Geochemical tracers of mantle processes

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Introduction

The mantle can be sampled directly only very rarely. Geochemists have thus come to rely heavily on mantle-derived magmas to study the composition and evolution of the mantle. Only those compositional features that are unaffected by magmatic processes are useful as tracers of mantle processes. These include radiogenic isotope ratios such as those of He, Sr, Hf, and Os, stable isotope ratios, and ratios of highly incompatible elements or elements of similar incompatibility, such as Ba/Nb or Pb/Ce. The term “incompatible” denotes a preference of the element for a melt over mantle minerals. Highly incompatible elements will partition entirely into the melt under most circumstances, so that the ratio of two such elements in a basalt will be virtually identical to that ratio in its source. This is also true to a lesser degree of ratios such as La/Sm and Zr/Nb, as Zr and Sm are not highly incompatible elements.

Of these mantle tracers, isotope ratios have proven the most useful because they are insensitive to magmatic processes and because they are functions of time. Thus the observation that two basalts have different 143Nd/144Nd ratios demonstrates not only that their mantle sources had different Sm/Nd ratios, but also that these differences existed for geologically long times (10⁸ years or more).

From a geochemical perspective, the mantle can be divided into 3 parts. The first is the mantle lithosphere. Most of the oceanic lithosphere is probably chemically similar to the depleted mantle, but the continental lithosphere is more heterogeneous. The second is the source of mid-ocean ridge basalts (MORB), often referred to as “depleted mantle”. It is characterized by incompatible element depletion and radiogenic isotope ratios indicating that this depletion must have occurred billions of years ago. The depleted mantle is more homogeneous than other mantle or crustal reservoirs. Most mantle geochemists and geophysicists believe this reservoir occupies the upper mantle beneath the lithosphere. There are some dissenters, most notably D. L. Anderson [e.g., Anderson et al., 1992], who places the depleted mantle within the seismic transition zone. The final part is the source of mantle plumes. Plumes must originate from some thermal boundary layer, such as the core-mantle boundary or the 660 km discontinuity.

In the following sections, I discuss the advances made in the last 4 years in understanding the evolution of these 3 parts of the mantle. I then discuss some significant advances in Re-Os isotope and noble gas isotope geochemistry. Finally, I discuss the identification of a common component in mantle plumes. Due to space limitations, this review focuses on large scale mantle processes, and does not consider U-series isotopic studies of magmatic processes or subduction zone processes. The reader is referred to Gill [1992] and Hawkesworth [1993], respectively, for reviews on these subjects.

Continental and Oceanic Mantle Lithosphere

The subcontinental lithosphere is extremely heterogeneous in its isotopic and trace element composition. This reflects the stable nature of the continental lithosphere: variations in parent-daughter ratios are preserved over long time periods and produce a larger range in radiogenic isotope ratios than in the asthenosphere, where convective mixing destroys heterogeneity. For example, Deng and Maccougal [1992] found that clinopyroxenes in xenoliths from Pleistocene volcanics of Inner Mongolia define a 1.64 Ga (1 Ga is 10⁹ years) isochron, which appears to record incompatible element depletion of this part of the mantle. Johnson and Beard [1993] found that basalts from the Rio Grande Rift define a shallower εNd – εHf slope than do oceanic basalts, which they attribute to prior incompatible element depletion through melting.

Several mechanisms may produce incompatible element enrichment of the subcontinental lithosphere, including metasomatism by hydrous fluids released from subducting lithosphere and freezing of mantle plume-derived or asthenosphere-derived melts. Such processes appear to have affected much of the lithosphere beneath the Western U.S. [Johnson and Thompson, 1991; Menzies et al., 1991; Fitton et al., 1991]. Within the last 5 Ma (Ma is 10⁶ yrs) depleted asthenosphere appears to have replaced lithosphere as the principal magma source within the basin and range [Fitton et al., 1991].

