Abstract

Igneous rocks comprise the bulk of the Earth, Moon, and other terrestrial planets. They form by the solidification of magma or "molten rock." This chapter introduces the reader to magma and igneous rocks. It deals with the nature of magma and some of the fundamental descriptive aspects of igneous rocks, primarily their mineralogy and textures. And finally, the reader is introduced to the concepts of igneous rock classification.

Magma: Fundamentals

Magma may be thought of as a molten rock. Magma is not often a simple liquid but rather a complex mixture of liquid, solid materials (crystals and rocks), and dissolved vapor. Magma's chemical composition and physical properties, such as viscosity and density, are all important factors that control magma's overall behavior—how fast or slow it will move, how explosive it will be, what minerals will form in it, and their grain sizes and physical interrelationships (i.e., *texture*). Below we discuss some of these aspects of magma.

Chemical Composition of Magma

In terms of chemical composition, all magmas (except rare *carbonatites* that are rich in a carbonate component) are silicate magmas in which the dominant component is silica (SiO_2), which generally comprises 45 % or more by weight. Alumina (Al_2O_3), with its abundance in common igneous rocks somewhere between 13 and 18 %, is a distant second (Table 2.1). SiO_2, TiO_2, Al_2O_3, Fe_2O_3, CaO, MgO, MnO, FeO, Fe_2O_3, Na_2O, and K_2O are often referred to as *major* oxides because together they comprise 99 % of any igneous rock. (Actually, the term "major element" is generally used, which is a bad practice that has been perpetuated through decades of geochemical practice.

This is rooted in the fact that these are analyzed as elements but expressed as oxides because the most abundant element, oxygen, is generally not analyzed.)

There are many other elements (e.g., V, Sr, Ni, La, Ce, Nd, Sm, Eu, Yb, Hf, etc.) whose abundances are generally much lower (parts per thousand or parts per million levels) and are hence referred to as *trace elements*. These elements are nonetheless important because they provide significant clues to the origin of igneous rocks.

Common Magma Types

As a starting point, it is useful to classify magmas into four broad groups, namely, *ultramafic*, *mafic*, *intermediate*, and *felsic* types (Table 2.2), based on relative abundances of SiO_2, $MgO + FeO + Fe_2O_3$ (called the *mafic* component), and total alkalis ($Na_2O + K_2O$). Equivalent terms for the above types are as follows: ultramafic magma = *komatiite magma*; mafic magma = *basaltic magma*; intermediate magma = *andesitic magma*; and felsic magma = *rhyolitic magma*.

It is apparent from Table 2.2 that the mafic component is minimal in felsic magmas and it increases toward ultramafic magmas. Alkalis and silica exhibit the opposite behavior. What minerals may form from a magma depend on the magma's chemical composition: for example, an ultramafic magma with its low SiO_2

Table 2.1 Major element composition of some igneous rocks

	Komatiite	Picrite	Basalt	Andesite	Rhyolite
SiO_2	45.26	43.68	49.20	57.94	72.82
TiO_2	0.33	2.19	1.84	0.87	0.28
Al_2O_3	6.74	9.83	15.74	17.02	13.27
Fe_2O_3	2.13	1.90	3.79	3.27	1.48
FeO	8.66	8.44	7.13	4.04	1.11
MnO	0.17	0.01	0.20	0.14	0.06
MgO	22.98	15.24	6.73	3.33	0.39
CaO	6.94	15.82	9.47	6.79	1.14
Na_2O	0.88	0.65	2.91	3.48	3.35
K_2O	0.05	0.38	1.10	1.62	4.30
P_2O_5	0.02		0.35	0.21	0.05
H_2O^+	3.41		0.95	0.83	1.10
H_2O^-	0.57	1.00	0.43	0.34	0.31
CO_2	0.84		0.11	0.05	0.08
Total	99.00	99.14	99.95	99.93	99.96

and very high MgO + FeO should be expected to crystallize minerals like olivine and pyroxenes, which have high MgO + FeO. Quartz (SiO_2) would *not* be expected to form in such a magma because formation of olivine and pyroxene crystals would use up all the available silica so that there will not remain any excess silica needed to form quartz. On the other hand, in a felsic magma, which is very high in silica and alkalis and poor in MgO + FeO, one should expect crystals of alkali feldspars [$(K, Na)AlSi_3O_8$] and quartz to form. Table 2.2 shows this direct correlation between magma's chemical composition and minerals that crystallize from it. This is the basis for *norm calculation*, which is further discussed in a later section.

Rock names are not only based on chemical composition but also on its grain size, which is determined by whether rock formed by quenching of lava (i.e., volcanic), by extremely slow cooling and solidification in a deep crustal pluton, or by relatively rapid crystallization in a shallow (hypabyssal) crustal intrusion, such as a dike or sill. The bottom part of Table 2.2 shows that

different names are given to rocks that have identical chemical composition but very different mode of occurrence or texture. For example, basalt and gabbro have the same chemical composition, but basalt is fine grained relative to gabbro, which is a coarse-grained plutonic rock.

Solid Component in Magma

Magma rises from its place of origin primarily by buoyancy due to its lower density (although other factors, such as volatile pressure, also contribute to magma's ascent) than that of the surrounding wall rocks. Magma does not usually rise directly through a single conduit from its source region to eruption; instead it may stall in some *magma chambers* that may be located at various levels in the crust or upper mantle (Fig. 2.1). Magma generally cools in these magma chambers such that crystals form and separate from it. In erupted lavas, such crystals, often with well-developed faces (they are called *euhedral* crystals), may be recognized as *phenocrysts* by virtue of their larger grain size relative to the surrounding glass or fine crystals that form due to rapid quenching of the lava.

Magma may also carry rock and mineral fragments broken off the conduit wall as it rises violently from the deep. Such wall rock and mineral inclusions are called *xenoliths* and *xenocrysts*, respectively (Fig. 2.2). The study of xenoliths and xenocrysts has proved to be particularly valuable because many of them come from depths in the Earth's upper mantle that cannot otherwise be sampled by drilling. Xenoliths and xenocrysts are often identified on the basis of their chemical composition, mineral assemblages, and disequilibrium reaction textures (such as rounded or *resorbed boundaries*), which develop due to their being out of chemical equilibrium with the host magma.

Table 2.2 First-order classification of magma and igneous rocks

| Magma type	Ultramafic	Mafic	Intermediate	Felsic
SiO_2	42–48	46–54	60–65	>70
MgO + FeO + MnO + Fe_2O_3	35–46	15–28	10–21	<3
$Na_2O + K_2O$	<1	2–3.5	3–6	5–10
Majo minerals	Olivine (generally dominant) + pyroxenes	Plagioclase + pyroxene	Pyroxene + plagioclase + amphibole	Alkali feldspar + quartz
Volcanic	Komatiite	Basalt	Andesite	Rhyolite
Hypabyssal	Komatiite	Diabase		
Plutonic	Peridotite	Gabbro	Diorite	Granite
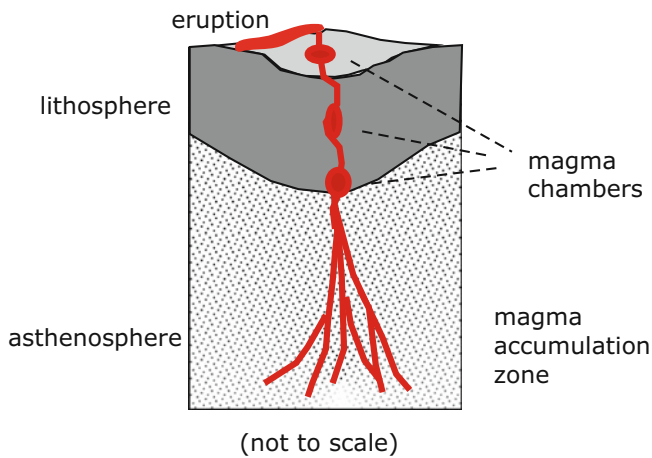
Fig. 2.1 Schematic diagram showing how magma collects to form larger veins and intrusions. On the way up the magma may reside in magma chambers at various depths, cool and crystallize, and then eventually erupt through a volcano

Structure of Silicate Magma

Magma's viscosity and density are important factors that determine how fast it will rise. These physical properties of magma are fundamentally related to its internal arrangement of atoms and molecules (Fig. 2.3). Magma's atomic structure may be understood by referring to the atomic structures of silicate minerals. The fundamental structural unit of all silicate minerals
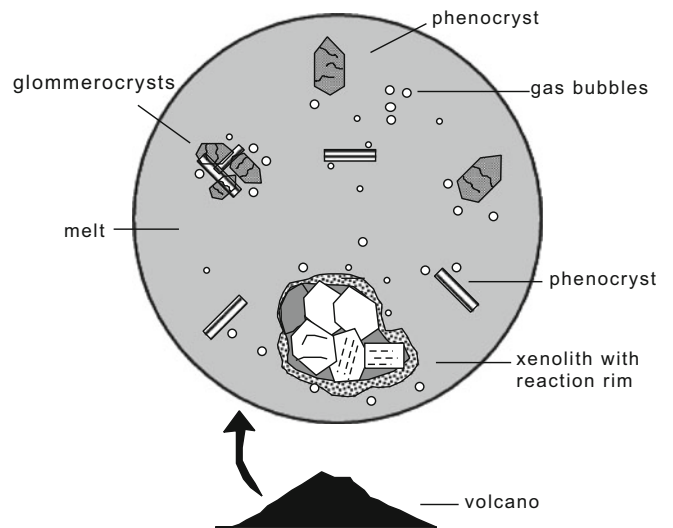

Fig. 2.2 Schematic drawing of a microscopic view (crossed polars) of a volcanic rock showing the presence of glass (isotropic), phenocrysts of two different minerals (olivine and plagioclase), cluster of phenocrysts (glomerocrysts), exsolved gas bubbles (preserved as vesicles), and a xenolith. Because the xenolith was not in equilibrium with the host lava, a reaction rim composed of different minerals developed between the host lava and the xenolith

is the $[SiO_4]^{4-}$ tetrahedron in which a small Si^{4+} cation at the tetrahedron's center is covalently bonded to four O^{2-} anions occupying its apices. In most silicate groups, for example, the chain silicates,
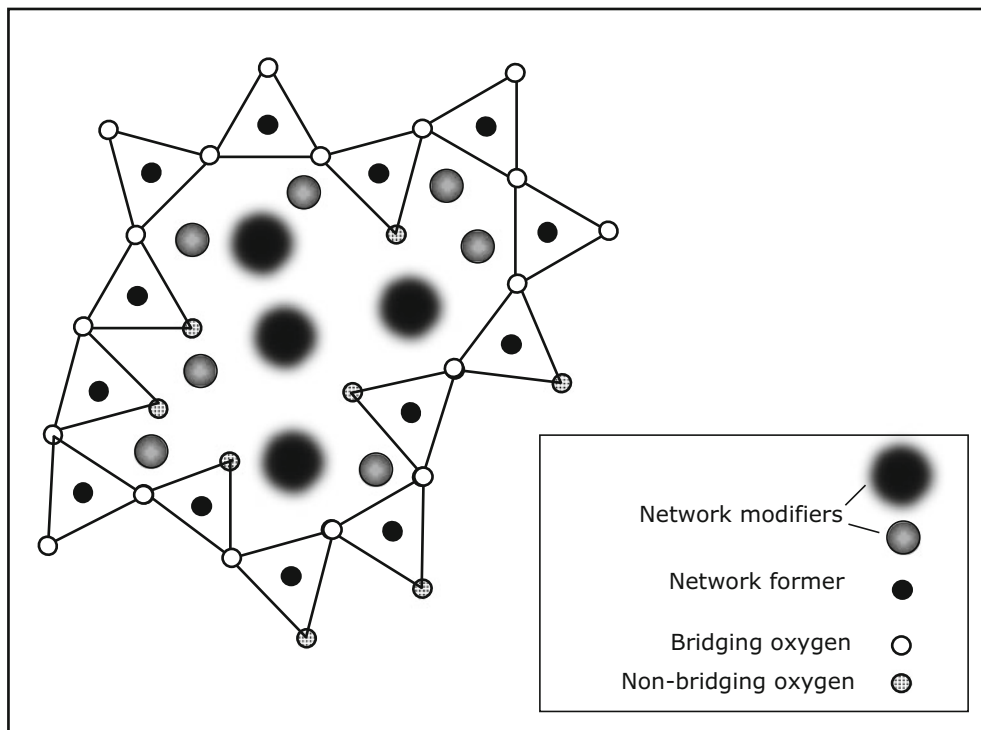

Fig. 2.3 Atomic structure of a silicate melt showing polymerized chains of silicate tetrahedra, network formers, and network modifiers

[SiO$_4$]-tetrahedral units share apical oxygen ions whereas the non-shared oxygen ions are bonded to other cations (Mg, Fe, Na, Ca, etc.) present in the mineral's structure. The shared versus non-shared oxygen ions are called *bridging* and *non-bridging* oxygen ions, respectively. Si and other fourfold coordinated cations (principally Al) that form the "backbone" of a tetrahedral network are called *network formers*. The cations (e.g., Mg, Ca, etc.) that usually occur in higher coordination and link the networks are called *network modifiers*. Silicate magmas, like silicate minerals, are composed of extensive networks of [SiO$_4$]-tetrahedral units with neighboring tetrahedra containing bridging and non-bridging oxygen ions. During melting of a silicate mineral or a rock, the covalent bonds between Si^{4+} and oxygen ions of each tetrahedron are generally too strong to break. On the other hand, the bonds between non-bridging oxygen ions and network modifiers do break, enabling these other cations to move around within the framework of polymerized [SiO$_4$] chains in a molten silicate (Fig. 2.3). As one would expect, SiO$_2$ content of a magma exerts strong control over the extent of polymerization of a melt—the greater the SiO$_2$, the more polymerized the melt.

Viscosity

Viscosity (η) of a magma is simply defined as its internal resistance to flow and is given as:

$$\eta = \sigma/\varepsilon$$

where σ and ε are applied *shear stress* and *rate of shear strain*, respectively. A convenient way to understand what viscosity means is to compare how syrup and water flows: syrup is said to be more viscous than water because it flows more slowly than the latter. Magmas or fluids in general can show a range of viscosity characteristics: *Newtonian* fluids exhibit a linear relationship between σ and ε and pass through the origin in a stress versus strain rate plot (Fig. 2.4). Crystal-free basalt magmas show Newtonian behavior. On the other hand a basalt or andesitic magma containing abundant crystals may behave like a *Bingham plastic*, that is, they may possess some finite yield strength and thus flow only when a certain threshold value of stress has been exceeded. Rhyolitic magmas exhibit *pseudoplastic* behavior in that it shows a nonlinear relationship between stress and strain rate. As should be expected, the extent of polymerization (i.e., how many silicate chains occur in the
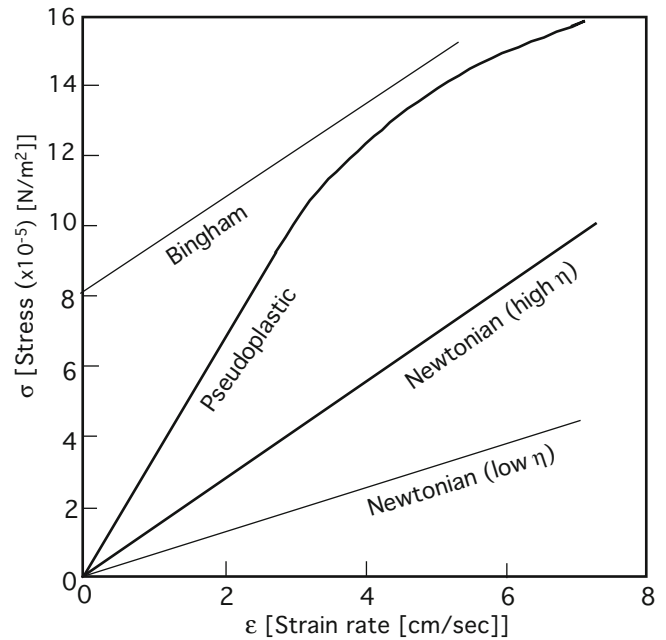

