

Chapter 3

Introduction to Silicate Melts and Magmas

3.1 Introduction

Igneous rocks form by crystallization of silicate melts.¹ As melt moves through the mantle and crust, it almost invariably carries crystals. These crystals may be minerals that crystallized from the melt as it cooled or they may be **xenocrysts**, foreign crystals incorporated into the melt from the rocks through which the melts ascended. This mixture of melt and crystals is called a **magma**.

Silicate minerals are highly ordered at the atomic scale, consisting of an oxygen framework in which the cations sit in two major types of sites: tetrahedral sites and octahedral sites. Tetrahedral sites contain Si (and to a lesser extent Al) and, depending on the mineral, they may occur as isolated tetrahedra or as linked tetrahedral (polymers) that form chains, sheets, or frameworks. The octahedral sites lie elsewhere in silicate structure and bond various tetrahedrally bonded polymers together. Cations typically found in octahedral coordination include Fe, Mg, Ca, and Na.

Silicate melts also have these structures at the molecular scale. When a silicate mineral melts, the long-range order disappears, but, unless the melt is taken to temperatures far above the liquidus, the short-range order is maintained. The octahedral and tetrahedral sites remain in the melt, but the polymers formed from the interlocked silica tetrahedra are discontinuous (Figure 3.1). The smaller ions, primarily Si, Al, and P, are called **network-forming ions** because they occupy the tetrahedral sites that help polymerize the melt. The larger ions (Fe, Ca, Mg, Na, etc.) are referred to as **network-modifying ions** because their presence will tend to depolymerize the melt. Because granites are rich in network-forming ions such as SiO_2 and Al_2O_3 , granitic melts consist of many linked tetrahedra and hence have a high degree of polymerization making them viscous. In contrast, rocks rich in network-modifying ions, such as basalt, are not strongly polymerized. Such melts will be more fluid than the granitic melts.

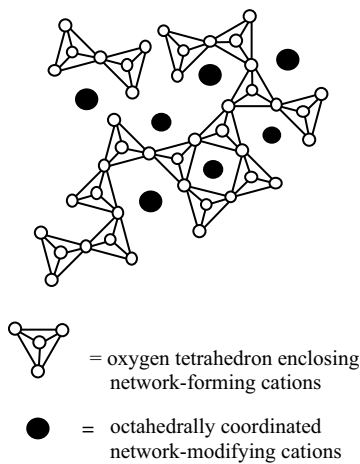


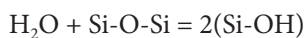
Figure 3.1 Schematic diagram showing the structure of a silicate melt. Small ions (Si and Al) occupy the silica tetrahedra and are called the network-forming ions; larger ions occur in sites outside the tetrahedral polymers and are called network-modifying ions.

3.2 The Role of Volatiles

The presence of volatile components (most commonly H_2O and CO_2) plays an important role in silicate melts. These volatiles not only affect the melting temperature of rock and the viscosity of the melt, they affect the processes that accompany melt ascent and solidification and provide an important mechanism for eruption of lavas.

3.2.1 Role of H_2O

Water plays a critical role in igneous petrology. Not only does water pressure determine the presence or absence of hydrous ferromagnesian silicates, the presence of water has a major effect on the properties of silicate melts. Water reacts with the bridging oxygens to break the silicate network (Figure 3.2) by a reaction such as:



Because of this behavior, H_2O is more soluble in the highly polymerized granitic melts than in the less polymerized basaltic melts. Furthermore, because solution of water into a melt depolymerizes the melt, addition of water to a silicate melt will decrease its viscosity.

The addition of water to a silicate melt also decreases the temperature of crystallization. The effect of water on the melting of albite is shown in Figure 3.3. Albite is a fairly good model for the melting of granite because both

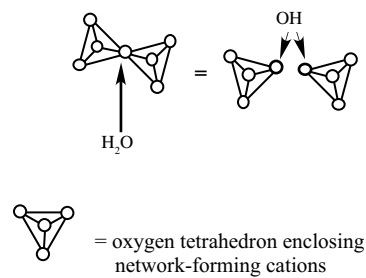


Figure 3.2 Diagram showing how the solution of H_2O into a silicate melt breaks silica polymers, replacing the bridging oxygen with OH.

albite and granite melts are dominated by network-forming elements. In a dry system, granite and albite will melt at very high temperatures (900°C or above). Increasing water pressure will cause melting to occur at progressively lower temperatures. This has an important effect on the behavior of melt in the crust. For example, a dry melt generated deep in the crust, such as at point A in Figure 3.3B, will become superheated as it rises in the crust because decompression leads it away from the dry solidus. Thus, it may be erupted at temperatures as high as 75°C above the solidus. In contrast, although a water-saturated melt (point B in Figure 3.3B) may form at low temperatures, it cannot move to much shallower levels without crystallizing. This means that any granitic melt emplaced into the crust or erupted on the surface must have been undersaturated in water at the depth where it was originally formed (point C, Figure 3.3B).

