

Generation of Magma

FUNDAMENTAL QUESTIONS CONSIDERED IN THIS CHAPTER

1. What is the role of the mantle in global magma generation?
2. How and where can solid rock be melted to generate magma?
3. How does the mineralogical and chemical composition of solid rocks in the upper mantle and lower continental crust dictate the composition of magmas generated by partial melting in these source regions?
4. How do partial melting conditions in source regions influence the composition of the generated magmas?

INTRODUCTION

The basic problems in understanding magmatic rocks are how, where, and why they are created and what accounts for their wide range of compositions, as exemplified in any variation diagram (e.g., Figure 2.4).

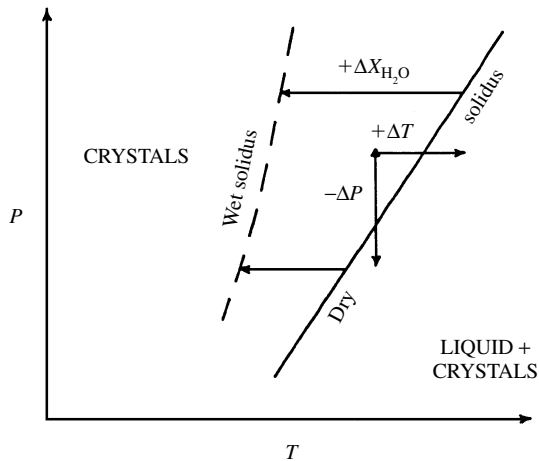
Clearly, active volcanoes manifest the occurrence of magma in the Earth. Basaltic lavas extruded from volcanoes have $T \sim 1200^\circ\text{C}$, whereas rhyolitic magmas commonly erupt at $900\text{--}700^\circ\text{C}$. These temperatures imply that magma has to be created at depths of *at least* the lower crust (45–35 km) or upper mantle (60 km), assuming a uniform geothermal gradient of 20°C and negligible cooling during ascent. But we know that the crust and mantle are generally solid rock all the way to the outer liquid metallic core because they transmit

seismic shear waves. So what are the anomalous conditions that explain how and why silicate magma is generated from solid silicate rock in the mantle and possibly the lower crust? Somehow, already hot but subsolidus rock must be perturbed to cause melting and magma generation.

Magma generation must be closely linked with global tectonics because active volcanoes occur where lithospheric plates are diverging and converging and where mantle plumes are actively rising, as under Hawaii (Figure 1.5 and Plate I). No active volcanoes occur in the central part of North America or Australia or in other huge areas around the globe.

* 11.1 MELTING OF SOLID ROCK: CHANGES IN P , T , AND X

Melting of a solid is conventionally associated with an increase in T ($+\Delta T$). However, phase diagrams in Chapter 5 show that melting can also occur at virtually constant T (isothermally) by decompression ($-\Delta P$) and by influx of volatiles ($+\Delta X_{\text{volatiles}}$) into already hot rock. In all three changes in state of the system, the rock already contains a significant store of thermal energy and becomes a melting **source rock** as $+\Delta T$, $-\Delta P$, and $+\Delta X_{\text{volatiles}}$ perturbations move the system above the solidus (Figure 11.1). Total melting of a source rock to its liquidus requires large, and geologically unrealistic, changes in these three intensive variables. Before total melting might occur, either the buoyant melt separates and moves out of the source or the partially melted rock becomes sufficiently buoyant to rise en masse, perhaps as a diapir, out of the source region and away from the perturbing changes.



11.1 Perturbations in P , T , or X_{water} move a subsolidus rock to above the solidus, causing partial melting. Potential source rock just below solidus represented by solid triangle. Influx of water, $+\Delta X_{\text{H}_2\text{O}}$, depresses the solidus to a “wet” solidus position (not water-saturated), placing the rock in the stability field of liquid + crystals. Perturbations of $-\Delta P$ and $+\Delta T$ move potential source rock to above the solidus.

As the causes of these perturbations are examined next it should be kept in mind that they may not act independently. For example, increase in T may be accompanied by an increase in volatile concentration where mantle-derived basalt magma intruded into the lower continental crust heats it and adds volatiles exsolved during crystallization. Addition of volatiles lowers melting temperatures, enhancing the effect of heating in melt production.

11.1.1 Temperature Increase, $+\Delta T$

An increase in T sufficient to melt solid rock in the Earth can occur in several ways.

Mass Movement of Rock or Magma. Heat transfer associated with convective or advective movement of rock or magma is an important means of raising rock temperatures above the solidus in large rock volumes. Two global plate tectonic regimes where this takes place include (Figure 11.2):

1. The descending oceanic lithosphere in subduction zones absorbs heat from the surrounding hotter mantle. Consequently, partial melting of the less refractory basaltic crust might occur in young, still hot subducting lithosphere, as contrasted with subduction of old, cold lithosphere. Shear heating of rapidly subducted lithosphere can also promote higher temperatures (discussed later).
2. Already hot, deep continental crust can be heated in excess of its solidus T by juxtaposition of hotter mantle-derived magma above subducting plates and upwelling decompressing mantle. Density constraints indicate that ascending basaltic magma can

be arrested at the base of the less-dense feldspathic crust, **underplating** it, or perhaps stagnating not far into it above the Moho (Section 9.1.1). The thermal budget for melting crust can be roughly approximated by considering deep continental felsic rock at a depth of 30 km on an average geothermal gradient of 20°C and therefore at $T = 600^\circ\text{C}$. To raise the rock T to its solidus of, say, 900°C and assuming the rock specific heat is 1.4 J/g deg requires $(1.4 \text{ J/g deg}) \times (900 - 600) \text{ deg} = 420 \text{ J/g}$ (equation 1.4). Melting 1 g of rock requires an additional 300 J/g, its approximate latent heat of melting. Thermal energy to accomplish this heating and melting totals about 720 J/g and can be supplied by less than 1 g of basalt magma initially at 1300°C . This assumes that the latent heat of crystallization of the basalt magma supplies about 420 J/g and cooling to 900°C supplies about $(1.4 \text{ J/g deg}) \times (1300 - 900) \text{ deg} = 560 \text{ J/g}$. An approximate rule of thumb is that the mass of continental felsic rock melted is about the same as the mass of basaltic magma heating it. Melting consumes relatively large amounts of thermal energy, moderating increasing T within the source rock.

Mechanical Work of Shearing. Work can be transformed into thermal energy, Q , according to

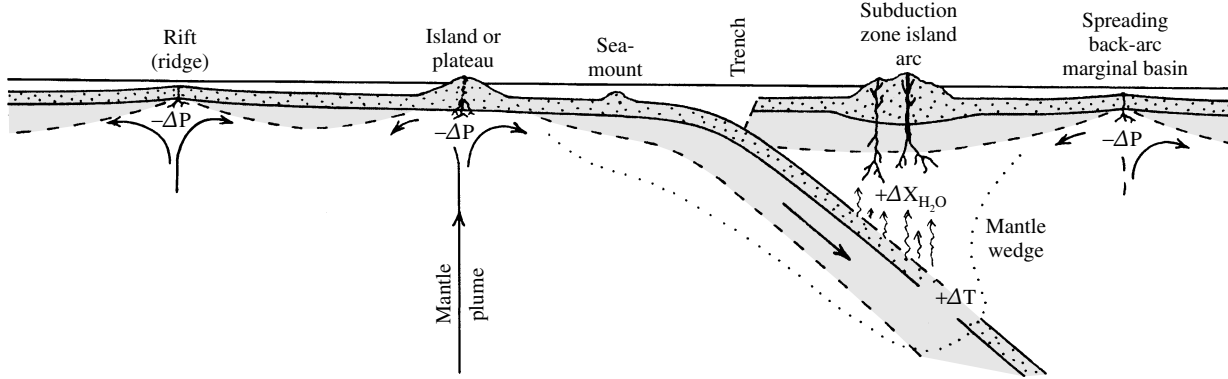
$$11.1 \quad Q = \tau \left(\frac{d\varepsilon}{dt} \right)$$

where τ is the shear stress required for threshold deformation and $d\varepsilon/dt$ is the strain rate (Section 8.2). For a shear stress of 50 MPa and a representative geologic strain rate of $5 \times 10^{-14}/\text{s}$, such as occurs during shortening in a mountain belt or convective mantle flow, $Q = 50 \times 10^6 \text{ Pa} [(kg/m^2)/Pa] \times [J/(kg m^2/s^2)] \times 5 \times 10^{-14}/\text{s} = 2.5 \times 10^{-6} \text{ J/m}^3 \text{ s}$. For 1 g of granite whose density is $2.7 \times 10^6 \text{ g/m}^3$, $Q = 6.8 \times 10^{-13} \text{ J/g s} \sim 2 \times 10^{-5} \text{ J/g y}$. This heating is so small it would require 36 My to melt granite initially at 600°C (see previous paragraph), assuming the stress is maintained and no heat is dissipated from the system.

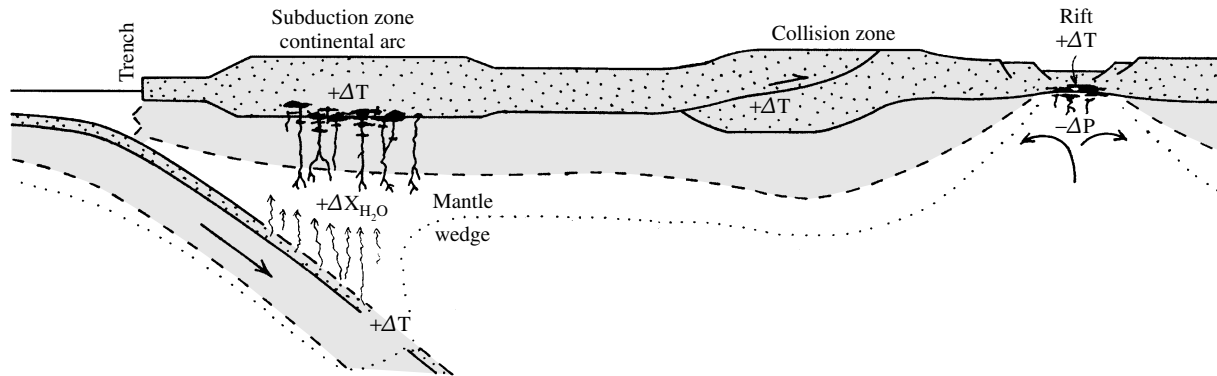
However, through a **thermomechanical feedback** phenomenon, shear heating might be sufficient to induce melting. Thus, shearing might become concentrated in certain zones, localizing more thermal energy production; the consequently higher T reduces the apparent viscosity, further localizing shear deformation, causing more heating and reduction of viscosity, and so on, until the melting T is reached. Such shear melting would tend to be localized in small volumes.

Local shear melting can occur at meteorite impact sites and along fault zones where strain rates are much higher.

OCEANIC LITHOSPHERE



CONTINENTAL LITHOSPHERE



- 11.2 Geologically plausible plate tectonic settings at which partial melts may be generated from otherwise solid rock by perturbations in P , T , or $X_{volatiles}$. Compare Figure 1.5. Lithosphere is shaded; oceanic and continental crust is stippled. Scale is grossly distorted. Decompression ($-\Delta P$) of rising solid mantle beneath oceanic ridge, continental rift, and in mantle plume partially melts peridotite. In subduction zones, water released (wavy arrows) by dehydration of hydrated oceanic crust capping subducting oceanic lithosphere rises into overlying mantle wedge; resulting $+\Delta X_{water}$ in peridotite causes partial melting as solidus T is depressed. Locally, young, hot, hydrated oceanic crust partially melts as a result of $+\Delta T$ caused by absorbed heat from surrounding hot asthenosphere. Ascending mafic magmas from volatile-fluxed mantle wedge underplate continental crust or rise slightly above the Moho, where they stagnate (black blobs), transfer heat to the already hot lower crust, and cause $+\Delta T$ and partial melting. A similar $+\Delta T$ occurs in a continental rift where basaltic magmas are derived from upwelling mantle underplate crust. Where continental crust is greatly thickened in collision zones, such as the present-day Himalayan, $+\Delta T$ results from processes such as radioactive heating, shearing, and adjustment to the geothermal gradient. Dotted line is highly schematic isotherm perturbed to shallower depths by rising mantle (heat source) and depressed to greater depth in the colder subducting lithosphere (heat sink).

Decay of Radioactive Isotopes. Thermal energy is generated by slow decay of K, Th, U, and other less abundant radiogenic isotopes in rocks. However, the rate of heat production is very low (granite about 3.4×10^{-5} J/g y, basalt 5.0×10^{-6} J/g y, peridotite 3.8×10^{-8} J/g y), so that, again, as with distributed shear heating, tens of millions of years would be necessary to raise the T of deep rock to its solidus, even if the heat produced did not leak out of the body in that time. Additionally, the first melting episode would extract much of the incompatible K, Th, and U from the source rock, so that subsequent heat generation would be greatly retarded. Therefore, radioactive decay *alone*, in most instances, cannot produce sufficient perturbation in T to cause magma generation.

Tectonic Thickening. Thrusting and folding thicken continental crust to >50 km in active orogenic (mountain) belts in subduction zones (Figure 11.2). This can lead to supersolidus temperatures in the deep crust. Several factors are intertwined in complex ways in this heating. Radioactive heat production in the thickened continental crust is enhanced, as is the insulating effect on the flux of heat from the mantle. The 600°C temperature at the base of a 30-km-thick crust is subsolidus for most potential source rocks, but, after thickening to, say, 50 km and readjustment to the same average geotherm of $20^\circ\text{C}/\text{km}$, the T at the base of the crust is 1000°C , which is well above the solidus of many crustal rocks. Optimal “incubation” depends on the rate of thickening and may be curtailed by prompt

Special Interest Box 11.1 Highly localized melting of rock independent of tectonic setting

Heat generation by mechanical friction accompanying movement at very high strain rates ($>10^{-2}/s$) along fault zones is locally sufficient to cause melting of rock (equation 11.1). Sheets of melted rock are commonly associated with cataclastic fabric (Figure 8.7) in shear zones because cataclasis is an essential precursor to frictional melting. The sheets are generally <1 cm thick. Rapid heat loss quenches the melt sheet to a vein of typically black glass called **pseudotachylite**. (Tachylite is a term once used for basaltic glass; however, not all shear-generated glass is basaltic.)

Locally extensive melting occurs at meteorite impacts where strain rates are $>10^6/s$; high-pressure silica polymorphs—coesite and stishovite—are also



11.3 Dark-colored pseudotachylite surrounding blocks of unmelted granite at McCreedy West in the north range of the 250-km-diameter, 1.85 Ga Sudbury, Ontario, impact structure. After the impacting meteorite excavated the crater in Archean granite and granite gneiss, faulting accompanying rebound and gravitational collapse generated pseudotachylite melts by comminution and frictional heating at high strain rates along the fault walls. (Photograph and caption courtesy of John G. Spray.)

created by the transient shock. Larger podlike or dike-like bodies of pseudotachylite as much as 1 km thick (Figure 11.3) are produced by the extreme comminution of fault walls during rebound and gravitational collapse of the transient impact cavity (Spray, 1998).

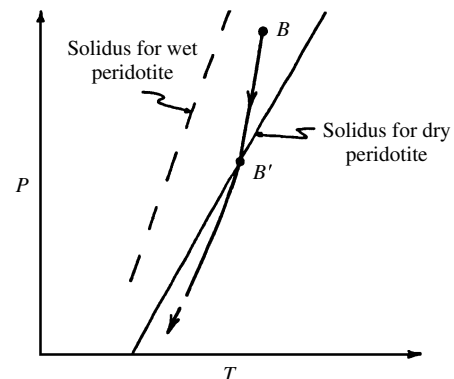
Many older pseudotachylites are devitrified to microcrystalline or cryptocrystalline aggregates. Devitrification textures can resemble those occurring in volcanogenic glass, for example, spherulitic texture. Fragments of unmelted rock are common within the glass; partially dissolved crystals are embayed. Flow layers, vesicles, and magmatic crystallization textures are locally evident.

An extremely rare and bizarre means of locally melting rock and unconsolidated sediment occurs through lightning strikes, producing **fulgurite**. These are irregularly shaped crusts or tubular glassy structures that can be as much as 40 cm long and 6 cm in diameter.

rapid uplift and erosion. Thermal modeling that takes into account these and other factors suggests optimal heating occurs on the order of a few tens of millions of years after orogenic thickening (e.g., Patiño Douce et al., 1990).

11.1.2 Decompression, $-\Delta P$

Because of the positive slope of the solidus curve of dry silicate systems in P - T space, adiabatic decompression of hot near-solidus rock can induce melting. Consider a solid parcel of mantle initially below the volatile-free solidus at B in Figure 11.4. As the parcel rises to shallower depth, it decompresses and cools adiabatically at a rate of about $0.3^\circ\text{C}/\text{km}$, or $1^\circ\text{C}/\text{kbar}$, along the path BB' , whose slope is steeper than the dry solidus. At B' , melting begins and the thermal energy required for the latent heat of fusion is drawn from the thermal energy



11.4 Schematic decompression melting. Note that decreasing depth is *downward* in diagram.

stored in the body. Hence, during continued decompression, the upwelling mantle cools at a more rapid rate at pressures less than B' than at pressures greater than B' . Thus, the P - T path of the ascending mantle closely follows the solidus. As the solidus has a slope of about $10^\circ\text{C}/\text{kbar}$, a decrease in P of 1 kbar (over about 3 km) results in an increment of T (corrected for adiabatic cooling) $\Delta T = 10 - 1 = 9^\circ\text{C}$. Multiplied by a specific heat, C_P of 1.4 J/g deg , this gives (equation 1.4) 12.6 J/g , sufficient to melt about 3% of rock with a latent heat of fusion of 420 J/g (McBirney, 1993, p. 243).

Special Interest Box 11.2 Decompression melting: belated acceptance of a two-century-old idea

Almost three-fourths of the Earth is covered by rocks of basaltic composition; most of these rocks constitute the oceanic crust that is several kilometers thick. Most basaltic magmas are generated by decompression melting, which therefore is the dominant magma-generating mechanism in the planet. Decompression melting in upwelling mantle beneath the globe-encircling system of oceanic spreading ridges and in ascending mantle plumes is accepted today as one of the most fundamental concepts in petrology. Surprisingly, acceptance was slow (Sigardsson in Sigardsson et al., 2000).

John Playfair in 1802 followed by George P. Scrope in 1825 speculated that a change in pressure might generate magma within the Earth. In the early nineteenth century, the pioneering work of Carnot, Clausius, and Clapeyron in the infant science of thermodynamics resulted in an exact expression for the change of the melting T of a solid substance with respect to change in P —the Clapeyron equation (3.13), $dT/dP = \Delta V/\Delta S$. Nicholas Desmarest had recognized in 1763 that the prismatic joint columns in the Claremont area of France were of volcanic origin, and subsequent observers correctly inferred that they formed by contraction during cooling. Thus, because a volume (and entropy) increase accompanied melting, the Clapeyron equation predicted an increase in the melting T of rock with increasing depth in the Earth. Hence, decompression of hot solid rock could lead to melting.

As early as 1881, Osmond Fisher proposed that convection currents in a molten interior (as then believed) ascended beneath the oceans and descended beneath continents. After the discovery of radioactivity near the turn of the 20th century, Arthur Holmes was one of the first to apply it for dating of minerals. Also being aware of the heat produced by radioactive decay, Holmes embraced convection as

a means to cool the Earth and in so doing proposed in 1928 a global mechanism remarkably like the modern paradigm of plate tectonics. But convection in mantle known to be solid, like Wegener's proposal of continental drift in 1912, were insights too preposterous at the time to be widely accepted.

