Annu. Rev. Earth Planet. Sci. 2002. 30:493–525
DOI: 10.1146/annurev.earth.30.091201.141236
Copyright © 2002 by Annual Reviews. All rights reserved

MANTLE MIXING: The Generation, Preservation, and Destruction of Chemical Heterogeneity

Peter E. van Keken[1], Erik H. Hauri[2], and Chris J. Ballentine[3]
[1]*Department of Geological Sciences, University of Michigan, Ann Arbor, Michigan 48109-1063; e-mail: keken@umich.edu*
[2]*Department of Terrestrial Magnetism, Carnegie Institution of Washington, 5241 Broad Branch Road, NW, Washington DC 20015; e-mail: hauri@dtm.ciw.edu*
[3]*Department of Earth Sciences, University of Manchester, Manchester M13 9PL, United Kingdom; e-mail: cballentine@fs1.ge.man.ac.uk*

Annu. Rev. Earth. Planet. Sci. 2002.30:493-525. Downloaded from arjournals.annualreviews.org
by UNIVERSITY OF OSLO on 03/30/06. For personal use only.

Key Words mantle geochemistry, plumes, ocean island basalts, mantle heterogeneity, seismology, noble gases, layered mantle, trace elements, radiogenic isotopes

■ **Abstract** Observations of the geochemical diversity of mid-oceanic ridge and ocean-island basalts have traditionally been attributed to the existence of large-scale mantle heterogeneity. In particular, the layered convection model has provided an important conceptual basis for discussing the chemical evolution of the Earth. In this model, a long-term boundary is assumed between a well-mixed and depleted upper mantle and a heterogeneous and more primitive lower mantle. The existence of high $^3He/^4He$ in ocean-island sources has been used to argue for the preservation of a primitive component in the deep mantle. Nevertheless, a primitive deep layer is difficult to reconcile with the abundant lithophile isotopic evidence for recycling of oceanic crust and the lack of preservation of primitive mantle. In addition, the widespread acceptance of geophysical evidence for whole mantle flow has made straightforward application of the layered convection model problematic. Model calculations show that whole mantle convection with present day heat flow and surface velocities is sufficiently vigorous to mix large-scale heterogeneity to an extent that is incompatible with the geochemical observations. Several concepts have been proposed in recent years to resolve the apparent conflicts between the various observational constraints and theoretical interpretations. The suggestions include the presence of deeper layering, preservation of highly viscous blobs, core mantle interaction, and strong temporal variations in mantle dynamics. Although these models generally appear to solve parts of the puzzle, at present no single model is able to account for all of the major observations. The reconciliation of conflicting evidence awaits improvements in observational and experimental techniques integrated with better model testing of hypotheses for the generation and destruction of mantle heterogeneity.

Annu. Rev. Earth. Planet. Sci. 2002.30:493-525. Downloaded from arjournals.annualreviews.org
by UNIVERSITY OF OSLO on 03/30/06. For personal use only.

INTRODUCTION

Historically, the two-layer mantle model has worked well to explain the main geochemical observations regarding the mantle distribution of radiogenic heat sources, noble gas and trace element concentrations, and the provenance of ocean-island and mid-oceanic ridge basalts. This conceptual model is implicit in almost all geochemical literature addressing the distribution of trace elements and the evolution of the mantle. However, the recent widespread acceptance of the various geophysical evidence for whole mantle convection and improved geochemical observations make a reevaluation of mantle mixing models necessary. We are facing the major challenge of reconciling the geochemical constraints for isolated mantle reservoirs with the geophysical observations of whole mantle convection. The fundamental rethinking of the role of the mantle convective system makes an update and expansion of previous review papers (Zindler & Hart 1986, Silver et al. 1988, Carlson 1994, Hofmann 1997) timely.

PHYSICAL AND CHEMICAL OBSERVATIONS OF MANTLE HETEROGENEITY

Geochemical Observations Supporting Models of Layered Mantle Convection

The long-term chemical heterogeneity of the mantle was demonstrated by the earliest surveys of radiogenic isotopes in oceanic basalts (Faure & Hurley 1963, Hedge & Walthall 1963, Gast et al. 1964). Subsequent detailed studies, particularly at Iceland (e.g., Hart et al. 1973; Schilling 1973, 1975; Zindler et al. 1979; Poreda et al. 1986), revealed several fundamental geochemical differences between mantle-derived magmas erupting at ridges and hot spots. These differences were consistent with a fairly simple model, in which the source of mid-oceanic ridge basalts (MORBs) had a prior history of melt extraction to form the continental crust, whereas the variable geochemistry of ocean-island basalts (OIBs) was explained by mixing of the depleted upper mantle with primitive mantle (cf. DePaolo & Wasserburg 1976, O'Nions et al. 1979, Allègre et al. 1979).

This conceptual model essentially defines three different reservoirs. Primitive mantle is identical to the bulk silicate Earth (BSE), which is the bulk Earth composition minus the core. Survival of a primitive reservoir requires therefore that it has remained a closed chemical system since the early accretion and differentiation of the Earth. Continental crust is the product of melt extraction from a portion of the original primitive upper mantle, which left the depleted upper mantle as the third reservoir.

The interpretation of noble gas isotopes has historically formed a compelling case for the preservation of primitive material in the mantle system. Because noble gases are highly incompatible and volatile, they are expected to be heavily outgassed during melting. Although heavier noble gases from the atmosphere may

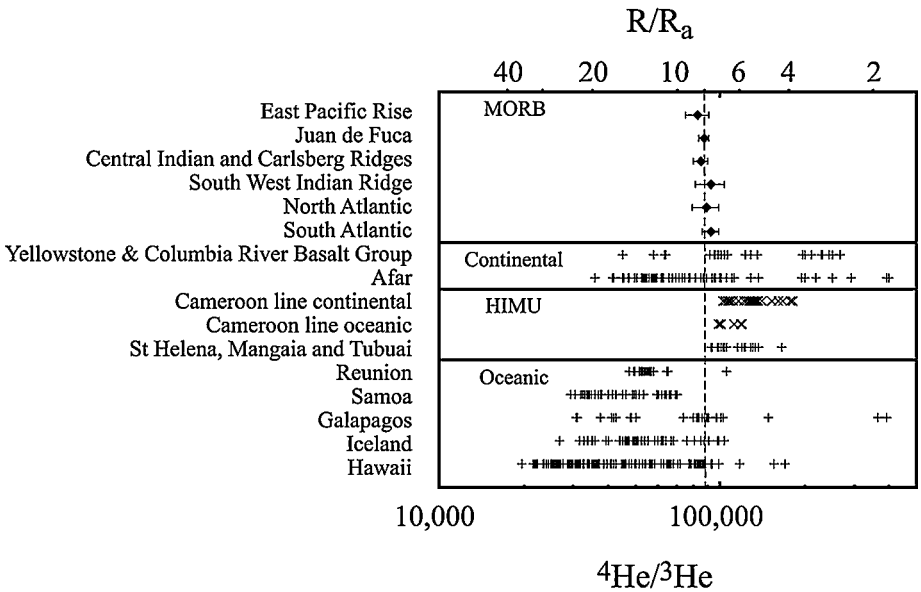


Figure 1 Summary diagram for $^4He/^3He$ from selected mid-ocean ridges, ocean islands, continental hot spots, and the Cameroon volcanic chain after Barfod et al. (1999). This figure illustrates the homogeneity of the worldwide mid-ocean ridge $^4He/^3He$ isotope source. With the exception of HIMU basalts containing radiogenic Pb, ocean islands range from mid-ocean ridge $^4He/^3He$ values to lower $^4He/^3He$ (higher R/R_a) and have been used as one of the principal pieces of evidence to support a layered mantle system.

form a small subducted component (e.g., Porcelli & Wasserburg 1995a), He is lost from the atmosphere with a half-life of some 10^6 years (e.g., Allègre et al. 1987, Torgersen 1989) and cannot be recycled. Although ^4He in the mantle is dominated by the α-decay of U and Th, ^3He is not produced in significant amounts by radioactive processes. ^3He emissions at mid-ocean ridges and hot spots can only be accounted for by the preservation of a volatile component in the mantle since its formation. Many hot spots have higher $^3He/^4He$ ratios than the almost uniform mid-ocean ridge values (Figure 1). We follow the common convention to define $R = {}^3He/^4He$ and express this value as a ratio to the present day value of air (R_a). The observation of high R/R_a has been used to argue for a primitive volatile rich source for these hot spots (Kurz et al. 1982, Allègre et al. 1987, O'Nions 1987). It has often been an assumption to equate this primitive volatile-rich source to a primitive mantle reservoir, as the preservation of ^3He and associated primitive noble gases such as Ne (Honda et al. 1991, Poreda & Farley 1992) appears to require isolation from surface melting.

The concentration and flux of noble gases provide additional constraints on understanding noble gas evolution in the mantle. The MORB source mantle He concentration can be calculated from the observed ^3He flux of ~1000 Mol ^3He/year

Annu. Rev. Earth. Planet. Sci. 2002.30:493-525. Downloaded from arjournals.annualreviews.org
by UNIVERSITY OF OSLO on 03/30/06. For personal use only.

Annu. Rev. Earth. Planet. Sci. 2002.30:493-525. Downloaded from arjournals.annualreviews.org
by UNIVERSITY OF OSLO on 03/30/06. For personal use only.

into the oceans (Lupton & Craig 1975, Farley et al. 1995), combined with the average production of basalt at ridges (Parsons 1981) and an assumption of the extent of MORB melting. An estimate of 20% partial melting at mid-oceanic ridges gives a MORB source mantle He concentration of 1.3×10^9 atoms ^{3}He/g. O'Nions & Oxburgh (1983) noted that the rate of outgassing of the associated radiogenic ^{4}He was similar to the amount produced in the MORB-source above 670 km. This flux is far lower than that predicted from the surface heat flow component produced by total mantle U + Th decay. The imbalance was explained by a boundary layer at 670 km that allowed heat but very little ^{4}He to pass through from a deeper radioelement-rich reservoir. These workers further suggested that this deep layer provided the source of the ^{3}He required to buffer the extensively outgassed upper mantle. A steady state model assuming diffusion of He only through this boundary layer was presented by Allègre et al. (1987). Concurrent work on the difference between the U/Th ratio derived from the Pb and Th isotopes in OIB and MORB resulted in a model that postulated a small but significant mass flux from the lower mantle into the upper mantle (Galer & O'Nions 1985). This in turn prompted the formulation of a similar steady state concept for the noble gases, with an advective transfer from the lower to upper mantle formulated first for He (Kellogg & Wasserburg 1990); followed by Ne and Ar (O'Nions & Tolstikhin 1994); and an elegant and complete formulation for He, Ne, Ar, Kr, and Xe that includes consideration of a return subduction flux for the heavy noble gases (Porcelli & Wasserburg 1995a,b).

The steady state models enable the ^{3}He concentration in the lower mantle to be calculated with the assumption that the lower mantle has primitive (or BSE) U + Th concentrations and has essentially remained a closed system for 4.5 Ga. From an assumed initial solar ^{3}He/^{4}He ratio of 120R_a, this reservoir has evolved to the observed ^{3}He/^{4}He ratio of $\sim$35R_a, which is equal to the highest values observed at both Hawaii and Iceland. This predicts a present day concentration of $\sim 7 \times 10^{10}$ atoms ^{3}He/g in the lower mantle (Allègre et al. 1987, Kellogg & Wasserburg 1990, O'Nions & Tolstikhin 1994, Porcelli & Wasserburg 1995a). This value is substantially higher than that of the MORB source and compares with the range found in CI chondrites of 7×10^{11} to 10^{14} atoms ^{3}He/g (Mazor et al. 1970).

A separate requirement for some form of compositional stratification in the mantle comes from consideration of Earth's heat loss and its radiogenic heat budget. Expressed in terms of the present day heat production, the BSE budget of U, Th, and K generates 19 TW (Van Schmus 1995). About 6–8 TW can be accounted for by the radiogenic heat production in the continental crust (Rudnick & Fountain 1995). The MORB-source is strongly depleted and would produce only $\sim$3 TW if it would occupy the entire mantle. This suggests a storage of radiogenic heat producing elements below the continental crust and MORB-source, possibly in a layer in the deep interior. A separate and more indirect argument is based on the heat balance. The present heat loss of 44 TW (Pollack et al. 1993) minus the BSE heat production and a reasonable estimate for the core mantle boundary heat flow of 3–7 TW (Buffet et al. 1996) leaves approximately 17–22 TW for secular cooling of the mantle, which translates to an average cooling rate of 110–130 K/Ga

Annu. Rev. Earth. Planet. Sci. 2002.30:493-525. Downloaded from arjournals.annualreviews.org
by UNIVERSITY OF OSLO on 03/30/06. For personal use only.

(e.g., Van Keken et al. 2001). This is much higher than the change in upper mantle temperature that is inferred from the geological record (up to 50 K/Ga; Abbott et al. 1993). A possible solution to this discrepancy is that past radiogenic heating is buffered in the deep mantle and not released efficiently to the surface (McKenzie & Richter 1981). This again requires some form of barrier to mantle flow.

In summary, noble gases support the existence of a significant reservoir in the mantle that has preserved high concentrations of volatiles since the formation of the mantle. Mass and heat balance considerations add weight to the idea of long-term preservation of an isolated and large reservoir. The simplest model to explain these observations is through layered mantle convection with almost perfect isolation between the top and bottom parts of the mantle. Although the layered model has been a useful abstraction of the mantle system, it has become increasingly clear that strict application of this model leads to substantial conflicts with various lines of geochemical and geophysical evidence. In the next sections, we review the evidence that necessitates a substantial reworking of the layered mantle model, first, by discussing tests of layered model predictions of lower mantle He, followed by the mantle mixing evidence from lithophile elements and an overview of the geophysical evidence for whole mantle flow.

PRESERVATION OF PRIMITIVE MANTLE?

Helium isotope measurements on oceanic basalts have demonstrated clearly that the upper mantle MORB source has evolved with high ratios of $U/^3He$ compared to hot spots, consistent with the history of upper mantle depletion recorded by $^{87}Sr/^{86}Sr$ and $^{143}Nd/^{144}Nd$ in MORB (e.g., DePaolo & Wasserburg 1976, O'Nions et al. 1979, Allègre et al. 1979, Kurz et al. 1982). The complement of this idea is that the undegassed and primitive volatile-rich component corresponds to the actual preservation of primitive mantle. However, several lines of evidence do not support this assumption. For example, within the assumption of layered convection, the noble gas data alone would appear to constrain the amount of exchange between the upper and lower mantle of 2%–4% over the age of the Earth, which is significantly less than that indicated by the Sr-Nd-Hf-Pb isotopes (~10%; Allègre 1997). More significantly, many studies have contributed to the recognition that most geochemical observations are not compatible with the preservation of primitive mantle. These include estimates of OIB-source ^3He concentrations that do not support the layered model predictions, the identification of enriched components in OIB sources, the widespread distribution of mantle sources (in both space and time) with homogeneous and nonprimitive ratios of key trace elements, and the recognition of a history of melt depletion in the high $^3He/^4He$ mantle source.

Helium Concentrations in the OIB-Source

A fundamental prediction from the layered mantle model is that the OIB-source is helium-rich and has a concentration of approximately 7×10^{10} atoms ^3He/g. It has

been widely noted that OIB samples in general contain lower gas concentrations than MORB (see review in Farley & Neroda 1998). This has in part led other workers to suggest that high $^3He/^4He$ in OIB reflects low ^4He rather than high ^3He concentrations (e.g., Anderson 1998a, Albarède 1998). A critical test of layered and alternative mantle models is the determination of OIB source ^3He concentration, which can be investigated by (*a*) combining the magma generation rate with the ^3He flux estimated from hot spot volcano CO_2 flux, (*b*) direct measurement of He concentrations in basalt samples and reconstruction to preeruption concentration, and (*c*) identification of isotopic mixing to resolve OIB end-member noble gas concentrations.

Unlike the mid-ocean ridge system, there is no data available to directly constrain the global OIB ^3He flux. At present, studies using ^3He and magma flux are limited to specific sites. For example, the summit CO_2 flux combined with observed $CO_2/^3He$ values and estimated magma production rates at Kilauea, Hawaii, provides a local estimate for the undegassed magma ^3He concentrations of 1.26×10^{10} atoms ^3He/g (Hilton et al. 1997). Assuming a 7% melt fraction, this would give a source concentration of 0.09×10^{10} atoms ^3He/g, some 80 times lower than the layered mantle prediction.

