

Theory of the Earth

Don L. Anderson

Chapter 11. Evolution of the Mantle

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Abstract:

I have now discussed the various kinds of magmas and refractory rocks that are important in the mantle, or at least the upper mantle, and some isotopic and seismic properties of the mantle. The visible rocks are the end, or present, product of mantle evolution. If these were our only source of information, we could come up with a fairly simple scheme of magma genesis, perhaps involving single-stage melting of a homogeneous, even primitive, mantle. It now seems unlikely that we will find the "Rosetta Stone" -- a rock fragment that represents "original Earth" or even the parent or grandparent of other rocks. Rocks and magmas represent products of complex multistage processes, just as the crust and mantle do. As we delve deeper into the Earth and further back in time, we depend more and more on isotopes and on modeling of planetary accretion and mantle processes.

Evolution of the Mantle

Rocks, like everything else, are subject to change and so also are our views on them.

—FRANZ Y. LOEWINSONLESSING

I have now discussed the various kinds of magmas and refractory rocks that are important in the mantle, or at least the upper mantle, and some isotopic and seismic properties of the mantle. The visible rocks are the end, or present, product of mantle evolution. If these were our only source of information, we could come up with a fairly simple scheme of magma genesis, perhaps involving single-stage melting of a homogeneous, even primitive, mantle. It now seems unlikely that we will find the "Rosetta Stone" — a rock fragment that represents "original Earth" or even the parent or grandparent of other rocks. Rocks and magmas represent products of complex multistage processes, just as the crust and mantle do. As we delve deeper into the Earth and further back in time, we depend more and more on isotopes and on modeling of planetary accretion and mantle processes.

A variety of studies have lent support to the concept of a chemically inhomogeneous mantle. The mantle contains at least two distinct reservoirs that differ in their large-ion-lithophile (LIL) contents and that have maintained their separate identities for at least 10^9 years. One reservoir, itself heterogeneous, has high values of Rb/Sr, Nd/Sm, U/Pb and, possibly, $^3\text{He}/\text{U}$. A variety of LIL-enriched basalts ranging from basanites, nephelinites and alkali basalts to tholeiites either form or evolve in, or become contaminated by, this reservoir. There may be several enriched reservoirs and several mechanisms of enrichment (subduction, trapping of melts or gas). The simplest hypothesis, however, is that there is one enriched reservoir that may be radially and laterally inhomogeneous, having been enriched at various times and by various mechanisms. The other major reservoir provides fairly uniformly depleted tholeiites (MORB) to the midocean ridges. The incompatible-element patterns

for MORB and enriched or alkali-rich basalts are complementary, and their ratios parallel the partition coefficient trend of garnet. This suggests that neither reservoir is primitive and that they are complementary products of a more primitive reservoir. The LIL-depleted reservoir appears to be garnet-rich, or fertile, relative to the enriched reservoir. The high density of garnet suggests that the depleted reservoir is the deeper one, and that ancient depletion/enrichment processes involved the upward transport of fluids that have been in equilibrium with garnet. The recycling of sediments and hydrothermally altered oceanic crust, too, has probably been important throughout most of Earth history. This material may contribute to the chemical and isotopic heterogeneity of the enriched source region, or it may be preferentially concentrated in the shallow mantle and act as a separate source.

From a geophysical and tectonic point of view the following possible sources can be identified. Some may be fertile sources of basalts, while others may simply interact with magmas generated elsewhere.

1. The continental shield lithosphere
2. The shallow mantle, sometimes called the asthenosphere.
3. The transition region.
4. Subducted sediments.
5. Subducted oceanic crust.

In addition, magmas may interact with the continental crust, the oceanic crust, the oceanic lithosphere, seawater and the atmosphere. Until all of these sources are evaluated, it seems premature to invoke such remote sources as the lower mantle and D".

MODELS OF THE MANTLE PETROGENESIS

Most models of petrogenesis assume that basic and ultrabasic magmas are formed by varying degrees of partial melting of fertile peridotite in the upper mantle. In these models the degree of partial melting and crystal fractionation and the depth of origin are the main variables in controlling the composition of the melt. Variations in volatile content and mineralogy of the source region, and crustal contamination are additional variables that have been used to explain the full range of magma compositions. Tholeiites are generally regarded as representing large degrees of partial melting of a shallow source, and the more K₂O-rich magmas, such as alkali basalts and nephelinites, are regarded as resulting from relatively small amounts of melting at deeper levels. The incompatible-element-enriched magmas with high LREE content, such as melilitites, nephelinites, basanites and kimberlites, are assumed to result from extremely small degrees of partial melting from a deep garnet-rich source region.

Isotopic studies require at least two source regions in the mantle. The most voluminous magma type, MORB has low LIL contents and low ⁸⁷Sr/⁸⁶Sr, ¹⁴³Nd/¹⁴⁴Nd and ²⁰⁷Pb/²⁰⁴Pb ratios. The Rb/Sr, Nd/Sm and U/Pb ratios in this reservoir have been low for at least 2.5 Ga, and perhaps were set by accretional differentiation. Since these ratios are high in melts and low in residual crystals, the implication is that the MORB reservoir is a cumulate or a crystalline residue remaining after the removal of a melt fraction. The enriched melt fraction is commonly assumed to efficiently enter the continental crust. The continental crust is therefore regarded as the only complement to the depleted MORB source. Mass-balance calculations based on this premise indicate that most of the mantle is undepleted or primitive, and the depleted reservoir is assumed to occupy most or all of the upper mantle. Since the continental crust is the only enriched reservoir in this model, magmas that have high LIL contents and ⁸⁷Sr/⁸⁶Sr ratios are assumed to be contaminated by the continental crust, or to contain a recycled ancient crustal component.

However, enriched magmas such as alkali-olivine basalts (AOB), oceanic-island basalts (OIB) and nephelinites also occur on oceanic islands and have similar LIL and isotopic ratios to continental flood basalts (CFB) and continental alkali basalts. Continental contamination is unlikely in these cases. Veins in mantle peridotites and xenoliths contained in alkali basalts and kimberlites are also commonly enriched and, again, crustal contamination is unlikely. In many respects these enriched magmas and xenoliths are also complementary to MORB (in LIL contents and isotopic ratios), suggesting that there is ancient enriched material in the mantle. Island-arc basalts are also high in LIL, ⁸⁷Sr/⁸⁶Sr, ¹⁴³Nd/¹⁴⁴Nd and ²⁰⁶Pb/²⁰⁴Pb, suggesting that there is a shal-

low and global enriched reservoir. Back-arc basin basalts (BABB) are closer to MORB in composition and, if the depth of the low-velocity zone and the depths of earthquakes can be used as a guide, tap a source deeper than 150 km. Many BABBs are intermediate in chemistry to MORB and OIB. This and other evidence indicates that the enriched reservoir is shallow and the depleted MORB reservoir is deeper.

Since the seismic velocities and anisotropy of the shallow mantle, above 220 km, are consistent with an olivine-rich aggregate, and since most mantle xenoliths are olivine-rich, it is natural to assume that the enriched reservoir is a peridotite. This peridotite can be infertile (low in Al₂O₃) yet enriched in LIL. Since the depleted reservoir has already lost a melt fraction, it should be depleted in garnet, unless it is an eclogite cumulate. Peridotites depleted in basalt have less Al₂O₃ and garnet than nondepleted or fertile peridotites.

The trace-element signatures of enriched magmas are consistent with derivation from a reservoir that has experienced eclogite fractionation or metasomatism by melts from an eclogite-rich source. Magmas from this reservoir will therefore be LREE-enriched, not because there is garnet in the residue but because the reservoir itself had experienced a prior stage of garnet removal or secondary enrichment by a fluid from a garnet-rich reservoir. This eliminates one of the arguments for derivation of LIL- and LREE-enriched magmas, such as alkali olivine basalts, by small degrees of partial melting from a deep garnet-rich peridotite layer. In general, diapirs from great depth will experience a greater degree of partial melting than diapirs originating at shallow depths. This is another argument against a shallow source for tholeiites. Extensive melting probably requires adiabatic ascent from a deep thermal boundary layer.

Kimberlites are among the most enriched magmas. Although they are rare, the identification of a kimberlite-like component in enriched magmas means that they may be volumetrically more important than generally appreciated.

The eclogite-fractionation, magma-mixing hypothesis for the evolution of the mantle (Anderson, 1982a,b,c) differs from conventional petrological assumptions in several ways: (1) The depleted source region is an eclogite-rich cumulate or a piclogite rather than a garnet peridotite; (2) enriched magmas are blends or hybrids of a depleted magma (MORB or picrite) and a melt from a shallow enriched peridotite reservoir; (3) the LIL pattern of enriched magmas, in particular LREE enrichment, is due to small degrees of partial melting of a garnet peridotite, or garnet-clinopyroxene fractionation from a picritic or tholeiitic magma. This pattern may also have been introduced into the source at some earlier time by metasomatic fluids.

There is general consensus that the depleted, or MORB, reservoir is depleted in the incompatible elements such as rubidium, neodymium, uranium and LREEs and that the very incompatible elements are depleted more than

the less incompatible elements, thereby giving low Rb/Sr, Nd/Sm and U/Pb ratios. The corresponding isotopic ratios indicate that the depletion event was ancient, perhaps dating back to the formation of the continental crust. The complementary enrichment events would likewise have been ancient. We use "depleted" to describe basalts and reservoirs that have low LIL contents and low LREE/HREE, Rb/Sr, Nd/Sm, U/Pb, $^{87}\text{Sr}/^{86}\text{Sr}$, $^{143}\text{Nd}/^{144}\text{Nd}$, $^{206}\text{Pb}/^{204}\text{Pb}$ and other ratios. A depleted reservoir can still be fertile, as it can provide basalts by partial melting. A garnet-clinopyroxenite cumulate, for example, can be depleted but fertile. Similarly, an enriched reservoir can be infertile, being low in CaO, Al_2O_3 , Na_2O and so on.

What are not agreed upon are the following:

1. The location of the MORB reservoir. It has been variously placed within the low-velocity zone, at the bottom of the low-velocity zone, and in the transition region, that is, anywhere between 220 km and 650 km. The homogeneity of MORB, on a global basis, the arguments that attribute it to large degrees of partial melting, the tendency of enriched interstitial fluids to migrate upward, and the possible contamination of the shallow mantle by subducted sediments and altered oceanic crust all argue for a relatively deep origin for the MORB reservoir. The time sequence of erupted magmas at island arcs, continental rifts and oceanic islands is consistent with a shallow enriched reservoir and a deeper LIL-depleted reservoir.
2. The composition of the MORB reservoir. The conventional petrological view is that MORBs result from large degrees of partial melting of a garnet peridotite. It now appears that midocean-ridge tholeiites are not primitive magmas but are the result of extensive crystal fractionation of a more picritic parental melt of a nonchondritic reservoir. Picrites imply that the parent peridotite be melted by 30 percent or more. The high degree of partial melting is required in order to generate high-MgO picritic magmas at shallow depths. The alternative point of view is that the depleted source region is a deep garnet-rich pyroxenite cumulate, such as a piclogite (olivine eclogite).
3. The nature of the depletion process. All workers agree that the source region of midocean-ridge tholeiites is depleted in the LIL elements and that this depletion was an ancient event. In the conventional view this depletion was the result of a small amount of partial melting and melt removal to the continental crust. Since the MORB source is obviously still fertile, in the sense that it can provide basalts, it must be garnet-rich in spite of having lost a basaltic component. The MORB source, in this model, must have remained homogeneous and must have escaped the early high temperatures implied by thermal history calculations. Alternatively, the depletion of the MORB source in a cooling Earth could be ex-

plained if it is a garnet-pyroxenite cumulate. The late-stage residual fluids in equilibrium with such a layer would be LIL- and LREE-enriched and, in this respect, kimberlitic. Such fluids, if buoyant, could form the proto-crust.

Fluids and melts are LIL-enriched, and they tend to migrate upward. Sediments and altered ocean crust, also LIL-enriched, re-enter the upper mantle at subduction zones. Thus there are several reasons to believe that the shallow mantle serves as a scavenger of incompatible elements, including the radioactive elements (U, Th and K) and the key tracers (Rb, Sr, Nd, Sm and, possibly, Pb and ^3He). The continental crust and lithosphere are commonly assumed to be the main repositories of the incompatible elements, but oceanic islands, island arcs and deep-seated kimberlites also bring LIL-enriched material to the surface. This is one reason for invoking a global upper-mantle enriched layer and for investigating reservoirs other than the continental crust in compiling inventories of the incompatible elements. Even a moderate amount of LIL in the upper mantle will destroy the arguments for a primitive lower mantle.

Recent metasomatism of an upper-mantle source region is sometimes invoked to explain LIL-enriched magmas, particularly those with low $^{87}\text{Sr}/^{86}\text{Sr}$ ratios. The source of the enriching fluid is seldom addressed. Recycling of sediments or remelting of subducted oceanic crust may explain some enriched magmas. These mechanisms do not explain the high $^3\text{He}/^4\text{He}$ ratios found in some oceanic-island basalts.

The continuum in trace elements and isotopic ratios between such enriched magmas as nephelinites and alkali-olivine basalts and depleted magmas such as MORB suggests that most mantle basalts represent mixtures between MORB, or its picritic parent, and partial melts from a shallow enriched reservoir. Mixing or contamination upon ascent is an alternative to recent metasomatism of the source reservoir.

