler & Hart's acronym for "Prevalent Mantle."

dredged from the underwater flanks of Savai'i. These isotopic compositions could not result from sediment assimilation because they are more extreme than any modern Pacific sediment. Furthermore, Pb/Ce ratios in these lavas are as high as 0.087 and Nb/U ratios are as low as 25. They argue on the basis of these observations that the Samoan plume contains a significant component of anciently subducted, continent-derived sediment, and it is difficult to imagine any alternative scenario. The Samoan plume appears to contain much more enriched, and perhaps unusually old, continent-derived sedimentary material than any modern subduction zone, judging from its extremely radiogenic isotopic compositions. Some basalts from the Society Islands also have high Pb/Ce and low Nb/U ratios, particularly those with high ^{87}Sr/^{86}Sr (Devey et al. 1990, White & Duncan 1996). It would seem, then, that plumes of the Society or EM II class (White 1985, Zindler & Hart 1986) do indeed contain a component of recycled oceanic crust and sediment.

Oxygen isotope ratios appear to confirm the presence of subducted surficial material in EM II plumes. Eiler et al. (1997) found that "EM2 lavas . . . are systematically outside of the range of normal upper mantle δ^{18}O values." Fractionation factors for oxygen isotopes are strongly temperature dependent, so that "igneous rocks whose oxygen isotopic compositions show significant variations from the primordial value ($\sim +6$) must either have been affected by low temperature processes, or must contain a component that was at one time at the surface of the Earth" (Taylor & Sheppard 1986). Eiler et al. (1997) conclude that their results are consistent with EM II plumes containing 2% to 6% sediment. Most other OIB did not show oxygen isotope ratios outside the normal mantle range. Notably, however, they found that lavas from HIMU plumes had generally low δ^{18}O, consistent with a source component that had reacted with water at elevated temperature, such as lower oceanic crust (e.g., Muehlenbachs & Clayton 1972). However, δ^{18}O in HIMU basalts is generally only slightly lower than the range of normal mantle, and Eiler et al. (1997) suggest that assimilation of lower oceanic crust might also provide an explanation.

In isotope space, OIB can be divided into several distinct groups (White 1985), which are generally referred to as PREMA, EM I, EM II, and HIMU (Zindler & Hart 1986), as illustrated in **Figure 8**. Weaver (1991) suggested that EM I, EM II, and HIMU basalts could be distinguished based on their trace-element geochemistry. However, Willbold & Stracke (2006) found that whereas HIMU basalts can indeed be distinguished from EM basalts, there are no systematic differences between EM I and EM II. They found that HIMU have lower Rb/Sr, Ba/La, K/La, Th/U, and Pb/Ce ratios, and higher U/Pb, Lu/Hf, and Nb/Rb ratios than EM basalts. The distinctions identified by Willbold & Stracke (2006) can be seen to some degree in **Figure 6**. Most oceanic basalts display negative Pb anomalies (the continental-crust and island-arc volcanics typically have positive Pb anomalies), but they are smaller for the EM basalts; the Samoan example actually has a slightly positive Pb anomaly. HIMU basalts, exemplified in **Figure 6** by the St. Helena sample, are enriched in Nb and Ta relative to K and Rb, and they have particularly strong depletion in Pb. The HIMU basalts also show a decrease in normalized abundance with increasing incompatibility for elements more incompatible than Nb.

Willbold & Stracke (2006) suggest that subduction of upper and lower continental crust can explain the isotopic signatures of EM II and EM I, respectively. Lower continental crust may return to the mantle through delamination in association with excessive thickening in orogeny (Kay & Kay 1993) or during rifting (Hanan et al. 2004). However, it may be difficult to separate lower continental crust from the underlying mantle lithosphere carrying a unique Os isotope signature that is, as noted above, not observed in OIB. Willbold & Stracke (2006) suggest instead that lower continental crust might be delivered to the mantle through subduction erosion. There is substantial evidence for such a process and evidence that the volume of continental material carried into the mantle in this way might exceed sediment subduction (e.g., von Huene & Scholl 1991, Clift & Vannucchi 2004).

