

MANTLE MODELS

THE TWO-RESERVOIR MANTLE MODEL

The initial interpretation of isotopic variations in the mantle (ca. 1975) was a two-reservoir model: an upper depleted mantle overlying a lower mantle that was 'primitive', or possibly enriched in incompatible elements. The idea that the lower mantle was primitive gained favor with the acquisition of Nd isotope data. The first Nd data obtained showed that Nd and Sr isotope ratios in oceanic basalts were well correlated and that Nd isotope ratios fell between typical MORB values of about $\epsilon_{Nd} = +10$ and the primitive mantle value of $\epsilon_{Nd} = 0$. Mixing between these two reservoirs could explain most of the isotopic variation seen in mantle-derived rocks. This sort of model is illustrated in Figure 18.1.

There were, and are, good arguments why the depleted reservoir should overlie the primitive one. First, it is generally thought the depleted reservoir acquires its characteristics through loss of a partial melt to form the crust. Obviously this reservoir should then be nearer the continental crust. In addition, depleted peridotite is less dense than undepleted peridotite. Second, the depleted reservoir seems to be sampled wherever rifting occurs, not only at major mid-ocean ridges, but also at smaller rifts. For example, the Cayman Trough, or Fracture Zone, is a transform fault in the Caribbean separating the South and North American Plates. Because of the nature of plate motion, there is a very small amount of spreading occurring within the Trough. Basalts erupted within the Trough are indistinguishable from those at the Mid-Atlantic Ridge. If the primitive reservoir overlay the depleted one and the depleted one were sampled only where major mantle convection currents carried it upward, we certainly would not expect to find it sampled in a place like the Cayman Trough. On the other hand, the deeper reservoir seems to be sampled exclusively, or nearly so, where there is independent evidence for major mantle upwelling in the form of mantle plumes. The geophysical evidence for this includes both gravity and elevation anomalies.

In a simple three reservoir model such as that pictured in Figure 18.1, it is possible to compute the relative masses of the depleted and primitive mantle if several parameters are known. The basic equations are simple mass-balance ones. For example, for the Nd isotopic system we may write the following mass balance equations: Since we assume that the bulk Earth has $\epsilon_{Nd} = 0$, we can write:

$$\sum_j M^j C^j \epsilon_{Nd}^j = 0 \quad 18.1$$

where M^j is the mass of the j^{th} reservoir, C^j is the concentration of Nd in that reservoir, and ϵ_{Nd}^j is the value of ϵ_{Nd} in that reservoir. We also assume the Sm/Nd is chondritic. We'll use $f_{Sm/Nd}$ to denote the relative deviation of the Sm/Nd ratio from the chondritic value, i.e.:

$$f_{Sm/Nd} = \frac{{}^{147}Sm/{}^{144}Nd - {}^{147}Sm/{}^{144}Nd_{CHUR}}{{}^{147}Sm/{}^{144}Nd_{CHUR}} \quad 18.2$$

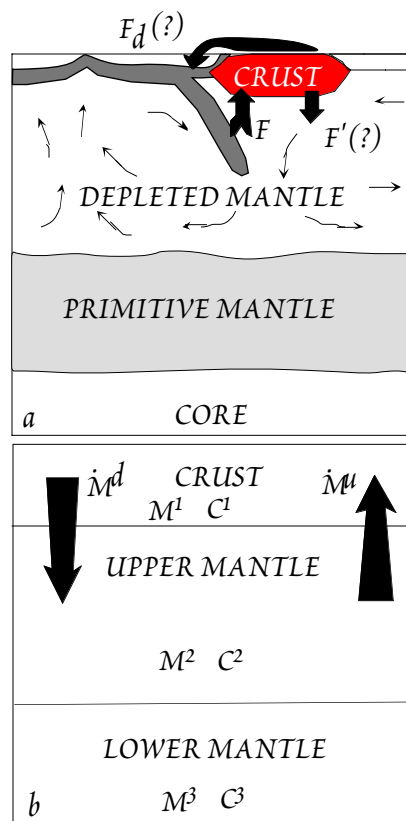


Figure 18.1 The two reservoir model of the mantle. The depleted mantle is the source of MORB and has $\epsilon_{Nd} = +10$, the lower mantle is primitive and has bulk Earth characteristics, e.g., $\epsilon_{Nd} = 0$.

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Then we may write a similar mass balance for the Sm/Nd ratio for the Earth:

$$\sum_j M^j C_{\text{Sm/Nd}}^j = 0 \quad 18.3$$

The mass balance for the Nd concentration is:

$$\sum_j M^j C_{\text{Nd}}^j = M^0 C_{\text{Nd}}^0 \quad 18.4$$

where M^0 is the mass of the silicate Earth and C_{Nd}^0 is the concentration of Nd in the silicate Earth. Finally, the masses of our three reservoirs must sum to the mass of the silicate Earth:

$$\sum_j M^j = M^0 \quad 18.5$$

The first equation simply states that the bulk-earth $\epsilon_{\text{Nd}} = 0$, the second that the Sm/Nd ratio is equal to the chondritic one, the third is the mass balance equation for Nd concentration (C_{Nd}), the fourth states that the masses of the three reservoirs must equal the total mass of the silicate earth (denoted by the superscript 0). We have implicitly assumed there is no Nd or Sm in the core. Assuming that the crust has grown from primitive mantle, then*

$$\epsilon_{\text{Nd}}^c = f_{\text{Sm/Nd}}^c Q T^c \quad 18.6$$

where T^c is the average age of the crust. If the Earth consists of only three reservoirs for Nd, namely the continental crust, depleted mantle, and primitive mantle, and if the depleted mantle and crust evolved from a reservoir initially identical to 'primitive mantle' then the mass balance equations 18.1, 18.3, and 18.4 must hold for crust and depleted mantle alone. In this case, equations 18.1, 18.4 and 18.6 can be combined to derive a relationship between the mass of the crust and the mass of the depleted mantle:

$$M^{\text{dm}}/M^c = \left(\frac{C_{\text{Nd}}^c}{C_{\text{Nd}}^0} - 1 \right) - \left(\frac{C_{\text{Nd}}^c}{C_{\text{Nd}}^0} \right) \frac{Q f_{\text{Sm/Nd}}^c T^c}{\epsilon_{\text{Nd}}^{\text{dm}}} \quad 18.7$$