Direct measurement of He concentration in Hawaiian basalts or estimates based on basalt CO_2 content and $CO_2/^3He$ ratios give similarly low source concentrations ($0.09–0.27 \times 10^{10}$ atoms ^3He/g, Torgersen 1989). This must be considered a minimum because of the effects of basalt degassing. Closed system vesiculation and subsequent loss provides an explanation for both noble gas, CO_2, and N_2 patterns in a gas-rich MORB glass sample (e.g., Sarda & Graham 1990, Javoy & Pineau 1991, Pineau & Javoy 1994). He, Ne, and Ar abundances that are corrected for atmosphere contamination (e.g., Ballentine & Barfod 2000, Ballentine et al. 2001) in a more extensive filtered MORB data set also reflect closed system vesiculation and subsequent gas loss (Moreira & Sarda 2000). The lower melt fraction, greater mean depth of melting, and shallower mean eruption depth at hot spots compared to mid-oceanic ridges result in a different style of degassing. Moreira & Sarda (2000) show, for example, that a filtered OIB data set agrees better with an open system Rayleigh degassing model. From model calculations, He concentrations in the OIBs are reduced by only 10% during the period of open system degassing. To account for the observed range in He concentrations, Moreira & Sarda (2000) propose that a secondary and unfractionating gas loss process also occurs. In this context, the highest observed concentrations place a lower limit on original magma (1.3×10^{10} atoms ^3He/g) and OIB source (0.09×10^{10} atoms ^3He/g) concentration. The latter is 74 times lower than the layered mantle prediction.

The third approach to determine the OIB source ^3He concentration has focused on the relationship between $^3He/^4He$ and Pb-Nd-Sr isotopes. In the case of simple two component mixing between the plume source and depleted MORB mantle (DMM), the mixing constant K, for the example of lead, is given by $K = (He/Pb)_{plume}/(He/Pb)_{DMM}$. If the plume Pb concentration is known, then the He and Pb concentration in DMM defines the He concentration in the plume source.

Annu. Rev. Earth. Planet. Sci. 2002.30:493-525. Downloaded from arjournals.annualreviews.org
by UNIVERSITY OF OSLO on 03/30/06. For personal use only.

Annu. Rev. Earth. Planet. Sci. 2002.30:493-525. Downloaded from arjournals.annualreviews.org
by UNIVERSITY OF OSLO on 03/30/06. For personal use only.

It is now clear that the depleted components in intraplate high $^3He/^4He$ (Hart et al. 1992) and low $^3He/^4He$ systems (Ballentine et al. 1997, Barfod et al. 1999) are not simple DMM. Some systems in ridge-centered hot spots, such as Iceland, can be ascribed to DMM-plume mixing. Hilton et al. (2000) use $^3He/^4He$ and $^{206}Pb/^{204}Pb$ correlation across the deeper section of the Reykjanes ridge to show that this defines a mixing line where $K = 1$ (Figure 2). In the limiting case where the Pb concentration of the plume is the same as that of DMM (Pb ~0.05 ppm), it follows that $He_{plume} = He_{DMM}$. The $^3He/^4He$ of the plume end-member is four times higher than the DMM component, and consequently, the ^3He concentration in the plume is at least four times higher than of the DMM source and has a ^3He concentration of 0.5×10^{10} atoms $^3He/g$ (14 times lower than the layered mantle estimate).

Because each technique of ^3He determination is subject to significant uncertainty, we have taken the approach of identifying the minimum OIB-source concentrations. Clearly, the CO_2 flux may undergo significant temporal variance (e.g., Hilton et al. 1998). Similarly, identification of the process causing secondary gas loss in OIB samples may enable sample types or collection procedures to be identified that return yet higher He concentrations. In the mixing determinations, the plume-source Pb concentration remains unknown. For example, assumption of primitive mantle concentrations (Pb = 0.185 ppm) increases the ^3He concentration estimate by a factor of 3.7 (Hilton et al. 2000). There has been no attempt to test the statistical acceptability of higher mixing constants, and therefore higher ^3He concentrations, without discarding data. Furthermore, we cannot discount the possibility that the OIB source has undergone gas loss before mixing. By themselves, none of these approaches can rule out the possibility of an OIB source with ^3He concentrations compatible with the layered mantle model. Nevertheless, three independent and different approaches define a coherent picture of minimum OIB ^3He concentrations that are over an order magnitude smaller than the predicted layered mantle model value. It is entirely possible that the OIB source has ^3He concentrations only marginally higher than the MORB source.

Enriched Mantle Sources

Geochemical studies of oceanic basalts at several hot spots such as Walvis Ridge-Tristan (e.g., O'Nions et al. 1977, Richardson et al. 1982, White & Hofmann 1982), Hawaii (e.g., Stille et al. 1983; Roden et al. 1984, 1994), Pitcairn (e.g., Woodhead & McCulloch 1989), Samoa (e.g., White & Hofmann 1982, Palacz & Saunders 1986, Wright & White 1987, Farley et al. 1992), and Cook-Australs (e.g., Palacz & Saunders 1986, Nakamura & Tatsumoto 1988, Woodhead 1996) have demonstrated the existence of enriched mantle components whose time-integrated Rb/Sr, Sm/Nd, and/or U + Th/Pb ratios were actually higher than primitive mantle. An important implication of these studies was that the apparent primitive compositions could be explained simply as various mixtures of enriched and depleted sources. Further survey work on hot spots revealed a large number of distinct hot spot components, which could be circumscribed by a small number of mantle

Annu. Rev. Earth. Planet. Sci. 2002.30:493-525. Downloaded from arjournals.annualreviews.org
by UNIVERSITY OF OSLO on 03/30/06. For personal use only.

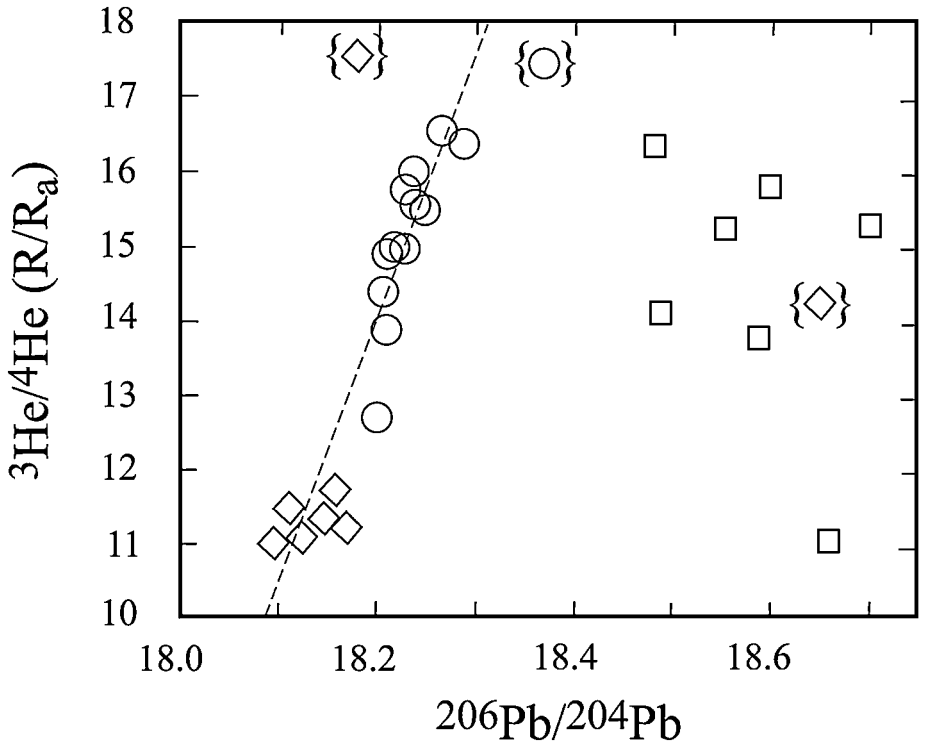


Figure 2 Assumed two component mixing between MORB-source mantle and the Iceland plume source after Hilton et al. (2000). Samples are distinguished by location on the Reykjanes Ridge (*diamonds* are south of 58.26°N, *circles* between 58.26° and 61.3°N, and *squares* north of 61.3°N). Bracketed samples are excluded by Hilton et al. (2000) from mixing consideration, as are samples north of 61.3°N that have shallow eruption depth and high water content. A linear plot defines a mixing constant with $K = 1$ ($r^2 = 0.87$, $n = 18$) and therefore a ^3He concentration in the Iceland plume source that is at least four times higher than the MORB-source mantle (see text). Mixing constants with $K > 1$ can be fit to the data, but this requires the exclusion of all data from samples south of 58.26°N.

end-members with enriched geochemical signatures (cf. White 1985, Zindler & Hart 1986, Allègre et al. 1987). Zindler & Hart (1986) showed that the entire isotopic database for the oceanic mantle could be described by the DMM and three enriched components (EM1, EM2, HIMU), with all other compositions created by mixing. Hart (1988) reviewed the possible mixing relationships between these various components. Parallel work on mantle xenoliths and peridotite massifs from the continental lithosphere revealed evidence for melt-fluid-rock interactions (so-called mantle metasomatism) that can create enriched compositions whose current (or time-evolved) isotopic compositions resemble many of the enriched

mantle end-members (cf. Zindler & Hart 1986 and references therein). Although these studies rarely showed any regional systematics or correlations, and displayed unusually large-amplitude heterogeneity at small spatial scales, it was apparent that none of these mantle samples had the composition of primitive mantle. Even mantle samples with apparently unmodified major element compositions have trace element and/or isotopic compositions quite unlike primitive mantle (Jagoutz et al. 1979; Zindler & Jagoutz 1988a,b; Galer & O'Nions 1989).

Constancy of Nb/U and Ce/Pb

Hofmann (1986) and Newsom et al. (1986) were the first to show that some trace element ratios in oceanic basalts are both constant and nonprimitive. The abundances of Nb, U, and Ce in the primitive mantle can be accurately estimated from their abundances in chondritic meteorites, and the Pb content of the primitive mantle is constrained from the U abundance and the well-understood isotopic evolution of Pb in the Earth. Constant values of Nb/U and Ce/Pb in mantle basalts derived over a large range in melting extent show that these elements behave in ways essentially identical to one another during partial melting, and thus represent the ratios of the mantle source. Constant values would be expected for an unmodified primitive mantle that is preserved over time, but the observed ratios are distinctly nonprimitive. Similarly, Sn is moderately siderophile with respect to Sm, but the Sn/Sm ratio is not fractionated during crust formation processes. Consequently, the observed Sn/Sm, which is lower than the primitive value, indicates that all mantle reservoirs have undergone a similar degree of fractionation caused by core formation processes (Jochum et al. 1993). Studies of ancient basalts and komatiites (Jochum et al. 1991, 1993) have shown that these constant ratios persist back at least 3.4 billion years. Mass balance considerations show that the conjugate budget of these elements resides largely in the continents, and the combined observations favor formation of the bulk of the Earth's continental crust in the first billion years of Earth history (Hofmann 1988). A fundamental conclusion of these studies is that no reservoir currently exists on the Earth that yields Nb/U, Nb/Th, or Ce/Pb ratios similar to that of the primitive mantle.

The Depleted Nature of the High-^{3}He Source

Several studies show clear correlations between He and the lithophile Sr-Nd-Pb isotopes (Kurz & Kammer 1991; Graham et al. 1992, 1996; Farley et al. 1992; Eiler et al. 1998; Kurz & Geist 1999; Barfod et al. 2001). Importantly, these studies vanquished the persistent suggestion from xenolith work that ^{3}He/^{4}He was somehow decoupled from other isotopes during melt generation. At the same time, Hart et al. (1992) made an important reevaluation of mixing arrays among hot spots. The accumulation of isotope data (including ^{3}He/^{4}He) on increasing numbers of hot spot basalts allows three new observations that lead to the recognition that all hot spots are geochemically heterogeneous. The first observation is that each hot spot can be described (to first-order) as a mixture of an enriched and a depleted

Annu. Rev. Earth. Planet. Sci. 2002.30:493-525. Downloaded from arjournals.annualreviews.org
by UNIVERSITY OF OSLO on 03/30/06. For personal use only.

Annu. Rev. Earth. Planet. Sci. 2002.30:493-525. Downloaded from arjournals.annualreviews.org
by UNIVERSITY OF OSLO on 03/30/06. For personal use only.

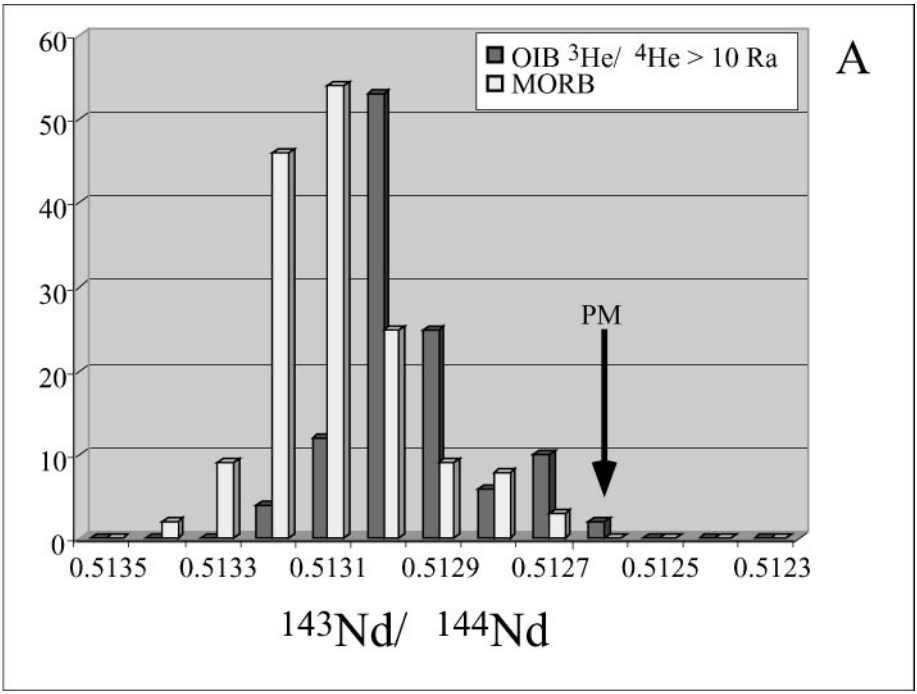
component; mixing arrays for individual hot spots are roughly linear in Sr-Nd-Pb isotope space. Second, all hot spots share a common depleted component that does not have the composition of the MORB source. This source was termed FOZO by Hart et al. (1992), and its appearance at hot spots around the globe requires either a single widespread component and/or a fundamentally important process coupled to the formation of hot spots (Hauri et al. 1994). Finally, the increased amount of helium isotope data allowed Hart et al. (1992) to recognize that the depleted FOZO component and the high-$^3He/^4He$ component were one and the same (Figure 3). Similar conclusions were reached by Hanan & Graham (1996) from a compilation of studies of hot spot-influenced mid-ocean ridges. These systematic observations showed that oceanic basalts with high $^3He/^4He$ ratios were nonprimitive by every other geochemical measure and in fact retained an isotopic history of depletion in Sr and Nd isotopes. The best estimate of the isotopic composition of the high-$^3He/^4He$ (FOZO) component is provided by Hauri et al. (1994), who show that the Sr-Nd isotope signature of FOZO is clearly depleted compared to primitive mantle, but not as depleted as that of the MORB source (Figure 4).

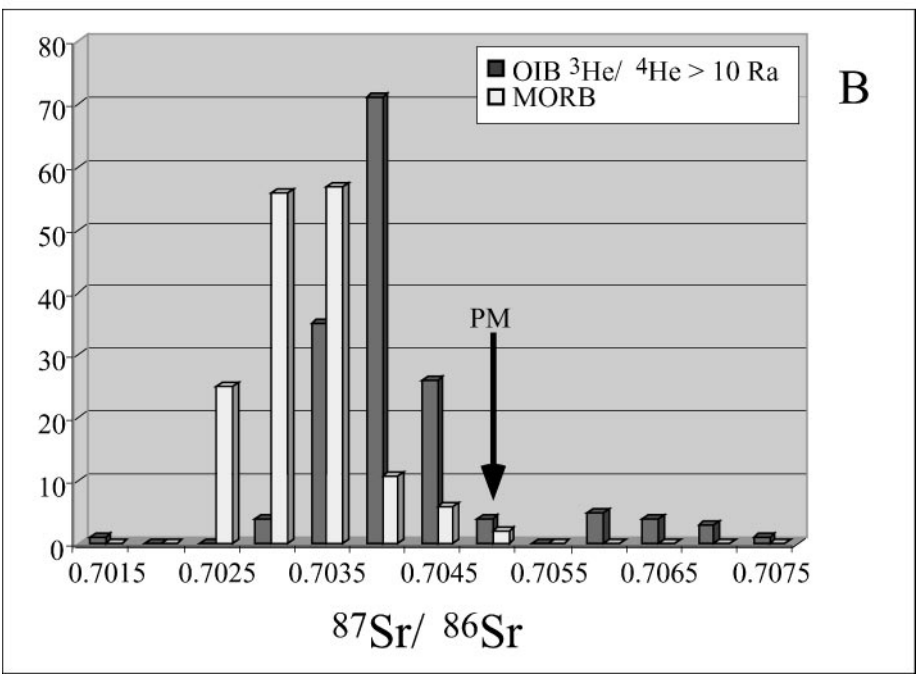
The evidence based on lithophile elements presented thus far suggests the possibility that mantle geochemistry is dominated by plate tectonic recycling of altered oceanic crust with minor additions from the continental lithosphere and possibly a primitive volatile source that may or may not reside in the mantle. This raises an intriguing question: How well does this model interpretation stand up to further tests based on the application of other isotopic measurements?

