

# Chapter 5

## Basalts and Mantle Structure

### 5.1 Introduction

Basalts are the most common rock type on the surface of Earth. The oceanic crust, which covers more than 70 percent of the surface of Earth, is composed of basalt and its intrusive equivalent, gabbro. Basalts dominate the rocks on oceanic islands and are also widespread on the continents. One of the major petrologic discoveries in the twentieth century was that basalts are partial melts of the mantle (Green and Ringwood, 1969). With this insight, basalts became more than simply interesting volcanic rocks: they took on significance as probes of the mantle. The chemistry of basalts, including their major and trace element compositions as well as their isotope geochemistry, provides direct evidence about the nature and composition the mantle that is difficult to obtain by other means. This chapter describes the petrology of basalts, the structure and composition of the mantle from which they are derived, and the various processes by which the mantle may partially melt to form basaltic magmas.

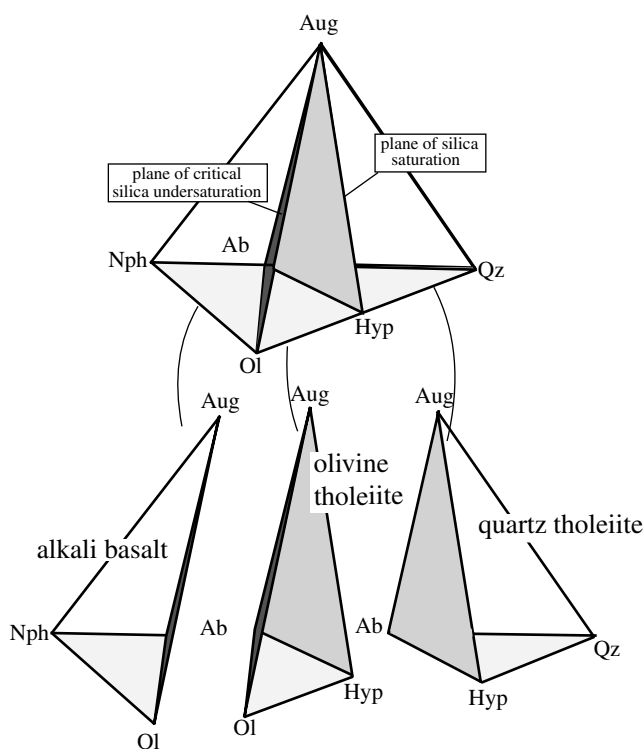
## 5.2 Basalt Petrology

### 5.2.1 Classification

Because basalts are typically fine-grained to glassy rocks, the most common classification is based upon normative (as opposed to modal) mineralogy. One of the best ways to visualize basalt chemistry is by use of the basalt tetrahedron (Yoder and Tilley, 1962) (Figure 5.1). The basalt tetrahedron has the apices of normative augite (Aug), quartz (Qz), nepheline (Nph), and olivine (Ol). Normative albite (Ab) plots one third of the way toward Nph from Qz and normative hypersthene (Hyp) plots midway between Ol and Qz. Because Hy never coexists with Nph, the Ol-Ab-Aug plane is always present. This plane is called the **plane of critical silica undersaturation** and separates Hyp-normative bulk compositions from Nph-normative bulk compositions. The Nph-normative basalts are called **alkali basalts**; the Hyp-normative basalts are called **tholeiites**. Because the Mg-rich olivine found in basalts never coexists with quartz, the Hyp-Ab-Aug plane is also important to basalt petrology. This plane is called the **plane of silica saturation** and it separates quartz-saturated tholeiites (i.e., **quartz tholeiites**) from olivine-saturated tholeiites (i.e., **olivine tholeiites**).