Without an alternative mechanism to relieve P in the interior of the hot Earth, petrologists in the first half of the twentieth century either equivocated or dismissed decompression melting as a viable process. As late as 1960, in their classic *Igneous and Metamorphic Petrology*, Francis J. Turner and John Verhoogen concluded that (p. 446) "Whether convection does occur in the mantle is not definitely known, nor is it known whether it could be effective in the upper part of the mantle where magmas are generated. Thus, although convection might lead to melting, it cannot be shown that it does, and the problem of generation of [basaltic] magma remains as baffling as ever."

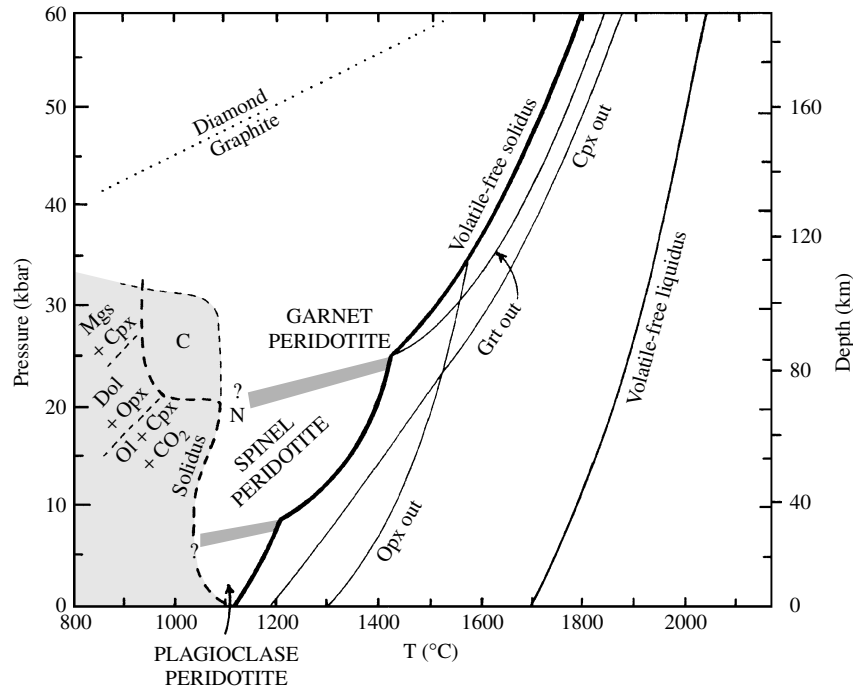
During the Cold War (1950s to 1970s), the U.S. government generously funded Earth scientists in order to learn more about the oceans. The result was the serendipitous discovery of seafloor spreading and plate tectonics that in turn lead to the confirmation of mantle convection and the opportunity for decompression of mantle peridotite to generate basaltic magma.

Decompression partial melting of mantle rock is the principal mechanism by which huge global volumes of basalt magma are generated at ocean ridges and above mantle plumes (Figures 1.1, 1.5, 11.2).

Some geologists believe that decompression of uplifting continental rock can cause partial melting, but it is likely of minor importance.

11.1.3 Changes in Water Concentration, $+\Delta X_{\text{water}}$

Among the possible changes in chemical composition of a solid rock system that could induce melting on a global scale, increase in water concentration (or pressure), $+\Delta X_{\text{water}}$ ($+\Delta P_{\text{water}}$), is the most significant. Very locally, an increase in CO_2 or other volatiles may induce melting in already hot rock. Even small increases in water concentration can significantly depress the solidus of silicate systems (Figures 5.2, 5.11, 5.29). This perturbation is most significant in subduction zones (Figure 11.2), where the descending wet oceanic crust dehydrates, liberating water that advects into the overlying wedge of peridotite mantle. In the crust, water (and lesser amounts of CO_2) resides in pore spaces in sediment and in cracks in thicker basaltic rock. More significantly, basaltic rock has been variably metamorphosed into hydrous mineral assemblages by hy-



11.5 Phase relations in mantle peridotite. Volatile-free phase relations for spinel lherzolite from Kilborne Hole, New Mexico. (Sample KLB-1; redrawn from Takahashi et al., 1993.) Subsolidus assemblage of garnet peridotite is garnet (Grt) + olivine (Ol) + orthopyroxene (Opx) + clinopyroxene (Cpx), for spinel peridotite is Spl + Ol + Opx + Cpx, and for plagioclase peridotite is Pl + Ol + Opx + Cpx. Note subsolidus transitions between the three peridotite facies are “smeared” (shaded bands) over a range of T and P because of solid solution. Thin-line curves denoting complete dissolution of Opx and Cpx into the melt with increasing T correspond to approximately 40% melting of the lherzolite. On the left, short dashed lines show phase relations of volatile-bearing peridotite with 0.3 wt.% H_2O plus <5 wt.% $MgCO_3$ (Mgs, magnesite) or $CaMg(CO_3)_2$ (Dol, dolomite) beneath its solidus (heavy dashed line). Shaded area indicates where amphibole is stable. N, approximate P - T region for nephelinite partial melt in spinel peridotite. Between about 20 and 30 kbar a Mg-Ca-Na carbonatite melt is stable with an amphibole peridotite above the solidus in the field labeled C. (Redrawn from Wallace and Green, 1988.)

drothermal activity at spreading ridges. As these hydrous minerals are heated in the descending crust to depths of hundreds of kilometers, sequential dehydration reactions (Section 5.7.2, Figure 5.31) release water.

In the lower continental crust, dehydration mineral reactions accompanying metamorphism might provide water to induce partial melting. Underplating mantle-derived magmas might also give off water and CO_2 as they crystallize.

* 11.2 MANTLE SOURCE ROCK

Most magma generated in the Earth is basaltic. Its source is decompressing peridotitic mantle upwelling beneath global ocean ridges and to a lesser extent in ascending mantle plumes (Figures 1.5 and 11.2). Basaltic magmas in subduction zones are generated in the mantle wedge overlying the subducting lithosphere. In most continental tectonic regimes, significant proportions of heat as well as mass for magma generation are also directly or indirectly derived from the mantle. In the young Archean Earth, a large proportion of the silic continents was created ultimately from partial

melts extracted out of the mantle. Hence: *The ultimate source of thermal energy and mass for production of magma in the Earth is the mantle.* Understanding the nature of the mantle source rock is therefore critically important.

Cosmological models relate the composition of the whole Earth to chondrite meteorites from which the Earth is presumed to have accreted at about 4.5 Ga. Their composition, on a volatile-free basis, is (Taylor and McLennan, 1985) as follows:

SiO_2	34.2 wt. %
TiO_2	0.11
Al_2O_3	2.44
FeO	35.8
MgO	23.7
CaO	1.89
Na_2O	0.98
K_2O	0.10
Total	99.2

Chondrites also contain water and carbon. Most of the Fe in the chondritic Earth segregated into its core shortly after or during accretion.

Measured seismic velocities in the upper mantle are compatible with a rock made of olivine, pyroxene, and garnet. These three phases are composed essentially of the five major chondritic chemical components: SiO_2 , Al_2O_3 , FeO , MgO , and CaO . Upper mantle densities constrained by isostatic calculations are about 3.35 g/cm^3 (increasing slightly with depth, Figure 1.3), which is more consistent with peridotite (about 3.3 g/cm^3) than with olivine-free, pyroxene-garnet rock (eclogite, 3.5 g/cm^3), although bodies of the latter might occur in minor amounts within the former.

Peridotite is an ultramafic rock made of Mg-rich olivine and lesser amounts of pyroxene, usually both Ca-Mg-rich clinopyroxene and Ca-poor, Mg-rich orthopyroxene; these three crystalline phases are a stable assemblage to a depth of about 410 km. Most peridotites contain more Al in the bulk rock than can be held in solid solution in pyroxenes and olivine, thus stabilizing a separate minor Al-rich phase, whose nature depends on P and less on T . At less than about 8 kbar (30-km depth), the stable Al-rich phase is plagioclase; from there to about 25 kbar (roughly 75 km, depending on crustal thickness), it is spinel; and at still higher P it is garnet (Figure 11.5). Transitions between these three peridotite assemblages are smeared over a range of P (depth) and T because the minerals are solid solutions. Garnetiferous peridotites equilibrated at depths greater than about 150 km and occurring as xenoliths in some kimberlitic rock also contain diamond. Diamonds commonly contain minute inclusions of minerals stable in the upper mantle (pyroxene, garnet). Extremely rare diamonds contain inclusions stable only at >670 km-depth in the deep mantle (Mg-Fe-Ca-Al perovskite, Figure 1.3), testifying to the depth from which at least some kimberlitic magmas carrying suspended diamonds are derived. At shallower depths where diamond is not stable, the stable C-bearing phase may be graphite if reducing conditions prevail or, if more oxidizing, a carbonate mineral or CO_2 found in fluid inclusions in mantle minerals.

Some petrologic information on the uppermost mantle is from dredged samples from upfaulted segments of the seafloor and from large tracts of *alpine* peridotite or peridotite *massifs* exposed in mountain belts. The bodies of peridotite are components of ophiolite—segments of oceanic lithosphere tectonically emplaced and exposed in mountain belts adjacent to oceanic trenches (see Section 13.6). Oceanic rocks commonly suffer overprinting metamorphic effects that have erased their primary character to varying degrees. However, ophiolites provide key stratigraphic and structural information that dredge samples and inclusions, described next, cannot.

11.2.1 Mantle-Derived Inclusions

Important compositional information on the mantle is provided by phaneritic **mantle-derived inclusions** of

peridotitic rock, usually hosted in alkaline basaltic and kimberlitic rock in thousands of localities worldwide (e.g., Nixon, 1987). These typically fresh and unaltered dense inclusions, also called **nodules**, or **xenoliths**, are fragments of rock entrained from near the source of the host magma or plucked from shallower conduit walls and lifted to the surface during rapid ascent. Mineral geobarometers indicate equilibration in the upper mantle. Inclusions are of two types, classified according to their clinopyroxene: Cr-rich diopside or Al-Fe-Ti-rich clinopyroxene.

Cr-Diopside Peridotite. The most common mantle-derived inclusion consists mostly of olivine with subordinate pyroxenes and smaller amounts of either spinel (Mg, Fe^{2+})(Cr, Al, Fe^{3+}) $_2\text{O}_4$ or pyrope garnet (Mg, Fe^{2+} , Ca) $_3$ (Al, Cr) $_2\text{Si}_3\text{O}_{12}$. The clinopyroxene is essentially diopside ($\text{CaMgSi}_2\text{O}_6$) that contains only a small weight percentage of alumina and Fe-oxide, but because of a few tenths of weight percentage of Cr_2O_3 is characteristically an emerald-green color (Plate V). Although constituting $<20\%$ and generally $<5\%$ of these inclusions, this distinctive green Cr-diopside serves as a characterizing mineral. Olivines and orthopyroxenes contain about 90 mol% of the Mg end member. Some of these peridotites contain less than 1% or so of hydrous minerals, including phlogopite and pargasite or richterite amphibole (Appendix A). Modal proportions of olivine and pyroxenes classify most inclusions as Iherzolite and harzburgite; fewer are dunite. A metamorphic fabric resulting from textural equilibration has created 120° triple-grain junctions (Section 6.4). Some inclusions show subtle, more pyroxene-rich layers. Some show a late imprint of shear deformation manifested in strained olivines that have *deformation bands* of contrasting optical extinction orientation in polarized light in thin sections.

Spinel-bearing peridotite inclusions are hosted almost exclusively in alkaline basalt, basanite, and nephelinite and very rarely in andesitic rocks in subduction zones. Host magmas are characteristically volatile-rich, are commonly explosive, and tend to ascend rapidly from their upper mantle source. The absence of inclusions in far more widespread tholeiitic basalts possibly stems from slower ascent to the surface and the more evolved nature of these magmas, which involves fractional crystallization of more primitive parents in crustal storage chambers; both factors inhibit lifting dense inclusions to the surface and promote their assimilation. Spinel peridotite inclusions are mostly of latest Cenozoic age. Whether from oceanic or continental regimes, they are not significantly different (McDonough, 1990), indicating that the uppermost mantle in these two global areas is similar.

Garnet peridotite inclusions derived from greater depths in the mantle are somewhat different chemically from spinel peridotites (Table 11.1). They are hosted in

Table 11.1 Average Worldwide Composition of Garnet Peridotite (Harzburgite) and Spinel Peridotite (Lherzolite) Inclusions

	GARNET		SPINEL					
SiO ₂	45.00	44.22	Li	1.5	Y	4.4	Mo	0.050
TiO ₂	0.08	0.09	B	0.53	Zr	21	Ru	0.0124
Al ₂ O ₃	1.31	2.28	C	110	Nb	4.8	Pd	0.0039
Cr ₂ O ₃	0.38	0.39	F	88	Ba	33	Ag	0.0068
FeO _t	6.97	8.47	S	157	La	2.60	Cd	0.041
MnO	0.13	0.14	Cl	53	Ce	6.29	In	0.012
MgO	44.86	41.60	Sc	12.2	Pr	0.56	Sn	0.054
NiO	0.29	0.27	V	56	Nd	2.67	Sb	0.0039
CaO	0.77	2.16	Cr	2690	Sm	0.47	Te	0.011
Na ₂ O	0.09	0.24	Co	112	Eu	0.16	Cs	0.010
K ₂ O	0.10	0.054	Ni	2160	Gd	0.60	W	0.0072
P ₂ O ₅	0.01	0.056	Cu	11	Tb	0.070	Re	0.00013
Total	100.00	99.97	Zn	65	Dy	0.51	Os	0.004
			Ga	2.4	Ho	0.12	Ir	0.0037
Mg-value*	92.0	89.8	Ge	0.96	Er	0.30	Pt	0.007
Olivine	68	62	As	0.11	Tm	0.038	Au	0.00065
Opx	25	24	Se	0.041	Yb	0.26	Tl	0.0012
Cpx	2	12	Br	0.01	Lu	0.043	Pb	0.00016
Spinel	—	2	Rb	1.9	Hf	0.27	Bi	0.0017
Garnet	5	—	Sr	49	Ta	0.40	U	0.00012

Major element and calculated modal composition in the first three columns. Trace element composition (in ppm) of spinel peridotite in last six columns.

*Mg-value = 100[MgO/(MgO + FeO)] molecular ratio.

Data from McDonough (1990).

more alkaline, more silica-undersaturated magmas, including kimberlites and lamproites (Sections 11.5 and 13.12.2). These same host magmas also locally contain xenoliths of **eclogite**, a rock of basaltic chemical composition that consists of a high-*P* assemblage of mostly green clinopyroxene solid solution (jadeite-rich omphacite) and purple-red Mg-rich (pyropic) garnet (Appendix A). These xenoliths could be derived from basaltic melts generated and entrapped in the mantle and crystallized to eclogite or from remnants of subducted, recrystallized oceanic crust.

A Brief Digression on Volatiles in Mantle Minerals. In the mid-20th century, geologists began to accept the notion that Cr-diopside peridotite inclusions represent samples of the upper mantle. But as this mantle rock contains no volatile-bearing minerals, as then thought, Oxburgh (1964) realized that generation of typically volatile-rich alkaline basalt and kimberlite magmas from this source would be impossible. Further difficulties arise because of the very low concentrations of K, Na, and many other incompatible elements in peridotite (Table 11.1); only by very small degrees of partial melting could such magmas be generated. Obviously, a source of volatiles and incompatible elements was needed in the upper mantle. Alternatively, these Cr-diopside peridotite inclusions do not represent the source of such magmas.

Soon after Oxburgh's seminal paper, geologists began seeing previously overlooked or ignored amphibole, mica, apatite, and other hydrous and volatile-bearing minerals in mantle-derived inclusions. In addition to water, amphiboles and micas contain several tenths of 1 wt.% F and about an order of magnitude less Cl. And these minerals contain substantial amounts of the missing K, Rb, and other incompatible elements. Sulfur is sequestered in sulfides that occur in some inclusions. Diamonds and micas contain hundreds of parts per million of N. Virtually pure CO₂-fluid inclusions occur in mantle olivines. Infrared spectroscopy reveals that nominally anhydrous olivine, pyroxene, and garnet in upper mantle peridotite actually contain measurable amounts of structurally bound water as (OH)⁻ or H⁺ (Bell and Rossman, 1992). Garnets generally have <60 ppm water but may have as much as 200 ppm, pyroxenes mostly 100–600 ppm, and olivines 100 ppm, increasing with depth.

The oxygen fugacity of the upper mantle is commonly appropriate to stabilize carbonate minerals, such as magnesite, MgCO₃, and dolomite CaMg(CO₃)₂. During even the most rapid decompression, as in ascending explosive kimberlite magma, carbonate minerals quickly decompose with release of CO₂; consequently, inclusions generally lack this mineral and some that did have suffered disaggregation so that associated

silicate grains have been dispersed as xenocrysts in the kimberlite.

Mantle volatiles could originate from two sources. Juvenile volatiles were derived from the primeval chondritic material from which the planet formed at 4.5 Ga. Volatiles liberated from subducting oceanic crust are introduced into the overlying mantle and perhaps, by some means, distributed more widely through the mantle. It is estimated that six times more water is introduced into the mantle over subducting oceanic crust than is delivered to the surface of the Earth in subduction-zone volcanism (Thompson, 1992).

Al-Fe-Ti-Rich Clinopyroxene Inclusions. Though not as abundant as Cr-rich diopside inclusions, Al-Fe-Ti-rich clinopyroxene inclusions are nonetheless widespread. They are especially significant in containing substantial amounts of volatile-bearing minerals enriched in incompatible elements (K, Rb, Ti, C, H, etc.). Some consist exclusively of volatile phases. These inclusions are petrographically more variable than Cr-diopside inclusions, into which they may locally grade, but are dominated by clinopyroxenes rich in Al, Fe, and Ti; because of their black, conchoidally fractured aspect in hand samples, these clinopyroxenes resemble obsidian. Modal proportions range widely among the following constituent minerals: clinopyroxene, olivine, orthopyroxene, high-Al/Cr spinel, magnetite, ilmenite, rutile, zircon, plagioclase (in low-*P* inclusions), carbonate minerals, Fe-sulfides, apatite, amphibole, and phlogopite. Note that the latter five contain volatiles and the latter three are hydrous. Wehrlites, olivine clinopyroxenites, and pyroxenites are common rock types (Figure 2.10b), together with mica- and amphibole-bearing varieties. Olivines are enriched in Fe and orthopyroxenes in Al, Fe, and Ti relative to these phases in Cr-diopside inclusions. Textures of inclusions are also variable, but magmatic ones are common, such as amphibole poikilitically enclosing other phases. Preservation of magmatic textures in inclusions suggests crystallization from melts not long before entrainment into the host magma; otherwise textural equilibration and development of metamorphic fabric would be expected to occur at the high *T* prevailing in the mantle.

Discrete **megacrysts**, up to several centimeters in diameter, of clinopyroxene, olivine, orthopyroxene, high-Al/Cr spinel, magnetite, ilmenite, rutile, zircon, amphibole, phlogopite, and plagioclase can occur with or without accompanying xenoliths in alkaline mafic host rocks.

