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B. Ronald Frost , Carol D. Frost
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Chapter 1

Introduction to Igneous Petrology

1.1 Introduction

Igneous petrology is the study of magma and the rocks that solidify from magma. Thus, igneous petrologists are concerned with the entire spectrum of processes that describe how magmas are produced, how they ascend through the mantle and crust, their mineralogical and geochemical evolution, and their eruption or emplacement to form igneous rocks. Igneous petrology requires a working knowledge of mineralogy. Readers who wish to review the characteristics of the major rock-forming igneous minerals will find a concise summary in the appendix, which emphasizes the identification of rock-forming minerals in hand sample and in thin section. In addition, the appendix includes descriptions of minerals found in minor abundance but commonly occurring in igneous rocks, including accessory minerals that contain trace amounts of uranium and are important geochronometers.

Before geologists can understand the origin of igneous rocks, they must classify and describe them. This chapter introduces the classification of igneous rocks using the mineralogical classification system recommended by the International Union of Geological Sciences (IUGS) Subcommission on the Systematics of Igneous Rocks, which has the advantage that it is relatively simple and can be applied in the field. For rocks that are too fine-grained to name using this classification, a geochemical classification can be employed instead. The simplest of these, the total alkali versus silica classification, is introduced in this text.

Finally, this chapter introduces basic terminology that describes the textural and structural features of igneous rocks. Descriptions of igneous textures document crystal shape, size, and the arrangement of the various minerals, glass, and cavities in the rock. Igneous structures are larger-scale features that are the result of rock-forming processes. The textures and structures preserved in igneous rocks provide information about their evolution, emplacement, and crystallization, all of which are fundamental goals of igneous petrology.

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1.2 The Scope of Igneous Petrology

All rocks are ultimately derived from magmas, which solidify to form igneous rocks. Consider, for example, the history of a shale. Such a rock is now composed of clay minerals. These clay minerals may have formed by weathering of a sedimentary rock that contained rock fragments and mineral grains. These components in turn may have been produced by erosion of a granitic gneiss. Before it was metamorphosed, this gneiss may have been a granodiorite, which is an igneous rock formed by crystallizing magma. As this example illustrates, the study of igneous petrology forms a foundation from which to study metamorphic and sedimentary rocks.

Igneous petrology is the study of the classification, occurrence, composition, origin, and evolution of rocks formed from magmas. The discipline can be divided into two components: **igneous petrography**[1], which is the description and classification of igneous rocks; and **igneous petrogenesis**, which is the study of the origin and evolution of igneous rocks. There are many different ways to approach the study of igneous petrology. *Field geology* is very important to the study of igneous petrology because important information is contained in the field relationships between rock units, the structure of an igneous rock, and its texture and physical appearance. For example, volcanologists depend heavily on their field observations during an eruption, and on the distribution of ash, lava, and other volcanic ejecta formed as the result of the eruption, to model the processes that occurred within a volcano before and during an eruption. *Laboratory identification* of the minerals in a thin section of an igneous rock, along with the chemical composition and age of a rock, are important means of classifying and relating it to other rocks with which it is spatially associated.

Another important way to study igneous rocks is through geochemistry. *Major-element geochemistry* can determine whether a suite of rocks is related through a process such as magmatic differentiation or mixing. *Trace-element geochemistry* is used to identify the role various minerals may have played as either crystallizing phases or residual phases in a suite of rocks. *Isotope geochemistry*, which can involve both radiogenic and stable isotopes, can determine if a suite of rocks formed from a single magma, or if a more complex, multi-source process was involved.

Because magmas that crystallize beneath Earth's surface are not observable and lavas erupted on the surface are hot and often dangerously explosive, geologists find it difficult to study the formation of igneous rocks directly. Therefore, *experimental petrology* is an important aspect of igneous petrology in which the pressures and temperatures required for igneous rocks to form and evolve are reproduced in the laboratory. For many rocks, a field and petrographic description does not provide conclusive proof of the process by which they formed. For these rocks, data gathered from experimental petrology are essential.

1.3 Classification of Igneous Rocks

One of the most tedious aspects of igneous petrography is the mastery of terminology. Innumerable, and often inscrutable, names have been applied to igneous rocks over the past few centuries as petrology grew in importance and sophistication. Much igneous terminology is arcane because in the early days of the science, petrologists did not have access to experimental data, phase diagrams, isotopic systems, or thermodynamic data and thus their work was mainly descriptive as opposed to quantitative. One way they described rocks was to name them. Among the more picturesque names is **charnockite**, which was named after the rock that formed the tombstone of Job Charnock, the founder of Calcutta (now Kolkata), India. Charnockite is a name given to an orthopyroxene-bearing granite, but there is no way to determine that from the origin of the name unless one was to desecrate Job Charnock's tombstone by sampling it for thin section and chemical analysis.

Unfortunately, like charnockite, most of the rock names that arose early in the development of igneous petrology do not provide much insight into the origin or evolution of the rock they describe. Many of the rock names based on type locality were given in the nineteenth or early twentieth century. Over time, geologists recognized the necessity of a more systematic rock classification scheme. In 1972, the IUGS Subcommission on the Systematics of Igneous Rocks published a rock classification system that has been widely adopted and use of many of the old rock names has been abandoned (Streckeisen, 1976; LeBas and Streckeisen, 1991; Le Maitre et al., 2005).