When a silicate melt crystallizes in the crust, the water originally bound into the melt structure is released as fluid. To understand the importance of this process, one must understand how the volume of water vapor changes as a function of pressure (Figure 3.4). At pressures above about 3,000 bars, the molar volume of H_2O in the melt is nearly the same as that in the fluid. Water exsolved from the melt at these pressures should disperse through the grain boundaries of the country rocks without much dilational effect. Melts emplaced at shallower levels (i.e., at lower pressures) will exsolve water that has a much higher molar volume. For example, melt crystallized at 1,000 bars will exsolve fluid with a molar volume more than three times larger than that of the H_2O dissolved in the melt. When such a melt crystallizes, the water released will hydrofracture the rock and produce a halo of veins around the intrusion. A melt crystallizing at 100 bars will

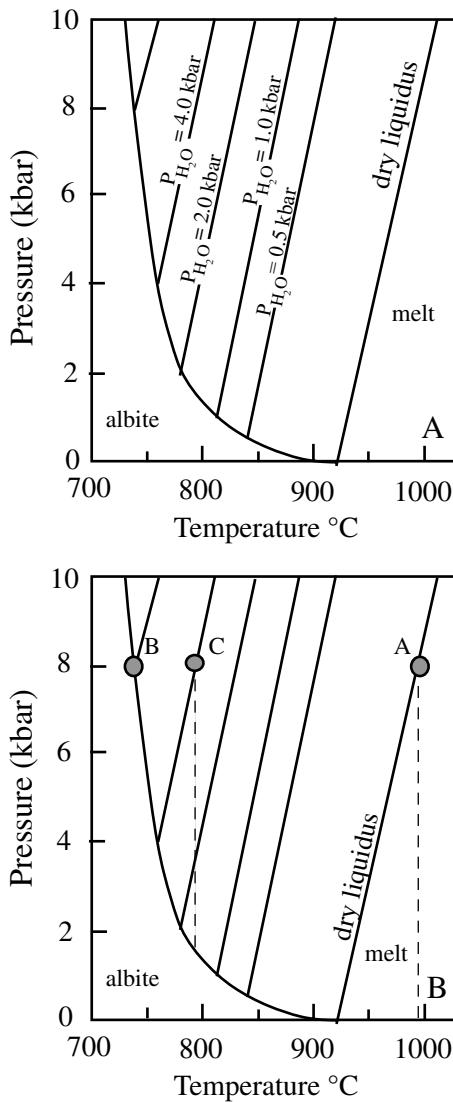


Figure 3.3 (a) Effect of water on the melting temperature of albite. (b) How the phase relations control the melting and movement of melts in the crust. The behavior of melts at points A, B, and C are discussed in the text. After Burnham (1979).

exsolve an aqueous fluid that occupies a volume nearly fifty times that of the H_2O dissolved in the melt. This volume change is so huge that melts crystallizing at shallow levels may exsolve water explosively, producing caldera-type eruptions.

3.2.2 Role of CO_2

Silicate melts contain dissolved water and carbon dioxide, as well as other volatile components such as halogens. Carbon dioxide may be the dominant volatile constituent

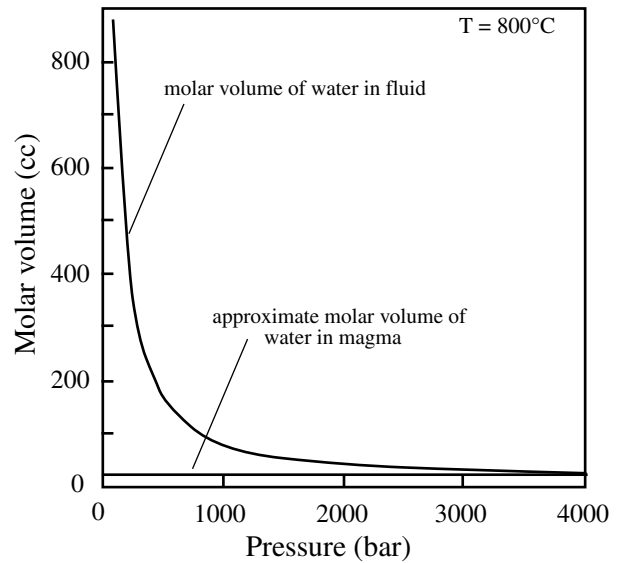


Figure 3.4 Comparison of molar volume of water in melt and in fluid at 800°C . Data from Burnham, Holloway, and Davis (1969).

