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Reconciling the geophysical and geochemical mantles: Plume flows, heterogeneities, and disequilibrium

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[1] Geophysical evidence and numerical models of mantle stirring imply the source of mid-ocean ridge basalts (MORBs) comprises most of the mantle, excepting only the D″ region and the "superpile" anomalies deep under Africa and the Pacific. Geophysical evidence is also strong that the mantle is heated substantially from within. Geochemical inferences of a strongly depleted MORB source are inconsistent with this picture because they would require the MORB source to be heated mainly from below and because they cannot accommodate all of the Earth's incompatible elements. Lacking any other large mantle reservoir, the MORB source is required to balance the global uranium budget, which implies a U concentration of about 10 ng/g, more than double recent estimates. The MORB source would then have been depleted only by a factor of two in highly incompatible elements, rather than four or more, relative to is primitive composition. Both geophysical and geochemical evidence support a heterogeneous, multicomponent MORB source. Surprisingly, former plume material may comprise 25% of the MORB source, and this alone could add 50–100% to previous inventories of incompatible elements. Previous geochemical estimates may also be less secure because of a continuing focus on the more common, more depleted MORBs, because of long chains of geochemical inference, and because of a reliance on peridotites that may not have equilibrated with the mean composition of the heterogeneous source. Mean compositions are of most geophysical relevance, rather than putative end-member compositions, but mean compositions will be difficult to estimate accurately because more enriched components are less common and more variable. Nevertheless, a reconciliation of geochemical and geophysical inferences seems possible.

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1. Introduction

[2] Despite active and constructive dialog between the geophysical and geochemical communities over the past fifteen years or so, and despite some success in explaining the age and character of refractory incompatible trace elements and their isotopes with geophysical models [*Christensen and Hofmann*, 1994; *Davies*, 2002; *Huang and*

Davies, 2007c], there are still basic incompatibilities between the geophysical and geochemical pictures of the mantle. Whereas the geophysical evidence is strong that the mantle is heated substantially from within [*Davies and Richards*, 1992; *Davies*, 1999b], the geochemical inference is that the source of the mid-ocean ridge basalts (MORBs) is strongly depleted in heat-producing elements [*Hofmann*, 2003]. Whereas there is also strong, if not always recognized, geophysical evidence that there is no layering of the mantle apart from the thin D″ region [*Grand et al.*, 1997; *Lay et al.*, 1998; *Davies*, 1999b], geochemical inferences imply a substantial deep-mantle reservoir that is not strongly depleted in incompatible elements [*Hofmann*, 2003]. There is a third basic discrepancy, though it will not be directly addressed here: most geophysical modeling of the thermal evolution of the mantle [e.g., *Davies*, 1993] requires more heat production than is inferred from geochemical and cosmochemical evidence [*McDonough and Sun*, 1995]. Geophysical issues involved in this question are debated elsewhere [*Korenaga*, 2006; *Silver and Behn*, 2008; *Davies*, 2009].

[3] In this paper the geophysical evidence and arguments are reiterated and explained in some detail, so as to emphasize their robustness and the fundamental incompatibility of the resulting geophysical picture with the common geochemical picture. These arguments derive from seafloor topography as well as from seismology. It is then argued that resolution of the discrepancies may lie in a more thorough consideration of the implications of a heterogeneous, multicomponent MORB source. In particular the roles of less common but more enriched components may be important, and disequilibrium resulting from heterogeneity may affect inferences. Heterogeneities from plumes may comprise a substantial fraction of all heterogeneities in the MORB source.

2. Geophysical Constraints on Mantle Structure and Heating

[4] Seismology has always provided some of the more detailed information about the structure of the Earth's interior, yet seismology has its limits and other evidence can provide complementary information. Seismic tomography is a difficult art, and its detection of subducted lithosphere in the mantle required global instrumental coverage, the detection of subtle effects among many competing effects, substantial computer power and clean data sets. It took more than twenty five years from when the possibility was first conceived to the first widely acknowledge detections [*Grand*, 1994; *Grand et al.*, 1997; *Widiyantoro*, 1997]. This was the evidence that persuaded many people that subducted lithosphere penetrates deep into the lower mantle and that the lower mantle is therefore not a separate reservoir from the upper mantle. Yet a strong argument to that effect had already been provided some years earlier [*Davies*, 1988]. This argument was based on the recognition that the mantle is convecting, and that the buoyancies that drive convection can also raise or lower the Earth's surface. The Earth's surface topography can therefore constrain the kind of convection occurring in the mantle and, because convection transports heat, it can constrain the modes of heating and quantities of heat transported in the mantle. Because layering would strongly affect convective heat transport, topography also provides a strong test for layering.

2.1. Mantle Heating

[5] The total rate of heat loss from the mantle is around 35 TW (1 TW = 1 terawatt = 10^{12} watt), not counting radiogenic heat in the continental crust that is lost directly to the surface [*Davies*, 1999b]. Several geophysical estimates, discussed below, indicate that only a small fraction of the mantle heat enters the mantle at its base, from the core. Around 9 TW can be attributed to slow cooling of the mantle, although there is some debate about this number [*Davies*, 1999b]. The implication is that the balance must be generated within the mantle. Here the geophysical constraints on this picture are briefly reviewed.

[6] *Davies* [1988] pointed out that the size of hot spot swells constrains the amount of heat being carried by mantle plumes, and concluded that plumes carry only about 6% of the mantle's heat budget. *Sleep* [1990] independently reached the same insight, made a more comprehensive inventory of hot spot swells and also concluded they carry around 6% of mantle heat budget. The result applies strictly only to plume tails. Allowing for the estimated frequency with which plume heads have risen [*Hill et al.*, 1992] increases the estimate by around half. Thus plumes are inferred to carry less than 10% of the mantle heat budget, or around 3.5 TW.

[7] More recent work, summarized below, has qualified this estimate, but for the moment the principle of the argument will be reviewed, as illustrated in Figure 1. Figure 1a depicts a layered mantle, but the argument applies equally if the

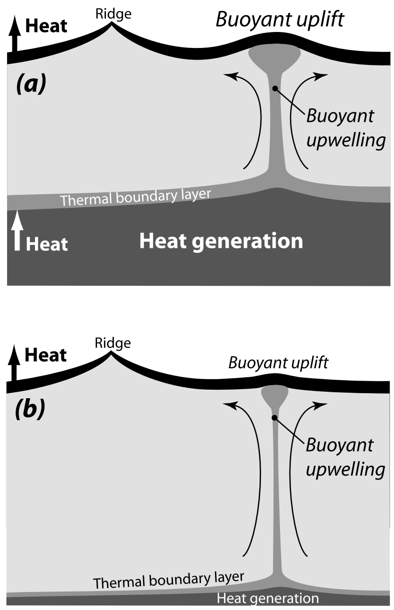

Figure 1. Sketches of relationship between heating and topography. (a) A thick lower layer supplies strong heating to the base of the upper layer, generating a strong buoyant upwelling and large topography. (b) A thin layer supplies only a small amount of heat, generating a weaker upwelling and small topography.

lower layer is the core. When convection is driven by heating from below, there are two thermal boundary layers, a cool one at the top (the lithosphere), and a hot one at the bottom. Buoyant upwellings will rise from the bottom boundary layer, and in the mantle the upwellings will have the form of plumes [*Davies*, 1999b]. When an upwelling reaches the top its buoyancy will lift the lithosphere, thus generating topography. The size of the topography is proportional to the amount of heat flowing up the plumes [*Davies*, 1999b], and therefore to the heat flowing from the lower thermal boundary layer into the plumes. If the mantle (or the mantle layer) were entirely heated from below, then plumes would carry 100% of the heat budget of the layer. The implication is then that hot spot swells would be comparable in amplitude and scale with the system of mid-ocean ridges.

[8] The last point may require further explanation. It follows from the fact that the top thermal boundary layer also generates topography. In this case the thermal boundary layer is generated by heat conducting out through the Earth's surface. This leaves a cool, strong layer (the lithosphere) that sinks (subducts) into the mantle, driving the ''plate mode'' of mantle convection [*Davies*, 1999b]. As a lithospheric plate moves away from a mid-ocean ridge, it cools and thermally contracts, which causes its surface to subside slowly. This is a form of negative topography (i.e., subsidence) caused by the negative buoyancy of the cool, dense plate. Our perception is that mid-ocean ridges stand high relative to old seafloor, but the physical process is really of old seafloor subsiding relative to its initial elevation (the elevation of zero-age seafloor, i.e., of mid-ocean ridges) [*Davis and Lister*, 1974; *Davies*, 1988, 1999b]. The mid-ocean ridges are thus the expression of the topography generated by the (negative) buoyancy of the top thermal boundary layer. The amplitude and extent of this topography is proportional to the heat flowing through the top thermal boundary layer, which is most of the mantle's heat budget.

[9] If the mantle were heated 100% from below, then the same amount of heat would flow through both boundary layers. Since the topography generated by each boundary layer is proportional to the heat flowing through it, the two forms of topography should be comparable.

[10] If it seems the two boundary layers together account for 200% of the mantle heat budget, the problem is resolved by recognizing that the two boundary layers play different and complementary roles. The bottom boundary layer is generated by heat flowing into the mantle, whereas the top boundary layer is generated by heat flowing out of the mantle. Thus plume-generated topography reflects heat entering the mantle and plate-generated topography reflects heat leaving the mantle.

[11] A glance at a map of the Earth's digitized topography shows that hot spot swells, the only topography identifiable as due to buoyant upwellings, are in no way comparable to the mid-ocean ridge system. This implies that the convecting mantle is heated only weakly from below. Therefore it must be heated substantially from within, presumably by radioactivity.

[12] Now returning to the heat carried by plumes, some careful modeling in recent years [*Labrosse*, 2002; *Bunge*, 2005; *Zhong*, 2006] has shown that the heat carried by plumes is greater in the deep mantle than in the shallow mantle. This is because the temperature excess of plumes is larger at depth. There are two contributions to this effect. One is that the ambient mantle has a slightly subadiabatic temperature gradient, as do the interiors of all convecting fluids. The other is that the gradient of an adiabat is proportional to its absolute temperature, so the hotter adiabat of a plume is steeper

than an ambient mantle adiabat. Combined, these effects yield a temperature difference between the plume and the surrounding mantle of 400–600°C at the bottom, versus 200–300°C at the top. Correspondingly, the excess heat transported by the model plumes in the deep mantle is about 2–3 times larger than in the shallow mantle. A realistically large viscosity difference between the plume and the ambient mantle is not yet possible in three-dimensional numerical models, because resolution is still limited: Labrosse's and Bunge's plumes are isoviscous, and Zhong's plumes are a factor of 10–30 less viscous, but realistic plumes would be 100–200 times less viscous [*Davies*, 2005]. There is a tendency in the models for larger viscosity variation to reduce the effect, so the actual variation of heat transport with depth is plausibly smaller than the models so far indicate. Here it will be taken to be a factor of 2, with an upper limit of a factor of 3. Thus plumes are inferred to be carrying 2–3 times 3.5 TW, i.e., 7–10 TW, from the lowermost mantle, though *Zhong* [2006] infers an upper limit of 14 TW.

