Chapter 1

Some *Fundamental Concepts*

In this initial chapter, I will generalize and set the stage for the more detailed chapters to come. After a brief introduction to igneous rocks, I will attempt to organize our approach to the subject of igneous petrology. Next, because magmas are created by the melting of material at depth, we will look in the broadest sense at the makeup and internal structure of the Earth, followed by a summary of current theories on how it got that way. Finally, we will review in a very general sense the mode of occurrence of the common igneous rocks. It will be assumed that you, the student, are familiar with the most basic concepts, such as the distinction between volcanic, plutonic, and pyroclastic igneous rocks, etc. You may want to briefly review the chapter on igneous rocks and processes in the text from your introductory geology course, as it will refresh your memory and provide an initial "big picture" as we proceed to refine the ideas. Reviewing a chapter on plate tectonics would also help in this regard.

1.1 INTRODUCTION

Igneous petrology is the study of melts (magma) and the rocks that crystallize from them. Origin by crystallization from a melt seems a simple enough criterion for considering a rock to be igneous. But we can only rarely observe the formation of igneous rocks directly, and then only for some surface lavas. We must therefore develop a separate set of observational criteria for determining that a rock is indeed of igneous origin. Such criteria will be developed further later on, but, by way of introduction, they include:

1) Field criteria. Intrusive igneous bodies commonly crosscut the "country rocks" into which they intrude, thereby truncating external structures, such as bedding or foliation. They may also exhibit some types of contact effects. When developed, a narrow, fine-grained **chilled margin** (or "chill zone") within the igneous body margin, or localized baking of the country rocks, are good indicators of an igneous origin for plutonic bodies. In addition, we have come to associate certain specific forms of rock bodies with an igneous origin. For example, a strato-volcano, a pahoehoe flow, a sill or laccolith, etc., have become associated with igneous processes, either by directly observing an igneous event, or by applying some of the criteria above. Field aspects of igneous rocks will be discussed further in Chapter 4.

2) Textural criteria. Petrography is the branch of petrology that deals with microscopic examination of thin sections, cut from rock samples and ground down to 0.03-mm thickness so they readily transmit light. By observing igneous rocks under the petrographic microscope, we have come to associate a specific **interlocking** texture with slow crystallization from a melt (Figure 1-1). When crystals are forming in a cooling melt, they usually develop a near-perfect crystal form, as the melt provides no obstruction to the preferred crystal shape. As the melt continues to cool, and more crystals form, they eventually begin to interfere with one another, and intergrow. The resulting interlocking texture shows interpenetrating crystals, much like a jigsaw puzzle. As will be discussed in Chapter 3, the relative development of crystal form, inclusions, and interpenetration can commonly be used to infer the *sequence* in which different mineral species crystallized.

Rapid cooling and solidification of a melt may result in another characteristic texture: **glassy** texture. When a melt solidifies too fast for ordered crystal structures to form, the result is a non-crystalline solid, or glass. Glass is readily recognized under the petrographic microscope by its isotropic optical character.

Because liquids cannot sustain substantial directional stresses, **foliations** rarely develop. A common textural criterion for distinguishing an igneous from a high-grade metamorphic crystalline rock in hand specimen is thus based on the isotropic texture (random orientation of



Figure 1-1 Interlocking texture in a granodiorite. From Bard (1986). Copyright © by permission Kluwer Academic Publishers.

elongated crystal) of the former. One must use caution, however, when applying this criterion, as some igneous processes, such as crystal settling and magmatic flow, can produce mineral alignments and foliations in igneous rocks.

Pyroclastic deposits (those resulting from explosive eruptions) can perhaps be the most difficult to recognize as igneous. Usually the magmatic portion has solidified and cooled considerably before being deposited along with a significant portion of pulverized pre-existing rocks caught in the explosion. The actual deposition of pyro-clastic material is in large part a sedimentary process, and hence the difficulty in recognition. In fact there is some debate among geologists as to whether pyroclas-tics should be considered igneous or sedimentary. They are igneous in the sense that nearly all of the material crystallized from a melt, even though a proportion may have been earlier volcanic deposits. This is the "pyro" part. They are sedimentary as well, in the sense that they represent solid particles deposited by a fluid medium: air, or sometimes water. This is the "clastic" part. Some geologists have wisely suggested that we avoid this hopeless debate by considering pyroclastics to be igneous going up, and sedimentary coming down!

In Chapter 3, we will discuss igneous textures in more detail, including both those seen in hand specimen, and those seen in thin section with the aid of the petrographic microscope.

As we initially consider the study of igneous rocks and processes, perhaps we should first consider what exactly it is that we want to know. The types of very "broad-brush" questions that we would expect to have answered might include the following: How are melts generated? What is melted, and where? What is produced by this melting? How do the melts so produced crystallize to igneous rocks, and what processes accompany this crystallization? In what way(s) do the liquid and solid portions evolve during the process of melting or crystallization? Does the large variety of igneous rock compositions now found at the Earth's surface result from different sources of melts, or can it be attributed to variations in the processes of melting and crystallization? Is there a relationship between igneous rock type and tectonic setting? If so, what controls this?

Finally we might ask, "What do we need to know to assess these?" In other words, what background and approach does a good modern petrologist need? I would suggest the following as imperative background:

1. A petrologist needs experience looking at rocks and textures. One cannot begin to study rocks without knowing how to recognize, describe, or ganize, and analyze them.
2. Experimental data (of synthetic and natural sam ples) are also necessary. We can best understand the generation and crystallization of melts by recreating these processes in the laboratory, sim ulating the conditions found at depth, and ana lyzing the results.
3. Some theory is also required, so we can organize and understand the experimental results better and apply those results beyond the exact compo sitions and conditions of the experiments. A bit of chemistry, including major elements, trace ele ments, and isotopes necessary, as is some ther modynamics. As we shall see, these techniques also help us characterize rocks and evaluate source regions and evolutionary processes.
4. We also need a knowledge of what comprises the Earth's interior and the physical conditions that exist there. Melts are created deeper than we can directly observe. If we want to know what is melted and how, we must review what is known about the Earth's interior, and how the constituents and conditions vary with tectonic setting.
5. Finally, we need some practical experience with igneous activity. A literature-based survey of the more common igneous rocks and processes in na ture provide a framework for all of the above, and give a more complete picture.

I hope that we will acquire the requisite skills and experience in the chapters that follow. Although it would

perhaps be preferable to develop each of the above skills sequentially, this is not practical, and may be impossible, as they are integrated in the scientific process. The student will get direct observational experience in the class laboratories, with the aid of Chapters 2 (classification) and 3 (textures). Chapter 4 is designed to give you a survey of igneous rocks in their field setting. This is largely a review of fundamental concepts. Next we get some theory on melt systems, and use that theory to look at some simple experimental systems. Then we will proceed to more complex natural systems and the (largely chemical) tools required to study them. Finally we will embark upon an overview of the most common igneous provinces and review the current state of modern theories on their development. This latter portion will require considerable application and even amplification of the theoretical skills we have acquired.

On a more grand scale, igneous, metamorphic, and sedimentary processes all contribute to a differentiated planet. Igneous processes are by far the most dominant in this regard, as they are largely responsible for the segregation of the crust from the mantle and for the origin of many natural resources. It will benefit us all to keep this large-scale differentiation in mind as we explore the more focused components in what follows.

1.2 THE EARTH'S INTERIOR

As mentioned above, virtually all igneous rocks originate by melting of material at some depth within the Earth. All terrestrial rocks that we now find at the Earth's surface were derived initially from the mantle, although some have gone through one or more cycles of subsequent sedimentary, metamorphic, and/or igneous processes since. If these rocks have an initial origin at depth, it follows that we need to know what makes up the Earth if we want to understand their origin more fully. The Earth's interior is subdivided into three major units: the *crust,* the *mantle,* and the *core* (Figure 1-2). These units were recognized decades ago, during the early days of seismology, since they were separated by major discontinuities in the velocities of *P* (compres-sional) and *S* (shear) waves as they propagate through those layers in the Earth (Figure 1-3).