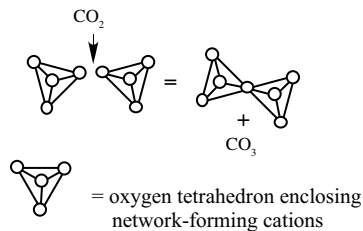
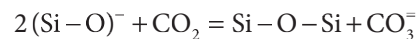


Figure 3.5 Diagram showing how the solution of CO_2 into a silicate melt enhances polymerization.

in melts at depth, whereas water may be incorporated into melts mainly as they pass through the crust. CO_2 -rich fluid inclusions are found in many volcanic and plutonic rocks. Therefore it is important to consider the effect of adding CO_2 on the viscosity of a melt. CO_2 dissolves into the melt by the following reaction:



As shown schematically in Figure 3.5, solution of CO_2 into a melt increases the polymerization of the melt.

Carbon dioxide has a greater solubility in melts with low polymerization, such as basalts, than in ones with few network modifiers, such as albite-quartz melts, because the

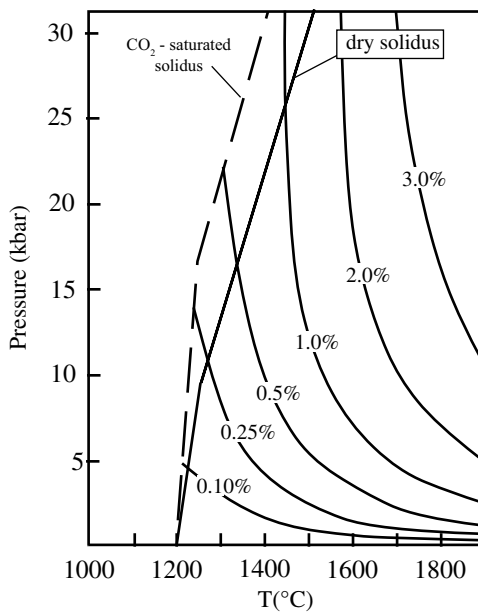


Figure 3.6 Solubility of CO_2 into tholeiitic basalt. After Spera and Bergman (1980).

latter have few nonbridging oxygens with which the CO_2 may bond. Like water, solubility of CO_2 into a melt lowers its melting temperature (Figure 3.6). However, the effect of CO_2 on the melting temperature of a silicate rock is far less extreme than that of water because CO_2 has a much lower solubility. CO_2 is most soluble in basaltic melts at high pressure. This means that as a basaltic melt rises in the crust it will exsolve CO_2 . By the time the melt erupts it will consist of a mixture of silicate melt and CO_2 -rich vapor. This vapor can become one of the major driving forces for eruption of basaltic melts, and indeed the most abundant volatiles emitted during volcanic eruptions are H_2O and CO_2 .

3.3 Physical Properties of Magma

3.3.1 Temperature

Magma temperatures are difficult to measure directly. The temperature of flowing lavas or lava lakes can be measured with a pyrometer or with a thermocouple, and these direct measurements are supplemented with experimental determinations of silicate melting temperatures in the laboratory. The temperature of intrusive magmas is even harder to determine, and for these petrologists rely on experimental results and temperatures calculated from the compositions of coexisting minerals.

Extrusion temperatures for lavas are mostly in the range of 800° to $1,200^\circ\text{C}$. The temperature of basaltic magma is higher than that of rhyolitic magma. Basalts are rarely much hotter than the temperature at which their first minerals crystallize; that is, they are rarely superheated. In contrast, some rhyolites may erupt at temperatures up to 200°C above their liquidus temperature. The temperature of intrusive magmas is probably lower than that of extrusive lavas. Water plays a role in this relationship because the solubility of water in silicate melts increases with increasing pressure. The increased water content of plutonic rocks has the additional effect of depressing liquidus temperatures to the extent that some granitic magmas may have been intruded at temperatures around 700°C .