[13] Recently the heat flow from the core has been constrained independently from observations of postperovskite transformations in the D″ region in the lowermost mantle. Because the pressure of the phase transformation is strongly temperature-dependent [*Hirose*, 2006], the geotherm in the lower thermal boundary layer may cross into and then out of the postperovskite stability field [*Hernlund et al.*, 2005]. There are indications from seismology of such a "double crossing," and they have been used to estimate a local heat flux of 35–85 mW/m^2, which would translate to a total heat flow of 7–17 TW if the temperature gradient were laterally uniform [*Lay et al.*, 2008]. However, these estimates are likely to be biased to higher values because the postperovskite phase may not occur at all where mantle temperatures are higher and the thermal gradient is lower. There are substantial uncertainties in the seismological observations and in the thermal conductivity, so this approach, while promising, does not yet give a strong constraint.

[14] Constraints on the heat flow from the core have also been deduced from the energy requirements of the geodynamo. These have been discussed recently by *Davies* [2007a] and *Lay et al.* [2008]. There are considerable uncertainties, particularly in the thermal conductivity in the core [*Davies*, 2007a] and in the energy dissipation associated with the dynamo. Lay et al. quote a range of 3–8 TW for the heat flow out of the core, but some earlier studies deduce higher values. Higher heat flows would imply rapid growth of the inner core [*Labrosse et al.*, 2001]. This in turn would require even higher heat flows in the past, because the dynamo is less efficient without inner core crystallization, and implausibly high core temperatures are then implied to have existed early in Earth history [*Buffett*, 2002]. To avoid this difficulty, it has been proposed that the core contains radioactive ^{40}K which would generate extra heat and prevent the core from cooling so rapidly [*Buffett*, 2002; *Nimmo et al.*, 2004]. However, the geochemical conditions required to sequester potassium into the core ought to have left other clear geochemical signatures that are not observed [*O'Neill and Palme*, 1998], so this possibility is implausible. *Davies* [2007a] presented thermal evolution models of the core in which all constraints were plausibly met and the present heat loss from the core is 5–7 TW.

2.2. Mantle Layering or Compositional Gradient

[15] Images from seismic tomography showing dramatically colored anomalies extending from subduction zones sometimes deep into the lower mantle required no explanation: as noted above, they provided strong evidence that subducted lithosphere penetrates deep into the lower mantle [*Grand*, 1994; *Grand et al.*, 1997; *Widiyantoro*, 1997]. The mass flow associated with subducting lithosphere, and the necessary counter flow back into the upper mantle, is sufficient to prevent the lower mantle from acquiring a distinct chemical (or thermal) signature from the upper mantle [*Davies*, 1999b].

[16] The topographic argument outlined above was also used by *Davies* [1988] to argue against the lower mantle being a separate reservoir of primitive composition, as earlier hypothesized [*Wasserburg and DePaolo*, 1979]. A primitive lower mantle would generate within it about two thirds of the mantle heat budget. This heat would have to conduct through the putative 660-km interface, would generate a hot thermal boundary layer at the base of the upper mantle, strong upper mantle plumes and thus topography of about two thirds the amplitude and extent of the mid-ocean ridge system. Later, higher-resolution numerical illustrations of this are given by *Davies* [1999b]. Topography on such a scale does not exist, as a glance at an image of the Earth's topography will

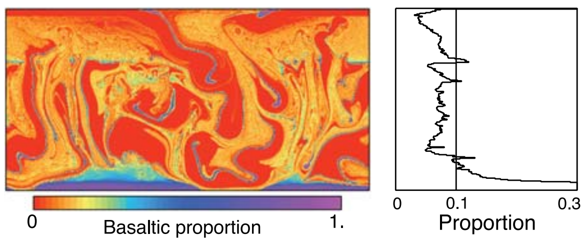

Figure 2. (left) Illustration of the accumulation of basaltic composition material at the base of the mantle, from an evolving numerical model [*Davies*, 2008]. (right) Profile of horizontally averaged proportion of basaltic component. Simulated oceanic crust (blue, Figure 2 (left)) forms from the melt zones near the top (red). Crust and depleted residual zones (red) subduct into the mantle where they are stirred by mantle convection. Some crust accumulates at the base because it is slightly denser. The accumulations form broad piles up to 800 km high in this snapshot, though their form is time-dependent. Although the proportion of basaltic composition rises steeply near the base (Figure 2, right), there is no significant large-scale vertical gradient through the main convecting part of the model.

confirm. This is a straightforward and robust argument, and it was corroborated by seismic tomography a decade later.

[17] The argument applies equally to the deep mantle layer proposed by *Kellogg et al.* [1999]. Since that layer would have to contain about half of the Earth's heat sources, it should generate plume-type topography about half the amplitude and extent of the mid-ocean ridges. Also *Tackley* [2002] argued that the undulating interface of the proposed layer should be easily detectable with seismic tomography.

[18] A variation on the deep-layer hypothesis is that the convecting mantle maintains vertical geochemical gradients within it, perhaps due to greater viscosity at depth [*Gurnis*, 1986]. However, more recent high-resolution numerical models of mantle stirring have shown that in the present mantle a substantial barrier is required for vertical chemical differences to accumulate [*van Keken and Ballentine*, 1998; *Davies*, 2007b]. (On the other hand some such dynamic stratification is plausible for the earlier, hotter mantle when denser compositional components could settle out of the lower-viscosity upper mantle [*Davies*, 2006, 2007b].)

[19] This is true even when there is a propensity to form a gradational transition into a D″-type layer, and has been confirmed in many models [*Christensen and Hofmann*, 1994; *Davies*, 2002, 2007b, 2008]. An example is shown in Figure 2 [*Davies*, 2008]. The vertical profile of horizontally averaged basaltic proportion shows no significant vertical gradient through most of the depth of the mantle, even though there is a strong gradient at the bottom where some of the tracers of basaltic composition accumulate. This is an important result from mantle stirring models, as it implies that the composition inferred from MORB compositions extends throughout the convecting mantle.

2.3. Summary

[20] The topographic constraint precludes a substantial fraction of the Earth's radioactive heat sources from being stored in a layer deeper than the interface from which mantle plumes rise. It also precludes a large heat flow from the core. The implied situation is depicted in Figure 1b. Hot spot swells are small in comparison with the mid-ocean ridge system. The plumes supporting the hot spot swells carry only a small fraction of the mantle heat budget. The D″ layer is thin, and generates only a small amount of heat, and only a modest amount of heat emerges from the core. The heat flows involved are quantified later in section 3.4.

3. Geochemical Implications of the Geophysical Mantle

[21] The geophysical evidence is strong that there is no barrier to flow at the 660 km seismic discontinuity, although there may be some temporary delaying of vertical flow. It is also strong that there is no other barrier to flow around 2000 km depth, as hypothesized by *Kellogg et al.* [1999]; seismological evidence may be equivocal, but the topographic constraint is strong. The only structures that are clearly resolved by seismology and that might have distinct composition are the D″ zone in the lowest 200–300 km of the mantle [*Lay et al.*, 1998] and the "superpiles" deep under Africa and the Pacific [*Ni et al.*, 2002; *Garnero*, 2004; *Simmons et al.*, 2007]. (The latter are often called "superplumes," but they bear no resemblance to plumes. They ought then to be called "anomalies" or, at most, since there is plausible evidence that they are slightly denser than normal mantle [*Simmons et al.*, 2007], "superpiles.") The implication is that the zone from which MORBs are drawn is the entire depth of the mantle except only for D″ and the superpiles.

[22] The MORB source zone is thus the main convecting mantle. Convection is sufficient to prevent significant vertical gradients from devel-

oping, as noted above, but this kind of convection is nevertheless not very efficient at stirring, so there may be considerable smaller-scale heterogeneity despite the convection [*Gurnis and Davies*, 1986; *Davies*, 2002; *Xie and Tackley*, 2004; *Huang and Davies*, 2007c]. Numerical models have by now been able to trace the stirring history through the age of the Earth [*Davies*, 2002; *Xie and Tackley*, 2004; *Davies*, 2008].

3.1. Remaining Primitive Fraction and Major Element Heterogeneity

[23] Mid-ocean ridges are the sites of mantle melting and the segregation of melt to form the oceanic crust, leaving a depleted zone below. The depth of main melting (i.e., the depth to the dry solidus) is about 60 km, but minor melting or melting of heterogeneities might occur as deep as 110 km [*Yasuda et al.*, 1994; *Spandler et al.*, 2008]. The present areal rate of seafloor spreading is 3 km^2/yr [*Parsons*, 1982]. The rate at which mantle mass is being processed through the MOR melting zone can be calculated as

$$\phi = \rho \mathrm{A_s d_m} \tag{1}$$

where ρ is the density of the upper mantle, A_s is the areal spreading rate and d_m is the melting depth. The time it would take to process one mantle mass, M, at this rate is then

$$\tau = \mathrm{M}/\phi \tag{2}$$

This can be called the mantle processing time [*Davies*, 2002]. With the above values, a density of 3300 kg/m^3 and a mantle mass of 4×10^{24} kg, τ = 4 Gyr.

[24] Because of higher radioactive heating in the past, the mantle would have been hotter, would have had a lower viscosity and so would have overturned faster. *Davies* [2002] estimated that, at present rates of overturn, it would have taken about 18 Gyr to accomplish the number of overturns that have actually occurred within 4.5 Gyr. A "model time," t_m, was defined to run in proportion to the rate of overturn. Davies also showed that the proportion, p, of the mantle that remains unprocessed declines exponentially with t_m:

$$\mathrm{p} = \mathrm{e}^{-\mathrm{tm}/\tau} \tag{3}$$

After t_m = 18 Gyr with τ = 4 Gyr, p = 0.011, so by this estimate only about 1% of the mantle remains primitive. This fraction would be even smaller if the melting depth was greater in the past, because the mantle was hotter. Thus we cannot expect a significant amount of primitive mantle to have survived in the MORB source.

