

Theory of the Earth

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Chapter 9. The Source Region

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Abstract:

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The Source Region

*Long is the way
And hard, that out of hell leads up to light.*

— JOHN MILTON

Magmas probably represent partial, rather than complete, melts of the basalt source region. Much or most of the source region is left behind when a melt is extracted from a partially molten source region. The composition, mineralogy and depth of the source region are all controversial. Little progress can be made without a simultaneous consideration of field and experimental petrology, isotopes and geophysics. Many plausible scenarios can be constructed from parts of the data, but most have little merit when tested against other sets of data. Magmas are just part of the story. The denser and more refractory parts of the mantle are the subject of this chapter.

BACKGROUND

The rocks of the mantle are compounds involving predominantly MgO, SiO₂ and FeO. These oxides account for more than 90 percent of the mantle. Intuitively, one might expect that the bulk properties of the mantle, such as density and seismic velocity and the locations of phase change discontinuities, are controlled only by compounds such as (Mg,Fe)SiO₃ and (Mg,Fe)₂SiO₄ and their high-pressure phases. Most discussions of mantle chemistry are based on this assumption. CaO and Al₂O₃ are minor constituents in the Earth as a whole, accounting for less than 8 percent of the mantle; however, they have an importance far beyond their abundance. In an alumina-poor mantle the mineralogy is dominated by olivine and pyroxene and their high-pressure forms such as β - and γ -spinel, SiO₂ (stishovite) and majorite. However, a small amount of Al₂O₃ completely changes the mineralogy. (Mg,Fe)O, CaO and SiO₂ combine with Al₂O₃ to form garnet, which may be the

dominant mineral between 400 and 650 km. Generally, changes in mineralogy have a greater effect on density and seismic velocities than small changes in composition of a given phase. CaO and Al₂O₃ change the melting point as well as the mineralogy and depths of phase changes. They therefore influence the evolution, stratigraphy and seismic profiles of the mantle. Some regions of the mantle may, in fact, be CaO- and Al₂O₃-rich.

The major shallow-mantle minerals, MgSiO₃ (pyroxene) and Mg₂SiO₄ (olivine), are unstable at high pressure and transform to denser phases in the spinel, garnet (majorite), ilmenite, perovskite, periclase and rutile structures. The small amount of FeO in the upper mantle readily substitutes for MgO and serves to increase the density, decrease the seismic velocities and lower the transition pressure for mantle phase changes. CaO and Al₂O₃ drastically alter the phase assemblages and the densities in the mantle. They also are major constituents of the basaltic or low-melting fraction of the mantle and serve to decrease the melting point. Al₂O₃ stabilizes the garnet structure, which means that the basaltic fraction of the mantle is denser than "normal" or Al₂O₃-poor upper mantle, but less dense than the perovskite-rich lower mantle in the pressure interval over which garnet is stable. Early differentiation of the Earth tends to separate the Al₂O₃-rich (basalt, eclogite) and Al₂O₃-poor silicates (peridotite), and the density differential is such that there should be little remixing. It is probable that the major geochemical reservoirs of the mantle, which have remained isolated for a large fraction of Earth history, also differ in Al₂O₃ content and intrinsic density. Reservoirs rich in Al₂O₃ and CaO are called "fertile."

We do not know the composition or mineralogy of the MORB reservoir. At the depth of MORB genesis, basalts

would crystallize as eclogite. Eclogite is a dense clinopyroxene-garnet-rich rock that is the high-pressure form of MORB and picrite, the postulated direct parent of MORB. The parent reservoir, in addition, probably contains orthopyroxene and olivine, the other major minerals of the mantle. The composition of the MORB reservoir can therefore range from eclogite to garnet peridotite, depending on depth and the extent of partial melting and crystal fractionation occurring prior to magma separation. The bulk of the material erupting at oceanic islands and as continental flood basalts may also come from either a garnet peridotite or eclogite source region. The source region for so-called hot-spot magmas and midocean-ridge basalts may, in fact, be the same if hot-spot magmas are contaminated in the shallow mantle prior to eruption.

The generally accepted view that basalts represent various degrees of partial melting, up to about 20 percent, of a garnet peridotite source region is based on several lines of evidence and assumptions (Yoder, 1976). Peridotites are a common rock type and are found as xenoliths in kimberlite and magma, in obducted sections of oceanic lithosphere and in dredge hauls at fracture zones. The seismic properties of the shallow mantle are generally consistent with peridotite. Garnet peridotites can form basalts by partial melting, and some peridotites appear to be the refractory residue remaining after basalt extraction. Other potential source materials, such as eclogite or pyroxenite, are not only rarer but are thought (erroneously) to require more extensive melting in order to provide basaltic magmas. Melt-crystal separation is believed to occur before such extensive melting can occur. These latter arguments only hold if the source region is identical in chemistry to the basalts it produces and if melt-crystal separation is more effective at the depth of initial melting than source buoyancy. Eclogites cover a wide range of compositions and they melt in a eutectic-like fashion, thus basalts can form from eclogites over a wide range of partial melting conditions.

There are several developments that have reopened the question of the nature and depth of the basalt source region. Seismic evidence strongly suggests that the source of midocean-ridge basalts is in the transition region, well below the low-velocity zone. The properties of the shallow mantle are, therefore, not relevant. The low-velocity zone itself is highly variable in depth, thickness and velocity. It locally extends below 400 kilometers. Midocean-ridge basalts are possibly derivative from picrites after extensive crystal fractionation and may, therefore, require more extensive melting than previously thought. Komatiites either require more than 60 percent melting, indicating that large amounts of melting are possible prior to melt separation, or that they formed at great depth; in either case high temperatures are implied. High melt densities make melt-crystal separation more difficult at great depth. If diapiric ascent is rapid, extensive melting can occur because of adiabatic decompression. Melt-crystal separation, or crystal fraction-

ation, may be restricted to depths less than 100 km, where diapiric ascent is slowed by high viscosities, high-strength lithosphere or buoyancy considerations. Finally, the elastic properties of the transition zone (see Chapter 4) are compatible with an eclogite-rich composition.

Basalts are chemically equivalent to garnet plus clinopyroxene (diopside and jadeite) and melting of these minerals is involved in basalt genesis. Experimental petrology, however, cannot tell us if the garnet and clinopyroxene are dispersed, as in peridotite, or how much olivine and orthopyroxene are in the source region. The basalt fraction of pyrolite, a hypothetical source peridotite, is arbitrary. Experimental petrology cannot constrain the bulk composition of the source region. The existence of meteorites rich in olivine has provided support for the argument that peridotite is the parental basalt material in the Earth. On the other hand, cosmochemical models of the Earth have more pyroxene and less olivine than most peridotite models for the mantle, and partial melting of meteorites does not yield liquids with appropriate iron-magnesium ratios: Meteorite olivine are generally much higher in FeO than mantle olivines. There is little doubt that there is abundant olivine in the mantle. The issues are whether it, and its high-pressure phases, are distributed uniformly throughout the mantle and whether it is the dominant mineral in the basalt source region.

Observations that have been used in support of an olivine-rich (>50 percent) source region for basalts—garnet peridotites are stable in the upper mantle, their bulk compositions are consistent with the materials forming the Earth, they are prominent among the recovered deep-seated samples, they are capable of yielding material of basaltic composition (Yoder, 1976)—deal with possible mineral assemblages and bulk compositions somewhere in the mantle rather than dealing directly with the basalt source region. The "conclusion" that garnet peridotite is the immediate parent of basaltic or picritic magma is actually a working hypothesis, not well established (Yoder, 1976). Melting of garnet and clinopyroxene, whether embedded in a peridotite or eclogite matrix, will form basalt. The eclogite itself may well have been derived from partial melting or crystal fractionation of peridotite at an earlier stage of Earth evolution. Yoder (1976) and Carmichael and others (1974) have given well-balanced treatments of the various possible source rocks. A source region with less than 50 percent olivine cannot be ruled out.

The effects of melt compressibility and matrix stiffness (reviewed in Chapter 7) are such that separation of melt is difficult at high pressure (Rigden and others, 1984). When sufficient melting occurs at depth, the whole source region may become unstable, bringing melt plus matrix to shallow levels where melt separation can occur. In a chemically stratified mantle the high temperature gradient in a thermal boundary layer causes melting to initiate in the deeper layer. Melt can be retained at depth if the overlying mantle has

high viscosity and low permeability. The high density, and possible high viscosity, of melts at depth also makes melt extraction more difficult than at shallower levels. The increased buoyancy of a partially molten source region may be more important than the density contrast between melts and residual crystals. Garnet exsolution at high temperature can also cause a deeper layer to become buoyant relative to a colder shallow peridotite layer. These effects serve to accentuate the instability of thermal boundary layers in a chemically stratified upper mantle. Such boundary layers may occur at the base of the upper mantle (near 650 km) and near 400 km. Of course, the instability breaks the strict layering, and such a mantle will exhibit both radial and lateral inhomogeneity, as suggested by the seismic results discussed in Chapter 10.

The previous chapter discussed the properties of the various types of magmas that emerge from the mantle. These **magmas** provide clues about the chemistry of some parts of the mantle but do not provide a unique composition or mineralogy. We do not even know what fraction of the mantle the basalts represent or from what depth they originated. Presumably, the parts that are left behind are denser and, possibly, more refractory. It is not clear that we have samples of the original source rock or even a representative sample of the residual material. Various attempts have been made to infer the properties of the basalt source region from rocks that are exposed at the surface, and we now turn our attention to observed and hypothetical rock types that may be important in the upper mantle.