Fig. 2.4 Stress–strain relationships and types of magma viscosities

magma, which is directly dependent on the abundance of SiO$_2$) exerts a strong control over viscosity: thus, rhyolite magma with its highest SiO$_2$ content is more viscous than andesite and basalt magma (Fig. 2.5).
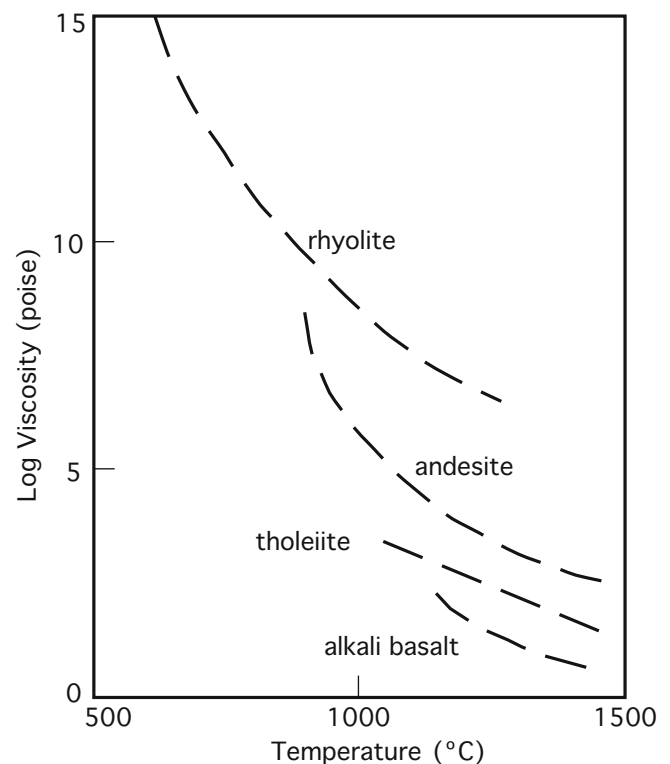

Fig. 2.5 Viscosities of four principal magma types at 1 atm pressure and as a function of temperature (McBirney 2007)
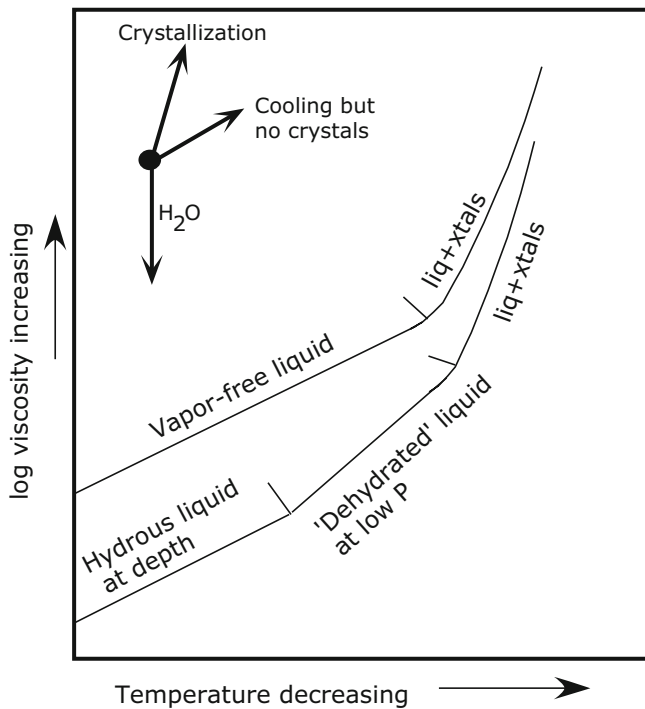
Fig. 2.6 Viscosities of hydrous versus anhydrous magmas as a function of devolatilization and partial crystallization (modified from Hess 1989). Hydrous magma has lower viscosity than anhydrous magma of broadly similar chemical composition. As a hydrous magma rises, it may expel ("exsolve") its dissolved volatiles (such as H_2O) due to lowering of pressure or temperature (discussed in a later chapter), and its viscosity may increase as a result. Appearance of crystals in great abundance may also sharply increase the viscosity of a magma

Temperature increase breaks down more silicate networks in the magma and thus lowers its viscosity. On the other hand, increase in crystal content in a cooling magma makes it progressively more viscous. Increase of dissolved H_2O, a network modifier, greatly lowers the viscosity of magma (Fig. 2.6).

Box 2.1: Viscosity, Density, and Velocity of Xenolith-Bearing Magma from Hawaii

The island of Oahu is composed of two large shield volcanoes and numerous small cinder cones. Lavas that erupted some 1–2 million years ago from the small cones contain mantle xenoliths. Using some simple viscosity-density values of the xenoliths and magma, it is possible to calculate a *minimum ascent velocity* of the magma. Note that these xenoliths are denser (density = 3.4 g/cm^3) than the magma (density = 2.8 g/cm^3) and therefore should have precipitated (or settled) out of the magma, but they were brought up because the magma was rising faster than the

Box 2.1 (continued)
xenoliths could settle. Using a simple law, called Stoke's law (see below), and assuming that the magma behaved as a Newtonian fluid, one can calculate the minimum ascent velocity of the magma that prevented the xenoliths from settling.

Stoke's law is given as

$$V = \frac{2gr^2 \left(\rho_{\text{cryst}} - \rho_{\text{liq}} \right)}{9\eta}$$

where
V = settling velocity of the xenolith in magma
g = acceleration due to gravity (assumed to be constant; 980 cm/s^2)
r = radius of the xenolith (2.5–10 cm)
ρ_{cryst} = density of xenolith (assumed to be 3.4 g/cm^3)
ρ_{liq} = density of liquid (assumed to be 2.8 g/cm^3)
η = viscosity (varies between 20 and 50 Pa.s (1 Pa.s = 10 Poise) for basalt magmas ± crystals)

Plugging these values into the equation, minimum ascent velocities of the host magma were calculated and plotted in Fig. 2.7 for two different sizes of xenoliths and two different magma viscosities. It is clear that viscosity has no effect on ascent velocity for small xenoliths; however, for larger sizes the effect is quite strong. Note that these estimates will increase by twice as much if the magma behaved like a Bingham fluid. However, field observations (such as a thin glassy coat of the host lava around the xenolith) on some of these xenolith-bearing magmas of Hawaii suggest that they were extremely fluid and may have behaved like a Newtonian fluid. The interested reader is referred to Spera (1980).
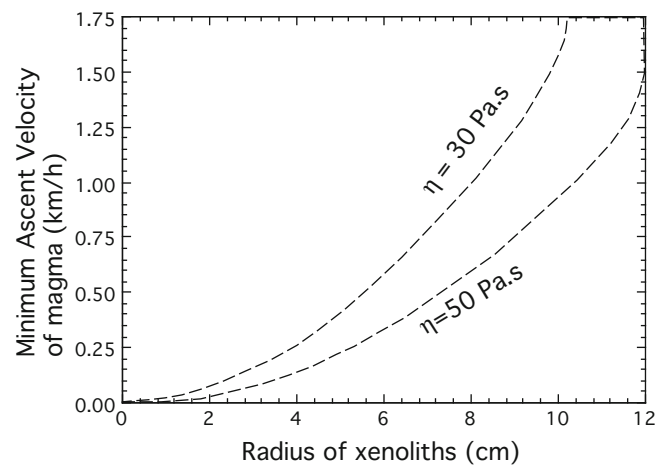

Fig. 2.7 Xenoliths and magma ascent rates

Density

Densities of different magmas have been measured in the laboratory, mostly at atmospheric pressure, and they vary between 2.2 and 3.1 g/cm^3 (Fig. 2.8). Density of magma is directly related to the abundance of the mafic (i.e., Mg + Fe) component in it. Thus, rhyolite magmas are less dense than andesites, and andesites are less dense than basalts. Magma rises from its point of origin by buoyancy, much like a helium-filled balloon rises through the atmosphere. This happens because the magma is lighter (i.e., less dense) than the surrounding wall rock.

Consider a simple example of buoyant rise of a basaltic magma from a depth of 60 km to the surface. We assume that the wall rock at ~60–40 km is peridotite with a density of 3.3 g/cm^3. The magma's density is assumed to be constant at 2.9 g/cm^3. The pressure on the magma and wall rock at any depth can be calculated from

$$P = \rho g h$$

where P is pressure (in GPa), ρ is density, and g is acceleration due to gravity (assumed to be constant with a value of 980 cm/s^2).

$$P_{rock} \text{ at 60 km} = (6,000,000 \times 3.3 \times 980)/10^{10}$$
$$= 1.94 \text{ GPa}$$

$$P_{magma} \text{ at 60 km} = (6,000,000 \times 2.9 \times 980)/10^{10}$$
$$= 1.70 \text{ GPa}$$

Therefore, the pressure difference (ΔP) of 0.24 GPa between the magma and wall rock makes the magma sufficiently buoyant to rise to the surface. In reality, however, this pressure difference may disappear when the magma reaches Moho, which is the seismic discontinuity that separates crust from the mantle. This is because crustal rocks, especially in continents, often have a lower density (~2.7 g/cm^3) than that of the magma; and therefore, the magma will not be buoyant when it reaches the Moho. Instead, it may be trapped at the Moho, where it would start cooling and crystallizing. After some amount of crystallization, the remaining differentiated magma may rise again when its density becomes less than that of its surrounding crust. Stolper and Walker (1980) showed that the vast majority of mid-oceanic ridge basalts (MORBs) undergo such density filtration and differentiation in magma chambers. Thus, they
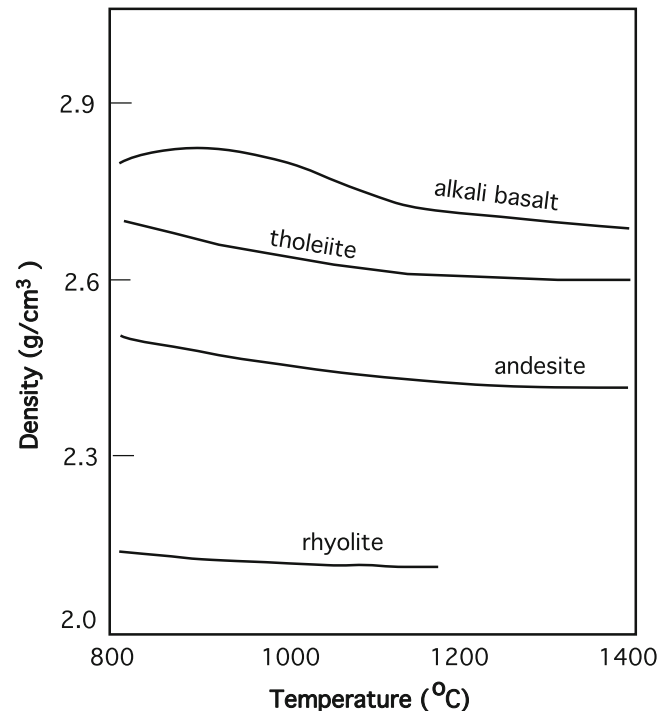

Fig. 2.8 Densities of magmas as a function of temperature and at atmospheric pressure

argued that erupted MORBs are not the *primary* magmas formed directly by partial melting of the mantle but are derived from them by crystallization differentiation. (The concept of primary magma will be discussed in a later chapter.)

Density of magma at atmospheric pressure can be calculated from its chemical composition and molar volume as follows:

$$\rho = \frac{\Sigma X_i M_i}{\Sigma X_i V_i}$$

where X_i, M_i, and V_i are mole fraction, gram molecular weight, and molar volume of an oxide. In a magma composed of many different oxide components SiO_2, Al_2O_3, TiO_2, etc., the mole fraction of any particular oxide, say, SiO_2, is calculated as follows: $X_{SiO_2} =$ moles of SiO_2/(moles of SiO_2 + moles of TiO_2+ moles of Al_2O_3 + etc.).

Expressed in more general terms, in a magma or a crystal composed of chemical components $a, b, c, \ldots i$, mole fraction of the component "i" may be given as

$$X_i = (m_i)/(m_a + m_b + m_c + \ldots m_i), \quad \text{or}$$
$$X_i = m_i/\Sigma m_{a \ldots i}$$

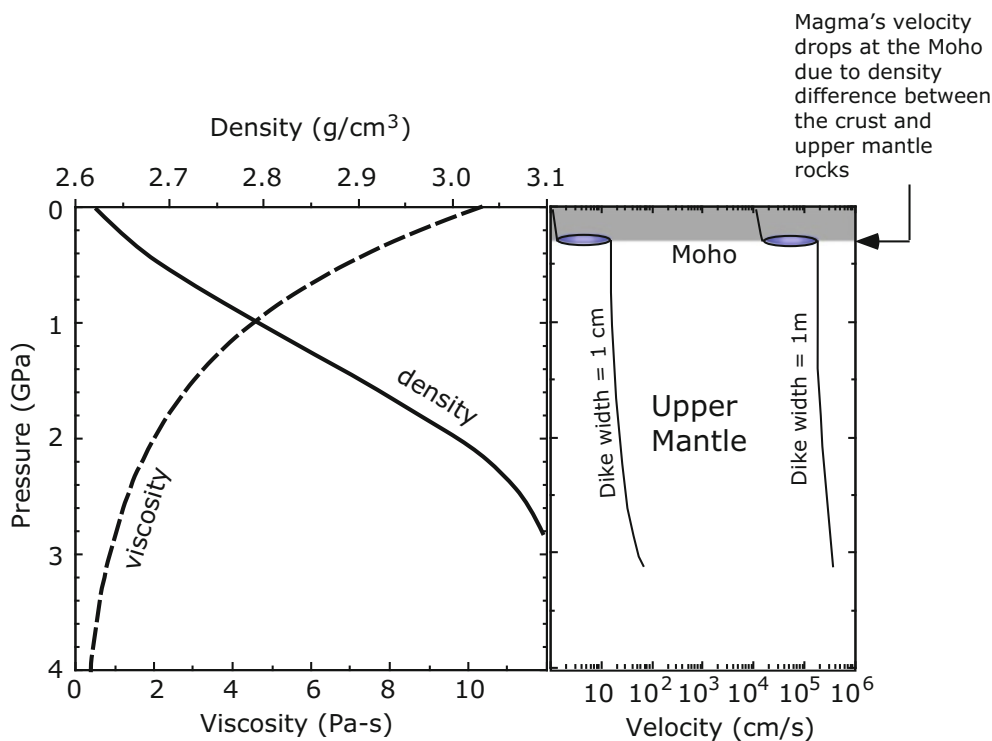where $m_{a \ldots i}$ are moles of $a, b, c, \ldots i$ components.

Fig. 2.9 Density and viscosity of tholeiitic basalt magma as a function of pressure (based on Kushiro 1980)

High-pressure measurements of density and viscosity have been very few (Fig. 2.9). Kushiro's (1980) experiments show that the density of a basaltic magma rises from 2.95 g/cm^3 at 1 atm pressure to about 3.5 g/cm^3 at a pressure of 3 GPa, whereas viscosity decreases drastically.

Earlier we noted that densities of magmas are generally less than those of mantle minerals. However, because magmas are more compressible than minerals, magma's density increases at a more rapid rate than minerals as pressure is increased. Some scientists have determined that ultramafic (komatiitic) magmas become denser than mantle minerals at 11–12 GPa (i.e., 330–360 km: Agee and Walker 1993; Ohtani et al. 1995). The occurrence of such a density crossover implies that magmas generated in the Earth's lower mantle will not leave its source area and instead will sink. The only way they could be buoyant is if some other component, such as dissolved volatiles, lowers their density.

Cooling of Magma and Textural Development

Magma formation and crystallization are in some ways different from how a pure metal, such as gold, melts or crystallizes. Pure metals have a fixed melting/freezing point, above which it is entirely molten and below

which it is solid: for example, note that gold has a single melting/freezing point, which is 1,064.18 °C. In contrast, laboratory experiments have shown that magma solidifies over a temperature range: basaltic magma, 1,230–980 °C; andesitic magma, 1,050–850 °C; and rhyolitic magma, 800–650 °C. This is what is to be expected based on what we will learn from the simple system phase diagrams in Chap. 3.

Figure 2.10 shows the crystallization sequence of minerals from a basalt magma, as determined in a series of equilibrium experiments in a laboratory furnace at atmospheric pressure. The temperature above which a magma is completely molten is referred to as its *liquidus*, and *solidus* is the temperature below which the melt is fully crystalline. In Fig. 2.10, the liquidus and solidus temperatures are 1,250 °C and 980 °C, respectively.