The study of basalts combined with experimental petrology allow one to estimate the temperatures and pressures at which basalts might have separated from their source region. The amount of melting involved and the composition of the source region, however, cannot be determined by these means. The emphasis on olivine-rich and peridotite source regions is based on the following arguments:

1. Peridotite is consistent with seismic velocities for the shallow mantle, and basalts come from the mantle. The seismic arguments for an olivine-rich shallow mantle include both the velocities and the anisotropy. However, peridotites rarely have either the major-element or trace-element chemistry necessary to provide basalts of the required composition. Most of these rocks are refractory

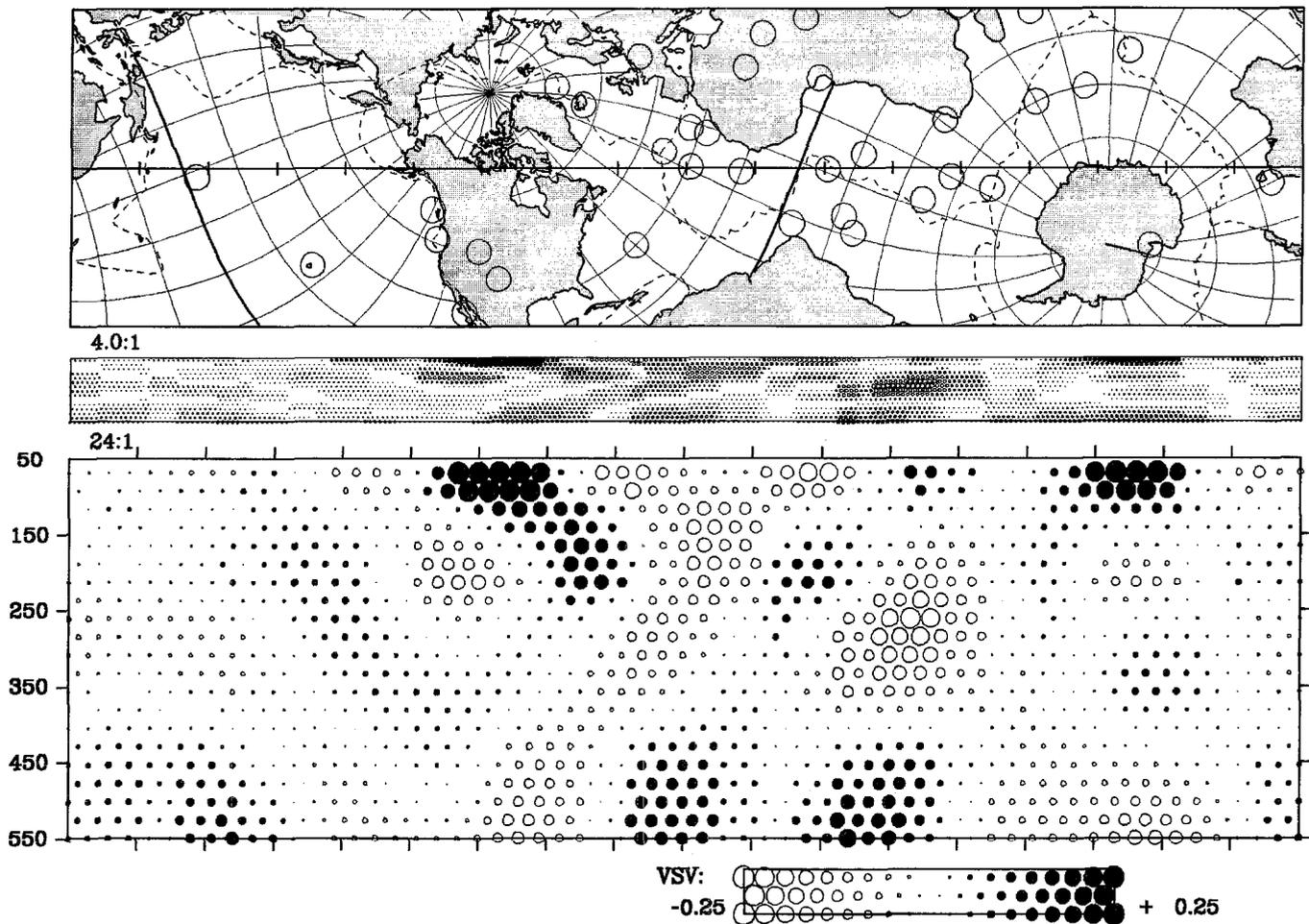
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FIGURE 11-1

Seismic shear velocities from 50 to 550 km depth (bottom) along great-circle path shown in center of top panel. Low velocities are open circles, fast velocities are filled circles. Vertical exaggeration is 24:1 in bottom panel and 4:1 in center panel. Note the slow velocities in the North and South Atlantic, where the cross section cuts across the Mid-Atlantic Ridge. Dashed lines are plate boundaries. Circles in upper panel are hotspots. (Model from Nataf and others, 1986.)

residuals or cumulates. In regions of very low seismic velocity, such as under tectonic regions and midocean ridges, the velocities are much lower than in peridotites. Partial melting or high-temperature grain boundary relaxation can reduce the velocities considerably, but when this occurs one can have almost any major-element chemistry for the matrix; that is, seismic velocities are no longer a constraint. The slow velocities associated with midocean-ridge upper mantle now appear to extend much deeper than 200 km, and, in some cases, into the transition region (Figure 11-1). Velocities in the transition region do not seem to be appropriate for peridotites. Thus, although arguments can be made that some parts of the shallow mantle have seismic velocities appropri-

ate for peridotite, the connection between these regions and the source regions for basalts has not been made.

2. The occurrences of garnet peridotite are appropriate to deep-seated environments. Garnet peridotite is stable in the upper mantle but, again, this does not prove that it occurs in the basalt source region or is the immediate parent for basalts. Most natural peridotites, in fact, are buoyant relative to fertile peridotites, eclogites and piclogite.
3. Garnet peridotites have close compositional relationships to meteorites. This is an argument that the average Earth or even the bulk composition of the mantle can be inferred from meteoritic abundances. Midocean-ridge

basalts clearly do not come from a chondritic source; the LIL elements are depleted in MORB relative to chondritic abundances. The MORB source is at least one generation removed from a chondritic ancestor. This argument, and many like it, confuse what may be in the mantle or what may be the average composition of the mantle with what is actually required of the immediate parent of basalts. A single-stage basalt forming process is implied whereas isotopic and trace-element data clearly require a multistage process.

4. Partial melts of natural samples of garnet peridotite at high pressure have basaltic compositions. Actually, partial melting of natural peridotites has not provided magmas with MORB compositions, particularly in the trace elements. Synthetic peridotitic aggregates (minus olivine) come close to matching inferred compositions of parental picrites, except for K_2O (Green and others, 1979). Melting of eclogites, of course, also gives basaltic composition melts, so the above argument, at best, is permissive rather than persuasive.
5. Melting of eclogites would have to be very extensive, and melt-crystal segregation would occur before such extensive melting can be achieved. Actually, eclogites provide basaltic melts over a wide range of melting, and large amounts of melting may be required for melt separation.

There is no doubt that garnet peridotites can and do come from the shallow mantle, and some regions of the upper mantle are probably mostly garnet peridotite. Some garnet peridotites can provide basalts by partial melting, as can pyrolite by definition. The average composition of the Earth is probably close to chondritic in major-element chemistry, and the mantle therefore contains abundant, although not necessarily predominant, olivine. By the same reasoning the mantle contains even more pyroxene and garnet. These arguments do not prove that the source region of the most abundant basalt type, midocean-ridge basalts, is garnet peridotite or that the regions of the mantle that appear to be peridotitic, on the basis of seismic velocities, are the regions where midocean basalts are generated.

Although some of the older ideas about the source regions, such as melting of a glassy or basaltic shallow source, can be ruled out, the possibility that a deep eclogite, or picritic eclogite, is the basalt source region cannot be ruled out. Olivine eclogite, or piclogite, the inferred composition of the transition region, is also a candidate source rock.

PETROLOGICAL EVOLUTION OF THE MANTLE

Most models of petrogenesis assume that so-called primary magmas are the result of varying degrees of partial melting

of peridotite. Abyssal tholeiites, because of their relative depletions in the LIL elements, have had a more complex history, although it seems likely that partial melting of a peridotite was involved at an early stage of their evolution. The trace-element inhomogeneity of the mantle (Chapter 13) plus the long-term isolation of the major reservoirs suggests that differentiation has been more effective in the long run than mixing. Mixing can be avoided in a chemically stratified mantle if the layers have a large intrinsic density contrast. Garnet has the highest density of any abundant upper-mantle mineral and therefore plays a role in determining the density of various regions of the mantle.

The conventional model for the origin of magmas might well be designated PM³ for "primitive mantle partial melt pyrolite model." In this model, tholeiitic basalts are considered to be primary, unfractionated melts resulting from about 20 percent melting of fertile garnet peridotite. In recent variations, oceanic tholeiites are treated as partial melts from a reservoir that has been depleted by removal of the continental crust. Continental and oceanic-island basalts (OIB) are assumed to be melts from a primitive undifferentiated lower mantle. The oceanic lithosphere is modeled as 6 km of basalt, the primary melt, and 24 km of depleted residual harzburgite (pyrolite = 1 part basalt, 4 parts depleted peridotite). Upon subduction, the lithosphere sinks to the core-mantle boundary (Ringwood, 1975). Except for a layer of depleted peridotite in the upper mantle, with perhaps some isolated blobs of eclogites, the mantle is uniform in composition and composed of pyrolite, which, by definition, can yield basaltic magmas by partial melting.

There are several problems with this model. There is increasing evidence that midocean-ridge tholeiites are not primary magmas but are the result of olivine fractionation from a more MgO-rich picritic parent. The harzburgite residue after removal of partial melt from a garnet peridotite is less dense than the parent and will remain in the upper mantle. Several billion years of seafloor spreading will fill up the entire upper mantle with this depleted residue. In addition, harzburgites do not appear to have been in equilibrium with MORB (Green and others, 1979).

The Moon differentiated early in its history and evolved into a series of cumulate layers. It is likely that the Earth did as well. Even if a large fraction of the heat of accretion was radiated away or was convected efficiently to the surface, it is difficult to construct geotherms that remain below the solidus of silicates during most of the accretion of the Earth (Chapter 1). Gravitational separation of the melt will concentrate the LIL elements and Al_2O_3 , CaO and SiO_2 into the upper mantle. In contrast to the conventional model, the "primitive mantle," which can yield basalt by partial melting, has already been processed into a depleted lower mantle and an upper mantle that on average is enriched. This processing occurs near the surface where material delivered by accretion, or by whole-mantle convection, crosses the melting zone. If olivine and orthopyroxene are the main residual phases, the incompatible-element en-

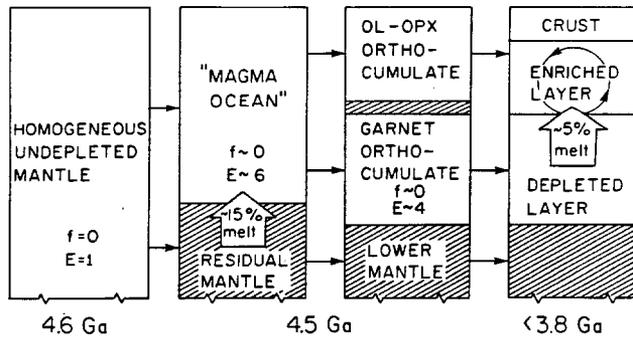


FIGURE 11-2

Differentiation of a planet during accretion and early high-temperature evolution. E is the enrichment of incompatible elements, relative to the starting materials. These elements have low crystal-melt partition coefficients and therefore readily enter the melt fraction. f is the fractionation factor and gives the ratio of two incompatible elements in the melt, expressed as the difference from the starting material. Very incompatible elements occur in the same ratio in melts as in the original, or primitive, material. Isotopic ratios of these elements will evolve at the same rate as in primitive material. A magma ocean will therefore be enriched but unfractionated. As the magma ocean crystallizes, the fractionating crystals will either float or sink, leaving behind an enriched, fractionated residual liquid layer. This fluid may permeate the shallow mantle, giving an enriched geochemical signature to this region, and to the continental crust. If melting and melt extraction are efficient in early Earth history, most regions of the planet will be depleted in incompatible elements. The large difference in crystallization temperature and density of olivine-orthopyroxene (ol-opx) and garnet means that mineralogically distinct regions can form in early Earth history.

richment E of the upper mantle will be relatively uniform at $1/F$ where F is the melt fraction (Figure 11-2). That is, for elements with low partition coefficients (such as Sm, Nd, Rb and Sr), the "primitive" upper mantle, a product of whole-mantle differentiation, will appear to be unfractionated (Chapter 8). Crystallization of a melt layer or magma ocean will lead to a series of cumulate layers, and fractionation of the LIL.