Thus the mass ratio of depleted mantle to crust can be calculated if we know the Sm/Nd ratio of the crust, the ϵ_{Nd} of the depleted mantle, and the concentration of Nd in the crust and in primitive mantle. Figure 18.2 shows a plot that shows the solutions of 18.7 as a function of T^c for various values of $\epsilon_{\text{Nd}}^{\text{dm}}$ obtained by DePaolo (1980). Most estimates of the average age of the crust are between 2 and 2.5 Ga, and $\epsilon_{\text{Nd}}^{\text{dm}}$ is about +10. Possible solutions for the ratio of depleted mantle to whole mantle are in the range of 0.3 to 0.5. A number of such mass balance calculations that included other isotopic systems as well were published between 1979 and 1980, all of which obtained rather similar results. Interestingly, the fraction of the mantle above the 650 km seismic discontinuity is roughly 0.33. The

* A note on notation:

The growth equation for $^{143}\text{Nd}/^{144}\text{Nd}$ is:

$$^{143}\text{Nd}/^{144}\text{Nd} = ^{143}\text{Nd}/^{144}\text{Nd}_i + ^{147}\text{Sm}/^{144}\text{Nd} (e^{\lambda t} - 1)$$

Since the half-life of ^{147}Sm is long compared to the age of the Earth, we may use the approximation:

$$e^{\lambda t} \approx \lambda t + 1$$

and hence:

$$^{143}\text{Nd}/^{144}\text{Nd} = ^{143}\text{Nd}/^{144}\text{Nd}_i + ^{147}\text{Sm}/^{144}\text{Nd} \lambda t$$

The equation may be transformed into epsilon notation, in which case it becomes:

$$\epsilon_{\text{Nd}} \approx \epsilon_{\text{Nd}}^i + Q_{\text{Nd}} f_{\text{Sm/Nd}} t$$

where ϵ_{Nd}^i is the initial value of ϵ_{Nd} (i.e., at $t = 0$), and Q are defined as:

$$Q_{\text{Nd}} = \frac{10^4 \lambda ^{147}\text{Sm}/^{144}\text{Nd}_{\text{CHUR}}}{^{143}\text{Nd}/^{144}\text{Nd}_{\text{CHUR}}}$$

Q_{Nd} is a constant with a value of 25.13 Ga^{-1} .

mass balance calculations suggested the seismic discontinuity was the chemical boundary between upper and lower mantle.

This simple model of the mantle began to unravel as additional Nd isotopic data were acquired. In particular, it is now clear that the Sr and Nd isotope data do not form a simple linear array, and that the $\epsilon_{\text{Nd}} = 0$ point is not the minimum value observed in basalts (Figure 16.1). It is apparent then that the variation observed in Sr and Nd isotope ratios is not simply a result of mixing between depleted and primitive mantle, and that reservoirs with time-integrated LRE enrichment must exist in the mantle. Furthermore, Pb isotopic data never had been consistent with such a model. Many investigators had ignored the Pb isotope system because they felt it may have been disturbed by loss of Pb to the core.

Mixing Lines on Isotope Ratio Plots

Before going further, it should be pointed out that scatter of the data from a linear array does not preclude a two-component model. This is because mixing lines on a plot of one isotope ratio against the another need not be straight. Indeed in the general case where one ratio is plotted against another, mixing lines will be curved. The degree of curvature is dependent of the ratio r :

$$r = \frac{X_2/Y_2}{X_1/Y_1} \quad 18.8$$

where X and Y are the denominators of the two ratios and subscripts 1 and 2 denote the two end members. The more this r deviates from 1, the higher the curvature of the mixing line. In the specific case of isotope ratios, the denominators are non-radiogenic isotopes whose abundance is essentially proportional to the abundance of the element. So for Sr and Nd isotope ratios in a mixture of components 1 and 2, the mixing line has a curvature given by:

$$r = \frac{\text{Sr}_2/\text{Nd}_2}{\text{Sr}_1/\text{Sr}_1} \quad 18.9$$

where Sr_1 is the concentration of Sr in component 1, etc. Only in the case where the Sr/Nd concentration ratios are the same will the line be straight ($r=1$).

Thus the scatter observed in Figure 16.3 could be due to variable Sr/Nd ratios. However, we must ask whether it is reasonable that the reservoirs could have variable Sr/Nd ratios but constant and uniform Sr and Nd isotopic compositions? The answer would seem to be no.

MULTI-RESERVOIR MANTLE MODELS

DePaolo and Wasserburg (1976) termed the linear correlation originally observed between Sr and Nd isotope ratios the "mantle array". Even though it is now clear that mantle does not always plot on the "mantle array", the term has survived, and is useful for reference. In the subsequent discussion, we will use "mantle array" to refer to those data that plot close to a line passing through $^{143}\text{Nd}/^{144}\text{Nd} = 0.51315$ ($\epsilon_{\text{Nd}} = +10$) and $^{87}\text{Sr}/^{86}\text{Sr} = 0.7025$ (typical depleted mantle) and $^{143}\text{Nd}/^{144}\text{Nd} = 0.51264$ ($\epsilon_{\text{Nd}} = +0$) and $^{87}\text{Sr}/^{86}\text{Sr} = 0.705$ (hypothetical primitive mantle).

If we consider where individual oceanic islands or island chains plot on various isotope ratio plots, we can see that there are some systematic features. For example, several islands, including St. He-

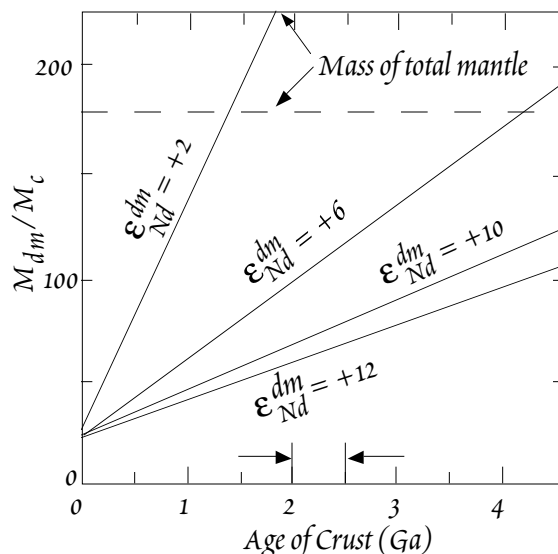


Figure 18.2 The relationship between ratio of mass of the depleted mantle to mass of the continental crust as a function of mean age of the crust calculated from equation 18.6 using various values of ϵ_{Nd} for the depleted mantle. The arrows at the bottom enclose the range of probable values for the mean age of the crust.