3.3.2 Heat Capacity and Heat of Fusion

Heat capacity, C_p , is defined as the amount of heat that must be added to raise the temperature of one gram of melt by one degree Celsius. Mathematically this is expressed as:

$$C_p = dH/dT$$

where dH is the amount of heat (enthalpy) that must be added and dT is the change of temperature. Another heat term involved in producing igneous melts is the heat of fusion, dH_f . This variable defines the amount of heat that must be added to a rock already at the melting temperature to produce one gram of melt. One unusual feature of silicate melts is the great difference in these two quantities. C_p is typically 0.3 cal/g deg for most rocks, whereas dH_f is typically $65\text{--}100 \text{ cal/g}$. This means it takes about the same amount of heat to melt a given mass of rock as it takes to raise its temperature by 200° or 300°C (McBirney, 1993). This huge heat of fusion makes ascending magmas an efficient means for moving heat through the crust. Crystallization of magmas in the deep crust releases heat necessary for high-grade metamorphism, whereas at shallow crustal levels, crystallization may provide the heat that drives hydrothermal circulation associated with many ore deposits.

3.3.3 Viscosity

The viscosity of a melt is a measure of its resistance to flow. It is a function of a number of properties, most importantly the composition of the melt, including the types and amounts of dissolved gases such as H_2O and CO_2 , and its temperature. The viscosity of *magmas* is more complicated

than that of silicate *melts* because most magmas contain crystals suspended in the melt. These crystals vary in size and shape, which also affect viscosity. The density of suspended crystals is particularly important: viscosity may be very different when crystals are abundant (greater than >40 percent by volume) than in more dilute suspensions of crystals in melt (Petford, 2009). Although quantification is difficult, viscosity is an important control on fundamental igneous processes including the rate of magma transport in dikes and sills.

Disregarding the complexities introduced by crystal-bearing magmas, it is possible to gain a sense of the variations in viscosities by considering melts of different compositions and temperatures. The compositional control on viscosity of a given melt may be predicted to an extent by determining the ratio of network-forming ions to network-modifying ions. The extent of polymerization is given by the ratio R , which is the total number of oxygen atoms divided by the sum of the network-forming ions in the melt:

$$R = O / (Si + Al + P)$$

Smaller values of R indicate more polymerized melts, which will be more viscous than melts with larger R values, which indicate melts that are less polymerized and therefore more fluid. The range of R extends from around two for pure silica melts to four for a pure forsterite melt.

Viscosity is also temperature dependent: the higher the temperature, the less viscous (more fluid) the melt. Temperature is related to viscosity (η) by the Arrhenius equation:

$$\eta = Ae^{E/RT}$$

where A is a constant, E is the activation energy for viscous flow, R is the gas constant, and T is temperature. Activation energy correlates to R value and is higher for more polymerized melts (Scarfe, 1973). Experimental measurements of the viscosity of melts with different values of R show that rhyolitic melts will have a viscosity nearly 1,000 times higher than that of a tholeiite at the same temperature (Figure 3.7). Figure 3.7 also shows that the viscosity of a melt increases as temperature falls; between 1,200°C and 1,150°C the viscosity of tholeiitic melt nearly doubles.

The differences in viscosity of silicate magmas affect the shapes and processes associated with volcanic rocks. For example, basaltic magmas, which have low R values and

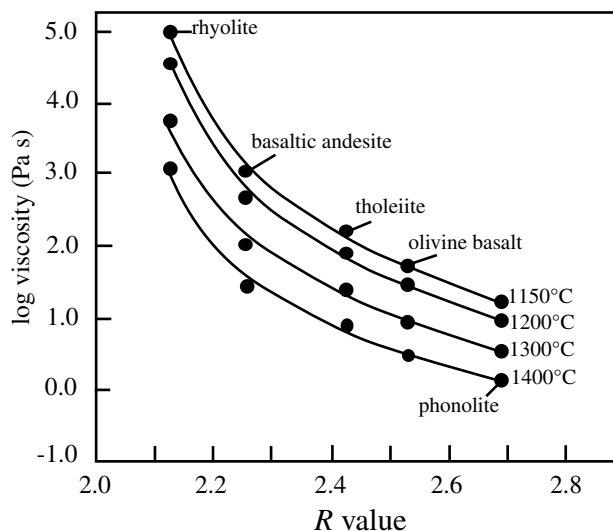


Figure 3.7 Relationship between viscosity, temperature, and composition of melts at 1 bar. After Scarfe (1973).

which are extruded at high temperatures, tend to be fluid (similar to ketchup; see Table 3.1) and may flow for long distances, forming shield volcanoes or flood basalts. In contrast, rhyolite magmas, which have high R values and are erupted at low temperature, tend to have high viscosity, and their behavior is more akin to Silly Putty (Table 3.1). Rhyolitic magmas may form steep-sided domes or, if the magma is too viscous for gas to escape, they may erupt explosively.