There are two basic approaches to the naming of rocks. A rock can be classified either according to the minerals it contains or by its chemical composition. The first

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approach has the benefit that geologists can name rocks in the field by identifying their mineralogy; however, it is not very helpful for classifying fine-grained rocks. Alternately, a chemical classification requires analytical data, and therefore is not useful in the field, but it does provide a means of naming fine-grained or glassy rocks. The compositions of most igneous rocks can be expressed in nine oxides: SiO_2, TiO_2, Al_2O_3, Fe_2O_3, FeO, MgO, CaO, Na_2O, and K_2O. These combine to form the major rock-forming igneous minerals, which include pyroxene, olivine, garnet, amphibole, mica, quartz, plagioclase, alkali feldspar, feldspathoid, magnetite, and ilmenite. Most rocks contain only a few of these minerals. The IUGS classification uses both mineralogical and chemical data but emphasizes classification on the basis of mineralogy.

1.3.1 Preliminary Classification

Igneous rocks are divided into the general categories of **plutonic**, **hypabyssal**, and **volcanic**, depending on their grain size. Plutonic rocks characteristically have coarse or medium grain sizes (>1 mm) and are inferred to have crystallized deep in the crust. Hypabyssal and volcanic rocks are fine-grained to glassy. Volcanic rocks crystallize at the surface and hypabyssal rocks crystallize at shallow depths, typically less than a kilometer. Because the grain size of an igneous rock is determined in part by the cooling rate of the magma, and this is a function both of magma temperature and the ambient temperature of the rocks into which the magma was emplaced, grain size generally increases with depth but there is no specific depth associated with the transition from plutonic to hypabyssal rocks.

In addition to classification according to grain size, the general composition of a rock can be described using the terms **felsic**, **mafic**, and **ultramafic**. Rocks that are rich in quartz, feldspars, or feldspathoids are light-colored and are called *felsic*. The term *felsic* combines parts of the words *fel*dspars (and *fel*dspathoids) and *si*lica. Darker-colored rocks, rich in ferromagnesian minerals, are called *mafic*. The term *mafic* reflects the enrichment of these rocks in *ma*gnesium and iron (*Fe*). Ultramafic rocks are essentially free of any felsic minerals.

1.3.2 IUGS Classification of Plutonic Rocks

Because plutonic rocks are relatively coarse-grained, so that their constituent minerals can be easily identified, either in hand specimen or in thin section, they are the most straightforward group of igneous rocks to classify. The IUGS classification is based on the abundance of the common minerals in volume percent (modal mineralogy, or **mode**), which are divided into five groups:

Q	quartz
A	alkali feldspar, including albite with up to 5 mole percent anorthite ($<An_5$)
P	plagioclase, with composition An_5 to An_{100}
F	feldspathoids (foid): nepheline, sodalite, analcite, leucite, cancrinite
M	mafic minerals: olivine, pyroxenes, amphiboles, micas, and opaque minerals, and accessory minerals such as zircon, apatite, sphene, allanite, garnet, and carbonate

Rocks containing less than 90 percent mafic minerals ($M<90$) are classified according to the amounts of Q, A, P, and F minerals they contain, whereas rocks containing more than 90 percent mafic minerals are classified according to the proportions of major mafic minerals. Felsic and mafic rocks typically have far less than 90 percent mafic minerals and ultramafic rocks far more.

Because rocks never contain both quartz and feldspathoids, felsic and mafic rocks can be classified in terms of three components, either QAP or FAP. Triangular plots of the three components are shown in Figure 1.1 along with the names assigned to rocks containing particular proportions of Q, A, P, and F minerals. However, some rocks are not uniquely defined by QAP or FAP alone. For example, both diorite and gabbro fall in the same portion of the QAP triangle. They are distinguished primarily on the basis of plagioclase composition: plagioclase in diorite is more sodic than An_{50}, whereas that in gabbro is more calcic. Because the IUGS classification does not consider the composition of the plagioclase, it cannot distinguish these two rock types. A third rock name is assigned to the gabbro/diorite portion of the QAP triangle: anorthosite. Anorthosite is a special name applied to rocks that contain more than 90 percent plagioclase. Because the IUGS classification is based only on the proportion of Q, A, P, and F minerals, it does not distinguish between rocks with only 10 percent ferromagnesian minerals and rocks with up to 90 percent ferromagnesian minerals. Therefore, anorthosite occupies the same part of the triangle as do the diorites and gabbros that have considerably higher mafic mineral contents. This classification scheme can be further specified by adding the names of the major mafic minerals present, with the most abundant placed closest to the rock name.