Significantly, Al-Fe-Ti-rich clinopyroxene assemblages are locally found as veins, to as much as several centimeters thick (Plate VI and Figure 11.6a), in Cr-rich diopside peridotite inclusions (e.g., Wilshire et al., 1988). Mineral barometers indicate that some veined inclusions are derived from depths of at least 170 km in

the mantle. Thicker veins may be the source of discrete megacrysts and xenoliths of the Al-Fe-Ti-rich clinopyroxene assemblage. Many veins are remarkably planar, indicating that the mantle host rock fractured in a brittle manner, probably hydraulically (Section 8.2.1), then the liquid hydraulic agent was emplaced into the fracture, forming the vein. Similar veins occur in large subaerial exposures of peridotite in ophiolite. Multiple generations of veins, one or more cutting earlier ones, are locally evident in both ophiolite and inclusions. Wall rock of Cr-diopside peridotite adjacent to a vein can be modified chemically and mineralogically for distances up to as much as several centimeters.

11.2.2 Metasomatized and Enriched Mantle Rock

Subsolidus modification of the chemical composition of a rock through the agency of invasive percolating liquids is called **metasomatism**. Rock volume can remain constant during this open-system replacive metasomatism. Two types may be defined: In **cryptic metasomatism** original solid-solution minerals remain but are changed in composition; for example, original clinopyroxenes may be made more Fe-rich, but subtle chemical changes may not be obvious, hence the designation *cryptic* (hidden) *metasomatism*. In **modal metasomatism** original minerals are replaced by entirely new minerals; an olivine-orthopyroxene assemblage (harzburgite) may be modally metasomatized to a new mineral assemblage of amphibole and clinopyroxene. Obviously, these two types of metasomatism can develop simultaneously in the same volume of rock; some original solid solution mineral compositions can be modified and other unstable minerals are replaced by new ones.

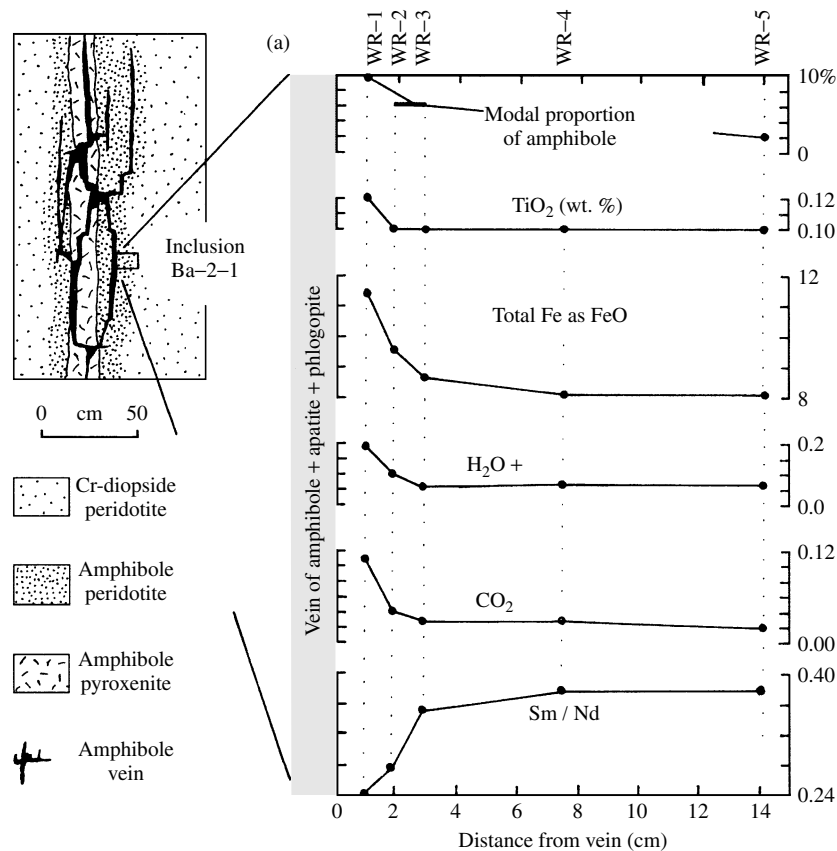
Metasomatism has been studied for at least a century in crustal metamorphic terranes and around ore deposits. However, Bailey (1970) first realized that xenoliths of Al-Fe-Ti-rich assemblages in alkaline volcanic rocks are records of mantle metasomatism and that they have a significant bearing on the generation of alkaline magmas. The rationale is that some highly alkaline mafic and ultramafic magmas cannot readily be derived from “normal” mantle of anhydrous, incompatible element-poor Cr-diopside peridotite. This sterile, or infertile, rock has to be metasomatically **enriched** in incompatible elements (Figure 11.6) before it can be a viable source of alkaline magmas that are enriched in such elements (see Section 11.5).

Enrichment details depend on the nature of the liquid, composition of the wall rock, and partition coefficients between the metasomatizing liquid and the minerals in the wall rock. Volatile-bearing minerals created by modal metasomatism—chiefly phlogopite, amphibole, apatite—harbor most of the incompatible elements—K, Ti, Al, Rb, Ba, Sr, H, F, Cl, and light rare earth elements. In their absence, incompatible elements are sequestered in clinopyroxene.

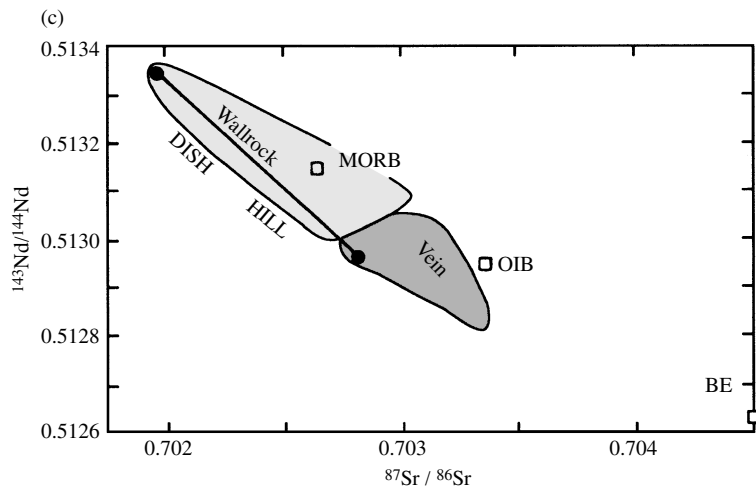
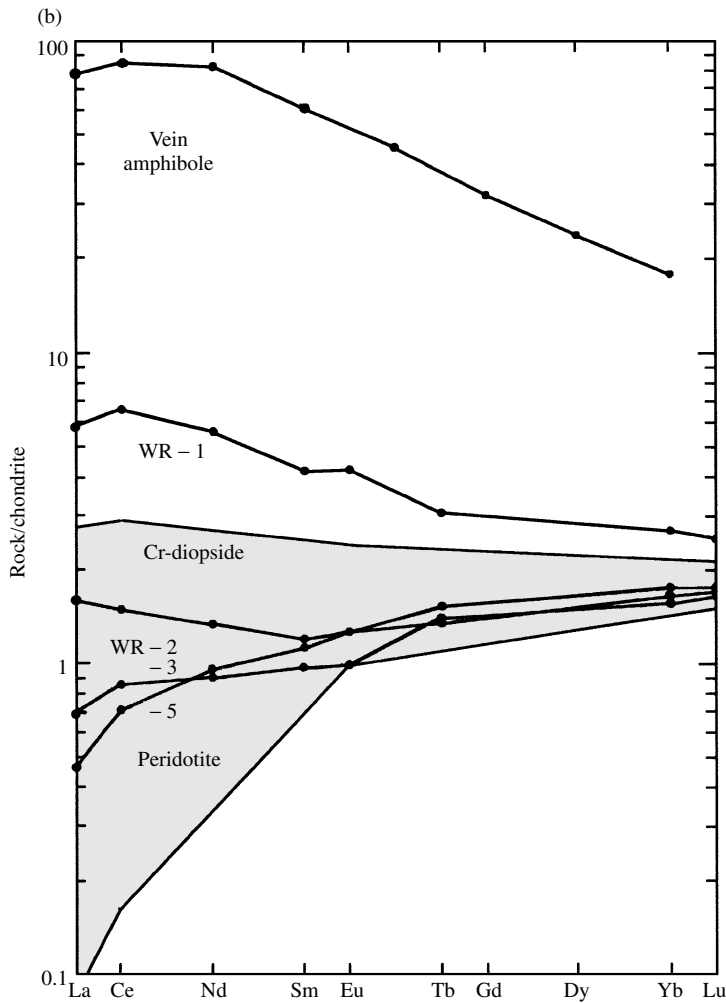
Possible metasomatizing liquids include volatile-rich silicate melts, carbonatite melts, and C-O-H fluids (see Section 4.2.1 for the distinction between fluid and melt). In most metasomatized mantle rocks it is uncertain whether a fluid or a melt was involved; in some cases it might have been both, or one type of liquid might have evolved into another. Some metasomatizing silicate melts are likely of Ti-K-rich alkaline basalt composition, judging from amphibole veins of roughly the same composition. Intraplate spinel peridotite xenoliths from the Kerguelen Islands in the southern Indian Ocean contain minute inclusions (a few tens of micrometers in size) in olivines and pyroxenes that consist of coexisting CO₂ fluid and silicate and carbonatite melt (now glass); still more minute quantities of minerals in these inclusions are Ti-rich kaersutite, diopside, rutile, ilmenite, and magnesite. Schiano et al. (1994) interpret these to be immiscible segregations of an initially homogeneous metasomatizing melt ($T \sim 1250^{\circ}\text{C}$) that invaded the mantle be-

neath the islands. On the other hand, minute glass inclusions in strongly depleted, cataclastic harzburgite xenoliths from a Philippine arc volcano are dacitic and contain relatively high weight percentages of Na (3–5), K (2–4), and H₂O (4–5), as well as high Cl (1600–7300 ppm), and S (as much as 2500 ppm). Melts were incorporated into growing host minerals (primary olivine and enstatite, secondary metasomatic amphibole and phlogopite) at about 920°C (Schiano et al., 1995). Regardless of the nature of the percolating metasomatic liquid or its ultimate origin, it contains significant concentrations of incompatible elements that react with minerals in Cr-diopside peridotite, creating enriched concentrations. Mg-rich, Ti-Al-poor olivine and orthopyroxene are especially unstable in and susceptible to replacement by metasomatic liquids that contain relatively high concentrations of Fe, Al, alkalis, and volatiles.

Both cryptic metasomatism and modal metasomatism depend on diffusion of ions driven by concentra-



11.6 Metasomatism of spinel lherzolite wall rock adjacent to a vein of kaersutite amphibole. Vein includes about 5% apatite plus 5% phlogopite. Chemical and isotopic data from composite inclusion *Ba-2-1* from Dish Hill, California. (Redrawn from Nielson et al., 1993.) (a) Left, hypothetical spatial relations in veined mantle peridotite where $17 \times 16 \times 9.5$ cm inclusion (small rectangle) might have been positioned (compare Plate VI). Right, select modal and elemental variations (in wt.%) in wall rock samples WR-1, WR-2, and so on, at indicated distances from vein. (b) Chondrite-normalized rare earth element (REE) diagram showing strong enrichment in vein amphibole relative to Cr-diopside peridotite, particularly of light REE, and progressively less enrichment in wall rock away from vein. (Redrawn from Menzies et al., 1987.) Shaded is range of Cr-diopside spinel lherzolites from Kilborne Hole, New Mexico (Redrawn from Irving, 1980.) (c) Isotopic ratios for Dish Hill sample shown by solid circles and connecting tie line. Wall rock and vein fields are for inclusion samples from California and Arizona. MORB, representative mid-ocean ridge basalt. OIB, ocean island basalt. (Redrawn from Wilshire et al., 1988).



11.6 (Continued).

tion gradients. However, because it is slow, the effective distance over which metasomatism can occur solely by diffusion is limited. Metasomatism is greatly enhanced if there is advective flow of liquid along grain boundaries and especially through fractures, after which slower intragranular diffusion can take place.

The significance of liquid flow in granular media goes far beyond metasomatic processes. Obviously, migration of melts generated by partial melting of peridotite is a crucial step in their segregation into larger masses that ascend and intrude overlying lithosphere as magmas. Flow of fluids during metamorphism is an-

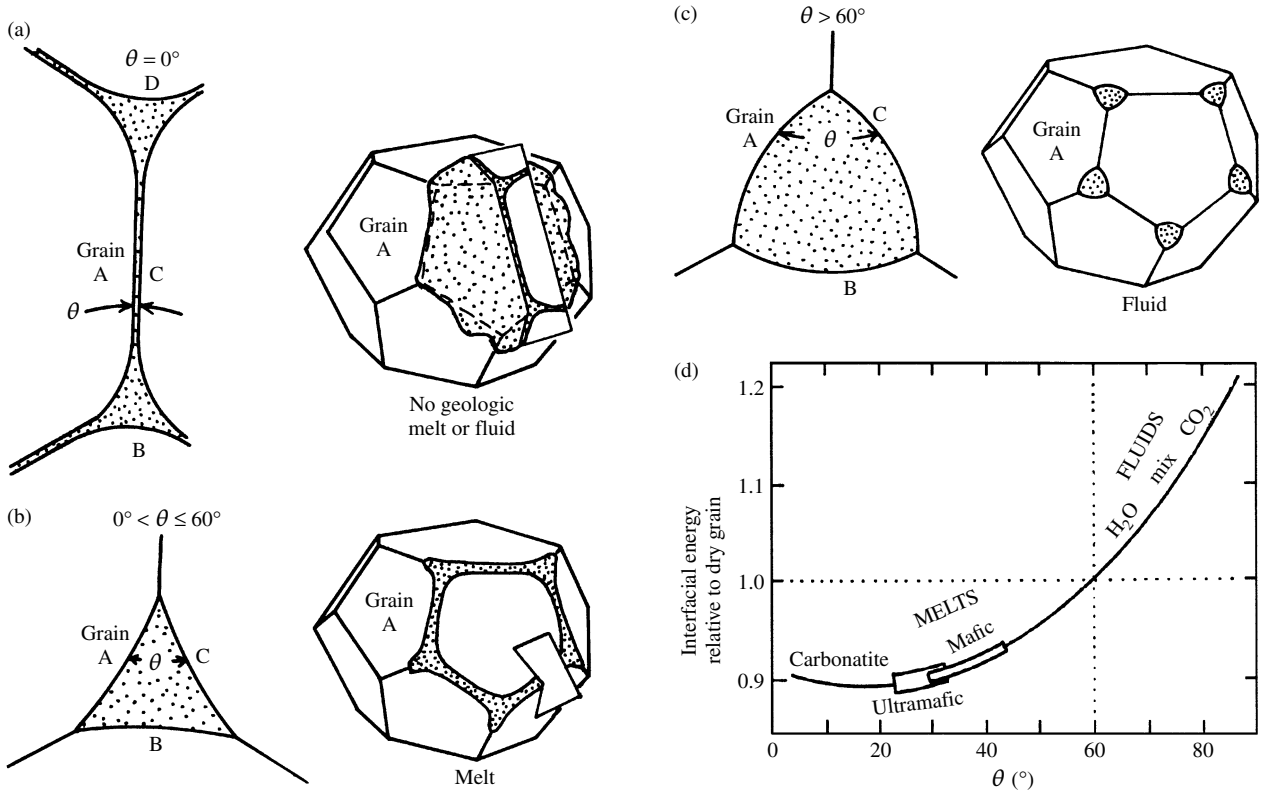
other subject of considerable importance. For these reasons, an extended discussion of liquid movement between mineral grains follows.

Distribution and Migration of Liquids in Granular Aggregates. Liquids can migrate through aggregates of mineral grains (rocks) by self-generated hydraulic fracturing (Section 8.2.1) or by **porous flow** (Section 8.5) that involves movement of the liquid along grain boundaries.

The principle of minimization of surface energy, or textural equilibration, is an important factor influencing the distribution and migration of very small volumes of liquid along grain boundaries in rocks (e.g., Watson et al., 1990). Governing principles are an extension of the discussion in Sections 6.4 and 7.4, which considered only crystal-crystal interfaces. Intersecting liquid/crystal interfaces (Figure 11.7) form an angle, θ , called the **dihedral angle**, which depends on the relative magnitudes of the crystal-crystal interfacial energy, γ_{CC} , and the liquid-crystal interfacial energy, γ_{LC}

$$11.2 \quad \theta = 2 \arccos \left(\frac{\gamma_{CC}}{2\gamma_{LC}} \right)$$

Interfacial energies, which depend on liquid and crystal compositions and on P - T conditions, can be readily measured by experiment. Three cases are significant (Figure 11.7). If $\theta = 0^\circ$, the liquid wets the entire crystal-crystal interface, forming a grain-boundary film. Apparently, no geologic phases exist for this case. If $\theta \neq 0$, no film exists, but interconnectivity of melt can still be maintained, not along grain surfaces, but rather along three-grain common edges, provided θ is less than or equal to 60° . The third case is for $\theta > 60^\circ$, where the liquid exists as minute pools at isolated three-grain corners. These geometries assume that there is a monomineralic (one-phase) crystalline aggregate and that surface energies of grains are isotropic. However, in the more realistic case of a polyphase peridotite in which grains, especially of the dominant olivine, have a preferred crystallographic orientation as a result of mantle flow, melt distribution can be some-



11.7 How very small volumes of liquid in a grain aggregate (rock) are distributed is a function of the relative interfacial energies of contacting phases (equation 11.2; see also Watson et al., 1990). The shape of reference grain A is dictated by the demands of minimal surface free energy in a one-phase grain aggregate (compare Figures 6.23 and 6.24). (a) For $\theta = 0^\circ$ the liquid wets the entire surface of grain A and its neighboring grains B, C, D, and so on, in the form of a thin film. Though theoretically possible, this has not been found to occur in geologic systems. Left-hand diagram shows distribution of liquid along grain boundaries in a plane perpendicular to their common edges. (b) For $0^\circ < \theta \leq 60^\circ$ melt is distributed in interconnecting strands with thread-lace scoria (Figure 6.25) is remarkable, but not unexpected for textural equilibration driven by minimization of surface energy. (c) For $\theta > 60^\circ$ small isolated packets of a fluid phase lie at grain corners. (d) Carbonatite melts and silicate melts of basaltic and ultramafic composition are stable along intergrain edges. H_2O - CO_2 fluids are stable along intergrain edges if in large volume whereas small volumes lie at grain corners.

what different from that depicted in Figure 11.7. Permeabilities (rates of liquid flow; Section 8.5) can be greater in anisotropic fabrics.

Figure 11.7d shows that CO₂ fluids are not interconnected until they make up about 8% of the aggregate as isolated pools merge. Water-rich fluids *may* have interconnectivity along three-grain edges. Mafic to ultramafic hydrous silicate melts have interconnectivity under virtually any upper mantle condition in peridotite even at melt fractions as low as about 1%. Carbonatite melts maintain interconnectivity at melt fractions as low as <0.1%.

Because of their low viscosity, melts are able to infiltrate along grain edges through mantle rock readily in the process of porous flow (McKenzie, 1985). The principal driving force, at least for larger volumes of melt, is the buoyancy of the less-dense melt. However, interconnecting strands of melt along three-grain edges can actually lower the interfacial energy in a grain aggregate, relative to the melt-free aggregate. Consequently, any available melt can be drawn spontaneously into an initially melt-free peridotite, reducing the energy of the system (Watson et al., 1990). Melts probably penetrate along grain boundaries by a coupled dissolution-precipitation process. Rates of porous flow are on the order of 1 mm/day for basaltic melts and perhaps a couple of orders of magnitude faster for aqueous fluids if their dihedral angles are <60°. Carbonatite melts also move fast.

In sheared peridotite xenoliths, grain size has been reduced and grains stretched into lenticular shapes, creating a penetrative foliation. Any melt in ductilely deforming peridotite would tend to collect along shear surfaces where the rate of melt migration might be enhanced, that is, flow permeability increased, relative to a static, nondeforming rock system.