There are two basic types of crust: *oceanic crust* and *continental crust.* The **oceanic crust** is thinner (about 10 km thick) and has an essentially basaltic composition. We will learn more about the composition, structure, and origin of the oceanic crust in Chapter 13. Because the process of plate tectonics is creating oceanic crust at mid-ocean ridges, and consuming it at subduction zones, the oceanic crust is continually being renewed and recycled. The oldest oceanic crust is in the southwest Pacific and

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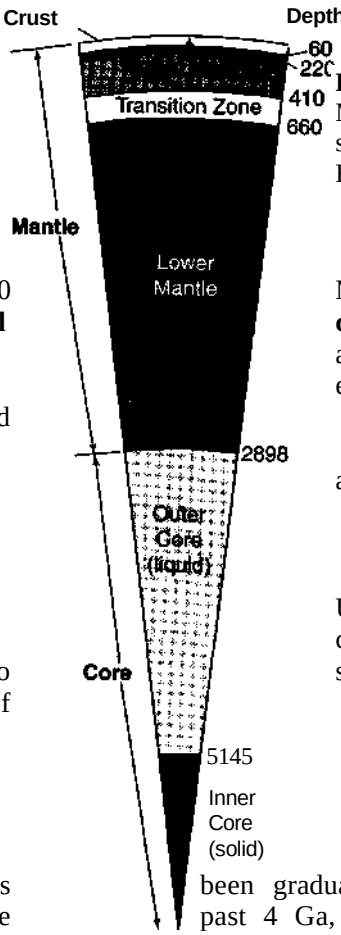


Figure 1-2 Major subdivisions of the Earth.

is about 160 Ma old. The **continental crust** is thicker (averaging about 36 km, but extending up to 90 km) and more heterogeneous, including all sorts of sedimentary, igneous, and metamorphic rocks. Unlike the oceanic crust, it is too buoyant to subduct. The amount of continental crust has been gradually increasing over the past 4 Ga, principally in the form of mantle-derived melts. Some continental crust is thus very old, whereas some is quite new. There will be lots more to say about the continental crust throughout this book. A *very* crude average composition of the continental crust would be represented by a granodiorite. The crust is too thin to represent on Figure 1-2. Even the thickest continental crust would be thinner than the uppermost line. It comprises about 1% of the volume of the Earth.

Immediately beneath the crust, and extending to nearly 3000 km, is the **mantle,** comprising about 83% of the Earth's volume. The boundary, or discontinuity, between the crust and mantle is called the Moho, or M discontinuity (shortened from Mohoroviac, the name of the

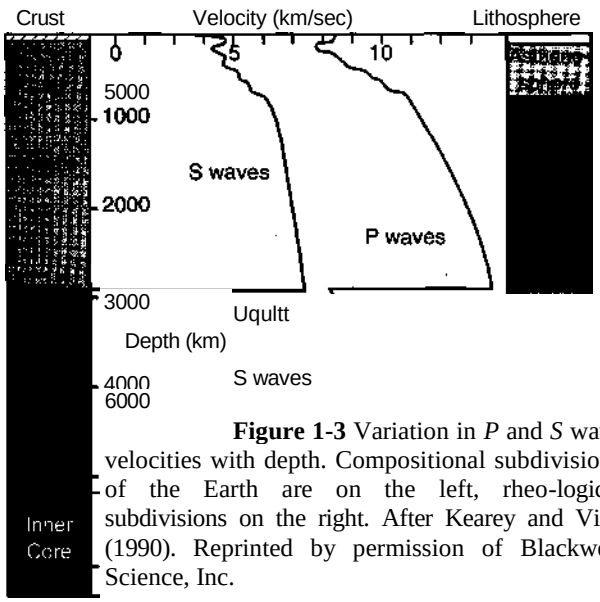


Figure 1-3 Variation in *P* and *S* wave velocities with depth. Compositional subdivisions of the Earth are on the left, rheo-logical subdivisions on the right. After Kearey and Vine (1990). Reprinted by permission of Blackwell Science, Inc.

Balkan seismologist who discovered it in 1909). At this discontinuity, the velocity of *P* waves increases abruptly from about 7 to over 8 km/sec. This results in refraction, as well as reflection, of seismic waves as they encounter the discontinuity, making it relatively simple to determine the depth. The mantle is composed predominantly of Fe- and Mg-rich silicate minerals. We will learn more about the petrology of the mantle in Chapter 10.

Within the mantle several other seismic discontinuities separate layers that are distinguished more by physical than chemical differences. The shallowest such layer, between 60 and 220 km, is called the **low velocity layer,** because within it, seismic waves slow down slightly, as compared to the velocity both above and below the layer (Figure 1-3). The slowness of seismic waves is unusual, because their velocities generally increase with depth since they propagate more readily through denser materials (just as sound travels more rapidly through water than through air). The reason seismic waves slow down in the low velocity layer is believed to be caused by 1 to 10% partial melting of the mantle. The melt probably forms a thin discontinuous film along mineral grain boundaries, which retards the seismic waves. The melt also weakens the mantle in this layer, making it behave in a more ductile fashion. The low velocity layer varies in thickness, depending on the local pressure, temperature, melting point, and availability of H_2O. We will discuss the origin of the low velocity layer further in Chapter 10.

Below the low velocity layer we encounter two more seismic discontinuities within the mantle. The 410-km discontinuity is believed to result from a phase transition in which olivine (a major mineral component of the mantle) changes to a spinel-type structure. At 660 km, the coordination of Si in mantle silicates changes from the familiar IV-fold to Vl-fold, as in the mineral perovskite. Both of these transitions result in an abrupt increase in the density of the mantle, accompanied by a jump in seismic velocities.

Below the 660-km discontinuity, the velocities of seismic waves increase fairly uniformly (Figure 1-3) until they encounter the **core.** The mantle/core boundary is a profound chemical discontinuity at which the silicates of the mantle give way to a much denser Fe-rich metallic alloy with minor amounts of Ni, S, Si, O, etc. The outer core is in the liquid/molten state whereas the inner core is solid. The composition of the inner and outer core is probably similar. The transition to a solid results from increased pressure with depth, which favors the solid state. S-waves cannot propagate through a liquid, because liquids cannot resist shear. Although 5-waves are only slowed by the thin liquid films in the low velocity layer, they disappear entirely as they reach the outer core (Figure 1-3). P-waves slow in the liquid core, and refract downwards, resulting in the seismic "shadow zone," a ring-like zone in which earthquake P-waves don't reach the surface of the Earth on the side away from which they originated.

An alternative way to consider the subdivisions of the Earth is based on rheological properties. Using this criterion, we may consider the crust and the more rigid portion of the uppermost mantle above the low velocity layer to behave as a strong coherent unit, and they are collectively called the **lithosphere.** The lithosphere averages about 70 to 80 km thick under the ocean basins, and 100 to 150 km thick under the continents. The more ductile mantle below this is called the **asthenosphere** (from the Greek *asthenes:* "without strength"). The lithosphere and asthenosphere are thus distinguished by their mechanical properties, not by composition or seismic velocity. Although they are not included in Figure 1-2, they are important in the theory of plate tectonics, because the ductility of the asthenosphere is thought to provide the zone of dislocation upon which the rigid lithospheric plates move. The mantle below the asthenosphere is commonly called the **mesosphere.** The exact asthenos-phere-mesosphere boundary should correspond to the transition from ductile to more rigid material with depth. The bottom of the ductile layer is poorly constrained. Most geophysicists believe that the asthenosphere extends to about 700 km depth. The nature of the mantle below this is not well known, but seismic waves which cross the mesosphere below 700 km are not greatly at-

tenuated, suggesting that this is a layer of high strength. The rheological subdivisions of the mantle are illustrated on the right side of Figure 1-3.

1.3 ORIGIN OF THE SOLAR SYSTEM AND THE EARTH

Now that we have some idea of what comprises the Earth, it is interesting to speculate on how it got that way. The following scenario summarizes the most generally accepted theories on the origin of the solar system. Although it will be presented as fact, this is only done so as to avoid the constant use of disclaimers and indefinite phrases. Please remember that this is only a collection of internally consistent ideas by which we explain what we now observe, although the extent of this consistency does lend credence to the models.