ULTRAMAFIC ROCKS

Ultramafic rocks are composed chiefly of ferromagnesian minerals and have a low silicon content. The term is often used interchangeably with "ultrabasic," although pyroxene-rich rocks are ultramafic but not ultrabasic because of their high SiO_2 content. Peridotites, lherzolite, dunite and harzburgite are specific names applied to ultramafic rocks that are chiefly composed of olivine, orthopyroxene, clinopyroxene and an aluminous phase such as plagioclase, spinel or garnet. Ultramafic rocks are dense and mainly composed of refractory minerals with high seismic velocities. Most of the shallow mantle is probably ultramafic in composition. Basic rocks, such as basalts, become dense at high pressure (for example, eclogite) and can have properties comparable to the more refractory peridotites. The main subdivisions of ultramafic rocks, along with the dominant minerals, are garnet peridotite (olivine, orthopyroxene, clinopyroxene, garnet), lherzolite (olivine, orthopyroxene, clinopyroxene), harzburgite (olivine, orthopyroxene), wehrlite (olivine, clinopyroxene), dunite (olivine \pm clinopyroxene), websterite (orthopyroxene, clinopyroxene) and eclogite (clinopyroxene, garnet). Some eclogites overlap basalts in their bulk

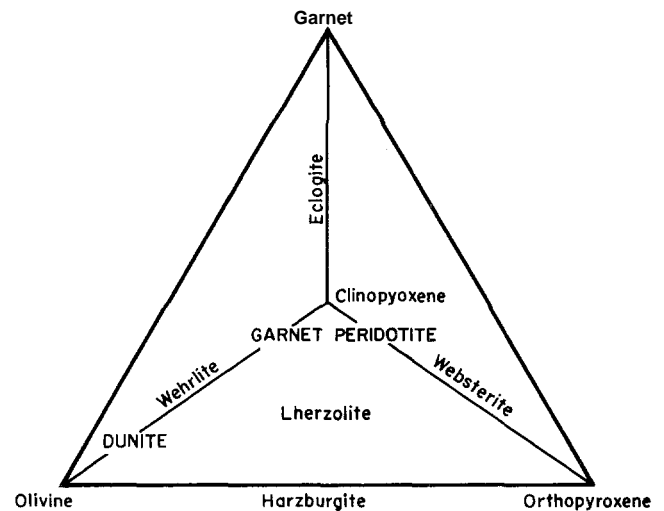


FIGURE 9-1
Nomenclature tetrahedron for assemblages of olivine, clinopyroxene, orthopyroxene and garnet. Dunites and garnet peridotites lie within the tetrahedron.

chemistry. The relationships between these rocks are shown in Figure 9-1.

Peridotites can represent

1. The refractory residue left after basalt extraction
2. Cumulates formed by the crystallization of a magma
3. Primitive mantle that can yield basalts by partial melting
4. Cumulates or residues that have been intruded by basalt
5. High-pressure or high-temperature melts (Herzberg, 1984, 1986)

Peridotites are divided into fertile or infertile (or barren) depending on their Al_2O_3 , CaO and Na_2O content. Fertile peridotites can be viewed as having an appreciable basaltic component. The terms "enriched" and "depleted" are often used interchangeably with "fertile" and "infertile" but have trace-element and isotopic connotations that are often inconsistent with the major-element chemistry. I use "infertile" and "barren" as attributes of a rock poor or very poor in CaO and Al_2O_3 and "depleted" for rocks poor in the incompatible trace elements and having low ratios of Rb/Sr , LREE/HREE and so on. These distinctions are necessary since some rocks are fertile yet depleted, for instance, the source region for midocean-ridge basalts. Table 9-1 gives compositions for representative ultramafic rocks.

Garnet lherzolites are composed mainly of olivine and orthopyroxene (Table 9-2). Olivine is generally in the range of 60 to 70 volume percent and orthopyroxene 30 to 50 percent. The average clinopyroxene and garnet proportions

TABLE 9-1
Compositions of Spinel and Garnet Lherzolites

Oxide	Spinel Lherzolite		Garnet Lherzolite
	Continental (avg. of 301)	Oceanic (avg. of 83)	
SiO ₂	44.15	44.40	44.90
Al ₂ O ₃	1.96	2.38	1.40
FeO	8.28	8.31	7.89
MgO	42.25	42.06	42.60
CaO	2.08	1.34	0.82
Na ₂ O	0.18	0.27	0.11
K ₂ O	0.05	0.09	0.04
MnO	0.12	0.17	0.11
TiO ₂	0.07	0.13	0.06
P ₂ O ₅	0.02	0.06	—
NiO	0.27	0.31	0.26
Cr ₂ O ₃	0.44	0.44	0.32

Maaløe and Aoki (1977).

are about 5 percent and 2 percent, respectively (Maaløe and Aoki, 1977).

The major oxides in peridotites and lherzolites generally correlate well (Figure 9-2 and Table 9-3). An increase in MgO correlates with decreases in SiO₂, Al₂O₃, CaO, TiO₂ and Na₂O and an increase in NiO. CaO, Al₂O₃ and Na₂O all go to approximately zero at an MgO content of about 48 weight percent. Cr₂O₃, MnO and FeO are roughly constant. The lherzolite trend can be explained by variable amounts of clinopyroxene and garnet.

Olivine- and orthopyroxene-rich rocks, presumably from the mantle, are found in foldbelts, ophiolite sections, oceanic fracture zones and, as xenoliths, in kimberlites and alkali-rich magmas. They are rare in less viscous magmas such as tholeiites. Olivine and orthopyroxene in varying proportion are the most abundant minerals in peridotites. These are dense refractory minerals, and peridotites are therefore generally thought to be the residue after melt extractions. Some peridotites are shallow cumulates deposited

from cooling basalts and are therefore not direct samples of the mantle. Alumina in peridotites is distributed among the pyroxenes and accessory minerals such as plagioclase, spinel and garnet. At higher pressure most of the Al₂O₃ would be in garnet. Garnet-rich peridotite, or pyrolite, is the commonly assumed parent of mantle basalts. This variety is fertile peridotite since it can provide basalt by partial melting. Most peridotites, however, have relatively low Al₂O₃ and can be termed barren. These are commonly thought to be residual after melt extraction. Al₂O₃-poor peridotites are less dense than the fertile variety and should concentrate in the shallow mantle. Given sufficient water at crustal and shallow mantle temperatures, peridotite may be converted to serpentinite with a large reduction in density and seismic velocity. Hydrated upper mantle may therefore be seismically indistinguishable from lower crustal minerals.

Lherzolites typically contain 60 to 80 percent olivine, 20 to 40 percent orthopyroxene, less than 14 percent clinopyroxene and 1 to 10 percent of an aluminous phase such as spinel or garnet. Spinel lherzolites, the lower-pressure assemblages, dredged from the ocean bottom are similar in composition to those found in alkali basalts and kimberlites on oceanic islands and continents. Garnet lherzolites, the higher-pressure assemblages, have lower Al₂O₃, CaO and FeO and are therefore denser than spinel lherzolites only when they contain appreciable garnet. They would become less dense at higher temperature, lower pressure or if partially molten.

The major-element chemistries of lherzolites vary in a systematic fashion. Most of the oxides vary linearly with MgO content. SiO₂, Al₂O₃, CaO, Na₂O and TiO₂ decrease with an increase in MgO. These trends are generally consistent with variable amounts of a basaltic component. However, the basaltic component, especially for the MgO-SiO₂ trend, is not tholeiitic or MORB. If lherzolites represent olivine-orthopyroxene-rich rocks with variable amounts of melt extraction or addition, this melt component is andesitic in major elements.

The major-element trends of lherzolites may also be controlled by melt-crystal equilibration at various depths in

TABLE 9-2
Mineralogy of Spinel and Garnet Lherzolites

Mineral	Spinel Lherzolite		Garnet Lherzolite	
	Average (wt. pct.)	Range (vol. pct.)	Average (vol. pct.)	Range (vol. pct.)
Olivine	66.7	65-90	62.6	60-80
Orthopyroxene	23.7	5-20	30	20-40
Clinopyroxene	7.8	3-14	2	0-5
Spinel	1.7	0.2-3	—	—
Garnet	—	—	5	3-10
Phlogopite	—	—	0.4	0-0.5

Maaløe and Aoki (1977)

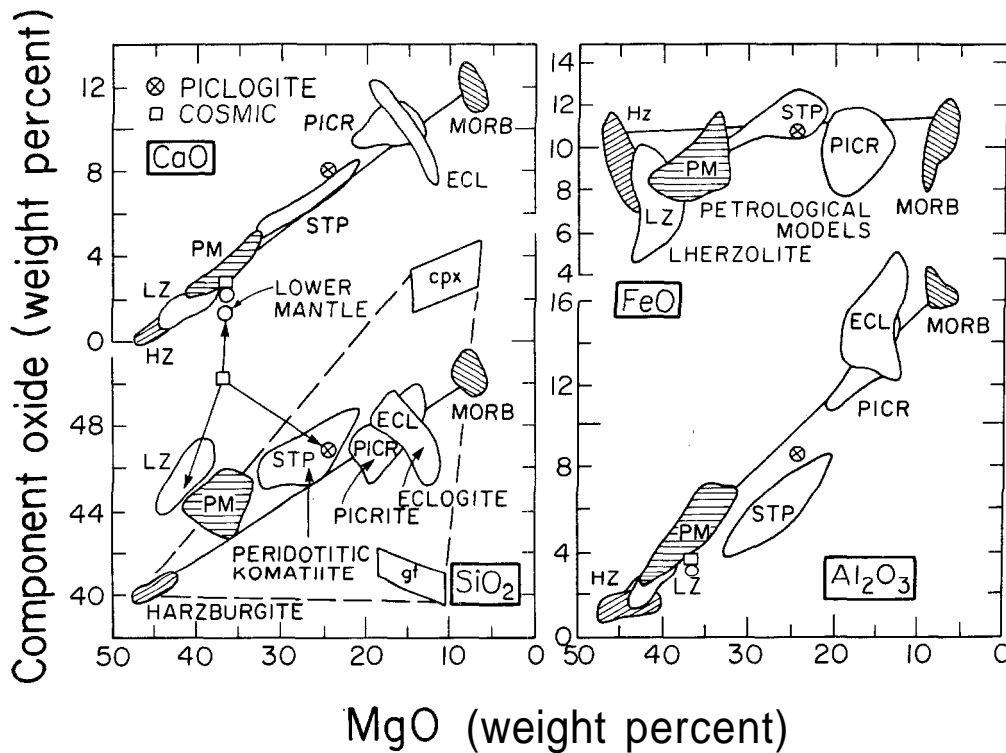


FIGURE 9-2

SiO_2 , CaO , Al_2O_3 and FeO versus MgO for igneous rocks. The basalt source region probably has a composition intermediate between basalt (MORB) and harzburgite. Most petrological models (PM) of the major-element chemistry of the source region favor a small basalt fraction. STP (spinifex textured peridotites) are high-temperature MgO -rich magmas. Picrites (PICR) are intermediate in composition between STP and MORB and may evolve to MORB by olivine separation. Picrites and eclogites (ECL) overlap in composition. Lherzolites (LZ) contain an orthopyroxene component, but the other rock types are mainly clinopyroxene + garnet \pm olivine. Squares represent estimates of primitive mantle composition based on a chondritic model. If the upper mantle is primarily lherzolite, basalt and harzburgite, the lower mantle (open dot) will be primarily orthopyroxene. The composition of the MORB source (piclogite model) probably falls between PICR and PM or STP.

the mantle. Lherzolites, and most other ultramafic rocks, are generally thought to be the refractory residue complementary to melts presently being extracted from the mantle. They differ, however, from primitive mantle compositions. In particular they contain more olivine and less orthopyroxene than would be appropriate for a chondritic or "cosmic" mantle. Upper-mantle lherzolites and basalts may be complementary to the lower mantle, representing melts from the original, accretional differentiation of the mantle. The MgO content of melts increases with temperature and with depth of melting. At great depth (>200 km) relatively low- MgO phases, such as orthopyroxene and garnet-majorite may remain behind, giving olivine-rich melts. The major-element trends in lherzolites may therefore represent trends in high-pressure melts (Herzberg, 1984; Herzberg and others, 1986).

