Chapter 25

Crystallization of the mantle

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

Franz Y. Loewinson-Lessing

The Earth is cooling and crystallizing. The mantle has evolved considerably from the magma ocean era to the plate tectonic era. Part of the evolution is due to igneous processes, and part is due to plate tectonic processes. 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 homogenous, even primitive, mantle. We could design simple one- and two-layer mantle models, such as were popular in the last century. It now seems unlikely that we will find the ‘Rosetta Stone,’ a rock or a meteorite fragment that represents ‘original Earth’ or even the parent or grandparent of other rocks. Rocks and magmas represent products of complex multistage processes, and they are mixtures of components with various melting histories. 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. Melts and rocks are averages of various components and processes. Sampling theory and the manipulation of averages are involved in this modeling. In addition to recycling, and an intrinsically heterogeneous mantle, there are igneous processes that cause chemical heterogeneity interior to the mantle.

Melt trapping and melt transport can create distinctive chemical and isotopic components. Chemical heterogeneity can be formed internal to the mantle.

Components vs. reservoirs revisited

A variety of studies have lent support to the concept of a chemically inhomogenous mantle. The mantle contains a variety of components that differ in major elements, intrinsic density, melting point, large-ion-lithophile (LIL)-contents and isotope ratios; they have maintained their separate identities for at least $10^9$ years. These components have been termed reservoirs but only very loose bounds can be placed on the sizes of these so-called reservoirs. They could be grain-size, or slab-size, depending on the ability of chemical species to migrate from one to another. A component can have dimensions of tens of kilometers, a typical scale for subducted or delaminated assemblages. The melting process gathers together these components, and partial melts therefrom. Some components or reservoirs have high values of Rb/Sr, Nd/Sm, U/Pb and $^3$He/U. These components are enriched, but for historic reasons they are sometimes called primitive, undegassed or more-primitive (than MORB).

Lithology of the upper mantle

Since the seismic velocities and anisotropy of the shallow mantle are consistent with an olivine-rich aggregate, and since most mantle xenoliths are olivine-rich, it has been natural to assume
that the shallow mantle, and the source region for various basalts, including OIB and NMORB, are the same and is a peridotite. Since the LIL-depleted components in the mantle have already lost a melt fraction, they should be depleted in garnet – and therefore infertile – unless they are eclogite cumulates, or delaminated lower continental crust. Peridotites depleted in basalt have less $\text{Al}_2\text{O}_3$ and garnet than nondepleted or fertile peridotites. But the depletion event may have involved a very small melt fraction, in which case the incompatible trace elements will be affected more than the major elements.

The traditional emphasis on homogenous olivine-rich and peridotite source regions for mantle magmas is based on the following arguments.

1. Peridotite is consistent with seismic velocities for the shallow mantle; basalts come from the mantle; the upper mantle is therefore mainly peridotite.
2. Garnet peridotite is stable in the upper mantle.
3. Garnet peridotites have close compositional relationships to meteorites.
4. Partial melts of natural samples of garnet peridotite at high pressure have basaltic compositions.
5. Eclogites are the high-pressure chemical equivalent of basalt and partial melting of eclogite does not recreate the composition of basalt.
6. Melting of eclogites would have to be very extensive, and melt–crystal segregation would occur before such extensive melting can be achieved.

These arguments are all suggestive rather than definitive. They do not rule out other lithologies for the upper mantle or for the source regions of at least some basalts. There is increasing evidence that large parts of the upper mantle are eclogitic or composed of garnet pyroxenite and are therefore more fertile than most peridotites. Most of the seismic information about the upper mantle is derived from seismic waves with wavelengths from 20 to 300 km. Much of the petrological information comes from midocean ridges and large volcanoes, which sample comparable size regions of the upper mantle. This averaging effect of geophysical and geochemical data can distort views regarding homogeneity of the mantle. But, there is no doubt that peridotites can and do come from the shallow mantle; some regions of the upper mantle and lithosphere are probably mostly peridotite. Even if basalts derive from eclogitic regions of the mantle they traverse and evolve in peridotitic surroundings.