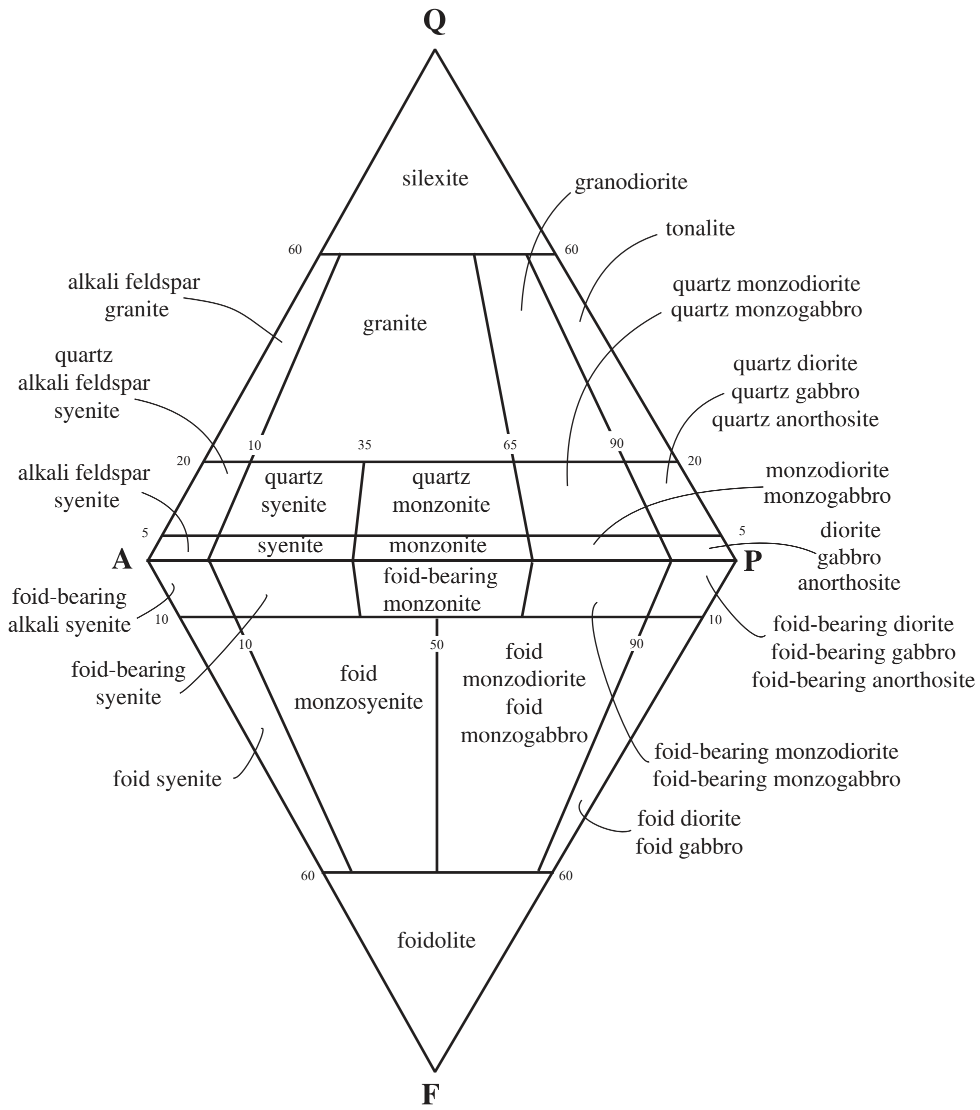

Figure 1.1 IUGS classification of plutonic rocks, based upon modal mineralogy. A = alkali feldspar, F = feldspathoid, P = plagioclase, Q = quartz. After Le Maitre et al. (2005).

For example, a biotite–hornblende tonalite contains more hornblende than biotite.

Mafic rocks can be further subdivided according to the proportion of plagioclase, orthopyroxene, clinopyroxene, olivine, and hornblende they contain (Figure 1.2). Strictly speaking, the term *gabbro* applies to a rock consisting of augite and calcic plagioclase, although the term is also broadly applied to any rock consisting of calcic plagioclase and other ferromagnesian minerals. For example, troctolite, a rock with olivine + calcic plagioclase, and norite, a rock with orthopyroxene + calcic plagioclase, are included in the gabbro family. Though not shown in Figure 1.2, rocks consisting of calcic plagioclase and hornblende are, quite logically, called hornblende–gabbros. Most gabbroic rocks contain between 35 and 65 percent mafic minerals. If they contain less than this, the rock name may be prefixed by *leuco-*, meaning light. If they contain more than 65 percent mafic minerals, they may be prefixed by *mela-*, meaning dark.

Ultramafic rocks contain little or no plagioclase and thus require their own classification scheme, based on ferromagnesian mineral content. Ultramafic rocks containing more than 40 percent olivine are called *peridotites*, whereas ultramafic rocks containing more than 60 percent pyroxene are called *pyroxenites* (Figure 1.3). Peridotites and pyroxenites are further divided depending on the relative proportions of orthopyroxene, clinopyroxene, and olivine. The presence of other mineral phases can be used to further specify the name of the ultramafic rock; for instance, lherzolite that contains garnet is called garnet lherzolite.

Charnockites (orthopyroxene-bearing granitic rocks), lamprophyres (mafic and ultramafic rocks with mafic

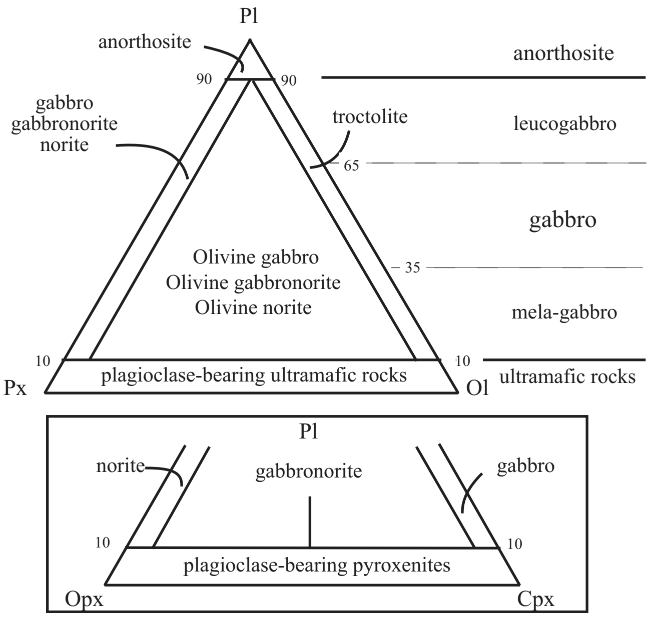

Figure 1.2 IUGS classification of gabbroic rocks. Ol = olivine, Pl = plagioclase, Px = pyroxene. Inset shows classification with regards to the type of pyroxene. Opx = orthopyroxene, Cpx = clinopyroxene. After Le Maitre et al. (2005).