Metasomatized Subcrustal Lithosphere. It is quite possible that porous flow of liquids, perhaps assisted by the grain-scale deformation just described, occurs in the convecting asthenospheric mantle. Very slight partial melting (<1%?) of the asthenosphere might create metasomatizing melts. Where strands of ascending buoyant liquids collect to form overpressured volumes, hydraulic fracturing can occur, allowing faster ascent. But because moving liquid of either form carries little heat relative to the larger mass of wall rock, the melt or fluid readily freezes as veins, unless there is enough volume of melt to continue to ascend as magma. Mantle wedges overlying subducting oceanic plates (Figure 11.2) are likely extensively veined (Davies, 1999). Uppermost mantle beneath continental rifts and where mantle plumes are operative is also likely metasomatically veined. After liquid production has ceased beneath a particular plate, it may begin again during the course of changing plate motions. Thick, relatively cool (low-geothermal-gradient) mantle underpinnings of

continents, especially shields that may be as old as the early Proterozoic, might have been metasomatized to varying degrees by repeated veining episodes. The Archean mantle may have been too hot to allow freezing of veins. Younger lithosphere in oceanic areas has also been metasomatized. This metasomatized subcrustal or mantle lithosphere, laced as it is with metasomatic veins, is believed to play a very important role in the generation of alkaline magmas, especially the highly alkaline, highly potassic magmas (Section 11.5).

* 11.3 GENERATION OF MAGMA IN MANTLE PERIDOTITE

Chapter 5 introduced concepts of the melting of multi-phase silicate rocks based on their crystal-melt equilibria. In this section, additional concepts, which relate to generation of a wide range of ultramafic and mafic melts from “normal” mantle sources, that is, Cr-diopside peridotite, are developed.

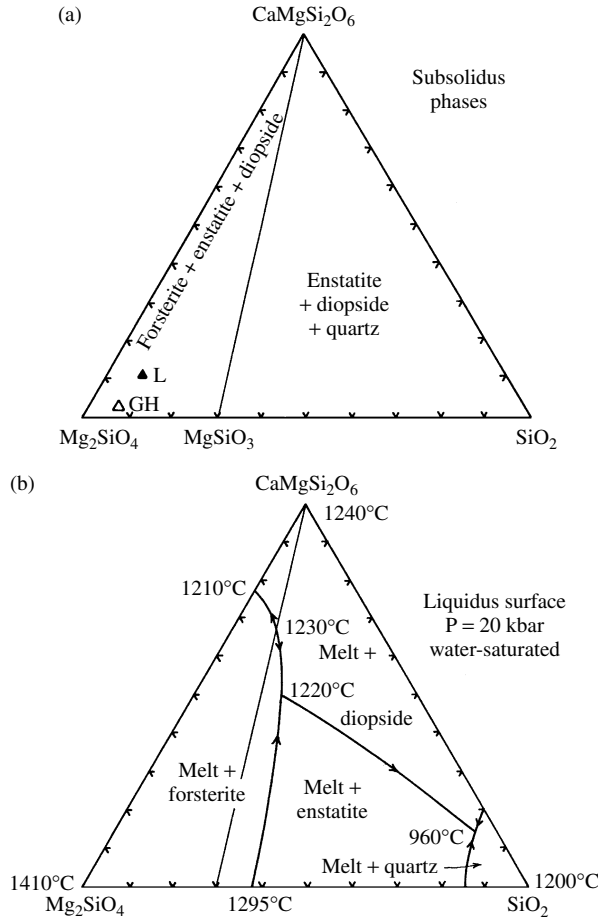
Although decompression melting is undoubtedly paramount in generation of basaltic melts from peridotite, it is convenient to examine partial melting that accompanies an increase in T using available T - X phase diagrams to elucidate pertinent concepts. The system Mg₂SiO₄-CaMgSi₂O₆-SiO₂ under water-saturated conditions at $P = 20$ kbar (Figures 11.8 and 11.9) will be used to model melting. The small amounts of Al, Ti, Fe, Cr, Mn, K, Na, and other minor and trace elements, mainly sequestered in solid solution in clinopyroxene, the small modal amounts of spinel or garnet in natural peridotites, and amphibole and mica in hydrous peridotite, are ignored at this point. The average worldwide spinel lherzolite (Table 11.1) with its somewhat depleted composition is used as a model of the source rock.

Generation of a melt that is less than the whole rock is referred to as **partial melting**. Different types of partial melting have been recognized, only two of which are considered here:

1. **Equilibrium (batch) partial melting** generates a melt that is in equilibrium with the **crystalline residue** as long as the two are in intimate communication and until the melt leaves. Equilibrium melting in a closed system is the reverse of equilibrium crystallization.
2. **Fractional partial melting** generates melt that is immediately isolated from the crystalline residue; no reaction relations occur between melt and crystals. As it is a disequilibrium process, it is not the reverse of fractional crystallization.

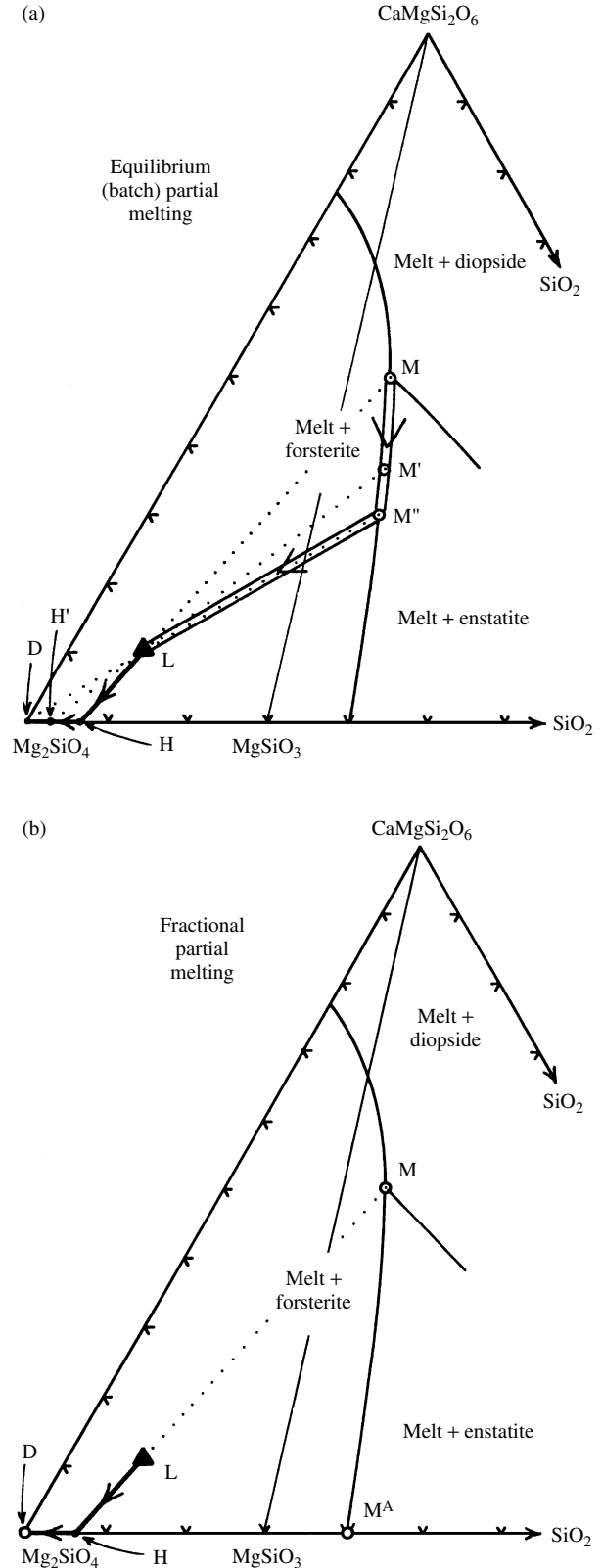
11.3.1 Equilibrium (Batch) Partial Melting of Lherzolite

In Figure 11.9a, the first melt to form with rising T in a lherzolite L has a composition at point M . This invariant point represents the only possible melt in the



11.8 Ternary system Mg_2SiO_4 - $CaMgSi_2O_6$ - SiO_2 . (a) Subsolidus assemblages. Average worldwide spinel peridotite, which is a lherzolite, *L*, and garnet peridotite, which is a garnet harzburgite, *GH*, from Table 11.1. (b) Projected liquidus features at $P = 20$ kbar (2 GPa; depth about 70 km) water-saturated. (Redrawn from Kushiro, 1969.)

ternary system that coexists in equilibrium with two pyroxenes plus olivine, the three major minerals constituting the lherzolite source rock. Point *M* at the confluence of three boundary lines in Figure 11.9a has a physical resemblance to the small pools of melt that form at mutual grain boundaries (Figure 11.7c) of two pyroxenes and olivine in a lherzolite; melts do not develop at olivine-olivine or pyroxene-pyroxene grain contacts. *M* is an incongruent melt that is substantially more enriched in silica and $CaMgSi_2O_6$ than the source. This reflects preferential melting of diopside and a lesser amount of incongruent melting of enstatite, which yields a more silica-rich melt plus residual olivine. If the melt *M* were to crystallize, it would form 7% quartz, 46% diopside, and 47% enstatite. Recalculation to 100% of the normative *Q*, *Di*, *Hy* (2.2, 17.9, and 21.5) in average Hawaiian tholeiitic shield basalt in Table 13.2 gives 5%, 43%, and 52%. This similarity in composition between the hypothetical model melt *M* and tholeiite magma ignores the neces-



11.9 Partial melting in the system Mg_2SiO_4 - $CaMgSi_2O_6$ - SiO_2 at $P = 20$ kbar water-saturated. See also Figure 11.8. (a) Equilibrium partial melting of average worldwide spinel lherzolite (*L*, solid triangle). Double line represents track of changing partial melt compositions. Heavy line between *L* and *D* represents changing composition of crystalline residues. (b) Fractional partial melting. Open circles at *M*, *M^A*, and *D* represent melts.

sity that there be sufficient K, Na, Al, Fe, and Ti in the real source to stabilize plagioclase and Fe-Ti oxides in the real tholeiitic magma at low pressures.

Continued melting of the source lherzolite simply yields more melt *M*, provided crystals of the two pyroxenes plus olivine are present in the crystalline residue. Invariant equilibrium persists and *T* and melt composition remain unchanged. In a *T*-increasing melting system, the latent heat of fusion consumes the increased thermal energy. The increased amount of melt is created by preferential dissolution of diopside, eventually shifting the composition of the crystalline residue to diopside-free harzburgite, *H*. The lever rule indicates that about 79% of the system is harzburgite residue and 21% is melt. That is, the **melt fraction** is 21%. With all of the diopside melted out of the source, so the residue is only olivine plus enstatite, further *T* increase can no longer generate a melt at the invariant point *M*. Instead the melt composition must track down the boundary curve on the liquidus surface between the stability fields of melt + forsterite and melt + enstatite, and away from the $\text{CaMgSi}_2\text{O}_6$ apex; this component in the melt is simply being diluted as enstatite dissolves. Melt *M'* on this boundary curve, for example, coexists in equilibrium with a harzburgite, *H'*. The tie line between the melt and crystalline residue *must* pass through point *L*, the composition of the initial source rock. The reason for this special geometric constraint is that in equilibrium melting the system remains closed and all melt-crystal tie lines are tethered at the bulk composition point. It may be noted that point *M'* is closer to this bulk composition than is *M*; the melt fraction in the system has increased to about 29%. Continued melting drives the melt composition to *M''*, where the melt fraction is about 33%; the crystalline residue is now pure forsterite, or dunite rock, *D*.

High-degree partial melts, 30%–40%, of an initial lherzolite source in the mantle contain high concentrations of dissolved olivine and are ultramafic picrite and komatiite. Low-degree partial melts, 5%–10%, are mostly of dissolved clinopyroxene (and spinel or garnet in real sources). The solid residue progressively loses spinel or garnet, clinopyroxene, and then orthopyroxene.

11.3.2 Fractional Partial Melting of Lherzolite

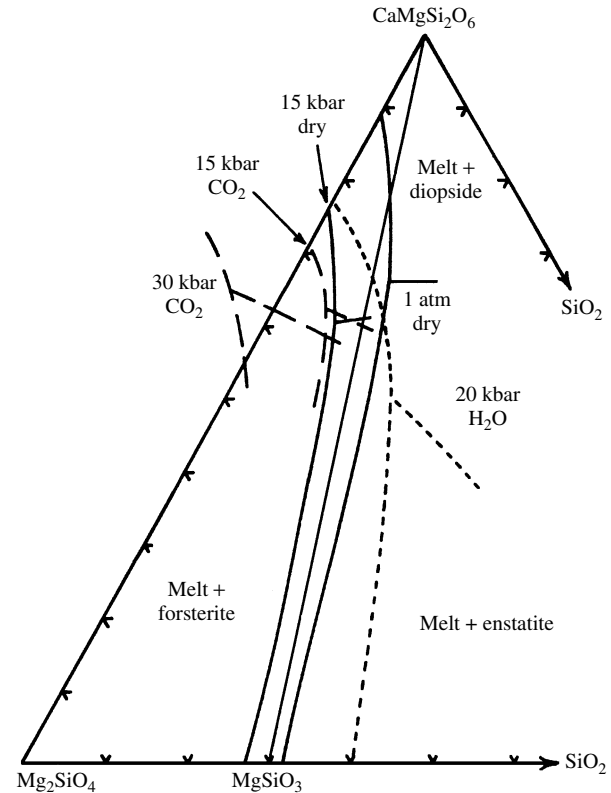
This manner of partial melting begins, as before, with generation of an initial melt *M* (Figure 11.9b). But each successive parcel of melt *M* generated is removed (fractionated) from the crystalline residue, which shifts directly away from *L* along a straight line projected from *M* to harzburgite, *H*, as before. However, once the crystalline residue becomes *H*, further partial melting yields different products from those in the equilibrium case. With a source rock at

H, the system has no “memory” of ever having been at *L*; it only “thinks for the moment” that the source is at *H*. With no diopside in *H*, only forsterite plus enstatite, the only possible melt that can coexist in the whole ternary system is melt M^A . But this melt requires a 75°C increase over melt *M*. Therefore, as more thermal energy is absorbed by the source rock after the last liquid *M* has been removed from the rock and as its last diopside is dissolved, no further melting can occur until sufficient heat has been absorbed to raise the source rock to 1295°C. At that *T*, melting resumes, generating melt M^A , chiefly by dissolving enstatite.

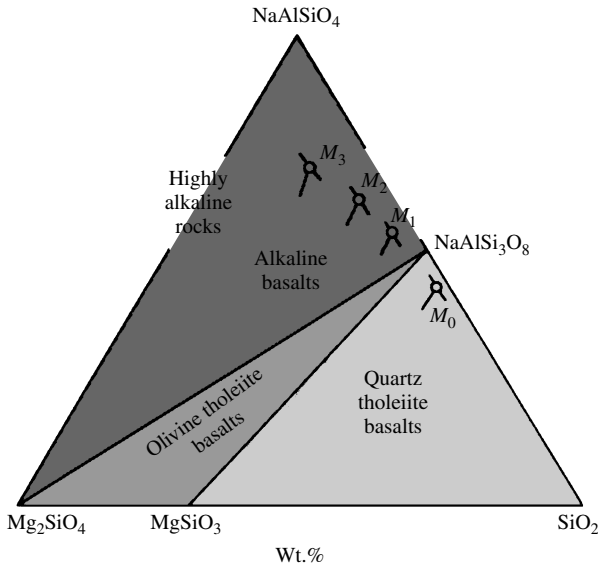
11.3.3 Factors Controlling Partial Melt Composition

Changes in *P* and volatile concentration significantly shift the position of cotectic boundary lines and invariant first-melt compositions in the model source systems shown in Figures 11.10–11.12.

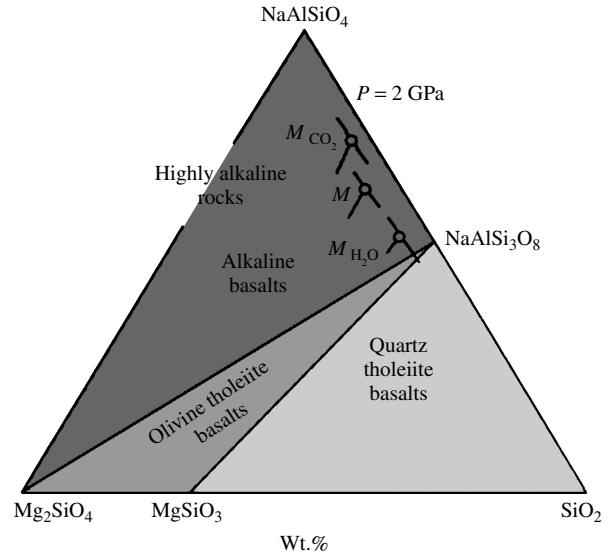
Increasing *P* shifts anhydrous invariant-point first melts toward lower silica and higher Na (alkali) compositions. In Figure 11.10, from a silica-saturated melt (modeling a quartz-normative tholeiitic basalt magma) at 1 atmosphere, the equilibrium shifts to an olivine-saturated melt (alkaline basalt magma) at 15 kbar.



11.10 Beginning-of-melting invariant points and associated liquidus boundary lines in the system Mg_2SiO_4 - $\text{CaMgSi}_2\text{O}_6$ - SiO_2 . Solid lines, volatile-free (dry); short dashed lines, water-saturated; long dashed lines, CO_2 -saturated. (Redrawn from Eggler, 1974.)



11.11 Shift of beginning-of-melting invariant points M_0 , M_1 , M_2 , and M_3 at 1 atm, 1, 2, and 3 GPa, under volatile-free conditions. Basaltic melts modeled by this ternary system are indicated. (Redrawn from Kushiro, 1968.)



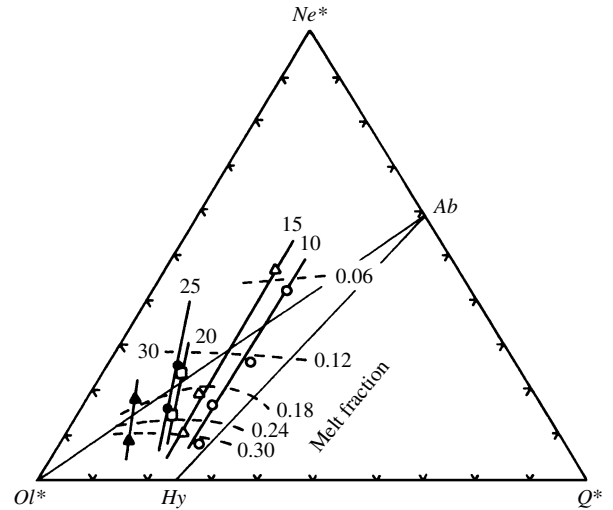
11.12 Shift of beginning-of-melting invariant points at 3 GPa under indicated volatile-saturated conditions. M is the volatile-free dry system. (Redrawn from Egglar and Holloway, 1977.)

In Figure 11.11, increasing P shifts first melts from silica-saturated (quartz tholeiite magma), to silica-undersaturated (alkaline basalt magma), and to highly silica-undersaturated (nephelinite and other highly alkaline magmas) at highest P . Partial melts contain more dissolved olivine at higher P . Hence, those developed from mantle peridotite at P in excess of about 3 GPa (about 100-km depth) have MgO content >15 wt.% and are picrites and komatiites, in contrast to the basaltic melts generated at lower P (Herzberg and O'Hara, 1998). Thus, when a high- P partial melt decompresses at shallower depths, olivine precipitates to maintain equilibrium. This phenomenon accounts, at least in part, for the common occurrence of olivine phenocrysts in extruded basalt magmas.