Although it is unlikely, and unnecessary, that the liquidus of peridotite is exceeded during accretion, the separation of melt from residual crystals in effect lowers the liquidus temperature from that of peridotite to that of basalt at low pressure and eclogite at high pressure. In the near-surface chill layer of a magma ocean, olivine and plagioclase are the low-pressure liquidus phases. At greater depth, clinopyroxene crystallizes. Since the liquidus of basalt and eclogite increases with depth faster than the adiabat, crystallization will also occur at the base of a magma ocean. At depths greater than 60 km, garnet and clinopyroxene crystallize from the melt. A deep eclogite cumulate layer is therefore one of the first products of crystallization of a deep (>50 km) magma ocean. The remaining fluid is depleted in Al_2O_3 , CaO, SiO_2 and Na_2O . Garnet and clino-

pyroxene between them retain FeO, MnO, yttrium, ytterbium, titanium, zirconium, and the heavy rare-earth elements (HREE). Olivine crystallizing at shallow levels sinks but will react with the silica-undersaturated fluid to form clinopyroxenes and garnet. It is buoyant relative to melt at high pressure. Olivine, therefore, is primarily a shallow mantle phase. The removal of eclogite from the melt means that there will not be much Al_2O_3 in the melt for formation of a thick, lunar-type anorthosite layer at the surface. The eclogite will probably be mixed with some olivine, which is also denser than the melt, at least in the shallow mantle.

The cumulate layers originally contain interstitial fluids that hold most of the incompatible elements. As crystallization proceeds, these melts migrate upward. Melts from an eclogite or olivine eclogite cumulate will be depleted in Al_2O_3 , CaO, Na_2O , yttrium, ytterbium, manganese, and the HREE. These are the characteristics of kimberlites. Removal of late-stage (kimberlite) intercumulus fluids from an eclogite-rich cumulate layer will deplete it and enrich the overlying olivine-rich layer. The enrichment, however, will be selective. It will be uniform in the very incompatible elements, giving primitive ratios of Rb/Sr, Sm/Nd and such, but will impart a pattern of depletion in the HREE, yttrium, sodium, manganese and so on since these are the eclogite-compatible elements. Partial melts from a shallow enriched reservoir will therefore appear to have a garnet-residual pattern, even if this reservoir contains no garnet. This pattern can be transferred to any MORB magmas interacting with this layer.

An eclogite-rich layer such as piclogite is normally denser than peridotite; therefore, the enriched, fertile and depleted, less fertile or infertile, reservoirs remain separate and isolated. However, convection in a chemically stratified system causes lateral variations in temperature, thermal boundary layers and deformation of the interfaces because of the buoyancy of the uprising currents. If this deformation raises a chemical boundary across the solidus, or if the temperature is perturbed by, for example, continental insulation, then partial melting can generate a buoyant diapir. Subsidiary reactions between garnet and clinopyroxene also occur at high temperature. This results in a temperature-induced density decrease much greater than can be achieved by thermal expansion. Adiabatic ascent of a diapir from a buried eclogite or piclogite layer (see Chapter 9) can lead to extensive melting because of the small amount of refractory component compared to olivine-rich (>60 percent) peridotites. Crystal settling apparently can be avoided in a rapidly rising diapir because of the high temperatures, temperature gradients and stresses (see Chapter 9). Depleted picritic melt, delivered to the shallow mantle, can evolve to tholeiites by shallow-level olivine fractionation and to alkali basalts by deeper clinopyroxene and garnet fractionation. Some of the melt may remain in the shallow mantle.

Ringwood's (1975) petrogenesis model assumes that tholeiites are primary magmas representing 20–30 percent

partial melting of pyrolite. Most of the mantle is assumed to be still capable of yielding a variety of basalts by varying degrees of partial melting. In the other extreme under the model presented here, all of the mantle has already been processed, and the melt products and their cumulates are concentrated in the upper mantle. The crystalline residue of the early differentiation, about 75 percent of the mantle, is now the lower mantle. Separation of upper and lower mantle may be irreversible. Transfer of material between the lower mantle and the upper mantle will be prohibited if the 650-km discontinuity is a chemical boundary with a sufficiently high intrinsic density contrast. The high melting temperature of the lower mantle and the small difference between melt density and crystal density at high pressure makes it unlikely that upper-mantle material can overcome a compositional density barrier by partial melting. The high FeO-content of the lower mantle, inferred from its density and from solar abundances (Chapter 1) is presumably a result of accretional fractionation and makes the Earth irreversibly differentiated.

With the recognition that tholeiites are not primary magmas, the temperatures and extent (or depth) of partial melting in the upper mantle have had to be revised upward. The maximum amount of melting of a mantle silicate is buffered by the latent heat of fusion, the specific heat and, most importantly, by the segregation of melt from crystal at high degrees of partial melting. High degrees of partial melting are probably restricted to the rising limbs of convection cells. The deep-mantle adiabat in these regions extends into the upper mantle, and temperatures well in excess of the solidus are feasible. With the higher heat productivity in the past, the heat is most efficiently removed if there are

more regions of upwelling or more rapid transfer of material through the upper-mantle melting zone. In either case, it is likely that large volumes of melts were generated in early Earth history, and melt/crystal separation is likely.

The process of planetary accretion and melting during accretion is akin to a zone-refining process. The surface of the planet, where the kinetic energy of accretion is turned into heat, acts as the furnace and refractory, "purified" material is fed into the planet. The incompatible elements and melts are preferentially retained near the surface. A deep magma ocean at any one time is not required, or even desirable. It is not desirable since if the surface melt layer is in equilibrium with dense phases such as perovskite, there should be anomalies in the trace-element patterns of upper-mantle materials. Large amounts of garnet extraction (—10 percent) and, presumably, majorite are allowed.

The crystallization behavior of a partial melt removed from a garnet peridotite has been discussed in detail by M. J. O'Hara and his coworkers (O'Hara and Yoder, 1967; O'Hara, 1968; O'Hara and others, 1975). Table 11-1 illustrates their scheme for generating tholeiites from primitive mantle. Early melting is likely to be extensive since large amounts of melt, 15–25 percent, occur in small temperature range just above the solidus. Eclogite extraction at moderate depth leaves a residual fluid deficient in CaO, Al₂O₃ and Na₂O with a high MgO and FeO content. Crystallization at shallow pressures gives an olivine cumulate and a residual fluid enriched in CaO, Al₂O₃, TiO₂, Na₂O and K₂O and low in MgO.

Table 11-2 gives a more detailed comparison of the possible products of mantle differentiation. The cosmochemical mantle model of Ganapathy and Anders (1974),

TABLE 11-1

Effect of Eclogite and Olivine Fractionation on Primitive Magma

Magma	SiO ₂	Al ₂ O ₃	FeO	MgO	CaO	TiO ₂	Na ₂ O	K ₂ O
1. Primitive	46.2	11.1	10.8	20.2	9.4	0.77	1.06	0.08
2. Extract	46.2	13.9	9.3	16.3	11.9	0.81	1.29	0.02
3. Picrite	46.2	8.3	12.3	24.1	6.9	0.74	0.83	0.14
Tholeiites								
4. Model	50.0	13.8	12.4	8.5	11.5	1.23	1.38	0.23
5. Hawaiian	50.0	14.1	11.4	8.6	10.4	2.53	2.16	0.39
6. Continental	50.6	13.6	10.0	8.5	10.0	1.95	2.90	0.54
7. Average oceanic	50.7	15.6	9.9	7.7	11.4	1.49	2.66	0.17

1. Possible primitive magma. The partial melt product of primitive mantle differentiation (O'Hara and others, 1975).

2. Eclogite extract (O'Hara and others, 1975).

3. Residual liquid after 50 percent eclogite (2) removal from primitive magma (1). This is a model picritic primary magma.

4. Residual liquid after a further removal of 40 percent olivine (Fo_{87.5}) from liquid (3).

5. Average Hawaiian parental tholeiite.

6. Continental tholeiite (Tasmania) (Frey and others, 1978).

7. Average oceanic tholeiite glass (Elthon, 1979).

TABLE 11-2

Composition of Mantle, Upper Mantle, Possible Picritic Parent Magmas and Eclogites

Material	SiO ₂	Al ₂ O ₃	FeO	MgO	CaO	TiO ₂	Na ₂ O	K ₂ O
Mantle and Upper Mantle Compositions								
1. Bulk mantle	48.0	5.2	7.9	34.3	4.2	0.27	0.33	
2. Residual mantle	48.3	3.7	7.1	37.7	2.9	0.15	0.15	
3. Pyrolite	45.1	3.3	8.0	38.1	3.1	0.2	0.4	
Possible Picritic Parent Magmas								
4. Eclogite extract	46.2	13.9	9.3	16.3	11.9	0.81	1.29	0.02
5. Oceanic crust	47.8	12.1	9.0	17.8	11.2	0.59	1.31	0.03
6. Tortuga dikes	47.3	13.6	9.8	17.6	9.6	0.79	0.89	0.06
7. High-MgO tholeiites	46.2 46.3	12.6 13.0	11.0 11.3	16.6 15.5	10.5 10.9	0.69 0.71	1.18 1.26	0.02 0.03
Kimberlite Eclogites								
8. Average	47.2	13.9	11.0	14.3	10.1	0.60	1.55	0.84
9. Roberts Victor	46.5	11.9	10.0	14.5	9.9	0.42	1.55	0.85

1. Bulk mantle composition (Ganapathy and Anders, 1974).
2. Residual after 20 percent extraction of primitive magma (line 1, Table 11-1).
3. This is an estimate of shallow mantle composition (Ringwood, 1975).
4. Possible eclogite extract from primary magma (O'Hara and others, 1975).
5. Average composition of oceanic crust (Elthon, 1979).
6. High magnesia Tortuga dike NT-23 (Elthon, 1979).
7. High magnesia-tholeiites.
8. Average bimineraleclogite in kimberlite.
9. Eclogite, Roberts Victor 11061 (O'Hara and others, 1975)

on line 1 of the table, uses the iron content (an element that condenses at high temperature) of the Earth and the observed heat flow to estimate the abundances of the refractory elements (the heat flow constrains the abundances of U and Th, two of the refractory elements). The relative abundance of the refractory elements is chondritic. The composition of pyrolite is given in line 3. Various candidates for picritic parent magmas are given in lines 4 to 7. The composition of eclogites in kimberlites are also given. These are possible cumulates from the primary melt and also possible parents for MORB. Note the similarity between the eclogite and picrite compositions.

These tables illustrate the plausibility of a large eclogite-rich cumulate layer in the mantle that has properties similar to those inferred for the picritic parent magma of MORB. An eclogite or basalt layer representing about 10–15 percent of the mantle can reconcile the major-element compositions inferred by cosmochemical, geophysical and petrological techniques.

Table 11-3 delineates a self-consistent petrological and geochemical model of the mantle and crust as it traces the enrichment and fractionation of the incompatible elements during the various stages of mantle differentiation. All results are given relative to the starting composition. The fractionation factors, for Rb/Sr and Sm/Nd, are also given for the products at various stages. The level of enrichment varies from stage to stage, but fractionation of rubidium relative to strontium and samarium relative to neodymium is only significant in the solid residues of partial melting

and in melts that represent only a small degree of partial melting (or late-stage residual fluids). Thus, an "enriched" reservoir can have unfractionated Rb/Sr and Sm/Nd ratios and can give "primitive" ratios of ⁸⁷Sr/⁸⁶Sr and ¹⁴³Nd/¹⁴⁴Nd. This fairly obvious point needs to be stressed because it has been overlooked by advocates of an undifferentiated primitive mantle.

In addition to demonstrating the effects of partial melting, crystal fractionation and redistribution of an enriched fluid from a crystallizing cumulate layer (mantle metasomatism), Table 11-3 also compares the composition of MORB, continental flood basalts, continental crust and kimberlites with the various products of this differentiation. In order to make this comparison I have normalized to the whole-mantle concentrations of Ganapathy and Anders (1974). Most other recent estimates of the abundances of refractory trace elements in the Earth are within 15 percent of these values and volatile-refractory ratios are within a factor of 2, generally much closer. The main assumption common to all estimates is that the large-ion incompatible refractories occur in the Earth in the same relative proportions as they occur in chondrites. This seems to be valid since these elements apparently do not undergo significant preaccretional, nonmagmatic fractionation.

These calculations indicate that the source reservoir for midocean-ridge basalts may be an eclogite-rich cumulate which has been depleted by removal of a kimberlite-like fluid. A deep eclogitic cumulate layer can become unstable at depth as it warms up, due to garnet-clinopyroxene

TABLE 11-3

Effect of Partial Melting and Eclogite and Peridotite Fractionation from Primary Melt,
Relative to Primitive Mantle

Material	K	Rb	Sr	La	Sm	Nd	Yb	$f_{\text{Rb/Sr}}$	$f_{\text{Sm/Nd}}$
1. Primitive mantle	1	1	1	1	1	1	1	0	0
2. Primitive melt	3.98	3.97	3.86	3.93	3.64	3.82	2.33	0.03	-.05
3. Primary eclogite cumulate	3.33	3.32	3.30	3.30	3.25	3.27	2.78	.01	-.01
4. Primary peridotite residual	6.58	6.31	6.10	6.45	5.28	6.02	0.53	.03	-.12
5. Depleted eclogite	1.06	1.10	2.45	1.77	2.84	2.57	2.83	-.55	.11
6. Model MORB	2.12	2.20	4.85	3.52	5.59	5.09	5.46	-.55	.10
7. Average MORB	2.9	1.5	3.4	2.5	6.1	4.9	6.5	-.56	.24
8. Enriched extract from eclogite	60.8	59.6	24.9	42.0	13.2	21.0	1.43	1.39	-.37
9. Continental crust	59.5	58.1	14.9	26.8	9.7	20.5	5.12	2.90	-.53
10. Kimberlite	41.3	75.6	27.5	211	34.2	65.9	2.79	1.75	-.48
11. Enriched reservoir	6.03	5.84	4.11	5.08	3.19	3.89	2.76	0.42	-.18
12. Model enriched tholeiite	30	29	20	25	14	18	7	0.45	-.22
13. Average continental tholeiite	25	18	13	22	18	22	12	0.38	-.18

Modified from Anderson (1982a).