3.3.4 Density

Density is an important property that affects the behavior of magmas in various ways: it is one of the factors controlling whether magmas rise through the crust, whether crystals settle out, and whether ions diffuse readily. The density of magmas varies from around 2.2 to 3.1 g/cm³. Density rises with increasing pressure; it falls with increasing temperature. In general, mafic magmas are denser than felsic ones, mainly because mafic magmas are typically richer in heavy oxides such as CaO and FeO, whereas felsic magmas are richer in lighter oxides such as SiO₂, Al₂O₃, and Na₂O.

3.4 The Ascent of Magmas

Most magmas are less dense than the surrounding country rocks, and therefore tend to ascend. The ascent velocity of a body of magma depends on its density and viscosity. At

Table 3.1 Viscosities of Magmas and Common Substances

Material	Viscosity (Pa s)	Wt. % SiO ₂	Temperature (°C)
ASE motor oil	2 × 10 ¹		20
Ketchup	~5 × 10		20
Basalt	10–10 ²	45–52	1,200
Peanut butter	~2.5 × 10 ²		20
Crisco®	2 × 10 ³		20
Andesite	~3.5 × 10 ³	58–62	1,200
Rhyolite	~10 ⁵	73–77	1,200
Rhyolite	~10 ⁸	73–77	800

Source: From Philpotts and Ague, 2009 and sources therein.

depth, magmas may ascend as diapirs, plastically deforming as they push aside the country rocks. The velocity of ascent is approximated by Stokes Law:

$$V = \frac{2g\Delta\rho r^2}{9\eta_w}$$

where V is the velocity, g = the force of gravity, r is diapir radius, $\Delta\rho$ is the density contrast between wallrock and magma, and the viscosity subscript refers to the wallrock (w). From this equation it is clear that velocity is greatest for magma bodies of large size and when magmas ascend through wallrocks with low viscosity.

Magmas may also rise through fractures, forming sheet-like intrusions known as dikes. In this situation, the buoyancy force causing the magma to rise is sufficient to fracture the rock. The fractures may be the site of multiple intrusions of magma, which may be of similar or contrasting composition. A set of subparallel dikes, usually composed of basalt, is referred to as a dike swarm (Map 1.1). Radial dikes may emanate from a central magma body, such as the conduit feeding a volcano (Figure 1.14). Magmas cool as they fill fractures to form dikes because dikes commonly have a very large ratio of surface area to volume. If enough magma travels along a fracture, the wall rock may be thermally eroded, and a more cylindrical conduit may form. Cylindrical geometry is much more favorable for transporting magma with a minimum of heat loss, and it is the common shape of kimberlite diatremes.



Figure 3.8 Stopped block of gabbro that sank through the magma chamber, where it deformed layers of anorthosite accumulating on the magma chamber floor. Laramie anorthosite complex, USA.

At shallow levels, magma ascent may be primarily by **stopping**. In this process, wallrocks are fractured and founder into the magma (Figure 3.8). The magma loses heat both to the wall rock and by assimilation of the stopped blocks. For this reason, it is unlikely that magmas ascend great distances by this process, and ascent by stopping may be limited to the uppermost few kilometers of the magma's rise.

Because not all magma bodies will be of the appropriate temperature, size, density, or viscosity to rise all the way to the surface of the earth, it is worth remembering that those magmas extruded on the surface may not be representative of magmas formed at depth.

3.5 Magmatic Differentiation

Magmatic differentiation refers to all the mechanisms by which a parent magma may give rise to igneous rocks of various compositions. These processes include the crystallization of mineral phases and separation of the residual liquid (fractional crystallization), separation of two melts (liquid immiscibility), mixing of separate magmas, and assimilation of country rock either as a solid or as a partial melt into the parent magma. It is also possible to produce a diversity of rock types by partially melting solid rock and removing the melt, leaving behind a refractory residue. Because magmas originate by processes of melting, these mechanisms will be examined first.

3.5.1 Partial Melting

Partial melting, also referred to as *partial fusion* or *anatexis*, is the process by which melt is produced in a proportion less than the whole. Partial melting can be considered by examining the two end member models introduced in the discussion of phase diagrams in Chapter 2:

Equilibrium melting. The partial melt that forms continually reacts and equilibrates with the remaining solid until the moment the melt is removed. Up to the point of segregation, the bulk composition of the system remains constant.