[25] By the same logic the proportion of subducted crust in the mantle, f, will exponentially approach a maximum $f_m = \phi_s/\phi$, where ϕ_s is the rate of subduction of oceanic crust:

$$\phi_\mathrm{s} = \rho_\mathrm{c} \mathrm{A_s d_c} \tag{4}$$

so

$$\mathrm{f} = \mathrm{f_m}\left(1 - \mathrm{e}^{-\mathrm{tm}/\tau}\right) \tag{5}$$

where ρ_c is the density of oceanic crust and d_c is its thickness. After 18 Gyr of model time f will be 99% of f_m, and with a crustal density of 2900 kg/m^3 and thickness of 7 km this gives f = 0.06. In other words the mantle should contain about 6% of subducted, unprocessed oceanic crust. As we will see in section 6.2, not all crust melted under ridges may have been removed from the mantle, so the total amount of crust-derived material may be about double this.

[26] This result quantifies a widespread expectation that, after several billion years of subducting oceanic crust plus underlying depleted mantle, the mantle will have substantial major element heterogeneity, with accompanying trace element heterogeneity.

3.2. Heterogeneity and Apparent Age of Refractory Incompatible Trace Elements

[27] MORBs have been known for some time to display significant isotopic heterogeneity, and ocean island basalts (OIBs) to display somewhat greater heterogeneity [e.g., *Hofmann*, 2003]. However, it is not true that MORBs are remarkably uniform, as used commonly to be said; this is also stressed by *Hofmann* [2003]. MORBs display about one third to one half the range of refractory element isotopic heterogeneity of OIBs, and most of the trends evident in OIBs are evident in muted form in MORBs. OIBs also have higher concentrations of refractory incompatible trace elements, and *Hofmann and White* [1982] argued that this was due at least in part to higher concentrations in OIB sources. They argued further that this could be explained if subducted oceanic crust accumulated in the D$''$ region at the base of the mantle and was subsequently brought to the shallow mantle by mantle plumes. A third key feature of the isotopic

Table 1. Uranium and Heat Generation Budgets[a]

Reservoir	Mass (10^{22} kg)	Mass (%)	U Concentration (ng/g)	U Mass (10^{15} kg)	Heat Generation (pW/kg)	Heat Generation (TW)
Average silicate Earth	400	100	20	80	5	20
Continental crust	2.6	0.65	1400 (900–1800)	36 (23–47)	350 (225–450)	9 (6–12)
D″	8.5	2.1	50 (50–80)	4 (4–7)	10.5 (10–15)	1
MORB source, i.e., rest of mantle	389	97	**10** (6.5–13)	**40** (26–53)	2.5 (1.5–3)	10 (7–13)

[a] Plausible ranges in parentheses. Bold numbers inferred from U mass balance. Heat generation productivity from *Stacey* [1992], assuming Th/U = 3.8 and K/U = 13,000. One pW = 1 picowatt = 10^{-12} W.

observations is that both MORB and OIB data form a quasi-linear array in a plot of ^{207}Pb/^{204}Pb versus ^{206}Pb/^{204}Pb and the slope of this array can be interpreted as an apparent age of about 1.8 Ga [*Chase*, 1981]. The implication of the latter observation is that the isotopic heterogeneities survive in the convecting mantle for around 2 Gyr.

[28] By now it has been shown that these observations can be explained in the kind of mantle outlined in section 3. *Christensen and Hofmann* [1994] showed the way by modeling subducted crust and depleted mantle with tracers, showing that some crustal material did accumulate at the bottom and that residence times would yield apparent ages similar to the lead observations, though their model was run for only 3.6 Gyr. They also obtained a range of isotopic heterogeneity comparable to observations, though *Davies* [2002] argued this depended on poorly constrained scales of tracer averaging and melt mixing. Subsequent modeling extended this approach to the full mantle Rayleigh number and full age of the Earth [*Davies*, 2002], to three dimensions [*Huang and Davies*, 2007a, 2007b, 2007c] and to fully evolutionary models [*Xie and Tackley*, 2004; *Davies*, 2008]. Important results are that residence times and apparent ages are controlled primarily by the processing time scale [*Huang and Davies*, 2007b], that plausible values of the processing time scale yield results compatible with the apparent lead ages of the mantle, and the D″ accumulation does not form readily under present mantle conditions, though it persists from earlier, hotter mantle conditions [*Davies*, 2008].

[29] Another significant point is that in these models the heterogeneities are removed mainly by remelting, rather than by being stirred down to a scale at which the material or melt is homogenized [*Kellogg and Turcotte*, 1990]. Removal by remelting is much less sensitive to factors such as the time dependence or dimensionality of the flow, whereas stirring and homogenization are strongly sensitive to such factors [*Christensen*, 1989; *van Keken and Zhong*, 1999].

3.3. Uranium Budget

[30] If the mantle is internally heated and unlayered, apart from D,″ then there is nowhere in the mantle to locate the radioactive heat sources other than in the MORB source, which comprises most of the mantle. This implication has been evident for some time, but there has been no obvious reason why the geochemical estimates of the MORB source composition might be inaccurate. Some such reasons will be developed below, but for now the implications of the geophysical constraints will be pursued further. The implications will be illustrated for uranium, which will serve as a representative of the heat-producing elements, and also of the refractory incompatible elements.

[31] The average uranium content of the silicate parts of the Earth is estimated to be 20 ± 4 ng/g [*McDonough and Sun*, 1995; *O'Neill and Palme*, 1998; *Lyubetskaya and Korenaga*, 2007]. (Concentrations are given as nanograms of uranium per gram of rock, rather than as "parts per billion." This removes the ambiguity of whether the ratio is by weight, volume or mole.) Table 1 lists the reservoir masses, concentrations and uranium masses for the reservoirs to be discussed in this section. Thus a total of 80 × 10^{15} kg of uranium in the Earth is implied by the average silicate concentration.

[32] A significant fraction of this uranium is now in the continental crust. Because the continental crust is very heterogeneous, estimates of its average uranium content have varied from 0.9 to 1.8 μg/g, with *Rudnick and Fountain* [1995] preferring 1.4 μg/g. The latter value implies nearly half of the Earth's total uranium (36 × 10^{15} kg) is in the continental crust (Table 1).

[33] The D″ region at the bottom of the mantle will also contain significant uranium. The structure inferred from seismology makes it clear this region involves more than just a thermal boundary layer [*Lay et al.*, 1998]. Both compositional variation and a phase transformation [*Fei et al.*, 2004] seem to be involved. The *Hofmann and White* [1982] proposal that D″ contains an accumulation of old subducted oceanic crust is consistent with this interpretation, and numerical models of mantle evolution support it [*Davies*, 2008], as noted above.

[34] An example from those evolving models is shown in Figure 2. The bottom accumulation covers much of the base of the model, though it has been separated into two parts by subducting lithosphere. The larger part extends to a height of about 800 km above the base of the model, though the form and height are time-dependent. These features plausibly resemble not only D″ but the "superpile" anomalies in seismic velocity under Africa and the Pacific [*Ni et al.*, 2002; *Garnero*, 2004; *Simmons et al.*, 2007]. Although the piles in Figure 2 are hundreds of kilometers thick, much of the accumulation is only mildly enriched in the basaltic component, and analysis of this model shows that only the lowest 130 km contains more than 30% basaltic composition. The total amount of recycled basalt is then estimated to be no more than a layer of pure basaltic component 100 km thick, and this may be taken as a reasonable upper bound. The average uranium content of oceanic crust is about 50–80 ng/g [*Hofmann*, 1988; *Sun and McDonough*, 1988; *Donnelly et al.*, 2004]. The implied amount of uranium in D″ is then only about 5% of the Earth's total complement (Table 1).

[35] The amount of uranium in the continental crust plus D″ is thus about 40×10^{15} kg (Table 1), just half of the silicate Earth's total. If we follow the geophysical logic that the balance of this uranium is in the MORB source, comprising the non-D″ mantle, then its concentration must be about 10 ng/g. Taking the various uncertainties into account, as given in Table 1, the value might fall between 6.5 and 13 ng/g. Previous geochemical estimates of the U content of the MORB source are around 3–4.7 ng/g [*Jochum et al.*, 1983; *Salters and Stracke*, 2004; *Workman and Hart*, 2005]. Thus the present representative value of 10 ng/g is 2–3 times larger than previous preferred estimates. The MORB source is depleted by only a factor of 2 relative to the primitive mantle composition, taken to be the average of the silicate Earth, rather than by a factor of 4–7.

[36] This amount of uranium in the MORB source implies a significant amount of material derived from subducted oceanic crust may be distributed through it, assuming that the more refractory peridotite contains little uranium. If, following Table 1, subducted oceanic crust has a uranium content of 50 ng/g and, then 20% of the mantle would have to be composed of this component in order to obtain a mean mantle concentration of 10 ng/g. If the oceanic crust concentration is 80 ng/g (Table 1), then subducted crust would comprise 12.5% of the mantle. The latter is close to the amount of crust (directly subducted plus internally recirculated) suggested earlier.

3.4. Heat Budget

[37] The radioactive heat production implied by the uranium budget is included in Table 1. Heat generated in the continental crust is conducted directly to the surface. Of the heat generated within the mantle, about 90% is generated in the MORB source, the balance coming from D.″

[38] A full consideration of the mantle's heat budget must consider heat coming from the core and the slow cooling of the mantle. Heat from the core is expected to be carried by plumes, but plumes will also carry the 1 TW of heat generated in D.″ However, not all the heat that starts to ascend in plumes reaches the upper mantle, as was discussed in section 3. According to that discussion, the heat transported by plumes in the upper mantle is about 3.5 TW, and the heat transported in the deep mantle would be about 7 TW, with an upper limit of about 10 TW. This would comprise about 1 TW from D″ (Table 1), with the remaining 6 TW coming from the core. Plausible thermal evolutions of the core are compatible with this rate of heat loss, as noted earlier [*Davies*, 2007a].

[39] We must also consider that the mantle is cooling. Cooling releases stored heat, and its relationship to the thermal boundary layers is similar to that of internal heating: the released heat leaves through the top thermal boundary layer, which is stronger as a result, but it does not affect the heat entering through the bottom thermal boundary layer.