Unserpentinized peridotites have seismic velocities and anisotropies appropriate for the shallow mantle. This situation is often generalized to the whole mantle, but seis-

mic data for depths greater than 400 km are not in agreement with that hypothesis. It is not even clear that peridotite has the proper seismic properties for the lower lithosphere. In the depth interval 200–400 km both eclogite and peridotite can satisfy the seismic data.

Garnet pyroxenites and eclogites are also found among the rocks thought to have been brought up from the mantle, and they have physical properties that overlap those of the ultramafic rocks. Garnet-rich pyroxenites and eclogites are denser than peridotites and should therefore occur deeper. The extrapolation of the properties of peridotites to the deep upper mantle, much less the whole mantle, should be done with caution. Not only do other rock types emerge from the mantle, but there is reason to believe that peridotites will be concentrated in the shallow mantle. The common belief that seismic data *require* a peridotite mantle is simply false. The V_p and V_s velocities in the upper 150 km of the mantle under stable shield areas are consistent with olivine-rich perido-

TABLE 9-3
Compositions of Peridotites and Pyroxenite

Oxide	Lherzolites			Dunite (4)	Pyrox- enite (5)	Peridotites		
	Spinel (1)	(2)	Garnet (3)			(6)	(7)	(8)
SiO ₂	44.15	44.40	44.90	41.20	48.60	44.14	46.36	42.1
Al ₂ O ₃	1.96	2.38	1.40	1.31	4.30	1.57	0.98	
FeO	8.28	8.31	7.89	11.0	10.0	8.31	6.56	7.10
MgO	42.25	42.06	42.60	43.44	19.10	43.87	44.58	48.3
CaO	2.08	1.34	0.82	0.80	13.60	1.40	0.92	
Na ₂ O	0.18	0.27	0.11	0.08	0.71	0.15	0.11	
K ₂ O	0.05	0.09	0.04	0.016	0.28	—	—	
MnO	0.12	0.17	0.11	0.15	0.18	0.11	0.11	
TiO ₂	0.07	0.13	0.06	0.06	0.83	0.13	0.05	
P ₂ O ₅	0.02	0.06	—	0.10	0.10	—	—	
NiO	0.27	0.31	0.26	—	—	—	—	
Cr ₂ O ₃	0.44	0.44	0.32	—	—	0.34	0.33	
H ₂ O	—	—	—	0.50	0.90	—	—	

(1) Average of 301 continental spinel lherzolites (Maaløe and Aoki, 1977).

(2) Average of 83 oceanic spinel lherzolites (Maaløe and Aoki, 1977).

(3) Average garnet lherzolite (Maaløe and Aoki, 1977).

(4) Dunite (Beus, 1976).

(5) Pyroxenite (Beus, 1976).

(6) High-T peridotites, South Africa (Boyd, 1986).

(7) Low-T peridotites, South Africa (Boyd, 1986).

(8) Extrapolated lherzolite trend (~0 percent Al₂O₃, CaO, Na₂O, etc.).

tite. Elsewhere in the shallow mantle both eclogite and peridotite are consistent with the data.

It should be kept in mind that the most abundant rock type by far that emerges from the mantle is tholeiite (equivalent to garnet plus clinopyroxene). A garnet peridotite with suitable compositions of garnet and clinopyroxene, particularly Na₂O and TiO₂ content, can certainly yield basalts by partial melting. The amount of melting, however, and the amount of residual olivine and orthopyroxene are unknown. A suitable natural garnet peridotite (pyrolite) has not yet been found that can provide the major- and trace-element abundances found in, for example, midocean-ridge tholeiites. Such a peridotite would have to have high Al₂O₃, Na₂O, CaO and TiO₂ and low concentrations of the incompatible elements. Pyrolite (next section) therefore remains a hypothetical mantle rock.

If picrites are the parent for tholeiitic basalts, then roughly 30 percent melting is implied for generation from a shallow peridotitic parent. If the parent is eclogitic, then similar temperatures would cause more extensive melting. However, generation of basaltic magmas from an eclogitic parent does not require extensive melting. Melts of basaltic composition are provided over a large range of partial melting.

Basalts and peridotites are two of the results of mantle differentiation. They both occur near the surface of the Earth and may not represent the whole story. They are also

not necessarily the result of a single-stage differentiation process. In the next chapter I discuss the possible evolution of the mantle, from its high-temperature birth to the formation of the geochemical reservoirs that we sample today. This complements the standard approach that attempts to take the observed rocks and work backward in time.

PYROLITE

The general idea of a peridotitic mantle as a source for basaltic magmas by partial melting goes back at least as far as Bowen (1928). Ringwood (1962) formalized the concept in a series of papers starting in 1962. He proposed the existence of a primitive mantle material that was defined by the property that on fractional melting it would yield a typical basaltic magma and leave behind a residual refractory dunite-peridotite. This he called "pyrolite" (pyroxene-olivine-rock). The initially proposed composition was one part basalt and four parts dunite, although this is arbitrary. Green and Ringwood (1967) considered that 20–40 percent melting will be necessary before liquid segregates and begins an independent existence, and this mechanical idea influenced their choice of basalt-to-dunite ratios. The stable phase assemblages in pyrolite, and the disposition of Al₂O₃, depend on temperature and pressure. Garnet pyrolite is essentially identical with garnet peridotite but is more fertile

than most natural samples (Ringwood, 1962). Ringwood earlier argued that garnet pyrolite would not yield, by direct partial melting, basaltic magma and proposed that partial melting of plagioclase pyrolite must be the main source of basalts. He attributed the low-velocity zone to the presence of plagioclase and assembled many arguments in favor of these conjectures. Plagioclase is not in fact stable at upper-mantle pressures, and Ringwood revised his pyrolite model and ideas about the basalt source region many times in the subsequent years.

In an early model garnet pyrolite was considered to be free of orthopyroxene. In 1963 Green and Ringwood calculated a pyrolite composition based on the assumption that it was three parts dunite plus one part of the averages of tholeiitic and alkali olivine basalt; it therefore contained some orthopyroxene. In 1966 Ringwood assumed that pyrolite was a three-to-one mix of Alpine-type peridotite and a Hawaiian olivine-tholeiite. Table 9-4 gives compositions of some of these pyrolite models.

The cornerstones of the pyrolite hypothesis were that the basaltic magmas that were thought to represent the largest degrees of partial melting, tholeiites, were primitive, unfractionated melts and that they were in equilibrium with the residual peridotite at the depth of melt separation. O'Hara (1968) argued that extensive crystal fractionation, at depth, operated on the parent magma and, therefore, basalts including tholeiites were not primary magmas. He suggested that partial melting and magma segregation occur at pressures greater than 30 kbar, rather than the 5–25 kbar in the Green-Ringwood models. Under these conditions the parent magma is picritic, and extensive olivine crystallization at low pressure produces tholeiitic magmas. He also

proposed that extensive eclogite fractionation at depths of order 100 km occurred. Ringwood (1975) argued strongly against O'Hara's model.

Jaques and Green (1980) on the basis of an extensive series of melting experiments on synthetic peridotites abandoned the earlier model of Green and Ringwood (1967) and Green (1971), which involved the segregation of olivine tholeiite magmas at about 30 km depth. Studies by Green and others (1979) and Stolper (1980) have shown that primary picritic magmas are likely parents for midocean-ridge basalts by separation of about 15 percent olivine at about 60–70 km depth and temperatures of about 1400–1450°C. Picritic and komatiitic magmas require about 40–60 percent melting of peridotite if the melting initiates at such shallow depths. Green and others showed that there can be no simple genetic relationship between the primary tholeiitic picritic magmas that are parental to midocean-ridge basalts and the residual harzburgite or accumulate peridotite of ophiolite complexes. Thus, these studies effectively pull the rug out from under the basic pyrolite assumptions. Nevertheless, the same authors proposed yet more pyrolite models that are mixtures of picrite and varying, arbitrary proportions of olivine (50–69 percent), orthopyroxene (0–25 percent) and spinel (0.5 to 1 percent). The new pyrolites are richer in Al_2O_3 and CaO than previous pyrolite models and less rich in TiO_2 , Na_2O and K_2O . Their preferred model generates the primary midocean-ridge basalts by 24 percent partial melting.

At this point one is reminded of Karl Popper's (1972) "conventionalist stratagem": "Some genuinely testable theories, when found to be false, are still upheld by their admirers by introducing ad hoc some auxiliary assumptions

TABLE 9-4
Compositions of Pyrolites and Garnet Peridotites (Weight Percent)

Oxide	Pyrolite				Garnet Peridotite	
	(1)	(2)	(3)	(4)	(5)	(6)
SiO_2	45.1	42.7	46.1	45.0	42.5	46.8
TiO_2	0.2	0.5	0.2	0.2	0.1	0.0
Al_2O_3	3.3	3.3	4.3	4.4	0.8	1.5
Cr_2O_3	0.4	0.5	—	0.5	—	—
MgO	38.1	41.4	37.6	38.8	44.4	42.0
FeO	8.0	6.5	8.2	7.6	3.8	4.3
MnO	0.15	—	—	0.11	0.10	0.11
CaO	3.1	2.1	3.1	3.4	0.5	0.7
Na_2O	0.4	0.5	0.4	0.4	0.1	0.1
K_2O	0.03	0.18	0.03	0.003	0.22	0.02

(1) Ringwood (1979), p. 7.

(2) Green and Ringwood (1963).

(3) Ringwood (1975).

(4) Green and others (1979).