Fractional melting. The partial melt is removed in infinitely small increments so that it cannot interact with the residual solid. The bulk composition of the solid continually changes because melt, which is of a different composition than the initial solid, is lost from the system.

Which of these end member processes most closely approximates partial melting in nature will depend on the ability of the melt to separate from the residual crystals. This in turn depends on many factors that control the microscopic geometry of the partial melt within the solid rock. Models of melt production and segregation in the mantle suggest melts can escape from the solid matrix after relatively small degrees of melting. Assuming that compaction drives expulsion of partial melt, MacKenzie (1984) suggested melt will be expelled once the melt represents 3 percent or more of the total volume. MacKenzie and O'Nions (1991) present

evidence based on the inversion of rare earth element concentrations that melt segregation in the mantle can occur when melt fractions are less than 1 percent. More recently, Rabinowicz and Toplis (2009) considered the combined effect of shear segregation related to the ductile flow of the mantle and compaction, and determined that, depending on the viscosity of the solid mantle, basaltic melts will be expelled at melt fractions between 3.5 and 10 percent.

3.5.2 Crystallization Processes

Crystallization of solid phases from a melt also presents opportunities to form rocks of a different composition than the original melt. Like partial melting processes, crystallization can be considered by examining the two end member models discussed in Chapter 2:

Equilibrium crystallization. Crystals remain in contact with the residual liquid after they form and continually react and equilibrate with the liquid. In this case, the bulk composition of the final solids are the same as the original melt composition, and no magmatic differentiation takes place.

Fractional crystallization. Crystals are removed from the residual liquid as soon as they are formed, either by gravitational settling or floating. In this process, the bulk composition of the remaining liquid changes as crystals form and are removed.

It is possible to segregate liquid from crystals by mechanisms other than gravitational separation. The liquid remaining before crystallization may be complete and can be squeezed out in a process called **filter pressing**. A buoyant liquid in a mush of loosely packed crystals may migrate to a zone of lower pressure, just as water is driven out of a pile of accumulating and compacting sediments. Another mechanism, **flow segregation**, may separate crystals from the remaining melt during flow through a dike or along the walls of a pluton. Nearest the contact with country rock, the velocity gradient is steepest and a zone of maximum shear is present. This shearing results in a force that tends to drive crystals out of the zone of maximum shear and toward the interior of the flow. Crystals are thus found in the areas of lowest velocity gradient; in a dike they may be concentrated in the center, and the margins of the dike are much finer grained.

3.5.3 Liquid-Liquid Fractionation

Magmas may become compositionally zoned by two different processes: boundary layer fractionation and immiscibility. **Boundary layer fractionation** results when the magma along the margins of the chamber acquires a density different from the rest of the magma, either by absorption of water from the country rocks or by partial melting or crystallization. This liquid then rises and collects under the roof of the magma chamber, or less commonly may sink and collect at the bottom. This process was important in the formation of the Bishop Tuff, an extensive, compositionally stratified deposit distributed over much of the western United States. The 0.77-million-year-old tuff is postulated to have been erupted from a compositionally zoned magma chamber beneath what is now Long Valley caldera in southern California (Hildreth, 1979). Because of the high viscosity of silicic magmas, crystal settling is unlikely to be an effective process of magmatic differentiation. Instead, Hildreth proposed that absorption of water from the wall rocks produced a water-enriched boundary layer that was less dense than the magma in the interior of the chamber. This magma rose to the top of the chamber, producing a compositionally stratified magma chamber. Another boundary layer process that may produce stratification results from cooling of magma along the walls and roof. Crystallization of the cooler magma along the margin of the chamber may leave a less dense interstitial liquid that rises along the walls toward the top, where it forms a stratified cap to the magma chamber (Sawka, Chappell, and Kistler, 1990).

Certain melt compositions may separate into two or more **immiscible** (i.e., unmixable) liquids. Immiscibility is thought most common in mafic melts, which may separate into a sulfide liquid and a mafic silicate liquid. Immiscibility is also likely to occur in alkaline melts rich in CO₂ and that may separate into a high-alkali silicate liquid and a carbonate-rich liquid. Iron-rich basaltic melts may separate into a felsic silica-rich liquid and a mafic iron-rich liquid. If the two liquids formed have very different densities, then they may separate very effectively, as oil does from water. If the magma is very viscous and crystal-rich, the two liquids may not separate as well, and small droplets be evident in the interstices between early crystallizing minerals. Evidence of this process is preserved as solid inclusions of (Na, K) Cl