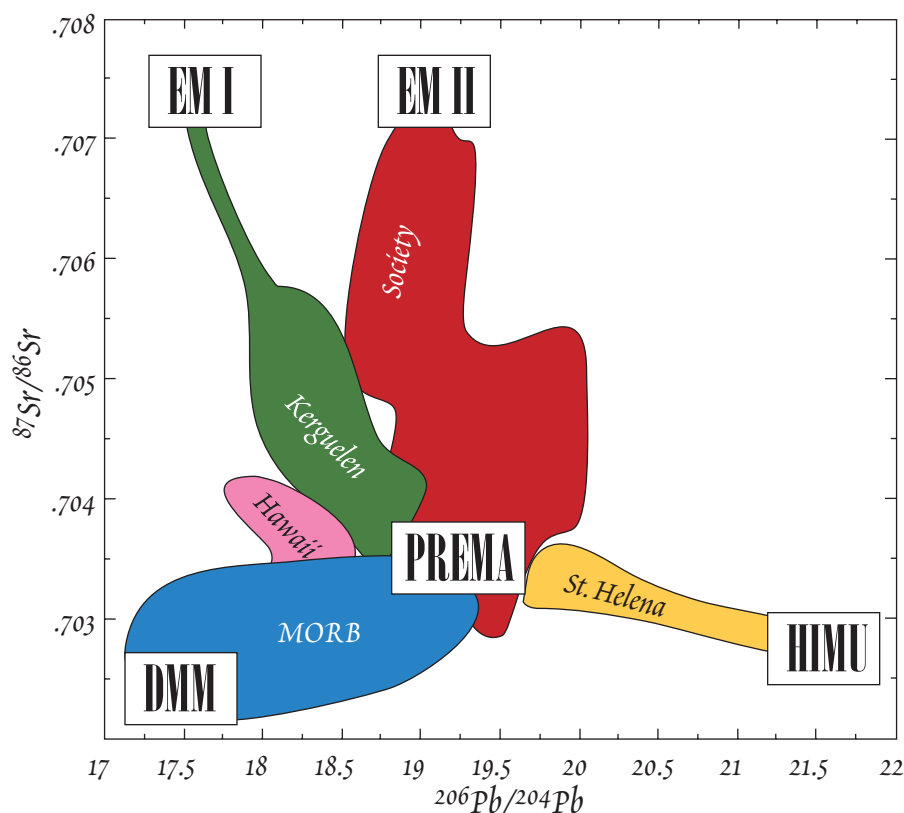


Figure 18.3. Five reservoir types of White (1985) and the components of Zindler and Hart (1986). They are essentially identical, except for Hawaii and PREMA (prevalent mantle). Other Zindler and Hart acronyms stand for high- μ (HIMU), enriched mantle I and II (EM I and EM II), and depleted MORB mantle (DMM). The main difference in interpretation is that whereas White argued that each reservoir type may consist of many reservoirs, but all had evolved through similar processes, Zindler and Hart (1986) argued that five distinct reservoirs exist, and that variations in isotope ratios result from mixing of these reservoirs.

St. Helena Island in the Atlantic and the Austral Chain in the Pacific, plot slightly below the Sr-Nd mantle array with $^{87}\text{Sr}/^{86}\text{Sr}$ about 0.7029, only slightly higher than MORB. Basalts from these same islands also plot below the Hf-Nd isotope correlation. In addition, they have remarkably radiogenic Pb, with $^{206}\text{Pb}/^{204}\text{Pb} > 20$. Following this kind of procedure, I found that oceanic basalts fall into 5 or so groups (White, 1985). It is reasonable to suppose that this reflects the existence of 5 reservoirs, or perhaps more accurately, 5 types of reservoirs within the mantle. Although this need not necessarily be the case, the idea has been accepted as a sort of working hypothesis by mantle geochemists (although it is unclear exactly how many classes there are, some prefer 4 or 6).

The next question to ask is what processes have led to the distinct identities of these reservoirs. For the MORB reservoir, this question is relatively easy to answer: removal of a partial melt accounts for the principal isotopic characteristics. Two of the reservoir types, called Kerguelen and Society by White (1985) but subsequently termed EM I and EM II ('enriched mantles 1 and 2') by Zindler and Hart (1986), have some characteristics of continental crust and sediment, and hence it is suspected that recycling of crustal material, via subduction, has been the principal process in the evolution of these reservoirs. It is not clear, however, why recycling should lead to two apparently distinct reservoirs. Numerous authors have suggested the St. Helena reservoir type, whose most distinctive characteristic is high Pb isotope ratios, and which Zindler and Hart (1986) called HIMU (for

high- μ), has acquired its distinctive isotopic characteristics through recycling of the oceanic crust. The basis for this argument is the effects of ridge-crest hydrothermal activity, which apparently removes Pb from the oceanic crust, but transfers seawater U (which is ultimately of continental crustal derivation) to the oceanic crust, effectively increasing its μ . For the most part, it must be admitted that we do not yet understand the evolution of the OIB source reservoirs. It does seem, however, that a number of processes have operated over geologic time to produce reservoirs in the mantle that are distinct from both depleted mantle and primitive mantle. The reservoir types of White (1985) and components of Zindler and Hart (1986) are shown in Figures 18.3 and 18.4.

We should note that the existence of multiple reservoirs in the mantle does not necessarily invalidate the mass balance models discussed above if the mass of the various OIB reservoirs is insignificant. Since the volume of OIB is small compared to MORB, this is certainly a possibility. However, these mass balance models also neglect the mass of the continental lithosphere, significant parts of which appear highly incompatible element enriched.