Volcanic rocks of Eastern China have isotopic characteristics that fall almost entirely within the range of oceanic island basalts [Basu et al., 1991]. Xenoliths in these basalts show a wider range of isotopic compositions than the basalts [Tatsumoto et al., 1992]. These authors suggested that the incompatible element enrichment recorded by these isotope ratios occurred through metasomatism by fluids released from subducting lithosphere in Precambrian time.

The oceanic mantle lithosphere can apparently also be metasomatized by infiltrating melts. Hauri et al. [1993] reported finding xenoliths in lavas from Samoa and Tubuai (Austral Islands) that contained secondary assemblages of clinopyroxene ± spinel ± apatite ± glass and were highly enriched in incompatible elements, with notable relative depletions of Ti, Zr, Sr, and Nb. This suggests they formed by reaction with or precipitation from a carbonatite melt. The Samoan xenoliths lie on an extension of the Sr-Nd isotopic array defined by Samoan basalts. Hauri et al. [1993] thought the metasomatizing carbonatite was produced by melting of the plume in both cases, probably an early, very low degree melt that migrated upward and reacted with the overlying lithosphere.

MORB and Depleted Mantle

The depleted upper mantle is perhaps the best understood of the mantle reservoirs, as MORB provide a global sample of it. Although it is relatively homogeneous in comparison with other mantle reservoirs, significant heterogeneity does exist within it. The Indian Ocean upper mantle is isotopically distinct from the upper mantle beneath the Atlantic and Pacific. The boundary between the Indian and Pacific upper mantle provinces, known from earlier work to be located within the Australian-Antarctic Discordance was found by Pyle et al. [1992] to be only 40 km wide and be migrating westward at about 25 mm/yr. In contrast to the sharp eastern boundary of the Indian province, the southwestern boundary appears to be more diffuse, and occurs between 17° E and 26° E along the Southwest Indian Ridge [Mahoney et al., 1992]. In the northeast Indian Ocean, Schilling et al. [1992] observed incompatible element ratios such as La/Sm and isotope ratios of Sr and Pb in basalts from the Gulf of Aden decreasing with distance from the Afar plume center, located at the junction of the Red Sea, Gulf of Aden, and African Rift. In addition to plume and depleted mantle components (the latter is Indian type), they also identified possible Pan-African continental lithosphere in the isotope systematics.
Hf isotopes in MORB have been somewhat problematic because they are more variable in MORB for a given $\varepsilon_{Nd}$ than are oceanic island basalts. This was confirmed by the first new Hf isotope data on MORB in nearly a decade [Salters and Hart, 1991]. Salters and Hart argue that this requires the presence of both residual garnet and clinopyroxene in the melting events that depleted the mantle in Hf relative to Lu and Nd relative to Sm.

Evolution of the Depleted Mantle

For the Sr and Nd decay systems, the depleted mantle and continental crust appear to be complimentary and the isotope data are consistent with a simple model in which the crust has been created by extraction of partial melts from the mantle. Pb isotope ratios are not easily accommodated in such models, however. Pb isotope ratios record an increase of U/Pb in the depleted mantle, whereas extraction of partial melts should decrease in this ratio. White [1993] concluded from a study of U, Th, and Pb in MORB that the U/Pb ratio in the depleted mantle is lower than the bulk Earth value, consistent with the incompatible depletion of this reservoir. Furthermore, the low value of U/Pb he inferred is consistent with continental mass balance. The problem then apparently lies with Pb isotope ratios. White [1993] concluded that the depleted mantle is an open system, arguing depletion of the upper mantle is maintained at approximately steady-state by a combination of extraction of partial melts and input of new material from mantle plumes.