SMOKING GUNS FOR SLAB RECYCLING: RE-OS AND OXYGEN

New isotopic systems have been developed and applied to measurements of oceanic basalts in recent years, with new insights gained particularly from measurements of hafnium ($^{176}Hf/^{177}Hf$), osmium ($^{187}Os/^{188}Os$), oxygen ($\delta^{18}O$), and neon ($^{21}Ne/^{20}Ne$, $^{22}Ne/^{20}Ne$). The five–end-member description of mantle variability, which includes MORB, FOZO, HIMU, EM1, and EM2 as established by Zindler & Hart (1986), Hart et al. (1992), and Hauri et al. (1994), has not required expansion or revision with the addition of data from these other isotope systems over the past decade (Hofmann 1997). However, new osmium and oxygen isotope data on the mantle

Figure 4 Histograms of (*A*) Nd and (*B*) Sr isotopes for oceanic island basalt samples (OIB) with $^3He/^4He$ ratios $>10R_a$ (*dark bars*) and MORB (*light bars*). OIB samples with high-$^3He/^4He$ have strongly peaked Sr (0.70388 ± 13, $2\sigma_M$) and Nd (0.51290 ± 2, $2\sigma_M$) isotope ratios that are depleted compared to primitive mantle values (*black arrows labeled PM*). A student's T-test analysis shows the MORB and high-$^3He/^4He$ OIB populations to be distinct in both Sr and Nd isotopes.





Annu. Rev. Earth. Planet. Sci. 2002.30:493-525. Downloaded from arjournals.annualreviews.org
by UNIVERSITY OF OSLO on 03/30/06. For personal use only.

Annu. Rev. Earth. Planet. Sci. 2002.30:493-525. Downloaded from arjournals.annualreviews.org
by UNIVERSITY OF OSLO on 03/30/06. For personal use only.

end-members have revealed the limitations of this five-component classification. For recent reviews of the rhenium-osmium and oxygen isotope systems, see Shirey & Walker (1998) and Eiler (2001).

Among the outstanding characteristics of the ^{187}Re-^{188}Os decay system are the large Re/Os ratio acquired by magmas and the complimentary low Re/Os ratio retained by the peridotite residues of mantle melting. The data show a near-perfect separation of Re/Os ratio that correlates directly with rock type, a characteristic not shared by any other isotope system. With sufficient time, these differences in Re/Os ratio translate into large differences in the ^{187}Os/^{188}Os ratio (Figure 5). This makes the ^{187}Os/^{188}Os system a direct tracer of the addition of mafic crust or melt to a mantle source (Hauri & Hart 1993), and Os isotopes have consequently been used to trace the involvement of recycled mafic crust in the sources of hot spots (see references in Shirey & Walker 1998). Seawater contamination and posteruptive radiogenic growth of ^{187}Os have unfortunately complicated similar studies of MORB and as a consequence the Os isotope composition of the oceanic upper mantle is best characterized by data on seafloor peridotites (Roy-Barman & Allègre 1994, Snow & Reisberg 1995).

The relative abundances of oxygen isotopes are dominated by low-temperature mass-dependent fractionation processes and in particular water-rock interactions (Eiler 2001). Recent technical advances using laser-based heating have significantly improved the accuracy of oxygen isotope analyses of silicates, resulting in precise estimates of the homogeneity of oxygen isotopes in upper mantle peridotite (Mattey et al. 1994). One of the results of these improvements, realized only in the past decade, has been a recognition that δ^{18}O data for oceanic basalts correlate with trace element abundances and radiogenic isotope tracers such as Sr, Nd, Pb, and Os isotopes (Eiler et al. 1997). The oxygen isotope variations directly trace those parts of the mantle that have interacted with water at or near the Earth's surface. Together, osmium and oxygen isotopes provide powerful tracers for the recycling of subducted crust into the sources of MORB and OIB (e.g., Lassiter & Hauri 1998, Eiler et al. 2000).

Osmium isotope data for oceanic rocks are shown in Figure 6 with corresponding Pb isotope data. These data show several defining characteristics. First, all of the OIB mixing arrays converge toward the FOZO component rather than the upper mantle, confirming the mixing topology emphasized by Hauri et al. (1994). Second, all OIB have variable and elevated ^{187}Os/^{188}Os, indicating mixing of peridotite and crustal components in their mantle sources. Finally, the enriched mantle sources (EM1, EM2, HIMU) are characterized by elevated ^{187}Os/^{188}Os. Often, these enriched components have δ^{18}O values that are different from upper mantle values (Eiler et al. 1997).

From these data, several observations emerge. Each hot spot data array in Figures 3 and 6 is consistent with mixing between a peridotitic component (FOZO, with low ^{187}Os/^{188}Os and high ^{3}He/^{4}He) and an enriched component with high ^{187}Os/^{188}Os and low ^{3}He/^{4}He, which can be identified as recycled mafic crust. Variations in the Sr, Pb, and O isotopic composition of these enriched components

Annu. Rev. Earth. Planet. Sci. 2002.30:493-525. Downloaded from arjournals.annualreviews.org
by UNIVERSITY OF OSLO on 03/30/06. For personal use only.

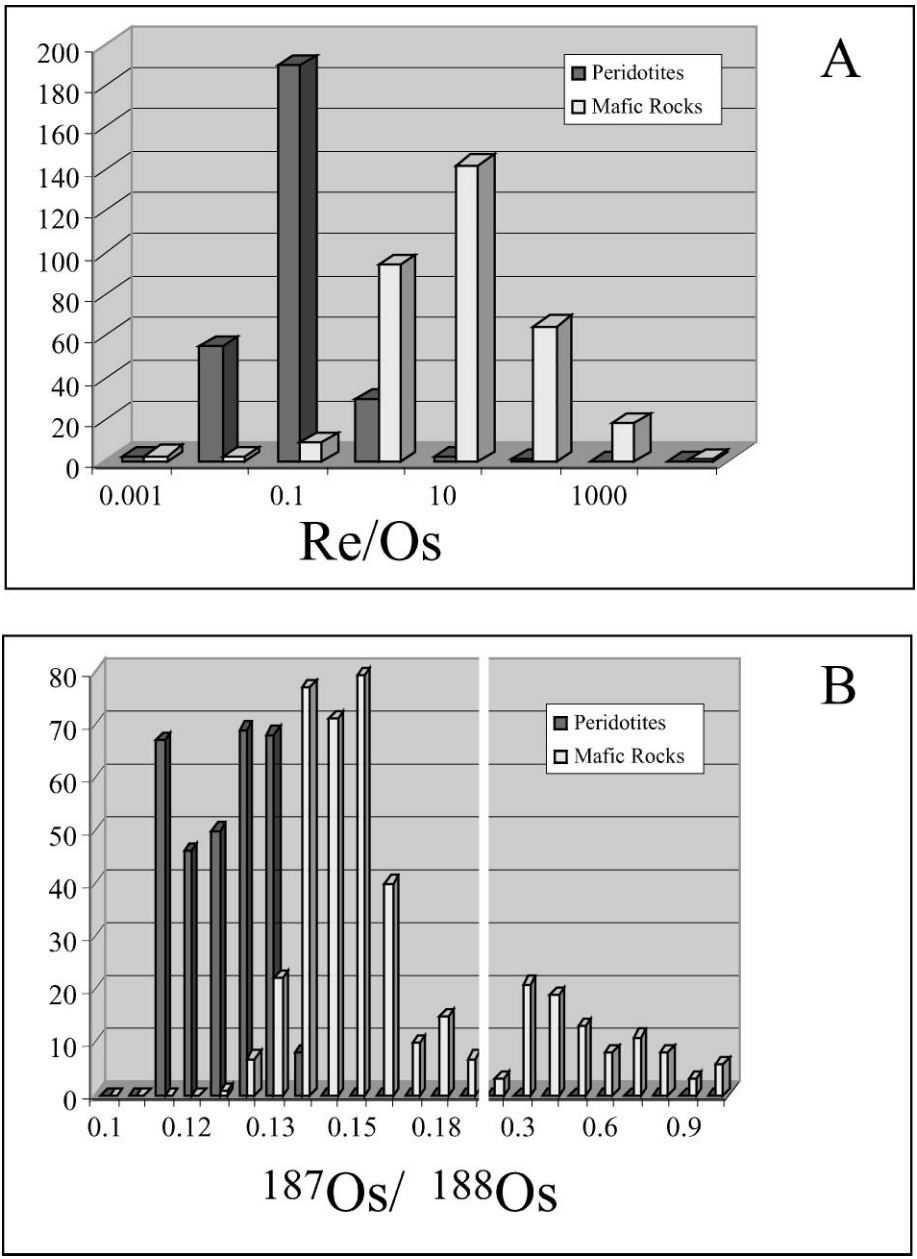


Figure 5 Histograms of (*A*) Re/Os ratio and (*B*) $^{187}Os/^{188}Os$ for peridotites (lherzolites, harzburgites, dunites) and mafic rocks (basalts, pyroxenites, eclogites). Note the change in scale in (*B*). The clear separation of Re/Os ratio between mantle and crustal rocks translates into large differences in Os isotopes with time. The peak in the distribution for mafic rocks in (*B*) is dominated by OIB, which are thought to derive from mixed peridotite-mafic sources (Hauri & Hart 1993, Hauri 1996).

Annu. Rev. Earth. Planet. Sci. 2002.30:493-525. Downloaded from arjournals.annualreviews.org
by UNIVERSITY OF OSLO on 03/30/06. For personal use only.

can be explained by the presence or absence of hydrous alteration and/or sedimentary cover. As a result, the hot spot data are not simply random mixtures of five mantle components, but instead are consistent with each hot spot containing a recycled mafic component that is likely to reflect a unique combination of water-rock alteration, subduction-zone processing, and aging in the convecting mantle.

Taken to its logical conclusion, this model implies both a complication and simplification of the five-component models of White (1985), Zindler & Hart (1986), and Hauri et al. (1994). Complexity arises because the strict number of mantle "components" is probably equal to the number of hot spots that display enriched isotopic signatures. Three oceanic examples of the EM1 end-member are expressed at the islands of Tristan da Cunha, Oahu (Hawaii), and Pitcairn. Basalts from each of these islands have similar Sr, Nd, and Pb isotope ratios, but distinct differences in $\delta^{18}O$ (Eiler et al. 1997, Harris et al. 2000) that strictly argue for different origins. If similar differences exist among other geochemical end-member examples, then it is not likely that the complete spectrum of isotopic heterogeneity can be described simply by mixing of a limited number of end-members.

One simplification is that whereas the true number of mantle components appears to be large, all hot spots display elevated $^{187}Os/^{188}Os$ which requires involvement of mafic sources; a likely origin for these mafic sources is subduction of oceanic crust. A Re-Os mass balance between the depleted upper mantle and continental crust shows that this latter reservoir contains only 10% of the amount of Re depleted from the upper mantle, a discrepancy that can be balanced by assuming a reservoir of oceanic crust in the mantle that amounts to 4%–10% of the mantle mass (Hauri & Hart 1997). Hot spot basalts with high $^{187}Os/^{188}Os$ are the obvious expression of this recycled crustal reservoir. As a result, the classical end-members EM1, EM2, and HIMU are probably not end-members in actual mantle composition but rather reflect end-members in geologic processes that can be identified in the oceanic plate tectonic cycle, such as crust formation, hydrous alteration, sediment addition, and subduction. This may be one reason why large-scale geophysical and geochemical anomalies such as the DUPAL anomaly (Hart 1984) correlate well with mantle return flow constrained by the prior history of subduction over the past 120 Ma (Castillo 1988, Lithgow-Bertelloni & Richards 1998).

Mixing Arrays and Mantle Concentrations

Data arrays in isotopic space usually reflect mixing between distinct mantle or melt components, and in certain cases these data arrays can be constrained by mixing relationships that result in curved mixing trajectories. The sense and degree of curvature is related to differences in the elemental abundance ratios of the two mixing components (cf. Langmuir et al. 1978), and as such, they can be used to estimate the relative abundances of Sr, Nd, Pb, He, and Os in various MORB and hot spot components. For isotopes of Sr, Nd, Pb, and Hf, mixing arrays are usually (but not always) sublinear, indicating similar relative abundances of these elements among all mantle components. For helium, such mixing calculations

usually suggest an apparent He depletion in the high-$^3He/^4He$ FOZO component when the calculations assume MORB as an end-member (Kurz et al. 1982, 1987; Farley et al. 1993). This paradox can be explained if (*a*) the mixing arrays represent mixing of melts, and (*b*) if plume magmas degas more than MORB magmas prior to mixing. This paradox may also be resolved if MORB is not an end-member. For Os, mixing arrays tend to be sublinear but allow low Os concentrations in the enriched sources, which is consistent with the presence of low-Os recycled crust (Hauri 1996, Hauri et al. 1996).

In order to estimate He or Os concentrations from isotopic mixing arrays, assumptions must usually be made about similarity in the relative abundances of Sr, Nd, or Pb in the two mixing end-members. Such assumptions may not be valid if the mixing components are lithologically very different (e.g., peridotite, recycled mafic crust, sediment) or if the mixing arrays reflect mixing of melts from sources with various proportions of these components. In this context, oxygen isotopes play a key role because the abundance of oxygen varies little ($\pm$15%) among various lithological components and can thus be assumed with confidence (James 1981). Mixing relationships with oxygen isotopes should thus directly constrain the absolute abundances of various elements in the end-members of mixing arrays. Hawaii is the only hot spot that has a sufficiently complete data set to perform such a calculation, and the results indicate that the Loihi component (high $^3He/^4He$) contains higher Os contents than does the enriched Koolau component, consistent with the latter being a manifestation of recycled oceanic crust (Hauri 1996, Eiler et al. 1996, Lassiter & Hauri 1998). However, the He-O correlations do not strongly constrain He concentrations.

The reevaluation of mixing arrays and the identification of a subduction component in hot spot basalts push toward whole mantle convection, rather than layering in the strict sense. Until recently, the geophysical observations of mantle structure would permit a layered model with a boundary at 670 km because of the presence of a strong seismic reflector and the cessation of earthquakes at that depth. The next section provides a review on how the geophysical picture of the mantle has changed in recent years, which also suggests a departure from the layered model to a form of whole mantle convection that allows for preservation of long-lived heterogeneity.

Annu. Rev. Earth. Planet. Sci. 2002.30:493-525. Downloaded from arjournals.annualreviews.org
by UNIVERSITY OF OSLO on 03/30/06. For personal use only.

GEOPHYSICAL OBSERVATIONS SUGGESTING WHOLE MANTLE FLOW

In the traditional layered mantle convection model, the strong seismic reflector at 670-km depth has generally been assumed to be the flow barrier. The strong density contrast (10%) and sharpness (less than 10 km; Paulssen 1988) support the interpretation of the 670-km discontinuity as a major chemical discontinuity. Improved experimental mineral physics has provided an important alternative view. For example, diamond anvil studies on the phase transformations in the Mg_2SiO_4-Fe_2SiO_4

Annu. Rev. Earth. Planet. Sci. 2002.30:493-525. Downloaded from arjournals.annualreviews.org
by UNIVERSITY OF OSLO on 03/30/06. For personal use only.

system at 1600°C identified the breakdown of the spinel phase into the denser perovskite and magnesiowüstite phases at 23 GPa (Ito & Takahashi 1989). More importantly, the experiments showed a very narrow transition between the phases, which is consistent with the seismological observations (Paulssen 1988). This effectively removes the need for a major chemical discontinuity in the Earth.