[40] The total rate of heat loss from the Earth is about 44 TW [*Sclater et al.*, 1980]. If 9 TW of this is lost directly from the continental crust (Table 1),

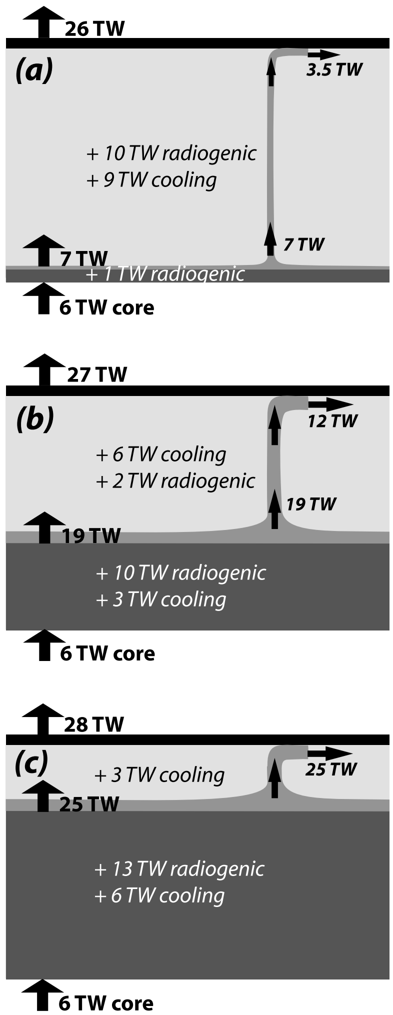

Figure 3. Heat budgets of various mantle models. (a) Preferred model: thin D″ region and a mildly depleted MORB source comprising most of the mantle. (b) Deep mantle layer containing about half of the Earth's refractory incompatible elements. (c) The old layered mantle: depleted upper mantle and primitive lower mantle. Only Figure 3a is consistent with observed hot spot swells. See section 3.4 for details.

then the heat loss from the mantle is about 35 TW. So far we have 6 TW from the core, 1 TW from D″ and 10 TW generated in the MORB source, totaling 17 TW. The difference is the net rate of heat loss from the mantle as it cools: 35 − 17 = 18 TW. Taking a specific heat of 1000 J/kg°C, this gives a rate of cooling of the mantle of 140°C/Gyr. This is rather higher than most estimates from thermal evolution calculations, which yield around 50–80°C/Gyr [*Davies*, 1993]. The difference could imply that the Earth is in a transient phase of faster cooling [*Silver and Behn*, 2008]. Such a transient could last several hundred million years without unduly affecting the mantle temperature in observable ways, but it is difficult to constrain. *Korenaga* [2006], on the other hand, argues the higher cooling rate is a long-term phenomenon, which raises questions about the compatibility of a hot mantle with Proterozoic and Archean tectonics. However, Korenaga's thermal evolution is based on some particular assumptions, especially concerning the radius of curvature of subducting plates. When assumptions more representative of observations are used, a more conventional thermal evolution is obtained [*Davies*, 2009].

[41] The heat budgets implied by the various versions of mantle layering are compared in Figure 3. They are based on a total radiogenic heat production of 20 TW (Table 1). Mantle cooling will be taken to contribute 9 TW, corresponding to a cooling rate of 70°C/Gyr, for the purpose of this illustration. Stronger cooling would only strengthen the point of these comparisons.

[42] The preferred mantle model is sketched in Figure 3a. Six TW is presumed to come from the core, and 1 TW of radiogenic heat is added to this in D.″ Plumes rise from the top of D,″ and therefore carry 7 TW of heat. This reduces to 3.5 TW as the plume arrives in the upper mantle, thus fitting the upper mantle heat flow inferred from hot spot swells [*Davies*, 1988; *Sleep*, 1990]. It is important to understand that only a very small fraction of this heat leaks out through the lithosphere, and the rest is delivered to the mantle interior; plumes are a result of heat entering the mantle. The other half of the initial plume heat is lost to the mantle interior on the way up. Thus the 7 TW from the core and D″ are delivered *into* the mantle. To this are added 10 TW of radiogenic heat in the mantle (Table 1) and 9 TW from cooling of the mantle, making a total of 26 TW. This amount of heat is removed from the mantle by the action of plate tectonics (the top boundary layer removes heat from the interior). In this model the MORB source is heated only 27% from below.

[43] The effect of the deep, thick layer proposed by *Kellogg et al.* [1999] is illustrated in Figure 3b. This layer was hypothesized to contain most of the incompatible elements that are not in the continental crust, so it would generate about 10 TW of radiogenic heat. It comprises about a third of the mantle mass, so it would contribute about 3 TW from cooling. Thus a total of 19 TW would emerge through the top interface of this layer. This is where plumes would originate, so they would carry 19 TW initially. They would lose perhaps one third of their heat rising through the rest of the mantle,

so delivering about 12 TW to the upper mantle. This would generate topography 3–4 times larger than the observed hot spot topography. The upper two thirds of the mantle would yield 6 TW from cooling and perhaps 2 TW of radiogenic heat, reflecting the strong depletion assumed for the MORB source. The plates would then remove the total of 27 TW from the mantle. In this model the MORB source is heated 70% from below.

[44] The old two-layered mantle is illustrated in Figure 3c. If the lower mantle were primitive it would account for about two thirds of the Earth's heat generation, and so 25 TW would be fed into upper mantle plumes, which would lose little rising through the upper mantle. They should therefore generate very large and obvious topography, seven times larger than the observed hot spot swells and not much smaller than the mid-ocean ridge system. In this model the MORB source is heated 89% from below.

[45] Thus the observed topography of the seafloor precludes a deep layer containing a substantial fraction of the Earth's heat sources, quite apart from seismological observations, which also do not favor mantle layering. This requires the mantle's heat sources to be mainly in the MORB source.

4. De Facto Geochemical Mantle

[46] Geochemical interpretations of the mantle have evolved along a quite different path. In the mid-1970s neodymium isotopes were regarded as an exciting new complement to strontium and lead isotopes because the primitive value of $^{143}Nd/^{144}Nd$ was known from meteorites. The most influential of the early papers on mantle-derived neodymium [*DePaolo and Wasserburg*, 1976] reported about a dozen values, many of which were higher than primitive, indicating incompatible element depletion of the mantle source. A couple of values were close to the primitive value, suggesting a primitive source remained in the mantle. Quite soon those values, which were from continental locations, were recognized as likely to be contaminated by continental crust, but the idea of a primitive mantle source had taken root.

[47] At that time the dominant geophysical view was that mantle convection was layered, with a boundary at the 660 km seismic discontinuity. The reasons were plausible but not compelling, and quite soon were challenged [*Davies*, 1977; *O'Connell*, 1977]. The idea of whole mantle convection began to compete with the dominant view, to be subsequently vindicated as described earlier. Nevertheless the neodymium data were interpreted, not unreasonably, in the context of a layered mantle and the ideas of a depleted upper mantle and a primitive lower mantle were quickly established [*Wasserburg and DePaolo*, 1979], despite their slender basis in observations. A correlation between neodymium and strontium isotopes was interpreted as a mixing line between material from the two reservoirs, and this gave rise to a third key idea, that the composition of the upper mantle is represented by the most depleted "end-member."

[48] By now the dominant geochemical view has been considerably modified. There is also, of course, significant debate among geochemists. Nevertheless the early two-layer interpretation and its corollary of a primitive lower mantle seems still to significantly channel geochemical thinking. As late as 1997 *Hofmann* [1997] still felt the need to emphasize the lack of evidence for primitive material, and the fact that key trace element ratios such as Nb/U are distinctly nonprimitive in OIBs and well as MORBs [*Hofmann et al.*, 1986; *Hofmann*, 2003]. Around the same time some interpretations of noble gases were still assuming the old two-layered mantle [*Kellogg and Wasserburg*, 1990; *Porcelli and Wasserburg*, 1995]. Workman and Hart were still doubting any depletion of the lower mantle in 2005 [*Workman and Hart*, 2005]. Meanwhile lead isotopes had been telling us for decades that all samples had been processed [*Davies*, 1984], though this seemed to gain little recognition.

[49] The early version of the layered mantle eventually yielded to the geophysical evidence for large mass flows through the 660 km seismic discontinuity cited earlier. A later version presumed the deep reservoir to be somewhat smaller, confined to the lower third of the mantle [*Kellogg et al.*, 1999], and to be enriched relative to the MORB source, though not necessarily primitive. The MORB source was still taken to be strongly depleted.

[50] Conventional mass balances of various incompatible elements have concluded that only around half of the mantle has been depleted of incompatible elements during the formation of the continental crust and atmosphere [*Allegre et al.*, 1996; *Hofmann*, 1997] and that, if the MORB source is strongly depleted, the balance of these elements must be sequestered somewhere deep in the mantle. However, the rigorous conclusion is that about half of the mantle's incompatible elements have been extracted from the mantle, and how the

remainder are distributed within the mantle is a separate question.

5. Heterogeneous MORB Source

[51] It has become progressively more recognized that the mantle must be heterogeneous because of continued subduction of heterogeneous lithosphere (depleted zone plus crust). It has become widely acknowledged that the mantle is indeed observed to be compositionally heterogeneous on many scales, in both major elements [*Zindler et al.*, 1984; *Niu and Batiza*, 1997; *Dosso et al.*, 1999] and trace elements [*Hofmann*, 1997, 2003]. Several issues need to be clarified.

[52] If the MORB source is heterogeneous, then the heterogeneity of MORBs presumably arises from that source, rather than from mixing of a depleted source with material from a putative enriched reservoir. In that case the composition of the MORB source is represented by the mean MORB composition, rather than by its most depleted end-member. A more comprehensive average MORB composition was apparently determined by *Su* [2002], but that source is an unpublished thesis [*Donnelly et al.*, 2004] that is now difficult to access.

[53] Even if heterogeneity is acknowledged, recent studies have still tended to focus on the more common, relatively depleted part of the MORB spectrum, i.e., on the mode of the distribution [*Salters and Stracke*, 2004; *Workman and Hart*, 2005]. However, as *Davies* [1984] noted long ago and *Hofmann* [2003] has recently reemphasized, there is no clear way to separate this "depleted" MORB from the broad tail of less common, less depleted MORB.

[54] From the point of view of reconciling the geochemical estimates of mantle heating with the geophysical, it is the total concentration of heat sources in the MORB source that is important, regardless of how they got there. Mass balances between the continental crust and the mantle also require the total concentration in the MORB source. Thus one should count the entire distribution, including the more enriched components, in determining the mean and the total. Determining the mean might then not be easy, because it may depend strongly on the less common, relatively enriched components. Furthermore, these may include not only subducted components but components added by plumes.

5.1. Plume Contributions

[55] The mass flow of plumes is less than that of subducted lithosphere, but only by about a factor of about three. The rate of subduction of seafloor is 3 km^2/yr [*Parsons*, 1982]. The strongly differentiated part of the lithosphere is about 60 km deep, so the volume of heterogeneous lithosphere added per year to the MORB source is around 180 km^3/yr. On the other hand the volumetric flow rate of the Hawaiian plume is estimated to be about 7.5 km^3/yr [*Davies*, 1992, 1999a] and the Hawaiian plume carries about 10% of the global plume flow, so the global volumetric flow rate of plumes into the upper mantle is about 75 km^3/yr, around 40% of the plate flow, or 30% of the combined flows into the MORB source.