(5) Boyd and Mertzman (1987).

(6) Boyd and Mertzman (1987).

or by re-interpreting the theory ad hoc in such a way that it escapes refutation. Such a procedure is always possible, but it rescues the theory from refutation only at the price of destroying, or at least lowering its scientific status."

Jaques and Green (1980), in common with most petrologists, have restricted their attention to rocks that can form basalts by a relatively small degree of partial melting since high degrees of partial melting may not be geologically feasible. They point to laboratory measurements showing strong crystal settling of olivine when melting exceeds 40 percent, implying that conditions in a convecting mantle are the same as in a static laboratory experiment. The real Earth has a cold, high-viscosity, buoyant, strong and relatively impermeable lithosphere. An important question, then, is how much melting can occur in the underlying convecting mantle that cannot easily lose its melt fraction to the surface? What are the likely products of partial melting in a convecting, differentiating mantle?

In a garnet-clinopyroxene-rich rock there is a small temperature differential between the onset of melting and the temperature of extreme melting. Adiabatic ascent of such a rock can lead to extensive melting. In contrast, in a garnet peridotite, the eclogite fraction is melted near the solidus but the bulk of the rock does not melt until much higher temperatures, near the liquidus. Therefore, garnet peridotites require a large temperature increase in order to melt extensively.

The pyrolite hypothesis was based entirely on major elements and on several arbitrary assumptions regarding allowable amounts of basalt and melting in the source region. It does not satisfy trace-element or isotopic data on basalts, and it violates chondritic abundances and evidence for mantle heterogeneity.

ECLOGITES

Hauy (1822) introduced the term "eclogite" for rocks composed of omphacite (diopside plus jadeite) and garnet, occasionally accompanied by kyanite, zoisite, amphibole, quartz and pyrrhotite. Natural eclogites have a variety of associations, chemistries, mineralogies and origins, and many names have been introduced to categorize these subtleties. Nevertheless, the term "eclogite" implies different things to different workers. To some eclogites mean metamorphic crustal rocks, and to others the term implies biminerally kimberlite xenoliths. The chemical similarity of some eclogites to basalts prompted early investigators to consider eclogite as a possible source of basalts but more recently has been taken as evidence that these eclogites are simply subducted oceanic crust or basaltic melts that have crystallized at high pressure. Some eclogites are demonstrably metamorphosed basalts, while others appear to be igneous rocks ranging from melts to cumulates. The trend

in recent years has been toward the splitters rather than the lumpers, and toward explanations that emphasize the derivative and secondary nature of eclogite rather than the possible importance of eclogite as a source rock for basaltic magmas.

Pyroxene-garnet rocks with jadeite-bearing pyroxenes are found as inclusions in alkali basalt flows as layers in ultramafic intrusions, as inclusions in kimberlite pipes, as tectonic inclusions in metamorphic terranes associated with gneiss and schist, and as inclusions in glaucophane schist areas. Jadeite-poor garnet clinopyroxenites are abundant in Salt Lake Crater, Hawaii. Only the eclogites from kimberlite pipes are demonstrably from great depth, that is, well into the upper mantle. Some of these contain diamonds.

Peridotites and lherzolites predominate over all other rock types as inclusions in diamond-bearing kimberlites. The presence of diamond indicates origin depths of at least 130 km, and other petrological indicators suggest depths even greater. In a few kimberlite pipes eclogites form the majority of inclusions. The overwhelming majority of peridotites and lherzolites are infertile, that is, very low in aluminum, sodium and calcium. If the distribution of rock types in kimberlite inclusions is representative of the source region, the majority of basaltic components in the upper mantle resides in eclogites rather than in an olivine-rich rock (Smyth and others, 1984).

Table 9-5 shows typical clinopyroxene and garnet compositions of eclogites and peridotites, a synthetic twomineral eclogite and for comparison, a typical MORB. Note that, in general, diopside plus garnet from peridotite does not approximate the composition of MORB. In most cases, however, omphacite and garnet from eclogite bracket MORB compositions, and therefore eclogite is a more appropriate source rock. Table 9-6 gives comparisons of the bulk chemistry of some South African eclogites and MORB and an estimate of the average composition of the oceanic crust. There is a close correspondence between the composition of kimberlite eclogites and the material in the oceanic crust.

Many kimberlite eclogites show signs of garnet exsolution from clinopyroxene, implying that there can be substantial changes in the density of eclogites as a function of temperature in the subsolidus region. Clinopyroxenes in eclogites have exsolved 20 percent or more garnet, implying a substantial increase in pressure or decrease in temperature (Smyth and Caporuscio, 1984). A pressure increase is unlikely. A representative eclogite can increase in density by 2.5 percent by cooling from 1350°C to 950°C at 30 to 50 kbar. The reverse process can happen as garnet plus clinopyroxene is elevated into a lower pressure regime along an adiabat. For example, the density of an eclogite can decrease by about 3 percent simply by rising 50 km. Thus, garnet exsolution caused by pressure release can accomplish more than a 1000°C rise in temperature, by thermal expansion, all in the subsolidus regime. This plus the low

TABLE 9-5
Typical Trace Element Concentrations (ppm) and Ratios in Eclogite and Peridotite Minerals

	Eclogites		Synthetic	Peridotites		MORB
	Omphacite	Garnet	Eclogite	Diopside	Garnet	
K	1164	337	820	615	296	700
Rb	0.565	1.14	0.7	2.1	1.45	0.4
Sr	249	8.25	95	337	5.50	110
Na	52244	74	12700	1332	2420	17300
Ti	4856	899	2500	659	959	5500
Zn	106	15	55	28	69	80
Rb/Sr	0.002	0.14	—	0.006	0.26	0.004
Sm/Nd	0.206	0.522	—	0.211	0.590	0.335
Rb/K	4.9*	33.8*	8.5*	34.1*	49.0*	5.7*

Basu and Tatsumoto (1982), Wedepohl & Muramatsu (1979).

* × 10⁻⁴.

melting point of eclogite means that rising eclogite-rich convection currents obtain considerable buoyancy.

The MORB source region has been depleted by removal of either a small melt fraction or a late-stage intercumulus fluid. The abundances of uranium, thorium and potassium in various mantle samples (given in Table 9-7) show that eclogite xenoliths found in kimberlite are often less rich in these elements than peridotites found in the same pipe, although both may have been contaminated by the kimberlite matrix. These elements are, nevertheless, higher than in primitive mantle. The ²⁰⁶Pb/²⁰⁴Pb ratios of eclogites are generally low, implying long-term depletion.

At one time eclogite was considered a possible parent material for basalt. In recent years the consensus has swung

strongly to the view that partial (—20 percent) melting of garnet peridotite provides the basaltic magmas. Cosmochemical and seismic data have been used in support of the latter hypothesis. However, the differentiation of the Earth into crust, depleted and enriched reservoirs and, in particular, the evidence for a depleted MORB reservoir indicate that a chondritic source is not appropriate for all, if any, mantle reservoirs, even if the average composition of the mantle is chondritic. An eclogite cumulate layer is one possible product of a differentiated chondritic Earth and a possible source for basaltic magmas.

The other arguments for a peridotite source have to do with the amount of melting that is considered plausible. Melt-crystal separation can occur easily at relatively shal-

TABLE 9-6
Comparison of Kimberlite Eclogites and Some Other Rock Types

Oxide (percent)	Eclogite		Picrite (3)	MORB (4)	Ocean Crust (5)
	(1)	(2)			
SiO ₂	45.2	47.2	44.4	47.2	47.8
TiO ₂	0.5	0.6	1.18	0.7	0.6
	17.8	13.9	10.2	15.0	12.1
Cr ₂ O ₃	0.4	—	0.22	—	—
	—	—	—	3.4	—
FeO	11.2	11.0	10.92	6.6	9.1
MgO	13.1	14.3	18.6	10.5	17.8
MnO	0.3	—	0.17	0.1	—
CaO	9.6	10.1	9.7	11.4	11.2
Na ₂ O	1.6	1.6	1.37	2.3	1.3
K ₂ O	0.03	0.8	0.13	0.1	0.03

(1) Bobbejaan eclogite (Smyth and Caporuscio, 1984).

(2) Roberts Victor eclogite (Smyth and Caporuscio, 1984)

(3) Picrite, Svartenhuk (Clarke, 1970).

(4) Oceanic tholeiite (MORB).

(5) Average oceanic crust (Elthon, 1979).

TABLE 9-7
Heat-Producing Elements in Mantle Rocks (ppm)

Rocks	Uranium	Thorium	Potassium
Eclogites			
Roberts Victor	0.04–0.8	0.29–1.25	83–167
	0.04–0.05	1.3	1833
Zagadochnaya	0.024		
Jagersfontein	0.07	0.17	1000
Bultfontein	0.07	0.31	1300
MORB	0.014	0.035	660
Garnet peridotite	0.22	0.97	663
Ultramafic rock	0.26	0.094	35
Primitive Mantle	0.02	0.077	151

Note: Measured values of $^{206}\text{Pb}/^{204}\text{Pb}$ for eclogites are 15.6–19.1; for primitive mantle it is —17.5.

low depths in stationary magma chambers but is less likely at depth or in rising decompressing diapirs. In particular, partial melts at depth can be trapped if the overlying layer is subsolidus and impermeable or if the melt density increases rapidly with pressure. More likely Rayleigh-Taylor instabilities occur in the deeper layer, and large diapirs with entrained melt rise through the shallow mantle. Pressure-release melting then can cause extensive melting in the diapir. The instability occurs when melting in the deeper layer is extensive enough that, by the elimination of garnet, it becomes buoyant relative to the shallow mantle. The top of an eclogite-rich zone underlying a peridotite region will be in a thermal boundary layer and therefore 100–200°C warmer than the shallower mantle. A density reversal would set in for about 10–20 percent melting in eclogite if the overlying material is peridotite. We refer to a basalt or eclogite-rich rock as a piclogite.