Vast improvements in observational geophysics have further reduced the viability of a present-day chemical barrier in the lower transition zone. In the past decade, a consensus for some form of whole mantle flow has emerged, where slabs drive the large-scale flow and penetrate to the deep mantle, if not to the core-mantle boundary. Of particular interest for the development of this model are observations of the fate of slabs after subduction. The cold, dense slabs provide the main driving force for mantle convection, and hence, the geometry of subducting slabs provides the main indication for the geometry of the overall mantle flow. Early efforts to constrain the flow used the distribution and source mechanisms of earthquakes in Wadati-Benioff zones. Traditionally, the termination of earthquakes at 670 km and the compressive nature of deep earthquakes (Isacks & Molnar 1971) have been used in support of the layered mantle convection model. Yet, the Wadati-Benioff zones are fairly linear belts in most subduction zones down to the maximum depth of seismicity which would be better explained with continued penetration of slabs into the lower mantle (e.g., Davies 1980). The cessation of the earthquakes below 670-km depth is then associated with low yield stress of the lower mantle (Karato 1989). The seismicity data consequently permit a model of whole mantle convection where slabs encounter some form of resistance going into the lower mantle. Several seismological studies supported the existence of aseismic extensions of upper mantle slabs to at least 1000–1200-km depth based on travel time perturbations (Creager & Jordan 1984, 1986) and waveform complexity (Silver & Chan 1986). Early global tomographic models (Dziewonski 1984, Dziewonski & Woodhouse 1987) showed a distinct pattern of high velocity in the deep mantle that could be caused by ancient slabs flattening at the base of the mantle. Model interpretations of the geoid and gravity above subduction zones showed that the geodetic signal could be explained well by slabs penetrating through the 670-km boundary with a viscosity increase into the lower mantle by a factor of 30–100 (Hager & Richards 1989). Similar suggestions for a high viscosity lower mantle arrive from geodynamic inversions (e.g., Mitrovica & Forte 1995). It is interesting to note that a model of mantle circulation, based on a reconstruction of past plate motions, can explain the global seismic tomography quite well when it is assumed that the lithosphere and lower mantle are more viscous than the upper mantle by approximately one or two orders of magnitude (Lithgow-Bertelloni & Richards 1998). This suggests that the large-scale structure of mantle convection could be largely determined by the sinking of slabs into a higher viscosity lower mantle.

The advances in high-resolution seismic tomography have been of particular interest in the discussion about the present-day convective state of the mantle. The quality of these models has been vastly improved in recent years by techniques that make use of detailed waveform modeling (Grand 1994), variable grid spacing

Annu. Rev. Earth. Planet. Sci. 2002.30:493-525. Downloaded from arjournals.annualreviews.org
by UNIVERSITY OF OSLO on 03/30/06. For personal use only.

(Fukao et al. 1992, Bijwaard et al. 1998), precise relocations of hypocenters (Van der Hilst et al. 1997, Engdahl et al. 1998), or combinations of various forms of seismological data to constrain both short and long wavelength seismic velocity variations (Ritsema et al. 1999). The tomographic models generally show linear features with high velocity in the lower mantle that correlate well with locations of past and present subduction (Figure 7). In some cases, such as at Tonga, buckling and retarding of the slab in the lower transition zone is visible, but in general, the 670-km discontinuity does not appear to prohibit slab penetration. This interpretation has been strengthened by the good agreement with paleogeographic reconstructions in several locations (e.g., Van der Voo et al. 1999). The fate of slabs in the lowermost mantle is still unclear. On the one hand, the high velocities observed near the core-mantle boundary under the convergent margins around the Pacific and below the Tethys suggest that slabs penetrate to this depth and flatten to form a "slab graveyard" (e.g., Wysession et al. 1995). On the other hand, most tomographic models show slabs fading out before they reach the lowermost mantle. This has been interpreted as being caused by a physical barrier to flow (Van der Hilst & Kárason 1999), although the interpretation is complicated by the relatively low resolution of tomographic structure in the lowermost mantle (e.g., Kárason & Van der Hilst 2001).

Seismological observations can also put constraints on the form of the return flow away from the slabs. The seismic signal associated with mid-oceanic ridges is generally found to be shallow (Zhang & Tanimoto 1992), and it is reasonable to assume that these divergent margins are passive features that form in response to the downward flow induced by the slabs. Active plume-like upwellings are suggested to underlie most regions of hot spot volcanism. So far, it has been difficult to image these upwellings seismically (Nataf 2000), but recent studies may have started spotting plumes in the upper (Wolfe et al. 1997, Shen et al. 1998) and lower mantle (Bijwaard & Spakman 1999, Goes et al. 1999, Ritsema et al. 1999; Figure 8). Of interest for deep Earth dynamics is the observation that the lowermost mantle under Africa and the Pacific is anomalous in shear velocity and density structure (e.g., Su et al. 1994, Ishii & Tromp 1999). Although the magnitude or even sign of correlations between velocity, density, and temperature is not yet clear (Masters et al. 2000, Romanowicz 2001), it appears that these regions are active upwellings that may be chemically distinct from the bulk of the lower mantle (Lithgow-Bertelloni & Silver 1998).

Of potential key interest to understand the chemical evolution of the mantle is the D" region that occupies the lowermost 200 to 300 km of the mantle. The structure of this layer has been the focus of a significant seismological effort in recent years (see reviews in Lay et al. 1998, Garnero 2000). This region of the mantle is very heterogeneous as expressed in seismic velocity and anisotropy and includes regions such as the ultra-low-velocity zone (Garnero & Helmberger 1998) that may indicate areas of strong core-mantle interaction.

In summary, the geophysical evidence for the present-day state of Earth's internal dynamics points to a form of large-scale convection in which most, if not

all, of the mantle participates. This raises an important question. How can such a mantle model be reconciled with the strong geochemical evidence for large-scale heterogeneity?

EVALUATING THE MIXING EFFICIENCY OF MANTLE CONVECTION MODELS

Annu. Rev. Earth. Planet. Sci. 2002.30:493-525. Downloaded from arjournals.annualreviews.org
by UNIVERSITY OF OSLO on 03/30/06. For personal use only.

The chemical structure of the mantle is altered dynamically by the creation and destruction of heterogeneity. Creation involves processes such as differentiation by melting, hydrothermal alteration of oceanic crust, subduction of sediments, and potentially interaction between the core and the mantle. Destruction of heterogeneity occurs by mixing, which depends critically on the vigor and planform of convection. Modeling of the efficiency of mixing provides a useful tool to investigate the feasibility of conceptual models of mantle structure and dynamics.

In a convective medium, mixing of two fluids occurs by stretching and folding of surfaces (Ottino 1989). In general, it is found that the surface area surrounding a given volume increases exponentially with time, with a growth rate that depends on convective vigor (Batchelor 1952, Ottino 1989). Hoffman & McKenzie (1985) modeled this stretching behavior using 2D models of mantle convection at moderate convective vigor and argued, by extrapolation, that whole mantle convection would destroy substantial heterogeneity well within the lifetime of the Earth. As a consequence, they favored a layered model of mantle convection to isolate large portions of the mantle from efficient mixing. Of particular interest is how variations of rheological and thermodynamical parameters in the mantle affect this conclusion. Gurnis & Davies (1986) argued that the inferred high viscosity of the lower mantle makes the lower portions of the mantle poorly mixed, whereas the upper mantle could be well mixed by plate tectonics.

In the past decade, we have seen renewed interest in mixing studies, which has been aided by the large increase in computational capability. This has made it possible to investigate mixing in 2D at Earth-like convective vigor and to start modeling mixing in the more appropriate 3D geometry. Van Keken & Ballentine (1998) provided a test of the hypothesis that a high viscosity lower mantle could be responsible for the low mixing efficiency of the mantle. They considered models that match the convective vigor of the present-day Earth as measured by surface heat flow and plate velocities. By modeling the $^3He/^4He$ evolution of the Earth in a 2D cylindrical model of mantle convection, in which radiogenic ingrowth and surface degassing were taken into account, they concluded that the high viscosity of the lower mantle alone is not sufficient to retard mixing. The preservation of a substantial, more primitive, and poorly mixed reservoir as indicated by the helium isotopes in OIB was not observed, although small-scale heterogeneity could survive over time periods of several hundreds of millions of years. Several independent studies (Ferrachat & Ricard 2001, Hunt & Kellogg 2001; P. Tackley and G. Davies, personal communication) have reached a similar conclusion.

Annu. Rev. Earth. Planet. Sci. 2002.30:493-525. Downloaded from arjournals.annualreviews.org
by UNIVERSITY OF OSLO on 03/30/06. For personal use only.

It is reasonable to assume that the decrease of thermal expansivity and increase of thermal conductivity with depth will cause more sluggish convection in the deeper interior. Also, the endothermic nature of the 670-km phase discontinuity can been used to argue for a partial separation of flow between the upper and lower mantle without requiring variations in bulk chemistry. The mixing behavior of whole mantle convection with phase transitions and depth-dependent expansivity and diffusivity was investigated by Van Keken & Ballentine (1999). Although the mixing rate in the lower mantle is low compared to that in the upper mantle, the requirement that the convective system is characterized by realistic values for surface heat flow and plate velocities causes the overall mixing to be fairly efficient. As a consequence, it can be argued that some form of chemical layering is still necessary to preserve the observed geochemical heterogeneity.

A similar conclusion was reached by reconsideration of the heat-helium imbalance (O'Nions & Oxburgh 1983), which suggests that proportionally significantly less ^{4}He is released at the Earth surface compared to the expected flux based on the mantle concentration of U and Th. It is possible that the differences in timescale between processes of rapid extraction of helium at a mid-oceanic ridge and the slow diffusive release of heat along the cooling oceanic lithosphere are responsible for large variations in He/heat. Van Keken et al. (2001) investigated this hypothesis by calculating the heat-He evolution over four Ga using a secular-cooling model of the Earth. By monitoring the release of heat and helium in time, they concluded that although excursions from the mean He/heat to the observed present-day ratio may exist, it would be rather fortunate if the present-day Earth is in such a state given the short duration and long interval between excursions.

An important question concerns the difference between mixing in 2D and 3D, due to the potentially large differences in convection planform and geometric differences in mixing behavior. Schmalzl et al. (1996) investigated the differences between 2D and 3D mixing in isoviscous models at comparable convective vigor. They argued for significant differences in mixing behavior and in particular that mixing in 3D was less efficient than in 2D. It should be noted that these experiments were limited to poloidal flow only. In general, a 3D flow pattern can be decomposed in poloidal and toroidal components. For example, the divergent flow at a mid-oceanic ridge is poloidal, whereas the strike-slip motion along a transform fault is toroidal. An intriguing demonstration of the influence of toroidal motion was provided by Ferrachat & Ricard (1998) who used a simple kinematically driven box model where the surface velocity was prescribed by a ridge with a transform fault. The mixing efficiency, as evaluated by Luyaponov exponents, showed that models with a transform fault allowed for a separation of the model into a well mixed region and poorly mixed areas that remained isolated from the large-scale flow over long time periods (Figure 9). It is important to note that kinematic mixing generally allows for better preservation of heterogeneity compared to fully dynamical models (Christensen 1989). Van Keken & Zhong (1999) evaluated the mixing efficiency in a 3D spherical model of present-day convective flow that was steady state but dynamically driven, using the slab buoyancy flow model of

Annu. Rev. Earth. Planet. Sci. 2002.30:493-525. Downloaded from arjournals.annualreviews.org
by UNIVERSITY OF OSLO on 03/30/06. For personal use only.

Lithgow-Bertelloni & Richards (1998). The resulting model matches the geometry and speed of the present-day plate motions reasonably well by assuming a depth-dependent rheology with a 30-fold increase of viscosity into the lower mantle. The combined poloidal and toroidal motion leads to particle paths that in a substantial portion of the mantle have a characteristic cork screw pattern and allow for large distances of transport (Figure 10). Similar to the box model of Ferrachat & Ricard (1998), this model demonstrates the existence of well-mixed regions next to poorly mixed regions. A related illustration of the mixing efficiency of the present day Earth based on the model of Van Keken & Zhong (1999) is provided in Figure 11, which shows how the Indian plate would be mixed in over time, assuming steady state convective flow. Although poloidal flow dominates the mixing in certain areas such as the Indian Ocean, the toroidal component of surface flow, which is particularly strong near Papua New Guinea, causes large-scale disruption of the plate and efficient transport of material into the Pacific circulation zone. Nevertheless, it is questionable if the type of isolation seen in steady state models can be argued to exist in the Earth. The intrinsic time-dependence of mantle convection should expose such isolated cells and mix them efficiently into the large-scale flow.

In summary, we are left with the conclusion that models dominated by whole mantle convection, even when moderated by phase changes or more sluggish convection in the lower mantle, cannot preserve large-scale heterogeneity over the age of the Earth. This conclusion becomes more severe when considering that most of the recent models use the present-day convective vigor as a measure. In general, this can be considered a conservative estimate because the convective vigor was most likely higher in the past due to the higher radiogenic heat production. Reconstructions of past plate motions (Lithgow-Bertelloni & Richards 1998) and heat flow (Sprague & Pollack 1980) also suggest that the convective vigor at present is moderate within the record since the Jurassic.

IDEAS FOR THE PRESERVATION OF HETEROGENEITY

Several suggestions have been put forward in recent years to try to reconcile the apparent conflict between the geophysical evidence for whole mantle convection (and by extrapolation a well-mixed mantle) and the geochemical evidence for substantial and long-lived heterogeneity. In this section, we review some of the recent concepts and provide a discussion of the strengths and weaknesses. In general, the proposed suggestions for preservation of chemical heterogeneity are variations on one of three themes: inefficient mixing due to physical properties, preservation of chemical heterogeneity by physical barriers, or temporal changes in Earth dynamics.

Convective Layering at 670 km Depth

Several recent studies prefer to retain a form of convection model with a mechanical boundary layer at the base of the transition zone or slightly deeper

Annu. Rev. Earth. Planet. Sci. 2002.30:493-525. Downloaded from arjournals.annualreviews.org
by UNIVERSITY OF OSLO on 03/30/06. For personal use only.

(Wen & Anderson 1997, Anderson 1998b). This requires the assumption that the tomographically observed high velocity extensions to upper mantle subduction zones are not due to a mechanical input from the upper mantle, but rather are caused by thermal coupling. The potential for thermal coupling has been demonstrated in modeling studies (e.g., Čižková et al. 1999), but this mechanism requires a strong mechanical decoupling as well as very efficient diffusive heat transport along the 670-km discontinuity, both of which appear to require very special conditions that are not apparent at other depths in the mantle. As argued by Kárason & van der Hilst (2000), it may also be quite difficult to explain the present-day mantle structure by thermal alignment along the mechanical boundary of slow moving structures in the lower mantle with rapidly changing structures in the upper mantle. An independent problem with layered mantle convection is the inefficiency of heat transport across the mechanical boundary layer, which would lead to a strong thermal boundary layer and potential overheating of the lower mantle, neither of which is observed (Spohn & Schubert 1982, McNamara & Van Keken 2000). Layered convection models also have implications that contradict several lines of geochemical evidence, as discussed earlier.

Interplanetary Dust Particles

An alternative secondary source of primitive noble gases that has been proposed is the subduction of interplanetary dust particles (IDPs; Sarda et al. 1988, Anderson 1993, Allègre et al. 1993). This would be a convenient way to explain the presence of substantial ^{3}He concentrations in the mantle while being consistent with the lithophile geochemistry that suggests that no primitive mantle has survived mantle mixing. However, the mass of IDP material is not sufficient to account for the ^{3}He concentration or flux requirements (Trull 1994, Stuart 1994), and it is not likely that the volatile content of IDPs survive dehydration and melting of the sediments upon subduction (Staudacher et al. 1989, Hiyagon 1994).

Deeper Layering

Kellogg et al. (1999) proposed the existence of a chemical boundary layer in the lower mantle at approximately 1600 km depth. In essence, this is a variation on the "classical" layered model where the boundary is moved to a region that is seismologically more poorly resolved than shallower levels and D". Several seismological studies also hint at chemical and/or thermal heterogeneity. The strengths of this model include the potential for preservation of chemical heterogeneity and a source for the missing radiogenic elements. It shares weaknesses with the layered convection model regarding thermal layering, the likely overheating of the lower layer, and requirements for even higher ^{3}He concentrations. It can be argued that these problems are more severe in this case due to the much smaller volume of the lower layer and the consequently much higher radiogenic heat production that needs to be assumed (McNamara & Van Keken 2000). The model conceptualizes large topography at the boundary layer to avoid detection by seismological

reflections, and this may also reduce some of the problems with the temperature distribution. Nevertheless, even with an irregular boundary, the large physical differences between the layers would lead to seismic scattering, which thus far has not been observed (Vidale et al. 2001).

Distributed Blobs

In a variation on chemical layering, it has been suggested that chemical heterogeneity can be distributed as isolated blobs that are mixed in with the large-scale structure (Davies 1984). This allows for distributed heterogeneity that is difficult to see by geophysical methods but does allow for storage of the various plume components and the missing radiogenic heat. Recently, a version of the blob model has been revived (Becker et al. 1999) in which it is assumed that these blobs have a higher viscosity than the surrounding mantle because it has been suggested such blobs do not mix efficiently in kinematically driven flows (Manga et al. 1996). The large surface-to-volume ratio of small blobs compared to that of a layer allows for efficient cooling of the blobs and hence the ability to retain high concentrations of radiogenic elements (Becker et al. 1999). Several questions about this model remain unanswered regarding the origin and survival mechanism of these blobs. It is puzzling how such blobs can remain neutrally buoyant and why such blobs would not be seen at mid-oceanic ridges, yet are sampled by dynamically much less important mantle plumes.