[56] Most of the plume heterogeneity will not be removed by melting, but will be stirred into the mantle [*Davies*, 1999b]. The volumetric eruption rate of the Hawaiian plume has been about 0.03 km^3/yr over the past 25 Myr [*Clague and Dalrymple*, 1989]. Even allowing for substantial subsurface emplacement of magmas, this implies that magmas represent only around 1% of the volume of the plume. If these magmas were derived by 5% melting of the plume [*Hofmann and White*, 1982], then only 20% of the plume would have melted, and the remainder would be stirred unaltered into the mantle.

[57] Thus we should expect around one quarter of mantle heterogeneity to derive from plumes. This implies that MORBs with plume-like signatures should not be excluded from estimates of average MORB composition, even for MORBs formed close to known plumes.

[58] A crude indication of the effect of including most MORB values is shown in Table 2, which lists abundances of Nb, Th and U. The first row shows simple raw averages from the category "All mid-ocean ridges" from the PetDB database. The other rows show comparison values for two estimates of "normal MORB" and two enriched MORBs. U in the all-MORB average is 3–5 times larger than previous MORB estimates. It may be that U values are less reliable because U concentrations are so low, so Th and Nb values have been included. Typical MORB ratios are Th/U = 2.6 and Nb/U = 47 [*Hofmann*, 2003], so these are suitable proxies. All-MOR Th values are 2–3 times larger than previous estimates, while Nb values are 1.6–2.6 times greater. These factors may be minima, because plume-affected ridge segments

Table 2. Average MORB Compositions

	Nb (μg/g)	Th (ng/g)	U (ng/g)
"All MOR"[a]	5.95	400	240
N-MORB[b]	2.33	120	47
EPR average[c]	3.79	200	80
E-MORB[b]	8.3	600	180
MARK E-MORB[c]	16.05	1100	305

[a] Raw averages from PetDB, http://www.petdb.org/.
[b] *Sun and McDonough* [1988].
[c] *Donnelly et al.* [2004].

Table 3b. MORB Source Enrichments due to Plume Components[a]

Plume Enrichment	MORB Source Enrichment
2	1.25
3	1.5
5	2
10	3.25

[a] Enrichment factors are relative to "depleted MORB mantle" inferred from "normal MORB." Plume material is assumed to comprise one quarter of MORB source material (see section 5.1).

may have been excluded from the "all-MOR" category of PetDB, though this is not clear from the PetDB summaries.

[59] A potentially more robust estimate of mean MORB source composition, though with rather large uncertainties at present, comes from the composition of OIBs, inferred to be produced by plumes. Table 3a shows the enrichments of various classes of OIBs relative to MORBs for Nb, Th and U. Hawaiian OIB is enriched by a factor of 2–3, whereas the other OIBs are enriched by factors of 10–40, with a mean of around 20. The higher enrichments are probably due in part to relatively low melt fractions [*Hofmann*, 2003], so they probably do not reflect the mean enrichments of the associated plumes. Table 3b shows the effect of various plume enrichments on the mean composition of the MORB source. If plumes are enriched by factors of 3–5 relative to the "depleted MORB mantle," then the mean mantle concentrations of incompatible elements is increased by factors of 1.5–2. Greater enrichments cannot be ruled out at this stage.

[60] The accuracy of the latter estimates could be considerably improved by carefully estimating the melt fraction and enrichment of each plume and combining them with estimates of the melt volume (from erupted volumes) and the plume flow rate (from hot spot swells) [*Davies*, 1999b], following the approach used at the beginning of this section.

Table 3a. Enrichments of Plume Components[a]

	Nb	Th	U
Hawaii/N-MORB	2.2	2.9	2.4
EM1/N-MORB	13	24	15
EM2/N-MORB	11	42	24
HIMU/N-MORB	17	28	18

[a] Values taken from *Hofmann* [2003, Figure 20]. N-MORB: "normal MORB"; EM1: enriched mantle 1; EM2: enriched mantle 2; HIMU: high $^{238}U/^{204}Pb$ [*White*, 1985; *Zindler and Hart*, 1986].

5.2. Other Enriched Contributions

[61] Relatively enriched contributions may also come from subducted sediments and from the sources of E-MORBs. Both of these tend to be excluded from inventories by the penchant for focusing on the "depleted MORB mantle." The quantity of sediment affecting the deep mantle is inferred to be small [*Hofmann*, 2003], but its strong enrichment means it needs to be carefully considered. On the one hand, even 0.5% of sediment enriched by a factor of 100 could add 50% to inventories. On the other hand, this component may already be included in plume inventories, especially in the EM categories that are inferred by some to include a sedimentary component [*Hofmann*, 2003].

[62] Sediments or some other enriched component(s) seem to contribute to MORBs as well, as evidenced by the occurrence of enriched MORBs (E-MORBs). These occur even in regions far from any obvious plume influence. They may be enriched by factors of a few to 40 over "normal MORBs" (Table 2 [*Hémond et al.*, 2006]). These are, according to Hémond et al., much less common than N-MORBs but by no means rare. Their origin is debated. *Donnelly et al.* [2004] propose that they come from mantle that was metasomatized at a subduction zone by fluids expelled from subducted lithosphere. On the other hand Hémond et al. argue their source can be subducted alkali basalts. In either case a more enriched component from the Earth's surface is invoked.

5.3. How Well Determined Is the "Depleted MORB Mantle"?

[63] Without wishing to detract from the considerable efforts and ingenuity of geochemists, it must be acknowledged that geochemical estimates of mantle composition depend on quite long chains of inference. Indeed it is remarkable that the

abundance of virtually every element can be estimated. Nevertheless uncertainties must accumulate through such chains of inference. It may also be true that particular assumptions still significantly affect results. The focus on the more depleted end of the compositional spectrum (the mode) is one example of this. Another may be the use of melting models that assume a homogeneous source and melt-solid equilibrium, when there is abundant evidence that the mantle is heterogeneous in major element composition and there are good reasons to suspect extracted melt does not fully equilibrate with the mean composition or a peridotitic component, as will be elaborated below.

[64] For example, in the recent estimate of *Salters and Stracke* [2004], the estimate of thorium content depends on no less than eight elemental ratios and one correlation connecting it to the major element composition. U and K require a further ratio each. Three of those elemental ratios are inferred from isotopic ratios, and this requires estimates of the average time for which the isotopic ratio has been accumulating, which is not particularly well known. The estimates are anchored by a melting model that connects the major element composition of MORB with its source, but does this melting model apply for a heterogeneous source? Most of the trace element contents are connected to major elements by a correlation between Lu and CaO in anhydrous spinel peridotites. However, this relationship could be different in the original, heterogeneous (and hydrous?) source, which might contain enriched eclogitic or pyroxenitic heterogeneities with significantly higher Lu content. This would increase the estimate of almost all the trace elements in their approach.

[65] Another recent estimate by *Workman and Hart* [2005] is based on the trace element composition of clinopyroxenes from abyssal peridotites. A different series of inferences is required to infer an original source composition. These begin with a reconstruction of whole-rock composition, which requires partition coefficients and modal abundances, then a melting model to estimate premelt composition, then a series of trace element ratios. Some element ratios are inferred from isotope ratios, and the latter are taken from a MORB compilation from which MORBs close to plumes and subduction zones have been excluded. The authors acknowledge a possible geographical bias, which would favor more depleted compositions. The robustness of this chain of inference to assuming a heterogeneous source needs to be evaluated. In particular, if enriched heterogeneities have been melted out of peridotites before they reach the surface, for which *Salters and Dick* [2002] adduce indirect evidence, then the residual peridotites could fail to record the incompatible elements in those heterogeneities.

[66] Thus it may be that the "depleted MORB mantle" is more enriched than these estimates, and both enriched heterogeneities and the mean composition of the MORB source would then be proportionately more enriched.

6. Melt Extraction From a Heterogeneous Mantle

[67] Over the past decade petrological and geochemical studies have addressed the question of melting in a heterogeneous source, in which subducted oceanic crust, existing as eclogite or pyroxenite in the upper mantle, is distributed through a more refractory peridotitic matrix. On the one hand, they have yielded evidence that melt from heterogeneities does contribute to MORBs and OIBs. On the other hand, they have produced arguments that melt from the heterogeneities often may not equilibrate with the surrounding peridotite. The same arguments suggest that significant amounts of melt may not be erupted at mid-ocean ridges, and thus may be trapped in the mantle and recirculate internally.

6.1. Reaction and Disequilibrium of Eclogite Melts

[68] There has been a number of experimental petrological studies of heterogeneous systems [e.g., *Kogiso et al.*, 2004; *Sobolev et al.*, 2007; *Spandler et al.*, 2008, and references therein]. Subducted oceanic crust usually exists as eclogite in the upper mantle, and it melts at lower temperatures than the peridotite matrix in which it is embedded [*Yasuda et al.*, 1994]. It will therefore melt first as mantle rises under a mid-ocean ridge and may produce more melt as a result [*Campbell*, 1998]. This melt will be more silicic, and will react with any peridotite it comes into contact with, and is likely to refreeze, yielding refertilized, hybrid lherzolites or pyroxenites [*Yaxley and Green*, 1998; *Pertermann and Hirschmann*, 2003; *Sobolev et al.*, 2007].

[69] The reaction of eclogite-derived melts with peridotite might produce a range of compositions, from lherzolite through pyroxenite, garnet pyroxe-

nite even to eclogite, depending on the relative proportions of melt and peridotite. However, *Sobolev et al.* [2005] propose that the product will be a relatively uniform pyroxenite, with approximately 50% inputs each from the eclogite melt and the peridotite. For the sake of conciseness the term hybrid pyroxenite will be used here.

[70] There is by now considerable observational evidence that some melts from eclogitic or pyroxenitic sources reach the surface without fully equilibrating to the homogeneous composition. *Takahashi et al.* [1998] concluded that the Columbia River flood basalts are derived from shallow melting of a pyroxenite source in a mantle plume head. *Sobolev et al.* [2005] argue that unusually high Ni and Si contents of Hawaiian shield basalts are consistent with derivation from a secondary, olivine-free pyroxenitic source produced by melt from recycled oceanic crust hybridizing with peridotite in the Hawaiian plume [*Hofmann and White*, 1982]. Others have argued for some time that small near-ridge seamounts are produced by melting from heterogeneities, plausibly of recycled oceanic crust, that only pass through the edge of the subridge melting zone [*Zindler et al.*, 1984; *Niu and Batiza*, 1997]. *Salters and Dick* [2002] show that abyssal peridotites from the southwest Indian ridge cannot explain the neodymium isotopes of nearby basalts without invoking a more enriched component, plausibly pyroxenite or eclogite, that has been completely melted out of the residual peridotites.