OPEN SYSTEM MODELS OF MANTLE EVOLUTION

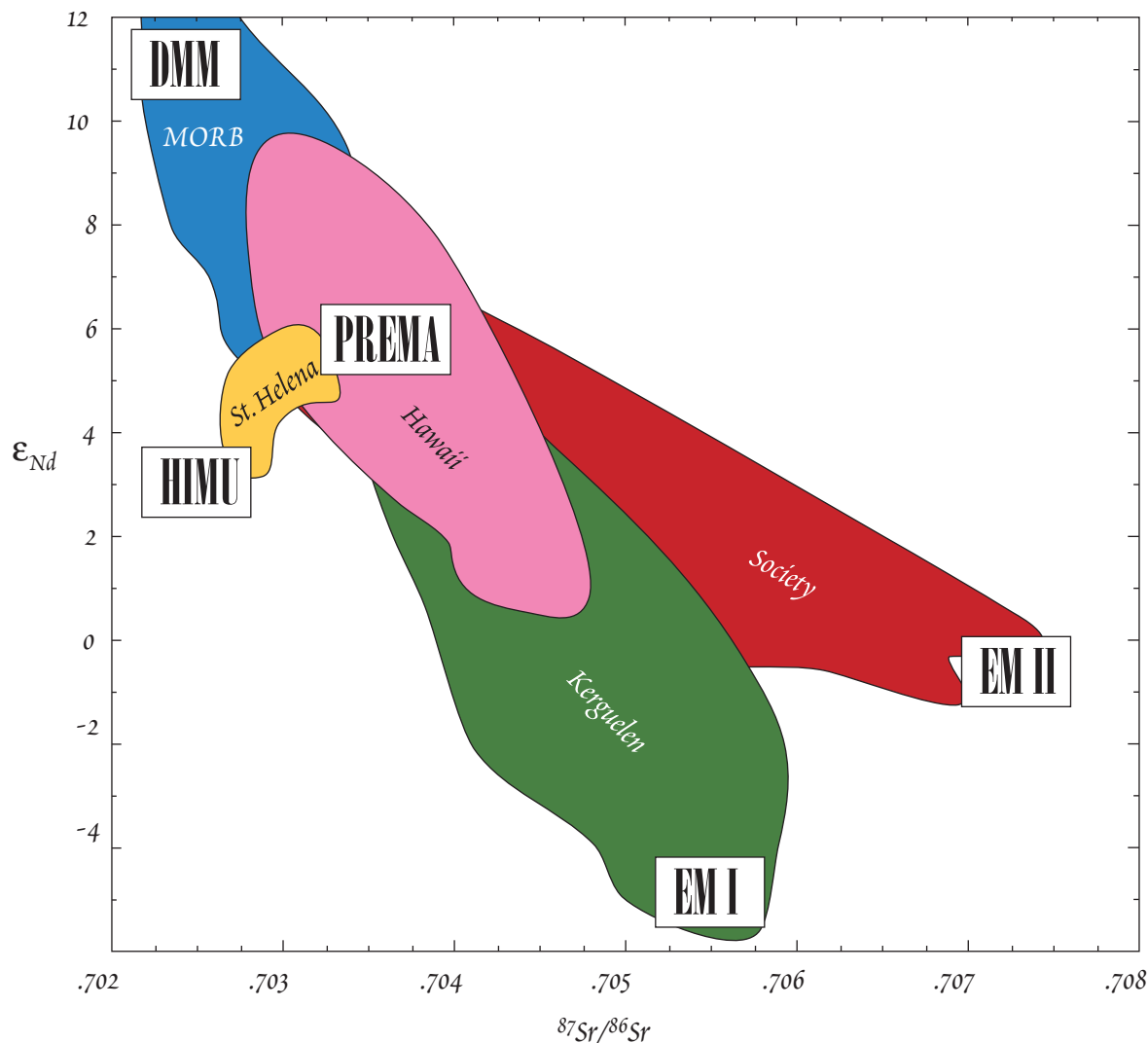


Figure 18.4. Five reservoir types of White (1985) and the components of Zindler and Hart (1986) in a plot of ϵ_{Nd} vs. $^{87}Sr/^{86}Sr$.

A radically different view of the mantle has been taken in papers by Galer and O'Nions (1985) and White (1993). The models we have discussed thus far assume that isotope ratios in mantle reservoirs reflect the time-integrated values of parent-daughter ratios in those reservoirs. Indeed, we devoted some time to the concept of time-integrated parent-daughter ratios in Lecture 16. Wasn't this, after all, what Gast said, that (among other things) an isotope ratios reflects the *time-integrated* parent-daughter ratio?

Indeed, what did Gast say? He said "*The isotopic composition of a particular sample of strontium... may be the result of time spent in a number of such environments. In any case, the isotopic composition is the time-integrated result of the Rb/Sr ratios in all past such environments.*" If for example, a sample of Sr from the depleted upper mantle (we'll adopt the acronym DUM* for this reservoir) had spent the past 4.55 Ga in that reservoir, its isotopic composition should indeed reflect the time-integrated Rb/Sr in that reservoir. But suppose that sample of Sr had spend only the last few hundred million years in the DUM? Its isotopic composition will be more of a reflection of the Rb/Sr ratios in the previous environments than in DUM. This is exactly the point made by Galer and O'Nions.

We have seen in previous lectures that the time integrated Th/U ratio is recorded by the $^{208}\text{Pb}^*/^{206}\text{Pb}^*$ ratio. Galer and O'Nions (1985) found that the average $^{208}\text{Pb}^*/^{206}\text{Pb}^*$ in MORB corresponded to a time-integrated Th/U ratio of about 3.75. The chondritic Th/U ratio, according to several compilations, is about 3.9. Since Th and U are both highly refractory elements, this should be the ratio of the bulk earth as well. The present-day Th/U ratio of the mantle source of a basalt can be deduced from Th isotope systematics, as we have seen. According to the compilation made by Galer and O'Nions, the Th/U ratio in DUM, based on Th isotope ratios in MORB, is about 2.5. That the present ratio is lower than the chondritic one makes perfect sense because Th is more incompatible than U, so we would expect this ratio to be low in DUM. Assuming the upper mantle started out with a chondritic Th/U ratio of 3.9 at 4.55 Ga, and has decreased through time to 2.5, the time-integrated