1. All values in table are normalized to primitive mantle. All calculations based on equilibrium partial melting or equilibrium fractional crystallization. Partition coefficients from Frey and others (1978). f is $(\text{Rb/Sr})-1$ or $(\text{Sm/Nd})-1$.

2. 25 percent melt, garnet peridotite residual.

3. 50 percent garnet; 50 percent interstitial fluid. Note that this is nearly unfractionated. This layer and its fluids evolve as nearly primitive material until melt extraction occurs. This is model MORB reservoir prior to depletion. This garnet-rich reservoir becomes depleted as it further crystallizes and expels the intercumulus liquids. The fluid at this point is enriched but not very fractionated.

4. Residual fluid after removal of eclogite (3) from primitive melt (2). This evolves into a shallow peridotite source region (above the garnet cumulate layer). It is somewhat enriched and becomes more enriched as fluid is added to it from the underlying crystallizing garnet cumulate layer.

5. Remove 75 percent of final 5 percent melt fraction from garnet cumulate (3). This is a model for the depleted (MORB) reservoir. The late-stage fluid is both enriched and fractionated.

6. Remove 50 percent olivine plus orthopyroxene from depleted eclogite (5).

7. Midocean-ridge tholeiite (MORB); combined range of uncertainties in MORB and mantle compositions is greater than ± 50 percent.

8. 5 percent melt extract from garnet cumulate (3). This is used to enrich the peridotite layer (4) and form continental crust (9). This enriched fluid can be compared with continental crust (9) and kimberlite (10).

9. Continental crust.

10. Kimberlite.

11. Enriched source region. Two-thirds of melt extract from garnet cumulative layer used to enrich the primary residual fluid (4); one-third removed to continental crust. Composition of enriched reservoir; 5 parts (4), 0.5 part (8), 5 parts peridotite.

12. 20 percent melt of (11). This is model tholeiite from enriched source region (continental tholeiite, ocean-island tholeiite). Alkali basalts and other continental basalts are generated from this reservoir by smaller degrees of partial melting. Alternatively, these basalts are mixtures of MORB and a melt from 11.

13. Continental tholeiite.

reactions or to partial melting. The seismic properties of the transition region are consistent with a picritic eclogite (piclogite) composition (Anderson and Bass, 1986). Furthermore, the seismic velocity anomalies under ridges can be traced to 400 km or deeper (Figure 11-1).

These calculations show that extensive quantities of garnet can be removed from the shallow mantle or, put another way, that the upper mantle could have been molten and in equilibrium with a large amount of garnet. Majorite, a high-pressure garnet-like form of pyroxene, probably has similar partition coefficients. Therefore a deep magma ocean, say 300 km deep, is not inconsistent with trace-

element data. Actually, the lower mantle can be depleted in LIL during accretion, by zone refining, even if the magma ocean is much shallower.

CHEMICAL STRATIFICATION OF THE MANTLE

Chemical stratification resulting from early differentiation of the mantle, upward removal of the melt and fractionation via crystal settling of the resulting magma is one way to explain the presence of separate and chemically distinct res-

ervoirs. In the first stage, probably during accretion, the incompatible elements (including Rb, Sr, Nd, Sm, and U) are concentrated into melts (zone refined) and the upper mantle is more or less uniformly enriched in these elements. As the upper-mantle magma layer cools, a garnet-rich cumulate, containing intercumulus fluids, forms at the base. Peridotitic cumulates or melts concentrate at shallower depths. The deep garnet-rich cumulate layer can have near-primitive ratios of Rb/Sr and Sm/Nd if it contains a moderate amount of interstitial fluid. Transfer of a late-stage melt (KREEP or kimberlite) is one mechanism by which this layer may become depleted and the complementary region enriched. For this type of model, the isotopic ratios will be a function of the crystallization (fractionation) history of the upper mantle and the history of redistribution of LIL-enriched fluids. It is possible that even the most depleted MORB is contaminated by magmas from the enriched reservoir. The MORB reservoir may therefore be more depleted than generally assumed.

To calculate the isotopic evolution of a reservoir, relative to primitive mantle, we need only know how Rb/Sr and Sm/Nd vary relative to primitive ratios. The entries in Table 11-3 are therefore all relative to primitive, undifferentiated mantle. Since the composition of the melt and the crystals are complementary, the composition of a cumulate of crystals and interstitial fluids will appear less fractionated than either component. Intercumulus material typically constitutes 35–50 percent of an orthocumulate (crystals plus frozen liquids).

The calculations displayed in Table 11-3 show that relatively "primitive" ratios of incompatible elements and, therefore, isotopic ratios, do not necessarily imply a primitive reservoir. In particular, orthocumulates are less fractionated than either of the components. The highly fractionated kimberlites may represent the late-stage fluids of an apparently primitive crystallizing cumulate layer. KREEP probably formed on the Moon by a similar process. Other evidence indicates that the kimberlite source region, which is probably below 200 km depth, is garnet- and clinopyroxene-rich. Reaction relations between olivine, orthopyroxene and the melt also favor a garnet-clinopyroxene-rich cumulate at depth. A garnet-clinopyroxene-rich residue, even after extensive partial melting, is possible at high pressure.

What is the fate of eclogite in the mantle? Ringwood's (1975) model assumes that it sinks to the core-mantle boundary and is removed from the system. Estimated densities as a function of depth for eclogite and garnet peridotite are shown in Figure 11-3. Eclogite and garnetite (garnet solid-solution) are denser than peridotite to depths at least as great as 500 km. On the other hand the post-spinel phases of olivine and the perovskite form of orthopyroxene are denser than garnetite or ilmenite eclogite. Eclogite-rich cumulates, or subducted eclogitic lithosphere, are therefore unlikely to sink into the lower mantle. Whether eclogite can sink below 500 km depends on temperature and the com-

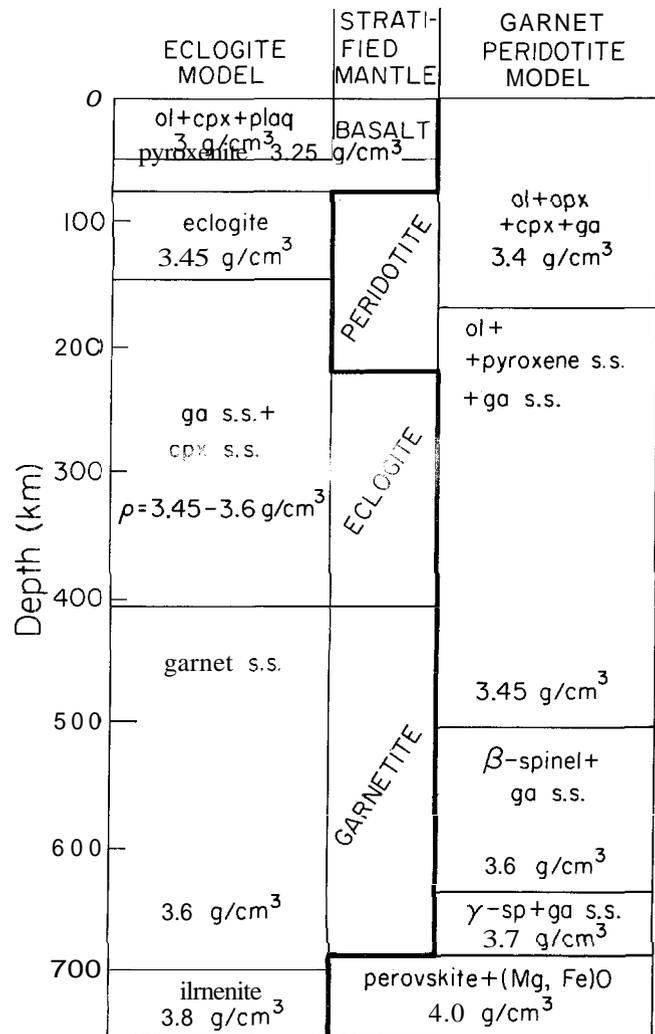


FIGURE 11-3 Approximate densities of basalt/eclogite (left) and garnet peridotite (right). Eclogite (subducted oceanic lithosphere or a cumulate in a deep magma ocean) is denser than peridotite until olivine converts to β -spinel. Below some 400 km the garnet and clinopyroxene in eclogite convert to garnet solid solution. This is stable to very high pressure, giving the mineralogical model shown in the center, the gravitationally stable configuration. The heavy line indicates that the bulk chemistry varies with depth (eclogite or peridotite).

pressibility and thermal expansivity relative to peridotite. The abrupt termination of earthquakes at 670 km suggests that oceanic lithosphere can penetrate to this depth. The sharpness of the discontinuity near 650 km is consistent with a chemical interface. This may be the lower boundary of the eclogite-rich layer. If eclogite represents 10 percent of the mantle, this would imply a thickness of about 200 km. The top of the eclogite layer would therefore be at a depth of 400 km but could rise to the shallow mantle when partially molten. Below some 400 km pyroxene dissolves in garnet to form a garnet solid solution (S.S.). This is

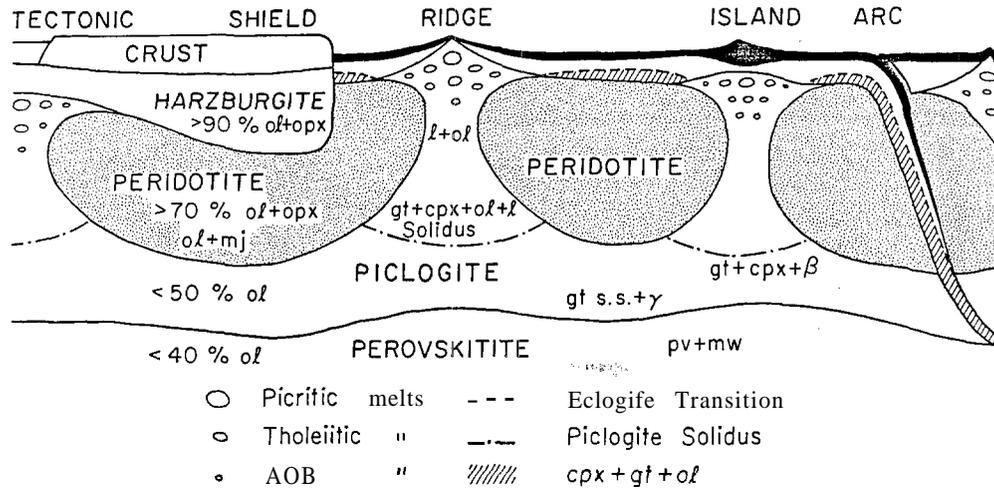


FIGURE 11-4

A possible configuration of the major rock types in the mantle. Peridotite accumulates in the shallow mantle because of its low density, particularly residual or infertile peridotite. Fertile peridotite, eclogite, olivine eclogite and piclogite are denser and are, on average, deeper. Partial melting, however, can reverse the density contrast. Upwellings from a fertile piclogite layer are shown under ridges, islands and tectonic continents. The isotopic and trace-element contents of the magmas are controlled by tectonic and shallow-mantle and crustal processes. The continental lithosphere is stable and is dominantly infertile harzburgite. It may have been secondarily enriched by upward migration of fluids and melts. The shallow mantle is predominantly infertile peridotite (stippled). This is underlain by piclogite, which, however, can rise into the shallow mantle if it becomes hot, partially molten or entrained. Melts from the piclogite layer also underplate the lithosphere, giving a dense, eclogite-rich base to the oceanic lithosphere, which can recycle into the transition region. The top parts of the subducted slab rejuvenate the shallow mantle with LIL. The slab is confined to the upper mantle, being too buoyant to sink into the lower mantle.

termed garnetite, the high-pressure form of eclogite. Piclogite contains olivine as well as eclogite and a piclogite layer can be thicker than an eclogite layer, for a given garnet content of the mantle.

The spatial relationship of the geochemical reservoirs is uncertain. If these were generated by petrological processes such as partial melting and crystal fractionation, the melts, residues, and cumulates would differ in bulk chemistry as well as in trace-element chemistry. The intrinsic density increases in the order basalt, picrite, depleted peridotite, fertile peridotite, eclogite. Basalts and picrites crystallizing or recrystallizing below about 50 km transform to eclogite, which is the densest upper-mantle assemblage. Garnet-poor and olivine-rich residues or cumulates are likely to remain at the top of the upper mantle since they are less dense than parental peridotites and do not undergo phase changes in the upper 300 km of the mantle. Eclogites are about 15 percent and 3 percent denser than basalts and fertile peridotites, respectively. With a coefficient of thermal expansion of $3 \times 10^{-5}/^{\circ}\text{C}$, it would require temperature differences of 1000–5000°C to generate similar density contrasts by thermal effects alone, as in normal thermal convection, or to overcome the density contrasts in these assemblages.