Increasing CO_2 pressure (fugacity) also decreases silica and increases alkali concentrations in first melts, reinforcing the effects of increasing P .

Increasing water pressure (fugacity) has the opposite effect, shifting invariant first-melt compositions to higher silica and lower alkali and $\text{CaMgSi}_2\text{O}_6$ concentrations.

Experimentally generated partial melts of a natural peridotite (e.g., Hirose, 1997) confirm the results of the simple ternary systems just described in shifting low- T first melts at increasing P to more silica-undersaturated compositions (Figure 11.13). But, in addition, the experiments on natural peridotite reveal the importance of yet another factor controlling melt compositions—the **degree of partial melting**, manifested by the melt fraction. Smaller-degree partial melts are nepheline-normative alkaline basalt compositions be-



11.13 Factors controlling the composition of experimentally generated partial melts of anhydrous spinel lherzolite KLB-1 from Kilborne Hole, New Mexico. Partial melts (circles, squares, and triangles) are increasingly enriched in normative olivine (depleted in normative quartz) with increasing P (10, 15, 20, 25, and 30 kbar). Small extents of partial melting yield nepheline-normative melts but increasing melt fractions (0.06, 0.12, 0.18, 0.24, 0.30) generated at increasing T have more normative hypersthene. $Ne^* = ne + 0.6 Ab$; $Q^* = Q + 0.4 Ab$; $Ol^* = Ol + 0.75 Hy$. Experiments by Falloon et al. (1999) on KLB-1 show that only the highest melting temperatures in these anhydrous melting experiments attained equilibrium; accordingly, they urged caution in interpreting melt compositions. However, the *pattern* of changing melt compositions with respect to degree of melting and P seems reasonable. (Redrawn from Hirose and Kushiro, 1993.)

cause they contain high concentrations of the incompatible elements Na and K as well as H, C, Ti, P, and others. These elements are not present in the simple model ternary systems just described but are sequestered in natural peridotite minerals, chiefly clinopyroxene. As the degree of melting increases at constant P , incompatible elements are diluted by increasing amounts of dissolved Si, Al, Fe, Mg, and Ca so that the melts become less alkaline, more silica-rich hypersthene- or even quartz-normative tholeiitic basalt compositions.

11.3.4 Modeling Partial Melting Using Trace Elements

Quantitative modeling of trace elements is another way of looking at the relation between melt composition and degree of partial melting. The composition of the source rock is also important in governing the trace element composition of partial melts, as may be appreciated from the bulk partition coefficient, D (equation 2.2 and Section 2.5.1).

Low degrees of partial melting corresponding to a small melt weight fraction, F , produce melts enriched in incompatible elements and depleted in compatible elements compared with the source rock. Larger degrees of melting reduce incompatible element concentrations but increase compatible element concentration until at $F = 1$ (100% melting) the melt has the same composition as the source. The type of melting, whether equilibrium (batch) or fractional, influences the trace element concentration in a partial melt, in addition to the source rock composition. Only equilibrium melting is considered here (but see Rollinson,

1993) because it may be the more likely case, but this point is controversial.

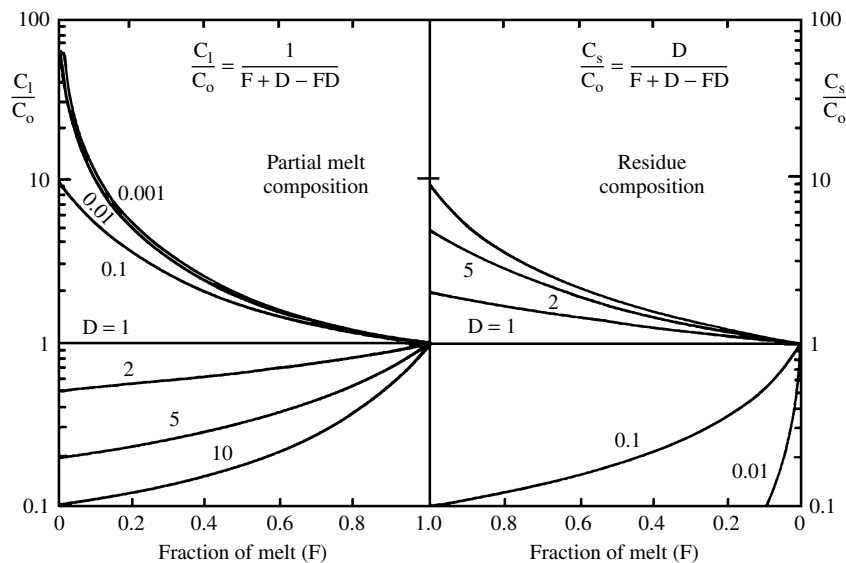
Equilibrium (batch) partial melting can be described by

$$11.3 \quad \frac{C_l}{C_0} = \frac{1}{(F + D - FD)}$$

and by

$$11.4 \quad \frac{C_s}{C_0} = \frac{D}{(F + D - FD)}$$

where C_0 is the initial concentration of an element in the solid source, C_l is the concentration of the same element in the partial melt, and C_s is the concentration in the unmelted solid residue. These equations are plotted in Figure 11.14 for several values of D as a function of F and melt, residue, and source compositions. For strongly incompatible elements, such as uranium in a basalt-peridotite system, D is very small (<0.001) and $C_l/C_0 \sim 1/F$; the concentration of the element in the partial melt depends essentially on F and approaches infinite concentration as F approaches 0. In the crystalline residue (equation 11.4), a highly incompatible element is strongly depleted with even very small degrees of partial melting. Relative to the source, small degree melts can have large changes in the ratio of two incompatible elements whose D differs significantly, say, 0.5 versus <0.01 . Batch partial melting produces melts that are not extremely depleted in the compatible elements. The lowest concentrations are proportional to $1/D$ for very small F . For example, if $D_{Ni} = 5$, the lowest concentration of Ni (at very small F) that can oc-



11.14 Plots of equations 11.3 and 11.4 showing trace element concentrations in partial melts (C_l) and residue (C_s) as a function of melt fraction, F , and bulk partition coefficient, D . Original concentration in source rock is C_0 . On left, curves for $D = 0.01$ and 0.001 are almost coincident and appear to be a single, thick line.

cur in a single batch of partial melt is one-fifth of the initial concentration in the solid source.

Figure 11.15 shows primitive mantle-normalized rare earth element (REE) patterns (Section 2.5.3) for a garnet lherzolite source and the batch partial melts that could be derived from it over a range of melt fraction. Smallest-degree partial melts (low F) are most enriched in the most incompatible elements. Higher-degree partial melts are progressively more like the source. Remember that in these REE diagrams the most incompatible elements are on the left and less incompatible elements are on the right. Also note that the presence of garnet in the residue causes the patterns to have a markedly negative slope and high light REE to heavy REE ratio (Figure 2.22); in other words, La/Yb is large (Figure 2.23).

11.3.5 Characteristics of Primary Magma

Petrologists do not universally agree on the composition of **primary magmas** that have moved to the surface unmodified and undifferentiated from their mantle peridotite source. It is, nonetheless, important to establish a baseline reference against which differentiated magmas discussed in the following chapter can be compared.

Erupted magmas that contain dense mantle peridotite inclusions are commonly believed to be primary because their upward transport must be too rapid to allow differentiation of the magma and dissolution of the inclusion. However, entrainment of pieces of the mantle after some differentiation and subsequent to segregation from the source rock cannot always be ruled out. In any case, hosting of mantle inclusions is only permissive evidence; primary magmas need not contain inclusions.

Other criteria for recognition of primary magmas are chemical, though there is little consensus on the values to be used. Melts segregated from mantle peri-

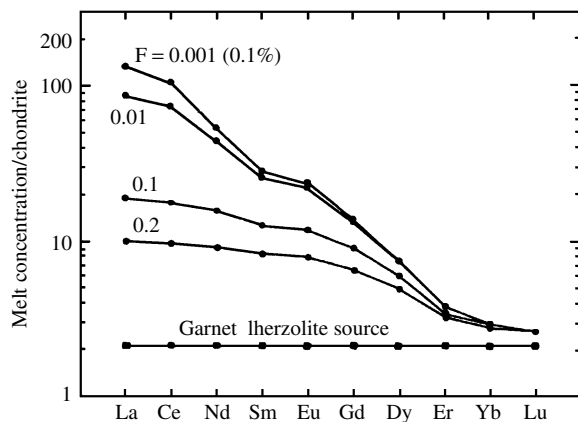
dotite are in equilibrium with olivine, about Fo₉₀. The Fe^{2+}/Mg distribution coefficient between this olivine and an equilibrium melt, or $(\text{Fe}^{2+}/\text{Mg})_{\text{olivine}}/(\text{Fe}^{2+}/\text{Mg})_{\text{melt}}$, is P -dependent but is approximately 0.3 ± 0.03 . This translates into an atomic ratio, called the **Mg number**, of $100 \text{ Mg}/(\text{Mg} + \text{Fe}^{2+}) = 68 - 75$, or a weight ratio $\text{FeO}/\text{MgO} = 0.4 - 0.7$ in the rock. These ratios for total Fe as FeO or Fe^{2+} are relatively insensitive to the degree of partial melting but are strongly influenced by fractional crystallization. In addition, 8 wt.% MgO, 400 ppm Ni, and 1000 ppm Cr are commonly used as lower limits for primary magmas.

A necessary (but again not sufficient) condition for a mafic volcanic rock to have solidified from a primary magma is that its composition lies at a beginning-of-melting invariant point (e.g., in Figures 11.9 and 11.10) at some mantle P . For a lherzolite source, this near liquidus high- P melt must be saturated or equilibrated with olivine, two pyroxenes, and either spinel or garnet. Most basaltic rocks have not solidified from magma so equilibrated, implying modifications after leaving the source, as recognized in early studies (Yoder and Tilley, 1962).

Metasomatized mantle sources do not necessarily contain olivine, or, if it is present, it is more Fe-rich than Fo₉₀. Primary magmas derived from such sources will not have the same chemical values as those derived from normal depleted peridotite.

Because of the uncertainties involved in establishing what is a primary magma, petrologists often attempt to decide which rock in a comagmatic suite under investigation solidified from a **primitive magma**, one whose composition was *least unmodified* by differentiation processes after leaving its source. Such a magma tends to have the largest Mg number and highest concentrations of Ni and Cr. However, caution must be exercised, for example, a basalt may have high Mg number, Ni, and Cr but petrographic examination might reveal unusually large amounts of olivine that could have accumulated by gravity settling, a type of differentiation process. Incompatible elements can be enriched in some differentiated magmas, but they can also be enriched in metasomatized mantle and in the primary magmas derived from them by partial melting.

Obviously, deciding whether a rock formed from a primary magma, or which rock solidified from a primitive magma, requires careful study.



11.15 Chondrite-normalized diagram for rare earth elements in melts generated by varying degrees of partial melting of a garnet lherzolite source. (Data from Rollinson, 1993.)

* 11.4 MAGMA GENERATION IN SUBARC MANTLE WEDGE

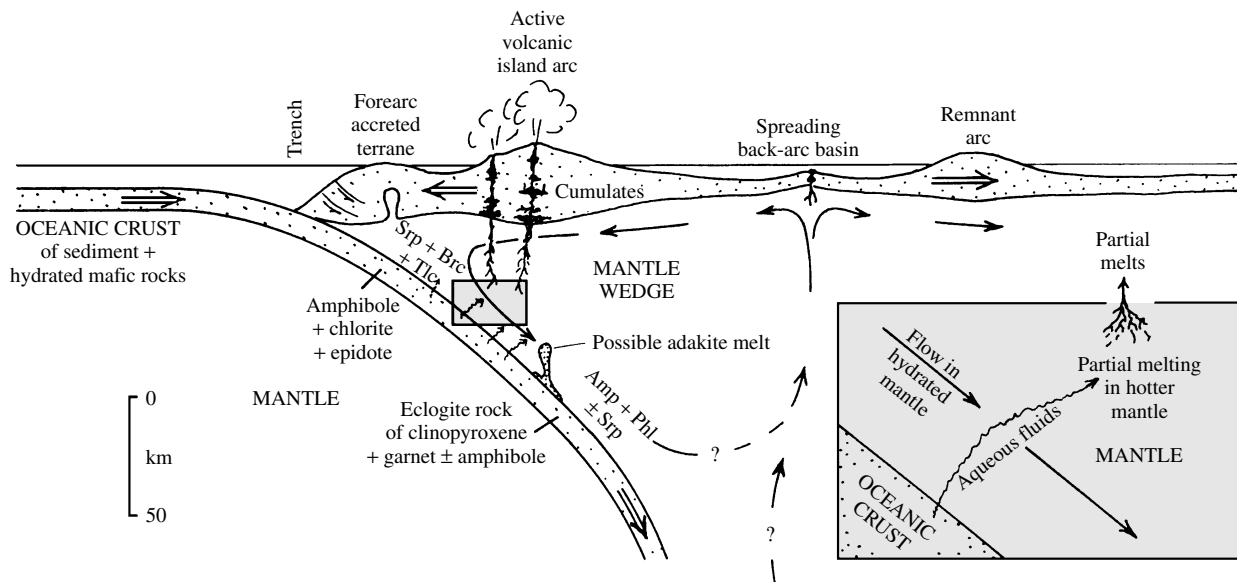
After oceanic rifts, the most voluminous global magma production is in subduction zones (Figure 1.1). **Arc magmatism** occurs where oceanic lithosphere descends

beneath overlying oceanic lithosphere in **island arcs** and beneath continental margins in **continental arcs**. Few, if any, mantle-derived magmas survive their journey through the crust, especially thick sialic continental crust, unscathed by contamination and other differentiation processes, making it difficult to understand their primitive character. This overprint, however, is highly variable, depending largely on the thickness of the crust. Island arc volcanic rocks, because they form on relatively thin and mafic crust, provide a logical starting point for inquiry into the nature of magma generation in the subarc mantle (see Davidson [1996] for a lucid summary).

Minor but widespread hornblende and biotite in volcanic arc rocks are stabilized by >3–5 wt.% water in magmas, in contrast with the essentially anhydrous mantle-derived magmas generated in other oceanic settings. As first pointed out by Coats (1962), dehydration (“drying out”) of the oceanic slab as it subducts and heats up likely produces relatively “wet” arc magmas. The water liberated from the slab promotes generation of hydrous partial melts in the overlying wedge of hot mantle peridotite (Figure 11.2).

11.4.1 Dehydration of Subducting Oceanic Crust

Subducting oceanic lithospheric plates consist of mantle peridotite and overlying oceanic crust (see Figure 13.1). This crust has 6–7 km of basaltic rock and an overlying veneer of sediment, variable proportions of which are scraped off into the trench and accreted onto the edge of the overlying plate (Figure 11.16). The crust contains variable amounts of volatiles, chiefly H₂O and CO₂. Carbonate minerals occur in deep marine sediment and in pore spaces and veins in basaltic rocks; overall CO₂ concentrations in oceanic crust are probably about 0.1 wt.%. The overall concentration of water in the crust is 1–2 wt.%, part of which is physically entrapped in pore spaces and cracks. However, an important amount is chemically bound in hydrous silicate minerals formed by moderate-*T* seafloor metamorphism of the basalt and peridotite at oceanic spreading ridges and by subsequent long-term low-*T* submarine “weathering” as the plate moves away. Hydrous Mg-Fe-Ca-Al silicates that replace primary magmatic minerals in basalt include epidotes, micas, amphiboles, serpentines, chlorites,



11.16 Highly schematic dynamics of an island arc-mantle wedge system. Double-line arrows indicate relative motion of crust (stippled). Subduction of hydrated oceanic crust liberates water in complex dehydration mineral reactions, forming ascending aqueous fluid solutions (wavy arrows). The forearc accreted terrane is built chiefly of scraped-off oceanic sediment and possible mafic-ultramafic rocks (ophiolite). Below this forearc, water liberated from the dehydrating crust hydrates the overlying peridotite in the mantle wedge to serpentine (\pm brucite \pm talc). Some of this buoyant mass of low-density serpentinized peridotite rises as diapirs (see diapir in figure) into the accreted terrane and may be extruded onto the ocean floor. As the descending crust continues to heat up, more water is liberated, and it transforms into drier, high-*P* eclogite, but amphibole may still be present. In exceptionally young, hot subducting crust, dehydration partial melting of amphibole eclogite may generate adakite melts that rise through the mantle wedge, possibly mixing with andesitic partial melts, before intruding into the arc crust or erupting. Hydrated mantle peridotite overlying the subducting crust may be dragged down in a “corner flow” by viscous coupling, releasing water as the low-*T* hydrous minerals partially dehydrate into amphibole + phlogopite \pm serpentine that are stable at higher *P-T* (Figure 11.18). Inset diagram in lower right (enlarged from the box in the main part of the diagram) shows that convective flow in mantle draws rising aqueous fluids laterally away from the crust into the hotter part of wedge where partial melting occurs. It should be noted in this inset diagram that temperatures increase in rocks from the lower left corner to the upper right: That is, there is an inverted thermal gradient.

Special Interest Box 11.3 Significance of the radiator on the front of the global magma engine

Sampling of oceanic lithosphere reveals widespread and locally intense alteration of the rocks at spreading ridges by seawater advecting through pervasive fractures. As it traverses through the hot crust, the seawater is heated and dissolves high concentrations of various mineral components from the hot rock. Where vented back into the ocean at T as much as 350°C these hydrothermal solutions form “black smokers” (Figure 11.17). Advective circulation of seawater at oceanic ridges has the following major impacts on global geology:

1. Circulation acts as a gigantic “radiator on the front of the global magma engine” (Sigurdsson et al., 2000, p. 5), providing for significant dissipation of heat from the interior of the Earth. It is estimated that seawater advects through the hot crust at a rate of about 1.5×10^{14} kg/y, or



11.17 Black smoker venting from a hot spring on the East Pacific Rise at a water depth of about 3 km. Dark-gray plume carries high concentrations of minute mineral precipitates, including Cu-Fe-Zn sulfides. Photograph taken by Robert D. Ballard from the deep submersible *Alvin* and furnished courtesy of Woods Hole Oceanographic Institution.

enough to recycle the entire mass of the oceans once every 5 My!

2. Seafloor hot springs support a unique biological community that is a factor in the global food chain.
3. The circulation system impacts the chemical composition of ocean water. Sinks and sources of mineral components are influenced by crust-seawater interactions. Components added by rivers and precipitated in marine sediments are not the only governing factors, as once thought.
4. Minerals, such as chalcopyrite (CuFeS_2) and sphalerite (ZnS), are deposited on the seafloor as the hydrothermal solutions cool at black smoker vents. Because of seafloor spreading, these potentially economic deposits are included in on-land ophiolite (Section 13.6) or are processed in the subduction “factory” where plates converge.
5. Primary anhydrous minerals in the mafic and ultramafic rocks of the oceanic lithosphere are transformed by seafloor metamorphism into hydrous, more oxidized minerals. As these hydrated rocks are heated in subducting slabs, the liberated water promotes magma generation in the overlying mantle wedge and is, therefore, chiefly responsible for global arc magmatism—the second most prolific on Earth (Figure 1.1).

chloritoid, prehnite, pumpellyite, talc, lawsonite, zeolites, and clay minerals. As the crust is compressed and slowly heated during subduction, physically entrapped water is first released at a depth probably not exceeding several kilometers. During further descent and heating, a series of endothermic, subsolidus dehydration reactions (Section 5.7.2 and Figure 5.31b) are “smeared” over a considerable range of T , P (depth), and volatile fugacities, progressively liberating water from the hydrous minerals (e.g., Liu et al., 1996). The smearing results from extensive solid solution in the hydrous phases, like the melting of plagioclase solid solutions that takes place over a range of T (Figure 5.13). P - T paths that a subducting slab take through time are highly variable, making generalizations difficult.