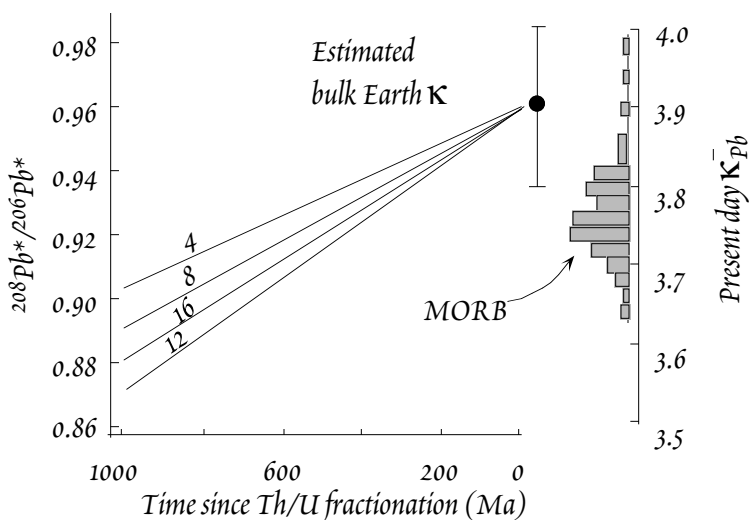


Figure 18.5. Evolution of $^{208}\text{Pb}^*/^{206}\text{Pb}^*$ is a system with Th/U = 2.5 assuming a starting Th/U of 3.9. $\bar{\kappa}_{\text{Pb}}$ is the time-integrated value of κ . Lines indicate various values of μ ranging from 4 to 16. Histogram on the right shows the values of $\bar{\kappa}_{\text{Pb}}$ in MORB. Comparison of these values with the evolution lines suggests a residence time for Pb in the upper mantle of 600 ± 200 Ma. From Galer and O'Nions (1985).

ratio should be somewhere in between these two values. Indeed, it is. However, the time-integrated value of 3.75 is surprisingly close to the initial value. This would imply in a simple evolutionary model of the mantle that the depletion in Th relative to U must have occurred relatively recently. Indeed, as illustrated in Figure 18.5, this depletion must have occurred only 600 Ma ago. This is a surprising result, and one that is inconsistent with other evidence. For example, Nd isotope ratios in ancient mantle-derived volcanic rocks suggests depletion of the upper mantle began early in Earth's history, as we shall see. Furthermore, the average age of the continental crust appears to be about 2-2.5 Ga. If the depleted mantle is the complimentary reservoir to the continental crust, time-inte-

* You may get the impression that to really succeed in mantle isotope geochemistry you need to be good at thinking up acronyms. As near as I can tell, this is true. This acronym is due to Claude Allegre.

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grated parent-daughter ratios should indicate a depletion age of about 2-2.5 Ga. Galer and O'Nions (1985) concluded that something was very wrong with conventional views of the mantle. They suggested that Pb now in the upper mantle had not resided there for long, that it was ultimately derived from a lower mantle reservoir that had a primitive (i.e., chondritic) Th/U ratio. In other words, the upper mantle had not evolved simply by losing melt fractions to the continental crust, but was a completely open system, with fluxes into it as well as out of it. They argued that the apparent depletion time of 600 Ma was in reality simply the residence time[‡] of Pb in the upper mantle.

Subsequently, most geochemists would now agree that the Th/U ratio for the bulk Earth is higher than 3.9, probably in the range of 4.0 to 4.2 (but perhaps as high as 4.3). However, additional Th isotope data on MORB indicates a lower present-day κ for the depleted mantle than estimated by Galer and O'Nions in 1985 (2.3 vs. 2.5), so the dilemma posed by Galer and O'Nions remains.

I (White, 1993) found that a similar problem arose with U/Pb ratios and came to similar conclusions as those of Galer and O'Nions. As we have seen, Pb isotope ratios in MORB rather surprisingly record a time-integrated value of μ that is higher than the bulk earth ratio (because they plot to the right of the geochron). However, I concluded from several lines of evidence that present value of μ in the upper mantle must be lower than bulk Earth. One of these lines of evidence involves solving a mass balance equation for the crust and upper mantle for U, Th, Pb and Ce. As we have seen, Pb is a volatile element, so its concentration in the bulk earth is not *a priori* known. It turns out, however, that the Pb/Ce ratio in MORB is constant. Since the other three elements are refractory, their concentrations in the bulk Earth can be assumed to be chondritic, the following relation can then be derived:

$$\mu_{DM} = 62.425 \frac{(\kappa_{BSE} - \kappa_{Crust})M_{Sys}U_{BSE}}{(\kappa_{DM} - \kappa_{Crust})(Ce_{BSE}M_{Sys} - Ce_{Crust}M_{Crust})\left(\frac{Pb}{Ce}\right)_{DM}} \quad 18.10$$

where $M_{Sys} = M_{Crust} + M_{DM}$, and the subscripts DM, Crust, and BSE denote depleted mantle, continental crust, and bulk silicate Earth respectively. The constant arises from terms for the fractional abundance of ^{204}Pb and conversion from ppm to molar units. When reasonable estimates for the various parameters are substituted into the right hand side of 18.9, one derives a value for μ_{DM} of 6 or less, whereas the best estimates for bulk silicate earth are around 8.

Thus the depleted mantle does appear to have a μ that is lower than bulk Earth, just as we would expect. But this low ratio has not been recorded by Pb isotope ratios. The obvious conclusion is that Pb in MORB could not have been present in the depleted mantle, and that it must be some flux of Pb to the upper mantle as well as out of it. I found that Pb isotope systematics were not consistent with this Pb being derived from some primitive mantle reservoir, as suggested by Galer and O'Nions. Using Galer and O'Nions estimate of the residence time of Pb in the upper mantle, I calculated the necessary flux using equation 18.11. It turns out that this flux can easily be supplied by mantle plumes, which clearly penetrate the upper mantle, and as we shall see, mix with it. Thus it appears to be mantle plumes that supply Pb, and probably other highly incompatible elements to the upper mantle, perhaps maintaining their concentrations in near steady-state. Figure 18.6 is a box model and schematic that illustrates some of the possible important fluxes through the mantle.

[‡] Residence time of some element i in a reservoir is defined as:

$$\tau = \frac{C_i M_i}{f_i} \quad 18.11$$

where τ is the residence time, C_i is the concentration of element i in the reservoir, M_i is the mass i in the reservoir, and f_i is the flux of i into or out of the reservoir. The residence time of Pb in the depleted mantle is the average time an atom of Pb will spend there between entering and leaving.