Simple Stokes' Law calculations show that inhomogeneities having density contrasts of 0.1 to 0.4 g/cm³ and dimensions of 10 km will separate from the surrounding mantle at velocities of 0.5 to 2.5 m/yr in a mantle of viscosity 10²⁰ poises. This is orders of magnitude greater than average convective velocities. Inhomogeneities of that magnitude will be generated by partial melting as material is brought across the solidus in the normal course of mantle convection. The higher mantle temperatures in the past make partial melting in rising convection cells even more likely and the lowered viscosity makes separation even more efficient. It seems unlikely, therefore, that chemical inhomogeneities can survive as blobs entrained in mantle flow for the long periods of time indicated by the isotopic data. Gravitational separation is more likely, and this leads to a chemically stratified mantle like that shown in Figures 11-4 and 11-5. The unlikely alternative is that the reservoirs differ in trace elements but not major elements. Small differences in bulk chemistry change the mineralogy and therefore the intrinsic density, and, in general, mineralogy is more important than temperature in generating density inhomogeneities.

The density differences among basalt, depleted peridotite, fertile peridotite, and eclogite are such that they can-

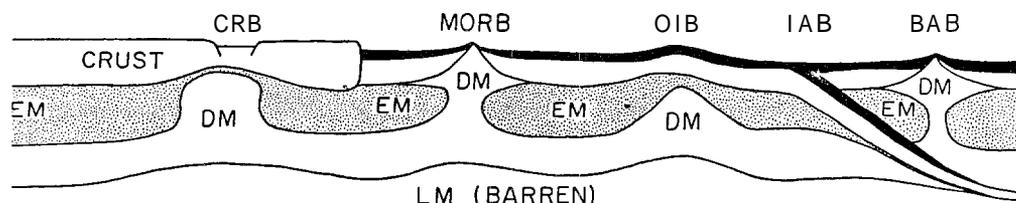


FIGURE 11-5

Possible configurations of geochemical reservoirs (DM, Depleted Mantle; EM, Enriched Mantle; LM, Lower Mantle; CRB, Continental Rift Basalts; MORB, Mid-ocean-Ridge Basalts; OIB, Oceanic-Island Basalts; IAB, Island-Arc Basalts; BAB, Back-Arc Basalts). In this model EM is heterogeneous and probably not continuous. It is isotopically heterogeneous because it has been enriched at various times. LM does not participate in plate tectonic or hotspot volcanism. Contrast this with the model of Wasserburg and DePaolo (1979), which has OIB and CFB coming from a primitive lower mantle, and that of Allegre and others (1982), which has OIB coming from the lower mantle and a uniform upper mantle providing MORB.

not be reversed by the kinds of temperature differences normally encountered in mantle convection. However, phase changes such as partial melting and basalt–eclogite involve large density changes. A picritic or pyroxenitic lithosphere, for example, will be less dense than fertile peridotite at depths shallower than about 50 km where it is in the plagioclase or spinel stability field. As the lithosphere thickens and cools, it becomes denser at its base than the underlying mantle and a potential instability develops. Similarly, if the temperature in a deep garnet-rich layer exceeds the solidus, the density may become less than the overlying layer. The large density changes associated with partial melting and the basalt–eclogite phase change may be more important in driving mantle convection than thermal expansion.

NATURE OF THE ENRICHMENT/DEPLETION PROCESS

Detailed modeling of the isotopic and trace-element characteristics of the enriched and depleted source regions of the mantle show that these regions are genetically related. The continental crust is also approximately complementary to the MORB source, but it was probably not derived by a single-stage partial melting event operating on primitive mantle. In detail, crust plus depleted mantle does not equal primitive or chondritic mantle. Although continental crust recycling via sediment subduction can increase the LIL and certain isotopic ratios in the shallow mantle, it cannot explain the high $^3\text{He}/^4\text{He}$ ratios of some enriched magmas and other characteristics of the enriched source region.

The melting of ancient subducted oceanic crust has also been proposed as the source of enriched magmas. The subducted crust, being at the top of the slab, may be exposed to high temperatures and, as evidenced by the seismic

activity, high stresses, and is probably fractured and possibly melted shortly after subduction. It is not clear how or where ancient subducted crust might be stored. It is doubtful that it would sink into the lower mantle, or even into the transition region. Partial melting of depleted MORB, in the eclogite stability field, gives magmas that are more LIL-enriched than MORB and, for small degrees of melting, can have LREE enrichment. The very high LIL contents and high $^3\text{He}/^4\text{He}$ ratios of enriched magmas cannot be explained by this mechanism. A source enriched in LIL, LREE, and $^{87}\text{Sr}/^{86}\text{Sr}$ as well as $^3\text{He}/^4\text{He}$ is required. The top part of the oceanic crust is altered by hydrothermal activity, and this is a more promising source if the altered part of the crust can survive the subduction process. However, the isotopic data require that any subducted oceanic crust contribution to the enriched reservoir must be ancient.

The emphasis on recycling and continental contamination as mechanisms of enrichment is based on an all-or-nothing philosophy such that when primitive mantle is melted it only forms continental crust and a complementary depleted (MORB) reservoir. In this philosophy present melts from the mantle are either primitive, from previously unfractionated mantle, or depleted. Most mantle magmas are, in fact, in the depleted quadrant of the mantle array (high ϵ_{Nd} , low $^{87}\text{Sr}/^{86}\text{Sr}$). However, MORBs are enriched in the lead isotopes and have $^3\text{He}/^4\text{He}$ ratios higher than atmospheric. This suggests that mantle magmas are mixtures of a depleted and enriched component, with the depleted component dominant.

Although the high LIL content of the crust and the argon-40 content of the atmosphere suggest efficient upward transport of volatiles and the incompatible elements, the high $^3\text{He}/^4\text{He}$ ratio of the mantle indicates that this process has not gone to completion. The process that put the LIL into the crust can also be expected to enrich the shallow mantle.

Semantics has played a subtle part in most petrological

models. It has long been recognized that midocean-ridge basalts are low in LIL and $^{87}\text{Sr}/^{86}\text{Sr}$ compared to continental magmas. Since MORB do not pass through continental crust, it has been common practice to refer to depleted basalts as mantle magmas and to assume that low LIL and $^{87}\text{Sr}/^{86}\text{Sr}$ is a characteristic of the mantle. High-LIL and high- $^{87}\text{Sr}/^{86}\text{Sr}$ basalts (and high $\delta^{18}\text{O}$) therefore are deemed crustal or crustal-contaminated. When enriched magmas were found on oceanic islands, it was difficult to accept these as mantle melts since, by definition, the mantle was depleted. Thus, some islands were assumed to rest on continental crust or to represent melting of recycled crustal material. As the number of enriched islands and seamounts increased, this became an untenable position, and the idea became popular that deep mantle plumes provided melts to the islands. Some of the early $^{143}\text{Nd}/^{144}\text{Nd}$ results for islands, kimberlites and continental flood basalts fell near the chondritic growth curve, and it was presumed then that hot-spots were the result of deep mantle plumes from primitive, unfractionated mantle. This ignored the lead-isotopic evidence. As data accumulated it became clear that island basalts ranged from very depleted to very enriched in $^{143}\text{Nd}/^{144}\text{Nd}$ and $^{87}\text{Sr}/^{86}\text{Sr}$ (and always enriched in $^{206}\text{Pb}/^{204}\text{Pb}$ and $^{207}\text{Pb}/^{204}\text{Pb}$). Some geochemists still maintained that the most enriched islands were continental fragments in spite of geophysical evidence to the contrary. Tristan da Cunha, for example, is one of the most enriched islands but lies close to the axis of the Mid-Atlantic Ridge. It is also low in $^3\text{He}/^4\text{He}$.

Although there are a variety of ways to enrich the shallow mantle by present tectonic processes, the isotopic evidence points toward ancient enrichment as well. We also, of course, must start forming the oceanic crust early on.

A crystallizing cumulate layer contains interstitial melt that becomes progressively more enriched and fractionated as crystallization proceeds. Fluids from such a layer may be the source of the original crust and the metasomatic fluids that have invaded the upper mantle at various times and may account for the progressive enrichment of the source region for oceanic-island and continental basalts.

A depleted reservoir can represent either a cumulate or a region of the mantle that has been subjected to melt extraction. Because of packing considerations and the fact that crystallization at depth is slower than crystal settling, an early cumulate probably retains a large amount of interstitial fluid. Removal of a late-stage melt is therefore also required in this case to generate a very depleted reservoir. The composition of the residual fluid in equilibrium with eclogite, as a function of crystallization, is shown in Figure 11-6. I have chosen 50 percent garnet and 50 percent clinopyroxene for the residue in this calculation. For both trace-element systems the fractionations increase rapidly as the residual melt fraction drops below about 20 percent (that is, above 80 percent crystallization). Kimberlites may represent such late-stage fluids. They appear to have been in

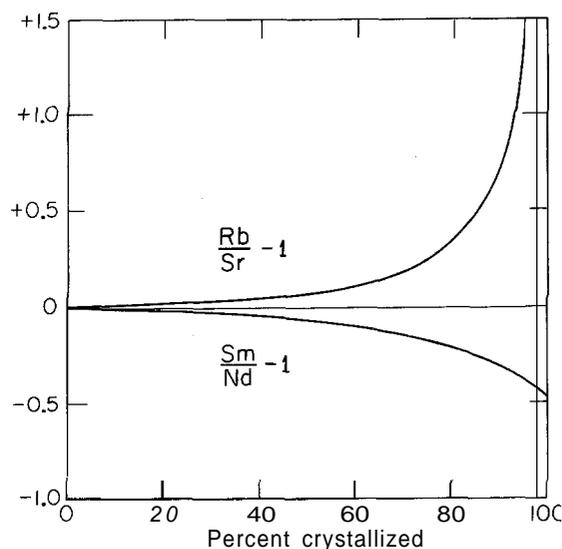


FIGURE 11-6

Variation of the normalized Rb/Sr and Sm/Nd ratios in the melt fraction of a crystallizing eclogite cumulate. Equilibrium crystallization is assumed. The fractionation factor in the melt increases as freezing progresses. If melt extracts from this layer are the enriching fluids for upper-mantle metasomatism, then enrichment will increase with time. Early melt extracts have nearly primitive ratios. The crystalline residue is 50 percent garnet and 50 percent clinopyroxene.

equilibrium with eclogite and often contain eclogite xenoliths. They also have LIL patterns that are complementary to MORB. From Figures 11-6 and 11-7 we see that progressive crystallization of an eclogite cumulate and removal of the melt at various stages is not efficient in fractionating Rb/Sr and Sm/Nd until the late stages. Isotopic evolution in the region enriched by the expelled fluids will start to deviate significantly from the primitive-mantle growth curve only at relatively late times. The more recently enriched reservoirs will have higher Rb/Sr and Nd/Sm ratios than ancient enriched and subsequently isolated reservoirs (Figure 11-7). The partition coefficients for lead are not well known, but the U/Pb ratios appears to behave similarly to the Rb/Sr ratio unless sulfides are involved. That is, the U/Pb ratio in the melt also should increase as crystallization proceeds. For a simple crystallization history of the depleting reservoir, the fractionation factor of the melt increases rapidly with time, for example as an exponential or power law of time (Figure 11-6).

The melt content of a cooling garnet- and clinopyroxene-rich silicate decreases rapidly within a few tens of degrees of the solidus. The average temperature of the mantle has decreased about 150°C in the past 3 billion years (Turcotte, 1980), and the cooling is roughly exponential with a characteristic time of about 2 Ga. Complete crystallization of the upper mantle therefore occurs on a time scale comparable to the age of the Earth, and extreme fractionation,

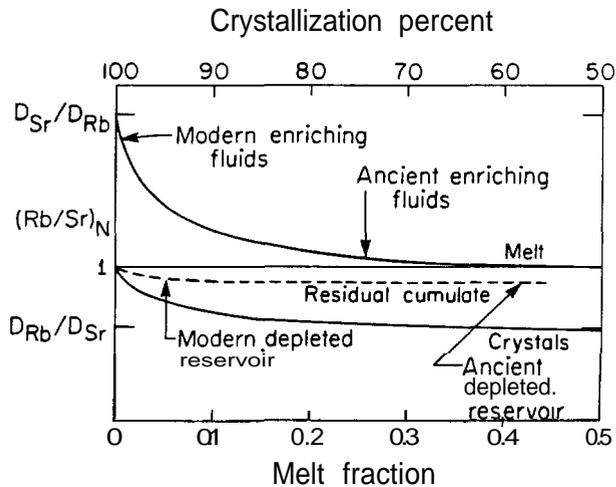


FIGURE 11-7

Composition of the melt phase and the residual crystals as a function of melt fraction for a crystallizing eclogite cumulate. A hypothetical residual cumulate is also shown. The residual cumulate is formed by extracting half of the melt fraction. This shows that the composition of the depleted reservoir is nearly constant with time. Therefore, progressive depletion is less important an effect than progressive enrichment. D_{Sr} and D_{Rb} are partition coefficients.

at least on a global scale, is restricted to the past 1 to 2 Ga. The distribution of oceanic ridges and hotspots is evidence that a large part of the upper mantle is still above or close to the solidus.