The major factor governing the depth of volatile mineral decomposition in subducting slabs is their thermal state (T distribution). For example, dehydration occurs at shallower depths in hotter slabs. This thermal state depends on many factors (Peacock et al., 1994):

1. The age of the incoming slab. Young lithosphere from a nearby spreading ridge is hotter.
2. The mass of previously subducted lithosphere. Previously cooled asthenosphere cannot heat the slab as fast.
3. Intensity of shear heating along the margins of the

more rigid lithospheric slab as it penetrates the more ductile asthenosphere. Faster-moving slabs create more heating (equation 11.1).

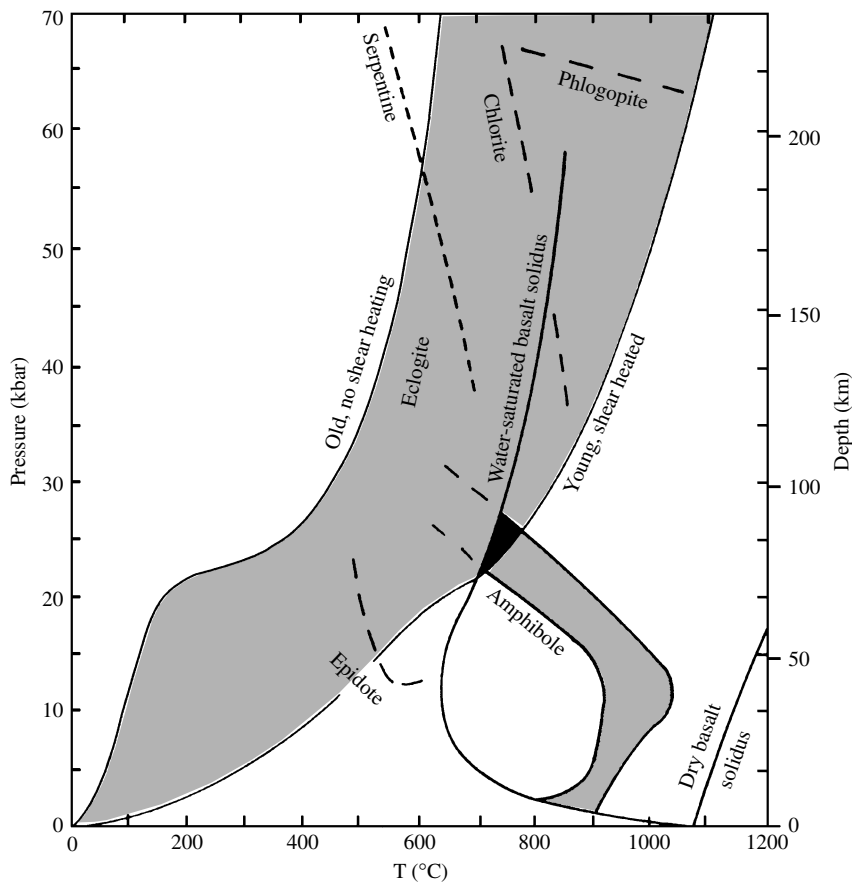
4. Slab dip: A slab that penetrates cooler mantle on a near-horizontal trajectory, as beneath the central Andes, does not heat as fast as a slab that penetrates hotter deeper mantle on a near-vertical path, as beneath the Mariana Islands in the western Pacific.
5. Vigor of convection in the mantle wedge: The basal part of the cooling wedge is partially coupled in a viscous manner with the descending slab and is dragged down with it, thus the slab does not contact hotter mantle (Figure 11.16).
6. Mineral reactions in the slab. Endothermic dehydration reactions cool the slab whereas exothermic hydration reactions heat it.

In short, older lithosphere created at more distant spreading plate junctures subducting at a slower speed without shear heating dehydrates at greater depths (Figure 11.18).

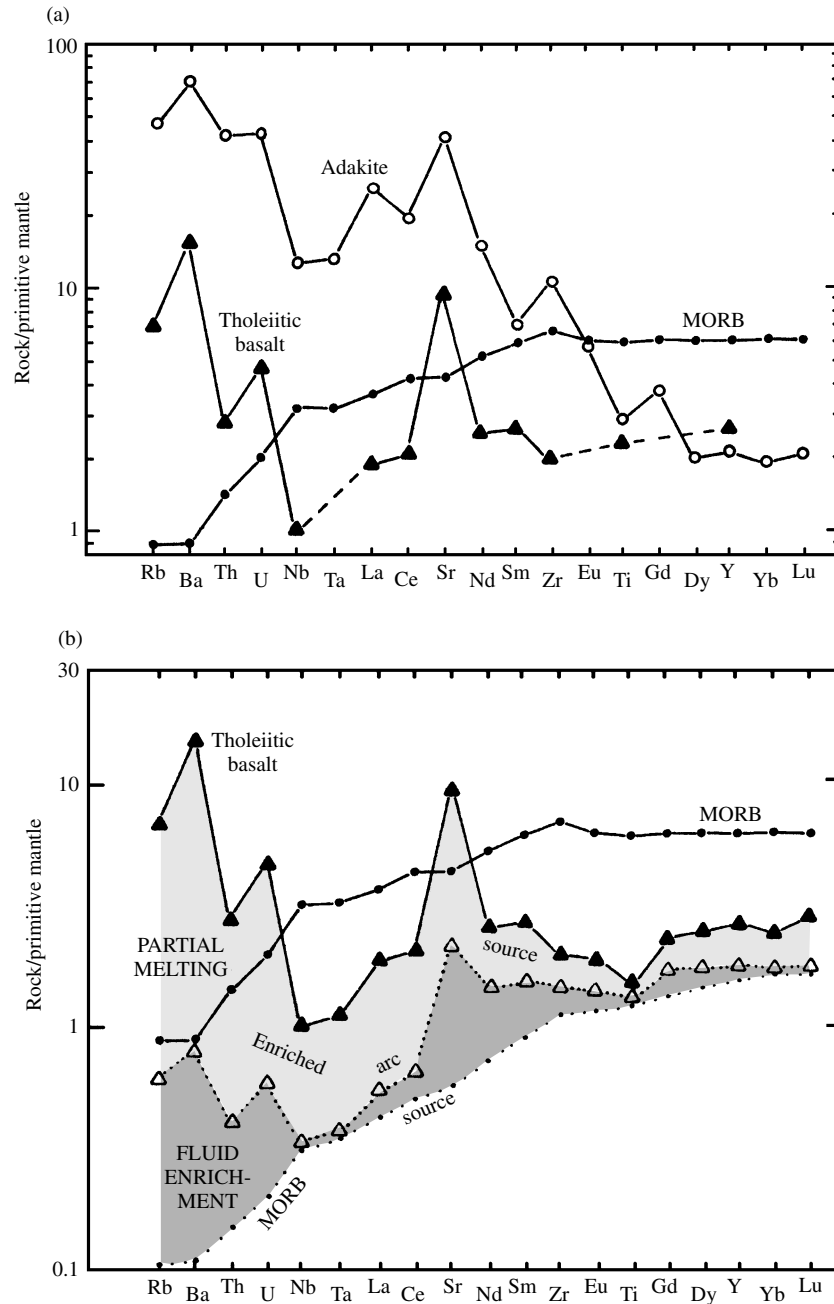
11.4.2 Magma Generation in the Mantle Wedge

Experiments on spinel lherzolite under water-undersaturated and water-saturated conditions at 10 kbar—conditions appropriate to the mantle wedge—yield basaltic andesite and andesite melts, respectively, for melt fractions of <0.23 at 1000–1050°C (Hirose, 1997). Evidence for such degrees of partial melting is found in very rare, strongly depleted harzburgite xenoliths hosted in island arc rocks (Maury et al., 1992).

Trace Element Arc Signature. Primitive island arc rocks have a characteristic arc signature in normalized trace element diagrams (Section 2.5.3) that is “spiky” in contrast to the smooth pattern of MORB and most other oceanic rocks (Figure 11.19). Large-ion-lithophile elements (LILEs), such as Rb, Ba, Th, U, and K, are enriched relative to rare earth elements (REEs), La through Lu, and high-field-strength elements (HFSEs), such as Nb, Ta, and Ti. The relative



11.18 Range of calculated P - T paths (light shading) of subducting oceanic crust. End-member subducting slabs are (1) old, relatively cold upon entry into mantle, and slow-moving and (2) young, relatively hot, and fast-moving so shear heating occurs. Dark shading indicates P - T conditions for dehydration melting of hydrous basalt systems that contain amphibole (compare Figure 5.11). Note the small, wedge-shaped P - T “window” (black), where basaltic crust in young, shear-heated subducting slabs is predicted to experience dehydration melting. Basalt at P greater than about 25 kbar is composed of an eclogite assemblage of garnet + clinopyroxene \pm amphibole. Upper P - T stability limit of minerals indicated by dashed lines. (Redrawn from Peacock et al., 1994.)



11.19 Trace-element arc signature in island arc rocks. (a) “Spiky” patterns for island-arc adakite and island-arc tholeiite basalt compared with smooth pattern for mid-ocean ridge basalt (MORB). (Data from Drummond et al., 1996; Wilson, 1987; Sun and McDonough, 1989.) (b) Differential enrichment of a hypothetical MORB source (lowermost dotted line) by fluid solutions liberated from the subducted oceanic crust (dark shading) produces an enriched mantle peridotite arc source. Partial melts of this arc source (light shading) yield the island-arc tholeiite basalt magma in (a). (Redrawn from Davidson, 1996.)

depletion in Nb, Ta, and/or Ti constitutes a **negative Nb-Ta-Ti anomaly** typical of arc rocks.

The origin of the arc signature has been controversial. The depletion in Nb-Ta-Ti has been explained by retention of a refractory phase in the source in which those elements are highly compatible, such as rutile. These HFSEs may also be sequestered in residual hornblende in the source (Drummond and Defant,

1990). Many petrologists now believe that aqueous fluids play a significant role in the arc signature. LILEs have low ionic potential (Figure 2.20) and are readily dissolved and transported in aqueous fluids at high P and T (Tatsumi and Eggins, 1995), whereas Nb, Ta, and Ti are less mobile. A basic working model for arc-magma generation, always subject to future revision as is any model, involves partial melting of subarc peri-

dotite that has been metasomatically enriched in the more soluble ions (Figure 11.19b). Dissolved components might be from the slab with possible additions scavenged from the subarc wedge. Migrating silicate melts do not have these selective element preferences and therefore, if present, cannot dominate as metasomatizing agents.

Whether oceanic sediment contributes to arc magmas has also been controversial. However, significant amounts (to as much as 100 ppm) of boron (B) in arc rocks and the discovery of cosmogenic ^{10}Be (Section 2.6.3) in some of them have proved that at least some sediment is subducted (e.g., Leeman, 1996) and may be melted in some arcs. B is enriched in ocean-floor sediment and altered oceanic crust, whereas in mantle rocks it is <1 ppm. B is sequestered in clay and other phyllosilicate minerals and is quite mobile in aqueous fluids created as these minerals are heated in the subducting slab and liberate water. Be is like B in many ways and the near-uniformity of $^{10}\text{Be}/\text{B}$ ratios within a particular arc suggests that the two elements are governed by the same process that transfers matter from the slab to the mantle wedge. B abundances and $^{10}\text{Be}/\text{B}$ ratios decrease in arc rocks away from the trench, consistent with progressive scavenging of B and Be from the slab during its descent.

11.4.3 Partial Melting of Subducted Basaltic Oceanic Crust: Adakite

In the early days of the plate tectonic “revolution” it was believed by many geologists that melting of the subducted basaltic crust yielded the copious volumes of magmas manifested in this regime. However, it was soon realized that the overlying water-fluxed peridotite wedge provided a more viable magma source. Nonetheless, theoretical studies revealed a small “window” in P - T -time-composition space where partial melting of subducted basaltic crust might occur in young, hot slabs. Figure 11.18 indicates a range of P - T paths that subducting plates might take. Only the youngest and, therefore, hottest lithosphere descending rapidly, therefore inducing greatest shear heating, is expected to experience partial melting. At the other extreme, old lithosphere created at distant spreading ridges subducting at a slow speed without shear heating follows a path far below the T of even a water-saturated basalt solidus. The basaltic crust in which partial melting is likely to occur has been mostly dehydrated and converted to a high- P hydrous eclogitic assemblage of pyropic garnet, jadeitic clinopyroxene, and amphibole (Figure 5.11).

The search for candidate rocks in arcs where the subducting slab is <25 Ma and might represent partial melts of this amphibole-bearing eclogite assemblage has focused on **adakite**, named for Adak Island in the Aleutian arc. Adakite is basically a dacite, lo-

cally an andesite, having unusually high concentrations of Al_2O_3 (>17 wt.%), Na, Sr, and Eu, but low Mg, Ti, Nd, Y, Yb, and $^{87}\text{Sr}/^{86}\text{Sr}$ relative to the widespread andesite-dacite-rhyolite suite in subduction zones (Drummond et al., 1996). The high Sr and Eu suggest a lack of plagioclase in the source or no fractionation of this phase. The elevated light REE/heavy

Special Interest Box 11.4 Catalina schist: An exposed sample of the crust-mantle wedge interface

The discussion in Section 11.4 regarding hydration of the subarc mantle wedge, associated metasomatism, and partial melting in it and the basaltic crust is largely based on inference. However, exposures on Santa Catalina Island southwest of Los Angeles, California, are consistent with the nature of the inferred crust-mantle wedge interface and afford an opportunity for verification by real rocks (Sorensen, 1988).

The rocks show clear evidence for fluid migration, metasomatism, and partial melting, although mineral barometers indicate these processes occurred at shallower depths (approximately 30 km) than those generally inferred (Figure 11.18). The rocks are part of an accreted terrane in the Cretaceous forearc of coastal California formed where oceanic lithosphere was subducted beneath the continent (Figure 11.16) and subsequently uplifted to the surface. Structurally lowest rocks metamorphosed at lowest T are overlain by rocks metamorphosed at increasing T ; this thermally inverted (relative to a normal geothermal gradient) sequence is presumed to be a sample of the crust-mantle wedge interface (see inset diagram in Figure 11.16). Metamorphosed gabbros and overlying seafloor clay rocks represent a segment of the oceanic crust. Overlying these is what appears to be a segment of the mantle wedge that consists of metasomatized peridotite and enclosed amphibole-eclogite blocks of tholeiitic basalt composition. The metasomatized peridotite consists of combinations of enstatite, anthophyllite, tremolite, talc, and quartz that indicate addition of water and Si and loss of Mg from the initial rock. (Widespread lower- T serpentine replaces these minerals.) The high- T hydration metasomatism culminated in partial melting of the basaltic rocks, producing thin dikes, stringers, and larger pods of plagioclase + quartz + muscovite. Had the partial melting occurred at $T \sim 20$ kbar or more (>70 km; see Figure 11.18), partial melts would have been adakitic (trondhjemitic) in composition.

REE ratios, specifically, high La/Yb or La/Y, unlike lower ratios and flatter REE patterns of most arc rocks (Figure 11.19a), are consistent with residual garnet in a source chemically like MORB (Figures 2.22 and 2.23). Whether partial melting is water-saturated or water-undersaturated is controversial, but the former allows for lower, more realistic temperatures and for rutile stability that would account for the observed HFSE depletion in adakites (Prouteau et al., 1999). On the other hand, the negative Nb-Ta-Ti anomalies and overall spiky aspect of the trace element pattern may indicate that the slab-derived adakite magmas were overprinted by an arc signature en route to the surface, likely involving some mixing with wedge-derived melts.

Some phaneritic rocks known as **trondhjemite** have a similar chemical signature to adakite and may have solidified from similar magmas intruded into the crust. Trondhjemites occur locally in Phanerozoic subduction zones but are most widespread in cratons of Archean rock.

* 11.5 GENERATION OF ALKALINE MAGMAS IN METASOMATICALLY ENRICHED MANTLE PERIDOTITE

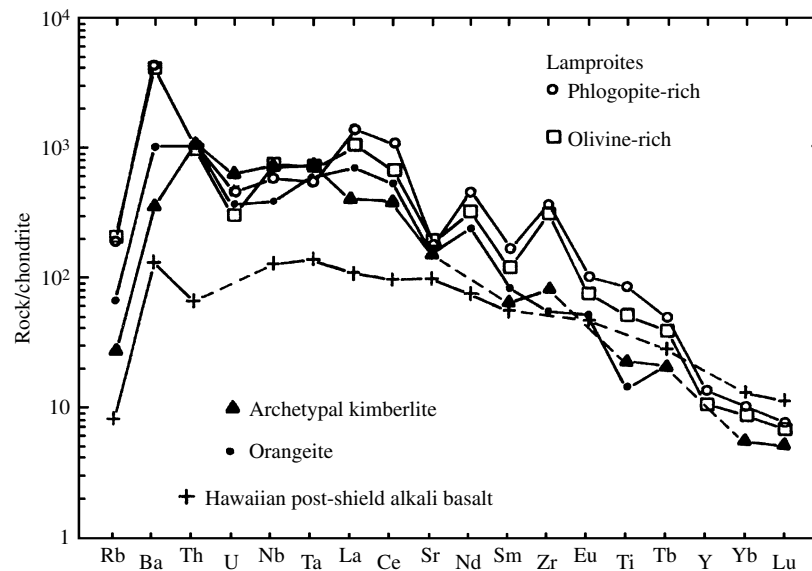
The origin of silica-undersaturated alkaline magmas has been debated for many decades. According to early hypotheses primary alkaline magmas were not generated in the mantle or lower crust; they were created from primitive subalkaline basaltic magmas via differentiation overprints. One proposal invoked conversion of silica-saturated magma to silica-undersaturated magma by assimilation of limestone—a hypothesis based on the eruption of leucite-rich magmas from

Vesuvius, Italy. These silica-undersaturated lavas carry abundant carbonate xenoliths picked up during their ascent through about 3 km of Mesozoic rocks. Silica metasomatism of the xenoliths created Ca-Si minerals, such as diopside, whereas complementary contamination and desilication of the host magma were believed to create silica-undersaturated magma. However, this theory failed to explain many aspects of alkaline magma generation on a global basis, not the least of which is the absence of enough limestone in the oceanic crust to produce widespread alkaline rocks of volcanic islands. Alkaline magmas in many tectonic settings, including continental rifts, have primitive chemical compositions that are demonstrably not the result of crustal contamination.

Yet another untenable hypothesized process is low-*P* crystal fractionation from subalkaline basaltic magma; residual melts are more silica-rich, not silica-undersaturated (Yoder and Tilley, 1962).

Thus, the basic necessity for alkaline magmas is a means of generating them as primary magmas, probably in the mantle because so many of the less evolved magmas are quite mafic, even ultramafic. Subsequently, these primitive alkaline magmas may be differentiated to alkaline daughter magmas, as in oceanic islands such as Tristan da Cunha in the Atlantic (Figure 2.16).

Silica-undersaturated primary magmas can be generated at relatively high pressures and CO₂ concentrations and especially by small degrees of partial melting of “normal” spinel or garnet peridotite. Small-degree (e.g., *F* ~ 1%) partial melts that are believed capable of segregating from their source (Mckenzie, 1985) are enriched in the most incompatible elements by factors of about 100 (Figure 11.15).



11.20 Chondrite-normalized trace element diagram for lamproites, archetypal kimberlites, and orangeites from Table 13.11 compared with oceanic island alkaline basalt from Table 13.2.

