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 $\varepsilon_{\text{Nd}} = +10$ and the primitive mantle value of $\varepsilon_{\text{Nd}} = 0$. 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 14.1.

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; 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 14.1, it is possible to compute the relative masses of the depleted and primitive mantle if several parameters are known. For example, for the Nd isotopic system we may write the following mass balance equations: Since we assume that the bulk Earth has $\varepsilon_{\text{Nd}} = 0$, we can write:

$$ \sum_j M_j C_j \varepsilon_{\text{Nd}}^j = 0 \quad 14.1 $$

where $M_j$ is the mass of the $j^{th}$ reservoir, $C_j$ is the concentration of Nd in that reservoir, and $\varepsilon_{\text{Nd}}^j$ is the value of $\varepsilon_{\text{Nd}}$ in that reservoir. We also assume the Sm/Nd is chondritic. We’ll use $f_{\text{Sm/Nd}}$ to denote the relative deviation of the Sm/Nd ratio from the chondritic value, i.e.:

$$ f_{\text{Sm/Nd}} = \frac{\text{147 Sm}^{\text{144 Nd}} - \text{147 Sm}^{\text{144 Nd}}_{\text{CHUR}}}{\text{147 Sm}^{\text{144 Nd}}_{\text{CHUR}}} \quad 14.2 $$
Then we may write a similar mass balance for the Sm/Nd ratio for the Earth:

$$\sum_j M_j C_j f_j \frac{\text{Sm}}{\text{Nd}} = 0$$

14.3

The mass balance for the Nd concentration is:

$$\sum_j M_j C_j = M^o C^o_{\text{Nd}}$$

14.4

where $M^o$ is the mass of the silicate Earth and $C^o_{\text{Nd}}$ in the concentration of Nd in the silicate Earth. Finally, the masses of our three reservoirs must sum to the mass of the silicate Earth:

$$\sum_j M_j = M^o$$

14.5

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 ($C_{\text{Nd}}$), 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 that there is no Nd or Sm in the core. Assuming that the crust has grown from primitive mantle, then

$$\epsilon_{\text{Nd}} = f \frac{\text{Sm}}{\text{Nd}} QT$$

14.6

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 14.1, 14.3, and 14.4 must hold for crust and depleted mantle alone. In this case, equations 14.1, 14.4 and 14.6 can be combined to derived a relationship between the mass of the crust and the mass of the depleted mantle:

$$\frac{M^d_m}{M^o} = \left( \frac{C^c_{\text{Nd}}}{C^o_{\text{Nd}}} - 1 \right) - \left( \frac{C^c_{\text{Nd}}}{C^o_{\text{Nd}}} \right) f \frac{\text{Sm}}{\text{Nd}} QT\epsilon_{\text{Nd}}$$

14.7

Thus the mass ratio of depleted mantle to crust can be calculated if we know the Sm/Nd ratio of the crust, the $\epsilon_{\text{Nd}}$ of the depleted mantle, and the concentration of Nd in the crust and in primitive mantle. Figure 14.2 shows a plot that shows the solutions of 14.7 as a function of $T^c$ for various values of $\epsilon^d_m$ obtained by DePaolo (1980). Most estimates of the average age of the crust are between 2 and 2.5 Ga, and $\epsilon^d_m$ 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,

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* A note on notation:
  The growth equation for $^{147}\text{Nd} / ^{144}\text{Nd}$ is:

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

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

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

and hence:

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

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

$$\epsilon_{\text{Nd}} = \epsilon_{\text{Nd}_{\text{i}}} + Q_{\text{Nd}} f \frac{\text{Sm}}{\text{Nd}}$$

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

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

$Q_{\text{Nd}}$ is a constant with a value of 25.13 Ga$^{-1}$. 
the fraction of the mantle above the 650 km seismic discontinuity is roughly 0.33. The 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 \( \varepsilon_{Nd} = 0 \) point is not the minimum value observed in basalts (Figure 13.2). 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 might have been disturbed by loss of Pb to the core.