Figure 2.10 shows that crystals and melt coexist within the temperature interval between solidus and liquidus. Crystals of different minerals form at different temperatures in the magma. Under equilibrium conditions of cooling, magma is slowly cooled such that crystals and melt can exchange chemical components as dictated by their crystal/melt equilibrium partition coefficients (explained later). This type of slow cooling can only occur in deep-seated magma chambers but not at Earth's surface, where erupted
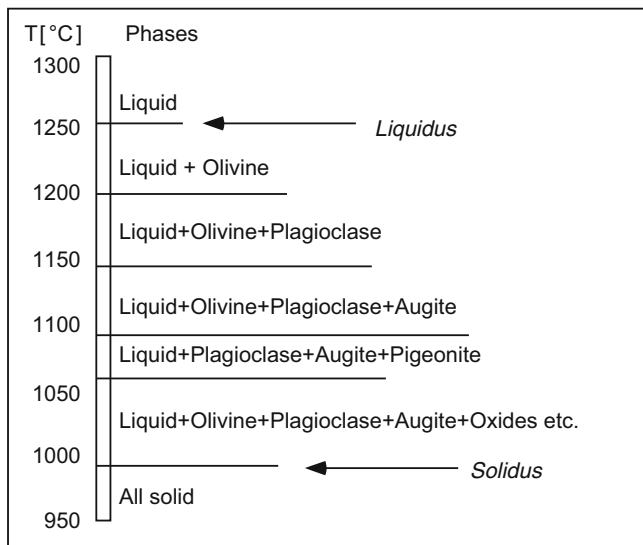
Fig. 2.10 Equilibrium crystallization of a basalt magma composition at atmospheric pressure as determined from laboratory experiments. This result is specific to the starting composition used, which includes the temperature at which the melt begins to crystallize (called the *liquidus* temperature), the temperature it is fully crystallized (called the *solidus* temperature), and the sequence in which different minerals appear at different temperatures. Other starting compositions will give different results
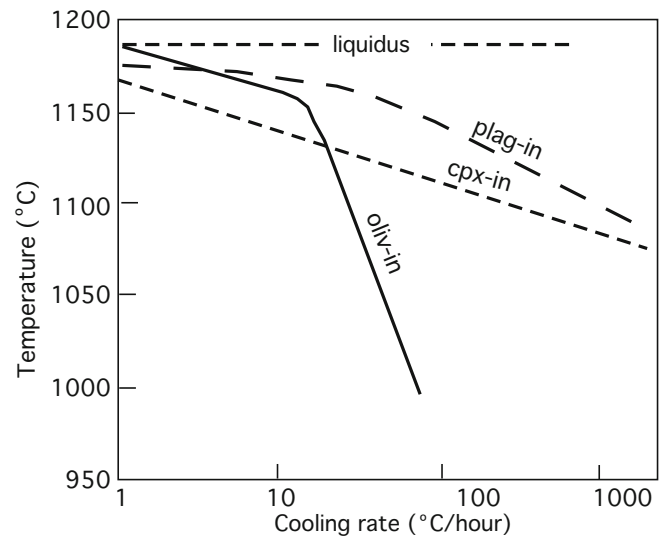

Fig. 2.11 Relationship between undercooling (ΔT = liquidus temperature–actual temperature), nucleation, and growth of crystals. Large undercooling (i.e., rapid "quenching") produces abundant nuclei, whereas small undercooling produces coarse crystals

Fig. 2.12 Crystallization of minerals from a basalt magma at different cooling rates. This figure shows that the sequence in which minerals crystallize may depend on the rate of cooling (after C.E. Lesher, with permission)

controls nucleation of mineral crystals, their chemical composition, and crystal morphology. An understanding of these relationships may be useful in interpreting cooling conditions of lava flows, for example.
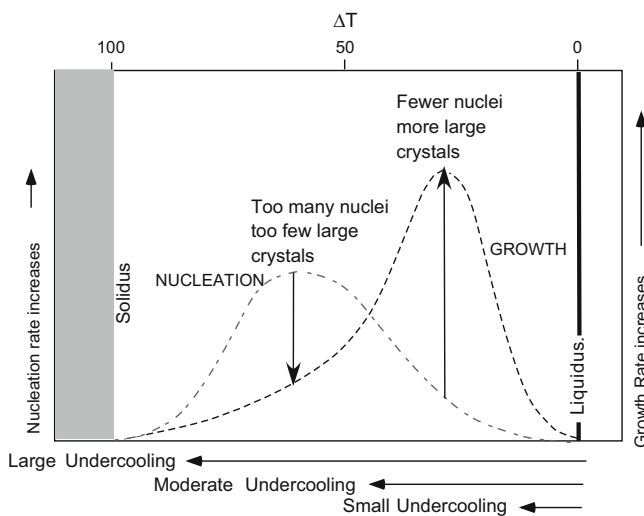
Figure 2.11 shows the relationship between nucleation, crystal growth, and undercooling. When the degree of undercooling is great, large number of nuclei will form; however, none of these nuclei will grow to form large crystals (e.g., arrow pointing downward). At small degrees of undercooling, fewer nuclei will form, but they will most likely grow to form larger crystals (arrow pointing upward). Rapid undercooling to subsolidus conditions will result in the formation of glass. Basaltic lava can be glassy or very fine grained, reflecting rapid undercooling. Cooling in shallow dikes can result in moderate undercooling and therefore result in a coarser-grained texture. Plutons approximate equilibrium conditions and experience very little undercooling, and therefore, plutonic rocks are coarse grained.

Undercooling may delay the nucleation of mineral phases to a great extent and, particularly, suppress the appearance of a phase altogether, as illustrated with a series of experiments on a basaltic starting melt (Fig. 2.12). Note that under equilibrium or extremely slow rate of cooling (less than 4°/h), olivine appears at the liquidus, followed by plagioclase and then clinopyroxene. Olivine's appearance (i.e., nucleation) is significantly suppressed at a cooling rate between about 10 and 100 °C/h such that it appears last in the

lava cools much faster, i.e., it is *undercooled*, as it rapidly loses heat to the atmosphere and the ground. The rate at which magma is cooled below its liquidus, expressed as °C/s, is referred to as rate of undercooling. The rate of undercooling is an important factor that
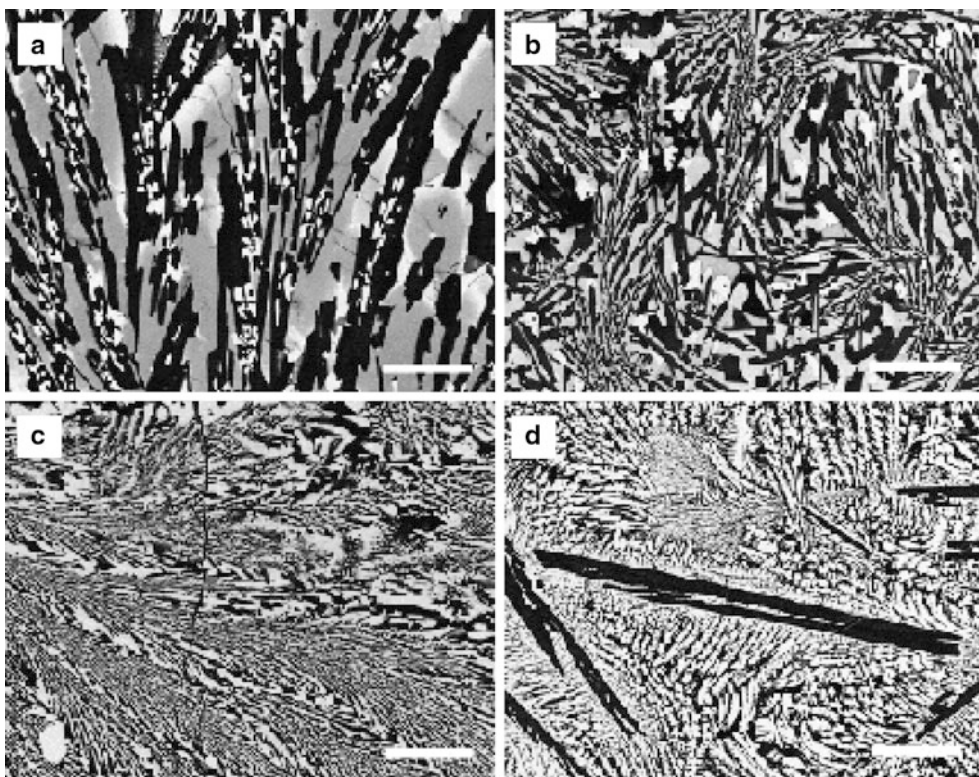
Fig. 2.13 Cooling rate versus run textures: cooling rate increases from (**a**) to (**d**). *Dark*, plagioclase, and *light*, pyroxene (back-scattered electron image: from Lesher et al. 1999)

crystallization sequence. Even more rapid cooling rate completely suppresses nucleation of olivine.

Morphology of crystals is also strongly affected by the rate of cooling. Crystals of olivine, pyroxene, and plagioclase are generally very skeletal or feathery at high cooling rates, whereas at slower cooling rates, these minerals assume their equilibrium shapes (i.e., lath to tabular shapes for plagioclase and euhedral-subequant shape for olivine and prismatic–subprismatic for pyroxene). Figure 2.13 shows the relationship between crystal size, morphology, and rate of cooling in four runs, in which (a) represents the slowest cooling rate (therefore, the coarsest grains) and (d) represents fastest undercooling, and plagioclase forms feathery dendritic growths here.

Undercooling is also related to compositional zoning (zoning refers to contrasting chemical compositions of different parts of the same crystal) in minerals, particularly in plagioclase (discussed in a later chapter). Slowly cooled plagioclase crystals in plutonic intrusions are generally unzoned, whereas plagioclase crystals in shallow intrusions and matrix of lava are strongly zoned. Pyroxenes also show strong chemical variation due to rapid, metastable crystallization due to rapid cooling.

> **Box 2.2: Experimental Petrology**
> *Experimental petrology* is a branch of petrology that attempts to simulate the deeper crustal and mantle environment in which magmas form and crystallize. The instrumental setup for experiments varies considerably based on whether one wishes to examine phase relationships at 1 atm or at higher pressures. Whereas one-atmosphere furnaces are useful for the purpose of simulating crystallization in lava lakes and shallow magma conduit, anyone trying to understand melting at upper mantle conditions or magma crystallization in the mantle or deep continental crust will need to use a piston-cylinder apparatus (0.3–4 GPa). Experiments at higher pressure (5–30 GPa) require a multi-anvil apparatus or a diamond-anvil apparatus for simulating the extreme pressures of the deep

Box 2.2 (continued)

lower mantle or the outer core of the Earth. Pictures and descriptions of such instruments may be found in a number of websites; the "Magma Factory" website of Tokyo Institute of Technology (http://www.geo.titech.ac.jp/lab/takahashi/takahashilab.e.html) is an excellent place to look at some of these instruments and what they do.

To study magma crystallization at atmospheric pressure, the starting material (powdered mixture of oxides, rock, or minerals) is placed in a suitable capsule (graphite, iron, platinum, etc.) or is compressed into a pellet and hung from a platinum wire and then inserted in a furnace in which the partial pressure of oxygen can be controlled. It is important to control oxygen pressure for any system that contains iron, because oxygen influences the proportion of Fe^{2+}/Fe^{3+} ratio in the melt, which in turn controls phase equilibrium behavior (particularly that of the oxide phases) of the melt during crystallization. It is then held at a certain temperature for a long time (hours to days) and then quenched by dropping the capsule/pellet into water or other liquid. Such long times are used to ensure that the run products have reached equilibrium. The run products are then examined with a petrographic microscope, scanning electron microscope, or an electron microprobe to determine the phases (mineral types and/or glass), their proportions (mode), and their chemical composition. Undercooling experiments are done in a furnace where the temperature controller can control the rate in which a run can be cooled. Having said all this, I should note that there is a great deal of variability in the ways experiments are designed and the instruments are chosen.

Experimental petrology continues to play a very important role in our understanding of how the Earth works by providing fundamental information on magma behavior at deeper crustal and mantle pressures. The greatest difficulty lies in "scaling" up such information gathered from tiny capsules to a scale of thousands of cubic kilometers within Earth. We will see examples of these difficulties in various other chapters in this book.

Crystal Size Distributions

The study of distribution of sizes of plagioclase crystals in basaltic rocks can greatly contribute toward understanding crystallization processes in magma chambers (Marsh 1988a, b; Cashman and Marsh 1988). An important aspect of the study of crystal size distributions (CSD) is its simplicity in the data collection process. In modern times when petrology/geochemistry research seems totally dependent on expensive instruments that only the very rich institutions can afford, application of CSD in petrology is an important breakthrough. All it requires is the use of a standard petrographic microscope to measure the grain sizes and then plotting crystal size (L, crystal length in mm) against population density, which is expressed as $\ln(n(L))$ where $n(L)$ is the number of crystals per volume of rock per crystal size. The interested student is urged to visit Professor Mike Higgins' website (http://geologie.uqac.ca/~mhiggins/CSD.html) and textbook (Higgins 2006), which are excellent resources for CSD research methods.

Basaltic lavas generally show two extreme types of textures—(a) porphyritic texture with coarse phenocrysts of plagioclase $\pm$ olivine $\pm$ augite crystals set in a fine-grained groundmass of tiny plagioclase laths, granules of augite, and Fe–Ti oxide minerals (i.e., magnetite and ilmenite) or (b) aphyric or phenocryst-free texture, in which the groundmass may be all glass or a fine-grained mixture of crystals or crystallites (crystals too small and are barely visible under the microscope with $10\times$ objective) and glass. CSDs of all such lavas have a negative slope in the log–linear plots (Fig. 2.14). The intercept of a CSD on the Y-axis (i.e., 0, y) represents the nuclei density, whereas the slope is a product of time for crystal growth and growth rate. The intercept and the slope of a CSD together reflect the histories of crystal population in a rock. A smooth CSD (Fig. 2.14a) reflects a common crystallization history ("single nucleation event" Marsh 1998) where the larger grains are simply older and plot further away from the (0,y) point. On the other hand, a CSD with two distinct linear segments with a break ("dog leg" CSD) reflects two distinct populations of crystals with two different nucleation histories (Fig. 2.14b). This may be a consequence of two mixed magma batches, for example. Other complications can occur that may change the form of the CSD: for example, in a study of Kiglapait intrusion, Labrador, Higgins (2006) noted that CSDs for plagioclase, olivine, and pyroxene look "humped" at the smallest grain size end and a sort of

"flattening" to "concave up" pattern of the CSD with increasing grain size (Fig. 2.14c). Higgins proposed that this is due to a textural coarsening process called Ostwald ripening in which smaller grains are dissolved and larger grains are enlarged due to growth from interstitial liquid between the larger grains.

Finally, we consider the CSDs of rocks in different parts of a magma chamber or a conduit in which a phenocryst-free magma enters and crystallizes *in situ* from walls inward (Fig. 2.15). If the textures of the rocks formed in different parts of such a chamber or conduit is not further complicated by crystal accumulation, wall rock assimilation, Ostwald ripening, and such, then the CSDs of rocks sampled from different parts of such intrusions may look like those schematically shown in Fig. 2.15.

Box 2.3: Magma Cooling, Crystal Percentage, and Size

Size and percentage of crystals of various minerals in an igneous rock depend upon several factors that include nutrient availability, nucleation rate, crystal growth rate, rate of cooling, crystallization energy, competition from other growing crystals of other phases, etc. Several of these factors are intimately dependent on one another, and therefore, one can use a simple formula, originally developed by Melvin Avrami in 1939–1941 and later modified by others, to understand how crystal percentage and size can vary as a magma body cools. Although the original equation was not developed to understand magma cooling and crystallization, Bruce Marsh of Johns Hopkins University first made use of the Johnson–Mehl–Avrami (JMA) equation to study magma crystallization. Here, we can use a simple exercise based on the JMA equation below and see how crystal size and percentage may occur in magmas (Marsh 1996):

$$\text{Crystal\%} = 100 \times \left[1 - \exp\left[-\frac{\pi}{3} \left(J_o (G_o)^3 \; t^4 \right) \right] \right]$$

In this equation J_o is the nucleation rate, G_o is the growth rate, and t is time.

Let us consider growth of plagioclase crystals in basalt magma. Following Marsh, we ignore complications that are embedded in the assumptions made in the application of the JMA equation and nonlinear variation in the nucleation

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Box 2.3 (continued)

and growth rates and use constant values of J_o of 10^3 cm^3/s and G_o of 10^{-10} cm/s for illustrative purpose.

As a side note, we explore how the above two values came about. They come from the examination of samples cored from a lava lake that formed in a small crater in Hawaii, called Kilauea Iki. In November 1959 Kilauea Iki erupted for 36 days, and the erupted lava formed a 120-m-thick lavas lake. This lake cooled slowly and became fully solidified in 2003 (source: Hawaii Volcano Observatory: http://hvo.wr.usgs.gov/volcanowatch/archive/2003/03_01_09.html). The lake was drilled multiple times from 1960 to 1988, temperatures were measured in situ, and drill core samples were examined for crystallization development. To many petrologists this lava lake represents a good natural analog of a near-surface magma chamber.