The most popular model for the origin of the universe has the Big Bang occurring between 12 to 15 Ga before present (b.p.). According to radiometric dating of meteorites, the solar system began to form about 4.56 Ga b.p. as a huge cloud of matter called the **solar nebula** (Figure 1-4). The nebula consisted mostly of molecular H_2 plus some He and minor Be and Li (the only products of the Big Bang). A further 2% comprises heavier elements, including some other gases and fine solid particles ("dust"), presumably created by nuclear synthesis reactions in earlier nearby stars and supernovae. The nebular cloud began to collapse slowly because of the gravitational interactions of its constituents. Because it was rotating, it flattened to a disk-like shape as a result of centrifugal forces, with 1 to 10% of the mass constituting the central disk. The balance between gravitational collapse, centrifugal force, and conservation of angular momentum resulted in the majority of the mass losing angular momentum and falling to the center of the disk, eventually to form the sun. Small meter- to kilometer-sized bodies, called **planetesimals,** began to form and grow in the nebula. The gravitational collapse of the mass and its compression generated considerable heat, eventually reaching the stage where nuclear synthesis (fusion) of hydrogen to helium became possible.

The first 100,000 years witnessed a very rapid evolution of the "proto-sun," accompanied by a high luminosity caused by the heat generated by the initial contraction. When the compression was nearly over, the sun entered the **T-Tauri stage,** characterized by less vigorous activity, lasting up to 10 Ma. The **solar wind,** a stream of charged particles, changed character during the T-Tauri stage and began to emanate radially outward from the sun, rather than spirally from the poles. The nebula lost about half of its initial mass during this stage.

Of the remaining material, 99.9% of the mass collapsed to form the sun, and the other 0.1 %, with the majority of the angular momentum, remained in the disk. The disk material had sufficient mass to contract to the median plane, where it eventually separated into localized accumulations that formed the planets. The process of planetary accretion took place within a strong temperature and pressure gradient generated by the early sun. As a result, the more volatile elements comprising the solid particles of the nebula evaporated in the inner, hotter portion of the solar system. The vapor particles were then stripped off by the intense T-Tauri solar wind, and condensed directly to solids further outward where the temperature was sufficiently low. Only the larger planetesimals survived this intense activity in the inner solar system. The actual condensation temperatures (and hence the distance from the sun at which condensation took place) depended upon the particular elements or compounds involved. Only the most refractory elements survived or condensed in the innermost zone, whereas the more volatile constituents were moved further outward. As a result, then, primarily of the temperature gradient and solar wind, the nebula experienced a chemical differentiation based on condensation temperatures. Refractory oxides such as $A1_2O_3$, CaO, and TiO_2 either failed to volatilize at all or condensed quickly in the innermost portions of the solar system. Fe-Ni metal alloys, Fe-Mg-Ni silicates, alkali metals and silicates, sulfides,

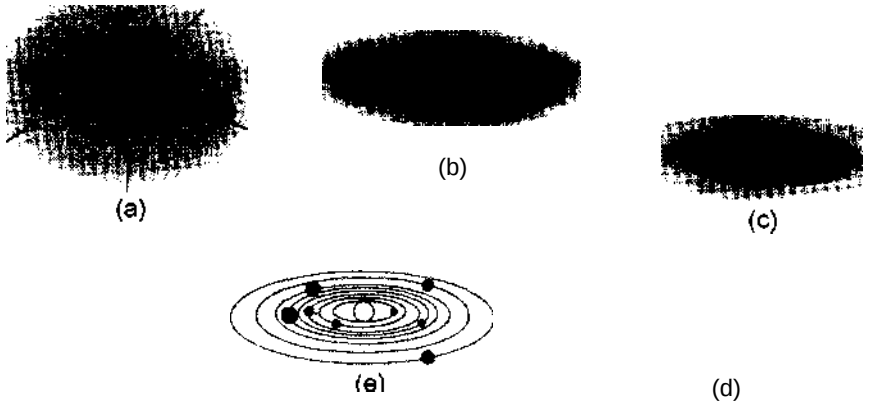


Figure 1-4 Nebular theory of the formation of the solar system, (a) The solar nebula condenses from the interstellar medium and contracts, (b) As the nebula shrinks, its rotation causes it to flatten to a disk (c), with most of the matter concentrated toward the center as the primordial sun (d). Outer solid particles condense and accrete to form the planets (d) and (e). From Abell et al. (1988). Copyright © by permission Saunders.

hydrous silicates, H_2O, and solids of ammonia, methane, etc., condensed and concentrated progressively outward. The distance beyond which the very volatile compounds such as water and methane condensed has been referred to as the **snow line.** Apparently, a gradient of decreasing pressure outward from the center of the nebula also had an effect, principally on the relative condensation temperatures of Fe metal vs. silicates, and thus on the Fe/Si ratio (and oxygen content) of the planets.

The condensed solids next continued to accrete as planetesimals. In the inner portion of the solar system, the more refractory planetesimals further accumulated and formed the **terrestrial** (Earth-like) planets (Mercury, Venus, Earth, and Mars) as well as the parent bodies that produced the present asteroids and meteorites. In the outer portions, beyond the snow line, the large **gaseous** planets formed. Pluto is anomalous in orbit and probably composition as well. It may be a moon of one of the planets that escaped into solar orbit or the nearest body of icy objects with orbits beyond Neptune.

From this very brief sketch, it seems clear that the composition of a planet is in large part the result of rather specific conditions that existed at a particular radial distance from the center of the solar nebula during the first 10 Ma of stellar evolution. The composition of the Earth is largely a result, then, of the nature of the ancient supernova that "seeded" the solar nebula with solid particles, and the evaporation/condensation processes associated with the temperature at Earth's particular distance from the T-Tauri sun. Thus we would not expect the Earth's composition to be equal to that of other planets or to that of the solar nebula as a whole.

The differentiation process that produced the chemical variation across the solar system was not perfectly efficient. The composition of the Earth is complex, containing some of every stable element, and not just those that could condense at our distance from the sun. Some of the varied constituents of the Earth, including the volatiles, were contained in the early planetesimals that were large enough to resist complete vaporization during the hot T-Tauri stage of solar evolution, whereas

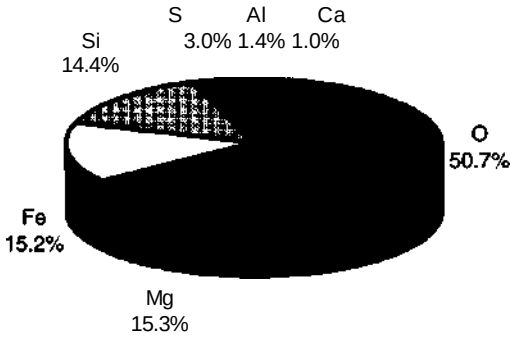


Figure 1-5 Relative atomic abundances of the seven most common elements that comprise 97% of the Earth's mass.

others may have been added later via impact of comet-like bodies from the outer solar system. Nonetheless, the process described above strongly favored the concentration of certain elements, and only seven elements now comprise 97% of the mass of the Earth (Figure 1-5). These elements are consistent with the solar abundances and condensates that we expect to have formed at the pressures and temperatures at the Earth's position in the nebular gradients described above.

1.4 DIFFERENTIATION OF THE EARTH

The planetesimals that now form the Earth probably accumulated in a sequential fashion caused by the gravitational accretion of denser ones first, concentrating Fe-Ni alloys and denser oxides toward the center. Whether or not the Earth got this "head start" toward further differentiation is hard to say, but it differentiated more extensively soon thereafter. This extensive differentiation resulted from heating, caused by gravitational collapse, impacts, and concentrated radioactive heat. Eventually the planet heated sufficiently to initiate melting at some shallow depth, probably beneath a solid crust that stayed cooler by radiating heat to space. Once melting began, mobility within the Earth increased. Denser portions of the melts moved downwards, whereas lighter portions rose. The gravitational energy released by this process probably generated enough heat to melt the entire Earth, with the possible exception of the outermost chilled layer. This layer may also have melted if there was sufficient gaseous atmosphere to retard radiant cooling.

The result of the process was for the Earth to separate into layers controlled by density and the chemical affinities of the elements that comprise it. The concept of chemical affinity will be developed further in Chapters 9 and 27, but, in simplest terms, we can say for now that element behavior is controlled by the configuration of electrons in the outermost shells, which effects their bonding characteristics.