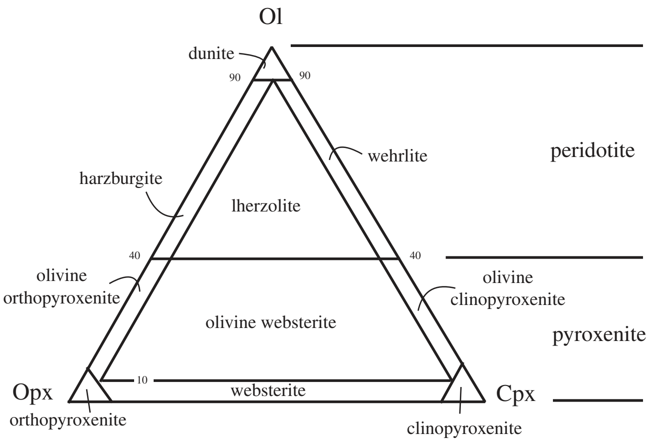

Figure 1.3 IUGS classification of ultramafic rocks. Ol = olivine, Opx = orthopyroxene, Cpx = clinopyroxene. After Le Maitre et al. (2005).

phenocrysts), carbonatites (igneous carbonate-rich rocks), and pyroclastic rocks have their own classification schemes (Le Maitre et al., 2005).

1.3.3 IUGS Classification of Volcanic and Hypabyssal Rocks

Whenever possible, the IUGS recommends that volcanic rocks be classified on the basis of modal mineralogy. The names for volcanic and hypabyssal rocks determined in this way are given in Figure 1.4. There are a few plutonic rock types for which there are no volcanic equivalents, such as anorthosite and ultramafic rocks. These plutonic rocks usually represent accumulations of crystals, and no liquid of that composition ever existed. The only ultramafic lava solidifies to form a rare rock called **komatiite**, which occurs almost exclusively in ancient Archean terrains. It is the volcanic equivalent of peridotite.

If the volcanic rocks are so fine-grained that minerals cannot be identified, then they must be classified on the basis of chemical composition. The IUGS has recommended that volcanic rocks be classified based upon their total alkali and silica contents (TAS) (LeBas et al., 1986) (Figure 1.5). The TAS diagram has as its *x*-axis the weight percent of SiO_2 of the rock, and as its *y*-axis the weight percent $Na_2O + K_2O$ of the rock. The diagram is then divided into 15 fields. Classification using this chemical approach gives rock names that are typically consistent with the names based on the QAPF diagram.

1.4 Igneous Textures

Petrologists use textures and structures to interpret how igneous rocks crystallized. The terms texture and structure are nearly interchangeable, although *texture* of a rock refers to the small-scale appearance of the rock: the size, shape, and arrangement of its constituent phases, including minerals, glass, and cavities. The *structure* of a rock refers to larger-scale features, recognizable in the field, such as banding, variations in mineral abundances, or jointing. Textures may provide information about cooling and crystallization rates and the phase relations between minerals and magma at the time of crystallization. Structures indicate the processes active during the formation of rocks and the mechanisms of differentiation.

1.4.1 The Crystallization of Igneous Melts

The formation of igneous rocks by crystallization of melts can be understood from basic thermodynamic principles. Thermodynamics is a field of chemistry that is designed to determine what assemblage of solids, liquid, and gas would be stable from a set of chemical compounds at a given temperature and pressure. In igneous petrology, thermodynamics predicts the minerals that will crystallize from a melt. A key observation of thermodynamics is that a chemical reaction such as crystallization can occur

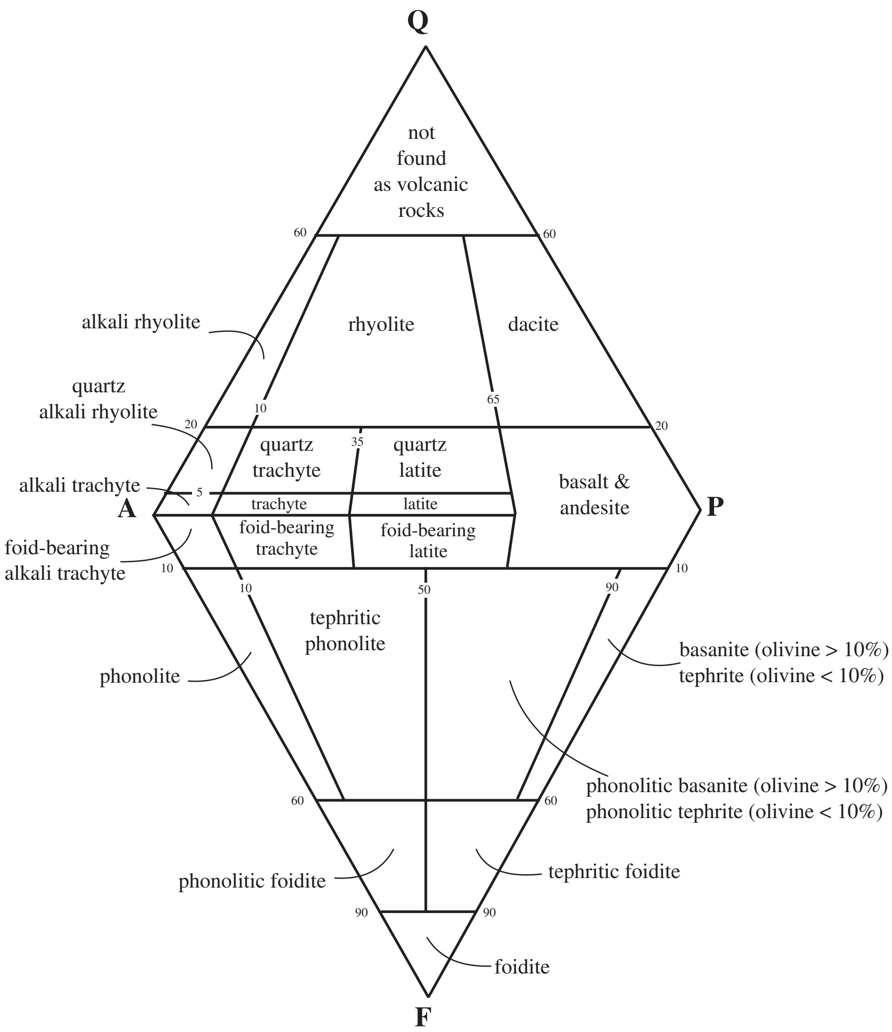

Figure 1.4 IUGS classification of volcanic rock based on modal mineralogy. Abbreviations as in Figure 1.1. After Le Maitre et al. (2005).