### 5.2.2 Chemistry and Petrography

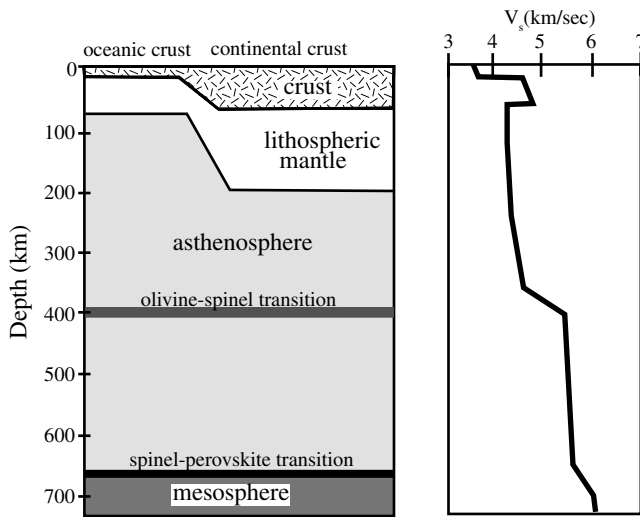
The normative differences between the two basalt types reflect their subtle differences in chemistry. As their name implies, alkali basalts are richer in alkalis ( $\text{Na}_2\text{O}$  and  $\text{K}_2\text{O}$ ) and poorer in CaO than are tholeiites. As such, they plot in the alkalic or alkali-calcic fields of Peacock (1931), whereas tholeiites are typically calcic or calc-alkalic. Important, alkali basalts have slightly lower silica contents than tholeiites (46–48 percent compared to 48–52 percent). Because alkali feldspars (for example, albite,  $\text{NaAlSi}_3\text{O}_8$ ) contain more silica than calcic feldspars ( $\text{CaAl}_2\text{Si}_2\text{O}_8$ ), crystallization of alkali feldspar will deplete silica from a rock more effectively than will crystallization of calcic feldspar. Thus the combination of relatively high alkalis and low silica explains why alkali basalts are nepheline normative. In addition to the differences in major element abundances, alkali basalts also tend to be richer in incompatible elements than tholeiites. As we noted in Chapter 4, incompatible elements are elements incompatible with crystallizing silicates but compatible with melt. These elements, such as  $\text{TiO}_2$ ,  $\text{Fe}_2\text{O}_3$ , and rare earth



**Figure 5.1** The basalt tetrahedron showing the differences in normative composition between alkali basalt, olivine tholeiite, and quartz tholeiite. After Yoder and Tilley (1962).

elements, are likely to be concentrated in the magma as it crystallizes or are likely to be the first elements to enter a melt when melting begins.

The chemical differences between tholeiites and alkali basalts are reflected in the following petrographic differences. Tholeiites typically contain olivine only as a phenocryst. The olivine commonly shows signs of resorption or reaction to pigeonite or hypersthene. Groundmass phases include pyroxenes and plagioclase. The augite in tholeiites is typically colorless, indicating that it is poor in ferric iron and titanium. Some quartz tholeiites may contain groundmass quartz or vesicles lined with a silica mineral, although in many quartz tholeiites the excess silica will be hidden in its glassy matrix. Alkali basalts contain olivine as both phenocrysts and groundmass. The augite tends to be pleochroic because it contains small amounts of ferric iron and titanium. There may be a late-stage alkali feldspar, often anorthoclase, in the groundmass. In most alkali basalts the normative nepheline is hidden in the residual glass, however, if the basalt is very alkalic, nepheline may appear in the groundmass. Such a basalt is called a **basanite**.



**Figure 5.2** The major layers of the crust and mantle, along with characteristic S-wave velocities for each layer.

As depicted in Figure 2.9, in the system nepheline – silica, albite is a thermal barrier. Melts on the silica side of the albite composition evolve to a quartz-bearing eutectic, whereas melts on the nepheline side evolve to a nepheline-bearing eutectic. This behavior extends to more complex silicate systems as well. Those melts that lie to the nepheline side of the olivine-albite-diopside plane in Figure 5.1 (i.e., alkali basalts) differentiate toward Nph-saturation, whereas those melts on the hypersthene side (i.e., tholeiites) differentiate toward Hyp- or Qz- saturation. As a result, alkali basalts and tholeiites follow very different differentiation paths. During differentiation, alkali basalts evolve to form nepheline-bearing rocks, such as phonolites or their plutonic equivalents, nepheline syenites (i.e., the rocks on the lower half of the IUGS diagrams shown in Figures 1.1 and 1.4). Tholeiites, in contrast, evolve toward silica saturation, forming residual magma with trachytic or rhyolitic composition.

## 5.3 Melt Generation from the Mantle

### 5.3.1 Mantle Composition

Because the mantle cannot be directly sampled, petrologists deduce its composition indirectly. The proxy evidence includes:

*Evidence from mantle-derived melts.* The compositions of partial melts derived from the mantle, particularly mid-ocean ridge basalts and ocean island basalts,

place important constraints on the composition of the mantle.