Further evidence of early mantle depletion has been published within the last 4 years. Collerson et al. [1991] reported a Sm-Nd isochron for a metamorphosed ultramafic suite from Labrador that gave an age of 3815 Ma and an initial $\varepsilon_{Nd}$ of +3.0. Collerson et al. interpreted these rocks as tectonically emplaced slices of Archean mantle. They are slightly light rare earth-enriched, in contrast to the positive initial $\varepsilon_{Nd}$ that indicates a light rare earth-depleted history. A second suite from the same region, which Collerson et al. [1991] interpret as meta-komatiites, is light rare earth-depleted yet has negative initial $\varepsilon_{Nd}$ (-2.1±2.3). There has also been a preliminary report of +4 initial $\varepsilon_{Nd}$ in the 3960 Ma Acasta gneisses in the Slave Province of Canada [Bowring et al., 1994].

$\varepsilon_{Nd}$ as high as +4 at 3800 Ma requires extensive light rare earth depletion of the mantle. This in turn requires the existence of a sizable light rare earth-enriched reservoir, perhaps volumetrically comparable to the present continental crust. However, pre-3800 Ma continental crust is extremely sparse. Instead of continental crust, some have focused on basaltic oceanic crust as the incompatible element-enriched compliment of the early Archean upper mantle. Galer and Goldstein [1991] pointed out, however, that the large degrees of melting associated with modern oceanic crust production do not sufficiently fractionate Sm and Nd to produce the depletion observed in the early Archean mantle. They suggested thick, alkaline oceanic crust was created by smaller degrees of melting in a regime where no seafloor spreading occurred.

A third possible explanation for the high $\varepsilon_{Nd}$ in early Archean rocks is that they are simply wrong, and there is increasing skepticism about their validity, e.g., Carlson [1994]. While the actual analytical data are not questioned, present day $^{142}\text{Nd}/^{144}\text{Nd}$ and $^{147}\text{Sm}/^{144}\text{Nd}$, not initial $\varepsilon_{Nd}$, are the measured quantities. Calculation of initial $\varepsilon_{Nd}$ from these data assumes the samples in question have been closed for their entire history and that the age is accurately known. The suspicion is that subsequent metamorphism has changed the Sm/Nd ratios. Nearly all early Archean rocks with high initial $\varepsilon_{Nd}$ have undergone subsequent metamorphism and the high initial $\varepsilon_{Nd}$ represent only one end of a spectrum of initial $\varepsilon_{Nd}$ values.

A closely related controversy is the question of $^{142}\text{Nd}$ anomalies in early Archean rocks. $^{142}\text{Nd}$ is produced by $\alpha$-decay of $^{142}\text{Sm}$, an extinct radionuclide with a half-life of 103 Ma. A $^{142}\text{Sm}/^{144}\text{Sm}$ ratio of 0.0078±0.0010 in the early Solar System has been established from meteoritic studies. If mantle depletion began before about 4.3 Ga, then excesses in $^{142}\text{Nd}$ due to $^{146}\text{Sm}$ decay should be observed in those samples with high positive initial $\varepsilon_{Nd}$. Harper and Jacobsen [1992] reported an average excess $^{142}\text{Nd}$ of 34±7 ppm for six replicates of an Isua, West Greenland sample that has an initial $\varepsilon_{Nd}$ of +4 at 3.81 Ga. This would appear to confirm that depletion of the upper mantle got underway very early. However, excess $^{142}\text{Nd}$ in other early Archean samples have not been found, nor has the Isua analysis been replicated in another laboratory. As a result, the existence of $^{142}\text{Nd}$ excesses remains controversial.

Mantle Plumes

Mantle plumes, most geochemists believe, provide a sample of the deep mantle, though the exact depth from which plumes rise remains uncertain. Schilling [1991] used the distinctive isotope and trace element ratios in plumes as “tracers of mantle dynamics, much as dyes are used to follow flow patterns.” He investigated 13 plumes located sufficiently close to spreading centers that they influenced the composition of basalt erupted on those spreading centers. Schilling used the width of the geochemical anomaly and the excess ridge elevation along the ridge axis in a simple plume source–ridge sink model to estimate the volume fluxes of these plumes. His estimates of fluxes for these plumes agreed within a factor of 5 with completely independent estimates made by Sleep [1980]. Schilling [1991] also calculated the excess temperature of the plumes and found they fell in the range of 160-280 K, again in reasonable agreement with other independent estimates. Given the uncertainty in Schilling’s temperature estimates (±50 K), plume temperatures appear to be remarkably uniform. Furthermore, excess temperature and plume geochemistry were unrelated, implying that all plumes had a common depth of origin with a regulated source of heat, which Schilling felt was the D’ layer (a region of anomalous seismic velocities at the base of the mantle).