Workman et al. (2004) proposed yet another variation on the recycling theme. They suggested that the unique geochemical signature of the EM II reservoir, and the Samoan plume in particular, results from the presence of subducted, metasomatized oceanic mantle lithosphere. Here again, the idea is that small-degree melts from the asthenosphere percolate upward into the oceanic mantle lithosphere where they freeze, enriching the lithosphere in incompatible elements. Workman et al. (2004) found that the best match occurs when slightly melt-depleted lithosphere is impregnated with 1.1% of a 0.5% melt of primitive mantle. However, the degree to which this actually matches the incompatible element geochemistry of EM II lavas has been questioned (Willbold & Stracke, 2006).

In summary, there is little consensus on the origin of the geochemical signatures of mantle plumes. A variety of ideas has been proposed, yet none seems entirely satisfactory. The common thread of all these ideas is, however, creation of heterogeneity in Earth's lithosphere and recycling of this material into the deep mantle through subduction and mantle convection.

LITHOLOGICAL HETEROGENEOUS MANTLE AND THE PROBLEM OF REJUVENESCENT VOLCANISM

The mantle was once viewed as homogenous peridotite, at least in terms of mineralogy and major-element chemistry. That view was perhaps a bit simplistic as populations of mantle xenoliths always contained a small but significant fraction of mafic materials such as eclogite and pyroxenite. Today, researchers recognize that such mafic materials are widespread, if not ubiquitous, and that they could influence isotopic and trace-element composition as well as play a critical role in basalt generation. Perhaps the first works to point out the potential geochemical significance of lithological heterogeneity were those of Hanson (1977) and Zindler et al. (1979). Allègre & Turcotte (1986) pointed out that continuous production and subduction of oceanic crust over billions of years was likely to produce a "marble cake" mantle with ubiquitous stringers of mafic material embedded in an peridotite matrix. Indeed, recent numerical simulations suggest that 97% of the mantle has been processed through the oceanic crust cycle (Huang & Davies 2007).

Over the past decade or so, the potential significance of lithological heterogeneity in mantle plumes (as well as in the MORB source) has been increasingly recognized (e.g., Hauri 1996). The melting of mafic material in the mantle has been investigated both experimentally (Hirschmann & Stolper 1996, Hirschmann et al. 2003, Pertermann & Hirschmann 2003, Kogiso et al. 2004, Pertermann et al. 2004) and from a theoretical thermodynamic perspective (Phipps Morgan 2001). Yet no clear consensus on its importance has been reached. Stracke et al. (1999) argued against the existence of garnet-pyroxenite or eclogite in the Hawaiian source on the basis of combined Hf-Nd-Th isotope and trace-element systematics. On the other hand, Reiners (2002) argued that

> . . . temporal-compositional trends in the form of decreasing incompatible elements and MgO, with increasing SiO_2, as the eruption proceeds . . . reflect sequential eruption of melt from increasing depths in mantle melting regions that are compositionally zoned because of different solidii and melt productivities of distinct peridotite and pyroxenite lithologic domains.

A particularly interesting, provocative, and consequently controversial idea was proposed by Sobolev et al. (2005), who argue on the basis of SiO_2 and Ni concentrations that the Hawaiian source is olivine-free. They suggest that eclogite, which originates as recycled oceanic crust, composes up to 30% of the Hawaiian plume. Because of its much lower solidus, the eclogite

melts at greater depth than peridotite. This melt then reacts with surrounding peridotite to form pyroxenite, which subsequently melts at shallower depth. Sobolev et al. (2007) argue that recycled oceanic crust in the form of eclogite stringers, and consequently this two-stage melting process, is pervasive; according to them, even MORB are composed of 10% to 30% pyroxenite-derived melts. The isotopic composition of the pyroxenite component, if it exists, is open to debate. Ito & Mahoney (2004, 2005), for example, assume that it is HIMU-like. On the other hand, Gurenko et al. (2009) found that in the Canaries the radiogenic Pb resides in the peridotite component whereas the pyroxenitic component has unradiogenic Pb and an enriched Sr and Nd isotopic signature.