Returning to the application of the JMA equation, we need to add one more factor that can relate crystal percentage to crystal size and that is the simple relationship between crystal size (L, cm) and growth rate (G_o): $L = G_o \cdot t$. When we do this, we find that (1) crystallization should be complete at the end of 10 years; (2) maximum crystal size should be 0.32 mm; and (3) crystal% versus time follows a sigmoidal profile, i.e., crystal% changes very little with time at the beginning and at the end of the crystallization process, but the crystal% rapidly increases in the middle of the crystallization process (Fig. 2.16). This gives us some basic understanding of kinetics of crystallization in magmas even though cooling rate and convective overturns, and input of fresh batches of magma, and magma mixing in natural magma bodies may greatly complicate the supply of nutrients that the crystals need for growth.

An interesting observation is that largest crystal of plagioclase in Kilauea Iki is about 0.3 mm long. This compares favorably with the size we calculated earlier using the JMA equation; however, the problem is that the observed crystals attained that size in 53 years! There are several ways to explain this "age" discrepancy, one of which is that the largest plagioclase crystals did not grow much in the last 33 years.
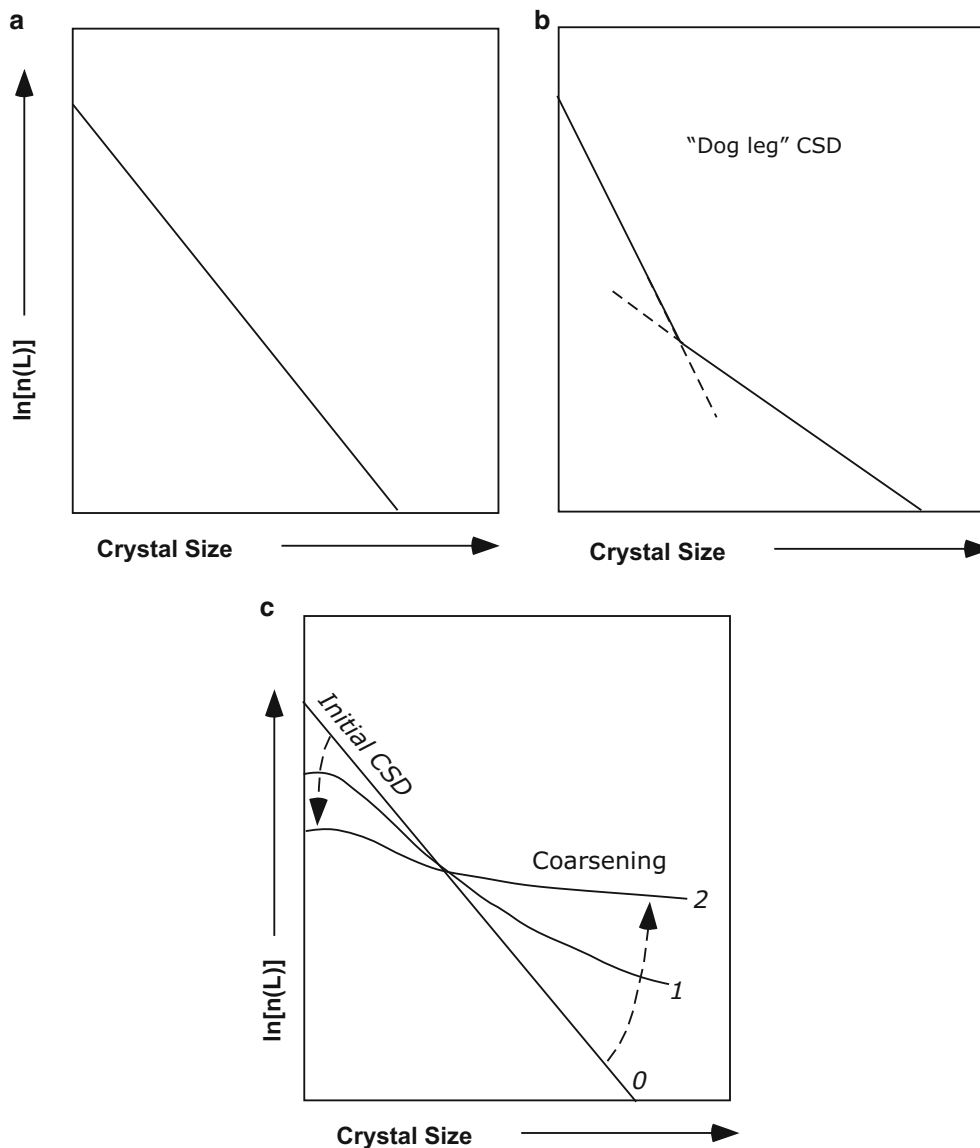

Fig. 2.14 Crystal size distribution (CSD) plots. (**a**) A simple CSD where all crystals in the rock are related by nucleation and growth history. (**b**) A kinked CSD with a "dog leg" expected of a mixed population of crystals with two distinct nucleation and growth histories. (**c**) "Flattening out" of CSD by coarsening ("Ostwald ripening") process in a large intrusive body

Igneous Rocks: Mode of Occurrence

Magma's transformation to an igneous rock via solidification can occur along many different pathways that depend primarily on the conditions of lithostatic pressure, volatile pressure, and rate of undercooling. Minerals that crystallize from it and the *texture* (which represents the physical relationship between the crystals or noncrystalline materials) provide significant clues to how an igneous rock may have formed from magma. As discussed in a later section, it is common for a batch of magma to go through several stages where they may be stored in magma chambers in the crust before eruption (Fig. 2.17). In some cases, the magma may completely solidify below the surface, and in other cases they may erupt to form lava. Thus, based on the depth of emplacement, igneous rocks can be divided into *intrusive*, meaning those that solidify below the surface, and *extrusive* types, which includes all the hardened eruptive products. To be sure, molten lava may not be the only product of an eruption, and extrusion of solid or solid plus gaseous products (called *pyroclastic materials* or *tephra*) is abundant in the case of intermediate and felsic eruptions. Mafic magmas generally erupt as lava and do not have such significant pyroclastic component. Below I describe some general features of volcanoes and intrusive igneous intrusions.
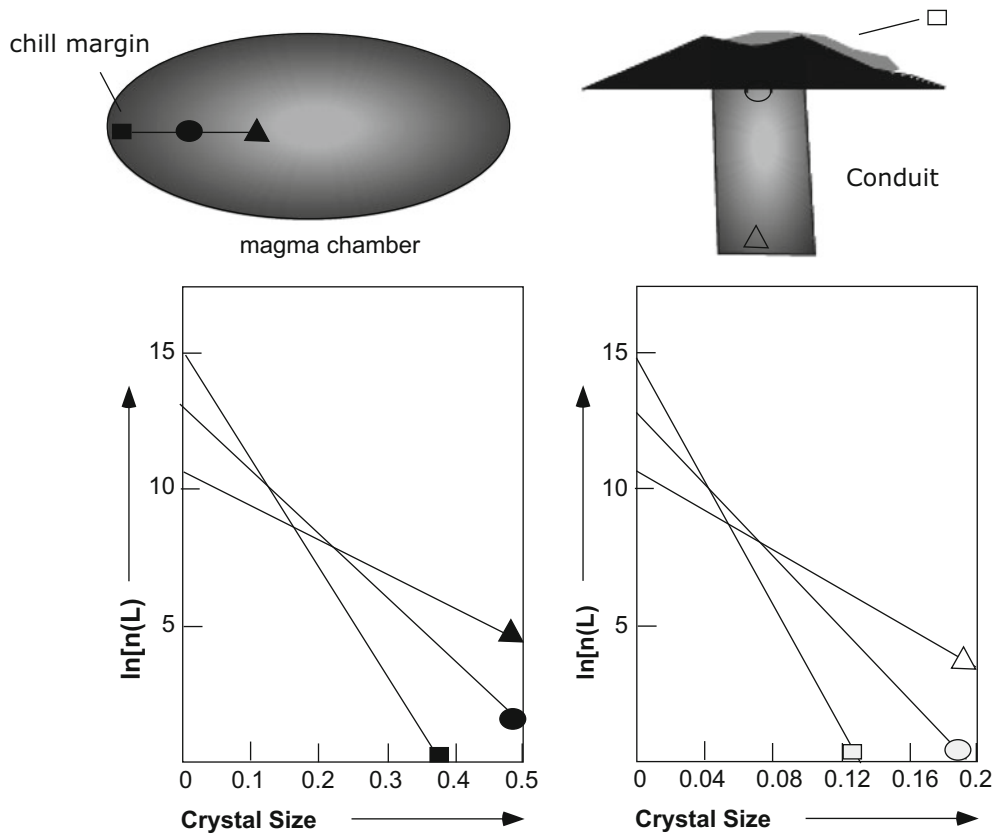
Fig. 2.15 *Left*: CSD of rocks in a large magma chamber. Magma closer to the margin would cool faster where as hotter interior would cool more slowly, resulting in different CSDs from margins to the interior of the intrusion. *Right*: lava quenches more rapidly, forming abundant finer crystals and fewer coarse crystals. In contrast, deep in the conduit magma cools more slowly and therefore forms larger crystals. This difference is reflected in distinct CSDs as well (redrawn from Zieg and Marsh 2002)
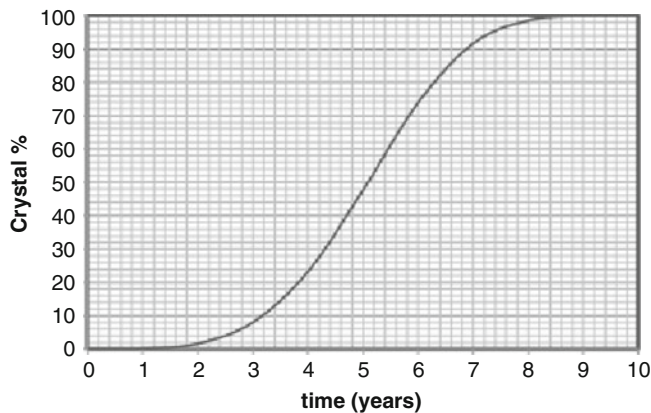
Fig. 2.16 Crystal% versus time as calculated from JMA equation

Volcanoes and Their Products

As discussed in Chap. 1, igneous activities predominantly occur along plate boundaries, and intraplate volcanism, such as the Hawaiian–Emperor chain in the middle of the Pacific plate, is rather rare on a global
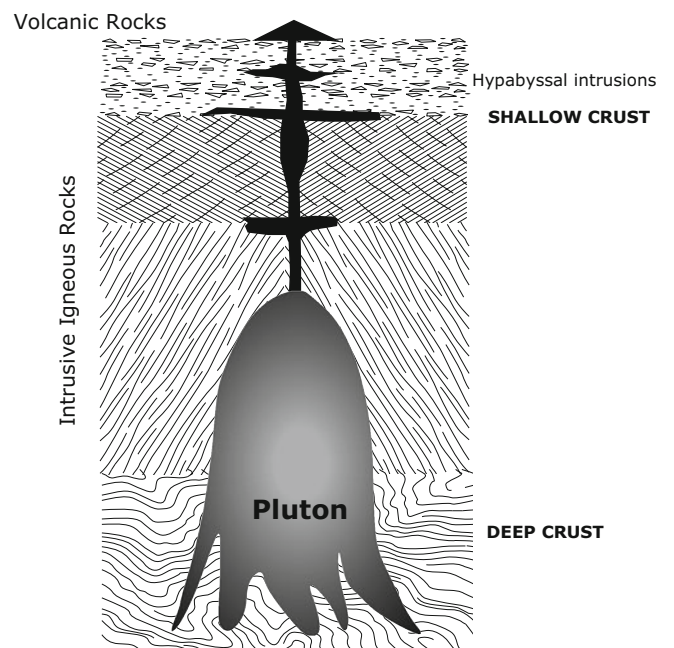
Fig. 2.17 Plutons, hypabyssal intrusions, and lava flow

Table 2.3 Volcano Explosivity Index (VEI)

VEI	Description	Ash plume height	Ejecta volume	Frequency	Example
0	Nonexplosive	<100 m	<10,000 m^3	Daily	Mauna Loa
1	Gentle	100–1,000 m	>10,000 m^3	Daily	Stromboli
2	Explosive	1–5 km	>1,000,000 m^3	Weekly	Galeras (1993)
3	Severe	3–15 km	>10,000,000 m^3	Yearly	Nevado del Ruiz (1985)
4	Cataclysmic	10–25 km	>0.1 km^3	≥10 years	Soufriere Hills (1995)
5	Paroxysmal	>25 km	>1 km^3	≥50 years	St. Helens (1980)
6	Colossal	>25 km	>10 km^3	≥100 years	Mt. Pinatubo (1991)
7	Super-colossal	>25 km	>100 km^3	≥1,000 years	Tambora (1815)
8	Mega-colossal	>25 km	>1,000 km^3	≥10,000 years	Toba (73,000 BP)

scale. Lava eruption can occur through volcanoes or through cracks (called *fissures*) in the crust or on the sides of a volcano. Some magmas erupt with great fury, spewing out large amounts of ash and other *pyroclastic materials*. Large pyroclastic eruptions can generate an ash column that can rise above 500 m in the atmosphere and circle the globe at more than 400 km per hour as part of the jet stream. For example, the June 1991 eruption of Mt. Pinatubo (Philippines) produced 5 km^3 of pyroclastics and a 400-km-wide ash cloud that rose 35 km and encircled the globe several times over a 2-year period, producing beautiful sunsets in the areas over which it passed and dropping Earth's annual mean temperature by 1 °F over a 2-year period.

Newhall and Self (1982) introduced the concept of a Volcano Explosivity Index (VEI) that is based on several factors, such as the volume of eruptive products, ash cloud height, and some qualitative observations, to categorize various volcanic eruptions (Table 2.3). A value of 0–8 is assigned to eruptions that range from "nonexplosive" to "mega-colossal" types.

The explosivity is directly related to how much gas (mainly steam) the magma carries when it nears the surface. Intermediate to felsic composition magmas generally contain a relatively greater amount of dissolved water that they inherit from their source region of melting. As these magmas ascend to shallow depths, they exsolve (i.e., they boil and release the vapor into the surroundings) the dissolved water— much like the release of pressurized CO_2 gas when a bottle of carbonated beverage is opened. This exsolved vapor blasts through the overlying column of rocks, producing a volume of tephra.

Explosive eruptions may also result from heating of groundwater (often referred to as *meteoric* water as opposed to *juvenile* water that comes from the magma itself) by a shallow magma body. The groundwater turns into steam, which has a much greater volume than water, and therefore, the steam pulverizes the

ground above it into tephra. This type of eruption is called *phreatic* eruption.

Tephra and Lava

Based on the size of individual particles, the tephra is classified into *ash* (particle diameter <4 mm), *lapilli* (4–32 mm), and *blocks* or *bombs* (>32 mm). Bombs typically have an ellipsoidal shape with flanges at the two sharp ends because they usually have a molten lava coating on a core of a rock fragment as they are thrown up in the air. The lava quenches to a glass as it drops back on Earth. Blocks have sharp edges and lack the roundness of bombs. *Tuff* is a type of rock that represents hardened pyroclastic materials.

Whether or not magma would flow out of a volcano as lava or barely rise to fill or partially fill the volcanic crater as a *dome* (a simple domal structure) depends on the magma's viscosity, which is in turn linked to its composition and dissolved vapor content. Felsic lava is so viscous that it normally does not easily form a lava flow and instead form domes inside a volcano. Mafic magmas are less viscous and therefore form lava flows. The most common type of mafic magma (called *tholeiite*, discussed later) is the least explosive, and their corresponding lavas can flow over great distances. In fact, some lava flows of the 16-million-year-old Grande Ronde formation of the Columbia River Basalt Group in Pacific Northwest (covering much of Washington and Oregon) are nearly 100 m thick and appear to have traveled hundreds of km from their eruption sites located close to the Washington–Idaho border to the Oregon coast.