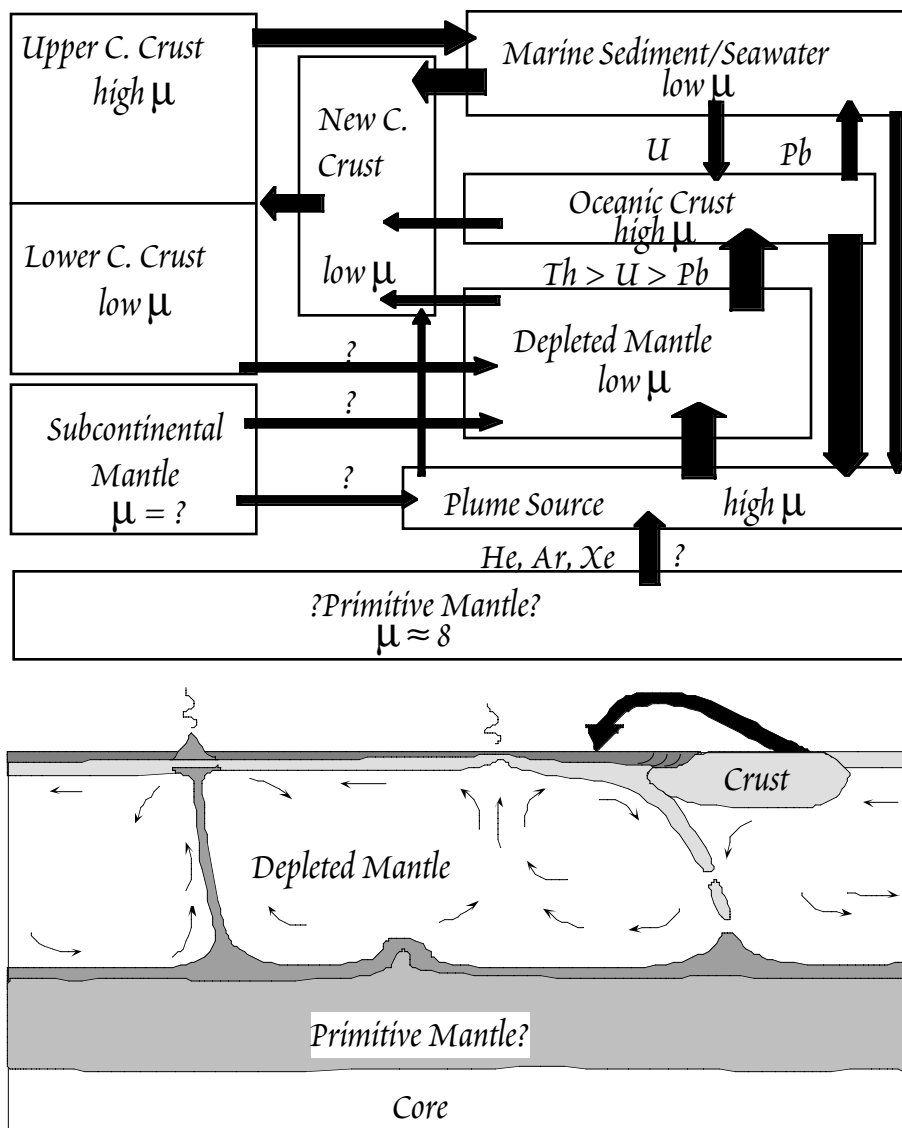


Figure 18.6. Box Model and corresponding schematic Earth illustrating the flow of U, Th and Pb through the Earth. Oceanic crust is created at mid-ocean ridges by partial melting of the depleted mantle. Th and U are partitioned into the melt to a greater extent than Pb. Hydrothermal exchange removes Pb from the oceanic crust, depositing it in sediment, and also deposits seawater U in the oceanic crust, resulting in an oceanic crust with a high μ . Some U, Th, and Pb are removed from the oceanic crust in subduction zones, but most remains and sinks to the boundary layer below the depleted mantle. Most marine sediment, which has a low μ , is accreted to continents or purged of U, Th and Pb in subduction zones. New continental crust, consisting of accreted sediment and juvenile island arc magma, has a low μ , but intracrustal differentiation produces an upper crust of high μ . Plumes form from the boundary layer of high- μ subducted oceanic crust. They mix with the depleted mantle, resupplying the depleted mantle with incompatible elements. Some flux of undegassed primitive mantle to the plume-source seems necessary to supply noble gases, but this has a trivial effect on the U-Th-Pb balance. The role of the subcontinental mantle lithosphere, as well as its Pb isotopic systematics, is uncertain. Size of the boxes does not correspond to the size of the reservoir.

GEOGRAPHIC VARIATION IN MANTLE ISOTOPIC COMPOSITION

An interesting question is whether geographic variations in mantle chemistry can be identified on a larger scale than that of individual volcanic island chains. The answer turns out to be yes. The first such geographic variation observed was on the Mid-Atlantic Ridge. Sr and Pb isotope ratios in MORB were observed to decrease with distance from Iceland and the Azores. Figure 18.7 illustrates Sr isotopic variations along the Mid-Atlantic Ridge. These variations were interpreted as 'contamination' of the asthenosphere by the Azores and Iceland mantle plumes. Somehow, the rising mantle plume mixes with asthenosphere through which it ascends, with the effect on isotopic compositions being noticeable up to 1000 km from the center of the plume. Similar effects have also been noted where even a ridge is located in proximity to a hot spot or mantle plume, including Easter Island, the Galapagos, and several of the islands in the South Atlantic and Indian Oceans.

These geographic variations are, however, recently imposed features of mantle plume dynamics. They do not necessarily imply mantle geochemical provinces. Is there evidence for such provinces, comparable to say tectonic provinces of the continents? The answer is again yes. Perhaps the first such 'province' to be identified was the Indian Ocean geochemical province. Data published as early as the early 1970's suggested MORB from the Indian Ocean were distinct from those of the Pacific and the Atlantic, having higher $^{87}\text{Sr}/^{86}\text{Sr}$ ratios. However, the scarcity and poor quality of data on Indian Ocean MORB left the issue in doubt for more than a decade. It was resolved with a flood of data on Indian Ocean MORB, beginning with a paper by Dupré and Allègre (1983). Dupré and Allègre found Indian Ocean MORB has higher $^{87}\text{Sr}/^{86}\text{Sr}$ ratios but lower $^{206}\text{Pb}/^{204}\text{Pb}$ ratios compared to MORB from other oceans. They also have high $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$ ratios for a given value of $^{206}\text{Pb}/^{204}\text{Pb}$ than other MORB. This is illustrated in Figure 17.1. Furthermore, these characteristics seem to be shared by many of the oceanic islands in the Indian Ocean. Subsequent work showed Indian Ocean MORB have low $^{143}\text{Nd}/^{144}\text{Nd}$ as well.