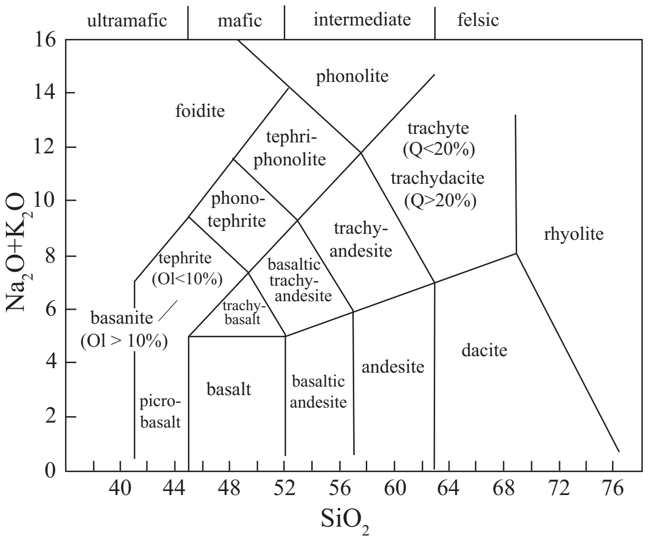

Figure 1.5 IUGS classification of volcanic rocks based on chemical composition, in weight percent oxide. Q = quartz, Ol = olivine. After Le Maitre et al. (2005).

spontaneously only if it evolves heat. Heat is tied up in minerals (and other substances) in two ways, **enthalpy** and **entropy**.

Enthalpy is the heat that is tied up in the chemical bonds of a substance. Because the bond strengths in each substance are different, a reaction from substance A to substance B will release (or consume) heat. *Entropy* is a measure of "randomness." It is an indication of how tightly bonded atoms are in the structure. The entropy of an element in a gas is much higher than the entropy of the same element in a melt, which in turn is higher than the entropy of that element in a solid. Similarly, the entropy of an element that is bonded by weak van der Waals' bonds is higher than that of the same element bonded by covalent bonds. The amount of energy released by a chemical reaction that can be used for chemical work is called **free energy**. It is symbolized by *G*, after J. Willard Gibbs, the person who formulated it, in

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1873. The equation for the change in free energy between the products and reactants in a chemical reaction (symbolized by the Greek letter Δ) is given as:

$$\Delta G = \Delta H - T\Delta S \tag{1.1}$$

This equation says that the amount of heat released by a reaction equals the amount of heat released by the change in bonding strength between the products and reactants minus the amount of heat consumed as a result of the change in configuration of the atoms. For any reaction there are three possibilities for the value of ΔG. It can be positive, in which heat must be added to make the reaction proceed. This means that the reactants in the reaction are stable. ΔG can be zero, in which the reactants and products are in **equilibrium**. The other possibility is that ΔG is negative, which means that the reaction may occur spontaneously. A spontaneous reaction may not take place even if $\Delta G < 0$ because there may be an energy barrier to overcome. In this case, the system is said to be **metastable**.

1.4.2 Crystal Size

For crystallization to begin during cooling of a silicate melt, the ΔG between the melt and the minerals in the solid state must be less than 0. However, crystallization of an igneous melt involves more than favorable thermodynamic conditions. It also involves **nucleation** of new crystals and growth of the crystals that have nucleated. Both nucleation and growth are dependent on **diffusion**, the rate at which ions of various elements can move through the melt. In general, the diffusion rate is faster for ions that are small or have low valence and slower for those that are larger and have high valence (Zhang, 2010). For example, Ca^{2+}, Mg^{2+}, or Fe^{2+}, which having the same charge and similar size, diffuse roughly at the same rate. The larger, singly charged Na^{+} ion will diffuse more quickly and Al^{3+} will diffuse more slowly. The diffusion rate of ions in a melt is also a function of the **viscosity** (resistance to flow) of the melt. Diffusion is faster in less viscous melts like basalt, and slower in viscous melts such as rhyolite (Zhang et al., 2010).

Consider a simple example such as crystallization of ice from water. When the temperature drops below 0 °C, ice crystals begin to form. Ice is more stable than water below 0 °C because ΔG for the reaction water $\rightleftharpoons$ ice is less than zero. Water doesn't freeze when the temperature is exactly 0 °C even though at this temperature water and ice are in equilibrium. At temperatures slightly below 0 °C, H_2O molecules in the water may bond together to make a tiny crystallite of ice. The first few H_2O molecules to bond together are metastable. Although energy is released when the molecules are bonded together, the ions that lie on the margins of the ice crystallite are not bonded to the ions of other ice molecules. Instead, they are bent back into the crystallite, producing energy that is called **surface energy**. The amount of energy released when a few molecules of ice combine together is the difference between the amount of free energy released because of the bonding of the molecules and the amount of surface energy from the unfilled bonds on the margins of the ice cluster. This difference will be large when the crystallite is small because the surface energy is large relative to the volume of the crystallite. The difference decreases with increases in crystallite size because the free energy released during the formation of a crystallite is a function of the volume of the crystal, whereas the surface energy is a function of the area. As the radius (r) of the crystallite increases, the volume increases by a factor of r^3 whereas the surface area increases by a factor of r^2. Thus, as the radius increases, the free energy of the crystallite increases at a much faster rate than the surface energy, which means that there is a certain radius at which the crystallite becomes stable. When this happens, the crystallite becomes a **nucleus**.