Goldschmidt (1937) proposed that the elements of the Earth tended to incorporate themselves into separate phases, analogous to the layers in ore smelting pots. We have inherited his terms:

- **Lithophile** ("stone-loving") elements form a light silicate phase.
- **Chalcophile** ("copper-loving") elements form an intermediate sulfide phase.
- **Siderophile** ("iron-loving") elements form a dense metallic phase.

A separate phase of **atmophile** elements may also have formed in the early Earth as a very minor ocean and atmosphere, but most of these light gaseous elements were not held by the Earth during the earliest stages and

escaped into space. Most of the oceans and atmosphere probably accumulated slowly later.

It is simple enough to determine the affinity of every element empirically, and use the results to predict the size (thickness) of each reservoir in the early differentiated Earth, but this approach doesn't work very well. For instance, Fe, which should be siderophile, occurs in all three phases. We must remember that the atoms are ionized, so the requirement of electric neutrality must be satisfied as well. We usually concentrate on cations, but the anions are equally important. For example, sulfur is obviously required to create a sulfide, so the amount of sulfur dictates the size of the chalcophile layer in smelting pots. Because there was not enough sulfur to satisfy all the chalcophile cations in the Earth, the excess chalcophile cations had to go elsewhere. Oxygen is the principal anion in silicate minerals. It combined with silicon for the lithophile layer, but other cations were required before neutrality was achieved. The most common minerals in the lithophile layer of the early Earth were probably olivine ($(Fe,Mg)_2SiO_4$), orthopyroxene ($(Fe,Mg)SiO_3$), and clinopyroxene ($Ca(Fe,Mg)Si_2O_6$). It was thus the relative abundance of oxygen that determined the thickness of the upper lithophile layer. The inner siderophile layer was determined by the excess of siderophile cations (mostly Fe) left over after neutrality was achieved with O and S. All the other elements, comprising the remaining 3% of the Earth's mass, went preferentially into one of these layers, in accordance with a particular atom's affinity.

As with the differentiation of the solar system, the Earth's differentiation was certainly not perfectly efficient: Not all of the elements are restricted to the predicted layer. Otherwise, we would never find such elements as gold (siderophile), copper (chalcophile), etc., at the Earth's surface today. This may be caused, in part, by a lack of complete equilibrium during the differentiation process, but (as we shall see in Chapters 9 and 27) even if equilibrium is attained, elements commonly partition themselves into different reservoirs in less than the most extreme proportions (not all into one reservoir).

After a few hundred million years, this molten, differentiated Earth cooled and mostly solidified to a condition similar to the planet we now inhabit, having a distinct temperature and pressure gradient with depth.

The lithophile, chalcophile, and siderophile layers are *not* to be confused with the present layers of the Earth: crust, mantle, and core. The core of the modern Earth is the siderophile layer, but the chalcophile component was most likely dissolved in the siderophile core and never separated as a distinct phase. Although such a phase does form in smelters, it is much less likely to do so at the high pressures involved in the core. The Earth is not a smelting pot. If a separate chalcophile phase did form, it might be an outermost layer of the outer core, but has not been detected seismically.

The mantle certainly represents the lithophile segregation, but what about the crust? Neither the oceanic nor the continental crust formed at this point by a large-scale differentiation event in the early Earth (although this may have happened for the plagioclase-rich highlands of the Moon). Rather, the Earth's oceanic crust has recycled many times in the past and the continental crust has evolved slowly over time. The processes by which the mantle differentiates to produce the crust are predominantly igneous in nature. We will return to this subject at several places later on.

1.5 HOW DO WE KNOW ALL THIS?

If you are now asking yourself how we can possibly know what has just been presented, you're approaching petrology with the right attitude. Theories, such as those concerning the origin of the universe, the solar system, and the Earth represent the best inferences we can make based on our interpretation of the data. The simplest explanation of all the data, without violating physical "laws" is preferred. The more varied the nature of the phenomena a theory explains, the more confidence we place in it. The scenario described above is consistent with the physical "laws" of celestial mechanics, gravity, nuclear synthesis, and so on. It is also consistent with our observations of seismic waves and the nature and make-up of the solar system. But rigorous evaluation of these criteria is well beyond the scope of this book. The scenario is intended only as background information, however. The data on the composition and layering of the Earth's interior in Figures 1-2 and 1-3 are the final result of the process and are presented as fact. These data are very important to the material that we will address in the pages ahead. After all, if igneous rocks are the products of melting at depth, it might be nice to know with some confidence what is being melted. For petrologic concerns, let's focus our skepticism here for a moment. We have not yet drilled a hole to the mantle (and will never do so to the core) in order to directly sample these materials. And our hypothetical mantle and core are far different than the materials we find at the Earth's surface. What evidence do we have to support the alleged composition and structure of our planet?

First, from careful measurements we can accurately determine the gravitational constant, and use that, plus the measured moment of inertia of the Earth, to calculate its mass, and from that, the average density. This places several constraints on the materials which make up the Earth. For instance, the average density of the Earth is approximately 5.52 g/cm^3. It is relatively easy to observe and inventory the chemical composition of the rocks exposed at the surface of the Earth. But the density of surface rocks is rarely greater than 3.0 g/cm^3. The Earth must thus contain a large proportion of material that is much denser than can be accomplished by com-

pression of surface-type rocks due to the increased pressure at depth.

One could come up with a variety of recipes for the dense material at depth, by mixing proportions of atoms of various atomic weights. However, such a random approach would better be guided by having some idea of which elements are more naturally abundant. The Earth must have formed from the solar nebula, so the composition of the nebula must provide us with significant clues to the makeup of our planet. The material that makes up the solar system can be analyzed from a distance by spec-troscopic means. Atoms can be excited by heat or particle interactions and emit characteristic light spectra when they return to their lower energy "ground state." The wavelength of the light that reaches Earth can be determined and related to the type of element or compound that emits it. By comparison with spectra of elements measured in the laboratory, the emitting atoms or molecules can be identified. The intensity of the spectral lines is proportional to their concentrations at the source. We thus get a good idea of what elements constitute the sun, other stars, even other planetary surfaces, and by analogy with these, our own planet.

Figure 1-6 illustrates the estimated concentrations of the elements in the solar nebula (estimated from certain meteorites, as discussed below). Note the logarithmic scale for concentration, which makes it easier to show the full range of abundances. Hydrogen is by far the most abundant element, as it made up most of the original nebula. Other elements (except He) were synthesized from H in the sun and other stars. The decrease in abundance with increasing atomic number (Z) reflects the difficulty of synthesizing progressively larger atoms. Another interesting feature that is clear from Figure 1-6 is the "saw-tooth" nature of the curve. This is in accordance with the "Oddo-Harkins rule," which says that atoms with even numbers are more stable, and hence more abundant than their odd-numbered neighbors.

We must assume that the elements that comprise the Earth are among the more common elements in Figure 1-6. For example, Fe, and to a lesser extent Mg and Ni, are much more abundant in the solar system than in the Earth's crust, so we might infer that these elements are concentrated elsewhere in the Earth. Fe is also dense enough to satisfy the Earth's high density requirement. In other words, using the data in Figure 1-6 as a starting point to model a planet with an average density of 5.52 g/cm^3 should lead us in the direction of the concentrations in Figure 1-5. Of course the process is complicated by such inhomogeneities as the radial differentiation of the solar nebula, and density variations and phase changes associated with increasing pressure in the Earth.

Seismic studies place further constraints on the materials that constitute the Earth. The velocities of *P* and *S* waves in various materials at elevated pressures and temperatures can be measured in the laboratory and compared to seismic velocities within the Earth, as determined from earthquakes or manmade explosions (Figure 1-3). In addition, reflection and refraction of seismic waves at discontinuities within the Earth provide direct evidence for the Earth's internal structure and the depths of the discontinuities that subdivide it into crust, mantle, outer core, and inner core, as well as other more detailed features.