Plumes generally produce linear island chains with only a few, and sometimes just one, volcanos active at any given time. The Galapagos are an exception, with at least 13 active volcanos arranged in a rectilinear pattern over an area of $10^5 \text{km}^2$. This fortunate circumstance allowed White et al. [1993] to produce a 2 dimensional geochemical map of the upper mantle in the region. They found isocone and incompatible element ratios defined a horseshoe pattern, with the most depleted values in the center of the Galapagos Archipelago and the more enriched values on the eastern, northern, and southern periphery. They concluded this pattern reflected thermal entrainment of asthenosphere by the plume as it undergoes velocity shear in the uppermost asthenosphere. Desonie et al. [1993] also invoked thermal entrainment by a sheared plume to explain isotopic variations in the Marquesas Archipelago.

Hawaii continues to be an area of intense study. Kennedy et al. [1991] found that post-shield lavas do not define a temporal trend to more depleted compositions, as has been found in other Hawaiian volcanos. Leeman et al. [1994] also found no simple temporal trend in the lavas of Kahoowale, but did find evidence of a cyclic pattern of geochemical variation, which may reflect variations in the contribution of plume and lithosphere to Kahoowale magmas. In contrast, Kurz and Kammer [1991] found distinct temporal trends in $^{87}\text{Sr}/^{86}\text{Sr}$, $^{206}\text{Pb}/^{204}\text{Pb}$, and $^{3}\text{He}/^{4}\text{He}$ in Mauna Loa lavas over the past 30,000 yrs and concluded 3 distinct mantle sources are
involved. Temporal changes in trace element ratios such as Pb/Ce and Nb/La suggest a systematic change in the composition of the source of Lanai Volcano [West et al., 1992].

Cheng et al. [1993] and Duncan et al. [1994] found that the temporal evolution of Tahiti follows the Hawaiian pattern in which eruption of alkaline magmas with depleted isotopic signatures follows a shield-building phase that consists of eruption of tholeiitic magmas with enriched isotopic signatures. The old shield-building lavas were created by the largest degrees of melting and have the most enriched isotopic signatures. As time passed, degree of melting decreased and isotopic signatures became more MORB-like. Duncan et al. [1994] argued that the temporal, compositional and isotopic variations reflect a progressive change from large degree melts derived from the hot plume core in the shield-building stage to smaller degree melts of the cooler plume sheath in late stages. The plume sheath in their view consists of material with more depleted composition that is viscously entrained during plume rise. In constrast, Hoernle et al. [1991] found no simple temporal pattern in data from Gran Canaria in the Canary Islands.

Plume magma composition may be influenced by a nearby ridge, as well as via versa. That is the conclusion Weis and Frey [1991] from a study of the basalts from the Ninetyeast Ridge, produced by the Kerguelen mantle plume. The Southern Kerguelen Plateau was created 115 Ma ago by the same plume. Because the Antarctic Plate has been nearly stationary, the Plateau was probably never far from the hotspot. Throughout this long 115 Ma history, the plume has maintained its distinctive isotopic characteristics [Weis et al., 1993].

Current theory holds that plumes initiate with large bulbous heads that produce flood basalts episodes when the heads reach the surface. Further evidence to support this idea was reported by Mahoney et al. [1991], Madagascar was over the Marion/Prince Edward plume when voluminous basalts erupted there about 90 million years ago, and Madagascar basalts bear some isotopic affinities to the Marion ones.