Volcano Types

Based on their geometrical form and relative size, volcanoes can be fundamentally of the following types: *shield volcano, stratovolcano, cinder or scoria cone, maar, tuff ring, and tuff cone* (Fig. 2.18). Shield
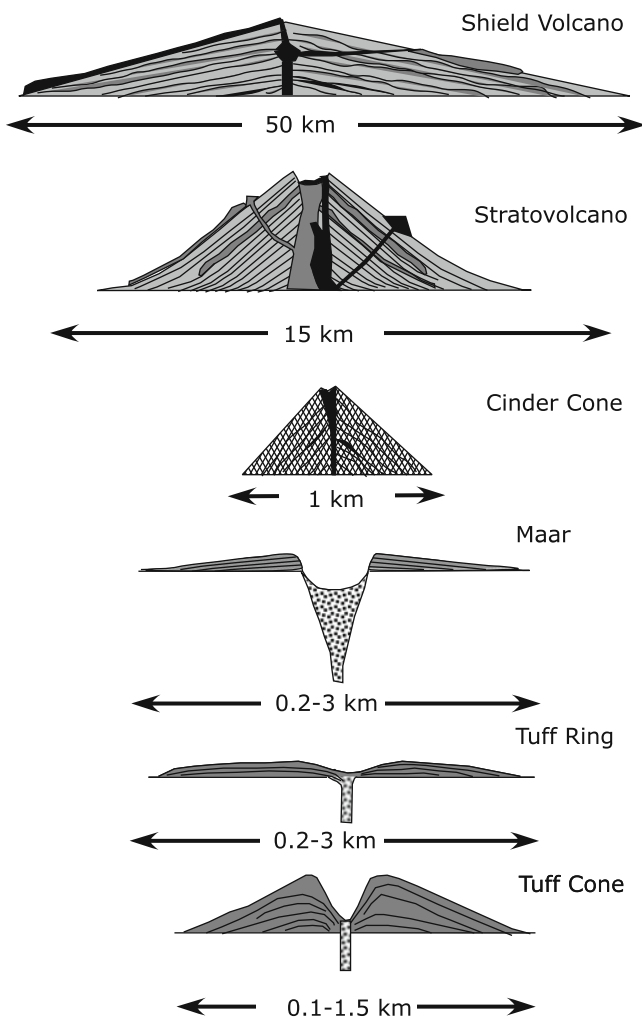
Fig. 2.18 Types of extrusive bodies

they develop structures that appear like folded cloth or entangled ropes. This type of lava is called *ropy* or *pahoehoe* lava (Fig. 2.19). The other type of lava that is associated with shield eruptions is *Aa lava*, which is also called *block lava* because they are composed of blocks and clinkers that travel at a very slow speed. The interior of block lava is generally partially molten and is pushed forward by the molten interior that gets replenished with magma supply from the volcano.

Stratovolcanoes (also called *composite cones*) are typically associated with subduction zones. They are highly explosive and are built of alternate layers of tephra and lava. Lavas are dominantly intermediate in composition, although basaltic and shoshonitic lavas may form a part of them as well (these names are discussed later). The tephra layers are usually associated with intermediate to felsic eruptions. Stratovolcanoes occur all along the continental margins around the Pacific Ocean, which is commonly referred to as the "Pacific Ring of Fire" (Fig. 2.20). Mt. St. Helens (NW USA), Pinatubo (Philippines), and Mt. Fuji (Japan) are all examples of stratovolcano (many spectacular pictures of these volcanoes may be found on the Web).

Cinder cones, maars, tuff cones, and tuff rings are much smaller than either shield or stratovolcanoes, and they can be found in all tectonic environments. They generally form by a single phase of explosive volcanism that can last a year to a few years. Cinder or scoria cones are so-called because they are generally composed of cinders or blocks of mafic lava, some ash, and other pyroclastics. Paricutin volcano in south-central Mexico is a famous example of a cinder cone. It erupted lava and tephra during 1943–1951.

The basalt *scoria* is full of vesicles that form in mafic lava due to escape of vapor bubbles. Sometimes these vesicles are refilled with minerals like zeolite, calcite, and quartz that form from hot water-rich fluids. Such refilled vesicles are called *amygdules*. Often these amygdular minerals grow to be very large (up to 1.3 m) and develop perfect crystal faces, and in such cases, they are called *geode*.

Pumice is similar to scoria in the sense that both are strongly vesicular; however, scoria is black and pumice is light colored because the glass in pumice is felsic in composition. Scoria sinks in water, but pumice floats because it contains many more vapor bubbles that make it less dense than water. Felsic magma is much more viscous than mafic magma, and therefore, it effectively traps the vapor bubbles, whereas in the latter they easily escape.

volcanoes are truly giant volcanoes with gently dipping flanks that take several tens of thousands of years to build with layers and layers of lava. They are best exemplified by the Earth's largest volcano—Mauna Loa (Hawaii)—which rises 4,100 m (13,444 ft) above sea level (Fig. 2.18). When its height is counted from the ocean floor to its peak, Mauna Loa stands 10,099 m (33,132 ft) tall, and by this measure, Mauna Loa is the tallest mountain on Earth—taller than Mt. Everest (8,850 m, 29,035 ft). The largest shield volcano in the terrestrial system is Olympus Mons on Mars (many pictures of Olympus Mons can be found on the web).

Shield volcanic eruptions are relatively "nonexplosive" because they erupt very fluid basaltic lava without emitting gases in any significant amount. Such lava flows can travel for several tens of kilometers. The lavas that come out of shield volcanoes are often so fluid that they can flow like a river, and as they harden,

Ropy Pahoehoe　　　　Aa lava

Fig. 2.19 Pahoehoe and Aa lava
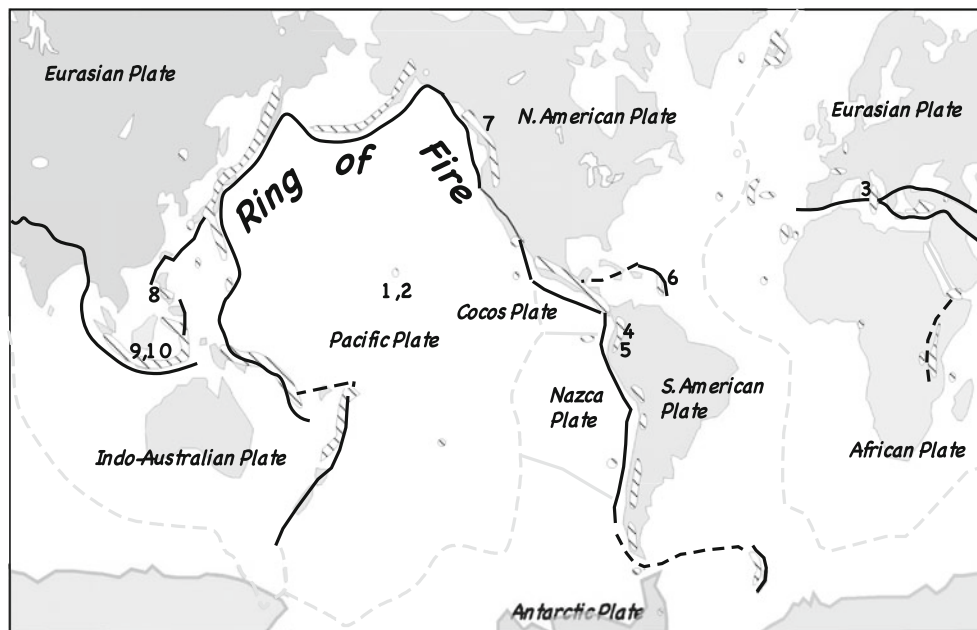
Fig. 2.20 Map of circum-Pacific ring of fire (reprinted with permission from US Geological Survey)

The term *tuff* is used to describe the rock that forms by compaction and recrystallization of pyroclastic materials. Maars, tuff rings, and tuff cones are all built of tephra and are often stratified (layered). They are usually created by phreatic eruptions. Maars have a much wider central depression than tuff cones and tuff rings (Fig. 2.18). The difference between a tuff cone and a tuff ring is in the height of the tuff rim.

Pyroclastic Deposits

Pyroclastic Fall Deposit. In an explosive eruption, pyroclastic materials of all sizes may rise up in a column, which is called a tephra column. The coarsest materials, such as blocks, bombs, and lapilli, along with ash fall back closer to the volcanic crater. As the ash column reaches the higher atmosphere, it spreads laterally to form a rapidly moving ash cloud. Some of the ash falls back to the Earth and forms a thick blanket on over a large area in the downwind direction from the volcano. Such deposits are collectively called *fall deposits* (Fig. 2.21).

Pyroclastic Flow Deposit. In some cases the tephra column may collapse and the tephra flows downslope at tremendous speeds (about 100–150 km per hour). Rapid pyroclastic flow downslope can also occur if the tephra blasts off the side of a volcano. This type of flow is called *nuée ardente* (glowing avalanche). Eventually the flow comes to rest at a distance away from the volcano.
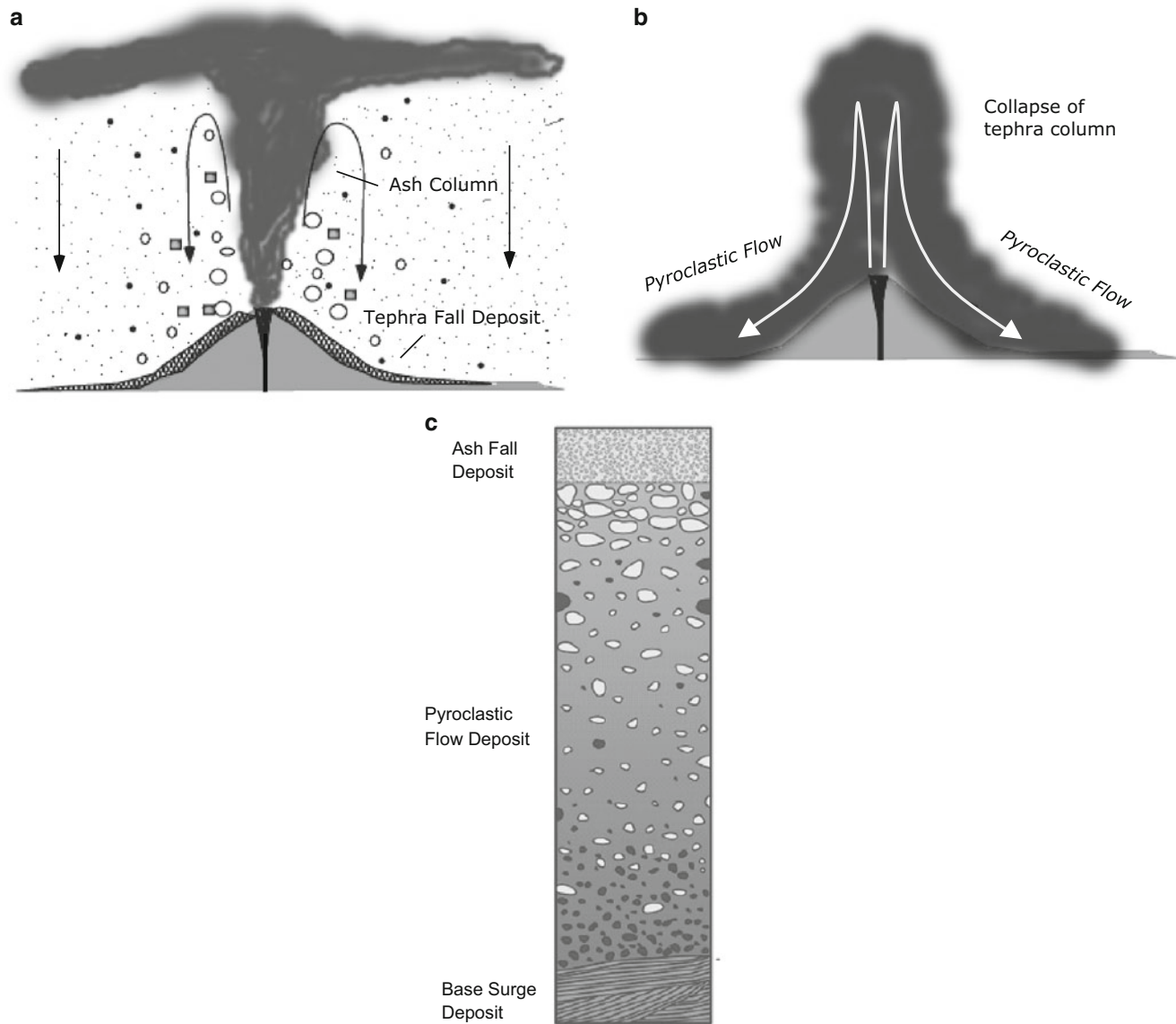
Fig. 2.21 Pyroclastic eruptions and their deposits. (**a**) Ash column and fall deposits. (**b**) Pyroclastic flow (nuee ardente). (**c**) Deposits from ash, base surge, and flow deposit

In recent US history, Mt. St. Helens produced a major pyroclastic flow via lateral blast. The deposit that forms from pyroclastic flow is called a pyroclastic flow deposit (Fig. 2.21c). *Ignimbrite* is a special class of pyroclastic flow deposit that is associated with silicic volcanism.

Base Surge Deposit. Some pyroclastic flows move at a tremendous speed while hugging the slope but cannot go too far from the volcano. Deposit formed from such flow is generally stratified with well-developed parallel and cross-stratification and is called a *base surge deposit*.

Lahar. It is an Indonesian term that refers to dangerous mudflows that are triggered by explosive eruption. Such mudflows are mixtures of volcanic ash, rain, or ice/snow melt water and can rapidly cover large areas.

Magma Emplacement in the Crust: Plutons and Hypabyssal Intrusions

Tectonic uplift and erosional processes sometimes expose magma's storage areas, called intrusions or intrusive bodies, in the crust. Study of these intrusive bodies tells us that they can be of various shapes and sizes. The term *pluton* is generally used to describe moderately large to really large intrusions that occur at depths greater than about 3 km (Fig. 2.22). At such deep levels, the crustal rocks are considerably hotter, and therefore, a magma emplaced in a pluton loses heat

Dike   Sill

←10 km→

Lopolith

Laccolith

Cone Sheet

Ring Dike

Batholith
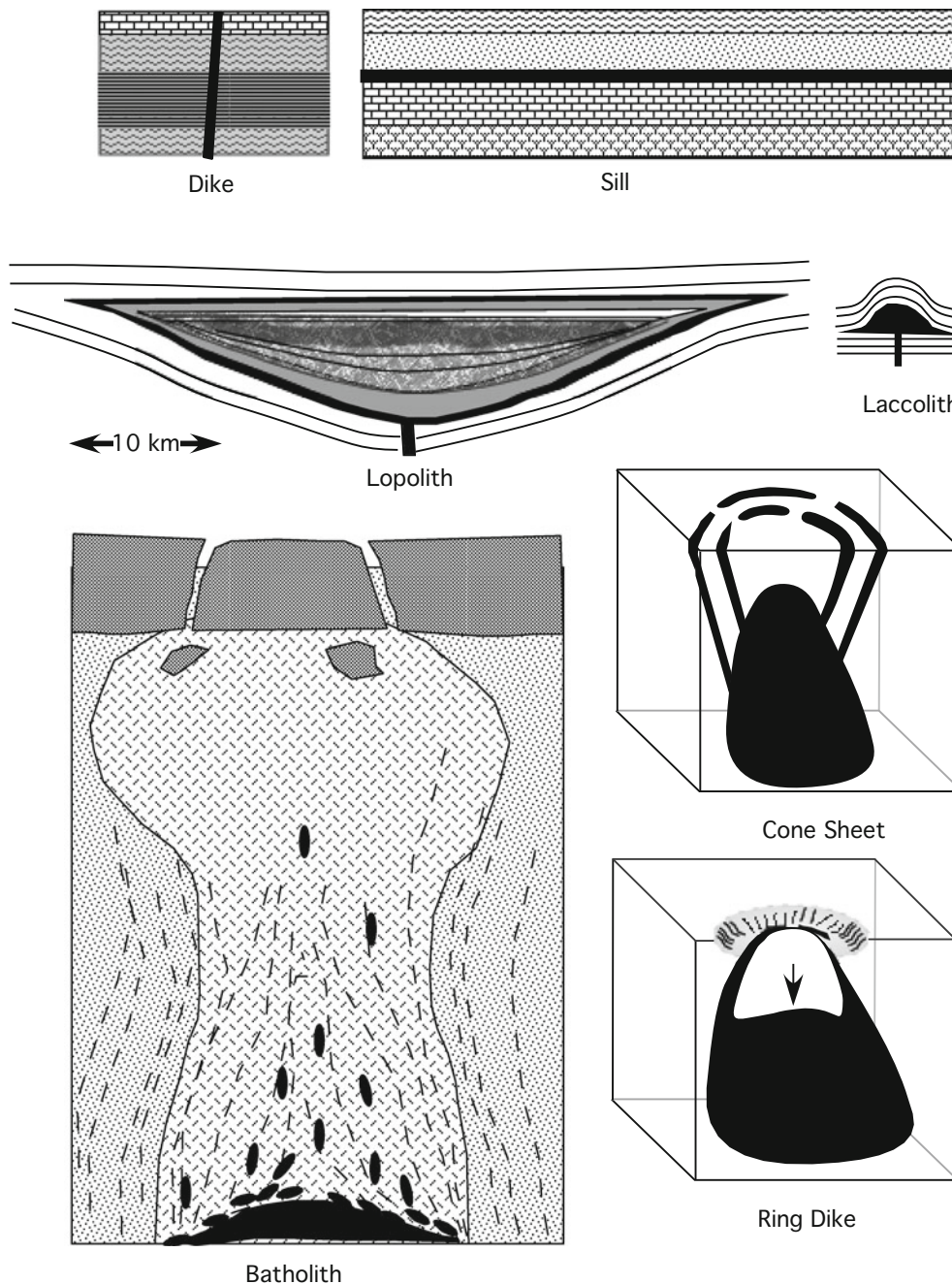
Fig. 2.22 Size and forms of igneous intrusions

very slowly to the surrounding crust. Really large plutons (hundreds of km in diameter) can take hundreds of thousands of years to completely solidify. Such slow cooling allows crystals to grow to large sizes while maintaining chemical equilibrium with the magma. Thus, plutonic rocks are coarse grained, free of glass, and the minerals in them are generally unzoned. Shallow level (0–3 km) intrusions are called *hypabyssal intrusions* and rocks in them have an intermediate grain size.