[71] Osmium isotopes provide some of the strongest evidence for the survival in erupted basalts of unequilibrated signatures from eclogites [*Kogiso et al.*, 2004; *Sobolev et al.*, 2005]. Osmium isotopes correlate nearly linearly with Sr, Nd and Pb radiogenic isotopes, and the sublinear correlations have been interpreted as indicating mixing between liquids, rather than reaction between a liquid and a solid. This would imply that the eclogite-derived melts survive their passage through the peridotite matrix until they reach the peridotite melting zone, or even near-surface magma chambers.

[72] *Kogiso et al.* [2004] have considered in some detail the physical circumstances in which eclogite-derived melts might survive both the initial melting process and then the passage through the peridotite matrix, taking account of the size of the eclogite body, diffusion rates in solids and liquids, and whether the melt is saturated or undersaturated in silica. Undersaturated melts will dissolve orthopyroxene in the peridotite, and this will tend to maintain porosity and to promote the infiltration of the melt into the peridotite, and thus the complete reaction of the melt. On the other hand saturated melts will crystallize upon contact with peridotite, and this will tend to reduce porosity and inhibit further infiltration of the melt into surrounding peridotite. This solidified pyroxenite would also inhibit diffusive interaction between the peridotite matrix and the melt forming within eclogite bodies.

[73] For cases in which infiltration of the melt into the surrounding peridotite is inhibited, *Kogiso et al.* [2004] estimate that pyroxenite bodies thicker than a few meters would retain their chemical identity during melting episodes. They also estimate that in bodies thicker than 1–10 m the melt would segregate internally from its solid residue without equilibrating with the surrounding matrix by diffusion. Thus significant volumes of melt might accumulate within an eclogite body prior to migration into the peridotite matrix. During migration, *Kogiso et al.* [2004] appeal to the formation of a solidified reaction barrier to insulate much of the melt from the peridotite it ascends through. They also note that the significant volumes of melt they infer might be capable of forming porous flow channels with high melt/rock ratios, such as are observed in ophiolite complexes [*Kelemen et al.*, 1997; *Lundstrom et al.*, 2000], and this might explain the observation of radiogenic Os in such channels [*Becker et al.*, 2001]. Finally, Kogiso et al. appeal to the observations summarized above that some eclogite-derived melt survives its passage through solid peridotite until it mixes directly with peridotite melts at shallower levels. These conclusions would seem to apply broadly also to the sequence proposed by *Sobolev et al.* [2005, 2007].

6.2. Melt Trapping and Recirculation

[74] Although some melt from eclogites may reach the surface, or at least shallow magma chambers where some of their distinctive signatures may survive, an implication of *Kogiso et al.*'s [2004] study is that other melt may react with the peridotite matrix and refreeze. Material close to the spreading axis is likely to ascend close to the Earth's surface and therefore to remelt and to be extracted. Any melt from heterogeneities that ascends into the zone where the peridotite matrix begins to melt is also likely to be extracted, as the peridotite melt will form a connected network through which the melt from heterogeneities can migrate.

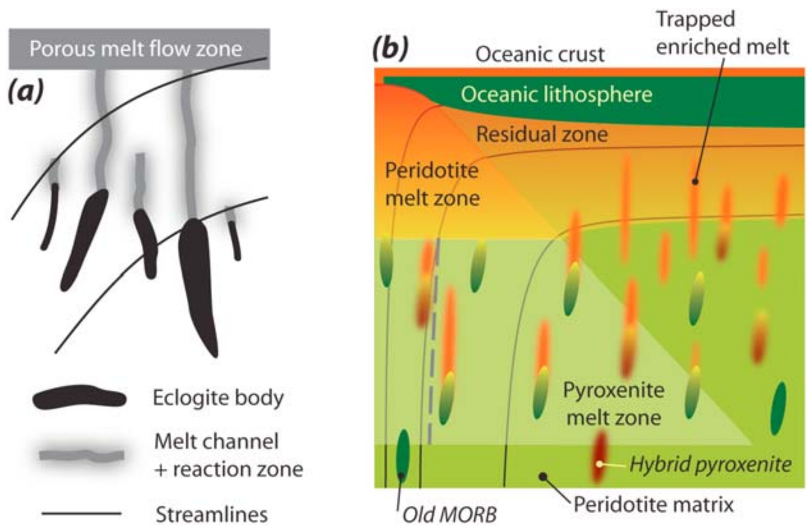

Figure 4. (a) Sketch of varied fates of melt from eclogite heterogeneities in a peridotite matrix. Melt from larger eclogite bodies may form channels that allow some of the melt to reach the peridotite melting zone, where porous melt flow predominates. Melt channels may require reaction zones to insulate them from further reaction with surrounding peridotite. Melt from smaller bodies may not travel far before refreezing and may be carried out of the melting zone by off-axis lateral flow under mid-ocean ridges. See section 6.2 for detailed discussion. (b) Sketch of heterogeneous melting under a ridge. Melting, shown in yellow, begins in eclogite bodies ("old MORB") well below the peridotite solidus. This melt rises (orange) from the larger bodies, some of it reacting with peridotite and refreezing, forming hybrid pyroxenite. Some eclogite melt rises into the peridotite melting zone and is extracted, thus contributing to oceanic crust. However, much of the melt passes to the sides of the extraction zone and remains in the mantle, trapping incompatible elements in the residual zone. Over time a population of hybrid pyroxenite heterogeneities will recycle and accumulate, and they also will contribute to melting. The heterogeneities are schematic and not to scale.

[75] However hybrid pyroxenite further from the spreading axis may not rise shallow enough to remelt, nor may it rise into the peridotite melting zone. This material may be carried laterally away without remelting. A sketch of how melt may be trapped is shown in Figure 4a. Melts from smaller eclogite bodies or silica-undersaturated melts are the most likely to be trapped in this way. The larger situation envisaged under spreading ridges is shown in Figure 4b. Much of the eclogite melting occurs outside the peridotite melting zone in disconnected pockets. Melt from larger eclogite bodies may migrate, but much of it may not reach the surface. Because the eclogite melting zone is deeper and wider than the peridotite melting zone, much of the eclogite melt, or its hybrid pyroxenite product, will pass outside the peridotite melting zone and may remain in the mantle. This effect will be enhanced if the peridotite matrix is more refractory than the "fertile peridotite" usually assumed in models of the melting of a homogeneous source. This is because the peridotite melting would begin shallower than 60 km and so be even less likely to capture the pyroxenite products.

[76] An implication of this picture is that substantial amounts of hybrid pyroxenite will be recirculated within the mantle, and over time some of it will return to MORB melting zones, potentially to be remelted. Thus there will be multiple generations of hybrid pyroxenite, and a significant population of it will accumulate, as foreshadowed in section 3.1. Thus also the material entering melting zones will have three main components, not two: peridotite residue, subducted oceanic crust and hybrid pyroxenite. Incompatible elements will be carried by both the subducted oceanic crust and the hybrid pyroxenite. The two types of heterogeneous inclusion are depicted in Figure 4b.

7. Conclusion

[77] Geophysical evidence weighs against significant layering or stratification of the mantle, and indicates the mantle is heated substantially from within. Together these constraints imply the MORB source comprises most of the mantle, and would require a mean mantle uranium content of about 10 ng/g, more than twice recent geochemical estimates. The mantle is expected to be heteroge-

neous on both geophysical and geochemical grounds. Heterogeneities are introduced into the MORB source mainly by subducting lithosphere, but around 25% come from plumes. Both subducted oceanic crust and plume material would form enriched heterogeneities dispersed through the MORB source. Some of these heterogeneities would be removed by melting under spreading centers, but some melt is likely to be trapped in the mantle and to recirculate as hybrid pyroxenite. Virtually all of the mantle will have been processed through ridge melt zones, so the mantle would be expected to comprise subducted oceanic crust (about 6%), plume material (about 25%), recirculating hybrid pyroxenite (possibly comparable to oceanic crust, so around 6%), and variably depleted residue from ridge melting (the balance, around 60%).

[78] Geochemical estimates of mantle composition have focused on the most common, relatively depleted material, and have taken key constraints from peridotites, from which enriched heterogeneities may have been melted out without fully equilibrating with the dominant peridotite. Melting models based on equilibrium melting of a homogeneous source may not accurately represent these situations. Plume-like material, or material adjacent to plumes or likely to include plume material, has been systematically excluded, but it may comprise as much as 25% of the MORB source. However, plume material is enriched in incompatible elements by anything from a factor of 3 to 40. Mean plume enrichment by a factor of 5 would be enough to double the mantle inventory of incompatible elements, and hence to satisfy the geophysical constraints on mantle heating. With contributions to the incompatible element inventory from subducted and dispersed oceanic crust and from recirculating hybrid pyroxenite, it would seem that the discrepancies between geophysical and geochemical inferences may be resolved.

[79] Geochemists have generally characterized the composition of relatively depleted mantle material, and then required a deep, less depleted reservoir to make up the balance of the Earth's incompatible refractory elements. In the picture developed here, the depleted and enriched materials do not comprise separate reservoirs, but rather are intermingled throughout the MORB source.

[80] If both subducted oceanic crust and hybrid pyroxenite contained 80 ng/g of U, then the inferred MORB source mean content of 10 ng/g would require 12.5% of the MORB source to be oceanic crust plus hybrid pyroxenite. This value is close to the above percentages deduced from dynamical arguments. The heterogeneities in plume-derived material derive ultimately mainly from subducted oceanic crust, in the present interpretation, so they would be included in these percentages.

[81] The isotopic characteristics of the refractory trace elements have already been shown to be compatible with geophysically based dynamical models [*Christensen and Hofmann*, 1994; *Davies*, 2002; *Xie and Tackley*, 2004; *Huang and Davies*, 2007c; *Davies*, 2008]. The differences between MORBs and OIBs are also explicable on the basis of those models, which yield an accumulation of subducted oceanic crust at the base of the mantle, plausibly forming the D″ region, as proposed by *Hofmann and White* [1982]. The picture developed here to account for mantle heating and refractory incompatible elements also leads to insights into the noble gases in the mantle, which are developed separately (G. F. Davies, Noble gases in the dynamic mantle, submitted to *Geochemistry, Geophysics, Geosystems*, 2009). Thus it may be possible to reconcile most of the geochemical observations with the geophysical picture of the mantle. The question of the apparent imbalance between the presently observed heat loss from the Earth and the heat generated by the cosmochemically inferred complement of U, Th and K is independent of the question of the distribution of heat sources, and is discussed elsewhere [*Davies*, 2009].