The average composition of the Earth is probably close to ordinary chondrites or enstatite meteorites in major-element chemistry, the mantle therefore contains abundant, although not necessarily predominant, olivine. By the same reasoning the mantle contains even more pyroxene plus garnet. The above arguments do not prove that the source region of the most abundant basalt types, 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 source regions, such as a glassy or basaltic shallow source, can be ruled out, the possibility that basalts involve eclogite, pyroxenite, recycled crust or cumulates, cannot be ruled out. Eclogite, garnet pyroxenite, peridotite–eclogite mixtures, or piclogite, are also candidate source ‘rocks’; the ‘grains’ in such ‘rocks’ can be tens of km in extent. Hand-specimen-sized rocks are a different scale from what volcanoes and seismic waves see.

The trace-element inhomogeneity of the mantle plus the long-term isolation of the various components suggests that differentiation has been more effective in the long run than mixing. Mixing can be avoided in a chemically inhomogeneous – or chemically stratified – mantle if the components are large chunks and/or have large intrinsic density and viscosity contrasts. Garnet has the highest density of any abundant upper-mantle mineral and therefore plays a role in determining the density of various components, and regions, of the mantle. However, the chemical heterogeneity may also be dispersed throughout the upper mantle. Eclogites come in a variety of compositions and densities; they all have low melting points compared to peridotites. Some eclogites have densities similar to some upper mantle peridotites; the density of eclogite is very
Table 25.1  Effect of eclogite and olivine fractionation on primitive magma

<table>
<thead>
<tr>
<th>Magma</th>
<th>SiO₂</th>
<th>Al₂O₃</th>
<th>FeO</th>
<th>MgO</th>
<th>CaO</th>
<th>TiO₂</th>
<th>Na₂O</th>
<th>K₂O</th>
</tr>
</thead>
<tbody>
<tr>
<td>1. Primitive</td>
<td>46.2</td>
<td>11.1</td>
<td>10.8</td>
<td>20.2</td>
<td>9.4</td>
<td>0.77</td>
<td>1.06</td>
<td>0.08</td>
</tr>
<tr>
<td>2. Extract</td>
<td>46.2</td>
<td>13.9</td>
<td>9.3</td>
<td>16.3</td>
<td>11.9</td>
<td>0.81</td>
<td>1.29</td>
<td>0.02</td>
</tr>
<tr>
<td>3. Picrite</td>
<td>46.2</td>
<td>8.3</td>
<td>12.3</td>
<td>24.1</td>
<td>6.9</td>
<td>0.74</td>
<td>0.83</td>
<td>0.14</td>
</tr>
<tr>
<td>Tholeiites</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>4. Model</td>
<td>50.0</td>
<td>13.8</td>
<td>12.4</td>
<td>8.5</td>
<td>11.5</td>
<td>1.23</td>
<td>1.38</td>
<td>0.23</td>
</tr>
<tr>
<td>5. Hawaiian</td>
<td>50.0</td>
<td>14.1</td>
<td>11.4</td>
<td>8.6</td>
<td>10.4</td>
<td>2.53</td>
<td>2.16</td>
<td>0.39</td>
</tr>
<tr>
<td>6. Continental</td>
<td>50.6</td>
<td>13.6</td>
<td>10.0</td>
<td>8.5</td>
<td>10.0</td>
<td>1.95</td>
<td>2.90</td>
<td>0.54</td>
</tr>
<tr>
<td>7. Average oceanic</td>
<td>50.7</td>
<td>15.6</td>
<td>9.9</td>
<td>7.7</td>
<td>11.4</td>
<td>1.49</td>
<td>2.66</td>
<td>0.17</td>
</tr>
</tbody>
</table>

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 (Fo₈₇₅) from liquid (3).
5. Average Hawaiian parental tholeiite.
7. Average oceanic tholeiite glass (Elthon, 1979).

Dependent on composition and temperature. If the mantle is chemically stratified, mixing will be less vigorous and chemically distinct components can survive.