Because a crystallite must become large enough to overcome the surface energy barrier before it can become a nucleus, it is difficult to nucleate a crystal from a melt when the temperature is close to the temperature of crystallization. However, nucleation becomes energetically favored if the melt is cooled below the crystallization temperature, a condition called undercooling. With low amounts of undercooling the ΔG between the crystals and melt is small. ΔG increases with increased undercooling, which means that the nucleation rate of crystals in a melt will increase as a melt is undercooled (Figure 1.6A). With a certain amount of undercooling, nucleation rate reaches a maximum. Nucleation rate drops with further undercooling because diffusion becomes increasingly difficult as the melt cools and becomes more viscous. At some temperature where the melt is nearly solidified, diffusion becomes very slow, and the molecules necessary for growth of a new crystal cannot move to a nucleation site.

Once a crystal has nucleated, its growth from a melt involves two steps: the diffusion of ions through the melt, and the adherence of these ions onto the growing face of the crystal. The diffusion rate through the melt will be

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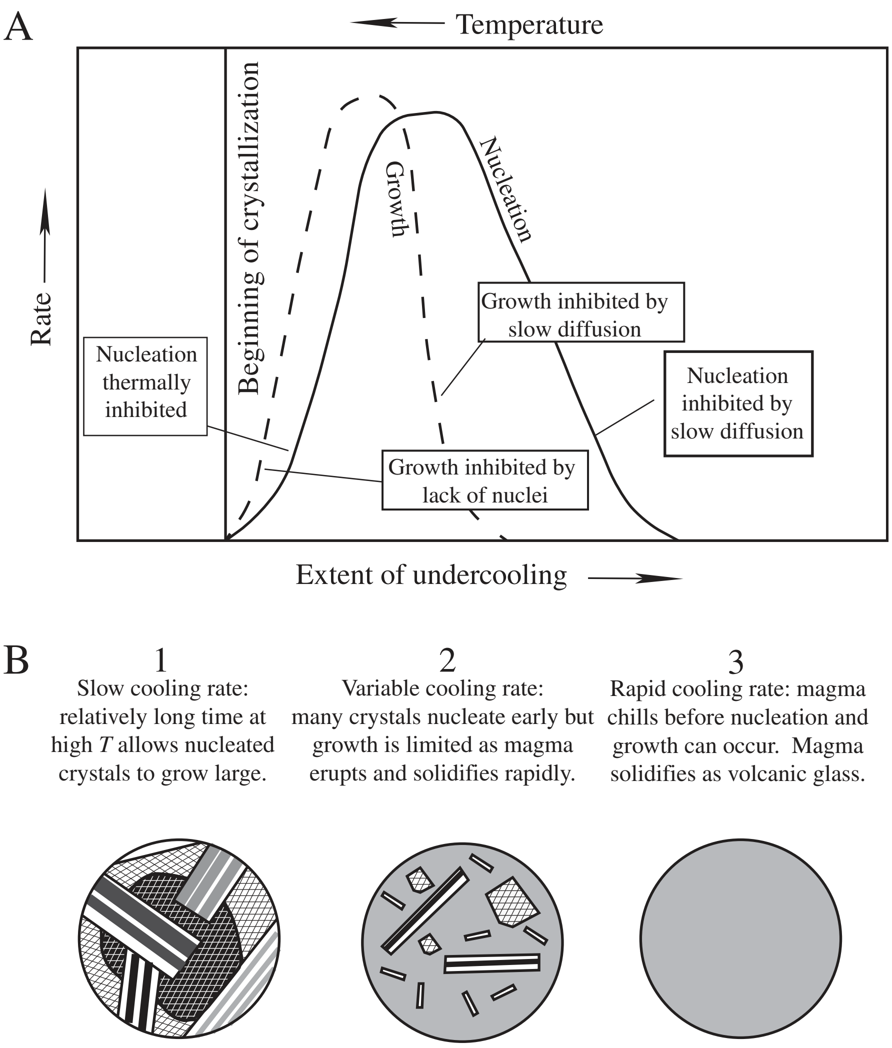

Figure 1.6 (A) Schematic representation of the interplay between growth and nucleation during cooling of melt. (B) Inset 1 shows textures formed in a plutonic rock in which the melt cooled slowly through temperatures at which the first crystals form. In these conditions growth is faster than nucleation, resulting in a rock with relatively few, large crystals. Inset 2 shows textures from a volcanic rock where the magma erupted after initial crystal nucleation and growth, and cooling took place rapidly, limiting further crystal growth. Inset 3 shows textures in a volcanic rock that cooled rapidly and solidified before nucleation and crystal growth could occur.

fastest at temperatures near the beginning of crystallization of the melt and will decrease with cooling as the melt becomes increasingly viscous. With low undercooling the adherence of ions, commonly called *nutrients*, to the face of the crystals will be slow, because only a few nuclei are present and hence only a limited crystal surface area is available on which the nutrients may adhere. The growth rate will increase with falling temperature as the crystals grow and an increasing area is available for the nutrients to populate. The maximum nucleation rate occurs at lower temperatures than the maximum growth rate because during nucleation the nutrients must migrate only short distances. In contrast, as the crystal grows and nearby nutrients are depleted, for the crystal to grow further nutrients must migrate longer distances. The total growth rate, therefore, will be zero at the temperature of melting, will rise with some degrees of undercooling, and will fall again as diffusion becomes increasingly difficult in the progressively more viscous, solidifying melt (Figure 1.6A).