*The composition of rocks of mantle origin.* Samples of rock that formed in the mantle can be found at Earth's surface and give important indications of the rocks that compose the upper mantle. Mantle rocks occur as xenoliths in basalts or kimberlites, as well as **ophiolites** (discussed further in Chapter 6), which represent pieces of the upper mantle and oceanic crust that have been thrust onto the continents.

*The composition of chondritic meteorites.* Chondritic meteorites have a similar composition to the bulk composition of Earth. The composition of the mantle can be estimated by taking the chondrite composition and subtracting those elements thought to make up Earth's core and crust.

*Geophysical evidence.* The geophysical properties of the mantle, in particular its density and seismic velocity, allow geologists to construct a fairly robust picture of mantle structure and place some constraints on composition.

These various kinds of evidence suggest the mantle has the composition of lherzolite: a peridotite dominated by olivine that contains both orthopyroxene and clinopyroxene (see Figure 1.3). An aluminous mineral is also present: either plagioclase, spinel, or garnet. Depth is the primary control determining which aluminum-bearing mineral is present; plagioclase forms at the shallowest levels, whereas garnet forms at greatest depth.

### 5.3.2 Crust and Mantle Structure

Geophysical evidence indicates the outer 700 kilometers of Earth consist of the following major layers (Figure 5.2):

*Oceanic or continental crust.* Oceanic crust is between three and ten kilometers thick, and continental crust is up to eighty kilometers thick. The base of the crust is defined by the Moho, the seismic discontinuity across which S-wave velocity increases from around 3.5 km/s in the crust to 4.5 km/s in the mantle.

*Lithospheric mantle.* Lithospheric mantle is the upper portion of the mantle that deforms brittlely. It is defined by relatively high S-wave velocities of around 4.5 to 5 km/s. It extends to about 80 kilometers depth beneath the oceans, and to around 200 kilometers beneath continents.

*Asthenosphere.* The asthenosphere extends from the base of the lithosphere to around 660 km depth. It is a relatively weak zone that deforms by creep. S-wave velocities

are lower than in the lithospheric mantle, and may be attenuated in part because of the presence of a partial melt. Together, the asthenosphere and lithospheric mantle compose the **upper mantle**.

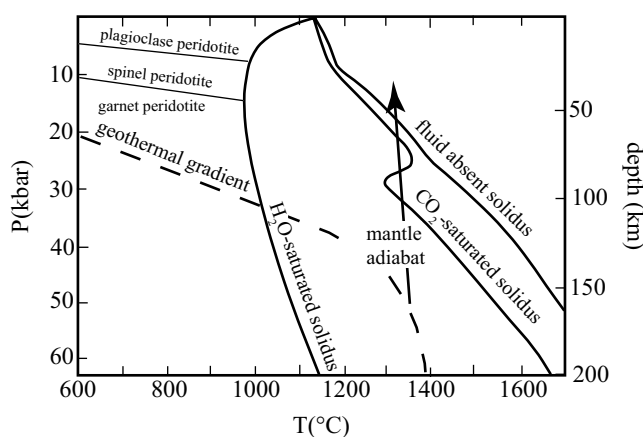
*Mesosphere.* The mesosphere is the part of the mantle below the asthenosphere, extending to the outer core. Its upper boundary is marked by a change in seismic velocity and density thought to correspond to a change in mineral structure in response to increasing pressure. The mesosphere is coincident with the **lower mantle**.

### 5.3.3 Mechanisms for Partial Melting of the Mantle

The normal temperatures encountered at increasing depth in the mantle (the dashed line indicating the geothermal gradient in Figure 5.3) are always below the solidus for fluid-absent lherzolite. Under ordinary circumstances, therefore, the mantle is solid. However, a number of phenomena can generate mantle melting. First, the normal geothermal gradient could be perturbed, so that it is locally hot enough to melt. This may occur beneath ocean islands at “hot spots” such as Hawaii.

Second, the temperature at which melting begins could be lowered by addition of a component to dry lherzolite. The addition of  $\text{CO}_2$  and/or  $\text{H}_2\text{O}$  to peridotite lowers the solidus significantly (Figure 5.3). This means that the addition of  $\text{CO}_2$  and even small amounts of  $\text{H}_2\text{O}$  can lower the solidus enough melt can be produced from lherzolite at the temperatures and pressures thought typical of the normal mantle thermal regime (Figure 5.3). Because subduction carries water-rich fluids along with oceanic crust into the mantle, this process is likely an important mechanism for adding fluids that depress the mantle solidus and trigger partial melting.