The normalized Rb/Sr ratios of a melt and residual crystals can be written

$$\begin{aligned} (\text{Rb/Sr})_m &= \frac{F + (1 - F)D_{Sr}}{F + (1 - F)D_{Rb}} \\ &= (\text{Rb/Sr})_{res} (D_{Sr}/D_{Rb}) = f_m + 1 \end{aligned}$$

where the D are solid-melt partition coefficients and F is the melt fraction. The range of $(\text{Rb/Sr})_m$ is from 1 for large F to D_{Sr}/D_{Rb} as F approaches zero. For an eclogite residue D_{Sr}/D_{Rb} is about 15, so the fractionation factor of the melt f_m , which is taken to be the enriching fluid, varies from zero for large degrees of melting to 14 for the final melt fraction. On the other hand, the total range off for the residual crystals is only from 0 to -0.8 . Considering that the residual cumulate may consist of interstitial melt as well as crystals, the total range is even smaller. As the cumulate freezes, continuously or episodically losing its fluids to the overlying layer it is enriching, it contains less of a more enriched fluid. The net result is a nearly constant f for the cumulate as it evolves. This may explain why depleted magmas require only a simple two-stage model to describe the evolution of their source reservoir and why depleted magmas (MORB) appear to come from a uniform reservoir. Most of the fractionation that a crystallizing reservoir experiences occurs upon the removal of the first batch.

The depleted reservoir becomes depleted by the removal of enriched fluids representing late-stage interstitial fluids or small degrees of partial melting (Figure 11-2). The enriched fluid is not necessarily removed to the continental crust; it can also serve to enrich the uppermost mantle. The enriched and depleted layers may also differ in major elements and mineralogy, possibly the result of crystallization and gravitational separation. This chemical stratification—that is, the difference in intrinsic density—makes it possible for the two source regions to maintain their isotopic identity. If the enriching fluid comes from a garnet-rich layer, it will impart an HREE-depleted pattern to the enriched layer. Since bulk chemistry controls the density and location of the source regions, it is not necessary to assume that one grows at the expense of the other except for a small change in thickness caused by the exchange of a small amount of melt. There also is no need to invoke a large primitive reservoir. Since the enrichment increases with time, there is no inconsistency between a currently highly enriched reservoir and isotopic ratios that depart only slightly from primitive mantle. Isotopic ratios reflect the time-integrated concentrations of the appropriate elements. The mantle need not be primitive even up to the times of the enrichment/depletion events. Partial melting of primitive mantle followed by crystallization and gravity separation gives upper-mantle source regions that, at least initially, have LIL ratios, including Rb/Sr and Sm/Nd ratios, similar to primitive mantle. Residual fluids in a cooling Earth become more fractionated with time.

The lead, neodymium and strontium isotopic data are consistent with progressive enrichment of the reservoir providing the enriched continental and oceanic-island magmas. The time scale of enrichment is of the order of 2 Ga. Such time-dependent enrichment can be accomplished if the enriching fluid is a late-stage residual melt of a deeper, slowly crystallizing cumulate layer. The trace-element pattern of the enriching fluid suggests that it was in equilibrium with garnet.

Partial melting of the mantle during accretion followed by melt separation, crystal fractionation and formation of upper-mantle cumulate layers is one model that can explain the observations. The layers evolve into the major upper-mantle reservoirs. The upward transfer of KREEP-like or kimberlitic material can explain the progressive depletion and enrichment of these reservoirs. Similar scenarios have been developed for the Moon.

Garnet precipitation below 50 km and olivine and orthopyroxene reaction relations with the melt lead to the formation of an eclogite-rich cumulate layer and a residual silica-undersaturated and alumina-poor melt that, upon further cooling, forms a shallower, less garnet-rich, peridotite cumulate layer. The lower mantle is formed of depleted residual crystals and is relatively cold because most of the radioactive elements are in the crust and upper mantle. If the lower mantle is "primitive" or undepleted in radioactive, then temperatures there will be much greater,

and temperatures in the upper mantle will be lower. In the primitive-lower-mantle scenarios of geochemistry, the large amount of heat produced by radioactivity gives a large increase of temperature across the upper mantle–lower mantle boundary and, therefore, a significant thermal boundary layer. In the depleted-lower-mantle scenario the temperature rise is much less.

THE ROLE OF MAGMA MIXING

When two magmas are mixed, the composition of the mix, or hybrid, is

$$xC_1^i + (1 - x)C_2^i = C_i^{\text{mix}}$$

where x is the weight fraction of magma 1, and C_1^i , C_2^i and C_i^{mix} are, respectively, the concentration of the i th element in magma 1, magma 2 and the mix. Mixing relations for elements are therefore linear.

The mixing relations for ratios of elements or isotopic ratios are more complicated. For example,

$$\begin{aligned} (\text{Rb/Sr})_{\text{mix}} &= \frac{x(\text{Rb/Sr})_1 + (1 - x)(\text{Rb/Sr})_2(\text{Sr}_2/\text{Sr}_1)}{x + (1 - x)(\text{Sr}_2/\text{Sr}_1)} \\ ({}^{87}\text{Sr}/{}^{86}\text{Sr})_{\text{mix}} &= \frac{x({}^{87}\text{Sr}/{}^{86}\text{Sr})_1 + (1 - x)({}^{87}\text{Sr}/{}^{86}\text{Sr})_2(\text{Sr}_2/\text{Sr}_1)}{x + (1 - x)(\text{Sr}_2/\text{Sr}_1)} \end{aligned}$$

These are hyperbolas, and the shape or curvature of the hyperbola depends on the enrichment factor E , in this case Sr_2/Sr_1 . Depleted magmas when mixed with an enriched magma can appear to be still depleted for some elemental and isotopic ratios, undepleted or "primitive" for others, and enriched for others, depending on the appropriate E . This simple observation can explain a variety of geochemical paradoxes. For example, many basalts are clearly enriched, relative to primitive mantle, in such ratios as La/Yb and ${}^{206}\text{Pb}/{}^{204}\text{Pb}$ but depleted in ${}^{87}\text{Sr}/{}^{86}\text{Sr}$.

Trace-element and isotopic data for magmas from the two major mantle reservoirs sometimes appear to be inconsistent. The incompatible elements and strontium and neodymium isotopes show that abyssal tholeiites (MORB) are from a reservoir that has current and time-integrated depletions of the elements that are fractionated into a melt. MORBs, however, have ${}^{206}\text{Pb}/{}^{204}\text{Pb}$ and ${}^{207}\text{Pb}/{}^{204}\text{Pb}$ ratios suggesting long-term enrichment in U/Pb as shown in Chapter 10. Alakli basalts and tholeiites from continents and oceanic islands are derived from LIL- and U/Pb -enriched reservoirs. Strontium and neodymium isotopic ratios, however, appear to indicate that some of these basalts are derived from unfractionated reservoirs and others from reservoirs with time-integrated depletions.

These inconsistencies can be reconciled by treating oceanic and continental basalts as mixtures of magmas from

depleted and enriched reservoirs. MORBs are slightly contaminated, depleted magmas, while oceanic-island and continental basalts are mixtures of MORB, or a depleted picritic parent magma, and an enriched end-member having trace-element patterns similar to potassic magmas such as kimberlites or nephelinites. The mixing relations are such that mixtures can be enriched in U/Pb , Rb/Sr , Nd/Sm or ${}^{206}\text{Pb}/{}^{204}\text{Pb}$ relative to primitive mantle, yet appear to have time-integrated depletions in ${}^{143}\text{Nd}/{}^{144}\text{Nd}$ and ${}^{87}\text{Sr}/{}^{86}\text{Sr}$.

A small amount of contamination by material from an enriched reservoir can explain the lead results for MORB (the "lead paradox"; see Chapter 10). Depleted basalts are more sensitive to lead than to rubidium or neodymium contamination. The ${}^{238}\text{U}/{}^{204}\text{Pb}$ of uncontaminated MORB may be about 7 compared to 7.9 for the primary growth curve and above 10 for the enriched reservoirs. Similarly, continental and oceanic-island basalts may represent mixtures of enriched and depleted magmas.

In previous sections I proposed that both of the major mantle magma sources are in the upper mantle and that both are global in extent. The shallower one is inhomogeneous, having been enriched (metasomatized) at various times by fluids from the deeper depleting layer. It may also be the sink of subducted sediments and hydrothermally altered oceanic crust. The deeper source, 400–670 km depth, is more homogeneous and more garnet-rich and provides depleted magmas that, however, become contaminated as they rise through the shallow enriched layer. Partial melting and diapiric ascent originate in the thermal boundary layers between chemically distinct layers. Magma mixing, prior to eruption, is an inevitable consequence of a chemically layered mantle with depleted magmas rising through an enriched uppermost mantle.

Abyssal tholeiites have ${}^{87}\text{Sr}/{}^{86}\text{Sr}$ and ${}^{143}\text{Nd}/{}^{144}\text{Nd}$ ratios that require that the time-averaged Rb/Sr and Sm/Nd ratios of their sources be lower and higher, respectively, than those estimated for the bulk Earth. These results can be explained by a general depletion in large-ion lithophile elements of the MORB reservoir, such as would be caused by extraction of a partial melt or a late-stage residual fluid from a crystallizing cumulate layer.

The lead isotopes for ocean islands can be interpreted in terms of a two-stage evolutionary model with a high ${}^{238}\text{U}/{}^{204}\text{Pb}$ ratio (μ ratio) for the second stage. The radiogenic nature of MORB, which by other measures appears to be depleted, has been attributed to continuous extraction of lead from the reservoir to the core, the lower crust or lower mantle. An alternative is that abyssal tholeiites are slightly contaminated by overlying high- μ material. This would require that enriched mantle underlies both oceans and continents.

Trace-element and isotopic data indicate that many basalts come from a source that is more enriched in LIL than primitive mantle. The presence of a large enriched reservoir in the mantle, perhaps forming contemporaneously with the continental crust, would materially affect models that as-

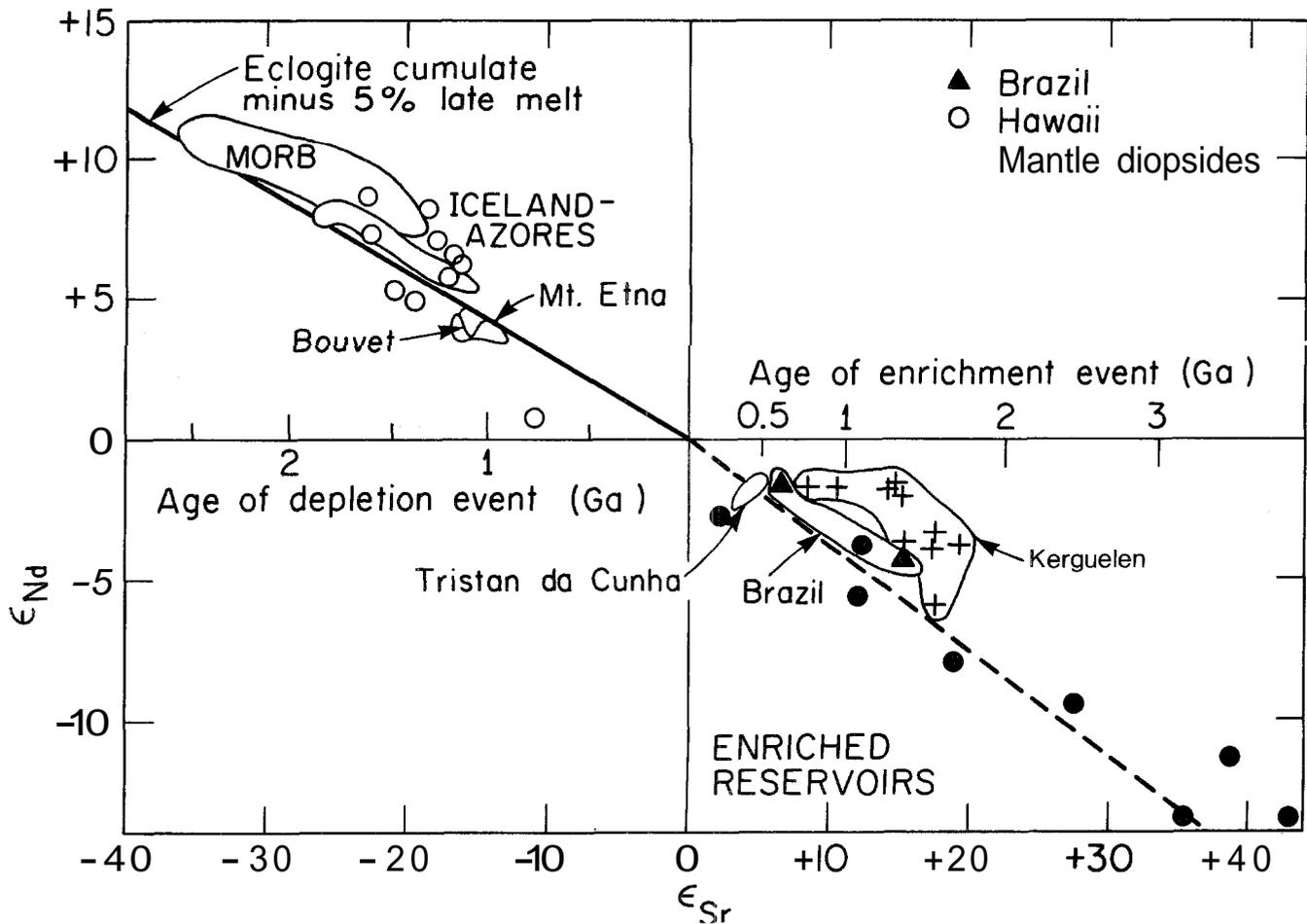


FIGURE 11-8

Neodymium and strontium fractionation trends for an eclogite cumulate (solid line) that has been depleted at various times by removal of a melt fraction. ϵ_{Nd} and ϵ_{Sr} are the $^{143}Nd/^{144}Nd$ and $^{87}Sr/^{86}Sr$ ratios expressed as fractional deviations in parts in 10^4 from those in a primitive undifferentiated reference reservoir. The dashed line is the complementary reservoir that has been enriched at various times by the melt extract from the eclogite layer. Reservoirs are unfractionated until the enrichment/depletion event and uniform thereafter. If present values of enrichment/depletion have been reached gradually over time or if these magmas are mixtures, then ages shown are lower bounds. "Depleted" OIB (upper left quadrant) may represent mixtures of depleted and enriched magmas.

sume that the continental crust is the only complement to the depleted reservoir. The presence, size, location and time history of enrichment of this reservoir are critical issues in discussions of mantle evolution. If this reservoir is shallow and global, it may be expected that MORB and partial melts from the enriched layer may be common.