Ocean Crust Recycling

A promising set of models addresses the formation and retention of mantle heterogeneity caused by recycling of oceanic crust (Hofman & White 1980, Silver et al. 1988, Christensen & Hofmann 1994, Coltice & Ricard 1999), which is strongly supported by the Re-Os evidence for a subduction component in hot spot basalts. The formation of basaltic crust at mid-oceanic ridges causes compositional heterogeneity due to processes that include garnet extraction from peridotite to form basalt and harzburgite, degassing upon melting, enrichment of the basaltic crust compared to the underlying mantle, and interaction with the hydrosphere. Upon subduction, the oceanic crust and depleted mantle can potentially form a long-lived reservoir at the base of the mantle. Modeling shows that such a recycling mechanism in a five reservoir model (bulk mantle, continental crust, atmosphere, recycled harzburgite, and recycled crust) can account for the observed variations in Rb/Sr and U/Pb/He system (Coltice & Ricard 1999) assuming that the "primitive" (high $^3He/^4He$) helium isotope signature is caused by the depleted characteristics of the harzburgite and hence low ^4He ingrowth (e.g., Albarède 1998). An additional attractive feature of this model is that it is based on plate tectonic processes without the introduction of ad hoc chemical heterogeneity or physical separation mechanisms. A possible drawback is that there is no clear solution for the missing heat producing elements, as the small volume of D" requires very high concentrations of radiogenic elements. This would lead to rapid destabilization. In addition, the U-depletion may not be strong enough to account for the high ^3He signature of

Annu. Rev. Earth. Planet. Sci. 2002.30:493-525. Downloaded from arjournals.annualreviews.org
by UNIVERSITY OF OSLO on 03/30/06. For personal use only.

many hot spots, which still would require ad hoc processes such as a helium flux from the core.

The Core

Driven by the continuous exposure of the deeper Earth by seismology, there is a logical and perhaps unavoidable tendency to push the locations for chemical heterogeneity deeper into the Earth. The core forms one such reservoir with several attractive characteristics. Some interaction between core and mantle has been suggested based on anomalies in ^{186}Os (Brandon et al. 1998) and the core may be the source for various other signals such as ^{3}He that have typically been interpreted to correspond to primitive mantle. In addition, the potential storage of radiogenic K in the core (e.g., Bukowinski 1976) would allow for a high core heat flow to compensate the missing heat in the present-day steady state heat balance. In a recent paper, Porcelli & Halliday (2001) conclude that given the current poorly defined noble gas partition data between silicate solid and a metallic liquid, it is not possible to discount the core as a sink for a significant mass of noble gases during core formation followed by later release into the convecting mantle system. At present, both the input and extraction of noble gases from the core require stringent and as yet untested requirements. These include the need for very high noble gas concentrations in the silicate mantle to partition sufficiently into the metal phase during core formation. The process of partitioning and release of both ^{3}He and ^{22}Ne has to occur at near solar ratios (Honda et al. 1991) without significantly affecting the trace element pattern of OIB or MORB. The only available observation in support of a core source for high ^{3}He/^{4}He is a weak correlation between ^{186}Os/^{188}Os and ^{3}He/^{4}He at Hawaii (Brandon et al. 1998). At present, the physics and chemistry of the core and core-mantle interactions are not well understood and suggestions for a core source may therefore be considered speculative rather than a viable alternative to the isolation suggested by oceanic basalt geochemistry.

Inefficient Mixing Due to Rheological Variations

The strong temperature- and stress-dependence of mineral deformation processes has been suggested as one key player in the preservation of chemical heterogeneity, by isolating regions with locally high viscosity due to low stress, low temperature, or low water content. In essence, this is similar to the "core-in-cell" model (Tatsumoto 1978, Schubert & Spohn 1981) in which it is assumed that the large-scale circulation is driven by transport through boundary layers while leaving the cores of the cells undisturbed, except when a plume, independent of the large-scale flow, travels through such a core and entrains the material that has been isolated. However, it is far from clear how the geometry of such cells can be retained in the time-dependent convection that characterizes the Earth. Some studies have pointed at the differences in mixing behavior of Newtonian and non-Newtonian fluids, which generally show less efficient mixing of the non-Newtonian models (e.g., Ten et al. 1997). This effect appears to be dominant at short length scales, and it is not clear if this is sufficient to retain large-scale heterogeneity over long

Annu. Rev. Earth. Planet. Sci. 2002.30:493-525. Downloaded from arjournals.annualreviews.org
by UNIVERSITY OF OSLO on 03/30/06. For personal use only.

timescales. An important challenge for geodynamical models remains the incorporation of realistic stress- and temperature-dependent rheology. This necessitates a self-consistent implementation of the physics governing plate breakup that remains unresolved, although much progress has been made in recent years (see review by Tackley 2000).

Recent Break-up of a Layered System

An important distinction between the geophysical and geochemical observations of mantle evolution is that geophysics provides us with a snapshot of the present day structure, whereas geochemistry provides a time-integrated history of the mantle system. As a consequence, it is permissible that the present-day configuration of whole mantle convection was attained only recently and that the Earth was characterized by layered mantle convection during most of its history (Allègre 1997). Although conceptually attractive, it is not clear how such a dramatic change in convective style would be achieved, nor how it would remain unnoticed in the surface geology. A breakup of the layered system can be predicted to dramatically increase the upper mantle temperature with large consequences for plate tectonics and extrusion temperatures of igneous rocks. However, this is not observed (e.g., Abbott et al. 1993).

Radiation and Conduction as Dominant Heat Transport Mechanisms in the Deep Mantle

An important argument against the deep storage of radiogenic elements in large concentrations is that such regions would heat up and convectively mix in with the large-scale flow. This can be counteracted by allowing for efficient cooling, for example, by increasing the surface-to-volume ratio such as in the blob model of Becker et al. (1999), or, more speculatively, by high efficiency of conduction and radiation. From a dynamical point of view, it would be attractive if the increase in temperature and pressure with depth allowed for a gradual transition between a "convective" and a "radiative zone" in the mantle, which would not be detectable by seismological methods and still allowed for efficient heat transport and the retention of chemical heterogeneity due to the much reduced convective vigor at depth. The reduced vigor may also play a role in stabilizing the large "superplume" areas in the lowermost mantle (Matyska et al. 1994). However, it is far from clear if conduction and radiation can dominate the heat transport in the deep Earth. Conductivity through lattice vibrations decreases with temperature and increases with pressure (Anderson 1989, Hofmeister 1999), but the overall depth-dependent increase through the mantle is estimated to be less than a factor of four, which is too small to dominate convective heat transport. The separate contribution through radiation increases strongly with temperature (e.g., Shankland et al. 1979), but decreases due to the effects of grain boundary scattering and the opacity induced by transition metals such as Fe. In general, the effect of radiation is considered small compared to that of conduction lattice vibrations (Katsura 1995, Hofmeister 1999).

Annu. Rev. Earth. Planet. Sci. 2002.30:493-525. Downloaded from arjournals.annualreviews.org
by UNIVERSITY OF OSLO on 03/30/06. For personal use only.

Annu. Rev. Earth. Planet. Sci. 2002.30:493-525. Downloaded from arjournals.annualreviews.org
by UNIVERSITY OF OSLO on 03/30/06. For personal use only.

SUMMARY AND OUTLOOK

The geochemical and structural state of the mantle today is the result of a complex interplay between the continuing processes of chemical differentiation and the dynamics of the geophysical environment that work to variably preserve or destroy the geochemical record of the heterogeneity. It is becoming increasingly obvious that geochemical models that ignore the dynamic environment in which they are set cannot provide a full account of the Earth's chemical evolution. It is also clear that geophysical models unconstrained by the mass balance and temporal information available from geochemistry will be incomplete. The greatest strength of the layered mantle model to date has been its simplicity, by keeping the number of mantle reservoirs to a minimum and yet accounting for many first-order geochemical observations. Similarly, dynamical models of mantle behavior are driven by fundamental principles of fluid dynamics and mineral physics, and in turn, reproduce first-order geophysical observations. These end-member models provide a clear frame of reference from which future models must evolve. Although we are left with the conclusion that no single model to date adequately addresses all the main constraints of the physio-chemical state and evolution of the Earth's mantle, recent approaches that integrate both the physics and chemistry of mantle behavior are making promising inroads toward the development of a more complete model.

Based on our current understanding, an important set of primary constraints can be formulated. These include: (*a*) the mass balance based on the bulk silicate Earth composition, (*b*) energy balance as expressed in thermal evolution models, (*c*) the observed diversity of OIB sources expressed in the lithophile and noble gas tracers as contrasted to the MORB homogeneity, (*d*) the petrological and structural evolution of the crust and upper mantle as recorded in the continents, (*e*) present-day structure of the mantle as observed by seismology, (*f*) time integrated constraints of chemical evolution provided by radiogenic isotope geochemistry, and (*g*) dynamical feasibility of conceptual models as tested by theoretical approaches.

It is clear that this list poses conflicts such as the need for an internal boundary layer suggested by the mass and heat balance and the seismological evidence for whole mantle flow, or the preservation of ^{3}He in mantle material that also records the effects of ~~surface~~ melting. Future generations of models need to be consistent with the main constraints and provide a resolution of the apparent conflicts that they pose. Within that context, the communication and creative interplay between geophysical and geochemical approaches aid the exposure of the deeper Earth and its role in the formation of oceanic basalts.

ACKNOWLEDGMENTS

We thank Wim Spakman and Jeroen Ritsema for providing their tomography models. Discussion with Larry Ruff, Henry Pollack, Stan Hart, Rick Carlson, Alex Halliday, Don Porcelli, Mark Rehkämper, Boris Kiefer, and Allen McNamara are gratefully acknowledged.

Visit the Annual Reviews home page at www.annualreviews.org

Annu. Rev. Earth. Planet. Sci. 2002.30:493-525. Downloaded from arjournals.annualreviews.org
by UNIVERSITY OF OSLO on 03/30/06. For personal use only.

LITERATURE CITED

Abbott DA, Burgess L, Longhi J, Smith WHF. 1993. An empirical thermal history of the Earth's upper mantle. *J. Geophys. Res.* 99:13835–50

Albarède F. 1998. Time-dependent models of U-Th-He and K-Ar evolution and the layering of mantle convection. *Chem. Geol.* 145:413–29

Allègre CJ. 1997. Limitation on mass exchange between the upper and lower mantle: the evolving convection regime of the Earth. *Earth Planet. Sci. Lett.* 150:1–6

Allègre CJ, Hamelin B, Provost A, Duprè B. 1987. Topology in isotopic multispace and the origin of mantle chemical heterogeneities. *Earth Planet. Sci. Lett.* 81:319–37

Allègre CJ, Othman DB, Polve M, Richard P. 1979. The Nd-Sr isotopic correlation in mantle materials and geodynamic consequences. *Phys. Earth Planet. Inter.* 19:293–306

Allègre CJ, Sarda P, Staudacher T. 1993. Speculations about the cosmic origin of He and Ne in the interior of the Earth. *Earth Planet. Sci. Lett.* 117:229–33

Anderson DL. 1989. *Theory of the Earth*. Boston, MA: Blackman Sci. 366 pp.

Anderson DL. 1993. Helium-3 from the mantle: primordial signal or cosmic dust? *Science* 261:170–76

Anderson DL. 1998a. A model to explain the various paradoxes associated with mantle noble gas geochemistry. *Proc. Natl. Acad. Sci. USA* 95:9087–92

Anderson DL. 1998b. The scales of mantle convection. *Tectonophysics* 284:1–17

Ballentine CJ, Barfod DN. 2000. The origin of air-like noble gases in MORB and OIB. *Earth Planet. Sci. Lett.* 180:39–48

Ballentine CJ, Lee DC, Halliday AN. 1997. Hafnium isotopic studies of the Cameroon line and new HIMU paradoxes. *Chem. Geol.* 139:111–24

Ballentine CJ, Porcelli D, Wieler R. 2001. Technical comment on 'Noble gases in mantle plumes' by Trieloff et al. (2000). *Science* 291:A2269

Barfod DN, Ballentine CJ, Halliday AN, Fitton JG. 1999. Noble gases in the Cameroon line and the He, Ne, and Ar isotopic compositions of HIMU mantle. *J. Geophys. Res.* 104:29509–27

Barfod DN, Lee DC, Ballentine CJ, Halliday AN, Hall C. 2001. Fractionation of helium and incompatible lithophile elements during partial melting under the Cameroon line. *Earth Planet. Sci. Lett.* In press

Batchelor GK. 1952. The effect of homogeneous turbulence on material lines and surfaces. *Proc. R. Soc. London Ser. A* 213:349–66

Becker TW, Kellogg JB, O'Connell RJ. 1999. Thermal constraints on the survival of primitive blobs in the lower mantle. *Earth Planet. Sci. Lett.* 171:351–65

Bijwaard H, Spakman W. 1999. Tomographic evidence for a narrow whole mantle plume below Iceland. *Earth Planet. Sci. Lett.* 166:121–26

Bijwaard H, Spakman W, Engdahl ER. 1998. Closing the gap between regional and global travel time tomography. *J. Geophys. Res.* 103:30055–78

Brandon AD, Walter RJ, Morgan JW, Norman MD, Prichard HM. 1998. Coupled Os-186 and Os-187 evidence for core-mantle interaction. *Science* 280:1570–73

Buffett BA, Huppert HE, Lister JR, Woods AW. 1996. On the thermal evolution of the Earth's core. *J. Geophys. Res.* 101:7989–8006

Bukowinski MST. 1976. Effect of pressure on physics and chemistry of potassium. *Geophys. Res. Lett.* 3:491–94

Carlson RW. 1994. Mechanisms for Earth differentiation: consequences for the chemical structure of the mantle. *Rev. Geophys.* 32:337–61

Castillo P. 1988. The DUPAL anomaly as a trace of the upwelling lower mantle. *Nature* 336:667–70

Annu. Rev. Earth. Planet. Sci. 2002.30:493-525. Downloaded from arjournals.annualreviews.org
by UNIVERSITY OF OSLO on 03/30/06. For personal use only.

Christensen U. 1989. Mixing by time-dependent convection. *Earth Planet. Sci. Lett.* 95:382–94

Christensen UR, Hofmann AW. 1994. Segregation of subducted oceanic crust in the convecting mantle. *J. Geophys. Res.* 99:19867–84

Čižiková H, Čadek O, van den Berg AP, Vlaar NJ. 1999. Can lower mantle slab-like anomalies be explained by thermal coupling between the upper and lower mantles? *Geophys. Res. Lett.* 26:1501–4

Coltice N, Ricard Y. 1999. Geochemical observations and one layer mantle convection. *Earth Planet. Sci. Lett.* 174:125–37

Creager KC, Jordan TH. 1984. Slab penetration into the lower mantle. *J. Geophys. Res.* 89:3031–49

Creager KC, Jordan TH. 1986. Slab penetration into the lower mantle beneath the Mariana and other island arcs of the northwest Pacific. *J. Geophys. Res.* 91:3573–89

Davies GF. 1980. Mechanics of subducted lithosphere. *J. Geophys. Res.* 85:6304–18

Davies GF. 1984. Geophysical and isotopic constraints on mantle convection: an interim synthesis. *J. Geophys. Res.* 89:6017–40

DePaolo D, Wasserburg GJ. 1976. Inferences about magma sources and mantle structure from variations of $^{143}Nd/^{144}Nd$. *Geophys. Res. Lett.* 3:743–46

Dziewonski AM. 1984. Mapping the lower mantle: determination of lateral heterogeneity in P velocity up to degree and order 6. *J. Geophys. Res.* 89:5929–52

Dziewonski AM, Woodhouse JH. 1987. Global images of the Earth's interior. *Science* 236:37–48

Eiler JM. 2001. Oxygen isotope variations of basaltic lavas and upper mantle rocks. In *Stable Isotopes in Igneous Processes*, ed. D Cole, JW Valley. Washington, DC: Mineral. Soc. Am. In press

Eiler JM, Farley KA, Stolper EM. 1998. Correlated helium and lead isotope variations in Hawaiian lavas. *Geochim. Cosmochim. Acta* 62:1977–84

Eiler JM, Farley KA, Valley JW, Hauri EH, Craig H, et al. 1997. Oxygen isotope variations in ocean island basalt phenocrysts. *Geochim. Cosmochim. Acta* 61:2281–93

Eiler JM, Farley KA, Valley JW, Stolper EM. 1996. Oxygen isotope constraints on the sources of Hawaiian volcanism. *Earth Planet. Sci. Lett.* 144:453–68

Eiler JM, Schiano P, Kitchen N, Stolper EM. 2000. Oxygen isotope evidence for the origin of chemical variations in the sources of mid-ocean ridge basalts. *Nature* 403:530–34

Engdahl ER, van der Hilst RD, Buland RP. 1998. Global teleseismic earthquake relocation with improved travel times and procedures for depth determination. *Bull. Seism. Soc. Am.* 88:722–43

Farley KA, Basu A, Craig H. 1993. He, Sr and Nd isotopic variations in lavas from the Juan Fernandez Archipelago, S. E. Pacific. *Contrib. Mineral. Petrol.* 115:75–87

Farley KA, Maierreimer E, Schlosser P, Broecker WS. 1995. Constraints on mantle He-3 fluxes and deep-sea circulation from an oceanic general circulation model. *J. Geophys. Res.* 100:3829–39

Farley KA, Natland J, Craig H. 1992. Binary mixing of enriched and undegassed (primitive?) mantle components (He, Sr, Nd, Pb) in Samoan lavas. *Earth Planet. Sci. Lett.* 111:183–99

Farley KA, Neroda E. 1998. Noble gases in the Earth's mantle. *Annu. Rev. Earth Planet. Sci.* 26:189–218

Faure G, Hurley PM. 1963. The isotopic composition of strontium in oceanic and continental basalt: application to the origin of igneous rocks. *J. Petrol.* 4:31–50

Ferrachat S, Ricard Y. 1998. Regular vs. chaotic mantle mixing. *Earth Planet. Sci. Lett.* 155:75–86

Ferrachat S, Ricard Y. 2001. Mixing properties in the Earth's mantle: effects of the viscosity stratification and of oceanic crust segregation. *Geochem. Geophys. Geosys.*, Vol. 2. Paper No. 2000GC000092

Fukao Y, Obayashi M, Inoue H, Nenbai M. 1992 Subducting slabs stagnant in the mantle transition zone. *J. Geophys. Res.* 97:4809–22

Annu. Rev. Earth. Planet. Sci. 2002.30:493-525. Downloaded from arjournals.annualreviews.org
by UNIVERSITY OF OSLO on 03/30/06. For personal use only.