PICLOGITE

Peridotites and pyrolite are rock types composed primarily of the refractory crystals olivine and orthopyroxene. Olivine is generally more than 60 percent, and both minerals together typically constitute more than 80 percent of the rock. Clinopyroxene and aluminous phases such as spinel, garnet and jadeite—the basalt assemblage—are minor constituents. At the other extreme are rocks such as clinopyroxenites and eclogites, which are low in orthopyroxene and olivine. Intermediate are rocks such as picrites, olivine eclogites and komatiites. Rocks that have less than 50 percent olivine plus orthopyroxene have been given the general name piclogite. In major elements they fall between dunites and basalts and between lherzolites and picrites. They contain a larger basaltic fraction than peridotites, although they contain the same minerals: olivine, orthopyroxene, diop-

side-jadeite and garnet. At high pressure they are denser than lherzolite, but at high temperature they can become less dense. Piclogites can represent garnet-rich cumulates or frozen high-pressure melts and can generate basaltic melts over a wide range of temperatures. They are not compositionally equivalent to basalts and do not require large degrees of partial melting to generate basaltic magmas. However, if piclogites are the source reservoir for mid-ocean-ridge basalts, the garnet must be mainly eliminated in order to satisfy the HREE data and in order to decrease the density of the reservoir so that it can rise into the shallow mantle. Therefore, clinopyroxene probably exceeds garnet in any reservoir with clinopyroxene rather than garnet as a near-liquidus phase.

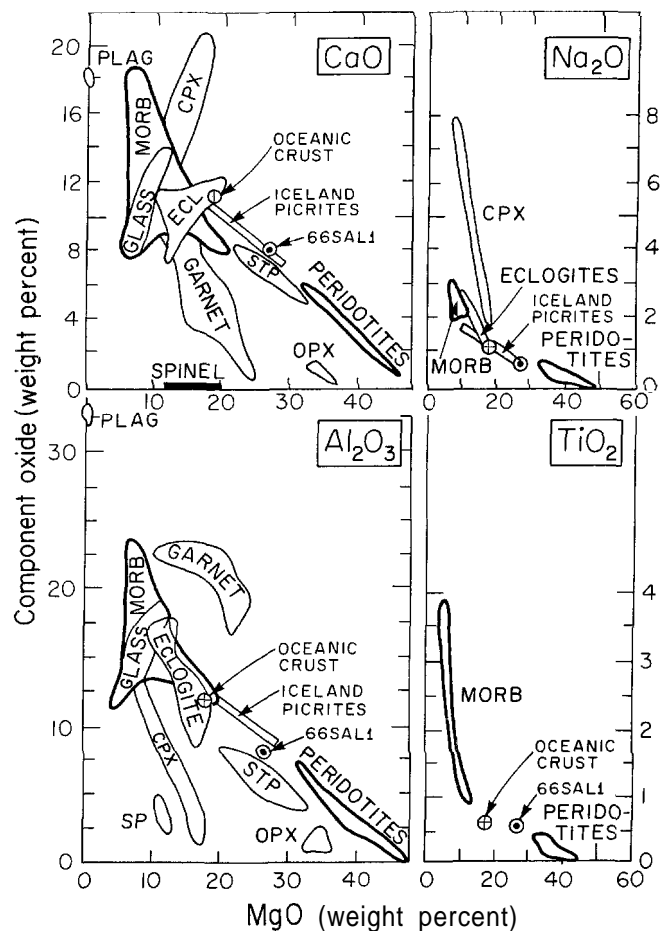
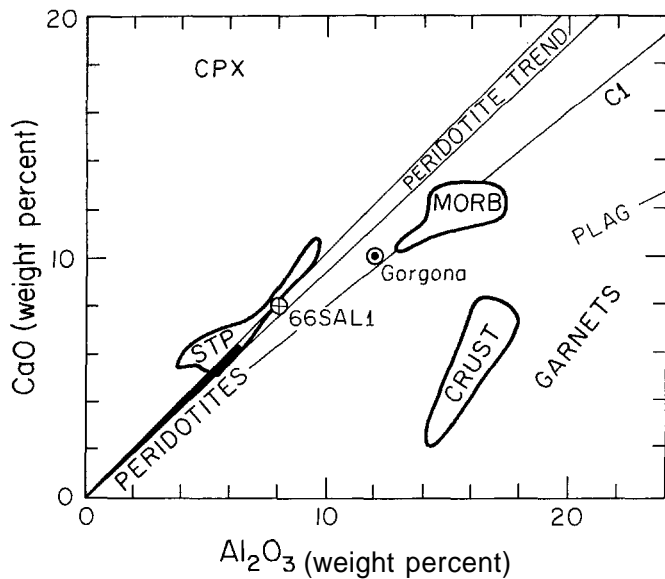


FIGURE 9-3
Major-element oxides of igneous rocks and rock-forming minerals. If peridotites represent residues after basalt extraction and MORB and picrites represent melts, and if these are genetically related, the MgO content of the basalt source is probably between about 20 and 30 percent, much less than most lherzolites or peridotites. 66 SAL-1 is a garnet pyroxenite from Salt Lake Crater, Hawaii, and falls in the composition gap between basalts and peridotites. It is a piclogite.


FIGURE 9-4

CaO versus Al_2O_3 for igneous rocks and minerals. Note that peridotites and STP are CaO-rich compared to C1 chondrites, and meteorites in general, while the continental crust is Al_2O_3 -rich. MORB falls on the C1 trend, possibly suggesting that it is an extensive melt of primitive mantle. Gorgona is an island in the Pacific that has komatiitic magmas. The solar CaO/ Al_2O_3 ratio is slightly higher than peridotites (Chapter 1).

Peridotitic komatiites (STP, for spinifex-texture peridotites) have major-element chemistries about midway between harzburgites and tholeiites and occupy the compositional gap between lherzolites and picrites (Figures 9-3 and 9-4). The present rarity of komatiitic magmas may be partially due to their high density and the thick lithosphere. Komatiites evolve to picrites and then to tholeiites as they crystallize and lose olivine, becoming less dense all the while. In a later chapter I show that the properties of the transition region (400–650 km depth) are consistent with less than 50 percent olivine and low orthopyroxene content, that is, piclogite.

Compositions of picrites, komatiites and a garnet pyroxenite (66 SAL-1) from Salt Lake Center, Hawaii are given in Table 9-8. These are all examples of rocks that fall between basalts and peridotites in their major-element chemistry. In terms of mineralogy at modest pressure they would be olivine eclogites or piclogites. They are all suitable source rocks for basalts.

DEPTHS OF MANTLE RESERVOIRS

Petrological data suggest that basalts separated from their parent material at depths of about 60 to 90 km. This is not necessarily the depth of the source region, although it is

generally assumed that melting initiates in the low-velocity zone and that this part of the mantle is in fact the source region. More likely, melting is caused by adiabatic ascent from greater depth, and the shallower depths are where cooling, fractionation and melt-crystal separation occur. Recent seismic results show that the low velocities under the East Pacific Rise extend to a depth of 400 km (Walck, 1984). Surface-wave tomographic results (Nataf and others, 1986) are consistent with this and indicate even greater depths under some ridge systems. In a convecting mantle, with ridge migration, the roots of midocean ridges are displaced relative to their surface expression, and this is shown clearly in the tomographic results. It is likely, therefore, that the ultimate source region of midocean-ridge basalts is in the transition region rather than in the low-velocity zone. The low-velocity zone is the location of final crystal-melt equilibration and MORB, or picrite, separation.

The V_p/V_s ratio, closely related to Poisson's ratio, generally increases with temperature, pressure and partial melting. The high V_p/V_s of the ridge structure, particularly between 90 and 150 km depth, suggests high temperature and, for the highest values, partial melting. The peculiar reversal of V_p/V_s under the shield LID at 150 km, associated with a decrease in K/ρ , can be explained by either a high orthopyroxene content or by anisotropy. In olivine aggregates the slow direction has low V_p/V_s and K/ρ . This zone has been sampled by kimberlites and seems to be predominantly highly deformed peridotite. This is probably the decoupling zone between the shield lithosphere and the underlying mantle. The rapid increase in velocity near 200 km implies a change in petrology.

In an accreting planet the upper mantle is above the solidus throughout most of the growth phase and the planet ends up zone refined, with melts and incompatible elements concentrated near the surface and dense, depleted residual crystals concentrated toward the interior (see Chapter 2). Even if a planet can avoid differentiation and gravitational separation during accretion, the relative slopes of the adiabat and the melting point assure that upper-mantle melting will occur in the rising limbs of convection cells. In olivine-rich peridotites it is unlikely that the liquidus is exceeded, but the separation of melt and crystals serves to remove the high-melting components so that a large part of the ancient upper mantle may actually have been completely molten. Freezing of the molten layer proceeds both from the top and the bottom, giving, on a Moon-size planet, plagioclase crystals that float to the surface and, in the case of an Earth-size planet, dense aluminous phases, such as garnet, which sink. In an accreting or crystallizing planet, refractory residual phases form the base of the system. Nearer the surface are layers of cumulate crystals, which represent the early crystallization products of the melt, and the late-stage residual melts, which contain most of the incompatible elements. The expulsion of late-stage intercumulus melts from deep cumulate layers is an alternative to small degrees of

TABLE 9-8
Compositions of Picrites, Komatiites and Garnet Pyroxenite

Oxide	Group II ¹		Komatiite		66SAL-1 ⁴
	Svartenhuk	Baffin Is.	S. Africa ²	Gorgona ³	
SiO ₂	44.4	45.1	45.7	45.3	44.8
TiO ₂	1.18	0.76	0.91	0.60	0.52
Al ₂ O ₃	10.2	10.8	7.74	10.6	8.21
Cr ₂ O ₃	0.22	0.27	—	—	—
FeO	10.92	10.26	—	10.9	9.77
MnO	0.17	0.18	0.23	0.18	—
MgO	18.6	19.7	19.43	21.9	26.53
CaO	9.7	9.2	8.04	9.25	8.12
Na ₂ O	1.37	1.04	0.61	1.04	0.89
K ₂ O	0.13	0.08	—	0.02	0.03
P ₂ O ₅	0.14	0.09	0.12	—	0.04
Fe ₂ O ₃	—	—	15.44	—	—
NiO	0.13	0.12	—	—	—
H ₂ O	2.33	2.14	3.46	—	—

¹Clarke (1970).

²Jahn and others (1980).

³Echeverria (1980).

⁴Frey (1980).

partial melting as a mechanism for depleting part of the mantle and forming the continental crust. In this scenario the concentration of volatiles and incompatible elements would decrease from the crust to the upper mantle to the transition region to the lower mantle, and each region would differ in major-element chemistry and mineralogy as well.