Phipps Morgan & Morgan (1999) proposed that mantle plumes contain 1% to 4% mafic material that melts completely or nearly so as mantle plumes near the surface. The residual peridotite matrix then forms the shallow asthenosphere, which flows toward divergent plate boundaries where it melts to form MORB. Ito & Mahoney (2004, 2005) suggest that, in fact, OIB and MORB sources are identical and consist of 90% depleted peridotite and variable proportions of enriched peridotite and pyroxenite. They assume that the enriched peridotite has a lower solidus and melts more readily than the depleted peridotite and that the pyroxenite melts even more readily than the enriched peridotite. They show that when this mix of material melts in buoyantly rising plumes under thick lithosphere, it is primarily the enriched peridotite and pyroxenite that melt. This produces OIB of various EM and HIMU flavors. When the same mix of material melts to greater extents in the passively upwelling environment of mid-ocean ridges, melts of the depleted peridotite dominate and the basalts produced have the chemical and isotopic signature of MORB. The significance of this idea is that plume and MORB reservoirs would not require separate regions of the mantle, but rather would intimately coexist throughout the mantle.

Many OIB mantle sources do seem to contain an incompatible-element-depleted component. A test of the Ito & Mahoney model is to ask whether this depleted component is identical to the MORB source. The depleted component in the Galapagos does indeed appear to be identical to the MORB source, but it appears to be thermally entrained asthenosphere rather than intrinsic to the plume (Richards & Griffiths 1989, White et al. 1993). A depleted component also appears to be present in the Iceland plume; whether this is identical to the MORB source and intrinsic to the plume is a matter of debate (e.g., Chauvel & Hémond 2000, Hanan et al. 2000). Fitton et al. (2003) argue that although the compositions of the MORB source and the depleted component in the Iceland plume overlap in many respects, they can be distinguished based on Nb/Y and Zr/Y ratios.

In mid-plate oceanic islands such as Hawaii, a depleted component often manifests itself in posterosional or rejuvenescent magmas that erupt long after—sometimes several million years after—the shield-building stage has ended (e.g., Sterns 1946, MacDonald 1968). In addition to Hawaii, posterosional volcanism has also been identified in at least Samoa, the Society Islands, Madeira, the Canaries, the Louisville Seamounts, Mauritius, and most recently, Jasper Seamount (Konter et al. 2009). In general, these cases mimic the Hawaiian pattern in which the posterosional phase has a distinctly more depleted isotopic signature than the shield-building phase (e.g., Chen & Frey 1983), despite their usually being much more incompatible-element-enriched. The latter feature most likely results from the generally highly alkaline posterosional magmas being generated by much smaller extents of melting than shield stage lavas.

Because posterosional magmas are isotopically distinct, the question of the nature of their source arises. Chen & Frey (1983) suggested it was oceanic lithosphere; White & Duncan (1996) suggested it was deep mantle viscously entrained by the plume. Energy considerations suggest, however, that the hot plume, rather than cooler surrounding material, is the most likely source of
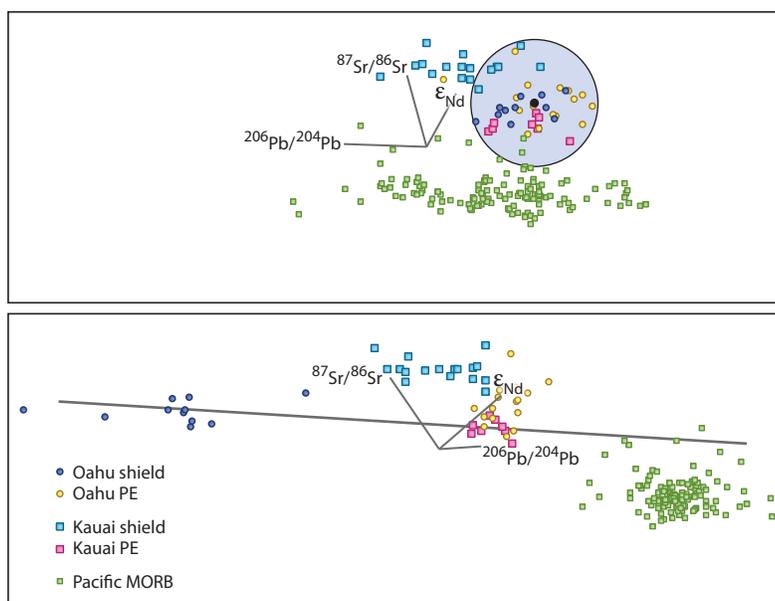
Figure 9