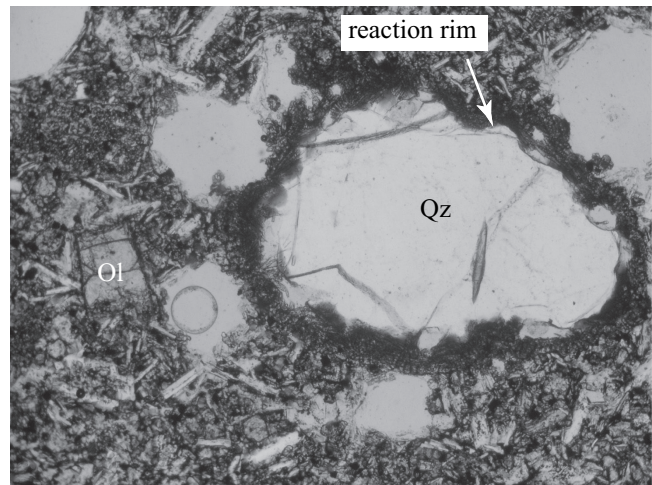
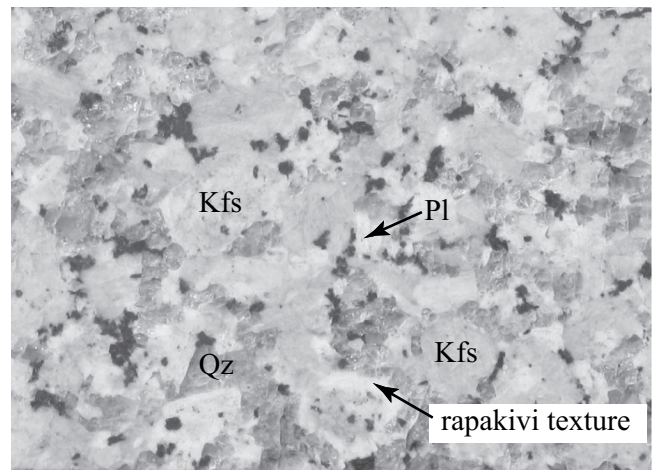


Figure 3.9 (A) Photo of alkali feldspar mantled with plagioclase (reverse zoning), a texture known as rapakivi texture. Mantled feldspars indicate a change in crystallization conditions. Ksp = alkali-feldspar, Pl = plagioclase, Qz = quartz. Sherman batholith, Wyoming, USA.. (B) Resorbed crystal of quartz (Qz) in an olivine-bearing (Ol) groundmass. The anhedral crystal shape and reaction rim around the quartz suggest that this crystal was entrained into a magma in which quartz was not stable.

in K-feldspar in the monzosyenites of the Laramie anorthosite complex, Wyoming, USA, which suggests these rocks formed from a magma that contained immiscible droplets of a chloride-rich melt (Frost and Touret, 1989).

3.5.4 Assimilation

Magmas may also change composition by assimilating material from their country rocks. *Assimilation*,

sometimes referred to as *contamination*, may occur in two ways:

Bulk assimilation occurs when blocks of wall rock are stoped into the magma and completely melt.

Assimilation of partial melts occurs when the wall rocks are heated to their solidus and begin to melt. It is important to remember that the composition of the first melts are usually different than the bulk composition of the rock, and hence the compositional effect on the magma is not identical to that from bulk assimilation.

Both assimilation of bulk rock and partial melt require thermal energy for melting. The melt necessarily becomes cooler, so much so that it may begin to crystallize. Crystallization releases heat of fusion, a positive feedback that helps assimilation proceed. Therefore assimilation is usually accompanied by crystallization (assimilation + fractional crystallization, often abbreviated as AFC).

Various textural features may evidence assimilation. Xenocrysts, crystals inherited from the country rock that would not normally be expected to crystallize from the melt, may be present. **Reversely zoned crystals**, such as plagioclase with calcic rather than sodic rims, suggest the melt has changed its composition and that the composition of the phases crystallizing from the melt has changed in a manner not normally associated with differentiation by simple fractional crystallization (Figure 3.9A). **Resorbed crystals**, those that show textural evidence of remelting may (but do not always) result from changes in melt composition, which may cause a previously crystallizing phase to be out of equilibrium with the contaminated melt (Figure 3.9B).