The terms *concordant* and *discordant* are used to describe the nature of contact between the intrusion and surrounding wall rocks (often called *country rocks*): an intrusion is said to be discordant when they cut through the wall rocks and concordant when they are parallel to the structures, such as the layers, in the country rock. Below is a list of names assigned to plutons and hypabyssal intrusions based on their size, shape, and relationships with the wall rocks.

Plutons (Fig. 2.22)

Laccoliths. Laccoliths are dome-shaped concordant plutonic intrusions. They acquire this form by pushing up the overlying rocks during magma emplacement.

Lopoliths. Plutons of mafic magmatic composition tend to have the form of a *lopolith*, which is a broad funnel- or keel-shaped intrusion whose exposed surface area could be as much as 64,000 km^2 (e.g., Bushveld intrusion of South Africa). Lopoliths generally have a gently "sinking" roof and a discordant, funnel-shaped bottom. Their large size allows magma to cool very slowly and developed layers of different rock types (discussed later).

Batholiths and Stocks. Felsic and intermediate magmas generally form batholiths, which are very large, discordant, and complex bodies of vast aerial extent. In terms of geometrical form, they can vary from large lenticular bodies to classic "bell jar" shapes. Their bottom is generally not exposed. Smaller batholiths are called *stocks*. Stocks are likely simple apophyses or "tongues" of larger batholiths below. A single batholith is generally not a single-phase intrusion but may grow in size over thousands of years by multiple phases of magma intrusion. *Mafic enclaves* or blobs and lens-like bodies of mafic rock are common in felsic batholiths. We will discuss more about batholiths in a later chapter.

Hypabyssal Intrusions

There is a large temperature contrast between the wall rocks and the magma when magma is emplaced at shallow depth. As a result, the magma forms a fine-grained or glassy margin along the contacts of a hypabyssal intrusion with the wall rock. Such a margin is called *chill* margin. Chill margin composition of an intrusion can give valuable clues to the nature of the original magma—its chemical composition, amount and composition of dissolved volatiles, and whether or not phenocrysts were present in the magma at the time of emplacement. The interior of hypabyssal intrusions is generally coarser grained, and the grain size may vary depending upon the rate of cooling of the intrusion. The coarsest grains tend to grow in the upper zone of an intrusion, where the grain size is not only due to slow cooling but also aided by concentration of volatiles. Occasionally, xenoliths of the wall rocks or larger crystals (megacrysts) or rocks (*cognate* xenoliths) formed from an earlier batch of magma may be found in an intrusion.

Dike. A dike is a tabular, discordant, intrusion. Dikes are commonly formed by basalt magmas and can vary from a few cm in width to tend of meters. The biggest dike on Earth is the Great Dyke of Zimbabwe. A system of parallel and cross-cutting dikes and sills, called a *dike swarm*, is often common in continental flood basalt provinces; for example, a dike swarm that fed many of the lava flows of the Columbia River Basalt Group is exposed close to the border of Washington and Idaho states in northwestern United States.

Sill. A sill is a tabular concordant intrusion. They can be several kilometers long and over a hundred meter thick. One of the best-known sills in the world is the *Palisades sill* whose outcrops are best exposed along New York–New Jersey State line on the New Jersey side of the Hudson River.

Ring Dikes and Cone Sheets. Ring dikes and cone sheets are a system of concentric dikes that form a ring-like structure on the surface (Fig. 2.23). The difference between the two is the dikes dip toward the center of the underlying pluton in a cone sheet, whereas they dip away from each other in a ring dike. Ring dikes form because of roof collapse when the magma pressure decreases in the underlying pluton. During such collapse, magma oozes out as dikes surrounding the collapse caldera. On the other hand, cone sheets form by magma pressure, which fractures the overlying roof rocks and magma then emplaces along such fractures.

Pegmatitic Intrusions. Pegmatites are extremely coarse-grained silicic rocks in which individual grains may grow to be several meters. They usually form lens-like intrusive bodies around a felsic batholith or stock. Such strong grain growth is attributed to crystallization in equilibrium with a vapor phase. Pegmatites are volumetrically almost insignificant when compared to basalts, andesites, and granites; however, they commonly host economically important ore minerals like gold and gemstones such as *emerald* (scientific name—Beryl).

Diatremes. These are essentially maars or tuff rings when viewed at the surface; however, their feeder dike is usually 1–2 m wide and reaches deep into the upper mantle. They are considered *a* special class of intrusion because they typically host diamond-bearing *kimberlite*s. Kimberlites are believed to be fluid-rich magma that rises from as deep as 300 km and carries the deepest mantle rocks found anywhere.

Elements of Petrography of Igneous Rocks

Physical description of the appearance of a rock is the very first step of any petrologic study. *Petrography* deals with such descriptive or observational aspects of a rock, and *petrogenesis* refers to processes that lead to the formation of an igneous rock. A good petrographic
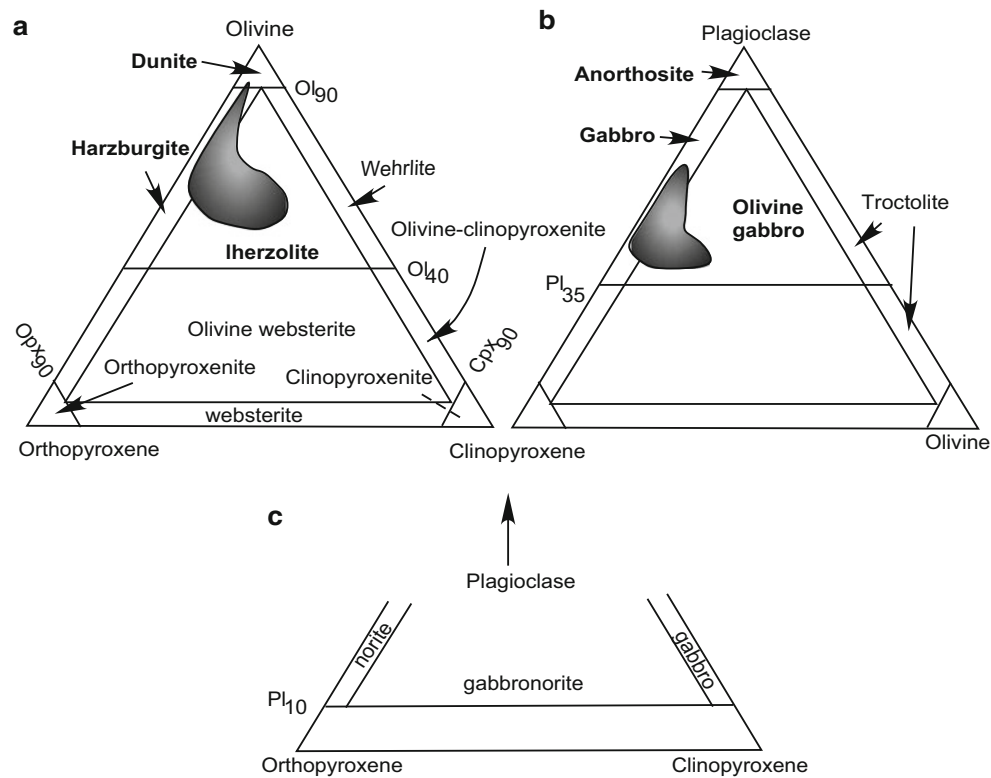
Fig. 2.23 Classification of mafic and ultramafic rocks

description of a rock should be such that it will afford the reader a clear mental picture of what the rock looks like without even looking at the actual specimen. It is also helpful to draw sketches of the rock or to take photomicrographs with a microscope.

There are two aspects of petrography—*mineralogy and texture*. A typical petrographic description must begin with the identification of the *essential minerals* (i.e., the dominant mineral components of the rock) and the *accessory minerals* (i.e., the minor mineral constituents). The next step is to provide a visual estimate of the *mode* (i.e., proportions of minerals). If necessary, one may wish to obtain quantitative modes by *point counting* under a microscope or image processing methods from back-scattered electron images of the rock obtained via a scanning electron microscope. Normally, a visual estimate is sufficient for routine descriptions.

In describing the texture of an igneous rock, the following aspects must be brought out:

Crystallinity

Crystallinity reflects the relative abundance and development of crystals in an igneous rock. When lava solidifies rapidly so much so that no mineral can nucleate, then the resultant volcanic rock is *glass*. The terms *glassy* and *holohyaline* are used to describe the texture of such glass. However, if the magma solidifies less rapidly, the resulting rock texture is called *aphanitic* if it is a mixture of glass and fine crystals called *crystallites* (which are so small that their individual grain outlines cannot be easily deciphered with a petrographic microscope) or *microlites* (the crystals are more visible but still extremely small). It refers to the extent to which a magma has crystallized. The terms *holocrystalline* (fully crystallized, generally medium to coarse crystals; somewhat analogous to the term *phaneritic*), *microcrystalline* (crystalline but crystals are very small (<1 mm) even under a microscope: also, *aphanitic*), *merocrystalline* (mixture of crystals and glass), and *holohyaline* (*glassy*) are generally used.

Sizes, Shapes, and Habits of Crystals of the Essential Minerals

In general, the terms *coarse* (>5 mm), *medium* (1–5 mm), and *fine* (<1 mm) are used to describe the overall grain size. In the case of very large (cm to m scale) crystals of pegmatites, the term "*pegmatitic*" is used to describe such crystals.

The terms *euhedral*, *subhedral*, and *anhedral* are used to describe, in a relative sense, how well-developed crystal faces are for each mineral phase. A mineral grain is said to be euhedral when all/most of its faces are smooth and well developed, and anhedral is just the opposite—none of the faces are well developed. A subhedral grain has some of its faces well developed. The shape of a grain is described by using such terms as *equant*, *subequant*, *prismatic*, *subprismatic*, *tabular*, *lath shaped*, etc.

In rapidly quenching lava, minerals often do not have the time to form well-developed crystals and instead form feathery, dendritic (i.e., like a tree), long, optically discontinuous crystals. This type of crystals is known as a *skeletal* crystal.

Specific Textural Terms for Description of Rocks in the Field

Texture or fabric of a rock is essentially a description of the geometrical relationships between individual mineral grains (and/or glass or other materials) of a rock. Below is a set of textural terms used in field description of a rock.

Glassy. Rapid quenching of liquid forms a glass. Therefore, the texture of a glass or a glass-rich rock is glassy. Obsidian is well known for its glassy appearance.

Spinifex. Ultramafic lavas of Archean age are characterized by this texture. In these lavas long (few cm), discontinuous, skeletal crystals of olivine rapidly grew toward the interior of the lava flow from the flow tops, giving the lava the appearance of inverted slabs of grass. The term spinifex is based on a type of grass found in southern Africa.

Aphanitic. The texture is called *aphanitic* if the rock is very fine grained so that individual grains are not discernible.

Phaneritic. This term refers to coarsely crystallized texture.

Porphyritic. Porphyritic texture is defined by the presence of conspicuously large crystals set in a groundmass of finer-grained minerals or glass or a mixture of glass and finer crystals. If the phenocrysts form clusters, then the texture is called *glommeroporphyritic texture*.

Aphyric. When a rock lacks any distinct phenocrysts, its texture is described as aphyric. This texture has held special significance in igneous petrology because aphyric textured rocks are generally thought to represent actual liquids.

Vesicular. This texture is common in basalt lavas, which contain numerous holes that were once filled with vapor bubbles that left during the eruption and solidification of the lava. It is also common for the vesicles to be subsequently filled with secondary minerals or mineraloids, such as quartz, amethyst, and calcite, when the texture is called *amygdular*

and the fill-in materials are referred to as *amygdules*. Some of these amygdules can be basketball sized (or even bigger) and have well-formed crystals projecting toward the interior of the hole. They are called *geodes*.

Pumiceous. This type of texture usually forms in felsic lavas where vesicles comprise more than 50 % of the rock (light colored) and such a rock, called *pumice*, floats in water. Basalt lava can also have more than 50 % vesicles, but it would not float in water because the rock is denser. Such a texture is called *scoriaceous* and the rock is called *scoria*.

Flow Banding or Trachytic. In flowing viscous lava, lath-like crystals of plagioclase (and other minerals) may orient themselves parallel to the flow direction, giving rise to what is known as flow banding. This texture is typical of trachytes but is also found in other volcanic rocks.

Spherulitic. This type of texture is formed by intergrowth of fine, radiating, needlelike crystals. It is common in some glassy rocks where the centers of individual spherules serve as nucleation centers for spherulitic crystallization. In obsidian, spherulites of quartz and feldspar can develop during devitrification (i.e., loss of water during cooling) of glass.

Ocellar. This is a texture found typically in lamprophyre dikes: individual *ocelli* consists of concentrically banded spheres wherein the bands consist of alternating light-colored and dark-colored minerals. They are unlike spherulites and may form liquid immiscibility or by late-stage melt emplacement into gas cavities.

Graphic/Granphyric/Eutectoid Intergrowth. This texture is defined by a hieroglyphic type intergrowth of quartz and alkali feldspar, usually formed during eutectic crystallization of these two minerals. The rock that has this distinctive texture is commonly called a *graphic granite*.

Microtextures

Textures, as viewed under the microscope, may be broadly divided into two groups—*inequigranular*, in which two or more distinct size groups of crystals occur, and *equigranular* (all grains are roughly equal in size) textures. If distinct size groups do not occur, and instead the grain size variation is a continuum between large and small sizes, then the term *seriate texture* is used. The terms used in field descriptions are also applicable to microscopic descriptions. As needed, additional expressions are used to describe the texture of a rock under the microscope.

Inequigranular Textures

Porphyritic Texture. As described above, in a porphyritic textured rock, phenocrysts are embedded in a finer

groundmass. It is a common texture among volcanic rocks. Some volcanic rocks may be aphyric, i.e., devoid of phenocrysts. Sometimes authors distinguish between megacrysts, phenocrysts, and microphenocrysts, based on whether it is visible in hand specimen (megacrysts and phenocrysts, the former being much larger) or only under the microscope (microphenocrysts). Some authors feel that the term megacryst is simply a phenocryst that grew to much larger size but essentially in the same magma (or related) in which it is found. The same authors would use the term *xenocryst* to refer to crystals that clearly was an accidental inclusion (such as wall rock fragment) in the magma in which it is found.

Poikilitic Texture. It refers to a texture in which small crystals are entirely enclosed by larger crystals, and the term *subpoikilitic* is used when such enclosure is only partial. The terms *subophitic* and *ophitic textures* are used to describe the textures of diabase (dolerite in European and Asian countries), a hypabyssal rock in which coarse crystals of augite partially (subophitic) to completely (ophitic) enclose lath shaped to tabular crystals of plagioclase. An oikocryst is an especially large type of crystal that grows late from the magma and encloses a large number of crystals.

Intergranular and Intersertal Textures. Cooling of mafic magma is such that earlier-formed crystals in a rock may trap melt between them. Such melt may rapidly quench later to glass, giving rise to what is called *intersertal* texture. The glass is often not preserved and may become *devitrified* (i.e., alter to clay minerals or mineraloids). On the other hand, the trapped liquid may have enough time to form smaller crystals of plagioclase, pyroxene, magnetite, ilmenite, apatite, and sulfides. Such a texture is called inter-granular texture.