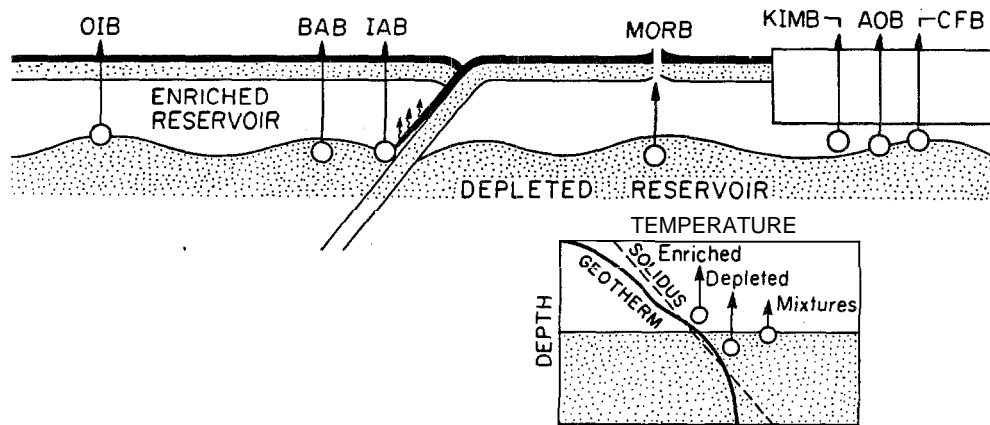
The origin of the oceanic crust and lithosphere is undergoing a dramatic paradigm shift. The conventional (pyrolite) theory is that midocean-ridge tholeiites are primary unmodified melts from the mantle, which form the oceanic crust, and the immediately underlying mantle is residual peridotite. The presumed parent is reconstructed by combining MORB and peridotite, and this is taken to be the composition of primitive mantle.

The average composition of the oceanic crust now appears, however, to be picritic, and tholeiites represent only the upper portions. The harzburgites underlying the oceanic crust in ophiolite sections appear not to be genetically related to the overlying basalts. The source region for MORB is not primitive but has experienced an ancient fractionation event that stripped it of its incompatible elements. A picritic magma is more refractory than tholeiite, and therefore the degree of partial melting and the required upper-mantle temperatures are much higher than previously thought. The source region for oceanic crust is remarkably homogeneous and has remained isolated from other parts of the mantle for a large fraction of the age of the Earth. This is known from isotopic ratios that show that this reservoir has been depleted in uranium, rubidium and samarium for over 10⁹ years. These observations all suggest that the MORB reservoir is large and deep, otherwise it would be quickly con-

taminated by subducted sediments, altered oceanic crust and the metasomatism that has affected the shallow mantle. The upper 200 km of the mantle is extremely heterogeneous, and enriched mantle appears to underlie both continents and oceanic islands. If the enriched source is deep, it must provide magmas that penetrate the depleted reservoir, without contaminating it, and emerge at such diverse regions as continental rifts and flood basalt provinces (CFB), oceanic islands (OIB) and island arcs (IAB). Figure 9-5 shows a possible configuration for the various reservoirs.

The source of the material that provides the distinctive trace-element and isotopic signature of hotspot magmas is controversial. In most discussions the source of energy and the source of material are not differentiated. It is possible that material rising from one depth level advects high temperatures to shallow levels, causing melting and the production of hotspot magmas from the material in the shallow mantle. It is not necessary that material melt itself in situ. In a stratified mantle, without transfer of material between layers, convection in a lower layer can control the location of hotspots in the overlying layer. In fact, the high temperature gradient in the thermal boundary layer between layers is the preferred source of magma genesis since the melting gradient is larger than the adiabatic gradient in a homogeneous convecting fluid (Figure 9-5). In a homogeneous mantle, if the melting point is not exceeded in the shallow mantle, then it is unlikely to be exceeded at greater depth since the melting point and the adiabat diverge with depth.

Modern midocean-ridge basalts are extremely depleted in the large-ion lithophile elements (LILE) such as rubidium,


FIGURE 9-5

A possible configuration of the enriched and depleted reservoirs; a continent, midocean ridge and subduction zone are shown schematically. In this model MORBs are relatively uncontaminated melts from a deep depleted reservoir, one that may become buoyant when hot or partially molten, and rise to the shallow mantle. Other basalts from the depleted reservoir are trapped by thick crust or lithosphere and interact with the enriched reservoir. Inset shows the effects of the thermal boundary between the reservoirs.

strontium, LREE and uranium. The low values of isotopic ratios such as $^{87}\text{Sr}/^{86}\text{Sr}$ show that the MORB source cannot be primitive; it must have been depleted in the LIL for at least 10^9 years. On the other hand the helium-3 in these basalts suggests the presence of a "primitive" component. The $^{206}\text{Pb}/^{204}\text{Pb}$ ratios are low compared to other basalts but still imply a high U/Pb compared to primitive or undifferentiated mantle. U/Pb should be decreased, just as Rb/Sr, in the processes that led to depletion of the MORB reservoir. These paradoxes can be understood if (1) partial melting processes are not 100 percent efficient in removing volatiles such as helium-3 from the mantle and (2) MORBs are hybrids or blends of magmas from a depleted reservoir and an enriched reservoir. In a chemically stratified mantle one reservoir is of necessity deeper than the other, and mixing or contamination upon ascent seems inevitable. Even with this mixing it is clear from the isotopes that there are at least two ancient magma reservoirs in the mantle, reservoirs that have had an independent history for much of the age of the Earth.

Although island-arc magmas and oceanic-island magmas differ in their petrography and inferred pre-eruption crystal fractionation, they overlap almost completely in the trace elements and isotopic signatures that are characteristic of the source region. It is almost certain that island-arc magmas originate above the descending slab, although it is uncertain as to the relative contribution of the slab and the intervening mantle wedge. The consensus is that water from the descending slab may be important, but most of the material is from the overlying "normal" mantle. It is curious that the essentially identical range of geochemical signa-

tures found in oceanic-island and continental flood basalts have generated a completely different set of hypotheses, generally involving a "primitive" lower-mantle source. If island-arc magmas originate in the shallowest mantle, it is likely that island and continental basalts do as well. Subduction may, in fact, be the primary cause of shallow mantle enrichment.

The main diagnostic differentiating some islands from most arcs is the high $^3\text{He}/^4\text{He}$ ratio of the former. High ratios, greater than atmospheric, are referred to as "primitive," meaning that the helium-3 content is left over from the accretion of the Earth and that the mantle is not fully outgassed. The word "primitive" has introduced semantic problems since high $^3\text{He}/^4\text{He}$ ratios are assumed to mean that the part of the mantle sampled has not experienced partial melting. This is not correct since it assumes that (1) helium is strongly concentrated into a melt, (2) the melt is able to efficiently outgas, (3) mantle that has experienced partial melting contains no helium and (4) helium is not retained in or recycled back into the mantle once it has been in a melt. As a matter of fact it is difficult to outgas a melt. Magmas must rise to relatively shallow depths before they vesiculate, and even under these circumstances gases are trapped in rapidly quenched glass. Although most gases and other volatiles are probably concentrated into the upper mantle by magmatic processes, only a fraction manages to get close enough to the surface to outgas and enter the atmosphere. High helium-3 contents relative to helium-4, however, require that the gases evolved in a relatively low-uranium-thorium reservoir for most of the age of the Earth. Shallow depleted reservoirs may be the traps for helium.

MANTLE METASOMATISM AND THE ENRICHED RESERVOIR

Kimberlites and alkali basalts are high in potassium, titanium, volatile components and large-ion-lithophile (LIL) elements. Several mechanisms have been proposed for the generation of highly enriched basalts:

1. They could represent small degrees of melting (2 percent or less) of mantle peridotite.
2. If the peridotite is already enriched in the LIL, then a larger degree of melting is allowed, and separation of very small melt fractions is not necessary. The enrichment event, however, must involve strongly fractionated fluids (high Rb/Sr, La/Sm and so on), and these must be the result of small degrees of partial melting or be a fractionated vapor phase.
3. The enriched fluid component could be the late-stage fluids, the last crystallizing fraction of a cumulate layer.
4. The enriched component could be introduced into the mantle by the subduction of sediments or altered oceanic crust. This cannot be the whole story since many enriched magmas from oceanic islands have high $^3\text{He}/^4\text{He}$ ratios.

The isotopic ratios of some enriched magmas indicate that the enriched component is ancient, although it could have been introduced into the magma, or the source region, relatively recently. A precursory metasomatic event is a currently popular hypothesis, but the source of the metasomatic fluid is seldom addressed. This mechanism is hard to distinguish from a mechanism that involves contamination or hybridization of a rising depleted magma with melts from an enriched source region.

Mantle-derived xenoliths from depths shallower than 170 km show evidence that titanium, alkalis, volatiles and LIL elements were introduced into the mantle, converting peridotite to assemblages bearing phlogopite, amphibole and clinopyroxene. These regions cannot be the source of LIL-depleted MORB.

Neodymium and strontium isotopic data indicate that significant isotopic heterogeneity occurs in the source regions of kimberlites and alkali basalts or that these basalts represent variable degrees of mixing between an enriched and a depleted magma. The combination of variable degrees of mixing and subsequent crystal fractionation can give suites of magmas that may not appear to be simple mixtures. The residual or cumulate assemblages will also display LIL and isotopic heterogeneity.

Metasomatism introduces ingredients such as iron, titanium and H_2O , which lower the solidus temperature, and also introduces the heat-producing elements potassium, uranium and thorium. Retention of interstitial fluids by a crys-

tallizing cumulate also introduces phases that are rich in these elements and that will yield them up early in a subsequent melting episode. These phases will probably be concentrated in the shallow mantle. Enriched material will also be concentrated in the shallow mantle by subduction.

The time sequence of eruption of alkali and tholeiitic basalts suggests that the enriched source is shallower than the depleted source. Extremely depleted MORB only occurs at mature continental and oceanic rifts, implying that the earlier basalts are contaminated. The shallow mantle may provide the "metasomatic" trace-element and isotopic signature, and the deeper source may provide the bulk of the magma.

The refractory rocks contained in magmas and kimberlites and exposed in mountain belts are harzburgites, dunites, orthopyroxenites and lherzolites, with harzburgites constituting a large fraction of the volume. Many of these, particularly the alpine peridotites, are depleted in the large-ion-lithophile elements, making them both infertile and depleted, as expected for refractory residues or cumulates. Most, however, are enriched in the incompatible elements, suggesting that they have been invaded by the upward migration of silicate melts or water-rich fluids. These melts introduce a kimberlite-like component into an otherwise refractory rock. Ultramafic rock masses exhibit a wide range of isotopic ratios, indicating a variety of ages, including ancient, for the enrichment events. Basalts passing through such peridotites pick up these LIL and isotopic signatures.