[82] A somewhat similar picture was developed by *Helffrich and Wood* [2001], based on seismological observations of small-scale heterogeneity in the mantle and on their expected abundance of subducted oceanic crust in the mantle assuming subduction has been occurring for several billion years. However, they did not address the inconsistency of their picture with conventional geochemistry, including the relationship between the high uranium abundance they inferred for the mantle (13 ng/g) and their value for the observed abundance in oceanic crust (65 ng/g). Nor did they note the topographic and other seismological constraints that more directly require higher heat generation in the MORB source.

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References

Allegre, C. J., et al. (1996), The argon constraints on mantle structure, *Geophys. Res. Lett.*, *23*, 3555–3557, doi:10.1029/96GL03373.

Becker, H., et al. (2001), Effects of melt percolation on the Re-Os systematics of peridotites from a Paleozoic convergent plate margin, *Earth Planet. Sci. Lett.*, *188*, 107–121, doi:10.1016/S0012-821X(01)00308-9.

Buffett, B. A. (2002), Estimates of heat flow in the deep mantle based on the power requirements for the geodynamo, *Geophys. Res. Lett.*, *29*(12), 1566, doi:10.1029/2001GL014649.

Bunge, H.-P. (2005), Low plume excess temperature and high core heat flux inferred from non-adiabatic geotherms in internally heated mantle circulation models, *Phys. Earth Planet. Inter.*, *153*, 3–10, doi:10.1016/j.pepi.2005.03.017.

Campbell, I. H. (1998), The mantle's chemical structure: Insights from the melting products of mantle plumes, in *The Earth's Mantle: Composition, Structure and Evolution*, edited by I. N. S. Jackson, pp. 259–310, Cambridge Univ. Press, Cambridge, U. K.

Chase, C. G. (1981), Oceanic island Pb: Two-stage histories and mantle evolution, *Earth Planet. Sci. Lett.*, *52*, 277–284, doi:10.1016/0012-821X(81)90182-5.

Christensen, U. R. (1989), Mixing by time-dependent convection, *Earth Planet. Sci. Lett.*, *95*, 382–394, doi:10.1016/0012-821X(89)90112-X.

Christensen, U. R., and A. W. Hofmann (1994), Segregation of subducted oceanic crust in the convecting mantle, *J. Geophys. Res.*, *99*, 19,867–19,884.

Clague, D. A., and G. B. Dalrymple (1989), Tectonics, geochronology and origin of the Hawaiian-Emperor volcanic chain, in *The Eastern Pacific Ocean and Hawaii*, edited by E. L. Winterer et al., pp. 188–217, Geol. Soc. of Am., Boulder, Colo.

Davies, G. F. (1977), Whole mantle convection and plate tectonics, *Geophys. J. R. Astron. Soc.*, *49*, 459–486.

Davies, G. F. (1984), Geophysical and isotopic constraints on mantle convection: An interim synthesis, *J. Geophys. Res.*, *89*, 6017–6040, doi:10.1029/JB089iB07p06017.

Davies, G. F. (1988), Ocean bathymetry and mantle convection: 1. Large-scale flow and hotspots, *J. Geophys. Res.*, *93*, 10,467–10,480, doi:10.1029/JB093iB09p10467.

Davies, G. F. (1992), Temporal variation of the Hawaiian plume flux, *Earth Planet. Sci. Lett.*, *113*, 277–286, doi:10.1016/0012-821X(92)90225-K.

Davies, G. F. (1993), Cooling the core and mantle by plume and plate flows, *Geophys. J. Int.*, *115*, 132–146, doi:10.1111/j.1365-246X.1993.tb05593.x.

Davies, G. F. (1999a), Geophysically constrained mantle mass flows and the ^{40}Ar budget: A degassed lower mantle?, *Earth Planet. Sci. Lett.*, *166*, 149–162, doi:10.1016/S0012-821X(99)00007-2.

Davies, G. F. (1999b), *Dynamic Earth: Plates, Plumes and Mantle Convection*, 460 pp., Cambridge Univ. Press, Cambridge, U. K.

Davies, G. F. (2002), Stirring geochemistry in mantle convection models with stiff plates and slabs, *Geochim. Cosmochim. Acta*, *66*, 3125–3142, doi:10.1016/S0016-7037(02)00915-8.

Davies, G. F. (2005), A case for mantle plumes, *Chin. Sci. Bull.*, *50*, 1541–1554, doi:10.1360/982005-918.

Davies, G. F. (2006), Gravitational depletion of the early Earth's upper mantle and the viability of early plate tectonics, *Earth Planet. Sci. Lett.*, *243*, 376–382, doi:10.1016/j.epsl.2006.01.053.

Davies, G. F. (2007a), Mantle regulation of core cooling: A geodynamo without core radioactivity?, *Phys. Earth Planet. Inter.*, *160*, 215–229, doi:10.1016/j.pepi.2006.11.001.

Davies, G. F. (2007b), Controls on density stratification in the early Earth, *Geochem. Geophys. Geosyst.*, *8*, Q04006, doi:10.1029/2006GC001414.

Davies, G. F. (2008), Episodic layering of the early mantle by the 'basalt barrier' mechanism, *Earth Planet. Sci. Lett.*, *275*, 382–392, doi:10.1016/j.epsl.2008.08.036.

Davies, G. F. (2009), Effect of plate bending on the Urey ratio and the thermal evolution of the mantle, *Earth Planet. Sci. Lett.*, doi:10.1016/j.epsl.2009.08.038, in press.

Davies, G. F., and M. A. Richards (1992), Mantle convection, *J. Geol.*, *100*, 151–206, doi:10.1086/629582.

Davis, E. E., and C. R. B. Lister (1974), Fundamentals of ridge crest topography, *Earth Planet. Sci. Lett.*, *21*, 405–413, doi:10.1016/0012-821X(74)90180-0.

DePaolo, D. J., and G. J. Wasserburg (1976), Inferences about mantle sources and mantle structure from variations of $^{143}Nd/^{144}Nd$, *Geophys. Res. Lett.*, *3*, 743–746, doi:10.1029/GL003i012p00743.

Donnelly, K. E., et al. (2004), Origin of enriched ocean ridge basalts and implications for mantle dynamics, *Earth Planet. Sci. Lett.*, *226*, 347–366, doi:10.1016/j.epsl.2004.07.019.

Dosso, L., et al. (1999), The age and distribution of mantle heterogeneity along the Mid-Atlantic Ridge (31–41°N), *Earth Planet. Sci. Lett.*, *170*, 269–286, doi:10.1016/S0012-821X(99)00109-0.

Fei, Y., et al. (2004), Experimentally determined postspinel transformation boundary in Mg_2SiO_4 using MgO as an internal pressure standard and its geophysical implications, *J. Geophys. Res.*, *109*, B02305, doi:10.1029/2003JB002562.

Garnero, E. J. (2004), A new paradigm for Earth's core-mantle boundary, *Science*, *304*, 834–836, doi:10.1126/science.1097849.

Grand, S. P. (1994), Mantle shear structure beneath the Americas and surrounding oceans, *J. Geophys. Res.*, *99*, 11,591–11,621.

Grand, S., et al. (1997), Global seismic tomography: A snapshot of convection in the Earth, *GSA Today*, *7*(4), 1–7.

Gurnis, M. (1986), The effects of chemical density differences on convective mixing in the Earth's mantle, *J. Geophys. Res.*, *91*, 11,407–11,419, doi:10.1029/JB091iB11p11407.

Gurnis, M., and G. F. Davies (1986), Mixing in numerical models of mantle convection incorporating plate kinematics, *J. Geophys. Res.*, *91*, 6375–6395, doi:10.1029/JB091iB06p06375.

Helffrich, G. R., and B. J. Wood (2001), The Earth's mantle, *Nature*, *412*, 501–507, doi:10.1038/35087500.

Hémond, C., et al. (2006), Origin of MORB enrichment and relative trace element compatibilities along the Mid-Atlantic Ridge between 10° and 24°N, *Geochem. Geophys. Geosyst.*, *7*, Q12010, doi:10.1029/2006GC001317.

Hernlund, J. W., et al. (2005), A doubling of the post-perovskite phase boundary and structure of the Earth's lowermost mantle, *Nature*, *434*, 882–886, doi:10.1038/nature03472.

Hill, R. I., et al. (1992), Mantle plumes and continental tectonics, *Science*, *256*, 186–193, doi:10.1126/science.256.5054.186.

Hirose, K. (2006), Postperovskite phase transition and its geophysical implications, *Rev. Geophys.*, *44*, RG3001, doi:10.1029/2005RG000186.

Hofmann, A. W. (1988), Chemical differentiation of the Earth: The relationship between mantle, continental crust, and ocea-