Finally, although we haven't visited the mantle or core for samples, we have had samples delivered to us at the surface (or so we believe). We will learn more about

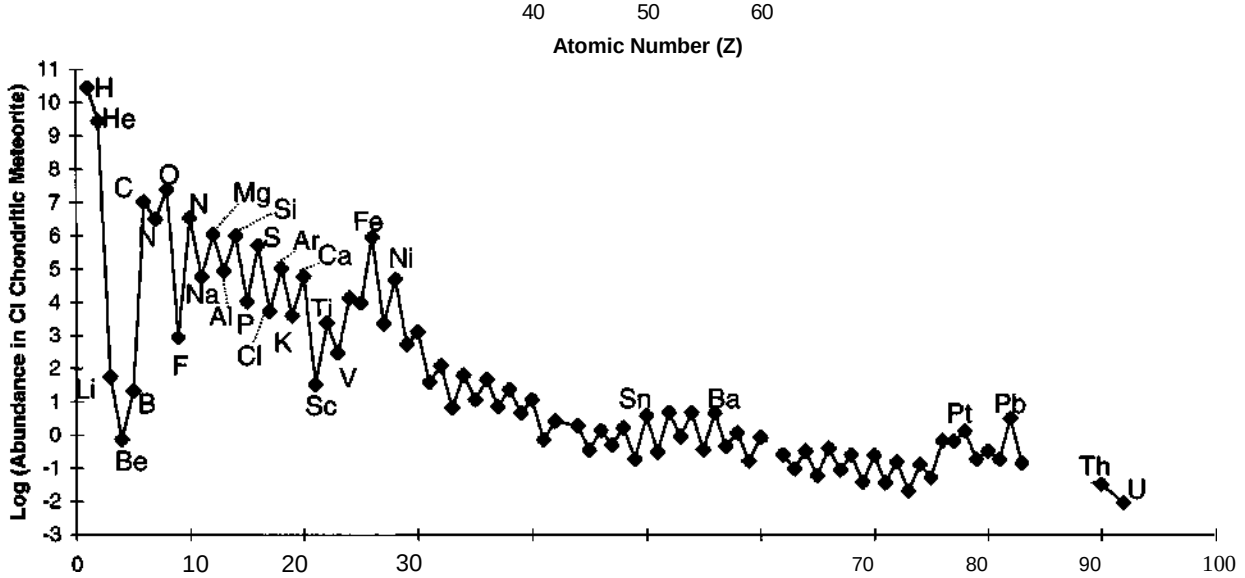


Figure 1-6 Estimated abundances of the elements in the solar nebula (atoms per 10^6 Si atoms). After Anders and Grevesse (1989). Copyright © with permission from Elsevier Science.

mantle samples in Chapters 10 and 19, but there are a number of rocks found at the surface that we believe to be of mantle origin. In many active and fossil subduction zones, slivers of oceanic crust and underlying mantle are incorporated into the accretionary prism. Thickening of the prism, followed by uplift and erosion, exposes the mantle-type rocks. Xenoliths of presumed mantle material are occasionally carried to the surface in some basalts. Deeper mantle material is believed to come to the surface as xenoliths in diamond-bearing kimberlite pipes (Section 19.3.3). The vast majority of samples found in all of these situations are olivine- and pyroxene-rich ul-tramafic rocks. When many of these samples are partially melted in the laboratory, they produce melts similar to natural lavas that we believe to be mantle-derived.

Because of its great density and depth, not a single sample of the Earth's core has reached the surface. However, we do believe that pieces of the core of other plan-etesimals have reached Earth in the form of some meteorites. We will briefly discuss meteorites in the next section, as they are quite varied, and provide important information about the composition of the Earth and solar system.

1.6 METEORITES

Meteorites are solid extraterrestrial objects that strike the surface of the Earth after surviving passage through the atmosphere. Most of them are believed to be fragments derived from collisions of larger bodies, principally from the asteroid belt between the orbits of Mars and Jupiter. They are very important, because many are believed to represent arrested early-to-intermediate stages in the development of the solar nebula that have not undergone subsequent alteration or differentiation like the Earth. They thus provide valuable clues to the make-up and development of the solar system. Meteorites have been classified in a number of ways. Table 1-1 is a simplified classification, in which I have combined several subclasses to give a general indication of the more important types and the percentages of each from observed falls.

Irons (Figure l-7b) are composed principally of a metallic Fe-Ni alloy, **stones** are composed of silicate minerals, and **stony-irons** (Figure l-7a) contain subequal amounts of each. Because stones look much like terrestrial rocks, they are commonly not recognized as meteorites, so irons (quickly recognizable by their density) tend to dominate museum collections. When we consider only specimens collected after an observed fall, however, stones comprise 94% of meteoritic abundance. Iron meteorites are believed to be fragments of the core of some terrestrial planets that have undergone differentiation into concentrations of silicate, sulfide, and metallic liquids, in the manner discussed above for the Earth. These

Table 1-1 Simplified Classification of Meteorites

Class		Subclass	# of Falls	% of Falls
Irons		All	42	5
Stony-irons		All	9	1
Stones	Achondrites	SNC's	4	8
		Others	65	
	Chondrites	Carbonaceous	35	86
		Others	677	

after Sears and Dodd (1988)

meteorites contain portions of siderophile (Fe-Ni alloy) and chalcophile (segregations of troilite: FeS) phases. The Fe-Ni alloy is composed of two phases, kamactite and taenite, which have exsolved from a single, homogenous phase as it cooled. The two phases are commonly intergrown in a cross-hatched pattern of exsolution lamellae called "Widmanstatten texture" (Figure l-7b). Stony-irons are similar to irons, but include a significant proportion of the silicate (lithophile) segregation mixed in. All irons and stony-irons are considered "differentiated" meteorites, because they come from larger bodies that underwent some degree of chemical differentiation. Meteorites however, display a large variation in the degree to which they represent differentiated portions of a planet. The parent bodies, most believed to be present in the asteroid belt, were of various sizes, and thus capable of different degrees of differentiation. In addition, the collisions that disrupted the parent bodies into meteoritic fragments have remixed, and even brecciated the material that we find in many meteorites.

Stones are further subdivided on the basis of whether or not they contain **chondrules** (Figure l-7c and d), nearly spherical silicate inclusions between 0.1 and 3 mm in diameter. At least some chondrules appear to have formed as droplets of glass that have subsequently crystallized to silicate minerals. Stones with chondrules are called **chondrites,** whereas those without are called **achondrites.** As with irons and stony-irons, achondrites are differentiated meteorites. Chondrites, on the other hand, are considered "undifferentiated" meteorites, because the heat required to permit melting and differentiation of a planet would certainly have destroyed the glassy chondrules. The small size of the chondrules indicate rapid cooling (<1 hr) requiring a cooler nebula at the time of their formation. They probably formed after condensation, but before formation of the planetesimals. Chondrites are thus considered to be the most "primitive" type of meteorites, in the sense that they are thought to have compositions closest to the original solar nebula (hence their use in estimating solar abundances in Figure 1-6). It has been suggested that all of the inner terrestrial planets formed from a material of average