Equigranular Textures

Equigranular textures are more common in plutonic rocks. Such overall textural terms as *panidiomorphic granular*, *hypidiomorphic granular*, and *allotriomorphic granular* are used when the bulk of the constituent grains are euhedral, subhedral, or anhedral, respectively. In large mafic (defined later) plutonic intrusions, layered structure on field to microscopic scales is common. Such rocks are called *cumulates*, and a whole range of textural terms are used to describe such rocks (described in a later chapter).

Other Textural Terms Used in Petrographic Description

Micropegmatitic or Microgranophyric. These terms refer to blebby to hieroglyphic intergrowths of quartz and alkali feldspar in the interstitial areas between coarse grains of mafic plutonic–hypabyssal rocks (gabbro and diabase).

Myrmekitic. Myrmekitic texture refers to an irregular intergrowth of quartz and oligoclase that "intrude" into an alkali feldspar grain. It is thought to be an igneous replacement texture.

Sieve Texture. Sieve texture specifically refers to plagioclase crystals with numerous glass (melt) inclusions giving the appearance of a plagioclase "sieve." They are found in volcanic or hypabyssal mafic rocks.

Mineral Zoning. Zoning in individual crystals is an important feature that is easily visible between crossed polars under a petrographic microscope. We have discussed normal, reverse, and oscillatory zoning in plagioclase in a later chapter. Pyroxenes are also known to show zoning of various types, in particular, sector zoning, patchy zoning, and hourglass zoning. In sector zoning different sectors of a grain go extinct at a time when the microscope stage is turned and with the polars crossed. In hourglass zoning, the sectors go extinct forming an hourglass pattern. In patchy zoning or wavy extinction, a mineral grain goes extinct in patches or in a continuous "wave" as the stage is turned.

Exsolution. Exsolution texture, defined by exsolved blebs or lamellae of one mineral in another, occurs in slowly cooled plutons of varied compositions. In felsic plutons, exsolution features are common—formed by exsolved plagioclase blebs or streaks in potash-rich feldspar (called *perthite*) and vice versa (*antiperthite*). In mafic plutons, exsolution of pigeonite and orthopyroxene lamellae in augite and exsolution of augite in crystals of orthopyroxene and pigeonite are relatively common.

Reaction Texture. Reaction texture develops when a mineral crystal becomes unstable in magma for whatever reason and starts reacting with the melt, and the products of the reaction form a mantle surrounding the crystals of the first mineral. Most commonly, reaction corona develops around a xenolith, i.e., a wall rock fragment, in magma. Formation of pyroxene rim around an olivine crystal in basalt magma is also well known.

Classifications of Igneous Rocks

A Simple "Field" Classification

Chemical composition of magma dictates what minerals may crystallize from it. The rate at which the magma cools, which is related to its depth and mode of emplacement, controls the sizes and shapes of crystals and its texture. Therefore, it is only natural that naming an igneous rock must take into account its

chemical composition, mineralogy, and texture (mode of emplacement). Table 2.2 shows such a general classification. A classification of this type is particularly useful in the field, where usually a pocket lens is the only tool one can use to identify a rock. There is a direct correlation between the hand specimen color of a rock and its content of *mafic minerals* (minerals with relatively high content of Mg and Fe—such as olivine and pyroxene): the greater the mafic mineral content, the darker the rock. Based on color of the hand specimen, igneous rocks are classified into *melanocratic (mafic minerals >60 %), mesocratic (mafic minerals 30–60 %), and leucocratic (mafic minerals <30 %) types*.

Brief Petrography of Some Common Igneous Rocks

Komatiites, Peridotites, and Pyroxenites

Komatiites are ultramafic lavas that erupted in the Archean (with only one exception in the Phanerozoic in Gorgona Island). The most distinctive feature of komaitiites is the presence of long skeletal crystals of olivine and/or pyroxene. As discussed earlier, its texture is called *spinifex*.

Peridotites are generally coarse-grained rocks, and their mode may vary a great deal. They are generally composed of ≥ 60 % olivine (Fo_{88}–Fo_{92}), ≤ 25 % aluminous (>5 wt% Al_2O_3) orthopyroxene, ≤ 15 % diopsidic clinopyroxene (with 4–8 wt% Al_2O_3 and 0.5–1.2 % Cr_2O_3), and an aluminous mineral (either plagioclase, spinel, or garnet). Textures of peridotites may vary tremendously from metamorphic types (foliated, cataclastic, etc.) to more hypidiomorphic granular. In a protoclastic texture, large deformed crystals (protoclasts) show granulation along grain boundary and are set in a groundmass composed of sheared, much smaller, grains. Exsolution in pyroxenes, triple-point junctions between grains, and subgrain boundaries (deformation lamellae) in olivine are generally common in these rocks. Most commonly, these rocks thus display metamorphic textures, which will become apparent upon comparison with metamorphic rock textures.

Pyroxenites are very coarse-grained rocks (i.e., individual mineral grains are visible with the naked eye) that are generally composed of ≥ 90 % pyroxene. Clinopyroxenites (i.e., pyroxenites in which clinopyroxene is dominant) are generally much more common than orthopyroxenites. Their grain size can vary from pegmatitic (extremely coarse) to coarse/medium. The very coarse-grained pyroxenites often have pyroxenes with spectacular exsolution structures. The exsolved phase may form lamellae to highly irregular, blebby shapes. The exsolved phases may be a pyroxene or an aluminous phase (garnet, spinel, or plagioclase). An aluminous phase, commonly a green-colored spinel, plagioclase, or garnet, is generally present in pyroxenites. Olivine is rare to absent. Phlogopite and a brown amphibole are generally present in minor amounts. The textures of these rocks are commonly igneous types, mostly hypidiomorphic granular; however, strongly recrystallized metamorphic textures with triple-point junctions are more common among the smaller-grained pyroxenites.

Basalt, Gabbro, and Diabase

These mafic igneous rocks are all dominantly composed of lath shaped to tabular crystals of plagioclase (commonly of An_{70}–An_{55} composition) and subprismatic to prismatic crystals of augitic clinopyroxene. These two minerals occur in roughly equal proportions in the rock. Pigeonite, and sometimes orthopyroxene, may also occur only in the tholeiitic/olivine–tholeiitic variety of these mafic rocks. Olivine's abundance usually varies between 0 and 15 %. Olivine may often be altered (partially or wholly to an iddingsite (an orange-colored clay or serpentine) but can be recognized by its relict shape. Magnetite and ilmenite are generally present, although their abundance is generally less than 5 %. Besides these, small globular forms of sulfides (usually chalcopyrite, pyrite–pyrrhotite, and pentlandite), formed due to sulfide–silicate liquid immiscibility, are also ubiquitous (<1 %). Tiny (generally <25 μm) euhedral crystals of apatite may also be found in the interstitial spaces.

The difference between the basalt, diabase, and gabbro is principally one of texture. Basalt is a fine-grained or glass-rich rock and is often porphyritic in nature. They may contain vesicles or amygdules. The most common phenocryst assemblages are olivine or olivine + plagioclase or olivine + plagioclase + augite. Intergranular and intersertal textures are common in the groundmass of porphyritic basalt. Diabase is medium grained and has a characteristic ophitic–subophitic texture. Gabbros are coarse grained, and the plagioclase and pyroxene crystals generally form coarse hypidiomorphic granular-type textures. Intergranular texture, defined by the occurrence of pyroxene, magnetite, and ilmenite crystals or microlites in the

interstitial spaces between plagioclase crystals, is also a common feature.

Andesite and Diorite

Andesites vary greatly in their modal composition. They generally exhibit porphyritic texture, dominated by strongly zoned plagioclase phenocrysts, euhedral-to-subhedral red-brown amphibole phenocrysts, and/or pyroxene phenocrysts. The plagioclase composition varies tremendously due to zoning, but the mean composition is $\sim An_{40}$. Cores of individual plagioclase crystals may be as calcic as An_{85}. Amphibole phenocrysts are often rimmed by pyroxene and opaque oxide mineral grains. One interpretation of these rims is that they formed due to dehydration breakdown reaction of the amphiboles as the lavas rose toward the surface. Zoned pyroxene phenocrysts of subcalcic augite, pigeonite, or orthopyroxene are often found. Olivine, biotite, and glass are generally uncommon.

Diorites (broadly used here to include diorite, monzodiorite, quartz diorite, and quartz monzodiorite as defined by the International Union of Geological Sciences, discussed later) are composed of plagioclase, amphibole, and pyroxenes ± quartz. They exhibit more equigranular, hypidiomorphic granular, texture. Plagioclase generally tends to dominate the mode. Granodiorites and other somewhat quartz-rich rocks have similar textures.

Granite and Rhyolite

By definition, the two minerals that dominate the compositions of these rocks are quartz and alkali feldspar. In hypabyssal intrusions and volcanic rhyolites, sanidine, a high-temperature polymorph of K-feldspar, is usually the dominant feldspar. The quartz phenocrysts in lavas are sometimes euhedral, but quartz grains in the groundmass are always anhedral and small. Extremely glassy dark-appearing (in hand specimen) equivalent of rhyolite is recognized as *obsidian*. *Rhyolites* are generally vesicle free. *Flow texture*, defined by curvilinear arrangement of alkali feldspar and quartz crystals, sometimes occurs in hypabyssal intrusions or crystal-laden lavas.

Granite and granodiorite typically form batholiths. Their texture is generally coarse-grained hypidiomorphic granular. They are composed of an alkali feldspar and quartz. Subhedral to euhedral albitic plagioclase crystals may also occur. Whether one or two types of feldspars occur in granite has significance in understanding their depths of origin (discussed further later). Perthites and antiperthites can be found in some granite group of rocks. Biotite and hornblende may occur, and their proportion may vary greatly between different granitic rocks. Granophyric or micropegmatite intergrowth is common in the interstices. Minor euhedral crystals of zircon, apatite, and sphene may also occur in the interstitial spaces.

Syenite and trachyte are similar to granite and rhyolite, respectively, except that they have lesser amounts of normative and modal quartz (note that nepheline syenite and phonolite are alkalic rocks and therefore have normative and modal nepheline). In general, syenites form small plutons and ring-dike complexes.

Rigorous Classifications

Many different classification schemes have been proposed in the literature. Among them, the one proposed by the *I*nternational *U*nion of *G*eological *S*ciences (more popularly known as the *IUGS Classification*, given in LeMaitre et al. 1989) for plutonic rocks is commonly used today. This classification is based on the observed proportion of mineral constituents (i.e., *mode*). The IUGS classification of volcanic rocks is not widely used because in such rock minerals are often not discernible. In the case of volcanic rocks, classification schemes based on chemical or normative (discussed later) compositions are generally used. In this book, we will adopt the IUGS classification of plutonic rocks and LeMaitre's chemical classification of volcanic rocks.

Classification of Plutonic Rocks

The IUGS classification is based on the modal abundance of quartz [Q], alkali feldspar [A], plagioclase [P], feldspathoids [F], and mafic minerals [M]. If the proportion of M, also known as the *color index*, exceeds 90 %, then it is an *ultramafic rock*. Rocks with color index less than 90 are classified in terms of their proportions of Q, P, A, and F contents (discussed later).

Classification of Ultramafic Rocks

Ultramafic rocks are dominantly composed of three minerals—olivine, orthopyroxene, and clinopyroxene. Figure 2.23a shows IUGS classification of ultramafic rocks in terms of *the olivine–orthopyroxene–clinopyroxene triangle*. *Dunite* is a rock with >90 % olivine.

Harzburgites are dominantly composed of olivine and orthopyroxene, with little or no clinopyroxene. *Lherzolites* contain more pyroxenes than a dunite and more clinopyroxene than a harzburgite. Because an aluminous phase (plagioclase, spinel, or garnet) is usually present in a lherzolite, an additional qualifier is added as a prefix to indicate what type of aluminous phase is present: if the aluminous phase is a spinel, then the rock is called a *spinel lherzolite*. Similarly, the names *garnet lherzolite* and *plagioclase lherzolite* are used when garnet or plagioclase, respectively, occurs. It is important to note that lherzolite is the predominant constituent of the Earth's upper mantle. The gray area in Fig. 2.23a shows where the bulk of the ultramafic rocks worldwide plot (gray-shaded area). Other ultramafic rocks, such as wehrlites, websterites, and clinopyroxenites, are much less abundant and are generally of igneous origin.

The majority of upper mantle lherzolites and harzburgites exhibit distinct *metamorphic textures*. It is therefore ironic to consider these important rock types as igneous rocks! Presumably, this tradition has continued because lherzolite is commonly accepted to be the source rock for basalt magmas, which form all of the oceanic crust and much of the continental crust! In such a case, *eclogite*, clearly a metamorphic rock composed of pyrope-rich garnet and an omphacitic clinopyroxene, must also be considered here since it too likely plays some role as a source rock for magmas.

In the literature, the terms peridotite and pyroxenite are loosely used to refer to olivine-rich versus pyroxene-rich ultramafic rocks. Peridotite thus broadly includes lherzolite, dunite, harzburgite, olivine websterite and wehrlite, whereas pyroxenites include mainly clinopyroxenite, orthopyroxenite, and websterite.

Mafic Plutonic Rocks

In this book a slightly modified version of the IUGS scheme of classification of mafic and felsic rocks is used. In this version many of the "nomenclature details" (which are essentially descriptive jargon) have been skipped because the focus of this book is on only those rocks that are most common.

Mafic plutonic rocks are classified on the basis of modal contents of olivine, plagioclase, and pyroxenes (Fig. 2.23b, c). Rocks containing >90 % plagioclase are called *anorthosite*. Anorthosite occurs as layers in plutonic layered intrusions and as massive, km-scale bodies of Precambrian age. Note that the "highlands" on the Moon are largely composed of anorthosite. *Gabbros* (plagioclase + clinopyroxene) and olivine gabbros (gabbros commonly with 5–15 % olivine) are most abundant mafic plutonic rocks (Fig. 2.23b). *Troctolite* (olivine + plagioclase rocks) occurs as layers in layered intrusions. Figure 2.23c shows that if the dominant pyroxene is an orthopyroxene (instead of clinopyroxene), then the orthopyroxene-bearing equivalent of a gabbro is called *norite*. Norites, *olivine gabbronorites*, generally occur in layered intrusions and also may be found associated with anorthosite massifs.

Felsic Plutonic Rocks

The modal proportions of quartz (Q), alkali feldspar (A), plagioclase (P), and feldspathoids ("foid," F) are used to name various rock types (Fig. 2.24). IUGS classification recognized a fundamental distinction between rocks that contain quartz and those containing feldspathoids. We will see in a later chapter that feldspathoids and quartz are not compatible, that is, they cannot occur in the same rock. Thus, IUGS classification has two triangles—QAP and FAP—which are shown to share the PA edge. In the field, more general terms are often used—for example, a rock containing more than about 20 % quartz is referred to as *granitoid*. An exact mode allows a more precise naming of the rock—for example, a rock will be called granite if it has 40 % Q, 40 % A, and 20 % P, because it will plot within the "granite" field. The IUGS classification recommends the use of a complex system of prefixes based on detailed modes; however, such detailed naming is often unnecessary.

The distinction between gabbro and diorite can only be made based on the proportion and nature of mafic minerals (gabbro being generally pyroxene rich and diorite being amphibole rich) and plagioclase composition (gabbro, $>An_{55}$; diorite, $<An_{50}$).

Classification of Volcanic Rocks

There are a number of classifications of volcanic rocks available in the literature. Among these, IUGS's TAS (stands for total alkalis, i.e., $Na_2O + K_2O$, and silica) classification and LeMaitre's classification (1989, Fig. 2.25) are commonly used. Both are based on total alkalis versus silica contents of volcanic rocks and have been shown to be quite useful (Cox et al. 1979). I have used the TAS classification throughout this book.

As far as mafic volcanic rocks are concerned, the most useful classification is a genetic classification that

1 = Quartz Syenite
2 = Quartz Monzonite
3 = Quartz Monzodiorite
4 = Quartz Diorite
5 = Syenite
6 = Monzonite
7 = Monzodiorite

8* - Diorite if M < 35%
       Gabbro ((Pl+Cpx)
       Anorthosite (Pl>90%)
       Norite (Pl+Opx)

9 = Foid-bearing syenite
10 = Foid-bearing monzonite
11 = Foid-bearing monzogabbro
       or monzodiorite
12 = Foid-bearing gabbro/diorite
13 = Foid syenite
14 = Foid monzosyenite
15 = Foid
monzodiorite/monzogabbro
16 = Foid diorite/gabbro
17 = Foidolite (nephelinite,
leucitite etc.)