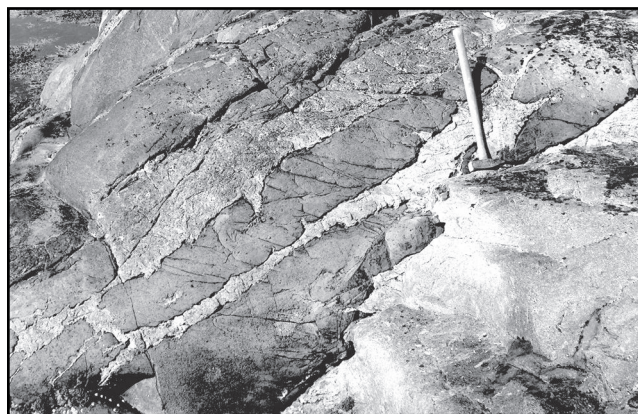


Figure 3.10 The Hortavær complex was constructed by multiple injections of diorite (dark lenses) intruded into a syenitic magma chamber (light-colored rock), north central Norway.

3.5.5 Magma Mixing

If two magmas are introduced into the same magma chamber, they may mix and form a magma of intermediate composition. The magmas may be independently derived, or may have formed by boundary layer fractionation and consequently rehomogenize. If the two magmas do not completely homogenize, then evidence of commingled melts may remain. Intrusion of denser mafic magma into a felsic magma chamber sometimes forms a sheet of mafic magma near the floor of the magma body. Thermal contrasts between the magmas cause the mafic magma to chill, forming lobate margins and other structures recording the presence of both mafic and felsic magmas within a chamber (Figure 3.10).

Summary

- Silicate melts are composed of network-forming ions, such as Si, Al, and P, that occupy tetrahedral sites. These form linked tetrahedra that polymerize the melt. Larger ions including Fe, Ca, Mg, and Na, form network-modifying ions that tend to depolymerize the melt. Granitic melts are richer in network-forming ions and are more strongly polymerized (and more viscous) than basaltic melts.
- The presence of H₂O in silicate melts tends to break the silicate networks and depolymerize the melt. Carbon dioxide has the opposite effect, and increases the polymerization of the melt.
- In general, mafic magmas are denser than felsic ones.
- Magma viscosity is a function of composition, temperature, and the kinds, quantities, and geometries of crystals present.

- Magmas are typically less dense than the surrounding rock, and hence have a tendency to rise. Magmatic ascent is an important mechanism to transport deep heat to the shallow crust.
- A diversity of igneous rocks can be produced by partial melting solid rock, extracting the melt, and leaving behind a more refractory residue. Partial melting may be fractional, in which infinitely small increments are removed from the remaining solid. Equilibrium melting describes the process by which partial melt continually reacts and equilibrates with the solid until melting is complete.
- A suite of igneous rocks with a variety of compositions may also form by equilibrium or fractional crystallization.
- Other processes that differentiate magmas include immiscible liquid-liquid fractionation, assimilation, and magma mixing.

Questions and Problems

Problem 3.1. Referring to the albite-H₂O system as an analog for granite (Figure 3.3), answer the following:

- How does increasing pressure affect the amount of H₂O that can be dissolved in a granitic melt?
- How does addition of H₂O affect the melting point of the granitic rock?
- What is the effect of dissolution of H₂O on the viscosity of the granitic melt?
- How does the ascent of a rising H₂O-saturated magma compare to that of a rising H₂O-undersaturated magma?

Problem 3.2. Aqueous fluids can be released during the late stages of crystallization of a magma in a magma chamber.

- Refer to Figure 3.3 to explain how crystallization at a constant pressure can lead a magma to become H₂O saturated.
- At what pressures might the aqueous fluid released hydrofracture the rock? Discuss your answer in terms of the molar volume of H₂O.
- Why might porphyritic texture result (like that shown in Figure 1.7D) when H₂O is lost from a magma body?

Problem 3.3. Use Stoke's Law to estimate the ascent velocity (in cm/s) of a magma diapir:

$$V = \frac{2g\Delta\rho r^2}{9\eta_w}$$

- Calculate ascent velocity for a 10 km diameter diapir with a density of 2.6 g/cm³ through mid-crustal wall rocks with a density of 3.0 g/cm³ and a viscosity of 10²¹ Pa s. (Recall that $g = 980 \text{ cm/sec}^2$ for Earth and $1 \text{ Pa s} = 10 \text{ g/cm s}$.)
- What will affect the ascent rate more: decreasing the diapir diameter to 1 km, or increasing viscosity of the wall rocks to 10²⁴ Pa s, a value representative of upper crustal rocks? Present calculations to support your answer. On the basis of these simple calculations, does diapiric ascent appear to be an effective mechanism for moving magmas through the upper crust?