Table 9-9 shows that trace-element-enriched peridotites have about the same patterns for some elements as kimberlite. The enriched (Q) component of both basalts and peridotites is similar to kimberlite.

RISE OF DEEP DIAPIRS

Melt extraction and crystal settling are important processes in igneous petrology. Basic melts apparently separate from magma chambers, or rising diapirs, at depths as great as 90 km. The basic law governing crystal settling, Stokes' Law, expresses a balance between gravitational and viscous forces,

$$V = \frac{2g\alpha^2\Delta\rho}{9\eta}$$

where a is the radius of a spherical particle, $\Delta\rho$ is the solid-liquid density contrast, η is viscosity of the melt and V is the terminal settling velocity. The same equation can be applied to the rise of diapirs through a mantle with Newtonian viscosity, with modifications to take into account nonspherical objects and non-Newtonian viscosity. Additional complications are introduced by turbulence in the magma chamber and the possibility of a finite yield strength of the magma. Both serve to keep small particles in suspension.

TABLE 9-9
Trace-Element Contents of Enriched and Depleted Peridotites and Kimberlite

	Peridotites		Kimberlite
	Enriched*	Depleted	
Ba	345	2.3	1000
Sr	132	0.5	740
Rb	19	0.05	65
La	7	0.18	150
Ce	20	0.69	200
Nd	11	0.56	85
Nb	16	0.4	110
Sc	10	14	15
Sm	1.9	0.27	13
Yb	0.42	0.37	1.2
Th	0.55	0.25	11.4
Ta	1.13	0.03	9
Hf	0.75	0.15	7

* RSI Mica parquise Iherzolite, Germany (Menzies and others, 1987).

Diapirs are usually treated as isolated spheres or cylinders rising adiabatically through a static mantle. Because of the relative slopes of the adiabat and the melting curve, diapirs become more molten as they rise. At some point, because of the increased viscosity or decreased density contrast, ascent is slowed and cooling, crystallization and crystal settling can occur. The lithosphere serves as a viscosity or strength barrier, and the crust serves as a density barrier. Melt separation can therefore be expected to occur in magma chambers at shallow depths.

In a convecting mantle the actual temperatures (adiabatic) diverge from the melting point as depth increases. Melting can therefore only occur in the upper parts of the rising limbs of convection cells or in thermal boundary layers. The additional buoyancy provided by melting contributes to the buoyancy of the ascending limbs. Although melts will attempt to rise relative to the adjacent solid matrix, they are embedded in a system that is itself rising and melting further. If broad-scale vertical convection is fast enough, diapirs can melt extensively without fractionating. The stresses and temperatures in the vicinity of rising plumes or diapirs are high, and these serve to decrease the mantle viscosity; thus rapid ascent is possible. In order to achieve observed equilibration temperatures and large degrees of partial melting, allowing for specific and latent heats, melting must initiate at depths of at least 200 km. The solidus temperature at this depth is at least 1900 K. The questions then are how fast does material rise between about 200 and 90 km, and is the material at 90 km representative of the deeper mantle source region or has it been fractionated upon ascent?

Viscosities in silicates are very stress- and temperature-dependent, and diapirs occur in regions of the mantle

that have higher than normal stresses and temperatures. Diapiric emplacement itself is a high-stress process and occurs in regions where mantle convection may have oriented crystals along flow lines. Diapirs may rise rapidly through such low-viscosity material. I have estimated that a 50-km partially molten diapir at a depth of 200 km could rise at a rate of about 40 cm/s (Anderson, 1981). Kimberlites travel an order of magnitude faster still. Crystal settling velocities are of the order of 10^{-2} cm/s. It appears therefore that deep diapirs can rise rapidly enough to entrain both melt and crystals. If the magma behaves as a Bingham fluid with interacting crystals and a finite yield strength or if the magma is turbulent, the Stokes settling velocities are upper bounds. Crystals in excess of 10 cm can be held in suspension with measured values of yield strength of basaltic melts.

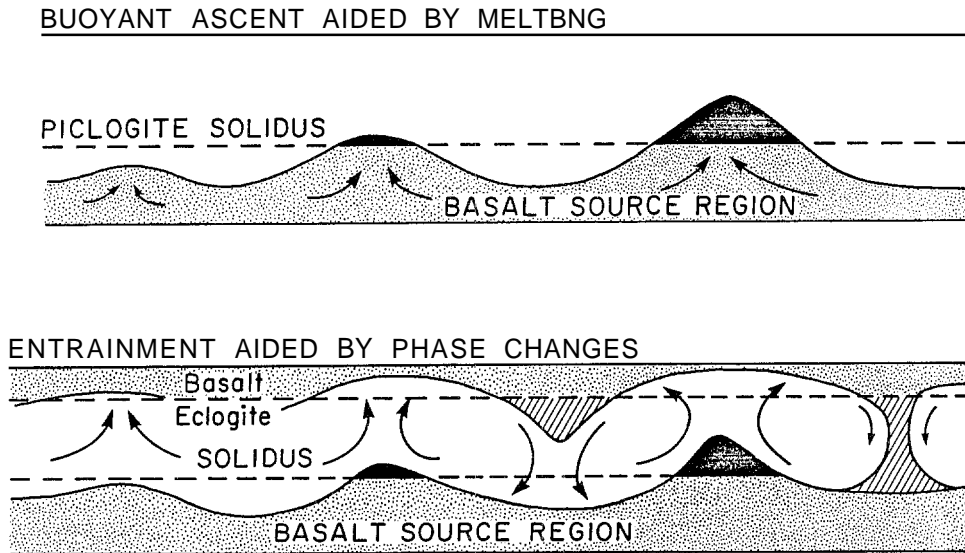
In addition, the density contrast between melt and crystals, $\Delta\rho$, is very small at great depth because of the change in composition and the high compressibility of the melt. At depth the melt content and the permeability are also low, and melt segregation is therefore very slow. The importance of crystal settling is not yet clear even in shallow, static cooling magma chambers. The situation is far more complex in a rising, melting and perhaps turbulent diapir.

In a chemically stratified mantle, for example residual peridotite over piclogite or fertile peridotite, there is a conductive thermal boundary between the convecting layers. In such a region the thermal gradient is in excess of the melting gradient, and melting is most likely to initiate at this depth. Partial melting in the deeper, denser layer causes a reduction in density, and a Rayleigh-Taylor instability can develop. Material can be lifted out of the piclogite layer by such a mechanism and extensive melting occurs during ascent to the shallower mantle. If the peridotite is hydrous, it can also melt and magma mixing is likely, particularly if the diapir is trapped beneath thick lithosphere. Unimpeded ascent of a picritic diapir to the surface yields relatively unfractionated, uncontaminated picritic-tholeiitic melts, as at midocean ridges. In midplate environments, such as Hawaii and other midplate hotspots, a similar melt will cool, fractionate and mix prior to eruption. Such a mechanism seems capable of explaining the difference between hotspot (ocean island, continental flood) basalts and MORB.

Material can leave a deep, dense source region by several mechanisms;

1. Melting in the thermal boundary at the bottom or the top of the layer because of the high thermal gradient in boundary layers compared to melting point gradients.
2. Melting, or phase changes, due to adiabatic ascent of hotter regions of layer and crossing of phase boundary.
3. Entrainment of material by adjacent convecting layer.

Some of these mechanisms are illustrated in Figure 9-6.

**FIGURE 9-6**

Methods for removing material from a deep, dense source region. In a layered mantle the interface between chemically distinct layers deforms due to convection and thermal expansion. This may induce phase changes, or partial melting, in the deeper layer, causing it to become buoyant. Material from the deeper layer may also be entrained.

TRACE-ELEMENT MODELING

The trace-element contents of basalts contain information about the composition, mineralogy and depth of their source regions.

When a solid partially melts, the various elements composing the solid are distributed between the melt and the remaining crystalline phases. The equation for equilibrium partial melting is simply a statement of mass balance:

$$C_m = \frac{C_o}{D(1 - F) + F} = \frac{C_r}{D}$$

where C_m , C_o and C_r are the concentrations of the element in the melt, the original solid and the residual solid, respectively; D is the bulk distribution coefficient for the minerals left in the residue, and F is the fraction of melting. Each element has its own D that depends not on the initial mineralogy but on the residual minerals, and, in some cases, the bulk composition of the melt. For the very incompatible elements ($D \ll 1$), essentially all of the element goes into the first melt that forms. The so-called compatible elements ($D \gg 1$) stay in the crystalline residue, and the solid residual is similar in composition to the original unmelted material. The above equation is for equilibrium partial melting, also called batch melting. The reverse is equilibrium crystal fractionation, in which a melt crystallizes and the crystals remain in equilibrium with the melt. The same equations apply to both these situations. The effective par-

tion coefficient is a weighted average of the mineral partition coefficients.

The Rayleigh fractionation law

$$C_m/C_o = F^{(D-1)}$$

applies to the case of instantaneous equilibrium precipitation of an infinitesimally small amount of crystal, which immediately settles out of the melt and is removed from further equilibration with the evolving melt. The reverse situation is called fractional fusion.

Nickel and cobalt are affected by olivine and orthopyroxene fractionation since D is much greater than 1 for these minerals. These are called compatible elements. Vanadium and scandium are moderately compatible. Rare-earth patterns are particularly diagnostic of the extent of garnet involvement in the residual solid or in the fractionating crystals. Note the high partition coefficients for the heavy REE, yttrium and scandium for garnet (Table 9-10). Melts in equilibrium with garnet would be expected to be low in these elements. MORB is high in these elements; kimberlite is relatively low.