A third mechanism that may produce melting is decompression of ascending mantle. Mantle material may ascend either as part of convection cells or as diapirs. The temperature gradient across the center of a convection cell is approximately adiabatic; that is, no heat is transferred in or out of the mass under consideration. For mantle materials, the adiabatic gradient is around  $0.3^\circ\text{C}/\text{km}$ , which means that mantle rising adiabatically does not cool appreciably as it ascends. By comparison, the melting point gradient of anhydrous mantle is much steeper (Figure 5.3). This relationship is shown by the arrow in Figure 5.3, which indicates the P-T path of a rising mantle diapir that originally lay on the mantle geotherm. As



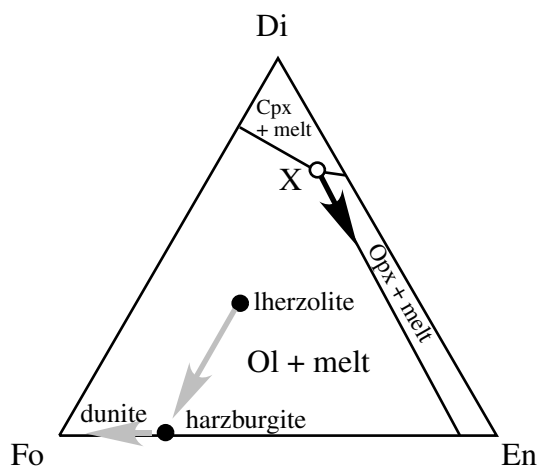
**Figure 5.3** Diagram showing how adiabatic decompression of the mantle (arrow) can lead to melting even if the mantle is dry. Modified from Philpotts and Ague (2009).

the diapir is decompressed adiabatically, it will melt when it crosses the fluid-absent solidus at around fifty kilometers depth, even if the mantle lacked any fluid component such as  $\text{CO}_2$  or  $\text{H}_2\text{O}$ . (If the mantle contains  $\text{CO}_2$  or  $\text{H}_2\text{O}$ , then melting begins at greater depth.) Because mid-ocean ridges are located above up-going limbs of convection cells, decompression melting is particularly important at mid-ocean spreading centers.

### 5.3.4 The Process of Mantle Melting

The composition of partial melts of the mantle and their residual solids can be illustrated in the simplified system Di-Fo-En (Figure 5.4). Lherzolite plots near the center of this ternary system. On heating, the first melt to form is of eutectic composition (point X). As melting proceeds, the residual solids will become progressively more olivine-rich (gray arrow in Figure 5.4). After about 25 percent melting, the diopside (as well as most of the aluminous phase, spinel or garnet) will have been completely incorporated into the melt, and the residue will consist only of olivine and orthopyroxene (i.e., the rock will be a harzburgite). If melting proceeds further, the melt composition will become increasingly enriched in the orthopyroxene component (black arrow in Figure 5.4), while the residue becomes enriched in olivine.

Production of basaltic melts therefore leaves the mantle enriched in olivine. This leads to the common terminology applied to peridotites. A **fertile** lherzolite (a lherzolite from which a basaltic melt can be extracted) contains abundant green (i.e.,  $\text{Al}_2\text{O}_3$ -rich) spinel as well as clinopyroxene that may be rich in minor components such as  $\text{TiO}_2$  and  $\text{Na}_2\text{O}$ . In a **depleted** lherzolite (a lherzolite from which a partial



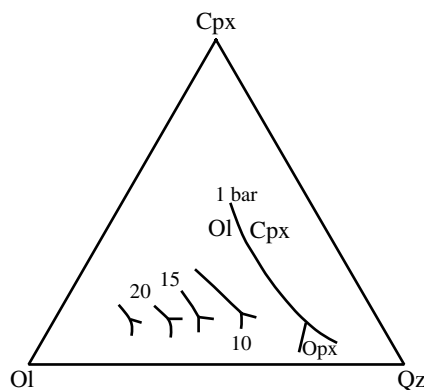
**Figure 5.4** Simplified phase diagram for the system Fo-Di-En at about 20 kbar. Black arrow shows the path followed by the melt during melting of lherzolite; gray arrows show the path followed by the residua.

melt has been extracted), the spinel is Cr rich and less abundant than in fertile lherzolite, and the clinopyroxene has lower  $\text{Na}_2\text{O}$  and  $\text{TiO}_2$  contents. Extremely depleted mantle rocks include harzburgite (a peridotite with little or no clinopyroxene) or dunite, which contains olivine with only minor amounts of orthopyroxene and clinopyroxene.