Oceanic plateaus also appear to be produced in this way. Isotopic affinities of Ontong-Java basalts and those from the Louisville Ridge seamounts are consistent with the Louisville plume, currently located near the intersection of the Eultan Fracture Zone and the Pacific-Antarctic Rise, having created the Ontong-Java and Manihiki Plateaus [Mahoney and Spencer, 1991; Mahoney et al., 1993]. Cretaceous basalts from the Nauru Basin have isotopic compositions displaced toward MORB compared to Ontong-Java basalts. Castillo et al. [1991] suggested that the former were produced at a spreading center influenced by the Ontong-Java/Louisville plume.

The South-Central Pacific is an unusual region in several respects. Mantle plumes appear to be particularly common, and anomalously shallow bathymetry suggests high mantle temperatures. Staudigel et al. [1991] found that extreme isotopic compositions, which characterize modern basalts from this region, also characterized Cretaceous basalts of the Magellan, Marshall and Wake seamounts, suggesting this “South Pacific Isotopic and Thermal Anomaly” has existed at least since the Cretaceous. The Line Islands, also mostly Cretaceous, show a particularly complex history [Garcia et al., 1993]. In some parts of the chain, three phases of volcanism occurred. The earliest basalts can be spatially, spatially, and compositionally associated with the Easter Island plume. However, the other phases of volcanism, which have similar isotopic characteristics, cannot be backtracked to a known hotspot.

It has been apparent for some time that there is an underlying simplicity to the isotope systematics of oceanic basalts: they can be divided into 4 or 5 “species” on the basis of isootope ratios [White, 1985]. Weaver [1991] found that these groups can also be distinguished on the basis of trace element ratios. As with isotopic ratios, no single ratio is diagnostic, but combinations of ratios are. Weaver found HIMU (HIMU derives from “high-μ”, where μ is 238U/204Pb) can be distinguished from EM I and EM II by K/Nb < 180, Ba/Nb < 5, Ba/La < 9 and Ba/Th < 80 (“EM” stands for “enriched mantle”). EM I has Ba/Th between 100 and 150, while EM II has Ba/Th < 85.

Re-Os Isotope Systematics

A significant development of the past 4 years has been the renewed effort at determination of Os isotope ratios. Mass spectrometry problems with Os isotopic analysis were largely overcome by development of a new analytical technique [Croser et al., 1991], so that very small quantities of Os can be accurately analyzed with relatively minor modification to a conventional mass spectrometer.

Studies using the older analytical techniques had already revealed much about Os isotope systematics. It appears that the bulk silicate Earth has a 187Os/186Os of about 1.05-1.10 [e.g., Martin, 1991]. This value is within the range of chondritic meteorites, a somewhat surprising result since Re and Os have very different silicate/metal partition coefficients. The depleted mantle appears to have 187Os/186Os somewhat lower than this, and oceanic island basalts somewhat higher 187Os/186Os, consistent with other isotope systematics. Os studies using the new analytical technique are now revealing the details of Os isotope variability in the mantle. Os isotope ratios from the Austral Islands, which have HIMU characteristics, have high and uniform 187Os/186Os (1.1-1.3) [Hauri and Hart, 1993; Reisberg et al., 1993], while Samoa and Tahaa (both EM I) fall in the range of 1.03-1.16. Samples from Kerguelen and Pitcairn (both EM I) have 187Os/186Os of 1.36-1.54. Hauri and Hart [1993] and Reisberg et al. [1993] concluded these results were consistent with plumes containing a component of recycled material.

Reisberg et al. [1993], however, raised a warning about Os isotope ratios in ocean island basalts. Most basalts from St. Helena and the Comores (both HIMU type islands) had 187Os/186Os in the same range as the Australs, but some samples have very much higher 187Os/186Os, up to 1.95. Reisberg et al. [1993] suggested these high ratios may reflect assimilation of Os-rich sediment, such as Mn oxides, during magma ascent. Because Os concentrations are so low, small amounts of assimilation could shift the 187Os/186Os ratio without having a detectable effect on Sr, Nd, and Pb isotope ratios.