However, the enrichment of incompatible elements in some primitive magmas that carry mantle xenoliths, and hence probably suffered little modification from their source, is too extreme for an origin involving small degrees of partial melting to be valid. These magmas are represented by rare, highly alkaline, generally highly potassic, mafic, and ultramafic rocks found in continental rifts and stable cratons called kimberlites, lamproites, kamafugites, and phlogopite-rich minettes (Section 13.12). They commonly contain mantle-derived xenoliths and locally the first two are host to diamonds, indicating rather immediate transport from mantle depths. Normalized concentrations of the most incompatible elements in these rocks (Figure 11.20) are more enriched by about an order of magnitude than smallest-degree ($F = 0.1\text{--}1\%$) partial melts of a garnet lherzolite source (Figure 11.15).

11.5.1 The Metasomatized Mantle Connection

For most petrologists, a metasomatically enriched mantle is the most reasonable, or at the very least, a contributing, source for alkaline magmas and is likely the only source for highly alkaline kimberlites, lamproites, and kamafugites. The suite of ultramafic inclusions commonly occurring in these primitive rocks includes metasomatized mantle of the Al-Fe-Ti-rich clinopyroxene type that contains abundant phlogopite and K-rich richterite amphibole (1 and 2 in Table 11.2). Data in Table 11.3 indicate only modest enrichment factors (<10) between representative lamproite and metasomatic xenolith, which suggest that the latter could well be a source of partial melts of the former. A near-liquidus assemblage of olivine-diopside-phlogopite in a phlogopite-rich minette suggests a metasomatized mantle source (Esperanca and Holloway, 1987). Lloyd et al. (1985) used a common phlogopite clinopyroxenite inclusion from Uganda in the east African rift as a source rock and obtained, after 20–30% partial melting under mantle conditions in the laboratory, a melt composition similar to that of the highly alkaline rocks hosting such inclusions.

The highly variable and exotic chemical compositions that are so typical of alkaline rocks are explained by the **vein-plus-wall-rock melting mechanism** (e.g., Foley, 1992). Alkaline magmas are viewed as partial melts of metasomatic veins that are enriched in volatiles and incompatible elements hybridized with variable amounts of partial melts from their less enriched peridotite wall rock. Initial melting of a mantle source laced with metasomatic veins generates very enriched, highly alkaline melts from the veins, as in the experiments of Lloyd et al. (1985). Because of extensive solid solution in the vein minerals, before the vein can be completely melted, components from the relatively depleted wall rock are dissolved and dilute the incompatible elements from the vein, producing a wide spectrum of melt compositions. The extreme composi-

Table 11.2 Chemical Analyses of Al-Fe-Ti-Rich Clinopyroxene Inclusions and Vein

	1	2	3	4
SiO ₂	48.31	41.82	43.46	42.1
TiO ₂	0.45	2.58	1.99	3.2
Al ₂ O ₃	1.14	11.16	11.70	13.0
Fe ₂ O ₃	6.88t	10.46t	2.62	—
FeO	—	—	7.19	8.1t
MnO	0.11	0.15	0.20	—
MgO	38.52	16.07	15.28	15.6
CaO	0.60	12.14	15.20	10.9
Na ₂ O	0.26	2.56	1.38	2.8
K ₂ O	1.17	2.01	0.32	1.3
P ₂ O ₅	0.06	0.33	0.06	—
LOI	2.30	0.34	—	—
Total	99.80	99.62	100.58	97.29
Sc	2	25		
V	40	261		
Cr	1603	851		
Co	85	59		
Ni	1505	308		
Cu	10	41		
Zn	50	65		
Cs	0.29	0.4		
Rb	47	47		
Ba	154	1442		
Th	1.21	3.16		
U	—	1.4		
Nb	9	60		
Ta	0.82	4.18		
Ce	23.8	80		
Sr	116	747		
Nd	7.7	39		
Sm	0.87	7.93		
Zr	106	215		
Hf	3.05	5.15		
Eu	0.23	2.44		
Tb	0.08	0.89		
Y	4	19		
Yb	0.16	1.26		
Lu	—	0.18		

Column 1: Richterite (amphibole)-phlogopite peridotite, xenolith, Bultfontein, South Africa. (Column 1 data from Menzies et al., 1987.) Column 2: Phlogopite-apatite-amphibole xenolith, Pulvermaar, Eifel region, Germany. (Column 2 data from Menzies et al., 1987.) Column 3: Amphibole clinopyroxenite xenolith, Geronimo volcanic field, Arizona. (Column 3 data from Kempton in Menzies and Hawkesworth, 1987.) Column 4: Amphibole vein in composite xenolith, sample Ba-2-1, Dish Hill, California. Analysis includes 0.29% Cr₂O₃. (Column 4 data from Wilshire et al., 1988.)

tional variability of alkaline magmas may, thus, reflect a widely ranging ratio of vein/wall rock components dissolved in the melts, compounded by the compositional variability of the metasomatic veins themselves.

Table 11.3 Comparison of Incompatible Element Concentrations (ppm) in Average Spinel Peridotite (Table 11.1), Metasomatic Phlogopite-Apatite-Amphibole Rock (Table 11.2), and Phlogopite-Rich Lamproite (Table 13.11)

	D	SPINEL PERIDOTITE	METASOMATIC XENOLITH	LAMPROITE	ENRICHMENT FACTOR	
					LAMPROITE/ PERIDOTITE	LAMPROITE/ METASOMATIC XENOLITH
Rb	0.015	1.9	47	457	241	10
Ba	0.010	33	1442	10,607	321	7
Nb	0.020	4.8	60	147	31	2
K	0.010	8300	16,683	79,680	10	5
La	0.010	2.6		348	134	
Ce	0.010	6.3	80	629	100	8
Sr	0.020	49	747	1296	26	2
Ti	0.080	540	15,480	37,740	70	2

D is the approximate bulk partition coefficient for a peridotite-basalt system (Table 2.5), enrichment factors are ratios of indicated concentrations.

Generation of alkaline magmas is intimately related to a metasomatically enriched lithospheric mantle (e.g., Tingey et al., 1991). Subcontinental mantle lithosphere has been separated from the underlying convecting asthenosphere since probably as long ago as the Archean-Proterozoic transition (2500 Ma), whereas in suboceanic realms it has only been separated since about 200 Ma at most. An enriched lithosphere accounts for two additional characteristics of alkaline rocks: highly variable radiogenic isotopic ratios and occurrence of the most highly alkaline rocks in continental settings. Because of relative compatibilities (Section 2.6.2), metasomatically enriched peridotite has higher Rb/Sr and lower Sm/Nd ratios. The time-integrated effect of this metasomatism is, therefore, to increase $^{87}\text{Sr}/^{86}\text{Sr}$ and decrease $^{143}\text{Nd}/^{144}\text{Nd}$. The highest $^{87}\text{Sr}/^{86}\text{Sr}$ and lowest $^{143}\text{Nd}/^{144}\text{Nd}$ are in lamproites and orangeites that occur on Precambrian cratons (see Figure 13.42), whereas alkaline rocks of oceanic islands, such as Hawaii, are less radiogenic because the elapsed time since metasomatic enrichment of the lithospheric source is less. Small-volume alkaline magmas typically develop in intraplate settings and continental rifts where upwelling mantle, locally plumes, raise temperatures in the enriched lithosphere or cause it to decompress through uplift and tectonic thinning. Either or both perturbations (Figure 11.2) lead to partial melting.

* 11.6 MAGMA GENERATION IN THE CONTINENTAL CRUST

Vast volumes of felsic magmatic rocks composed mostly of two feldspars and quartz occur in continental arcs (e.g., western North and South America) but are only of trivial volume in oceanic island arcs (e.g., west-

ern Aleutians and Tonga). This simple fact emphasizes the need for preexisting continental crust in the generation of voluminous felsic magmas because the essential difference between these subduction zones bordering the Pacific Ocean is the nature of the crust in the overriding lithosphere. Because feldspars and quartz are near-liquidus phases in most felsic systems at crustal and upper mantle pressures (Figure 5.25), felsic magmas cannot be generated from peridotite in which olivine and pyroxenes are near-solidus phases. In addition, large volumes of felsic magma are not created by differentiation of vast amounts of basalt magma in oceanic areas.

Significant felsic magma generation can occur in three continental regimes, including continental arcs, where additional heat and possible fluxing aqueous solutions are available to perturb an otherwise subsolidus status in the lower crust. These three regimes are listed in order of decreasing volume of magma generation (Figure 11.2):

1. Continental arcs overlying subducting oceanic lithosphere where large volumes of ascending high- T mafic magma rising from the underlying mantle wedge are buoyantly blocked in and/or underplate the lower crust; additional thermal energy for partial melting is produced by this heat source (Section 11.1.1)
2. Continental rifts or areas above rising mantle plumes where underplating mantle-derived basalt magma provides additional heat for melting
3. Thickened crust in continent-continent collision zones, such as the Himalayan, heated by adjustments to the geothermal gradient

Magma generation might also occur as continental rock just below its subsolidus is decompressed

as a result of rapid isostatic uplift and erosion or is fluxed by water liberated from nearby rocks experiencing metamorphic dehydration reactions, in (3), or from crystallizing mantle-derived magmas, in (1) and (2).

Many of the concepts of magma generation by partial melting for peridotite-basalt systems are applicable to continental rock-granite systems. However, the deep continental source is far more heterogeneous with respect to modal, mineral, and chemical composition than the mantle peridotite source, even allowing for its variable metasomatism. Moreover, silicic partial melts are cooler and orders of magnitude more viscous, making melt-residue segregation difficult, if not impossible, in reasonable geologic time scales. Consequently, there is greater opportunity for overprinting diversification processes to modify magmas slowly rising from their source. Source heterogeneity makes trace element modeling of the degree of melting and composition of the source rock more difficult and ambiguous (Harris and Inger, 1992).

11.6.1 Partial Melting of Continental Source Rocks

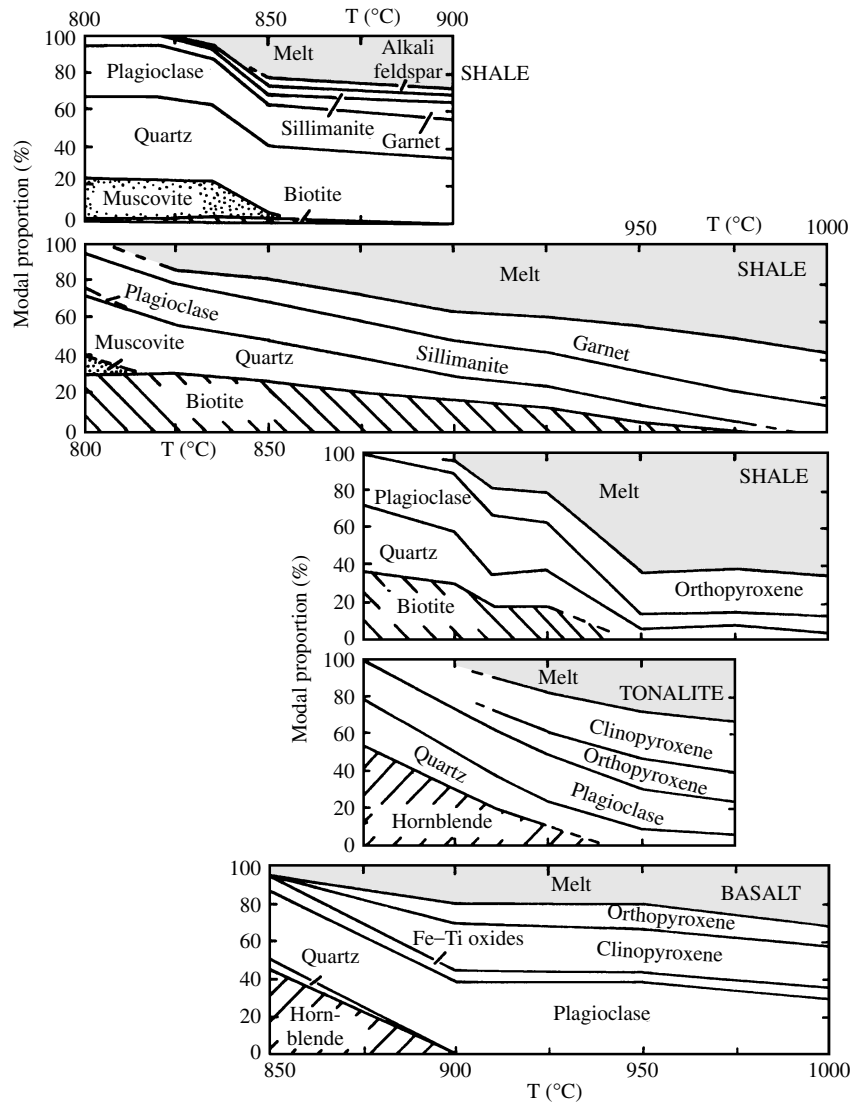
The continental crust is compositionally heterogeneous on most scales. Original rock types in the crust include a variety of sandstones, shales, and carbonate rocks plus rhyolite, dacite, andesite, basalt, and their phaneritic equivalents. All of these potential source rocks are metamorphosed because of deep burial and elevated temperatures where partial melting might take place. Constituent minerals are highly variable proportions of plagioclase, alkali feldspar, quartz, amphibole, micas, and generally lesser amounts of Fe-Ti oxides, carbonates, garnet, Al_2SiO_5 polymorphs, chlorite, epidote, and many other, less common metamorphic minerals. Among the major elements, most of the Mg, Fe, and considerable amounts of Al, Ca, and Na are sequestered in amphibole and to some extent in biotite, which together with muscovite harbors K. Significantly, these three minerals are hydrous. Accessory minerals, such as zircon and apatite, contain relatively large concentrations of trace elements, especially REEs.

Partial melting, sometimes called **anatexis**, of continental rocks generating felsic magmas can theoretically occur in strictly anhydrous systems that lack water as a separate phase as well as hydrous minerals. However, because solidus temperatures are relatively high, such melting is limited. But another reason why anhydrous melting rarely occurs is the widespread occurrence in potential source rocks of hydrous minerals. Partial melting of rocks that contain hydrous minerals in the absence of a separate aqueous fluid phase is referred to as **dehydration melting** (also called **fluid-absent, water-deficient, water-undersaturated, low-water fugacity, or low-water activity melting**). The only water in dehydra-

tion melting is chemically bound within micas and amphiboles, or other usually minor or accessory hydrous minerals, such as epidote and apatite. Partial melting of a source rock containing hydrous minerals results in their decomposition and liberation of water, together with other mineral components, forming a hydrous melt. Quartz, feldspar, and other anhydrous minerals are dissolved in this melt. A complementary less hydrous or anhydrous crystalline residue coexists with the melt. Partial melting may also occur, generally at relatively lower temperatures, under water-saturated conditions (also called **water-excess** or **high-water activity melting**). This extraneous water can be derived from, for example, nearby decomposing hydrous minerals below solidus temperatures or from crystallizing mantle-derived magmas that become water-saturated. There is no doubt that *water—in one form or another—is essential to generation of large volumes of felsic magma in the continental crust.*

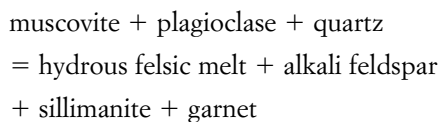
Source-Rock Fertility and Melt Fraction. Because felsic melts are highly viscous, small melt fractions cannot readily segregate from the crystalline residue in geologically reasonable times (McKenzie, 1985). Also, small melt fractions cannot provide sufficient buoyancy to mobilize the entire melt-crystal mass so it can ascend, for example, as a magma diapir. Extensional tectonic environments or other circumstances may facilitate magma movement. A fundamental constraint in felsic magma generation in the deep continental crust is that large melt fractions, perhaps >20%, are necessary to provide for a buoyant volume sufficiently large that it is capable of ascending toward the surface. This contrasts strongly with the minimum melt fractions (1% or so) for segregation of basaltic melts. Therefore, source-rock fertility and intensive variables that enhance melt productivity are of paramount importance in creating a mobile mass of felsic magma capable of rising through the crust. **Source-rock fertility** is the potential amount of components available to yield a melt, in this case of felsic composition. For example, one way to compare relative fertilities is to compare the amount of minimum-*T* granite components (Figures 5.24–5.26; subequal proportions of normative quartz, orthoclase, and albite) in potential source rocks. In Hawaiian tholeiitic basalt (Table 13.2) it is ~7 wt.%, whereas in diorite (Table 2.2) it is ~31 wt.%.

Figure 11.21 shows melt fractions and modal compositions of crystalline residues resulting from dehydration melting of metamorphosed clay rocks (shales) and mafic rocks under presumed equilibrium conditions. These two rocks represent widespread sedimentary and magmatic rock compositions, respectively, in the continental crust. Muscovite-rich, metamorphosed shale source rocks begin melting at 800–825°C by

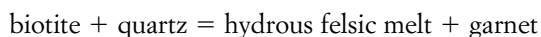


11.21 Experimental dehydration partial melting of metamorphosed shale, tonalite, and basalt source rocks. Top four panels at 10 kbar (about 37-km depth). (Redrawn from Patiño Douce and associates, e.g. Patiño Douce and Harris, 1998.) Bottom panel at 7 kbar (Redrawn from Beard and Lofgren, 1991).

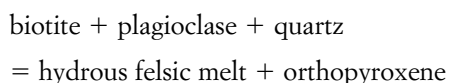
breakdown of muscovite. By 850°C, all of the muscovite has been consumed, producing about 20% hydrous melt, by the reaction



In metamorphosed shales containing biotite, this phase decomposes over a wider and higher range of T than muscovite, yielding steadily increasing melt fractions. Two reactions are

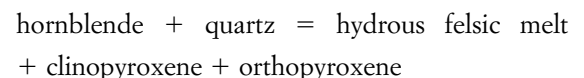


and

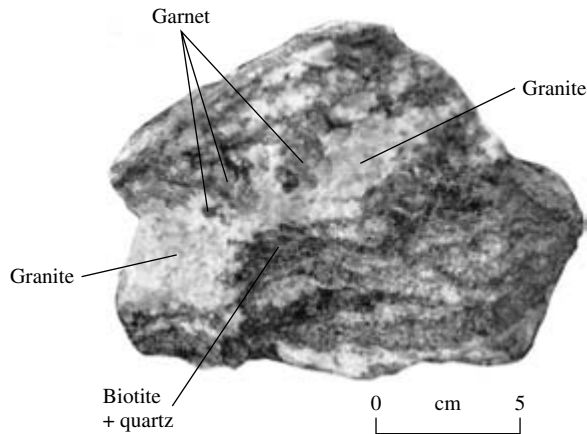


An example of the first reaction is shown in Figure 11.22.

Muscovite-free source rocks that contain abundant biotite or hornblende begin melting near 875–900°C, consuming them by 950°C. Hornblende breaks down by the reaction



Significantly increased melt productivity is produced by dehydration melting of an intimate mixture of a two-mica rock and a tonalite (Pl + Hbl + Qtz) compared to the melt fractions generated by melting each source rock alone. The enhanced fertility of a composite source, which may be quite realistic for the heterogeneous continental crust, apparently occurs because K migrates into the tonalite from the two-mica



11.22 Hand sample of pelitic **migmatite**. Dehydration melting of biotite and dissolution of quartz yielded garnet and hydrous felsic melt that is now crystallized to alkali feldspar and quartz.

rock and Na in the reverse direction, creating a composite source composition nearer the thermal minimum in the granite system (Figures 5.24–5.26). Granitic rocks and feldspathic sandstones can potentially yield large volumes of partial melt for the same reason, although dehydration melting of the generally minor amounts of biotite and amphibole in these rocks may not supply much water to flux dissolution of the felsic minerals.