Q = Quartz, A = Alkali feldspar and albitic plagioclase,
P = Plagioclase
F = Feldspathoids (Foid), M = Mafic minerals

*This classification scheme is for plutonic rocks
containing <90% M.*
For rocks with >90%M, the classification for ultramafic
rocks should be used.

Fig. 2.24 QAPF classification of felsic plutonic rocks. The *upper triangle* (*gray*) is applicable to rocks with modal quartz (i.e., SiO_2 saturated). All silica-undersaturated rocks containing feldspathoids plot in the *lower triangle* (SiO_2 undersaturated)

was proposed by Yoder and Tilley (1962). This classification (popularly known as the "*basalt tetrahedron*"; discussed in a later chapter) is based on calculated mineral molecules (or *norm*) and has been known to be most useful in understanding the petrogenesis of basalt magmas.

Three ultramafic volcanic rocks, namely, kimberlites, lamproites, and komatiites, have not been discussed in terms of the above classification and are briefly discussed here. *Kimberlites* are a class of rare

potassic volcanic ultramafic rocks that contain phenocrysts of phlogopite, altered olivine, abundant xenoliths, and xenocrysts of the mantle, mixed with a very fine-grained material. Kimberlites have been a popular subject because many contain deep mantle xenoliths and are diamond bearing.

Lamproites are a rare group of volcanic to hypabyssal rocks that are unusually rich in K_2O and vary from ultramafic compositions through mafic to more evolved compositions. Leucite and sanidine are generally

Fig. 2.25 LeMaitre's et al. (1989) classification of volcanic rocks

present in the highly evolved lamproites. Olivine lamproites are ultramafic and contain phenocrysts of olivine (Fo$_{91-94}$), diopside, phlogopite, and potassic amphibole in a fine-grained groundmass dominantly composed of diopside, phlogopite, and phlogopite.

Komatiites have been discussed earlier and are ultramafic lavas with spinifex structure. The presence of komatiites in the Archean attests to very high temperatures within the Earth's shallow upper mantle.

CIPW Norm Calculation

Plutonic and hypabyssal rocks are medium to coarse grained, and so one can easily identify their constituent minerals and count their mode. On the other hand, volcanic rocks are generally too fine grained or glassy so that visual inspection does not allow a comparison of such rocks with their plutonic equivalent. In 1903, Cross, Iddings, Pirsson, and Washington realized that a way to put volcanic rocks on a more "equal footing" with plutonic rocks would be to calculate the proportions of a set of standard minerals (called *normative minerals*) from the chemical composition of the rock (detailed procedure given in appendix). This procedure, which may seem daunting, has almost become a standard practice over several decades, and there are websites that one can visit to perform the calculations. The mineral percentages so calculated from a chemical analysis are referred to as CIPW norm, as opposed to the *mode*, which refers to actually counted mineral percentages from a thin section of the rock specimen using various microscopic techniques. Also, it should be noted that CIPW method calculates normative minerals in terms of weight percent; however, if molecular percent of such normative minerals is calculated, then it is called a *Niggli norm*.

Commonly, a relatively smaller number of normative minerals are used, which generally are apatite (*ap*), ilmenite (*il*), magnetite (*mt*), feldspar (composition given in terms of orthoclase (*or*), albite (*ab*) and anorthite (*an*) components), diopside (*di*—composition is further broken down in terms of wollastonite (*wo*), enstatite (*en*), and ferrosilite (*fs*) components), hypersthene (*hy*—composition may be further broken down into *en* and *fs* components), olivine (*ol*—composition

A Few Examples That Illustrate the Basic Principle of CIPW Norm.
The underlying premise in developing the CIPW scheme of norm calculation is as follows: (1) A magma, if cooled slowly in a crustal environment, would form a rock containing proportions of crystals of certain chemical compositions; and (2) because minerals contribute to the chemical composition of a rock, one could "invert" its chemical composition to determine the proportion of mineral molecules (with ideal chemical formulae). This logic of norm calculation may be illustrated with a few examples as follows:

Example 1

Consider a "rock" that is composed only of CaO, MgO, Al_2O_3, and SiO_2 (table below). Its chemical composition is given in Column (2).

1	2	3	4	5	6	7	8	9	10	11	12
	Wt%	Mol. Wt.	Mole prop.	An	Di	Remaining			Moles	Mol. Wt.	CIPW
SiO_2	49.09	60.0843	0.817	0.395	0.153	0.270		An	0.197	278.208	54.81
Al_2O_3	18.12	101.96	0.178	0.197	0.000	0.000		Di	0.076	216.552	16.54
MgO	15.45	40.304	0.383	0.000	0.076	0.307		Hy	0.233	100.388	23.39
CaO	15.33	56.079	0.273	0.197	0.076	0.000		Ol	0.037	140.692	5.21
										Total	99.94

Column (3) simply gives the molecular weights of each oxide. Column (4) lists the mole proportions of each oxide in the "rock" (Column 4) obtained by dividing Column (2) by Column (3). At this stage we are in a position to distribute the appropriate moles of each oxide to form the possible mineral moles.

We first distribute all of Al_2O_3, an equal amount of CaO, and twice as much SiO_2 to make anorthite ($CaO \cdot Al_2O_3 \cdot 2SiO_2$) moles (Column 5). We then assign the remaining CaO, an equal amount of MgO, and twice as much SiO_2 to make diopside ($CaO \cdot MgO \cdot 2SiO_2$; Column 6). At this stage we check for whether the remaining SiO_2 (S) is less than, equal to, or greater than the remaining MgO (M). There are three possibilities to consider:

If $S = M$, then one can only make enstatite ($MgO \cdot SiO_2$).

If $S > M$, then one has to make enstatite and the excess SiO_2 will form quartz.

If $S < M$, i.e., silica is deficient, then the result has to be a mixture of enstatite and forsterite ($2MgO \cdot SiO_2$) because forsterite requires much less silica than enstatite.

According to Column 7, $S = 0.270$ and $M = 0.307$. That is, $S < M$. Now the question is how we divide M and S among forsterite and enstatite moles?

Let us suppose SiO_2 in enstatite $(SiO_2)^{en} = x$. Then $(MgO^{en}) = x$.

If we assume SiO_2 in forsterite $(SiO_2^{fo}) = y$, then $MgO^{fo} = 2y$.

$S = x + y = 0.270$ and $M = x + 2y = 0.307$. From these equations, we obtain x (i.e., enstatite moles) $= 0.233$ and y (forsterite moles) $= 0.037$. We convert all the moles to wt% norm by multiplying them with their respective molecular weights (see below). These are the same columns as columns (9) through (12) in the table above.

	Moles	Mol. wt.	CIPW NORM
an	0.197	278.208	54.81
di	0.076	216.552	16.54
hy	0.233	100.388	23.39
ol	0.037	140.692	5.21
			99.94

Example 2

Consider a second example (below) of slightly greater complexity. This is a rock composed only of normative olivine and hypersthene. The table below shows the wt% contents of FeO, MgO, and SiO_2 in the rock. As before, we first obtain mole proportions of each oxide.

	Wt%	÷	Mol. wt.	=	Mole prop.
SiO_2	43.62		60.3		0.723
FeO	11.40		71.85		0.159
MgO	44.82		40.3		1.112

A Few Examples That Illustrate the Basic Principle of CIPW Norm (continued)

After calculating the moles, we are at a loss as to how to distribute them between normative olivine [2(MgO, FeO)·SiO$_2$] and hypersthene [(MgO, FeO)·SiO$_2$], since both are composed of MgO, FeO, and SiO$_2$! We add FeO to MgO and treat it as *M*. Following our previous example,

$$\text{For hypersthene,} \quad (SiO_2)^{hy} = x = (MgO + FeO)^{hy}$$

$$\text{For olivine,} \quad (SiO_2)^{ol} = y \quad \text{and} \quad (MgO + FeO)^{ol} = 2y$$

Therefore, in the present example,

$$\text{hy moles} = x = 0.175$$

$$\text{ol moles} = y = 0.548$$

At this stage of calculations, we know that in hypersthene MgO + FeO = 0.175; however, we do not know MgO and FeO values individually, and we need to know those values if we are to obtain CIPW norms, which require multiplication by appropriate molecular weights. Here we make the assumption, as required by the CIPW calculation procedure, that the MgO/(MgO + FeO) ratios of both minerals are equal to that available:

$$[MgO/(MgO + FeO)]^{hy} = [MgO/(MgO + FeO)]^{ol} = [MgO/(MgO + FeO)]^{available}$$

We know from above calculations that

$$[MgO/(MgO + FeO)]^{available} = 1.112/1.271 = 0.875$$

Therefore,

$$[MgO/(MgO + FeO)]^{hy} = [MgO/(MgO + FeO)]^{ol} = 0.875$$

We already know (MgO + FeO)^{hy} = 0.175 and therefore easily calculate (MgO)^{hy} = 0.875 × 0.175 = 0.153 and (FeO)^{hy} = 0.175 − 0.153 = 0.022. These numbers can now be used to obtain CIPW norm of enstatite and ferrosilite components in normative hypersthene in the same manner as in our first example:

	Mole prop.	En	Fs
SiO$_2$	0.175	0.153	0.022
FeO	0.022		0.022
MgO	0.153	0.153	

CIPW normative en = 0.153 × 100.38 (mol. wt. of enstatite) = 15.36
CIPW normative fs = 0.022 × 132.15 (mol. wt. of ferrosilite) = 2.91

Similarly, olivine has 0.548 SiO$_2$ and total MgO + FeO = 1.096.

	Mol. #	Fo	Fa
SiO$_2$	0.548	0.4795	0.0685
FeO	0.137		0.137
MgO	0.959	0.959	

CIPW normative fo = 0.4795 × 140 (mol. wt. of fo) = 67.13
CIPW normative fa = 0.0685 × 204 (mol. wt. of fa) = 13.97

Adding up all the normative components, en + fs + fo + fa, we get a total of 99.37.

may be further broken down into forsterite (*fo*) and fayalite (*fa*) components), nepheline (*ne*), and quartz (*q*). There are websites available that perform norm calculations, and here I will not carry out the series of steps needed to perform a complete norm calculation. Instead, I will use a couple of examples to illustrate the principle of norm calculation.

The following important assumptions are made in norm calculation:

1. Hydrous phases are ignored and the magmas are assumed to be "dry."
2. Mafic minerals are assumed to be free of Al_2O_3. Alumina content is used to calculate the feldspars or feldspathoids.
3. Mg/Fe ratio of all Mg- and Fe-containing minerals is assumed to be the same.
4. The procedure recognizes the incompatibility of quartz and feldspathoids (e.g., nepheline, leucite) based on the well-founded observation that quartz and feldspathoids can never occur in the same rock under equilibrium conditions. Thus, quartz and nepheline never appear together in the norm of any rock.

None of these assumptions are particularly "way off base." However, like any simplifying procedures, the norm calculation procedure also has its limitations, for example, ignoring hydrous phases renders this method to be of limited application to andesitic rocks. Perhaps the greatest value of norm calculation is in the classification of basaltic rocks, as will be seen in a later chapter.

Utility of Norm Calculation: The Bottom Line

In this modern world of high-tech instruments and high-precision trace element and isotopic analyses, the student may wonder if there is any value to calculating norm. I believe that the real advantage of calculating norm is that it allows a way to compare volcanic rock compositions, specifically basaltic rocks, to phase diagrams. Such information is useful in deducing melting, crystallization, and mixing histories of individual magmatic suites, as we will see in a later chapter.

O'Hara (1965) was one of the pioneers who demonstrated the power of plotting normative compositions of basaltic rocks on phase diagrams in making petrogenetic inferences. Noting that the components CaO, MgO, Al_2O_3, and SiO_2 comprise more than 90 % of mafic and ultramafic rocks and contain all the major minerals that form upper mantle peridotites and those that crystallize in a basalt magma, he reasoned that converting normative compositions of natural basalts to these four components would allow a direct comparison of the phase relationships in this system with natural mafic and ultramafic rocks. This comparison can then lead to a more quantitative understanding of basalt magma generation and crystallization at various pressures. The utility of plotting normative compositions of basalts in various phase diagrams will become clearer as we delve into the issues of basalt magma formation and crystallization in later chapters.

Summary

1. Magma can be classified into ultramafic, mafic, intermediate, and felsic types based on their chemical composition. Mg and Fe decrease from ultramafic to felsic while Si and (Na + K) increase.
2. In terms of atomic structure, silicate magma is principally composed of polymerized SiO_4 chains and network modifier cations.
3. Density increases from felsic to ultramafic magma, whereas viscosity decreases. Dissolved volatiles dramatically decrease viscosity of magma. Increasing presence of crystals increases viscosity.
4. Undercooling of the magma controls crystal size, morphology, and sequence in which different phases appear.
5. Crystal size distribution or CSD is a powerful way to understand the magma's cooling history.
6. Volcanoes can be of many different types—ranging from large shield volcanoes to tuff cones and scoria cones. Mafic lavas are in general the least viscous and form the largest volcanoes with long flowing lava flows. Felsic volcanoes form smaller volcanoes.
7. Intrusions are of two fundamental types—plutons and hypabyssal intrusions. Batholiths, stocks, and lopoliths are various types of plutons, whereas dikes, sills, laccoliths, ring dikes, and cone sheets are all various types of hypabyssal intrusions.
8. Plutonic rocks are classified on the basis of mineralogy and texture. Volcanic rocks are classified on the basis of their major element chemistry.
9. CIPW norm classification is a useful way to convert chemistry into potential mineral proportions.

Exercises

1. In the following sketch, a magma chamber has formed a 100-m-thick boundary layer of troctolite, i.e., a rock composed of olivine and plagioclase crystals. There is no layering in the rocks that comprise the boundary layer.
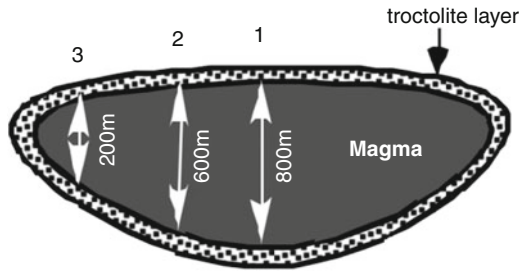
If a spherical block of 150 cm diameter breaks off from the roof at each of the sites 1, 2, and 3, then how long will it take for the blocks to reach the bottom at each site? Assume that the densities of troctolite and magma are 3.0 and 2.8 g/cm^3, respectively. Newtonian viscosity of the magma = 50 Pa.s. [*Answer: site 1: 2.72 h, site 2: 2 h, site 3: 0.7 h*]

2. An experimental study by I. Kushiro has given the following data on relationship between pressure (P), density, and viscosity of crystal-free basalt magma.

P (GPa)	Viscosity (Pa-s)	Density (g/cm^3)
0	9.8	2.62
1	4.7	2.79
2	2	3.02
3	1	3.1

Plot these data on two separate graphs and obtain mathematical relationships between (a) pressure and viscosity and (b) pressure and density. (c) Calculate the viscosity of crystal-free basalt magma at 4 GPa. [Answer to (c): 0.45 Pa-s]

3. Assuming that the basalt magma rises from 120 to 50 km through a dike of 1 m diameter (d), calculate its velocity (v in cm/s) at the following depths: 100, 80, 70, and 50 km using the following equation (discussed later in Chap. 5): velocity $(v) = [d^2 \cdot \Delta\rho \cdot g]/64\eta$ where $\Delta\rho$ = density of wall rock (density of magma), g = acceleration due to gravity, and η = magma viscosity.

[Assume 1 GPa = 30 km]. Calculate density and viscosity of magma using the equations below. Densities of wall rocks are given in the table below.

Viscosity (Pa-s) = $0.8356e^{(-0.77P)}$	
Density(g/cm^3) = $-0.035P^3 + 0.135P^2 + 0.07P + 2.62$	
Depth (km)	Density of wall rock (g/cm^3)
0–50	2.0
51–70	3.04
71–80	3.30
81–100	3.30

Plot the calculated velocities (cm/s) on X-axis and depth on Y-axis. At what depth will the magma stop rising?

4. Calculate how long (in days) it will take for a fully molten magma body to completely crystallize, assuming a constant nucleation rate of 10^{-3} cm^3/s and growth rate of 10^{-10} cm/s. Keep the calculation range within 1 and 120 days and 20 day intervals. Plot the curve that describes crystal% versus time (days). Within what range of days is the solidification rate the highest? (Refer to Box 2.3 to solve this problem.)