Gable CW, O'Connell RJ, Travis BJ. 1991. Convection in three dimensions with surface plates; generation of toroidal flow. *J. Geophys. Res.* 96:8391–405

Galer SJG, O'Nions RK. 1985. Residence time of thorium, uranium and lead in the mantle with implications for mantle convection. *Nature* 316:778–82

Galer SJG, O'Nions RK. 1989. Chemical and isotopic studies of ultramafic inclusions from the San Carlos volcanic field, Arizona: a bearing on their petrogenesis. *J. Petrol.* 30: 1033–64

Garnero EJ. 2000. Heterogeneity of the lowermost mantle. *Annu. Rev. Earth Planet. Sci.* 28:509–37

Garnero EJ, Helmberger DV. 1998. Further structural constraints and uncertainties of a thin laterally varying ultralow-velocity layer at the base of the mantle. *J. Geophys. Res.* 103:12495–509

Gast PW, Tilton GR, Hedge CE. 1964. Isotopic composition of lead and strontium from Ascension and Gough Islands. *Science* 145:1181–85

Goes S, Spakman W, Bijwaard H. 1999. A lower mantle source for central European volcanism. *Science* 286:1928–31

Graham DW, Castillo PR, Lupton JE, Batiza R. 1996. Correlated He and Sr isotope ratios in South Atlantic near ridge seamounts and implications for mantle dynamics. *Earth Planet. Sci. Lett.* 144:491–503

Graham DW, Jenkings WJ, Schilling JG, Thompson G, Kurz MD, Humphris SE. 1992. Helium isotope geochemistry of mid-oceanic ridge basalts from the South Atlantic. *Earth Planet. Sci. Lett.* 110:133–47

Grand SP. 1994. Mantle shear structure beneath the Americas and surrounding oceans. *J. Geophys. Res.* 99:11591–621

Gurnis M, Davies GF. 1986. The effect of depth-dependent viscosity on convective mixing in the mantle and the possible survival of primitive mantle. *Geophys. Res. Lett.* 13:541–44

Hager BH, Richards MA. 1989. Long-wavelength variations in Earth's geoid: physical models and dynamical implications. *Philos. Trans. R. Soc. London Ser. A* 328:309–27

Hanan BB, Graham DW. 1996. Lead and helium isotope evidence from oceanic basalts for a common deep source of mantle plumes. *Science* 272:991–95

Harris C, Smith HS, LeRoex AP. 2000. Oxygen isotope composition of phenocrysts from Tristan da Cunha and Gough Island lavas: variation with fractional crystallization and evidence for assimilation. *Contrib. Mineral. Petrol.* 138:164–75

Hart SR. 1984. A large-scale isotope anomaly in the Southern Hemisphere mantle. *Nature* 309:753–57

Hart SR. 1988. Heterogeneous mantle domains: signatures, genesis and mixing chronologies. *Earth Planet. Sci. Lett.* 90:273–96

Hart SR, Hauri EH, Oschmann LA, Whitehead JA. 1992. Mantle plumes and entrainment: isotopic evidence. *Science* 26:517–20

Hart SR, Schilling JG, Powell JL. 1973. Basalts from Iceland and along the Reykjanes Ridge: Sr isotopic geochemistry. *Nature* 246:104–7

Hauri EH. 1996. Major element variability in the Hawaiian mantle plume. *Nature* 382:415–19

Hauri EH, Hart SR. 1993. Re-Os isotope systematics of EMII and HIMU oceanic island basalts from the south Pacific Ocean. *Earth Planet. Sci. Lett.* 114:353–71

Hauri EH, Hart SR. 1997. Rhenium abundances and systematics in oceanic basalts. *Chem. Geol.* 139:185–205

Hauri EH, Lassiter JC, DePaolo DJ. 1996. Osmium isotope systematics of drilled lavas from Mauna Loa, Hawaii. *J. Geophys. Res.* 101:11793–806

Hauri EH, Whitehead JA, Hart SR. 1994. Fluid dynamic and geochemical aspects of entrainment in mantle plumes. *J. Geophys. Res.* 99:24275–300

Hedge CE, Walthall FG. 1963. Radiogenic strontium-87 as an index of geologic processes. *Science* 140:1214–17

Hilton DR, McMurty GM, Goff F. 1998. Large variations in vent fluid CO_2/He-3 ratios

Annu. Rev. Earth. Planet. Sci. 2002.30:493-525. Downloaded from arjournals.annualreviews.org
by UNIVERSITY OF OSLO on 03/30/06. For personal use only.

signal rapid changes in magma chemistry at Loihi seamount, Hawaii. *Nature* 396:359–62

Hilton DR, McMurty GM, Kreulen R. 1997. Evidence for extensive degassing of the Hawaiian mantle plume from helium-carbon relationships at Kilauea volcano. *Geophys. Res. Lett.* 24:3065–68

Hilton DR, Thirlwall MF, Taylor RN, Murton BJ, Nichols A. 2000. Controls on magmatic degassing along the Reykjanes Ridge with implications for the helium paradox. *Earth Planet. Sci. Lett.* 183:43–50

Hiyagon H. 1994. Retention of solar helium and neon in IDPs in deep sea sediment. *Science* 263:1257–59

Hoffman NRA, McKenzie DP. 1985. The destruction of geochemical heterogeneities by differential fluid motions during mantle convection. *Geophys. J. R. Astron. Soc.* 82:163–206

Hofmann AW. 1986. Nb in Hawaiian magmas: constraints on source composition and evolution. *Chem. Geol.* 57:17–30

Hofmann AW. 1988. Chemical differentiation of the Earth: the relationship between mantle, continental crust, and oceanic crust. *Earth Planet. Sci. Lett.* 90:297–314

Hofmann AW. 1997. Mantle geochemistry: the message from oceanic volcanism. *Nature* 385:219–29

Hofmann AW, White WN. 1980. The role of subducted oceanic crust in mantle evolution. *Carnegie Inst. Wash. Yearb.* 79:477–83

Hofmeister AM. 1999. Mantle values of thermal conductivity and the geotherm from phonon lifetimes. *Science* 283:1699–706

Honda M, McDougall I, Patterson DB, Doulgeris A, Clague DA. 1991. Possible solar noble-gas component in Hawaiian basalts. *Nature* 349:149–51

Hunt DL, Kellogg LH. 2001. Quantifying mixing and age variations of heterogeneities in models of mantle convection: role of depth-dependent viscosity. *J. Geophys. Res.* 106:6747–59

Isacks B, Molnar P. 1971. Distribution of stress in the descending lithosphere from a global survey of focal mechanism solutions of mantle earthquakes. *Rev. Geophys. Space Phys.* 9:103–74

Ishii M, Tromp J. 1999. Normal mode and free-air gravity constraints on lateral variations in velocity and density of Earth's mantle. *Science* 285:1231–36

Ito E, Takahashi E. 1989. Post-spinel transformations in the system Mg_2SiO_4-Fe_2SiO_4 and some geophysical implications. *J. Geophys. Res.* 94:10637–46

Jagoutz E, Palme H, Baddenhausen H, Blum B, Cendales M, et al. 1979. The abundance of major, minor and trace elements in the Earth's mantle as derived from primitive ultramafic nodules. *Proc. Lunar Sci. Conf.* 10:2031–50

James DE. 1981. The combined use of oxygen and radiogenic isotopes as indicators of crustal contamination. *Annu. Rev. Earth Planet. Sci.* 9:311–44

Javoy M, Pineau F. 1991. The volatiles record of a popping rock from the Mid-Atlantic Ridge at 14N—chemical and isotopic composition of gas trapped in the vesicles. *Earth Plan. Sci. Lett.* 107:598–611

Jochum KP, Arndt NT, Hofmann AW. 1991. Nb-Th-La in komatiites and basalts: constraints on komatiite petrogenesis and mantle evolution. *Earth Planet. Sci. Lett.* 107:272–89

Jochum KP, Hofmann AW, Seufert HM. 1993. Tin in mantle-derived rocks: constraints on Earth evolution. *Geochim. Cosmochim. Acta* 57:3585–95

Kárason H, van der Hilst RD. 2000. Constraints on mantle convection from seismic tomography. In *The History and Dynamics of Global Plate Motions*, ed. MA Richards, RG Gordon, RD van der Hilst, pp. 277–88. Washington, DC: Am. Geophys. Union

Kárason H, van der Hilst RD. 2001. Tomographic imaging of the lowermost mantle with differential times of refracted and diffracted core phases. *J. Geophys. Res.* 106:6569–87

Karato S. 1989. Plasticity-crystal structure systematics in dense oxides and its implications for the creep strength of the Earth's deep

Annu. Rev. Earth. Planet. Sci. 2002.30:493-525. Downloaded from arjournals.annualreviews.org
by UNIVERSITY OF OSLO on 03/30/06. For personal use only.

interior: a preliminary result. *Phys. Earth Planet. Int.* 55:234–40

Katsura T. 1995. Thermal diffusivity of olivine under upper-mantle conditions. *Geophys. J. Int.* 122:63–69

Kellogg LH, Hager BH, van der Hilst RD. 1999. Compositional stratification in the deep mantle. *Science* 283:1881–84

Kellogg LH, Wasserburg GJ. 1990. The role of plumes in mantle helium fluxes. *Earth Planet. Sci. Lett.* 99:276–89

Kurz MD, Garcia MO, Frey FA, O'Brien PA. 1987. Temporal helium isotopic variations in volcanic rocks from Loihi seamount and the island of Hawaii. *Geochim. Cosmochim. Acta* 51:2905–14

Kurz MD, Geist D. 1999. Dynamics of the Galapagos hotspot from helium isotope geochemistry. *Geochim. Cosmochim. Acta* 63:4139–56

Kurz MD, Jenkins WJ, Hart SR. 1982. Helium isotopic systematics of oceanic islands and mantle heterogeneity. *Nature* 297:43–46

Kurz MD, Kammer DP. 1991. Isotopic evolution of Mauna Loa volcano. *Earth Planet. Sci. Lett.* 103:257–69

Langmuir CH, Vocke RD, Hanson GN, Hart SR. 1978. A general mixing equation with applications to Icelandic basalts. *Earth Planet. Sci. Lett.* 37:380–92

Lassiter JC, Hauri EH. 1998. Osmium isotope variations in Hawaiian lavas: evidence for recycled oceanic lithosphere in the Hawaiian plume. *Earth Planet. Sci. Lett.* 164:483–96

Lay T, Williams Q, Garnero EJ. 1998. The core-mantle boundary layer and deep Earth dynamics. *Nature* 392:461–68

Lithgow-Bertelloni C, Richards MA. 1998. The dynamics of Cenozoic and Mesozoic plate motions. *Rev. Geophys.* 36:27–78

Lithgow-Bertelloni C, Silver PG. 1998. Dynamic topography, plate driving forces and the African superswell. *Nature* 395:269–72

Lupton JE, Craig H. 1975. Excess He-3 in oceanic basalts—evidence for terrestrial primordial helium. *Earth Planet. Sci. Lett.* 26:133–39

Manga M. 1996. Mixing of heterogeneities in the mantle: effect of viscosity differences. *Geophys. Res. Lett.* 23:403–6

Masters G, Laske G, Bolton H, Dziewonski A. 2000. The relative behavior of shear velocity, bulk sound speed, and compressional velocity in the mantle: implications for chemical and thermal structure. In *Earth's Deep Interior: Mineral Physics and Tomography from the Atomic to the Global Scale*, ed. S Karato, AM Forte, RC Liebermann, G Masters, L Stixrude, pp. 63–87. Washington, DC: Am. Geophys. Union

Mattey D, Lowry D, Macpherson C. 1994. Oxygen isotope composition of mantle peridotite. *Earth Planet. Sci. Lett.* 66:388–406

Matyska C, Moser J, Yuen DA. 1994. The potential influence of radiative heat-transfer on the formation of megaplumes in the lower mantle. *Earth Planet. Sci. Lett.* 125:255–66

Mazor E, Heymann D, Anders E. 1970. Noble gases in carbonaceous chondrites. *Geochim. Cosmochim. Acta* 34:781–824

McKenzie D, Richter FM. 1981. Parameterized convection in a layered region and the thermal history of the Earth. *J. Geophys. Res.* 86:11667–80

McNamara AK, van Keken PE. 2000. Cooling of the Earth: a parameterized convection study of whole vs. layered models. *Geochem. Geophys. Geosyst.*, Vol. 1.

Mitrovica JX, Forte AM. 1995. Pleistocene glaciation and the Earth's precession constant. *Geophys. J. Int.* 121:21–32

Moreira M, Sarda P. 2000. Noble gas constraints on degassing processes. *Earth Planet. Sci. Lett.* 176:375–86

Nakamura Y, Tatsumoto M. 1988. Pb, Nd and Sr isotopic evidence for a multicomponent source for rocks of Cook-Austral Islands and heterogeneities of mantle plumes. *Geochim. Cosmochim. Acta* 52:2909–24

Nataf HC. 2000. Seismic imaging of mantle plumes. *Annu. Rev. Earth Planet. Sci.* 28: 391–417

Newsom HE, White WM, Jochum KP, Hofmann AW. 1986. Siderophile and chalcophile element abundances in oceanic basalts, Pb

Annu. Rev. Earth. Planet. Sci. 2002.30:493-525. Downloaded from arjournals.annualreviews.org
by UNIVERSITY OF OSLO on 03/30/06. For personal use only.

isotope evolution and growth of the Earth's core. *Earth Planet. Sci. Lett.* 80:299–313

O'Nions RK. 1987. Relationships between chemical and convective layering in the Earth. *J. Geol. Soc. London* 144:259–74

O'Nions RK, Evensen NM, Hamilton PJ. 1979. Geochemical modeling of mantle differentiation and crustal growth. *J. Geophys. Res.* 84:6091–101

O'Nions RK, Hamilton PJ, Evensen NM. 1977. Variations in 143Nd/144Nd and 87Sr/86Sr ratios in oceanic basalts. *Earth Planet. Sci. Lett.* 34:13–22

O'Nions RK, Oxburgh ER. 1983. Heat and helium in the Earth. *Nature* 306:429–31

O'Nions RK, Tolstikhin IN. 1994. Behaviour and residence times of lithophile and rare gas tracers in the upper mantle. *Earth Planet. Sci. Lett.* 124:131–38

Ottino J. 1989. *The Kinematics of Mixing: Stretching, Chaos and Transport*. Cambridge: Cambridge Univ. Press. 364 pp.