Further problems with this model arise when we consider the new \( ^{142}Nd/^{144}Nd \) data suggesting that the Sm/Nd ratio of the Earth, or at least the observable part of it, is not chondritic. As we have seen, Boyet and Carlson (2005) calculated that the observable Earth has Sm/Nd that is 5-10% higher than chondritic, corresponding to \( \varepsilon_{Nd} \) of +5 to +10. This changes the mass balance calculations considerably. If the Sm/Nd is 5-10% higher than chondritic, the crust would have to be extracted from the entire mantle for the mantle to have \( \varepsilon_{Nd} \) of +10. If the observable Earth has Sm/Nd 8 to 10% greater than chondritic, then the \( \varepsilon_{Nd} \) values of MORB can be entirely explained by early creation of the early depleted reservoir (EDR) rather than extraction of continental crust.

If that's the case, where does the continental crust come from? One possibility is that the continental crust has been extracted from the mantle lithosphere that underlies it, i.e., the subcontinental lithosphere. Assuming the subcontinental mantle lithosphere extends uniformly to a depth of 200 km, then the subcontinental mantle lithosphere comprises less than 6% of the mass of the mantle. Is this enough mantle to form the crust? We can answer this by reframing the question somewhat: would this volume of mantle have sufficient Nd to account for that in the crust? We can calculate a minimum volume primitive mantle from which the continental crust would have to be extracted simply by comparing the Nd concentrations of crust and primitive mantle. Assuming a Nd concentration of 1.25 ppm and 20 ppm for the primitive mantle and continental crust respectively, then 10% of the Earth’s Nd is in the continental crust. Thus the crust must have been extracted from a volume of mantle larger than that of the continental lithosphere, even if Nd were quantitatively removed from it. Yet another alternative is that that reservoir is so depleted in its basaltic component that it can no longer melt and hence is not sampled by volcanism – in effect a hidden reservoir.

**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 \( r \).
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:

\[ r = \frac{86\text{Sr}}{144\text{Nd}}_2 / \frac{86\text{Sr}}{144\text{Nd}}_1 \]

where \( 86\text{Sr}_1 \) is the concentration of \( 86\text{Sr} \) in component 1, etc. Only in the case where the Sr/Nd concentration ratios are the same will the line be straight (r=1).

Since the range in isotopic compositions of Nd and Sr is usually small, \( 86\text{Sr} \) is very nearly a constant fraction of total Sr and \( 144\text{Nd} \) is very nearly a constant fraction of total Nd. Thus equation 14.9 can be rewritten as:

\[ r = \frac{\text{Sr}_2 / \text{Nd}_2}{\text{Sr}_1 / \text{Nd}_1} \]

The scatter observed in Figure 13.3 could arguably 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**

Wasserburg and DePaolo (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. 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 lead to the distinct identities of these reservoirs. For the MORB reservoir, this question is relatively easy to answer: extraction of partial melt. Two of the reservoirs 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. White and Hofmann (1982) suggested EMII could be produced by recycling of oceanic crust and sediment and EM I by oceanic crust alone, but it is unclear whether oceanic crust is ever subducted without sediment, and consequently not clear 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 also been produced by 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 14.3 and 14.4.

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. Mass balance models also neglect the mass of the subcontinental mantle lithosphere. Although significant parts of the subcontinental lithosphere appear highly incompatible element enriched, it is probably not a significant reservoir of incompatible elements.

**OPEN SYSTEM MODELS OF MANTLE EVOLUTION**

A 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 the previous lecture. Wasn’t this, after all, what Gast said, that (among other things) an isotope ratio reflects the time-integrated parent-daughter ratio?

Figure 14.4. Five reservoir types of White (1985) and the components of Zindler and Hart (1986) in a plot of $\varepsilon_{Nd}$ vs. $^{87}\text{Sr}/^{86}\text{Sr}$. 
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 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, $\kappa$ 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 $\kappa$ of 3.9 at 4.55 Ga, and has decreased through time to 2.5, the time-integrated 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 14.5, 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 2.25 Ga. If the depleted mantle is the complimentary reservoir to the continental crust, time-integrated parent-daughter ratios should indicate a depletion age of about 2.25 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

* 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.
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. Galer and O’Nions may have underestimated the value of $\kappa$ in the Earth; it is very possible that the value for the bulk Earth is higher than 3.9, perhaps in the range of 4.0 to 4.2 (and even as high as 4.3). However, this only extends the residence time to about 1 to 1.2 Ga, still much less than the average age of continents, so their point remains valid.