Part of this gravitational stratification will be irreversible. The coefficient of thermal expansion is high at low-pressure and high-temperature. This means that temperature can overcome intrinsic density differences. However, at high pressure, this is no longer possible and deep dense layers may be trapped. At lower mantle conditions, a chemically distinct layer with an intrinsic density contrast of ~1% can be stable against convective over-turn and mixing. Crystallization of a melt layer or magma ocean leads to a series of cumulate layers, and fractionation of the LIL.

Cumulate layers originally contain interstitial fluids that hold most of the incompatible elements. As crystallization proceeds, these melts may migrate upward. Melts from an eclogite or olivine eclogite cumulate have 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 apparently 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.

Access to deep layers

Convection in a chemically stratified system causes lateral variations in temperature, and deformation of the interfaces because of the buoyancy of the upwelling 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 buoyant diapirs, even in a dense eclogite-rich layer. Subsolidus 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 an eclogite blob, or a diapir from a buried eclogite or piclogite layer, can lead to extensive melting. Crystal settling and melt extraction can be avoided in a rapidly rising diapir because of the high temperatures, temperature gradients and stresses, and the surrounding envelope of subsolidus peridotite.

**Formation of fertile regions in the mantle**

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; refractory, 'purified' material is fed into the planet. The incompatible elements and melts are preferentially retained near the surface. A deep magma ocean or whole planet melting – 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 high-pressure phases such as perovskite, there should be anomalies in the trace-element patterns of upper mantle materials that are not observed.

Tables 25.1 and 25.2 and Figure 25.1 illustrate a petrological evolutionary scheme for the mantle. Early melting is likely to be extensive since large amounts of melt, 15-25%, occur in small temperature range just above the solidus. Eclogite and basalt extraction leave a peridotitic residue deficient in the basaltic elements, Ca, Al and Na. Olivine cumulates and enriched residual fluids are complements to the eclogite and basalt regions. Table 25.2 gives a more detailed comparison of the possible products of mantle differentiation.

These tables illustrate the plausibility of eclogite-rich regions in the mantle. An eclogite or basalt layer, or distributed blobs, representing about 10-15% of the mantle reconciles the major-element compositions inferred by cosmochemical, geophysical and petrological techniques, and can account for such elements as Nb, Ti and Zr.

![Figure 25.1](image.png)

**Fig. 25.1** 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 unfraccionated. 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. The large difference in crystallization temperature and density of olivine-orthopyroxene (ol-opx), garnet, plagioclase and so on, means that mineralogically distinct regions can form in early Earth history.

The source regions for some basalts may be eclogite-rich cumulates or blobs that have been depleted by removal of a Kimberlite-like fluid. Eclogitic layers or blobs become unstable at depth as they warm up. Garnet-clinopyroxene reactions and partial melting contribute to the buoyancy.

**Early chemical stratification of the mantle**

Chemical stratification resulting from early differentiation of the mantle, upward removal of the melt and fractionation via crystal settling during cooling is one way to explain chemically distinct reservoirs. 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 shallow
Table 25.2 | Composition of mantle, upper mantle, picrites and eclogites