nic crust, *Earth Planet. Sci. Lett.*, *90*, 297–314, doi:10.1016/0012-821X(88)90132-X.
Hofmann, A. W. (1997), Mantle chemistry: The message from oceanic volcanism, *Nature*, *385*, 219–229, doi:10.1038/385219a0.
Hofmann, A. W. (2003), Sampling mantle heterogeneity through oceanic basalts: Isotopes and trace elements, in *Treatise on Geochemistry,* vol. 2, *The Mantle and Core*, edited by R. W. Carlson, pp. 61–101, Elsevier, Amsterdam.
Hofmann, A. W., and W. M. White (1982), Mantle plumes from ancient oceanic crust, *Earth Planet. Sci. Lett.*, *57*, 421–436, doi:10.1016/0012-821X(82)90161-3.
Hofmann, A. W., et al. (1986), Nb and Pb in oceanic basalts: New constraints on mantle evolution, *Earth Planet. Sci. Lett.*, *79*, 33–45, doi:10.1016/0012-821X(86)90038-5.
Huang, J., and G. F. Davies (2007a), Stirring in three-dimensional mantle convection models and implications for geochemistry: 2. Heavy tracers, *Geochem. Geophys. Geosyst.*, *8*, Q07004, doi:10.1029/2007GC001621.
Huang, J., and G. F. Davies (2007b), Stirring in three-dimensional mantle convection models and its implications for geochemistry: Passive tracers, *Geochem. Geophys. Geosyst.*, Q03017, doi:10.1029/2006GC001312.
Huang, J., and G. F. Davies (2007c), Geochemical processing in a three-dimensional regional spherical shell model of mantle convection, *Geochem. Geophys. Geosyst.*, *8*, Q11006, doi:10.1029/2007GC001625.
Jochum, K. P., et al. (1983), K, U and Th in mid-ocean ridge basalt glasses and heat production, K/U and K/Rb in the mantle, *Nature*, *306*, 431–436, doi:10.1038/306431a0.
Kelemen, P. B., et al. (1997), A review of melt migration processes in the adiabatically upwelling mantle beneath oceanic spreading ridges, *Philos. Trans. R. Soc. London, Ser. A*, *355*, 1–35, doi:10.1098/rsta.1997.0010.
Kellogg, L. H., and D. L. Turcotte (1990), Mixing and the distribution of heterogeneities in a chaotically convecting mantle, *J. Geophys. Res.*, *95*, 421–432, doi:10.1029/JB095iB01p00421.
Kellogg, L. H., and G. J. Wasserburg (1990), The role of plumes in mantle helium fluxes, *Earth Planet. Sci. Lett.*, *99*, 276–289, doi:10.1016/0012-821X(90)90116-F.
Kellogg, L. H., et al. (1999), Compositional stratification in the deep mantle, *Science*, *283*, 1881–1884, doi:10.1126/science.283.5409.1881.
Kogiso, T., et al. (2004), Length scales of mantle heterogeneities and their relationship to ocean island basalt geochemistry, *Geochim. Cosmochim. Acta*, *68*, 345–360, doi:10.1016/S0016-7037(03)00419-8.
Korenaga, J. (2006), Archean geodynamics and the thermal evolution of the Earth, in *Archean Geodynamics and Environments*, *Geophys. Monogr. Ser.*, vol. 164, edited by K. Benn, J.-C. Mareschal, and K. C. Condie, pp. 7–32, AGU, Washington, D. C.
Labrosse, S. (2002), Hotspots, mantle plumes and core heat loss, *Earth Planet. Sci. Lett.*, *199*, 147–156, doi:10.1016/S0012-821X(02)00537-X.
Labrosse, S., et al. (2001), The age of the inner core, *Earth Planet. Sci. Lett.*, *190*, 111–123, doi:10.1016/S0012-821X(01)00387-9.
Lay, T., et al. (1998), The core-mantle boundary layer and deep Earth dynamics, *Nature*, *392*, 461–468, doi:10.1038/33083.
Lay, T., et al. (2008), Core-mantle boundary heat flow, *Nat. Geosci.*, *1*, 25–32, doi:10.1038/ngeo.2007.44.
Lundstrom, C. C., et al. (2000), A geochemically consistent hypothesis for MORB generation, *Chem. Geol.*, *162*, 105–126, doi:10.1016/S0009-2541(99)00122-9.
Lyubetskaya, T., and J. Korenaga (2007), Chemical composition of Earth's primitive mantle and its variance: 1. Method and results, *J. Geophys. Res.*, *112*, B03211, doi:10.1029/2005JB004223.
McDonough, W. F., and S. Sun (1995), The composition of the Earth, *Chem. Geol.*, *120*, 223–253, doi:10.1016/0009-2541(94)00140-4.
Ni, S. D., et al. (2002), Sharp sides to the African superplume, *Science*, *296*, 1850–1852, doi:10.1126/science.1070698.
Nimmo, F., et al. (2004), The influence of potassium on core and geodynamo evolution, *Geophys. J. Int.*, *156*, 363–376, doi:10.1111/j.1365-246X.2003.02157.x.
Niu, Y., and R. Batiza (1997), Trace element evidence from seamounts for recycled oceanic crust in the eastern equatorial Pacific mantle, *Earth Planet. Sci. Lett.*, *148*, 471–483.
O'Connell, R. J. (1977), On the scale of mantle convection, *Tectonophysics*, *38*, 119–136, doi:10.1016/0040-1951(77)90203-7.
O'Neill, H. S. C., and H. Palme (1998), Composition of the silicate Earth: Implications for accretion and core formation, in *The Earth's Mantle: Composition, Structure and Evolution*, edited by I. N. S. Jackson, pp. 3–126, Cambridge Univ. Press, Cambridge, U. K.
Parsons, B. (1982), Causes and consequences of the relation between area and age of the ocean floor, *J. Geophys. Res.*, *87*, 289–302, doi:10.1029/JB087iB01p00289.
Pertermann, M., and M. M. Hirschmann (2003), Partial melting experiments on a MORB-like pyroxenite between 2 and 3 GPa: Constraints on the presence of pyroxenite in basalt source regions from solidus location and melting rate, *J. Geophys. Res.*, *108*(B2), 2125, doi:10.1029/2000JB000118.
Porcelli, D., and G. J. Wasserburg (1995), Mass transfer of helium, neon, argon and xenon through a steady-state upper mantle, *Geochim. Cosmochim. Acta*, *59*, 4921–4937, doi:10.1016/0016-7037(95)00336-3.
Rudnick, R. L., and D. M. Fountain (1995), Nature and composition of the continental crust: A lower crustal perspective, *Rev. Geophys.*, *33*, 267–309, doi:10.1029/95RG01302.
Salters, V. J. M., and H. J. B. Dick (2002), Mineralogy of the mid-ocean-ridge basalt source from neodymium isotopic composition of abyssal peridotites, *Nature*, *418*, 68–72, doi:10.1038/nature00798.
Salters, V. J. M., and A. Stracke (2004), Composition of the depleted mantle, *Geochem. Geophys. Geosyst.*, *5*, Q05B07, doi:10.1029/2003GC000597.
Sclater, J. G., et al. (1980), The heat flow through the oceanic and continental crust and the heat loss of the Earth, *Rev. Geophys.*, *18*, 269–312, doi:10.1029/RG018i001p00269.
Silver, P. G., and M. D. Behn (2008), Intermittent plate tectonics?, *Science*, *319*, 85–88, doi:10.1126/science.1148397.
Simmons, N. A., et al. (2007), Thermochemical structure and dynamics of the African superplume, *Geophys. Res. Lett.*, *34*, L02301, doi:10.1029/2006GL028009.
Sleep, N. H. (1990), Hotspots and mantle plumes: Some phenomenology, *J. Geophys. Res.*, *95*, 6715–6736, doi:10.1029/JB095iB05p06715.
Sobolev, A. V., et al. (2005), An olivine-free mantle source of Hawaiian shield basalts, *Nature*, *434*, 590–597, doi:10.1038/nature03411.
Sobolev, A. V., et al. (2007), The amount of recycled crust in sources of mantle-derived melts, *Science*, *316*, 412–417, doi:10.1126/science.1138113.
Spandler, C., et al. (2008), Phase relations and melting of anhydrous K-bearing eclogites from 1200 to 1600°C and 3 to 5 GPa, *J. Petrol.*, *49*, 771–795, doi:10.1093/petrology/egm039.

Stacey, F. D. (1992), *Physics of the Earth*, 3rd ed., 513 pp., Brookfield, Brisbane, Queensl., Australia.

Su, Y.-J. (2002), Database applications to the petrogenesis of ocean ridge basalts, Ph.D. thesis, Columbia Univ., New York.

Sun, S., and W. F. McDonough (1988), Chemical and isotopic characteristics of oceanic basalts: Implications for mantle composition and processes, in *Magmatism in Ocean Basins*, edited by A. D. Saunders and M. J. Norry, *Geol. Soc. Spec. Publ.*, *42*, 313–345.

Tackley, P. J. (2002), Strong heterogeneity caused by deep mantle layering, *Geochem. Geophys. Geosyst.*, *3*(4), 1024, doi:10.1029/2001GC000167.

Takahashi, E., et al. (1998), Origin of the Columbia River basalts: Melting model of a heterogeneous plume head, *Earth Planet. Sci. Lett.*, *162*, 63–80, doi:10.1016/S0012-821X(98)00157-5.

van Keken, P. E., and C. J. Ballentine (1998), Whole-mantle versus layered-mantle convection and the role of a high-viscosity lower mantle in terrestrial volatile evolution, *Earth Planet. Sci. Lett.*, *156*, 19–32, doi:10.1016/S0012-821X(98)00023-5.

van Keken, P. E., and S. Zhong (1999), Mixing in a 3D spherical model of present day mantle convection, *Earth Planet. Sci. Lett.*, *171*, 533–547, doi:10.1016/S0012-821X(99)00181-8.

Wasserburg, G. J., and D. J. DePaolo (1979), Models of Earth structure inferred from neodymium and strontium isotopic abundances, *Proc. Natl. Acad. Sci. U. S. A.*, *76*, 3594–3598, doi:10.1073/pnas.76.8.3594.

White, W. M. (1985), Sources of oceanic basalts: Radiogenic isotopes evidence, *Geology*, *13*, 115–118, doi:10.1130/0091-7613(1985)13<115:SOOBRI>2.0.CO;2.

Widiyantoro, S. (1997), Studies of seismic tomography on regional and global scale, Ph.D. thesis, Aust. Natl. Univ., Canberra, A.C.T., Australia.

Workman, R. K., and S. R. Hart (2005), Major and trace element composition of the depleted MORB mantle (DMM), *Earth Planet. Sci. Lett.*, *231*, 53–72, doi:10.1016/j.epsl.2004.12.005.

Xie, S., and P. J. Tackley (2004), Evolution of U-Pb and Sm-Nd systems in numerical models of mantle convection and plate tectonics, *J. Geophys. Res.*, *109*, B11204, doi:10.1029/2004JB003176.

Yasuda, A., et al. (1994), Melting phase relations of an anhydrous mid-ocean ridge basalt from 3 to 20 GPa: Implications for the behavior of subducted oceanic crust in the mantle, *J. Geophys. Res.*, *99*, 9401–9414, doi:10.1029/93JB03205.

Yaxley, G. M., and D. H. Green (1998), Reactions between eclogite and peridotite: Mantle refertilisation by subduction of oceanic crust, *Schweiz. Mineral. Petrogr. Mitt.*, *78*, 243–255.

Zhong, S. (2006), Constraints on thermochemical convection of the mantle from plume heat flux, plume excess temperature, and upper mantle temperature, *J. Geophys. Res.*, *111*, B04409, doi:10.1029/2005JB003972.

Zindler, A., and S. Hart (1986), Chemical geodynamics, *Annu. Rev. Earth Planet. Sci.*, *14*, 493–570, doi:10.1146/annurev.ea.14.050186.002425.

Zindler, A., et al. (1984), Isotope and trace element geochemistry of young Pacific seamounts: Implications for the scale of upper mantle heterogeneity, *Earth Planet. Sci. Lett.*, *70*, 175–195, doi:10.1016/0012-821X(84)90004-9.