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 $\mu$ 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 $\mu$ 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 derived:

$$\mu_{DM} = 62.425 \frac{(\kappa_{BSE} - \kappa_C) M_{BSE} U_{BSE}}{(\kappa_{DM} - \kappa_C)(Ce_{BSE} M_{BSE} - Ce_C M_C)(Pb/Ce)_{DM}}$$

where $M_{BSE} = M_C + M_{DM}$ and the subscripts DM, C, and BSE denote depleted mantle, continental crust, and bulk silicate Earth respectively. The constant arises from terms for the fractional abundance of $^{206}$Pb and conversion from ppm to molar units. When reasonable estimates for the various parameters are substituted into the right hand side of 14.11, one derives a value for $\mu_{DM}$ of 6 or less, whereas the best estimates for bulk silicate earth are around 8.

Thus the depleted mantle does appear to have a $\mu$ 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 be 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 14.11. 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.

**Geographic Variations in Mantle Isotopic Composition**

An interesting question is whether geographic variations in mantle chemistry can be identified on a larger scale that 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 14.6 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 lo-

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$^\ddagger$ Residence time of some element $i$ in a reservoir is defined as:

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

where $\tau$ 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.
cated 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. The 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 \( \frac{\text{Sr}}{\text{Sr}} \) 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 \( \frac{\text{Sr}}{\text{Sr}} \) ratios but lower \( \frac{\text{Pb}}{\text{Pb}} \) ratios compared to MORB from other oceans. They also have high \( \frac{\text{Pb}}{\text{Pb}} \) and \( \frac{\text{Pb}}{\text{Pb}} \) ratios for a given value of \( \frac{\text{Pb}}{\text{Pb}} \) than other MORB.

![Figure 14.6. Variation of \( \frac{\text{Sr}}{\text{Sr}} \) in MORB along the Mid-Atlantic Ridge. From White et al. (1976).](image)

![Figure 14.7. 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).](image)
This is can be seen in Figure 13.7. Furthermore, these characteristics seem to be shared by many of the oceanic islands in the Indian Ocean. Subsequent work that 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 $^{208}\text{Pb}/^{204}\text{Pb}$ come mainly from a belt centered at about $30^\circ$ S. Hart named this feature the DUPAL anomaly (after Dupré and Allègre). He defined the DUPAL isotopic signature as having higher $\Delta$Sr ($\Delta\text{Sr} = ([^{87}\text{Sr}/^{86}\text{Sr} - 0.7030] \times 10^4$) and high $\Delta8/4$ and $\Delta7/4$. The value of $\Delta8/4$ and $\Delta7/4$ 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:

$$\Delta8/4 = \left(\frac{^{208}\text{Pb}^{204}\text{Pb} - ^{208}\text{Pb}_{\text{NHRL}}^{204}\text{Pb}}{^{208}\text{Pb}_{\text{NHRL}}^{204}\text{Pb}} \right) \times 100$$

$$\Delta7/4 = \left(\frac{^{207}\text{Pb}^{204}\text{Pb} - ^{207}\text{Pb}_{\text{NHRL}}^{204}\text{Pb}}{^{207}\text{Pb}_{\text{NHRL}}^{204}\text{Pb}} \right) \times 100$$

where

$$^{207}\text{Pb}_{\text{NHRL}}^{204}\text{Pb} = 15.627 + 1.209 \times ^{207}\text{Pb}^{204}\text{Pb}$$

$$^{208}\text{Pb}_{\text{NHRL}}^{204}\text{Pb} = 13.491 + 0.1804 \times ^{208}\text{Pb}^{204}\text{Pb}$$

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 14.7.

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 in often found there. Nevertheless, it is clear that there is something anomalous about this region.

The Galapagos Archipelago provides another 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 volcanoes that have all been active over the past 2 or 3 million years. Combining Nd isotope ratio determined on basalts from these volcanoes as well as data from previous geochemical studies of the Galapagos

![Figure 14.8. Contour map of $\varepsilon_{\text{Nd}}$ variation in the mantle beneath the Galapagos. Contouring is based on average $\varepsilon_{\text{Nd}}$ from 21 volcanos, whose locations are shown by solid dots (Locations were corrected for plate motion since time of eruption).](image-url)
Spreading Center (GSC) just to the north, White et al. (1993) produced the contour map of Nd isotope ratios shown in Figure 14.8. 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 $\varepsilon_{\text{Nd}}$ values occur, and a region in the center of the archipelago in which high $\varepsilon_{\text{Nd}}$ values occur. This pattern was unexpected. From what was observed along the MAR (Figure 14.6), one might expect $\varepsilon_{\text{Nd}}$ 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 14.9.