Melt-crystal equilibrium is likely to be much more complicated than the above equations imply. As semi-molten diapirs or melt packets rise in the mantle, the equilibrium phase assemblages change. The actual composition of a melt depends on its entire previous history, including shallow crystal fractionation. Melts on the wings of a magma chamber may represent smaller degrees of partial

TABLE 9-10
Crystal/Melt Partition Coefficients

	cpx	opx	ol	gt	plag	plag
La	0.02	0.0005	0.0005	0.004	—	0.02–0.11
Ce	0.04	0.0009	0.0008	0.021	0.10	0.02–0.11
Nd	0.09	0.0019	0.0013	0.087	0.05	0.02–0.07
Sm	0.14	0.0028	0.0019	0.217	0.05	0.02–0.04
Eu	0.16	0.0036	0.0019	0.320	1.50	0.14–0.39
Tb	0.19	0.0059	0.0019	0.70	—	—
Ho	0.195	0.0089	0.0020	1.4	—	—
Yb	0.20	0.0286	0.0040	4.03	0.04	0.006–0.03
Lu	0.19	0.038	0.0048	5.7	—	0.006–0.037
Sc	3.1	1.1	0.25	6.5	—	—
V	1.5	0.3	0.09	0.27	—	—
Co	1.2	2	1.3–6.5	2	—	—
Ni	2–4	3–5	3.8–35	0.8	—	—
Rb	0.05	0.02	0.01	0.02	0.04	0.019–0.36
Sr	0.165	0.016	0.016	0.014	1–4.4	1.2
Y	0.20	0.009	0.002	1.4	—	—
Ba	0.001	0.001	0.0001	0.0015	0.15–0.4	—
	0.08	0.01	0.01	0.04	0.35	—
U	0.04	0.006	0.0025	0.04	—	—
K	0.002	0.001	0.0002	0.001	0.1	0.18
He	—	—	<0.07	—	—	—
Ne	0.5	—	0.006–0.08	—	—	—
Ar	0.7	—	0.05–0.15	—	—	—
Kr	0.14	—	<0.15	—	—	—
Xe	<0.5	—	<0.3	—	—	—
Estimated Rock/Melt Partition Coefficients						
	Th	0.0004		P	0.015	
	Ta	0.001		Hf	0.015	
	Nb	0.0015		Zr	0.015	

Frey and others (1978), Wedepohl and Muramatsu (1979), Chen and Frey (1985), Hiyagon and Ozima (1986).

melting than those on the axis. Melts representing a variety of melting conditions may mix prior to eruption. It is therefore difficult to state that a melt represents a given degree of partial melting from a given depth. Nevertheless, extremely fractionated and enriched melts must, at some point in their history, be the result of very small degrees of partial melting or large degrees of crystal fractionation.

Solid trace elements form ionic or covalent bonds with the surrounding ions forming the crystalline lattice. Solid elements can be substituted into a lattice if they have similar ionic charge and radius as the ions in the lattice. For a given crystal the partition coefficient is a strong function of ionic radius; garnet and clinopyroxene, for example, are able to retain elements with ionic radii close to Al and Ca while olivine and orthopyroxene are not.

Noble gases are likely to be accommodated in a crystal as neutral ions, and therefore the mechanism of noble gas retention is quite different from that for solid elements. Neutral ions are likely to be trapped in dislocations, grain boundaries, vacancies and interstitial sites. These ions are

subject to the electric field of surrounding ions and undergo electric polarization. The heavier noble gases have a large electronic polarizability and may be held more readily in a lattice, in spite of their larger size. Measured partition coefficients between olivine and basalt melt are $D(\text{He}) \leq 0.07$, $D(\text{Ne}) = 0.006\text{--}0.08$, $D(\text{Ar}) = 0.05\text{--}0.15$, $D(\text{Kr}) \leq 0.15$ and $D(\text{Xe}) \leq 0.3$ (Hiyagon and Ozima, 1986). The partition coefficients for diopside are up to an order of magnitude higher, although data are lacking for helium. No data are available for garnet, which might be an important trap for helium in the upper mantle. Note that these partition coefficients for the rare gases are larger than for some of the incompatible solid trace elements.

The so-called incompatible elements are those that have low (< -0.1) crystal-liquid partition coefficients in olivine and orthopyroxene. Most of the elements listed in Table 9-10 are incompatible elements. Nickel and cobalt are compatible elements, being retained by the major mantle minerals ($D > 1$). Scandium is a compatible element for all pyroxenes and garnet and moderately incompatible for ol-

ivine. Note that the heavy rare-earth elements (HREE) are compatible in garnet and moderately compatible in clinopyroxene. This is why eclogite fractionation, or retention of garnet in the source rock, is effective in fractionating the REE pattern.

References

- Anderson, D. L. (1981) Rise of deep diapirs, *Geology*, **9**, 7–9.
- Basu, A. R. and Tatsumoto, M. (1982) Nd isotopes in kimberlites and mantle evolution, *Terra Cogn.*, **2**, 214.
- Beus, A. A. (1976) *Geochemistry of the Lithosphere*, MIR Publications, Moscow, 366 pp.
- Bowen, N. L. (1928) *The Evolution of the Igneous Rocks*, Princeton University Press, 332 pp.
- Boyd, F. R. (1986) In *Mantle Xenoliths* (P. Nixon, ed.) Wiley, New York.
- Boyd, F. R. and S. A. Mertzman (1987) In *Magmatic Processes* (B. O. Mysen, ed.) The Geochemical Society, University Park, Pennsylvania, 500 pp.
- Carmichael, I. S. E., F. Turner and J. Verhoogen (1974) *Igneous Petrology*, McGraw-Hill, New York, 739 pp.
- Chen, C. and F. A. Frey (1983) Origin of Hawaiian theiite and alkali basalt, *Nature*, **302**, 785.
- Clarke, D. B. (1970) Tertiary basalts of Baffin Bay; possible primary magma from the mantle, *Contrib. Mineral. Petrol.* **25**, 203–224.
- Echeverria, L. M. (1980) *Tertiary Komatiites of Gorgona Island*, Colombia, Carnegie Instn. Wash. Ybk., **79**, 340–344.
- Elthon, D. (1979) High magnesia liquids as the parental magma for ocean floor basalts, *Nature*, **278**, 514–518.
- Frey, F. A. (1980) The origin of pyroxenites and garnet pyroxenites from Salt Crater, Oahu, Hawaii, trace element evidence, *Am. J. Sci.*, **280-A**, 427–449.
- Frey, F. A., D. H. Green and S. D. Roy (1978) Integrated models of basalts petrogenesis: A study of quartz tholeiites to olivine melilitites from southeastern Australia utilizing geochemical and experimental petrological data, *J. Petrol.*, **19**, 463–513.
- Green, D. H. (1971) Composition of basaltic magmas as indicators of conditions of origin; application to oceanic volcanism, *Phil. Trans. R. Soc.*, **A268**, 707–725.
- Green, D. H., W. O. Hibberson and A. L. Jaques (1979) Petrogenesis of mid-ocean ridge basalts. In *The Earth: Its Origin, Structure and Evolution* (M. W. McElhinny, ed.), 265–295, Academic Press, New York.
- Green, D. H. and A. E. Ringwood (1963) Mineral assemblages in a model mantle composition, *J. Geophys. Res.*, **68**, 937–945.
- Green, D. H. and A. E. Ringwood (1967) The genesis of basaltic magmas, *Contrib. Mineral. Petrol.*, **15**, 103–190.
- Hauy, R. J. (1822) *Minéralogie*, **4**, 548.
- Herzberg, C. T. (1984) Chemical stratification in the silicate Earth, *Earth Planet. Sci. Lett.*, **67**, 249–260.
- Herzberg, C. T., (1986). *Eos, Trans. Am. Geophys. Union* **67**, 408.
- Hiyagon, H. and M. Ozima (1986) *Geochim. Cosmochim. Acta*, **50**, 2045.
- Jaques, A. and D. Green (1980) Anhydrous melting of peridotite at 0–15 Kb pressure and the genesis of tholeiitic basalts, *Contrib. Mineral. Petrol.* **73**, 287–310.
- Maaløe, S. and K. Aoki (1977) The major element composition of the upper mantle estimated from the composition of Iherzolites, *Contrib. Mineral. Petrol.* **63**, 161–173.
- Menzies, M., N. Rogers, A. Zindle and C. Hawkesworth (1987) In *Mantle Metasomatism* (M. A. Menzies, ed.) Academic Press, New York.
- Nataf, H.-C., I. Nakanishi and D. L. Anderson (1986) Measurements of mantle wave velocities and inversion for lateral heterogeneities and anisotropy, Part III, Inversion, *J. Geophys. Res.*, **91**, 7261–7307.
- O'Hara, M. J. (1968) The bearing of phase equilibria studies in synthetic and natural systems on the origin and evolution of basic and ultrabasic rocks, *Earth Sci. Rev.*, **4**, 69–133.
- Popper, K. (1972) *Conjectures and Refutations*, 4th ed., Routledge and Kegan Paul, London.
- Rigden, S. S., T. J. Ahrens and E. M. Stolper (1984) Densities of liquid silicates at high pressures, *Science*, **226**, 1071–1074.
- Ringwood, A. E. (1962) Mineralogical constitution of the deep mantle, *J. Geophys. Res.*, **67**, 4005–4010.
- Ringwood, A. E. (1966) Mineralogy of the mantle. In *Advances in Earth Science*, 357–399, MIT Press, Cambridge, Mass.
- Ringwood, A. E. (1975) *Composition and Petrology of the Earth's Mantle*, McGraw-Hill, New York, 618 pp.
- Ringwood, A. E. (1979) *Origin of the Earth and Moon*, Springer-Verlag, New York, 295 pp.
- Smyth, J. R. and F. A. Caporuscio (1984) Petrology of a suite of eclogite inclusions from the Bobbejaan Kimberlite; II, Primary phase compositions and origin. In *Kimberlites* (J. Kornprobst, ed.), 121–131, Elsevier, Amsterdam.
- Smyth, J. R., T. C. McCormick and F. A. Caporuscio (1984) Petrology of a suite of eclogite inclusions from the Bobbejaan Kimberlite; I, Two unusual corundum-bearing kyanite eclogites. In *Kimberlites* (J. Kornprobst, ed.), 109–119, Elsevier, Amsterdam.
- Stolper, E. (1980) *Contrib. Mineral. Petrol.* **74**, 13.
- Walck, M. C. (1984) The P-wave upper mantle structure beneath an active spreading center; the Gulf of California, *Geophys. J. Roy. Astron. Soc.*, **76**, 697–723.
- Wedepohl, K. H. and Y. Muramatsu (1979) The chemical composition of kimberlites compared with the average composition of three basaltic magma types. In *Kimberlites, Diatremes, and Diamonds* (F. R. Boyd and H. O. Meyer, eds.), 300–312, American Geophysical Union, Washington, D.C.
- Yoder, H. S., Jr. (1976) *Generation of Basaltic Magma*, National Academy of Science, Washington, D.C., 265 pp.
- Zindler, A. and E. Jagoutz, *Geochim. Cosmochim. Acta* (in press).