Melt Composition. Incongruent felsic partial melts are generally relatively enriched in Si, Na, K, and water and depleted in Ti, Mg, Fe, and Ca relative to the source rock. These melts may be silica-oversaturated with normative Q , even though the source rock may be barely silica-saturated or even silica-undersaturated. Partial melts are peraluminous, metaluminous, or even peralkaline (Section 2.4.4), depending on the composition of the source, intensive parameters, and degree of melting. These same conditions dictate the composition of the crystalline residue, which in turn influences melt composition; for example, melts in equilibrium with residual garnet or plagioclase are less aluminous than if these Al-rich minerals are entirely dissolved and are depleted in either heavy REE (garnet) or Eu (plagioclase) because these trace elements are compatible.

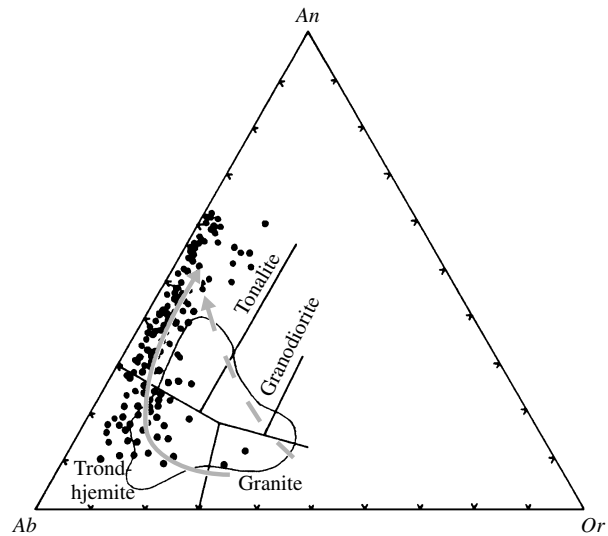
Partial melting of source rocks that contain two feldspars and quartz just above their solidus predictably generates melts of granite composition at the thermal minimum in the granite system (Figures 5.24–5.26). Source rocks with only two, one, or none of these three crystalline phases yield partial melts farther from the minimum composition.

Partial melting of basaltic rocks at $P = 5\text{--}32$ kbar under water-deficient conditions so that amphibole remains stable generates granite melts near the solidus, where melt fractions are less than a few percent but trondhjemite for melt fractions near 10%.

At still higher degrees of melting, tonalite compositions are generated, essentially bypassing granodiorite (Figure 11.23). In contrast, progressive partial melting of basaltic rocks over about the same range in P but water-saturated conditions generates granite, granodiorite, and finally, at highest degrees of melting, tonalite, bypassing trondhjemite.

11.6.2 “Alphabet” Granitic Magmas: Contrasting Sources

Contrasting Devonian granitoid batholiths and minor associated felsic volcanic rocks in the Lachlan fold belt of eastern Australia are postulated to have had two distinctive sources (Chappell and White, 1992). **S-type granites** were derived from magmas generated by partial melting of sedimentary source rock composed in part of metamorphosed clay minerals. Weathering of feldspars and other rock-forming minerals at the surface of the Earth yields chemically differentiated products, chiefly Na in seawater, Ca and Sr in limestones, and Al-rich clay minerals in shales. Consequently, partial melting of metamorphosed shales that contain aluminous micas, Al_2SiO_5 polymorphs, cordierite, garnet, and so on (Figure 11.21, top three panels), generates strongly peraluminous magmas (Section 2.4.4). With >1 wt.% normative corundum, S-type rocks contain Al-rich minerals, including one or more of andalusite, sillimanite, garnet, tourmaline, cordierite, and muscovite in addition to biotite, two feldspars, and quartz.



11.23 Experimentally generated partial melts of basaltic source rocks at 5–32 kbar plotted in terms of normative feldspar components (Figure 2.9). Full-line shaded arrow shows that smallest-degree dehydration partial melts (water-deficient conditions) are granite but more advanced melting generates trondhjemite and then tonalite melts. Water-saturated partial melts follow dashed shaded arrow with advancing degree of melting and are never trondhjemite. Outline encloses Archean tonalite-trondhjemite-granodiorite (TTG) suite. (Data from Rapp, 1997; Springer and Seck, 1997.)

Ilmenite rather than magnetite is the typical Fe-Ti oxide. Inclusions (enclaves) are of the same aluminous minerals but have higher concentrations of mafic phases. Whole-rock silica is generally >63 wt.%.

In contrast, **I-type granites** are believed to have been derived from magmas generated by partial melting of mafic and intermediate igneous rocks (Figure 11.21, bottom two panels). These mostly metaluminous granites contain more Na and Ca than S-type granites, which stabilize amphibole and commonly titanite, in addition to possible biotite, alkali feldspar, and quartz. Cogenetic rocks may be as mafic as gabbro, having as little as about 50 wt.% silica. Mafic inclusions are predominantly amphibole with plagioclase and lesser amounts of biotite, titanite, and clinopyroxene.

The postulated contrasting sources in the Lachlan fold-belt rocks spawned intense research on and debate about these and other granitoid rocks worldwide (e.g., Zen, 1988). Despite the important new focus on felsic magma sources, attempts to apply the eastern Australian S- and I-type labels elsewhere have caused considerable confusion (Miller et al., 1986). Their use implies that the source is known, but in practice this information is generally not verifiable. S-type is *not* synonymous with peraluminous because peraluminous magmas can also be generated from igneous source rocks. In addition, the composition of virtually every granitic rock reflects not only its source rock but a host of overprinting differentiation effects that acted on the primary magma. Collins (1996) finds that the Lachlan rocks have three distinct Sr-Nd-Pb-O isotopic sources: Mantle-derived mafic magma, late Proterozoic mafic arc rocks of the lower crust, and midcrustal metasedimentary rocks. I-type magmas were produced by mixing of the first two and S-type magmas by contamination of I-type magmas with metasedimentary partial melts.

A-type granites are a third, diverse group of felsic rocks typically created in **anorogenic** areas and having diverse sources and origins (see Section 13.9).

11.6.3 Crystalline Residues

Partial melting of mica-amphibole-bearing metamorphic source rocks in the deep continental crust generates complementary hydrous felsic melts and refractory mafic to ultramafic crystalline residues.

In contrast with the simple olivine-pyroxene residue in mantle sources, continental residues are mineralogically heterogeneous and complex. Unlike mantle residues, new crystalline phases may be created during partial melting of felsic continental rocks (Figure 11.21). Many trace elements are not dispersed among major minerals as in peridotitic systems, for which equations such as 11.3 and 11.4 apply, but instead are sequestered in accessory minerals such as zircon, and monazite, whose stabilization depends upon sufficient activities of elements normally present in trace concen-

trations (e.g., Zr and Ce). These accessories may or may not be stable phases in the residue.

Three potential residues of partial melting have been discerned in deeply eroded roots of mountain belts and as xenoliths of deep-seated rock in volcanic rocks. Metamorphic rocks of the granulite facies consist of an anhydrous assemblage of plagioclase + pyroxene \pm quartz \pm garnet \pm sillimanite \pm cordierite (compare Figure 11.21). **Migmatite** consists of layers, pods, and irregularly shaped masses of granite that are mingled with mafic metamorphic rock made largely of plagioclase, amphibole, biotite, pyroxene, and other refractory minerals (Figures 11.22 and 11.24). Migmatites are commonly interpreted to be partially melted rock in which the granitic partial melt has not segregated from the complementary residue. Some petrologists believe migmatites represent residues of production of granitoid plutons, but other possibilities exist. Garnet pyroxenites and eclogites, some containing amphibole, are the third residue.

Another possible relation between complementary partial melt and crystalline residue is represented in inclusions (granular aggregates) and single isolated crystals collectively called *restite* by Australian petrologists (e.g., Chappell et al., 1987). **Restite** crystals and inclusions are pieces of the residue caught up in a larger volume of melt. The whole mass forms a mobile body of magma that has sufficient buoyancy to ascend, perhaps diapirically, from the source into shallower crust. Magmas containing at least 60% restite might represent en masse mobilization of virtually the entire source volume, whereas magmas containing less restite might reflect some degree of melt-residue segregation.

Although intuitively reasonable, to what extent the restite concept actually applies to real felsic magmas is quite controversial and several questions may be posed. How much of a real felsic rock is restite? Zircon grains, commonly complexly zoned and having ages greater



11.24 Outcrop of deformed **migmatite**. Lower Proterozoic metamorphic belt, Tolstik Peninsula, Karelia, Russia. Felsic rock is concentrated in former shear zones. Camera lens-cap for scale. (Photograph courtesy of Michael Brown.)

than the age of the host magmatic rock, can be accepted as refractory restite, provided a xenocrystic derivation can be ruled out. But what of embayed calcic cores crowded with other mineral inclusions in plagioclase grains? Are these cores, and like ones occurring in other minerals, partially resorbed restite on which rim material was precipitated from the crystallizing melt? Or are they a result of mixing of magmas of contrasting composition? Are mafic inclusions of essentially the same minerals as the host rock, but in different proportions, pieces of restite equilibrated with the melt fraction? Or are they cognate inclusions ripped from the early crystallized margin of the magma chamber, equilibrated xenoliths of wall rock, recrystallized mantle-derived basaltic magma that powered the felsic magma system, or pieces of mixed and crystallized magma? Arguments in favor of the restite concept include the reluctance of restite material to segregate from melt of high apparent viscosity and the opportunity for restite grains to serve as sites for heterogeneous nucleation; widespread inclusions of relatively refractory apatite and zircon crystals in biotite and hornblende may be examples.

It should be noted that huge volumes of complementary mafic residues should underlie batholiths of felsic rock in continental mountain belts. However, geophysical evidence for these dense batholithic roots is often ambiguous. But Ducea and Saleeby (1998) present evidence from garnet pyroxenite xenoliths in Miocene volcanic rocks for a 70-km-thick keel of mafic-ultramafic rocks underlying the 30-km-thick Sierra Nevada, California, granitoid batholith (Figure 9.16; see also Section 13.7.2).

11.6.4 Melt Segregation

At least three factors that influence segregation of partial melts from residue can be identified:

1. Volume of melt produced: This is related to source fertility and melting temperatures. In pelitic sources, modest melt fractions, say, <10–20%, from muscovite breakdown may produce granite segregations in migmatite, but larger melt fractions related to biotite and/or hornblende breakdown may be required for complete melt segregation from the source rock or en masse mobilization.
2. Melt process: Dehydration melting produces a larger volume increase than water-saturated melting and so is more effective in creating porosity in the source rock.
3. Deformation of the source rock during melting. Granite segregations are commonly found along shear zones (Figure 11.24) in deformed migmatites where mobile melt might have been drawn into local lower-pressure sites by dilatancy pumping (Section 8.2.3). It is difficult to be sure of cause and

effect in such features because the shearing may be focused in the parts of the rock that contain granite melt.

11.6.5 Felsic Magma Generation and the Mantle Connection

The large volume of felsic magma in arc batholiths and ignimbrite fields (see Section 13.7) demands a significant advective input of heat from the mantle into continental source rocks to accomplish partial melting (Section 11.1.1). Production of granite magma by extreme fractional crystallization of basaltic magmas does locally occur, but the proportion (<10%) is too little to account for the observed volume, unless there are gigantic hidden volumes of complementary mafic-ultramafic rocks. In deep crust already at near-solidus temperatures of perhaps 500–600°C, heat transferred from arrested intrusions and underplatings of mantle-derived basaltic magma at temperatures to as much as about 1400°C makes *partial melting thermally unpreventable*. Added to this heat input is the fluxing effect of volatiles liberated from these magmas as they become fluid-saturated during crystallization.

Evidence for a thermochemical mantle connection to felsic magma generation includes the following:

1. Isotopic compositions of many felsic rocks that indicate a significant mantle component derived from basaltic magmas that hybridize with partial melts of continental rock
2. Thick, tilted sections of lower crustal rock exposed in orogens that reveal kilometer-scale basaltic magma underplatings (e.g., Sinigoi et al., 1995)
3. Fragmented synplutonic mafic dikes (Figure 11.25)
4. Blobs of mafic rock in felsic hosts (Figure 8.25), which resemble shapes of lava pillows formed in water and suggest that hotter mafic magma invaded and quenched against cooler felsic magma

Widespread mafic inclusions (Figure 7.48) hosted in granodiorite, tonalite, and dioritic plutons (Didier and Barbarin, 1991) have different possible origins, i.e., are polygenic, but some are derived from (3) and (4).

SUMMARY

Global magma generation is a manifestation of a convecting mantle in a cooling still-hot Earth. The mantle and underlying core are the source of the thermal energy. Directly or indirectly, the mantle provides most of the mass for magma production. Convection in the cooling mantle via plumes and moving lithospheric plates allows magma generation by perturbing the P , T , and composition of potential solid source rock in the upper mantle as well as the lower crust. Incongruent



11.25 Fragmented synplutonic mafic dike. Pativilca pluton, Llanachupan, Pativilca Valley, Ancash, Peru. Height of cliff approximately 250–300 m. Dike intruded almost crystallized magma, which then became mobilized, disrupting the mafic sheet. (Drawing by K. Lancaster; Reproduced with permission of W. S. Pitcher from Pitcher, WS, 1997. *The nature and origin of granite*. New York, Chapman & Hall.)

partial melting generates melts enriched in lower-melting- T components and incompatible elements, leaving a more refractory crystalline residue. Extraction of basaltic partial melts from the peridotitic mantle over billions of years of geologic time is responsible for the differentiation of the crust from the mantle and continent growth.

The upper mantle is mostly Cr-diopside peridotite that has been locally metasomatized and veined by C-O-H fluids and silicate and carbonatite melts. These metasomatizing liquids have created assemblages of Al-Fe-Ti-rich pyroxenes, amphiboles, micas, apatite, and other materials, thereby enriching the relatively infertile, depleted peridotite in volatiles and incompatible elements. From this heterogeneous, variably enriched mantle a wide compositional spectrum of primary magmas has been generated by variable degrees of melting over a range of depth and volatile concentrations. Primary magmas range from ultramafic (komatiite, picrite, kimberlite) to mafic (alkaline and sub-alkaline basaltic) to intermediate (basaltic andesite, andesite). Tholeiitic basalt magmas generated by partial melting of upwelling decompressing mantle at the ocean ridge system constitute the largest volume of magma on Earth. Smaller volumes of more alkaline magmas are generated by lesser degrees of melting, likely of enriched source rocks. Highly alkaline magmas likely represent partial melts of metasomatic veins with little but variable wall rock contribution. Primary basaltic and andesitic magmas are generated by water-fluxed peridotite in subduction zones.

Source rocks in the continental crust are more heterogeneous than mantle rock. In addition to abundant feldspars and quartz, the presence of amphiboles, micas, and many aluminosilicates makes partial melt-

ing more complex. Melts and crystalline residues are more variable. Most melting probably occurs under water-deficient (dehydration) conditions in which the only water in the source-rock system is bound in hydrous minerals. Partial melt compositions range from granite to granodiorite to tonalite. These melts are highly viscous and are unlikely to segregate from their residuum as easily as more mafic mantle melts. Partially melted, buoyant crustal sources may rise en masse as diapirs.

CRITICAL THINKING QUESTIONS

- 11.1 Discuss the relative volumes of magma generated in the Earth by perturbations in P , T , and volatile concentrations (see Figure 1.1). Is increase in T responsible for the largest volume of magmas?
- 11.2 Critique the effectiveness of decompression melting of uplifted and eroded continental crust in comparison with decompression melting of upwelling mantle at ocean ridges and in plumes. What is the principal factor that allows melting to occur?
- 11.3 If T increases steadily into the Earth, why aren't magmas generated more abundantly at greater depths?
- 11.4 Color is generally an unsound basis for classification of rocks. Nonetheless, the two principal types of mantle-derived inclusions in basaltic rocks have strikingly different colors that can be used in identification. Discuss and justify the rationale for this distinction.
- 11.5 Explain how Al-Fe-Ti-rich veins can be the source of megacrysts and inclusions made of the same mineral assemblages.

- 11.6 Why are incompatible elements sequestered in some mantle minerals (amphiboles, clinopyroxenes) and not others (olivine)?
- 11.7 Why is there an “inverted” thermal gradient from the subducting crust into the overlying mantle in the area of the inset diagram in Figure 11.16?
- 11.8 Review the possible compositions of primary magmas that can be generated from mantle sources and indicate the conditions under which they form.
- 11.9 The oldest known alkaline rocks on Earth are about 2.7 Ga, whereas older rocks to as much as 4.0 Ga include tholeiitic basalts and a variety of silica-saturated felsic rocks. Suggest one reason for the apparent absence of alkaline magmas in the young Earth.
- 11.10 Discuss the differences between primary and primitive mantle-derived magmas. How can they be accurately identified?
- 11.11 Contrast partial melting in lower continental crust and upper mantle.
- 11.12 What distinctive major- and trace-element signatures, if any, might be used to distinguish dehydration partial melts of muscovite-bearing from hornblende source rock?
- 11.13 Explain why S-type granites typically have ilmenite rather than magnetite and have low $\text{Fe}_2\text{O}_3/\text{FeO}$ ratios relative to I-type. (*Hint*: Is there something in shale source rocks that would cause their magmas to have low oxygen fugacity?)
- 11.14 How might juvenile primeval versus recycled surface water be distinguished in mantle minerals?
- 11.15 Section 11.2.1 indicated that eclogite xenoliths from the mantle might have two origins. How might these be distinguished in a particular xenolith?
- 11.16 Propose an alternate interpretation of the mafic blocks in granitoid pluton shown in Figure 11.25. What evidence might you seek to decide between your interpretation and that cited in the figure caption?
- 11.2 Estimate the amount of water cycled annually into the mantle via subducting oceanic crust that is entering the subduction zone at a rate of 5 cm/y. Compare this value with the amount of water exsolved from solidifying mantle-derived magmas intruded into and extruded onto the crust. Show all calculations and indicate and justify the numbers used. Discuss your results.
- 11.3 Draw the $\text{Mg}_2\text{SiO}_4\text{-SiO}_2$ binary phase diagram for 20-kbar anhydrous and for 20-kbar water-saturated from Figures 11.8b and 11.10. Compare with Figure 5.8. Discuss whether enstatite melts congruently or incongruently at 20 kbar.
- 11.4 Compare equilibrium and fractional partial melting of spinel lherzolite (Figure 11.9) by making diagrams in which T is plotted on the vertical axis and melt fraction on the horizontal axis. Indicate compositions of melts and crystalline residues at critical points in the diagrams. Discuss contrasts in these two styles of partial melting.
- 11.5 Partially melt, in both equilibrium and fractional manners, a rock composed of 30% diopside, and equal amounts of quartz and enstatite in the system in Figure 11.9. Describe in detail the compositions of crystalline residues and melts and plot their paths in the ternary.
- 11.6 Compare the time a basaltic melt would take to ascend buoyantly 100 km via porous flow through mantle peridotite versus through a dike 1 cm in width (see equation 9.1). Compare these times with the time for porous flow of an aqueous fluid rising 100 km above a subducting oceanic crust. Indicate all assumptions made and values of parameters chosen. Discuss the implications of your calculated times.
- 11.7 Write a mineralogical reaction whereby lherzolite is metasomatized by silica-bearing aqueous fluid to an assemblage of enstatite + anthophyllite + tremolite + talc + quartz. See Special Interest Box 11.4.

PROBLEMS

- 11.1 The minimal amount of water in the upper mantle to a depth of 670 km may be roughly