**THE ORIGIN OF MANTLE PLUMES AND THE COMMON COMPONENT IN PLUMES**

Determining how the various geochemical reservoirs of the mantle have evolved is among the most vexing problems in geochemistry. The principal observation to be explained is that mantle plumes invariably have less depleted isotopic signatures than MORB, and the isotopic compositions of some indicate net enrichment in incompatible elements. As we saw in the previous lecture, mantle plumes were initially thought to consist of primitive mantle (e.g., Schilling, 1973). As we found, mixing between primitive and depleted mantle can explain the Sr and Nd isotopic compositions of some plumes, but virtually none of the Pb isotope data can be explained this way, nor are the trace element compositions of OIB consistent with plumes being composed of primitive mantle. Indeed, although ‘primitive mantle’ has proved to be a useful hypothetical concept, no mantle-derived basalts or xenoliths have appropriate compositions to be ‘primitive mantle’ or derived from it. It is possible that no part of the mantle retains its original, primitive, composition (on the other hand, to have survived, primitive mantle must not participate in volcanism and other such processes, so the absence of evidence for a primitive mantle reservoir is not evidence of its absence).
Hofmann and White (1982) suggested mantle plumes obtain their unique geochemical signature through deep recycling of oceanic crust (Figure 14.10). Partial melting at mid-ocean ridges creates oceanic crust that is less depleted in incompatible elements than the depleted upper mantle. The oceanic crust is apparently inevitably subducted as virtually none is preserved at the surface, so it clearly is recycled back into the mantle. The question is what becomes of it? Hofmann and White noted that once oceanic crust reaches depths of about 60 km it converts to eclogite, which is denser than peridotite. Because it is rich in Fe and garnet-forming components, it remains denser than peridotite at all depths greater than 60 km (except, perhaps, just at the 660 discontinuity due to the negative Clapeyron slope). Thus it will sink to the base of the convecting region. If the mantle is chemically stratified, with a Fe-rich lower mantle, oceanic crust would sink to a thermal boundary layer at the 660 discontinuity. If the entire mantle convects as a single unit, that is if it is not chemically stratified, ocean crust will sink to base of the mantle, becoming embedded in thermal boundary layer there (D’). Hofmann and White originally suggested radioactive heating would ultimately cause it to become buoyant. However, heat conducted into it from below, from either the lower mantle or the core, is likely a more important heat source. In any case, upon sufficient heating, it rises, melting near the surface to create intraplate volcanoes.

As we shall see in a subsequent lecture, sediment appears often, if not always, to be subducted along with the oceanic crust. This subducted sediment would also contribute to incompatible element enrichment of plumes. Varying amounts, types, and ages of subducted sediment may be responsible for some of the geochemical variety displayed by plumes. Since sediment is ultimately derived from the continents, recycling of oceanic crust, continental crust, mantle plumes, and oceanic island basalts may all be part of a grand geochemical cycle. Tectonic erosion of continental crust in subduction zones and delamination of continental crust may be alternative mechanisms for deep recycling of continental crust.

Because the major element chemistry of OIB is often similar to that of MORB, it seems unlikely plumes could be composed entirely of recycled oceanic crust. Presumably, they consist primarily of peridotite, with a subordinate fraction of oceanic crust. However, because the oceanic crust has much higher incompatible element concentrations than peridotite, it provides most of the isotopic and incompatible element “flavor” of plumes.

Trace elements provide some evidence that some plumes contain a recycled sediment component. The Pb/Ce ratio is particularly useful indicator of the presence of sediment for several reasons. First, the Pb/Ce ratio is comparatively uniform in MORB and many OIB. Second, the Pb/Ce ratio is an order of magnitude higher in sediments than in the mantle (typically, Pb/Ce is greater than...
Lecture 14

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An alternative origin for mantle plumes was proposed by McKenzie and O’Nions (1983). They noted
the common evidence for incompatible element enrichment in the subcontinental lithosphere (which
we discuss in the next section) and suggested this material may occasionally become delaminated. Be
cause it is cold, it would also sink to the deep mantle. As in the case of the Hofmann and White model,
it would be stored in a thermal boundary layer, heated, and rise in the form of mantle plumes. Howev
er, as we shall see in the next section, recent studies have shown that the Os isotope composition of
the subcontinental lithosphere is quite distinctive, and quite different from that of mantle plumes. Th
suggests that “delaminated” subcontinental lithosphere does not contribute to mantle plumes. Be
cause mantle plumes come in several geochemical varieties, it is possible that both mechanisms operate.
Indeed, other as yet unknown processes may be involved as well.