Palacz ZA, Saunders AD. 1986. Coupled trace element and isotope enrichment in the Cook-Austral-Samoa Islands, Southwest Pacific. *Earth Planet. Sci. Lett.* 79:270–80

Parsons B. 1981. The rates of plate creation and consumption. *Geophys. J. R. Astron. Soc.* 67:437–48

Paulssen H. 1988. Evidence for a sharp 670-km discontinuity as inferred from P-to-S converted waves. *J. Geophys. Res.* 93:10489–500

Pineau F, Javoy M. 1994. Strong degassing at ridge crests—the behavior of dissolved carbon and water in basalt glasses at 114N, Mid-Atlantic ridge. *Earth Planet. Sci. Lett.* 123:179–98

Pollack HN, Hurter SJ, Johnson R. 1993. Heat loss from the Earth's interior: analysis of a global data set. *Rev. Geophys.* 31:267–80

Porcelli D, Halliday AN. 2001. The possibility of the core as a source of mantle helium. *Earth Planet. Sci. Lett.* 192:45–56

Porcelli D, Wasserburg GJ. 1995a. Mass transfer of helium, neon, argon and xenon through a steady-state upper mantle. *Geochim. Cosmochim. Acta* 59:4921–37

Porcelli D, Wasserburg GJ. 1995b. Mass transfer of xenon through a steady-state upper mantle. *Geochim. Cosmochim. Acta* 59: 1991–2007

Poreda R, Schilling JG, Craig H. 1986. Helium and hydrogen isotopes in ocean ridge basalts north and south of Iceland. *Earth Planet. Sci. Lett.* 78:1–17

Poreda RJ, Farley KA. 1992. Rare gases in Samoan xenoliths. *Earth Planet. Sci. Lett.* 133:129–44

Richardson SH, Erlank AJ, Duncan AR, Reid DL. 1982. Correlated Nd, Sr and Pb isotope variation in Walvis Ridge basalts and implications for the evolution of their mantle source. *Earth Planet. Sci. Lett.* 59:327–42

Ritsema J, van Heijst HJ, Woodhouse JH. 1999. Complex shear wave velocity structure imaged beneath Africa and Iceland. *Science* 286:1925–28

Roden MF, Frey FA, Clague DA. 1984. Geochemistry of tholeiitic and alkalic lavas from the Koolau Range, Oahu, Hawaii: implications for Hawaiian volcanism. *Earth Planet. Sci. Lett.* 69:141–58

Roden MF, Trull TW, Hart SR, Frey FA. 1994. New He, Nd, Pb and Sr isotopic constraints on the constitution of the Hawaiian plume: results from Koolau Volcano, Oahu, Hawaii, USA. *Geochim. Cosmochim. Acta* 58:1431–40

Romanowicz B. 2001. Can we resolve 3D density heterogeneity in the lower mantle? *Geophys. Res. Lett.* 28:1107–10

Roy-Barman M, Allègre CJ. 1994. $^{187}Os/^{186}Os$ ratios of mid-ocean ridge basalts and abyssal peridotites. *Geochim. Cosmochim. Acta* 58:5043–54

Rudnick R, Fountain DM. 1995. Nature and composition of the continental crust: a lower crustal perspective. *Rev. Geophys.* 33:267–309

Sarda P, Graham DW. 1990. Mid-ocean ridge popping rocks: implications for degassing at ridge crests. *Earth Planet. Sci. Lett.* 97:268–89

Sarda P, Staudacher T, Allègre CJ. 1988. Neon isotopes in submarine basalts. *Earth Planet. Sci. Lett.* 91:73–88

Annu. Rev. Earth. Planet. Sci. 2002.30:493-525. Downloaded from arjournals.annualreviews.org
by UNIVERSITY OF OSLO on 03/30/06. For personal use only.

Schilling J-G. 1973. Icelandic mantle plume: geochemical evidence along the Reykjanes Ridge. *Nature* 242:565–71

Schilling J-G. 1975. Rare-Earth variations across 'normal segments' of the Reykjanes Ridge, 60°-53°N, mid-Atlantic ridge, 29°S, and East Pacific Rise, 2°-19°S, and evidence on the compostion of the underlying low-velocity layer. *J. Geophys. Res.* 80:1459–73

Schmalzl J, Houseman GA, Hansen U. 1996. Mixing in vigorous, time-dependent three-dimensional convection and application to Earth's mantle. *J. Geophys. Res.* 101:21847–58

Schubert G, Spohn T. 1981. Two-layer mantle convection and the depletion of radioactive elements in the lower mantle. *Geophys. Res. Lett.* 8:951–54

Shankland TJ, Nitsan U, Duba AG. 1979. Optical absorption and radiative heat transport in olivine at high temperature. *J. Geophys. Res.* 84:1603–10

Shen Y, Solomon SC, Bjarnason IT, Wolfe CJ. 1998. Seismic evidence for a lower mantle origin of the Iceland plume. *Nature* 395:62–65

Shirey SB, Walker RJ. 1998. The Re-Os isotope system in cosmochemistry and high-temperature geochemistry. *Annu. Rev. Earth Planet. Sci.* 26:423–500

Silver PG, Carlson RW, Olson P. 1988. Deep slabs, geochemical heterogeneity, and the large-scale structure of mantle convection: investigation of an enduring paradox. *Annu. Rev. Earth Planet. Sci.* 16:477–541

Silver PG, Chan WW. 1986. Observations of body wave multipathing from broadband seismograms: evidence for lower mantle slab penetration beneath the Sea of Okhotsk. *J. Geophys. Res.* 91:13787–802

Snow JE, Reisberg LC 1995. Os isotope systematics of the MORB mantle: results from altered abyssal peridotites. *Earth Planet. Sci. Lett.* 133:411–21

Spohn T, Schubert G. 1982. Modes of mantle convection and the removal of heat from the Earth's interior. *J. Geophys. Res.* 87:4682–96

Sprague D, Pollack HN. 1980. Heat-flow in the Mesozoic and Cenozoic. *Nature* 285:393–95

Staudacher T, Sarda P, Richardson SH, Allègre CJ, Sagna I, Dmitriev LV. 1989. Noble gases in basalt glasses from a Mid-Atlantic Ridge topographic high at 14 degrees N; geodynamic consequences. *Earth Planet. Sci. Lett.* 96:119–33

Stille P, Unruh DM, Tatsumoto M. 1983. Pb, Sr, Nd and Hf isotopic evidence of multiple sources for Oahu, Hawaii basalts. *Nature* 304:25–29

Stuart FM. 1994. Speculations about the cosmic origin of He and Ne in the interior of the Earth—comment. *Earth Planet. Sci. Lett.* 122:245–47

Su W, Woodward RL, Dziewonski AM. 1994. Degree 12 model of shear velocity heterogeneity in the mantle. *J. Geophys. Res.* 99:6945–80

Tackley PJ. 2000. Mantle convection and plate tectonics: toward an integrated physical and chemical theory. *Science* 288:2002–7

Tatsumoto M. 1978. Isotopic composition of lead in oceanic basalt and its implication to mantle evolution. *Earth Planet. Sci. Lett.* 38:63–87

Ten AA, Yuen DA, Podladchikov YY, Larsen TB, Pachepsky E, Malevsky A. 1997. Fractal features in mixing of non-Newtonian and Newtonian mantle convection. *Earth Planet. Sci. Lett.* 146:401–14

Torgersen T. 1989. Terrestrial helium degassing fluxes and the atmospheric helium budget—implications with respect to the degassing processes of continental crust. *Chem. Geol.* 79:1–14

Trull T. 1994. Influx and age constraints on the recylced cosmic dust explanation for high $^3He/^4He$ ratios at hotspot volcanoes. In *Noble Gas Geochemistry and Cosmochemistry*, ed. J Matsuda, pp. 77–88. Tokyo: TerraPub

Van der Hilst RD, Kárason H. 1999. Compositional heterogeneity in the bottom 1000 kilometers of Earth's mantle: toward a hybrid convection model. *Science* 283:1885–88

Van der Hilst RD, Widiyantoro S, Engdahl ER. 1997. Evidence for deep mantle circulation

Annu. Rev. Earth. Planet. Sci. 2002.30:493-525. Downloaded from arjournals.annualreviews.org
by UNIVERSITY OF OSLO on 03/30/06. For personal use only.

from global tomography. *Nature* 386:578–84

Van der Voo R, Spakman W, Bijwaard H. 1999. Mesozoic subducted slabs under Siberia. *Nature* 397:246–49

Van Keken PE, Ballentine CJ. 1998. Whole-mantle versus layered mantle convection and the role of a high-viscosity lower mantle in terrestrial volatile evolution. *Earth Planet. Sci. Lett.* 156:19–32

Van Keken PE, Ballentine CJ. 1999. Dynamical models of mantle volatile evolution and the role of phase transitions and temperature-dependent rheology. *J. Geophys. Res.* 104:7137–51

Van Keken PE, Ballentine CJ, Porcelli D. 2001. A dynamical investigation of the heat and helium imbalance. *Earth Planet. Sci. Lett.* 188:421–34

Van Keken PE, Zhong S. 1999. Mixing in a 3D spherical model of present day mantle convection. *Earth Planet. Sci. Lett.* 171:533–47

Van Schmus WR. 1995. Natural radioactivity of the crust and mantle. In *A Handbook of Physical Constants, AGU References Shelf 1*, ed. TJ Ahrens, pp. 283–91. Washington, DC: Am. Geophys. Union

Vidale JE, Schubert G, Earle PS. 2001. Unsuccessful initial search for a midmantle chemical boundary with seismic arrays. *Geophys. Res. Lett.* 28:859–62

Wen LX, Anderson DL. 1997. Layered mantle convection: a model for geoid and tomography. *Earth Planet. Sci. Lett.* 146:367–77

White WM. 1985. Sources of oceanic basalts: radiogenic isotopic evidence. *Geology* 13:115–18

White WM, Hofmann AW. 1982. Sr and Nd isotope geochemistry of oceanic basalts and mantle evolution. *Nature* 296:821–25

Wolfe CJ, Bjarnason IT, VanDecar JC, Solomon SC. 1997. Seismic structure of the Iceland mantle plume. *Nature* 385:245–47

Woodhead JD. 1996. Extreme HIMU in an oceanic setting: the geochemistry of Mangaia Island (Polynesia), and temporal evolution of the Cook-Austral hotspot. *J. Volcan. Geotherm. Res.* 72:1–19

Woodhead JD, McCulloch MT. 1989. Ancient seafloor signals in Pitcairn Island lavas and evidence for large amplitude, small length-scale mantle heterogeneities. *Earth Planet. Sci. Lett.* 94:257–73

Wright E, White WM. 1987. The origin of Samoa: new evidence from Sr, Nd and Pb isotopes. *Earth Planet. Sci. Lett.* 81:151–62

Wysession ME, Valenzuela RW, Zhu AN, Bartko L. 1995. Investigating the base of the mantle using differential travel times. *Phys. Earth Planet. Inter.* 92:67–84

Zhang YS, Tanimoto T. 1992. High-resolution global upper mantle structure and plate tectonics. *J. Geophys. Res.* 98:9793–823

Zindler A, Hart S. 1986. Chemical geodynamics. *Annu. Rev. Earth Planet. Sci.* 14:493–571

Zindler A, Hart SR, Frey FA, Jakobsson SP. 1979. Nd and Sr isotope ratios and rare earth element abundances in Reykjanes Peninsula basalts: evidence for mantle heterogeneity beneath Iceland. *Earth Planet. Sci. Lett.* 45:249–62

Zindler A, Jagoutz E. 1988a. Lead isotope systematics in spinel lherzolite nodules. *Chem. Geol.* 70:58–68

Zindler A, Jagoutz E. 1988b. Mantle cryptology. *Geochim. Cosmochim. Acta* 52:319–33

Annu. Rev. Earth. Planet. Sci. 2002.30:493-525. Downloaded from arjournals.annualreviews.org
by UNIVERSITY OF OSLO on 03/30/06. For personal use only.

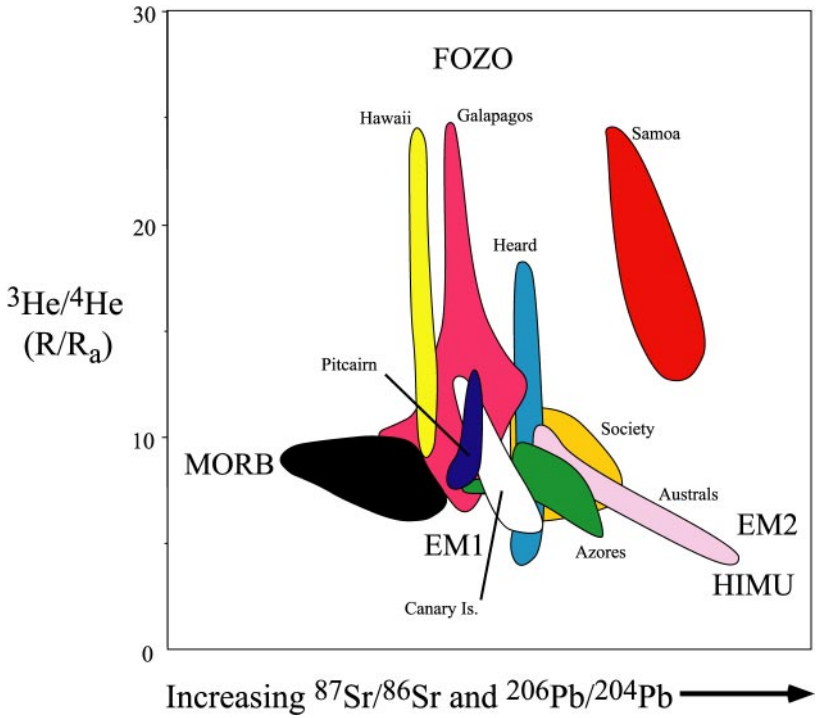
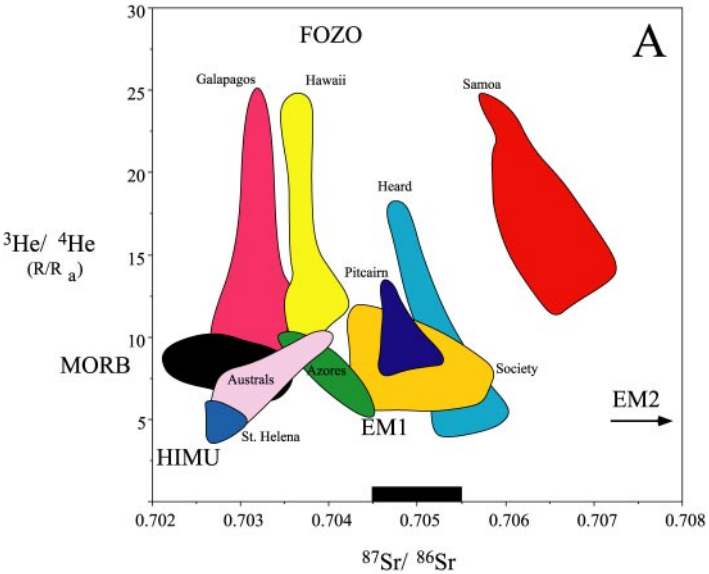


Figure 3 Two-dimensional projection of a 3D plot of Sr, Pb, and He isotopes in oceanic basalts, viewed with the Sr-Pb plane in the horizontal. Mantle end-members from Zindler & Hart (1986) and Hauri et al. (1994). The diagram has been rotated about the He axis so that both Sr and Pb isotope values increase to the right. Contours enclose data for individual hot spots and for MORB. Data are averaged by island, volcano, or geologic formation, except for Heard, Samoa, and MORB for which all data are plotted. The data converge toward a component high in $^3He/^4He$ (FOZO), with Sr and Pb isotope ratios distinct from MORB. Several less heterogeneous localities that are not shown (e.g., Easter, Bouvet, Iceland) plot toward the center of the diagram at high $^3He/^4He$ ratios. Data and references can be found in the PETDB (http://petdb.ldeo.columbia.edu/petdb/) and GEOROC (http://georoc.mpch-mainz.gwdg.de/Start.asp) geochemical databases.

Annu. Rev. Earth. Planet. Sci. 2002.30:493-525. Downloaded from arjournals.annualreviews.org
by UNIVERSITY OF OSLO on 03/30/06. For personal use only.



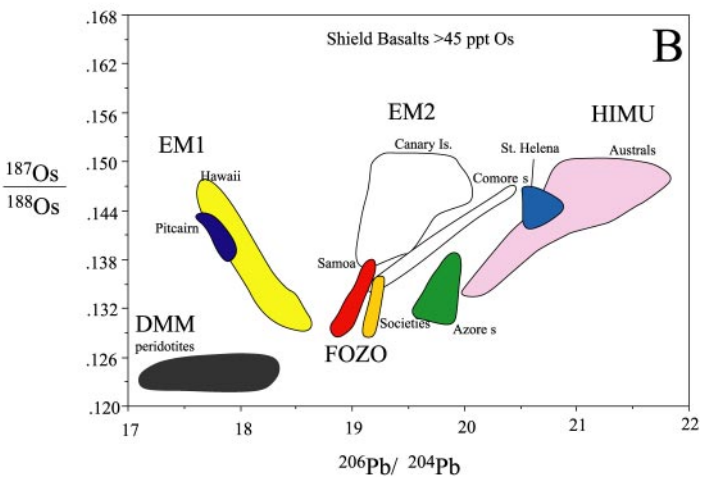


Figure 6 (*A*) Sr-He and (*B*) Os-Pb isotope correlations for oceanic basalts. All OIB show elevated Os isotope ratios, indicative of the presence of mafic material in their sources. Both data plots show elongated mixing arrays converging on the FOZO component.