<table>
<thead>
<tr>
<th>Material</th>
<th>SiO₂</th>
<th>Al₂O₃</th>
<th>FeO</th>
<th>MgO</th>
<th>CaO</th>
<th>TiO₂</th>
<th>Na₂O</th>
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<tr>
<td>Mantle and Upper Mantle Compositions</td>
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<td></td>
<td></td>
<td></td>
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<td></td>
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<tr>
<td>1. Bulk mantle</td>
<td>48.0</td>
<td>5.2</td>
<td>7.9</td>
<td>34.3</td>
<td>4.2</td>
<td>0.27</td>
<td>0.33</td>
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<tr>
<td>2. Residual mantle</td>
<td>48.3</td>
<td>3.7</td>
<td>7.1</td>
<td>37.7</td>
<td>2.9</td>
<td>0.15</td>
<td>0.15</td>
<td></td>
</tr>
<tr>
<td>3. Pyrolite</td>
<td>45.1</td>
<td>3.3</td>
<td>8.0</td>
<td>38.1</td>
<td>3.1</td>
<td>0.2</td>
<td>0.4</td>
<td></td>
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<tr>
<td>Possible Picritic Parent Magmas</td>
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<td></td>
<td></td>
<td></td>
<td></td>
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<td></td>
<td></td>
</tr>
<tr>
<td>4. Eclogite extract</td>
<td>46.2</td>
<td>13.9</td>
<td>9.3</td>
<td>16.3</td>
<td>11.9</td>
<td>0.81</td>
<td>1.29</td>
<td>0.02</td>
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<tr>
<td>5. Oceanic crust</td>
<td>47.8</td>
<td>12.1</td>
<td>9.0</td>
<td>17.8</td>
<td>11.2</td>
<td>0.59</td>
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<td>0.03</td>
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<td>6. Tortuga dikes</td>
<td>47.3</td>
<td>13.6</td>
<td>9.8</td>
<td>17.6</td>
<td>9.6</td>
<td>0.79</td>
<td>0.89</td>
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<tr>
<td>7. High-MgO tholeiites</td>
<td>46.2</td>
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<td>11.0</td>
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<td>0.69</td>
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<tr>
<td>Kimberlite Eclogites</td>
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</tr>
<tr>
<td>8. Average</td>
<td>47.2</td>
<td>13.9</td>
<td>11.0</td>
<td>14.3</td>
<td>10.1</td>
<td>0.60</td>
<td>1.55</td>
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<tr>
<td>9. Roberts Victor</td>
<td>46.5</td>
<td>11.9</td>
<td>10.0</td>
<td>14.5</td>
<td>9.9</td>
<td>0.42</td>
<td>1.55</td>
<td>0.85</td>
</tr>
</tbody>
</table>

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 bimineralic eclogite in kimberlite.

Mantle is more-or-less uniformly enriched in these elements (Figure 25.1). As the magma layer or magma ocean cools, cumulates, containing intercumulus fluids, form. Peridotitic cumulates at shallower depths. Cumulate layers can have near-primitive ratios of Rb/Sr and Sm/Nd if they contain a moderate amount of interstitial fluid. Transfer of late-stage melts (KREEP or kimberlite) is one mechanism by which parts of the mantle become depleted and other regions 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.

A mechanism for creating an LIL-depleted but still fertile reservoir involves an early thick basalt layer in the relatively cold surface thermal-boundary layer. Small degrees of melt can be removed and still leave the basalt fertile. As the Earth cools this basalt layer converts to eclogite and sinks into the mantle, creating a depleted but fertile reservoir.

What is the fate of eclogite in the mantle? Some models assume that it sinks to the core-mantle boundary and is removed from the system; others assume that it is in the transition region or at the base of over-thickened crust. Estimated densities as a function of depth for eclogite and garnet peridotite are shown in Figure 25.2. Some eclogites and garnetites (garnet solid solutions) 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 below 650 km unless they are very garnet-rich and very cold. Whether eclogite can sink below 500 km depends on composition, temperature and the compressibility and thermal expansivity relative to
peridotite. Seismic tomography and the abrupt termination of earthquakes near 670 km depth suggests that oceanic lithosphere can penetrate to this depth.

Intrinsic density increases in the order basalt, picrite, depleted peridotite, fertile peridotite, garnet-rich eclogite. Some eclogites are less dense than some peridotites. Basalts and picrites crystallizing or recrystallizing below about 50 km can become denser than other upper-mantle assemblages. 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. The zero-pressure density of a typical eclogite is about 15% and 3% denser than basalts and fertile peridotites, respectively. With a coefficient of thermal expansion of $3 \times 10^{-5} ^\circ C$, 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. Some eclogites are less dense than some mantle peridotites at depths greater than some 200 km. If the intrinsic density of the deep mantle is only 1% greater than the shallower layers, it could be permanently trapped because of the very low expansivity at high pressure.