Most oceanic islands show some variability in their isotopic compositions, defining elongated arrays
on plots of isotope ratios. Such elongated arrays suggest mixing. This raises the rather obvious ques
tion of what is mixing with what. In a few cases, the Comores are a good example, the elongate arrays
seems to reflect mixing between different plume reservoirs. The Comores data defines a trend in iso
topic space that appears to be the result of mixing between an EMI and a HIMU component. In other
cases, such as the Galapagos, the plume is clearly mixing with the depleted mantle. However, in man
cases, the cause of the isotopic variation is less clear.

Hart et al. (1992) plotted oceanic basalt isotope data in three dimensions, with axes of \(^{87}\text{Sr}/^{86}\text{Sr}\),
\(^{143}\text{Nd}/^{144}\text{Nd}\), and \(^{206}\text{Pb}/^{204}\text{Pb}\) (Figure 14.12). Principal component analysis confirmed that 97.5% of the
variance in the oceanic basalt isotope data could be accounted for by these ratios (leav
ing 2.5% to be accounted for by \(^{207}\text{Pb}/^{204}\text{Pb}\),
\(^{206}\text{Pb}/^{204}\text{Pb}\), and \(^{176}\text{Hf}/^{177}\text{Hf}\)). They found that
most of the data plotted within a tetrahedron defined by the hypothetical end members
EM1, EM2, HIMU, and DMM. They also noticed that many arrays were elongated toward
the base of this tetrahedron on the DMM-
HIMU join. From this they concluded that in
many, if not most cases, mantle plumes appear
to mixing with a previously unidentified component, which they named “FOZO” (an
acronym for Focal Zone), that has the ap
proximate isotopic composition of \(^{87}\text{Sr}/^{86}\text{Sr} = 0.7025, \epsilon_{\text{Nd}} = +9\), and \(^{206}\text{Pb}/^{204}\text{Pb} = 21.5\). They
suggested that FOZO is the isotopic composi
tion of the lower mantle and that plumes ris
ing from the core mantle boundary entrain
and mix with this lower mantle material. It is
unclear, however, whether such a composi
tion for the lower mantle can be fitted to rea

Figure 14.11. \(^{207}\text{Pb}/^{204}\text{Pb}\) and \(^{206}\text{Pb}/^{204}\text{Pb}\) in basalts from the Societies Islands studied by White and Duncan (1996). A calu
culated mixing line between depleted mantle and sediment passes through the data. Also shown are es
imated \(^{207}\text{Pb}/^{204}\text{Pb}\) ratios of average continental crust and bulk silicate Earth (BSE).
reasonable isotope mass balances for the Earth. A rather similar idea was presented by Farley et al. (1992), who point out that this additional component, which they called “PHEM”, seems to be associated with high $^3\text{He}/^4\text{He}$. White (1995) concurred with these ideas, but argued that the $^{87}\text{Sr}/^{86}\text{Sr}$ of FOZO is higher, and the $\varepsilon_{\text{Nd}}$ lower, than estimated by Hart et al. (1992) and probably closer to the values chosen by Farley et al. (1992). Hanan and Graham (1996) used $\text{Pb}$ and $\text{He}$ isotope ratios to deduce yet another potential common component of plumes, which they called “C”. The “C” composition of Hanan and Graham is similar to the PHEM of Farley et al. (1992) and may be just another name for the same thing. “C” and “PHEM” occur in the interior of the tetrahedron shown in Figure 14.12. Hanan and Graham (1996) argued that “C” is the principal component of plumes, the other components just add “flavor”. The “C” and PHEM compositions would have $\varepsilon_{\text{Nd}}$ of about +5. This is fairly close the value the accessible Earth, or the bulk silicate Earth should have if it has a $^{147}\text{Sm}/^{144}\text{Nd}$ ratio of about 0.206, as by Boyet and Carlson (2005). This is also close the modal (i.e., most common) value of $\varepsilon_{\text{Nd}}$ for oceanic island basalts. This composition may be something approximating a mantle composition that has been minimally modified since either the Earth formed or it initially differentiated and may be quite abundant in the deep Earth. In other words, it might be something close to “primitive mantle”.

REFERENCES AND SUGGESTIONS FOR FURTHER READING