Hart (1984) noticed that oceanic basalts with high $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$ ratios for a given value of $^{206}\text{Pb}/^{204}\text{Pb}$ come mainly from a belt centered at about 30° S. Hart named this feature the DUPAL anomaly (after Dupré and Allègre). He defined the Dupal isotopic signature as having higher ΔSr ($\Delta\text{Sr} = [^{87}\text{Sr}/^{86}\text{Sr} - 0.7030] \times 10^4$) and high $\Delta 8/4$ and $\Delta 7/4$. The value of $\Delta 8/4$ and $\Delta 7/4$ are percent deviations from what Hart defined as the Northern Hemisphere Regression Line, regression lines through the $^{208}\text{Pb}/^{204}\text{Pb}$ — $^{206}\text{Pb}/^{204}\text{Pb}$ and $^{207}\text{Pb}/^{204}\text{Pb}$ — $^{206}\text{Pb}/^{204}\text{Pb}$ arrays for northern hemisphere data:

$$\Delta 8/4 = \left[\frac{^{208}\text{Pb}/^{204}\text{Pb} - ^{208}\text{Pb}/^{204}\text{Pb}_{\text{NHRL}}}{^{208}\text{Pb}/^{204}\text{Pb}_{\text{NHRL}}} \right] \times 100 \quad 18.12$$

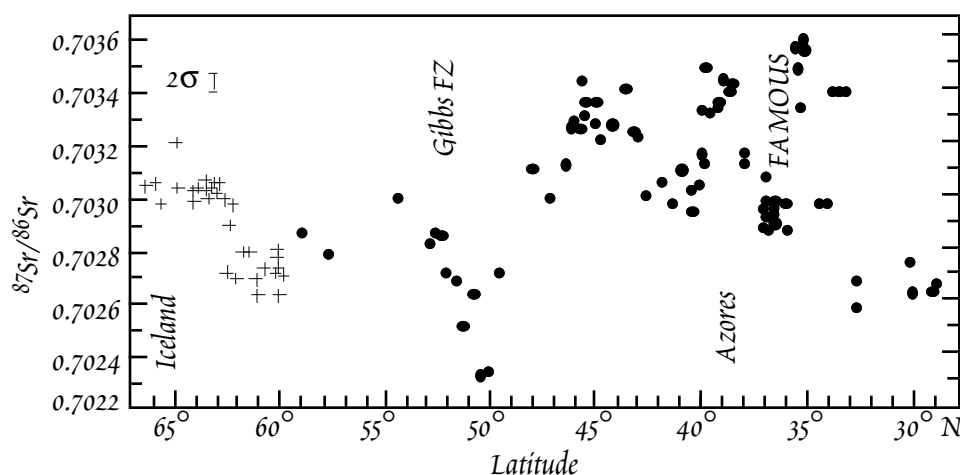


Figure 18.7. Variation of $^{87}\text{Sr}/^{86}\text{Sr}$ in MORB along the Mid-Atlantic Ridge. From White et al. (1976).

$$\text{and} \quad \Delta 7/4 = \left[\frac{{}^{207}\text{Pb}}{{}^{204}\text{Pb}} - \frac{{}^{207}\text{Pb}}{{}^{204}\text{Pb}}_{\text{NHRL}} \right] \times 100 \quad 18.13$$

$$\text{where} \quad \frac{{}^{207}\text{Pb}}{{}^{204}\text{Pb}}_{\text{NHRL}} = 15.627 = 1.209 \frac{{}^{206}\text{Pb}}{{}^{204}\text{Pb}} \quad 18.14$$

$$\text{and} \quad \frac{{}^{207}\text{Pb}}{{}^{204}\text{Pb}}_{\text{NHRL}} = 13.491 = .1084 \frac{{}^{206}\text{Pb}}{{}^{204}\text{Pb}} \quad 18.15$$

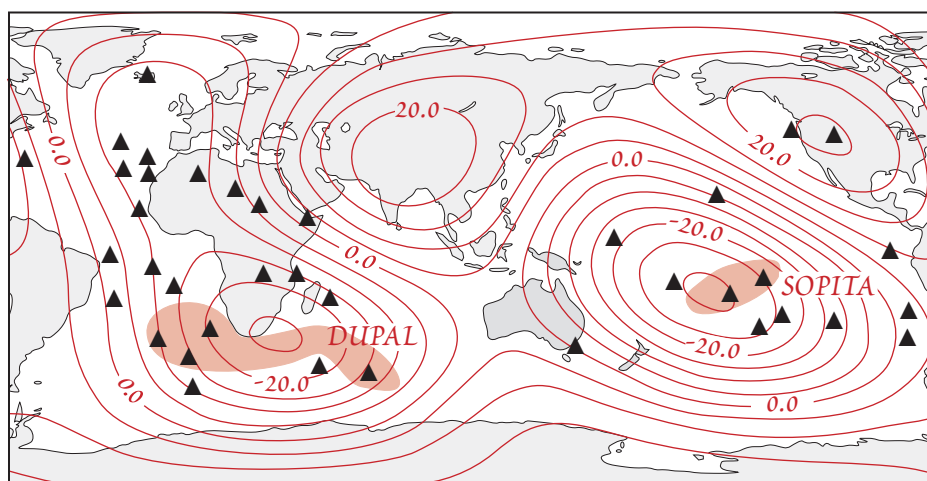


Figure 18.8. Map showing the distribution of mantle plumes (triangles), P-wave velocity anomalies (m/sec) averaged over the whole lower mantle (red lines), and location of the DUPAL and SOPITA isotope anomalies (pale red regions). Mantle plumes are located in regions of slow lower mantle seismic velocities, implying high temperatures. After Castillo (1989).