Simple Stokes' Law calculations show that inhomogeneities having density contrasts of 0.1 to 0.4 g/cm$^3$ 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 $10^{20}$ poises. This is orders of magnitude faster than average convective velocities. Inhomogeneities of that magnitude are 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 25.3 and 25.4. The unlikely alternative is that the reservoirs differ in trace elements but not major elements, intrinsic density or melting point. Small differences in bulk chemistry change the mineralogy and therefore the intrinsic density and melting point; 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 cannot be reversed by thermal expansion and 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 crust or 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 or crust thickens and cools, it becomes denser at its base than the underlying mantle and a potential instability develops. If the crust in island arcs, batholiths or compressional mountain belts gets thicker than about 50 km it is prone to delamination. Similarly, if the temperature in a deep garnet-rich layer or blob exceeds the solidus, the density may become less than the overlying layer. The large density changes associated with partial melting, delamination and the basalt-eclogite phase change may be more important in driving mantle convection than thermal expansion. In the deep mantle, chemical variations are more important than temperature variations in controlling density and density contrasts.

**Enriching fluids**

The composition of a residual fluid in equilibrium with eclogite, as a function of crystallization, is shown in Figures 25.5 and 25.6. The fractionations increase rapidly as the residual melt fraction drops below about 20% (that is, above 80% crystallization).

Kimberlites may represent such late-stage or residual fluids. They appear to have been in equilibrium with eclogite and often contain eclogite xenoliths. They also have LIL patterns that are complementary to MORB. Isotopic evolution in regions enriched by expelled small-degree melts will deviate significantly from the
primitive-mantle growth curve (Figure 25.6). The U/Pb ratio appears to behave similarly to the Rb/Sr ratio unless sulfides are involved. 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 25.5).

The distribution of oceanic ridges and hotspots suggests 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 where

$$\frac{(Rb/Sr)_{m}}{(Rb/Sr)_{res}} = \frac{F + (1-F)D_{Sr}}{F + (1-F)D_{Rb}} = f_m + 1$$

the $D$ are solid-melt partition coefficients and $F$ is the melt fraction. As the cumulate freezes, continuously or episodically losing its fluids to the overlying or underlying layer, it contains less of a more enriched fluid. The net result is a nearly constant $f$ for the cumulate as it evolves. Most of the fractionation that a crystallizing reservoir experiences occurs upon the removal of the first batch of melt.

Depleted reservoir become depleted by the removal of fluids representing late-stage interstitial fluids or small degrees of partial melting (Figure 25.1). The enriched fluid is not necessarily removed to the continental crust; it can also enrich the uppermost mantle and lithosphere. The enriched and depleted layers may differ in major elements and mineralogy, possibly the result of crystallization and gravitational separation. 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, or a cooling cumulate, become more fractionated with time.

Partial melting of the mantle during accretion, melt separation, crystal fractionation and formation of cumulate layers is one model
that can explain the geophysical and geochemical observations. The transfer of KREEP-like or kimberlitic material can explain the depletion and enrichment of various lithologies. Similar scenarios have been developed for the Moon.

The role of magma mixing

Melting regions in the mantle involve large degree melts toward the center and low-degree melting on the wings and at greater depth. Melts pond beneath the crust or lithosphere, collect in magma chambers, and mix or blend or hybridize. When two magmas are mixed, the composition of the mix, or hybrid, is

\[ xC_i^1 + (1 - x)C_i^2 = C_i^{\text{mix}} \]

where \( x \) is the weight fraction of magma \( i \), and \( C_i^1 \), \( C_i^2 \) and \( C_i^{\text{mix}} \) are, respectively, the concentration of the ith 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,

\[
\begin{align*}
\frac{(Rb/Sr)^{\text{mix}}}{(Sr_{87}/Sr_{86})_{\text{mix}}} &= x\left(\frac{Rb}{Sr}\right)_1 + (1 - x)\left(\frac{Rb}{Sr}\right)_2 \left(\frac{Sr_{87}/Sr_{86}}{Sr_{87}/Sr_{86}}\right)_1 \\
\end{align*}
\]

These are hyperbolas, and the shape or curvature of the hyperbola depends on the enrichment factor \( E \). 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 \( E \). This simple observation can explain a variety of geochemical paradoxes. For example, many basalts are clearly enriched, relative to primitive mantle, in some isotope and trace element ratios but enriched in others. Note that trace element and isotope ratios cannot simply be averaged; mixtures are weighted averages. Likewise, the means and variances of ratios have little meaning without information about the concentrations.