Noble Gas Studies

Noble gases are of interest because they provide information about the Earth’s degassing history. In MORB, the geochemical effects of ridge plumes is clearly demonstrated in He isotope ratios [e.g., Graham et al., 1992; Poreda et al., 1993]. Among oceanic islands, the temporal trends observed in Mauna Loa [Kurz and Kammer, 1991] contrast with temporal homogeneity on Reunion [Graham et al., 1990]. In the Juan Fernandez Islands, basalts from the shield-building stage have more primitive He than post-shield alkalic lavas [Farley et al., 1993], a pattern reminiscent of Hawaii.

For the other noble gases, there is now evidence that atmospheric contamination may be more of a problem than had been previously supposed. Farley and Poreda [1993] found that all oceanic islands as well as MORB define 20Ne/22Ne near the solar value, but plumes and depleted mantle have different 21Ne/22Ne (21Ne is produced in the mantle by 18O(α,n)21Ne and 22Mg(α,n)21Ne reactions). Contamination with atmospheric Ne during or after eruption then produces the observed arrays. Farley and Craig [1992] concluded Ar isotope variability in olivine phenocrysts from Juan Fernandez was due to dilution of...
radiogenic mantle Ar with assimilated atmospheric Ar.

Other studies have begun to resolve the relationship of He isotope ratios to other isotope ratios. For example, it is now clear that the St. Helena, or HIMU, “species” is characterized by low 3He/4He [Graham et al., 1991]. In Samoa and the Galapagos, the highest 3He/4He ratios are associated with intermediate Sr, Nd, and Pb isotope ratios [Graham et al., 1993; Farley et al. 1992]. Farley et al. [1992] proposed that high 3He/4He ratios were associated with a component they called PHEM (“primitive helium mantle”).

The Common Component of Plumes

There was much interest in the 1980’s in the “phylogeny” of plumes, i.e., defining “species” and end members and deducing their origin. The focus in the last few years has been on a possible common component in plumes. Hart et al. [1992] applied principle component analysis to a subset of the oceanic basalt isotope data and found that three eigenvectors can account for 97.5% of the variance, confirming that mantle plumes can be modeled as mixtures of just a few components. Most of the data thus plot within a tetrahedron in 87Sr/86Sr–143Nd/144Nd–206Pb/204Pb space. Viewing the data in this way, Hart et al. [1992] concluded that most arrays converge on a region they referred to as the “Focus Zone” or “FOZO” at the base of the tetrahedron (Figure 1). FOZO has an isotopic composition similar to that of MORB, except that the Pb is more radiogenic than MORB. FOZO, the PHEM component of Graham et al. [1992], and the “C” (common) component of Hanan and Graham [1994] may well be one and the same. Conceptually, all refer to a component that is common to many plumes and that has intermediate Pb isotope ratios and high 3He/4He.

There is some merit to the idea of a common component to many plumes. Many plumes, as well as MORB, define arrays that extend towards intermediate compositions on isotope-isotope plots. This is evident, for example, in the plot of 87Sr/86Sr vs. 206Pb/204Pb shown in Figure 2. Except for the Cook-Australis (which may be the products of several plumes), no volcanic island chain appears to cross this middle region, suggesting at least the possibility that a composition in the middle is actually an end-member.

Hart et al. [1992] argued that FOZO represents the compo-

![Figure 1. Three dimensional plot of selected oceanic basalt data in 87Sr/86Sr, 143Nd/144Nd, and 206Pb/204Pb space. DMM is “Depleted MORB Mantle”. Adapted from Hart et al. [1992].](image)

![Figure 2. 87Sr/86Sr vs. 206Pb/204Pb for selected oceanic basalts. A&P MORB is Atlantic and Pacific MORB. Dark gray region in the center indicates the apparent composition of the common component of plumes, variously designated PREM, FOZO, and “C”.](image)

References