Figure 11-8 shows the correlation in differentiation for MORB, oceanic islands, and some continental basalts and mantle diopsides. The depletion and enrichment ages are calculated for a simple two-stage model for the development of the enriched and depleted reservoirs. The Rb/Sr and Sm/Nd ratios are assumed to be unfractionated up to the age shown and then fractionated to values appropriate

for the depleted and enriched reservoirs. Subsequent isotopic evolution occurs in these fractionated reservoirs. For this kind of model the MORB reservoirs were apparently depleted and isolated at times ranging from 1.5 to 2.5 Ga, and the enriched reservoirs (giving magmas in the lower right quadrant) were enriched between 0.5 and 1.8 Ga. If the enrichment has been progressive, that is, Rb/Sr and Nd/Sm increasing with time, the start of enrichment could have been much earlier. Progressive enrichment, of the type often discussed for the uranium-lead system, could explain the low $\epsilon_{Nd}-\epsilon_{Sr}$ values found for continental and oceanic-island basalts with high Rb/Sr and Nd/Sm ratios.

The data shown in Figure 11-8 may also be interpreted

in terms of mixtures of magmas from depleted and enriched reservoirs. In fact, the compositions of alkali olivine basalts, basanites and continental tholeiites are bracketed by MORB and potassium-rich magmas such as nephelinites for most of the major and minor elements as well as for the isotopes. This supports the possibility that many continental and oceanic-island basalt types are mixtures.

Contamination of MORB seems to be a plausible way of reconciling the lead isotopic results with those from the other isotopic systems. Mixing alone, however, cannot explain the large spread of lead and U/Pb in MORB that accompanies the relatively restricted range in $^{206}\text{Pb}/^{204}\text{Pb}$. Crystal fractionation and hydrothermal alteration operating at the ridge-crest environment may also be involved.

It appears that, in general terms, the isotopic and trace-element results for continental and oceanic-island basalts can be understood in terms of magma mixing. A conceptual model is as follows. Consider a stratified mantle with a shallow enriched layer and a deeper (>200 km) depleted layer. Assume that partial melting in the thermal boundary layer of the depleted layer initiates diapiric ascent. The upward advection of hot isotherms initiates melting in the shallower layer. The initially erupted magmas will be the most enriched, and subsequent magmas will have more of the depleted component. The final eruptive product is MORB, which as it wanes is increasingly contaminated by its passage through and evolution in the enriched upper mantle. In this model, magma mixing is unavoidable and there is a specific temporal sequence of eruption; enriched magmas occur in the initial and waning stages. Enriched magmas would also be expected on the flanks of the surface expression of the diapir. Active seamounts and islands away from oceanic ridges would be expected to be the most enriched unless the depleted oceanic lithosphere is remelted and contributes to the erupted magmas. This model is illustrated in Figure 11-9.

The mixing parameters of Table 11-4 (see also Figure 10-4) indicate how several apparent geochemical contradictions can be resolved. If we adopt the point of view that all oceanic-island basalts are contaminated MORBs, then mixtures involving more than about 60 to 75 percent MORB will appear to represent magmas having time-integrated depleted ratios of Nd/Sm and Rb/Sr, respectively, that is, depleted ϵ_{Nd} and ϵ_{Sr} . On the other hand, the current Sm/Nd and Rb/Sr ratios of the mixture will appear enriched until the mixture is more than 78 to 94 percent of the depleted end-member. Therefore, contaminated MORB can exhibit LREE and Rb/Sr enrichment but have apparent time-integrated depletions, as is the case for most oceanic-island and continental flood basalts. With the parameters given, only magmas with more than about 94 percent MORB will appear depleted by all the above measures, and even purer MORB is required to exhibit $^{206}\text{Pb}/^{204}\text{Pb}$ depletion. On the other hand, magmas must contain less than about 60 percent MORB component in order to appear enriched by all mea-

asures. Magmas containing 20 to 60 percent MORB component have major- and trace-element compositions characteristic of the basanite-alkali olivine basalt suite.

The isotopic ratios of end-member MORB are greatly affected by small degrees of contamination, contamination that is probably unavoidable if MORB rises through, or evolves in, continental lithosphere or enriched upper mantle. Basalts at anomalous ridge segments show clear signs of contamination, as in T- and P-MORB (transitional- and plume-type MORB), and normal MORB may simply show less obvious signs of contamination. Since relatively depleted magmas are so sensitive to contamination, it seems unlikely that they can be used to define secondary isochrons. On the other hand, enriched magmas such as oceanic-island basalts are less sensitive to "contamination." The fact that MORB contains high $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{206}\text{Pb}/^{204}\text{Pb}$ ratios, in spite of being a "depleted" magma, is called the "lead paradox" (see Chapter 10 and Figure 10-4 in particular).

The dichotomy between high Nd/Sm in some alkali basalts coupled with depleted $^{143}\text{Nd}/^{144}\text{Nd}$ ratios ($\epsilon_{\text{Nd}} > 0$) requires that the Nd/Sm enrichment be recent. This implies either recent metasomatism of the depleted source region or recent magma mixing. If deep diapirs from the depleted reservoir cause the melting in the enriched layer, then magma mixing would be an expected prelude to eruption, and enriched magmas exhibiting isotopic evidence for long-term depletion could be easily understood. Magmas containing 10–20 percent contaminant will still appear isotopically depleted for Nd and Sr. Using values from Table 11-4, the Rb/Sr ratio of such mixtures is 0.037 to 0.052, enriched relative to primitive mantle and similar to ratios in oceanic-island and continental basalts. The Sm/Nd ratio is 0.30 to 0.34, also enriched compared to MORB and similar to primitive ratios. Such mixtures will appear to exhibit long-term enrichment in the lead isotopic systems.

MELTS FROM THE ENRICHED RESERVOIRS

So far I have considered a single enriched end-member that may represent a small degree of partial melting from an enriched refractory peridotitic layer in the upper mantle. I have speculated that melting in this layer is initiated by the rise of a diapir from a deeper depleted layer. Even if the enriched layer is homogeneous, the partial melts from this region will have variable LIL contents and ratios such as Rb/Sr, Sm/Nd, and U/Pb, which depend on the extent of partial melting. Magma mixtures, therefore, may appear to require a range of enriched end-members. A plot of an isotopic ratio versus a ratio such as Rb/Sr, Sm/Nd or La/Sm may exhibit considerable scatter about a two-component mixing line even if the end-members are isotopically homogeneous. An example is shown in Figure 11-10. The en-

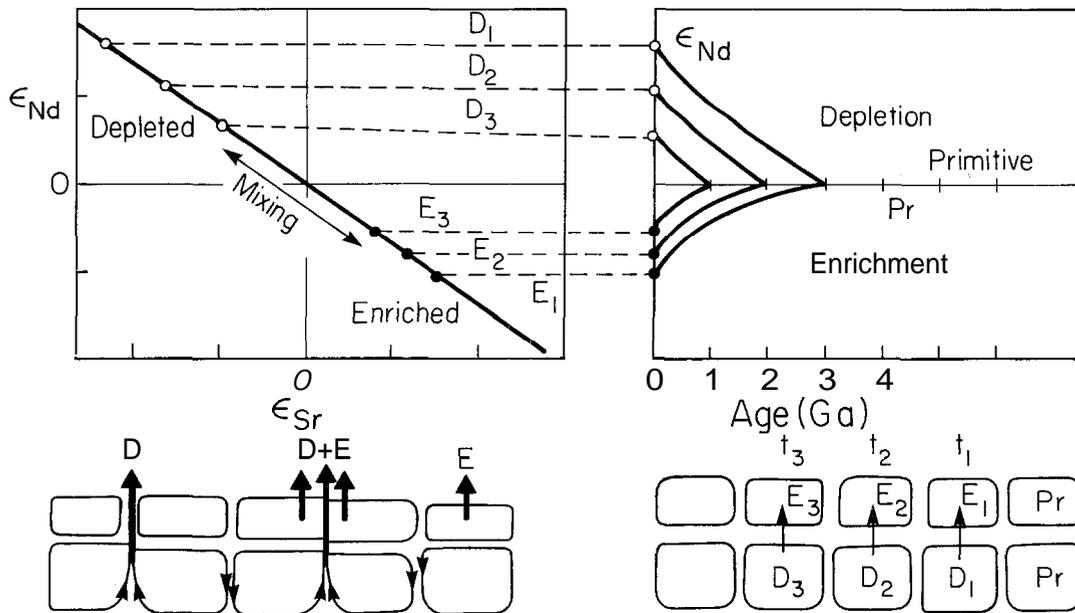


FIGURE 11-9

Illustration of isotopic growth in a two-layer mantle. The lower layer is formed of heavy cumulates, perhaps at the base of a magma ocean. As it freezes it expels enriched fluids to the shallow layer (E), thereby becoming depleted (D). As time goes on, and crystallization proceeds, the melts become more enriched and more fractionated. Isotopic growth is more rapid for the parts of the shallow mantle enriched at later times, but the earlier enriched reservoirs (E₁) have had more time for isotopic growth. The mantle array can be interpreted as the locus of points representing magmas from different aged reservoirs or as a mixing array between products of enriched and depleted reservoirs, or some combination. Melts from D may be contaminated by E if they cannot proceed directly to the surface. If E is a trace-element-enriched, but infertile, peridotite, then D may be the main basalt source region, and enriched basalts may simply represent contaminated MORB. The enriched component in E may be kimberlitic. The deeper layer transfers its LIL upward, forming a depleted layer (D) and a complementary enriched layer (E). The growth of ϵ_{Nd} in the depleted and enriched cells (upper right) combined with similar diagrams for ϵ_{Sr} generates the mantle array (upper left). Magma mixing reduces the spread of ϵ -values, decreasing the apparent ages of the depletion/enrichment events. Layer D may be the transition region.

riched source has ϵ_{Nd} and Rb/Sr as shown on the lower dashed line. The Rb/Sr of partial melts from this reservoir are also shown. The solid lines are mixing lines between these melts and a melt from the depleted reservoir having properties estimated for "pure MORB." The dashed lines are labeled by the fraction of MORB in the mixture. The data points are representative compositions of various basalts; most fall in the field representing 50 to 95 percent "MORB" and an enriched component representing 2 to 20 percent melt from the enriched reservoir. The Rb/Sr ratio may also be affected by crystal fractionation and true heterogeneity of one or both of the two source regions.

The hypothesis that oceanic and continental magmas represent mixtures of melts from ancient enriched and depleted reservoirs in the mantle appears capable of explaining a variety of geochemical data and resolving some isotopic paradoxes. The hypothesis explains the lead paradox

and other apparently contradictory trace-element and isotopic evidence for enrichment and depletion of the various mantle reservoirs. LIL-enrichment associated with time-integrated depletion, usually attributed to mantle metasomatism, is a simple consequence of mixing relations. Most oceanic-island and continental flood basalts may be "contaminated" MORB. If these basalts are mixtures, then the ϵ_{Nd} ages of their parent reservoirs will be underestimated, that is, the sources are more enriched or depleted in their isotopic ratios than the hybrid magmas. This may reconcile the geological evidence for ancient and rapid continental growth with the apparently conflicting isotopic evidence for the history of mantle reservoirs. The close proximity in time and space of enriched and depleted magmas in all tectonic environments—continental rifts, oceanic islands, fracture zones, midoceanic rifts and arcs—supports the concept of a stratified mantle. In a chemically stratified mantle the in-

TABLE 11-4
Mixing Model for Basalts

Parameter	MORB	Contaminant*	
Pb	0.08 ppm	2 ppm	
U	0.0085 ppm	0.9 ppm	
U/Pb	0.10	0.45	
²³⁸ U/ ²⁰⁴ Pb	7.0	30.0	
²⁰⁶ Pb/ ²⁰⁴ Pb	17.2	19.3, 21	
Rb	0.15 ppm	28 ppm	
Sr	50 ppm	350 ppm	
⁸⁷ Sr/ ⁸⁶ Sr	0.7020	0.7060	
Sm	2 ppm	7.2 ppm	
Nd	5 ppm	30 ppm	
¹⁴³ Nd/ ¹⁴⁴ Nd	0.5134	0.5124	
Enrichment Factors[†]			
Pb	25.0	U/Pb	4.5
Sr	7.0	Rb/Sr	26.7
Nd	6.0	Sm/Nd	0.60

*Assumed composition of contaminant. This is usually near the extreme end of the range of oceanic-island basalts.