The textures observed in igneous rocks reflect the interplay between nucleation, crystal growth, and the rate of cooling. If the magma cools slowly and crystallizes in an environment where the rate of growth was faster than nucleation, then it would solidify as a rock with large crystals, as for example 1 in Figure 1.6B. Such a rock would be called **phaneritic.** If the magma cools so that during crystallization it resides at conditions where nucleation was faster than growth, then it would consist of many small crystals. If the magma then erupted and solidified, the matrix would be composed of glass (example 2 in Figure 1.6B). Such a texture would be called **aphanitic.** If the melt chilled quickly before nuclei could form it would solidify to a **glassy** texture where there are no crystals at all (example 3 in Figure 1.6B).

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Another variable that affects grain size is the presence of volatile components or elements, such as H_2O or F, that decrease the viscosity of the melt and, hence, enhance the diffusion of essential elements to the face of a growing crystal. Melts with an abundance of these elements may crystallize extremely coarse-grained crystals, forming an intrusive rock called **pegmatite**. Pegmatites may have grain sizes up to a meter or more.

1.4.3 Crystal Shape

Petrologists use the shape of crystals and how the various minerals are arranged in an igneous rock to decipher the crystallization history of a rock. A mineral growing in a melt will tend to have grain boundaries that are **euhedral**, that is, they are bounded by well-formed crystal faces. The thin section of nepheline basalt shown in Figure 1.7A is composed of euhedral crystals of augite and olivine, contained in a fine-grained matrix. The textures shown in the thin section suggest that the augite and olivine began to crystallize from the melt and had grown to sizes of 1–5 mm before the lava erupted. The fine-grained matrix indicates that the melt in which the crystals were entrained chilled quickly and solidified as volcanic glass. A close examination of Figure 1.7A shows that the matrix