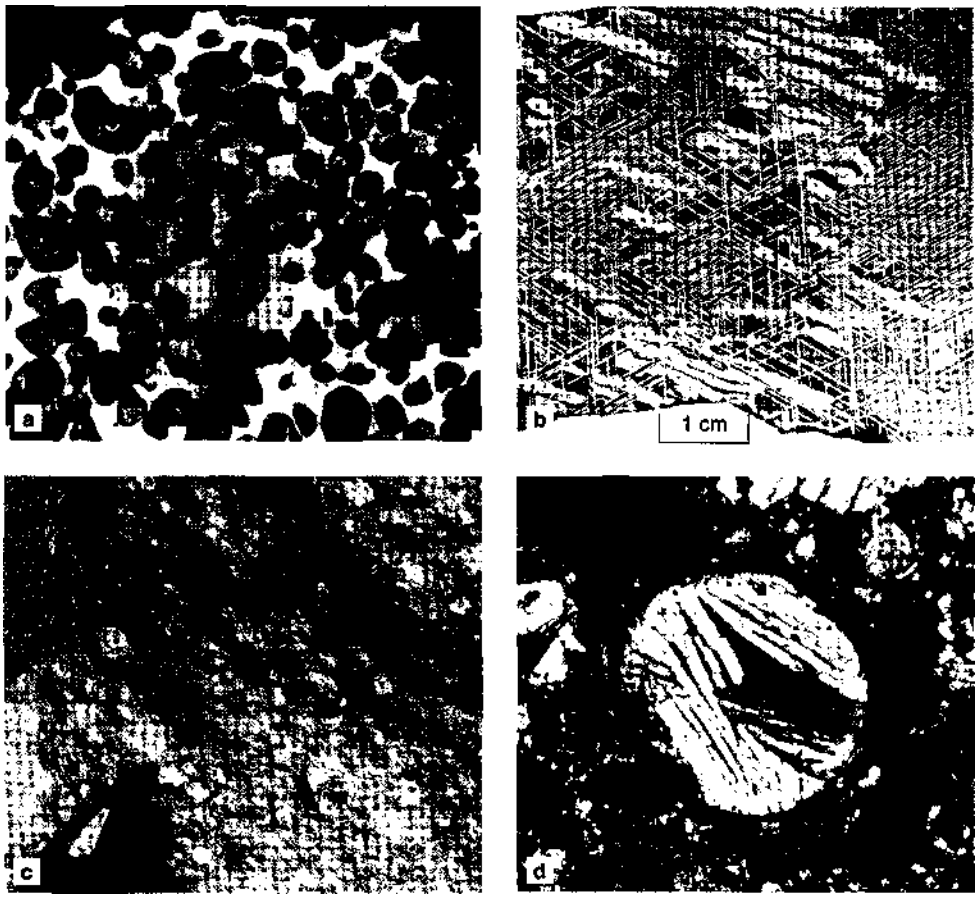


Figure 1-7 Meteorite textures, (a) Polished section of the stony-iron Springwater (Saskatchewan, Canada) meteorite (light is iron, dark is ~ 5mm olivines). © Courtesy Carleton Moore, Arizona State University, **(b)** Widmanstatten texture in the Edmonton (Kentucky) iron meteorite. © Courtesy John Wood and the Smithsonian Astrophysical Observatory, (c) Chondrules in the Allende chondrite meteorite (Mexico). Ball pen tip for scale, **(d)** Photomicrograph of a 0.5 mm diameter chondrule from the Dhajala (India) chondrite meteorite. The chondrule is composed of skeletal plates of olivine in a finer matrix including crystal fragments and glass. Other chondrules may be dominated by orthopyroxene, or mixtures of several minerals, including sulfides. Chondrules may be very fine grained or even totally glassy in meteorites that have suffered little or no reheating. Chondrite photographs courtesy of the Geological Museum, University of Copenhagen ©.

chondritic composition. This has led to the Chondritic Earth Model (CEM). The model provides a close fit to the composition of the Earth for most elements, but there are a few important differences. For example, the Earth is much denser, and must have a higher Fe/Si ratio than chondrites. Models such as the one presented above, based on condensation temperatures as a function of distance from the sun, are much better for explaining the chemical composition of the planets (particularly their variations) than is assuming some meteorite represents them all.

Further subdivision of meteorites is based on their textures and/or mineral content. There is considerable variety in the overall ("bulk") composition, as well as in the mineralogy. Over 90 minerals have been found among the stony meteorites, some of which are not found elsewhere on Earth. Some meteorites appear to come from the moon and neighboring planets. The "SNC" meteorites, for example, appear to be from Mars. Given this variety, the study of meteorites can provide us with valuable information on the chemical composition of the solar system and its constituents.

1.7 PRESSURE AND TEMPERATURE VARIATIONS WITH DEPTH

We now have a good idea of what comprises the Earth, and how it got there. If we are to proceed to an understanding of melts (and later of metamorphism) we should next attempt to understand the physical conditions (pressure and temperature) that occur at depth, so that we may appreciate how these materials respond and behave. As depth within the Earth increases, both pressure and temperature increase as well. Pressure increases as a result of the weight of the overlying material, whereas temperature increases as a result of the slow transfer of heat from the Earth's interior to the surface.

1.7.1 The Pressure Gradient

The pressure exerted in a ductile or fluid medium results from the weight of the overlying column of the material. For example, the pressure that a submarine experiences at depth is equal to the weight of the water above it, which is approximated by the equation:

$$P = pgh \qquad (1\text{-}1)$$

where P is pressure, p is the density (in this case, that of water), g is the acceleration caused by gravity at the depth considered, and h is the height of the column of water above the submarine (the depth). Because water is capable of flow, the pressure is equalized, so that it is the same in all directions. The horizontal pressure is thus equal to the vertical pressure (the axis along which the imaginary column of water would exert itself). This equalized pressure is called **hydrostatic** pressure. Near the surface, rocks behave in a more brittle fashion, so they can support unequal pressures. If the horizontal pressures exceed the vertical ones, rocks may respond by faulting or folding. At depth, however, the rocks also become ductile, and are capable of flow. Just as in water, the pressure then becomes equal in all directions, and are termed **lithostatic** pressures. Equation (1-1) will also then apply, with p being the density of the overlying rock. The relationship between pressure and depth is complicated because density increases with depth as the rock is compressed. Also, g decreases as the distance to the center of the Earth decreases. A more accurate approach would be to use a differential form of the P-depth relationship, complete with estimates of the variation in g and p, and integrate it over the depth range. However, the changes in g and the density of a given rock type are relatively minor in the crust and upper mantle, and also tend to offset each other, so Equation (1-1) should suffice for our needs. Only when the rock type changes, as at the Moho, would a different value of p be required. One need only calculate the pressure to the base of the crust using an appropriate average crustal density, and continue with depth using a density representative of the mantle.

For example, a reasonable estimate of the average density of the continental crust is 2.8 g/cm^3. To calculate the pressure at the base of 35 km of continental crust, we need only substitute these data into Equation (1-1), being careful to keep the units uniform:

$$P = \frac{2800\text{kg}}{\text{nr}} \quad 9.8\text{m} \quad 35{,}000\ \text{m}$$

$$s' = 9.6 \text{ X } 10^8 \text{ kg/(m s}^2\text{)} =$$

$$9.6 \text{ X } 10^8 \text{ Pa « 1 GPa}$$

See front inside cover for units and constants if necessary. This result is a good average pressure gradient in the continental crust of 1 GPa/35 km, or about 0.03 GPa/km, or 30 MPa/km. Because of (upward) round-off, this gradient is also suitable for oceanic crust. A representative density for the upper mantle is 3.35 g/cm^3, resulting in a mantle pressure gradient of about 35 MPa/km. These are numbers worth remembering, as they provide a good way to interpret pressures in the phase diagrams to come. Figure 1-8 shows the variation in pressure with depth using the Preliminary Reference Earth Model (PREM) of Dziewonski and Anderson (1981).

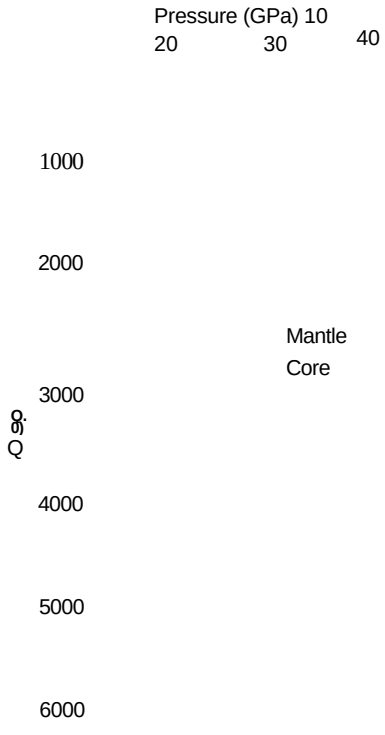


Figure 1-8 Pressure variation with depth. From Dziewonski and Anderson (1981).

1.7.2 The Temperature Gradient

Determining the **geothermal gradient,** the temperature variation with depth, is much more difficult than doing so for pressure, as there is no simple physical model analogous to Equation (1-1). There are models, however, based on methods of heat transfer, that must be constrained to conform to measured heat flow at the surface (or in "deep" drill holes and mines). There are two primary sources of heat in the Earth.

1. **Cooling.** Heat developed early in the history of the Earth from the processes of accretion and gravitational differentiation described in Section 1.4 has been gradually escaping since that time. This set up an initial temperature gradient once the planet solidified and began to cool. Some con tinued gravitational partitioning of iron in the inner core may contribute some heat as well.
2. Heat generated by the **decay of radioactive iso topes.** For reasons that will be discussed in Chap ters 8 and 9, most of the more radioactive elements are concentrated in the continental crust. Radioactive decay produces 30 to 50% of the heat that reaches the surface.