### 5.3.5 Origin of Tholeiitic versus Alkali Basalts

As noted at the beginning of this chapter, there are two main basalt types, tholeiitic and alkalic. It is natural to ask how melting of the mantle can produce basalts of varying compositions. Possible explanations include:

1. *The alkalic and tholeiitic basalts come from two different sources with different compositions.* As discussed in Chapter 9, mantle-derived, alkaline magmas display a wide range of compositions, from hyper-potassic magmas, such as those of the Roman province of Italy, to highly sodic magmas, such as those from the East African Rift. It is appealing to invoke a heterogeneous mantle to explain this broad compositional spectrum. If the extreme compositional range observed in alkaline rocks does reflect heterogeneous mantle, then the same heterogeneity may also explain lesser compositional differences, such as those that distinguish tholeiites and alkali basalts.
2. *Both alkali basalts and tholeiites come from the same kind of source but they represent melting at different pressures or different degrees of partial melting.* Most



**Figure 5.5** Pseudoternary projection from plagioclase on to the olivine-diopside-quartz plane showing how location of the basalt eutectic changes with increasing pressure. Modified from Elthon (1989).

alkali basalts differ only slightly in composition from tholeiites, so a substantial difference in source composition is not required. Furthermore, in some places, like Hawaii, the basalt types grade from alkalic to tholeiitic during the eruptive history of a volcanic center. To explain this, many petrologists argue that alkali basalts and tholeiitic basalts come from a single mantle source. Evidence supporting this argument derives from the fact that augite in the mantle is the major source of  $\text{Na}_2\text{O}$ ,  $\text{K}_2\text{O}$ ,  $\text{TiO}_2$ , and other incompatible elements enriched in alkali basalts. As noted earlier, augite is the first silicate depleted from melting of lherzolite. If  $\text{Na}_2\text{O}$  is extracted preferentially from this pyroxene during early stages of partial melting of the mantle, then the first melts are alkaline. As melting proceeds, the magmas become progressively more calcic, approaching tholeiite in composition.

Another possible explanation for the origin of the different basalt compositions relates to the pressure of melting. Figure 5.5 shows the effect of pressure on the olivine-orthopyroxene-clinopyroxene-plagioclase eutectic as projected from an aluminous phase (plagioclase, spinel, or garnet). This diagram, which is called a **pseudoternary** diagram, is read in a similar way to Figure 2.15 as long as the projected phase – plagioclase (or spinel or garnet at higher pressure) – is always present. Figure 5.5 shows that increasing pressure moves the eutectic away from silica toward olivine, meaning melts generated at high pressure will likely have less silica than those produced at lower pressure. Thus alkali basalts may be generated from the

same mantle as tholeiites either by lower degrees of partial melting, or at higher pressure, or both.

## 5.4 Environments where Magmas are Generated

Igneous activity observed today is confined to relatively few tectonic environments:

*Constructive plate margins.* These are divergent plate boundaries, such as mid-ocean ridges and back-arc spreading centers, where mantle upwells, decompression melting occurs, and magma is emplaced into the rift. Magmatism in this environment is described in Chapter 6.

*Destructive plate margins.* These are convergent plate boundaries that are either ocean-ocean or ocean-continent collision zones. In these collisions, water subducted

into the mantle via the downgoing, hydrated plate induces melting in the overlying mantle. The resulting magmatism forms ocean islands and continental margin arcs discussed in Chapter 7.

*Oceanic intraplate regions.* These manifest as islands and sea floor plateau that decorate the ocean floor and that were probably caused by hot spot magmatism. Oceanic intraplate magmatism is described in Chapter 6.

*Continental intraplate regions.* Within-plate continental magmatism produces igneous rocks that manifest a substantial range in composition because the magmas form by a number of processes and because the rocks that partially melt are compositionally varied reflecting a range of mantle and continental sources. Continental intraplate volcanism and plutonism are the subjects of Chapters 8 and 9, respectively.