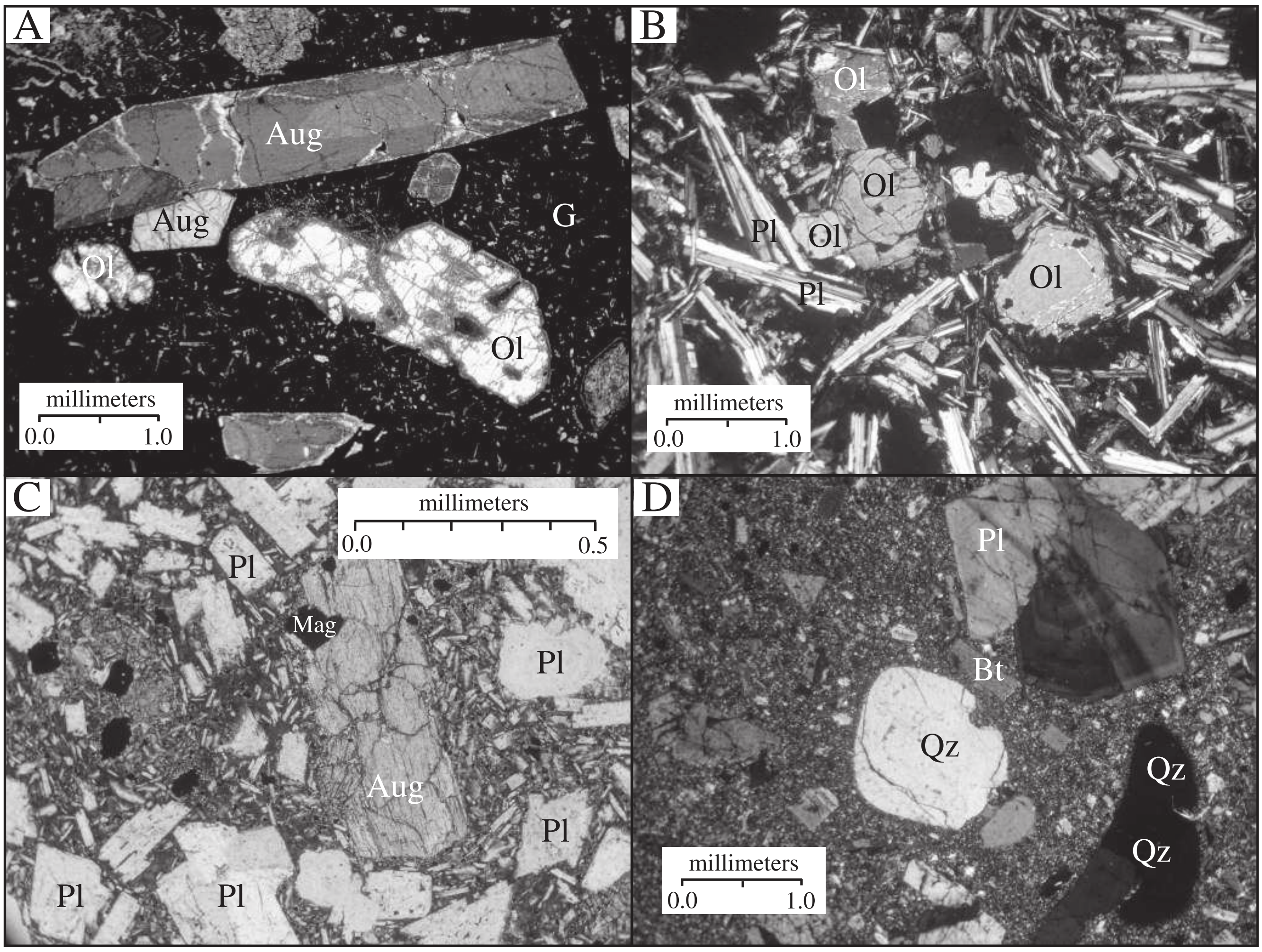

Figure 1.7 Photomicrographs showing textures in volcanic rocks. (A) Glassy nepheline basalt containing phenocrysts of olivine (Ol), augite (Aug), and glass (G), erupted near the Kaiserstuhl, southern Germany. Crossed polarized light (XPL). (B) Olivine tholeiite containing phenocrysts of olivine (Ol) and plagioclase (Pl) in a matrix of fine-grained olivine, augite, plagioclase, and glass, from the Snake River Plain, Idaho, USA. XPL. (C) Andesite with phenocrysts of augite (Aug) plagioclase (Pl) and magnetite (Mag) in matrix of fine-grained plagioclase, augite, and glass, from Soufriere volcano, St. Vincent. Plane-polarized light (PPL). (D) Dacite consisting of quartz (Qz), plagioclase (Pl), and biotite (Bt) in a matrix of quartz, plagioclase, and glass. XPL.

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is not all glass; a few extremely small grains of augite and plagioclase are also present. These probably nucleated shortly after the basalt erupted but before it completely solidified.

A crystal that is relatively large compared to the minerals composing the matrix of an igneous rock is called a **phenocryst**. In Figure 1.7A, the contrast in size between the phenocrysts and the matrix is obvious. However, few igneous rocks have a matrix so dominated by glass. More typically, the matrix undergoes some degree of crystallization. For example, the basalt shown in Figure 1.7B contains phenocrysts of equant olivine and elongate plagioclase in a matrix of finer-grained olivine, augite, plagioclase, and glass. Relations are similar in the andesite shown in Figure 1.7C, except the plagioclase in the andesite is stubbier than the plagioclase in the basalt. Phenocrysts of quartz may occur in highly siliceous melts, such as dacite (Figure 1.7D) and rhyolite, and the presence of quartz phenocrysts is one way to identify these rocks in the field.

Many of the textures characteristic of volcanic rocks also help petrologists interpret plutonic rocks. The early crystallizing minerals form a matrix of interlocking euhedral grains, in a texture called **cumulate** texture. The minerals that formed later are constrained to grow in the interstices of these cumulus grains. These are called **postcumulate** grains and they are **anhedral**, which means they are not bounded by crystal faces. Examples of cumulate texture are shown in Figure 1.8A, a gabbro consisting of cumulus plagioclase and magnetite and postcumulus augite, and in Figure 1.8B, a pyroxenite with cumulus orthopyroxene and postcumulus plagioclase. Some granitic rocks contain tabular plagioclase or potassium–feldspar; for example, the granodiorite shown in Figure 1.8C contains distinctly tabular plagioclase. The plagioclase has the same stubby aspect ratio as plagioclase of similar composition in the volcanic rock shown in Figure 1.7C. The concentric zoning in this plagioclase records changes in composition as the plagioclase grain grew in the granodioritic melt.

In some plutonic rocks, the magma solidifies after relatively coarse-grained minerals have formed, making a rock called a **porphyry**. This rock has a texture that is characterized by euhedral grains dispersed in a finer-grained matrix (Figure 1.8D). A porphyritic texture tells a geologist that the rock underwent a complex cooling history. First, it cooled slowly, during which time the phenocrysts grew, followed by sudden cooling that caused the rapid solidification of the rest of the melt.

1.5 Igneous Structures

Igneous rocks exhibit a wide variety of forms. Mafic volcanic rocks occur mostly as flows; felsic volcanic rocks may also form flows, but also commonly form **pyroclastic** rocks, or rocks fragmented while still hot. Hypabyssal rocks may form as **lava domes**, **dikes**, or **sills**, and plutonic rocks occur as **plutons** and **batholiths**, as well as dikes and sills.

1.5.1 Structures in Volcanic Flows

Lava flows may range in thickness from less than a meter to more than 10 meters. Mafic lava flows are often divided into two types: blocky lava is known as **aa** (Figure 1.9A), and massive lava with a ropey surface is called **pahoehoe** (Figure 1.9B). Pahoehoe texture forms on relatively hot lavas but as the lava cools, the surface breaks apart, making aa. These names are etymologically Hawaiian; abundant lava flows in Hawaii allowed native Hawaiians ample opportunity to develop a terminology, comparing the textures of the flows. In cross-section, many flows, particularly those that ponded before completely crystallizing, show **columnar jointing** (Figure 1.9C). Columnar jointing forms by contraction that cracks the rock as heat dissipates from upper and lower margins of the flow. The vertically oriented columns, which are typically hexagonal in cross-section, commonly form a relatively wide set of columns at the base of the flow and a colonnade of narrower columns at the top.

Where basalts erupt under or flow into water, they form **pillows** (Figure 1.9D). The magma that contacts water is chilled and quenches, forming a distinctive lobate, or “pillow,” shape. As lava continues to flow, it breaks the solidified crust of the initial pillow to form another lobe. A pillow basalt is constructed of hundreds of these nested lobes. In cross-section, the pillows have a rounded top and a tail that points downward. Pillow basalts are diagnostic of subaqueous volcanism and because they are well preserved in the geologic record, they allow geologists to identify underwater eruptions that may be up to billions of years old.

Commonly, gas bubbles exsolved from the magma gather at the top of a flow. Solidification of the melt will produce a rock pocked by holes from these exsolved gas bubbles. The holes are called **vesicles**, and they are key evidence of lava flows because gas bubbles are unlikely in plutonic rocks. Vesicles are also important markers of the