Trace-element and isotope data for magmas sometimes appear to be inconsistent. The incompatible elements and strontium and neodymium isotopes show that some abyssal tholeites (MORB) are from a reservoir that has current and time-integrated depletions of the elements that are fractionated into a melt. MORBs, however, have Pb-isotopic ratios suggesting long-term enrichment in U/Pb. Alkaline basalts and tholeiites from continents and oceanic islands are derived from LIL- and U/Pb-enriched reservoirs. Strontium- and neodymium-isotope ratios, however, appear to indicate that some of these basalts are derived from un-fractionated 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, or components, and as blends of small-degree and large-degree melts. MORBs can be thought of as slightly contaminated, depleted magmas, or as large-degree melts of a heterogenous source plus some small-degree melts. Oceanic-island and continental basalts are different, less homogenous, mixtures.
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 an enriched component or material from an enriched reservoir can explain the lead results for MORB (the ‘lead paradox’). Depleted basalts are more sensitive to lead than to rubidium or neodymium contamination. Similarly, continental and oceanic-island basalts may represent mixtures of enriched and depleted magmas, or small- and large-degree melts.

Figure 25.7 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 (Figure 25.8). 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 1.5–2.5 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, the start of enrichment could have been much earlier. The data shown in Figure 25.7 may 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.

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, 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); normal MORB may simply show less obvious signs of contamination. Magmas containing 10–20% contaminant will still appear isotopically depleted for Nd and Sr. Such mixtures will appear to exhibit long-term enrichment in the lead isotopic systems. Clearly, the mixing idea can be extended to multiple components, and to systems that are cooling and fractionating as they mix.
CRYSTALLIZATION OF THE 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 $\varepsilon_{Nd}$ in the depleted and enriched cells (upper right) combined with similar diagrams for Sr generates the mantle array (upper left). Magma mixing reduces the spread of values, decreasing the apparent ages of the depletion/enrichment events. Layer D may be the transition region. The cumulate layer could also be lower crustal cumulates.

**Fig. 25.8** 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 $\varepsilon_{Nd}$ in the depleted and enriched cells (upper right) combined with similar diagrams for Sr generates the mantle array (upper left). Magma mixing reduces the spread of values, decreasing the apparent ages of the depletion/enrichment events. Layer D may be the transition region. The cumulate layer could also be lower crustal cumulates.

**Melts from enriched or heterogenous mantle**

Even if the enriched parts of the mantle are homogenous, their partial melts will have variable LIL contents and ratios such as Rb/Sr, Sm/Nd and U/Pb that 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 homogenous. An example is shown in Figure 25.9.

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 50–95% MORB and an enriched component representing 2–20% 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.
evidence for enrichment and depletion of the various mantle reservoirs. LIL-enrichment associated with time-integrated isotopic depletion is a simple consequence of mixing relations. Hotspot, oceanic-island and continental flood basalts are mixtures of a variety of components, including recycled oceanic crust, delaminated continental crust, 'contaminated' MORB and so on. If basalts are mixtures, the isotopic ages of their parent components or reservoirs will be underestimated, the sources are more enriched or depleted in their isotopic ratios than the hybrid magmas. 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 – and the diversity of small-scale samples, supports the concept of a heterogenous mantle. Part of the heterogeneity is due to plate-tectonic processes, as discussed in earlier chapters, and part is due to igneous petrology, the transfer around of small degree melts. The final stage of eruption involves commingling and homogenization of melts. Mantle convection and stirring probably has little to do with the homogeneity of midocean-ridge basalts and observed mixing arrays.

Fig. 25.9 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 hypothesis that oceanic and continental magmas represent mixtures of melts from diverse lithologies – enriched and depleted components – appears capable of explaining a variety of geochemical data and resolving some isotopic paradoxes. The hypothesis explains apparently contradictory trace-element and isotopic