## Summary

- Basalts are classified as alkali basalts, quartz tholeiites, and olivine tholeiites based on their normative mineralogy.
- Alkali basalts evolve toward nepheline saturation and form phonolites and nepheline syenites. Tholeiites evolve toward silica saturation and form trachytes and rhyolites.
- The mantle is composed of an upper lithospheric mantle, which overlies the asthenosphere and mesosphere.
- The temperatures and pressures in the mantle encountered along a typical geothermal gradient are always below the solidus for dry melting of the mantle, so the mantle is normally solid.
- Partial melting of the mantle produces basaltic magmas. Melts are generated by perturbing the normal geothermal gradient to raise the temperature of the mantle, lowering the melting point by adding water or CO<sub>2</sub> or other components, or by bringing the mantle to shallower depths and producing melt by decompression.
- Alkali basalts and tholeiites could come from different mantle sources, or form from the same mantle source by different degrees of melting, or melting at different pressures.
- The tectonic environments that generate the greatest volume of magma are at constructive plate margins, both at mid-ocean ridges and in back-arc spreading centers (Table 5.1). Subduction zones are the second most voluminous sites of magmatism, followed by oceanic intraplate regions where ocean islands and plateau are formed. Lesser volumes of magma form within continental plates, but this tectonic setting creates the most variety of igneous rock compositions.

**Table 5.1** Global Rates (km<sup>3</sup>/yr) of Cenozoic Magmatism

Location	Volcanic rocks	Plutonic rocks
Constructive plate boundaries	3.0	18.0
Destructive plate boundaries	0.4–0.6	2.5–8.0
Continental intraplate regions	0.03–0.1	0.1–1.5
Oceanic intraplate regions	0.3–0.4	1.5–2.0
Global total	3.7–4.1	22.1–29.5

Sources: Crisp (1983) and McBirney (1993)

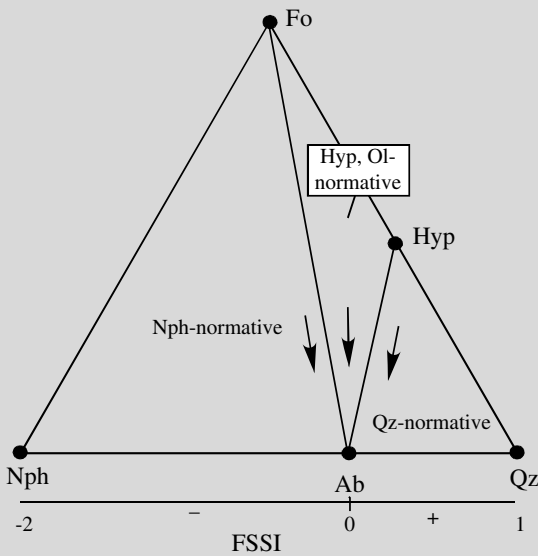
## Questions and Problems

**Problem 5.1.** Describe three mechanisms by which the mantle may partially melt. Give the tectonic environments in which each of these mechanisms may operate.

**Problem 5.2.** What are the mineralogical and chemical differences between alkali and tholeiitic basalt? What rock types represent the extreme differentiates of each?

**Problem 5.3.** What rocks that can be collected at Earth's surface provide the best information about the composition of the mantle? Explain your answer.

**Problem 5.4.** Use the figure below (from Frost and Frost, 2008) to relate the FSSI index (Chapter 4) to the basalt tetrahedron (Figure 5.1). Give the range of FSSI for alkali basalt, olivine tholeiite, and quartz tholeiite.



## Further Reading

McBirney, A. R., 2007, *Igneous petrology*, 3rd ed. Jones and Bartlett, Boston, Chapter 1.

McKenzie, D. and Bickle, M. J., 1988, The volume and composition of melt generated by the extension of the lithosphere. *Journal of Petrology*, 29, 625–79.

Philpotts, A. R. and Ague, J. J., 2009, *Principles of igneous and metamorphic petrology*, 2nd ed., Cambridge University Press, Cambridge, Chapter 23.

Wilson, M., 1989, *Igneous petrogenesis: A global tectonic approach*. Unwin Hyman, London, Chapter 3.