Two views of shield-stage and posterosional-stage data for Oahu and Kauai in three-dimensional
$^{87}Sr/^{86}Sr$-ε_{Nd}-$^{206}Pb/^{204}Pb$ space. The upper view is parallel to a line drawn from the Oahu shield data
through the Oahu posterosional data. The lower view is rotated perpendicular to the upper view. The line
does not pass through the Pacific MORB field. Abbreviation: PE, posterosional.

posterosional magmas. This would suggest that such plumes intrinsically contain both enriched
and depleted components that can melt separately.

Along these lines, Bianco et al. (2005) and Paul et al. (2005) proposed that the Hawaiian and
Reunion mantle plumes, respectively, are lithologically heterogeneous with a relatively
incompatible-element-enriched mafic component embedded within a depleted peridotite. Ac-
cording to these models, the pyroxenite melts more or less completely in the plume stem as it rises
beneath the growing shield volcano. Subsequently, owing to either lithospheric flexure or buoy-
ant lateral spreading of the plume, small-degree melts of the depleted peridotite are generated
and erupt as the posterosional lavas. Paul et al. (2005), however, demonstrated that the depleted
component in the Reunion plume is isotopically distinct from the MORB source. This same point
can be demonstrated for Hawaii. **Figure 9** shows two views of isotopic data from two Hawaiian
volcanoes and Pacific MORB in $^{87}Sr/^{86}Sr$-ε_{Nd}-$^{206}Pb/^{204}Pb$ space. The first shows the data viewed
parallel to the trend defined by the shield and posterosional data from Oahu; the second shows
the data perpendicular to that trend. It can be seen that the trend formed by these data does not
pass through the Pacific MORB field, nor does the trend formed by the Kauai data. Consequently,
the MORB source is not the depleted end-member in the source of posterosional volcanics of the
Hawaiian plume. The same point can be made in the same way for Samoa and the Society Islands.
Thus whereas mantle plumes do commonly contain depleted components, they are generally, if
not always, isotopically distinct from the MORB source. This argues against the idea of Ito &
Mahoney (2004, 2005) that OIB and MORB sources are identical. The MORB and plume reser-
voirs thus appear to occupy different regions of the mantle.

SUMMARY POINTS

1. The geochemistry of OIB provides strong support for Morgan's plume theory in several ways. First, OIB are usually, but not always, geochemically distinct from MORB. Second, olivine geothermometry of OIB suggests they are derived from mantle that is up to several hundred degrees hotter than that beneath ridges. Finally, (^{230}Th/^{238}U) and (^{231}Pa/^{235}U) suggest upwelling velocities of 2 to 6 cm year^{-1} within the melting region for low-buoyancy flux plumes and much higher velocities for Hawaii.

2. Geochemistry provides no clear evidence for where plumes originate. New experiments on the behavior of noble gases mean that He and other noble-gas isotope data do not require a deep primordial layer. Whereas ^{186}Os/^{188}Os from Hawaii, Gorgona, and Siberia may suggest a core component in some plumes, ^{186}Os/^{188}Os in most OIB do not depart from normal mantle values. This, and the lack of tungsten isotopic anomalies, suggests there is an alternative explanation for the anomalous ^{186}Os/^{188}Os.

3. The geochemical signature of OIB originated in the upper mantle and crust through melting. This material was subsequently transported into the deep mantle, probably by familiar plate-tectonic processes. Melting in the deep mantle would produce a very different signature that has not been observed. A convincing case can be made for sediment subduction and recycling in the Samoan plume. The detailed origin of the geochemical signature of other plumes is more problematic. Both the oceanic crust–recycling and subcontinental lithosphere–recycling hypotheses are not fully satisfactory in all respects.

4. Plumes are, in general, chemically heterogeneous and often contain both incompatible-element-enriched and depleted components. However, the depleted component of plumes is distinct from the MORB source, and may indeed be unique in each plume. It seems likely that mantle plumes are lithologically heterogeneous as well, with a small but significant mafic component in a dominantly peridotite matrix. Additional research must be done before the consequences of this lithological heterogeneity are fully understood.

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