Once generated, heat is transferred from hotter to colder regions by any of four processes, depending on the nature of the material involved in the transfer:

1. If a material is sufficiently transparent or translu cent, heat can be transferred by **radiation.** Radia tion is the movement of particles/waves, such as light or the infrared part of the spectrum, through another medium. This is the principal way a lamp loses heat, or how the Earth loses heat from its surface into space. It is also the way we receive heat energy from the sun. Heat transfer by radia tion is not possible within the solid Earth except possibly at great depth, where silicate minerals may become hot enough to lose some of their opacity to infrared radiation.
2. If the material is opaque and rigid, heat must be transferred through **conduction.** This involves the transfer of kinetic energy (mostly vibrational) from hotter atoms to cooler ones. Heat conduc tion is fairly efficient for metals, in which electrons are free to migrate. This is why you can burn your self if you handle an iron bar that has one end in a fire. Conduction is poor for silicate minerals.
3. If the material is more ductile, and can be moved, heat may be transferred much more efficiently by **convection.** In the broadest sense, convection is the movement of material due to density differences caused by thermal or compositional variations. For the present purposes, we shall consider the type of convection that involves the expansion of a material as it heats, followed by the rise of that material due to its gain in buoyancy. This convection explains why it is hotter directly above a candle flame than beside it. The air is heated, it expands, and rises because it is now lighter than the air around it. The same thing can happen to ductile rocks or liquids. Convection may involve flow in a single direction, in which case the moved hot material will accumulate at the top of the ductile portion of the system (or a cooled density current will accumulate at the bottom). Convection may also occur in a cyclic motion, typically in a closed cell above a localized heat source. In such a *convection cell* the heated material rises and moves laterally as it cools and is pushed aside by later convective matter. As the upper portions of the system heat up, the material pushed to the margin is cooler and sinks toward the heat source where it becomes heated and the cycle continues.
4. **Advection** is similar to convection, but involves the transfer of heat with rocks that are otherwise in motion. For example, if a hot region at depth is uplifted by tectonism, induced flow, or erosion and isostatic rebound, heat rises physically (although passively) with the rocks.

Convection may work well in the liquid core and in the somewhat fluid asthenospheric mantle, and may be responsible for the high heat flows measured at mid-ocean ridges. It is also a primary method of heat transfer in hydrothermal systems above magma bodies or within the upper oceanic crust, where water is free to circulate above hot rock material. Beyond these areas, however, conduction and advection are the only available methods of heat transfer.

The transfer of heat is a very important concept in petrology, as it controls the processes of metamorphism, melting, and crystallization, as well as the mechanical properties of Earth materials. Several petrologic processes, from explosive volcanism to lava flows and pluton emplacement, are critically dependent upon maintaining a heat budget.

Heat flow is relatively high in newly generated oceanic crust and in orogenic areas, where magma is rising by convection and/or advection to shallower depths. However, this high heat flow settles down to a "steady-state" conductive pattern after about 180 Ma in the oceanic lithosphere (away from the ridge), and after about 800 Ma in the continents (Sclater et al. 1980). The mathematical models are more complex than we care to deal

with but are based on conductive models fit to the measured near-surface values of heat flow. Because of the concentration of radioactive elements in the continental crust, we can expect the continental geotherm to be greater than the oceanic geotherm, but the two should converge at greater depths in the mantle (below the zones of radioactive element concentration). Estimates of the steady-state heat flow from the mantle range from 25 to 38 mW/m^2 beneath the oceans to 21 to 34 mW/m^2 beneath the continents. Heat flow is commonly measured in heat flow units (HFU), for which 1 HFU = 41.84 mW/m^2.

The steady-state geotherms (calculated by Sclater et al. 1980) are shown in Figure 1-9. There is a range for each, based on the upper and lower heat flow estimates. Note that the geotherm is higher for the continents than for the oceans, but there is some overlap between the two, resulting from the uncertainty in near-surface heat-flow ranges. The differences are restricted to the crust and uppermost mantle, and the oceanic and continental curves converge by about 150 to 200 km. The gradients continue to steepen with depth. Below about 200 to 250

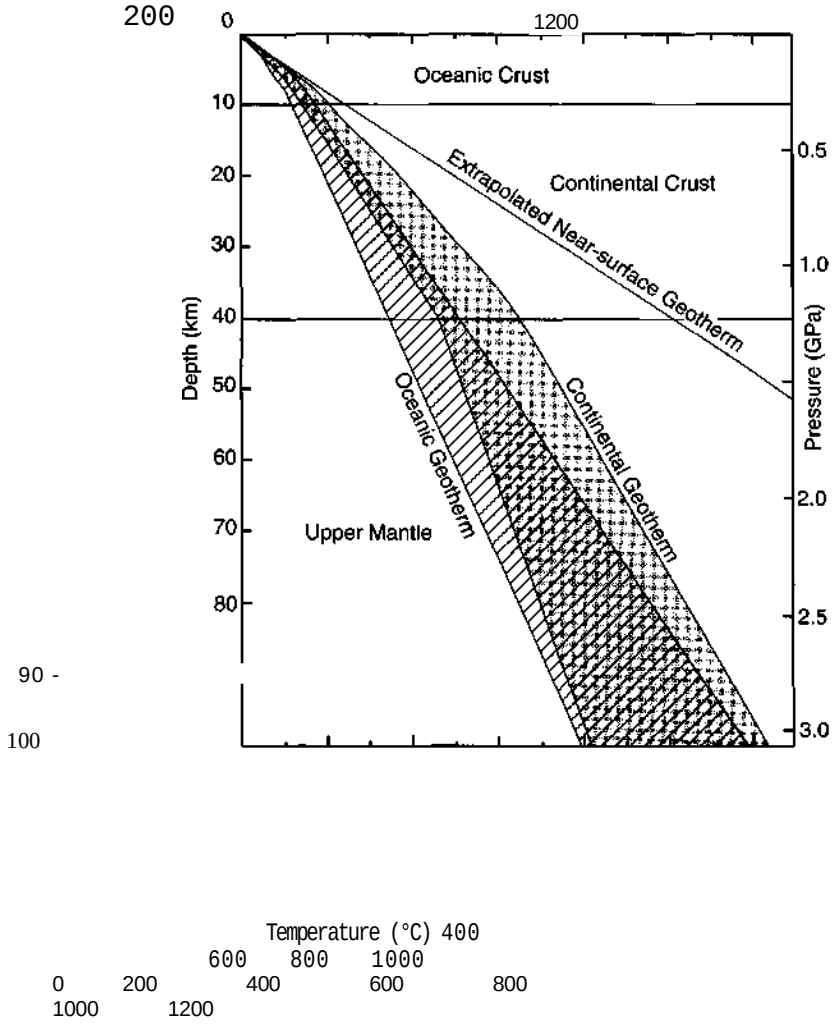


Figure 1-9 Estimated ranges of oceanic (hatched) and continental (solid) steady-state geotherms to a depth of 100 km (the approximate base of the lithosphere) using upper and lower limits based on heat flows measured near the surface. After Sclater et al. (1980).

km mantle convection results in a near adiabatic (constant heat content) heat flow and a geothermal gradient of approximately 0.3°C/km (10°C/GPa). The gradient becomes nearly vertical in the metallic core, where the thermal conductivity and convection (in the liquid portion at least) increases.

Heat transfer is also important when we consider the cooling of a body of magma. The rate at which a body cools depends upon a number of variables, including the size, shape, and orientation of the body; the existence of a fluid in the surroundings to aid in convective heat transfer; and the type, initial temperature, and permeability of the country rocks. An example of heat models applied to a cooling magma body will be discussed in Section 21.3.1.

1.8 MAGMA GENERATION IN THE EARTH

With this preamble we might next address the problem of magma generation. **Petrogenesis** is a good general term in igneous petrology for the generation of magma and the various methods of diversification of such magmas to produce igneous rocks. Most magmas originate by melting in the Earth's mantle, but some show evidence of at least a partial crustal component. Plate tectonics plays a major role in the generation of several magma types, but other types seem to result from processes at greater depths in the mantle than are influenced directly by plate tectonics. Figure 1-10 is a very generalized summary of the principal types of magmas and their geologic setting. We will study the processes and products in much more detail in Chapters 13 to 20.