[†]Ratio of concentration in two end-members.

dividual reservoirs can maintain their isotopic identity. In a layered, convecting mantle the average temperature gradient is adiabatic except in the thermal boundary layer between reservoirs. Melting is most likely to initiate in these regions of high temperature gradient, permitting the ascent and mixing of magmas from previously isolated reservoirs. Partial melting in the deeper layer, caused by deformation of the boundary or continental insulation, may be the trigger for diapiric ascent. If the solidus temperatures in the two reservoirs are similar, the geotherm will cross the solidus first in the lower part of the thermal boundary layer, that is, in the deeper layer (see Figure 9-5).

The geochemical homogeneity of MORB is one reason why it is desirable to place it deep, so it can be homogenized by convection and isolated from subducted sediments and altered oceanic crust. The heterogeneity of OIB and CFB and the requirement of several enriched end-members to satisfy all the various kinds of data is one reason why this reservoir appears to be shallow. The enrichment and heterogeneity of the OIB reservoir is also intermediate between the crust and the MORB reservoir.

MAGMA OCEAN

The separation of the mantle into distinct geochemical reservoirs probably occurred during accretion and possibly via a magma ocean. A magma is a molten or extensively molten rock that may contain suspended crystals. It behaves as a high-viscosity fluid. An ocean is a liquid that covers extensive areas of a planet. When the surface is below the freezing point, the ocean may be covered by a solid cap. As an

ocean freezes the components separate according to their densities and crystallization temperatures. The stability of a freezing ocean depends on the variation of temperature and composition with depth. Thus, a layer of molten rock containing suspended, settling or rising crystals and overlain by a cap of the lighter crystals can be called a magma ocean. The freezing of a magma ocean results in a layered stack of crystals, sorted according to their crystallization temperatures and densities.

A magma ocean is transient because the surface or atmospheric temperatures are undoubtedly less than the interior temperature of a planet. On the other hand, during the accretional stage of planetary evolution, energy can be delivered to the surface and interior much faster than it can be removed by convection and conduction. The extreme case is a large impact that can melt a large fraction of the mantle as well as eject material from the Earth. It is likely that the Earth experienced many giant impacts, and therefore the mantle was partially molten many times and at all stages in its evolution.

Since the settling of crystals in a freezing magma ocean is likely to be much faster than the freezing of the ocean itself, the result is a chemically stratified planet, during and after accretion. The high melting points of olivine and orthopyroxene, the buffering provided by the latent heats of melting, the high density of these refractory crystals relative to melts in the outer part of a planet, and the increase of melting temperature with pressure mean that low-pressure melting and melt-solid separation will dominate and that separation of basaltic material from peridotitic material is likely to be an efficient process. The effect of pressure on crystallization temperatures means that freezing in a adiabatic magma ocean will proceed principally from

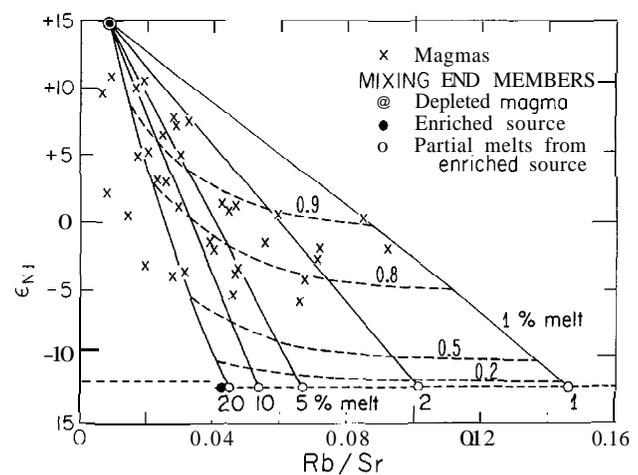


FIGURE 11-10

Mixing relations for a depleted magma and partial melts from an enriched peridotite reservoir. The solid lines are mixing lines, and dashed lines give the fraction of the depleted component.

the bottom, forcing residual melts and their incompatible elements toward the surface. Thus, the minor elements as well as the major elements are redistributed during accretion and the subsequent freezing of the magma ocean. It is difficult to imagine a scenario that will avoid this chemical stratification or allow a large portion of the mantle to retain cosmic abundances of the elements.

The transient nature of the ocean and the possibility that it may be largely buried or discontinuous should not divert attention away from the extensive and widespread melting that must occur in the accretion of a terrestrial planet.

The extreme concentration of trace elements in both the lunar and terrestrial crusts and the thick anorthositic crust on the Moon provide strong geochemical and petrological support for the concept of a magma ocean. The absence of a correspondingly thick crust on the Earth is simply a reflection of the higher pressures on the Earth. Plagioclase, the chief component of anorthosite, floats on magma but it crystallizes only at low pressure.

At low pressure there is a very large melting interval, more than 500°C, between the solidus and the liquidus of four-phase rocks consisting of olivine, orthopyroxene, clinopyroxene and an aluminous phase such as plagioclase, spinel or garnet. Olivine is the most refractory component and the calcic, aluminous and alkali-bearing phases are the most fusible. The large melting interval plus the large difference in density between melts and the near-liquidus phases, olivine and orthopyroxene, means that the basaltic and refractory components will have time to separate during accretion and the subsequent crystallization phase of Earth evolution. If melting ever extends deeper than about 200 km, melts may become denser than the refractory crystals and settle downward. Orthopyroxene transforms to denser phases, spinel plus stishovite or majorite, depending on the Al_2O_3 content, below about 300 km. This plus the olivine- β -spinel transformation at 400 km restricts a deep melt zone to between 200 and 300 or 400 km. This deep melt zone crystallizes much more slowly than near-surface melts. Garnet may settle as crystallization proceeds, because of its high density, but garnet and clinopyroxene coprecipitate over a narrow temperature interval and are more likely to stay together than melt and olivine-orthopyroxene, which differ substantially both in density and crystallization temperature.

The high density of garnet and majorite relative to mafic and ultramafic melts and olivine means that the transition region and lower mantle will be enriched in Si and have a higher Si/Mg ratio than the shallow mantle. If the pressure at the base of the magma ocean was greater than 100 kbar, the molten zone would act as a gravitational filter with orthopyroxene, as majorite, sinking to the base and olivine rising. As the planet grows, the majorite transforms to perovskite, thus explaining the seismic evidence for a

perovskite-rich lower mantle. If the magma ocean is deeper than about 400 km, any residual olivine that has not already floated to shallow depths will transform to β - and γ -spinel, phases that are probably denser than the melt at these depths. With this scenario most of the material in the lower mantle is the refractory residue remaining after melt-crystal separation and is therefore orthopyroxene-rich and depleted in the incompatible elements. Although garnet-majorite and perovskite can probably accommodate a variety of trace elements, it is the partitioning of elements between the phases and melts at lower pressures that sets the trace-element contents of the phases that presently exist in the high-pressure forms. It is likely therefore that the transition region and lower mantle are depleted in the incompatible elements.

The elastic properties of the lower mantle depend mainly on the Mg/Si ratio. For iron-free systems this ranges from 2 for olivine stoichiometry to 1 for orthopyroxene and garnet stoichiometries and 112 for diopside. For a chondritic mantle the ratio is slightly greater than 1. This reflects both the low olivine content, relative to upper-mantle peridotites, and the high FeO content. The high Mg/Si ratio of the upper mantle is partially due to the removal of basalt ($\text{CaMgSi}_2\text{O}_6$ plus $\text{Mg}_3\text{Al}_2\text{Si}_3\text{O}_{12}$). This also reduces the CaO and Al_2O_3 content of peridotite that has experienced basalt extraction. The amount of basalt at the Earth's surface is not enough to account for the difference in Mg/Si between peridotite and cosmic abundances, and the amount of CaO and Al_2O_3 in chondrites is not adequate to balance the Mg/Si if observed peridotites are the complement to basalt. The conclusion is that there is abundant basalt and orthopyroxene buried at depth in the mantle. One way to separate orthopyroxene from olivine is at the base of a deep (>300 km) magma ocean, where majorite, the liquidus phase, is denser than olivine.

A globe-encircling magma ocean has interesting implications regarding the early evolution of the mantle. Such an ocean probably existed throughout most of the accretional phase of planetary formation and may even have been regenerated by large late impacts. The rate of freezing of such an ocean depends on the temperature of the early atmosphere, which may have been quite dense and hot, and the rate at which heat is conducted through the base of the ocean. The temperatures in the ocean will be nearly adiabatic. The melting curve is much steeper. The ocean will therefore experience pressure freezing at its base and will crystallize upwards. The earliest crystals are olivine, orthopyroxene-majorite or orthopyroxene-perovskite, depending on depth. The composition of the residual fluid therefore also changes with depth. Deep liquids below 400 km will be MgO-rich, while shallower liquids will be SiO_2 -rich. In general, these crystals are denser than the residual melts except for a possible density crossover between olivine and melt at depths between some 200 and 300 km. The

liquids will also be enriched in CaO, Al₂O₃ and the incompatible elements. The processing of infalling material by a magma ocean is an efficient mechanism for separation of elements on the basis of the melting temperatures and densities of their host materials. A growing Earth with a deep magma ocean will result in an olivine-rich upper mantle and a depleted, infertile, orthopyroxene-rich lower mantle. Continued freezing will deposit clinopyroxene and garnet at depth and feldspar in the upper layers. The lighter SiO₂-rich and alkali-rich melts may even form a separate liquid layer at the surface, which eventually freezes to form the proto-crust. Such fluids would form by the crystallization of an olivine cumulate layer and, if lighter than the bulk of the magma ocean, will rise to the surface.

The end result is a chemically stratified mantle with the incompatible elements strongly enriched toward the surface and the olivine-rich melts and crystals concentrated in the upper mantle. The deeper mantle would be enriched in orthopyroxene.

The chemical stratification of the magma ocean and the sorting of crystals and melts results in a stratified mantle that cools much more slowly than a homogeneous fluid. Deeply buried melt and partial melt layers will cool and crystallize extremely slowly. Convection cools a planet rapidly only if all the material in the interior can be brought to the near surface to lose its heat or if surface cooled material is circulated back into the interior. This can only happen in a homogeneous material or in a stratified system if the density contrasts between strata can be overcome by temperature or phase changes. Otherwise, cooling at depth is slow, even in convecting regions, because the convected heat must be removed by conduction through each thermal boundary layer. There is the interesting possibility that truly primitive melts may still exist somewhere in the deep interior.

SUMMARY OF CONSTRAINTS ON MANTLE EVOLUTION

Many models of petrogenesis assume that large parts of the mantle have never experienced differentiation. A primitive, or undepleted or pyrolite lower mantle is the cornerstone of many models. The alternative point of view I have developed in this chapter is that the Earth, including the mantle, is extensively differentiated, the result primarily of high temperatures early in Earth history. The following observations are particularly relevant and have been adopted as constraints on mantle evolution models:

1. A large part of the upper mantle is close to or above the solidus as evidenced by the distribution of oceanic ridges, hotspots, and island arcs and seismic velocities.

2. The upper mantle has cooled appreciably in the past 4.5 Ga, indicating that extensive partial melting was even more important in early Earth history.
3. The most voluminous magma type is tholeiite, which probably represents large (>15–20 percent) degrees of partial melting; picrites and komatiites represent even larger degrees of melting.
4. Although enriched magmas such as nephelinites and alkali-olivine basalts are considerably less abundant than oceanic and continental tholeiites, they often occur in close proximity to tholeiites in continental interiors, oceanic islands, and along ridge crests. Magmas exhibiting extreme LIL-enrichment, implying small degrees of partial melting, are rare now and may have been rarer in the Precambrian.
5. Inhomogeneities in the mantle having density contrasts of greater than about 1 percent and dimensions greater than about 10 km have a lifetime much less than the convective overturn of the mantle or the age of the mantle reservoirs.
6. Bodies much smaller than the Earth have apparently experienced extensive partial melting and differentiation early in their history.
7. Since the melting gradient increases faster with depth than the adiabatic gradient, melting in the mantle must begin in thermal boundary layers, found near chemical interfaces in the mantle.
8. The intrinsic density differences between the various products of mantle differentiation, such as basalts, depleted peridotite, and eclogites, are much greater than density variations associated with thermal expansion for reasonable variations in temperature.
9. Komatiites imply that high temperatures existed in the Archean mantle and that deep, or extensive, melting was common.
10. The isotopic compositions of MORB and the atmosphere indicate that the Earth was efficiently (but not 100 percent) outgassed early in its history and that subsequent outgassing (such as ⁴⁰Ar) has also been relatively efficient.

A mantle evolution model involving early and extensive partial melting, cooling, crystal fractionation and gravitational stratification seems to have the potential of satisfying a variety of geochemical data. An eclogite-rich cumulate layer at depth is the terrestrial equivalent of the lunar anorthosite layer. High-temperature magmas and magmas representing relatively large degrees of partial melting are currently being delivered to the Earth's surface. Convection calculations indicate that the mantle has cooled appreciably in the past 3 Ga. Extensive melting and crystal fractionation were therefore probably important processes

in early Earth history. Isotopic data show that differentiation, or separation, rather than homogenization has been the rule throughout Earth history.

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