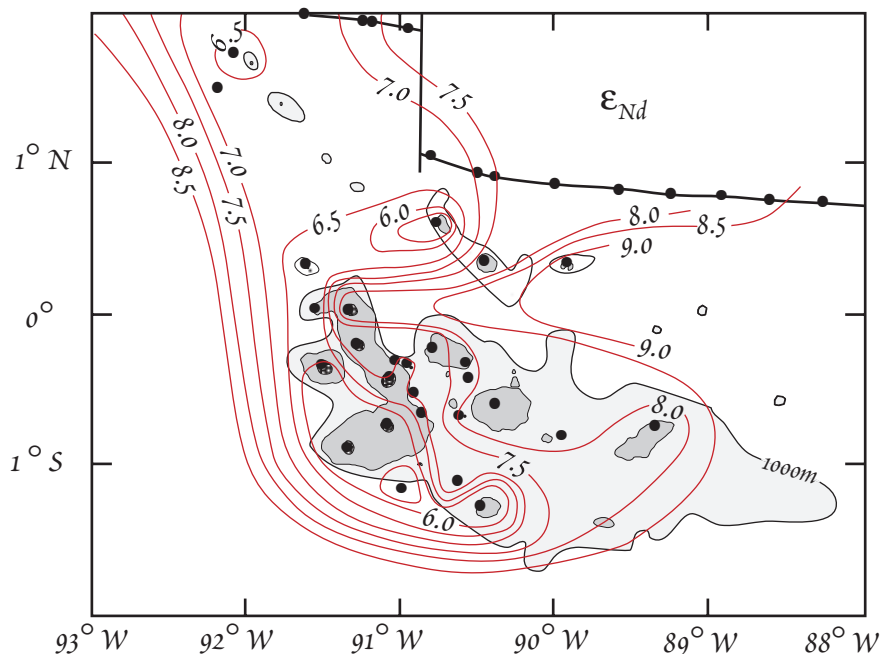


Figure 18.9. Contour map of ϵ_{Nd} variation in the mantle beneath the Galapagos. Contouring is based on average ϵ_{Nd} from 21 volcanoes, whose locations are shown by solid dots (Locations were corrected for plate motion since time of eruption).

Subsequently, Castillo (1989) suggested that Hart's "DUPAL anomaly" actually consisted of two separate regions: the DUPAL in the Indian Ocean, and the "SOPITA" (South Pacific Isotope and Thermal Anomaly) in the South Pacific and pointed out they correspond to regions of slow mantle seismic velocities, which in turn imply high mantle temperatures. Castillo's map is shown in Figure 18.8.

Interestingly, the Dupal characteristic is shared by both Indian Ocean OIB and MORB, but this does not seem to be the case in the Atlantic and Pacific. The Dupal signature has not been observed in Atlantic or Pacific MORB, except in the immediate vicinity of the Tristan da Cunha mantle plume in the south Atlantic. An additional question relates to sampling coverage. Nearly two-thirds of oceanic island occur in this belt, so it is not surprising that a particular chemistry is often found there. Nevertheless, it is clear that there is something anomalous about this region.

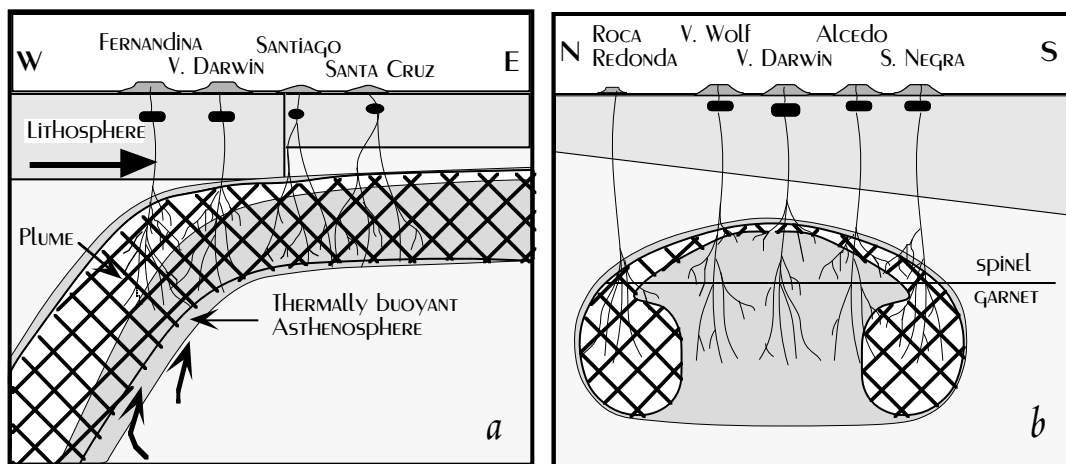


Figure 18.10. Cartoon illustrating the sheared plume model. Stippled pattern represents lithosphere, cross-hatched pattern is original plume material, grayed patterned is asthenosphere, darker gray is thermally buoyant asthenosphere. a.) east-west cross section beneath the center of the archipelago, b.) north-south cross section at the longitude of Isabela.

The Galapagos Archipelago provides another recent example of geographic variation of isotope composition in the mantle. The Galapagos provide an unusually favorable opportunity for producing a geochemical map of the mantle because they consist of 20 or so volcanos that have all been active over the past 2 or 3 million years. Combining Nd isotope ratio determined on basalts from these volcanos as well as data from previous geochemical studies of the Galapagos Spreading Center (GSC) just to the north, White et al. (in press) produced the contour map of Nd isotope ratios shown in Figure 18.9. The contours reflect regional geochemical variations in the mantle below.

The contouring reveals a horseshoe-shaped region around the western, northern, and southern periphery of the archipelago in which low ϵ_{Nd} values occur, and a region in the center of the archipelago in which high ϵ_{Nd} values occur. The high ϵ_{Nd} values are more typical of MORB than of oceanic island basalts. This pattern was unexpected. From what was observed along the MAR (Figure 18.7), one might expect ϵ_{Nd} to decrease radially from the center of the archipelago.

The pattern in the Galapagos may reflect the fluid dynamics of plume-asthenosphere interaction. Laboratory experiments have shown that a thermal plume (i.e., one that rises because it is thermal buoyant rather than chemically buoyant) will entrain surrounding asthenosphere if it is bent by asthenospheric motion. This is because the surrounding asthenosphere is heated by the plume, as a result, it also begins to rise. This interpretation is illustrated in a Figure 18.10.

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