Annu. Rev. Earth. Planet. Sci. 2002.30:493-525. Downloaded from arjournals.annualreviews.org
by UNIVERSITY OF OSLO on 03/30/06. For personal use only.

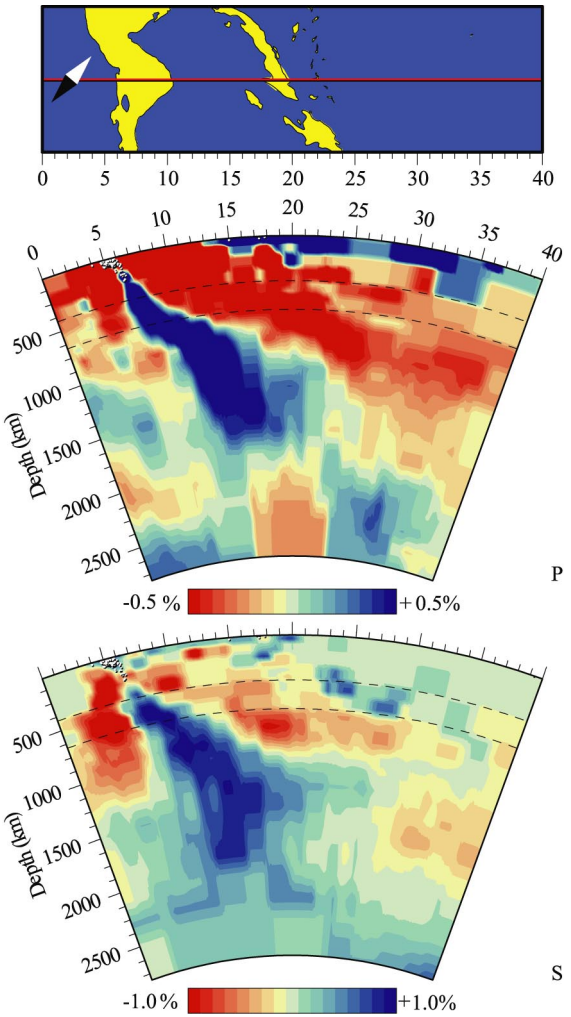


Figure 7 Seismic tomography provides strong evidence for the subduction of slabs into the lower mantle. This cross section through the mantle underneath Mexico and Cuba (*A*) shows the P and S wave tomographic models of Bijwaard et al. (1999). The color palettes are adjusted to focus on the lower mantle extension of the upper mantle subduction zones. Image courtesy of Wim Spakman.

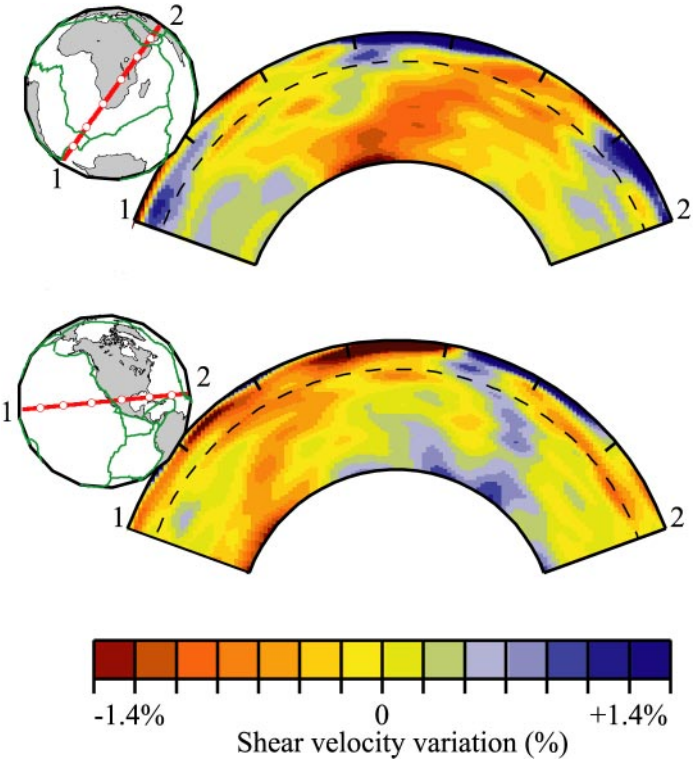


Figure 8 Tomographic models are increasingly capable of identifying low velocity anomalies that are most likely linked with convective upwellings away from subduction zones. This figure shows cross sections through the tomographic model S20RTS (Ritsema et al. 1999) that illustrate the potential existence of deep seated anomalies that feed the African (*A*) and Pacific (*B*) hot spot regions. Image courtesy of Jeroen Ritsema.

Annu. Rev. Earth. Planet. Sci. 2002.30:493-525. Downloaded from arjournals.annualreviews.org
by UNIVERSITY OF OSLO on 03/30/06. For personal use only.

Annu. Rev. Earth. Planet. Sci. 2002.30:493-525. Downloaded from arjournals.annualreviews.org
by UNIVERSITY OF OSLO on 03/30/06. For personal use only.

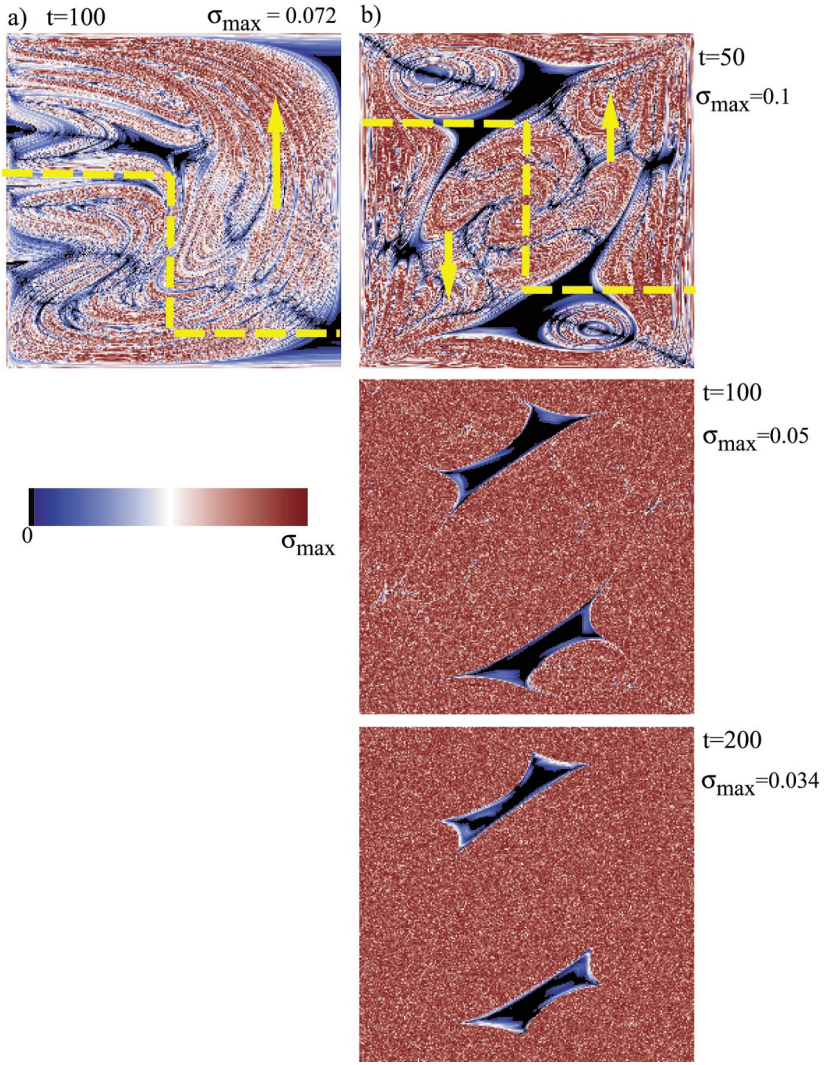


Figure 9 Toroidal effects in 3D convection can strongly modify conclusions based on 2D mixing models. These images illustrate the strong differences in the extent of mixing in a 3D Cartesian model of mantle convection driven by a surface flow that is prescribed by a ridge and a transform fault as in Ferrachat & Ricard (1998). The arrows in (*A*) and (*B*) indicate the surface geometry. Shown are the finite time Luyaponov exponents, σ, of particles that originated at mid-mantle depth. Higher σ corresponds to higher mixing efficiency. The models show that islands of laminar mixing ($\sigma \approx 0$) exist and get more defined as the mixing continues. The figures are based on a reevaluation of the models in Ferrachat & Ricard (1998) using the spectral-finite difference method of Gable et al. (1991) at high resolution (128^3).

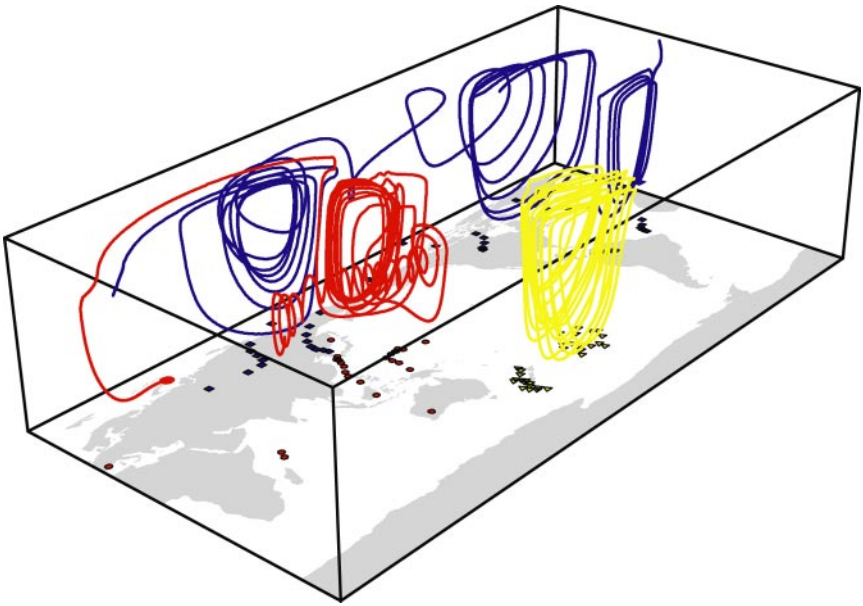


Figure 10 Mixing in a 3D spherical model of present day mantle convection shows different scales of mixing, ranging from simple poloidal flow (*yellow trace*) to cork screw patterns that are influence by toroidal energy at the surface (*blue* and *red traces*). The traces represent a particle trajectory for 4 Ga in the 3D spherical mixing model of Van Keken & Zhong (1999) and are represented in a Cartesian projection with a geographic map at the base of the box. The symbols on the map represent the intersections of the particle traces with a sphere at mid-mantle depth.

Annu. Rev. Earth. Planet. Sci. 2002.30:493-525. Downloaded from arjournals.annualreviews.org
by UNIVERSITY OF OSLO on 03/30/06. For personal use only.

Annu. Rev. Earth. Planet. Sci. 2002.30:493-525. Downloaded from arjournals.annualreviews.org
by UNIVERSITY OF OSLO on 03/30/06. For personal use only.

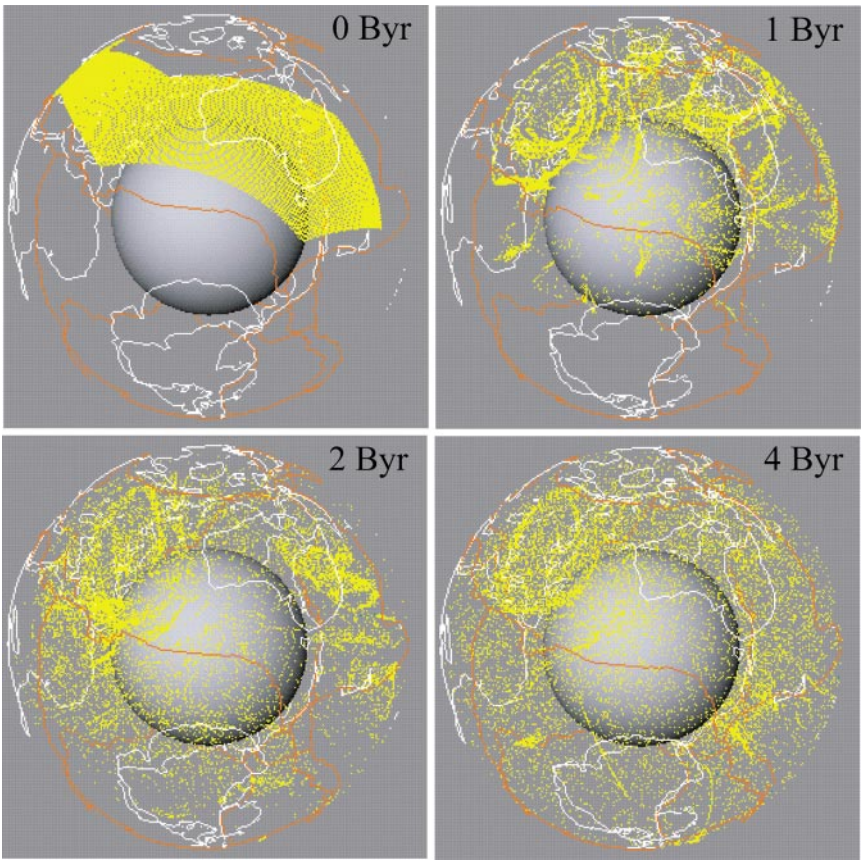


Figure 11 Evolution of the Australian-Indian plate in the steady state mixing model of Van Keken & Zhong (1999). Snapshots of the particles that composed the plate at time 0 are shown at 1, 2, and 4 Ga.

Annual Review of Earth and Planetary Sciences
Volume 30, 2002

CONTENTS

Frontispiece—*Vladimir Keilis-Borok* xvi

EARTHQUAKE PREDICTION: STATE-OF-THE-ART AND EMERGING POSSIBILITIES, *Vladimir Keilis-Borok* 1

MODELING COMPLEX, NONLINEAR GEOLOGICAL PROCESSES, *Greg A. Valentine, Dongxiao Zhang, and Bruce A. Robinson* 35

DATING THE TIME OF ORIGIN OF MAJOR CLADES: MOLECULAR CLOCKS AND THE FOSSIL RECORD, *Andrew B. Smith and Kevin J. Peterson* 65

MODERN INTEGRATIONS OF SOLAR SYSTEM DYNAMICS, *A. Morbidelli* 89

IMPLICATIONS OF EXTRASOLAR PLANETS FOR UNDERSTANDING PLANET FORMATION, *Peter Bodenheimer and D.N.C. Lin* 113

SCALING OF SOIL MOISTURE: A HYDROLOGIC PERSPECTIVE, *Andrew W. Western, Rodger B. Grayson, and Günter Blöschl* 149

STREAMFLOW NECESSARY FOR ENVIRONMENTAL MAINTENANCE, *Peter J. Whiting* 181

PETROLOGY OF SUBDUCTED SLABS, *Stefano Poli and Max W. Schmidt* 207

GEODYNAMO SIMULATIONS—HOW REALISTIC ARE THEY?, *Gary A. Glatzmaier* 237

MODERN IMAGING USING SEISMIC REFLECTION DATA, *Michael C. Fehler and Lianjie Huang* 259

PRELUDE TO THE CAMBRIAN EXPLOSION, *James W. Valentine* 285

PLUTO AND CHARON: FORMATION, SEASONS, COMPOSITION, *Michael E. Brown* 307

GEOLOGIC STRUCTURE OF THE UPPERMOST OCEANIC CRUST CREATED AT FAST-TO INTERMEDIATE-RATE SPREADING CENTERS, *Jeffrey A. Karson* 347

VOLCANOES, FLUIDS, AND LIFE AT MID-OCEAN RIDGE SPREADING CENTERS, *Deborah S. Kelley, John A. Baross, and John R. Delaney* 385

MANTLE MIXING: THE GENERATION, PRESERVATION, AND DESTRUCTION OF CHEMICAL HETEROGENEITY, *Peter E. van Keken, Erik H. Hauri, and Chris J. Ballentine* 493

Annu. Rev. Earth. Planet. Sci. 2002.30:493-525. Downloaded from arjournals.annualreviews.org
by UNIVERSITY OF OSLO on 03/30/06. For personal use only.

FOSSIL PLANTS AS INDICATORS OF THE PHANEROZOIC GLOBAL CARBON CYCLE, *D.J. Beerling and D.L. Royer* 527

INDEXES
Subject Index 557
Cumulative Index of Contributing Authors, Volumes 20–30 583
Cumulative Index of Chapter Titles, Volumes 20–30 586

ERRATA
An online log of corrections to *Annual Review of Earth and Planetary Sciences* chapters (if any, 1997 to the present) may be found at http://earth.annualreviews.org/

Annu. Rev. Earth. Planet. Sci. 2002.30:493-525. Downloaded from arjournals.annualreviews.org
by UNIVERSITY OF OSLO on 03/30/06. For personal use only.