The most voluminous igneous activity occurs at divergent plate boundaries. Of these, mid-ocean ridges (location 1, Figure 1-10) are the most common (Chapter 13). Beneath the ridge, the shallow mantle undergoes partial melting and the resulting basaltic magma rises and crystallizes to produce the oceanic crust. If a divergent boundary is initiated beneath a continent (location 2), a similar process takes place. The resulting magmatism, particularly at the early stages of continental rifting, is commonly alkaline, and typically shows evidence of contamination by the thick continental crust (Chapter 19). If the rift continues to develop, oceanic crust will eventually be created in the gap that forms between the separating continental fragments. The result will be a new ocean basin and igneous activity similar to location 1.

The oceanic plate created at mid-ocean ridges moves laterally and eventually is subducted beneath a continental or another oceanic plate. Melting also takes place at these subduction zones. The number of possible sources of magma in subduction zones is far more numerous than at ridges, and may include various components of mantle, subducted crust, or subducted

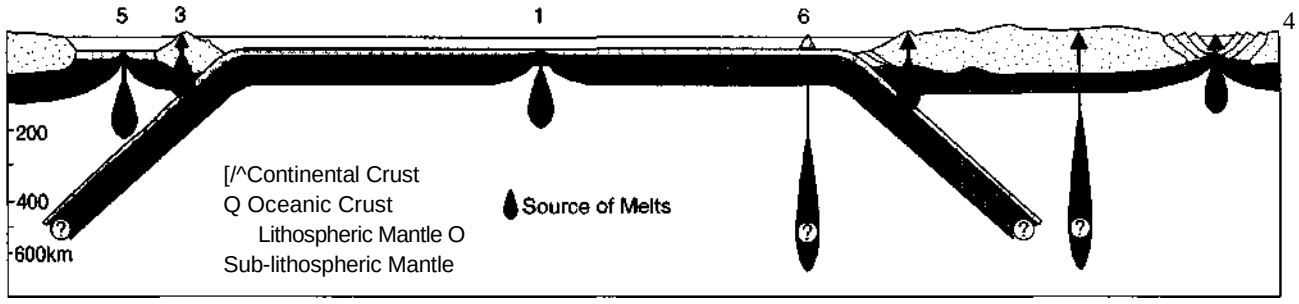


Figure 1-10 Generalized cross section illustrating magma generation and plate tectonic setting.

sediments. The types of magmas produced are correspondingly more variable than for divergent boundaries, but andesites are the most common. If oceanic crust is subducted beneath oceanic crust (location 3), a volcanic island arc forms (Chapter 16). If oceanic crust is subducted beneath a continental edge (location 4), a continental arc forms along the "active" continental margin (Chapter 17). A continental arc is generally more silica rich than is an oceanic arc. Plutons are also more common in continental arcs, either because the melts rise to the surface less efficiently through the lighter continental crust, or because uplift and erosion is greater in the continents and exposes deeper material.

A different, and slower, type of plate divergence typically takes place behind the volcanic arc associated with subduction (location 5). Most geologists believe some sort of "back-arc" extension is a natural consequence of subduction, probably created by frictional drag associated with the subducting plate. Such drag pulls down part of the overlying mantle, requiring replenishment from behind and below. Back-arc magmatism is similar to mid-ocean ridge volcanism. Indeed, a ridge also forms here, and oceanic crust is created and spreads laterally from it. Back-arc spreading, however, is slower, volcanism is more irregular and less voluminous, and the crust created is commonly thinner than in the oceans. At times, rifting occurs behind a continental arc, and the volcanic portion separates from the continent as a marginal sea forms by back-arc spreading. Such a process is believed to have separated Japan from the Asian mainland. At other times, such a process seems to initiate, and then it mysteriously ceases. The result may just be a graben structure, or plateau-type basalts may form prior to cessation of activity.

Although magmatism is certainly concentrated at plate boundaries, some igneous activity also occurs within toe plates, both oceanic (location 6) and continental (location 7). Ocean islands such as Hawaii, the Galapagos, the Azores, etc., all form by volcanism within the oceanic *plates (Chapter* 14). The products are usually basaltic but are commonly more alkaline than ridge basalts. The reason for this type of igneous activity is much less obvious than it is for plate margins, because our plate tectonic paradigm is of little use in these mid-plate regimes. The source of the melts is also less clear, but appears to be deep, certainly well into the asthenosphere. Several of these occurrences exhibit a pattern of igneous activity that gets progressively younger in one direction. The direction correlates well with plate motion in a manner that suggests the plate is moving over a stationary "hot spot" or mantle "plume," with the most recent activity occurring directly over the plume. Intraplate activity within continental plates is much more variable than that within the oceans (Chapter 19). It is composi-tionally variable, but usually alkaline, and occasionally extremely so. This reflects the more complex and heterogeneous continental crust, and sub-continental mantle as well. Some of the most unusual igneous rocks, such as kimberlites and carbonitites, occur within continental provinces. The term **igneous-tectonic association** refers to these broad types of igneous occurrence, such as mid-ocean ridge, island arc, or intra-continental alkalic systems. We will address these associations in the later chapters of the igneous section.

SUGGESTED FURTHER READINGS

The Interior of the Earth

Anderson, D. L. (1992). The Earth's interior, in *Understanding the Earth* (eds. G. C. Brown, C. J. Hawkesworth, and P. C. L. Wilson). Cambridge University Press. Cambridge.

Brown, G. C., and A. E. Mussett (1993). *The Inaccessible Earth.* Chapman & Hall. London.

Jacobs, J. A. (1975). *The Earth's Core.* Academic Press. London.

Jones, J. H., and M. J. Drake (1986). Geochemical constraints on core formation in the Earth. *Nature,* 322, 221-228.

Montagner, J.-P., and D. L. Anderson (1989). Constrained reference mantle model. *Phys. Earth Planet. Sci. Lett.,* 44, 205-207.

Wyllie, P. J. (1971). *The Dynamic Earth: Textbook in Geo-sciences.* John Wiley and Sons. New York.

The Origin of the Earth and Solar System

Brown, G. C. and A. E. Mussett (1993). *The Inaccessible Earth.* Chapman & Hall. London.

Newsom, H. W., and J. H. Jones, (1990). *Origin of the Earth.* Oxford University Press. New York.

Taylor, S. R. (1992). *Solar System Evolution: A New Perspective.* Cambridge University Press. Cambridge.

Taylor, S. R. (1992). The origin of the Earth. In G. C. Brown, C. J. Hawkesworth, and M. S. Matthews, (eds.) *Understanding the Earth.* Cambridge University Press. Cambridge.

Weaver, H. A., and L. Danly, eds. (1989). *The Formation and Evolution of Planetary Systems.* Cambridge University Press. Cambridge.

Wood, J. A. (1979). *The Solar System.* Prentice-Hall. Engle-wood Cliffs, N.J.

Meteorites

Goldstein, J. L, and J. M. Short (1967). The iron meteorites, their thermal history and parent bodies. *Geochim. Cos-mochim. Acta,* **31,**1733-1770.

Kerrige, J. F., and M. S. Matthews, eds. (1988). *Meteorites and the Early Solar System.* University of Arizona Press. Tus-con.

MacSween, H. Y. (1987). *Meteorites and their Parent Planets.* Cambridge University Press. Cambridge.

Wasson, J. T. (1985). *Meteorites.* W. H. Freeman. New York.

Pressure and Temperature Gradients in the Earth

Sclater, J. G., C. Jaupart, and D. Galson (1980). The heat flow through oceanic and continental crust and the heat loss of the Earth. *Rev. Geophys. Space Sci.,* **18,**269-311.

Williamson, E. D., and L. H. Adams (1923). Density distribution in the Earth. /. *Wash. Acad. Sci.,* **13,** 413-428.

Chapter 2

Classification and Nomenclature of Igneous Rocks

2.1 INTRODUCTION

This chapter is designed to assist the student in laboratories with hand specimen and thin-section identification. The preferred method for classifying any rock type, igneous, sedimentary, or metamorphic, is based on *texture* and *composition* (usually mineralogical composition). Textural criteria are commonly considered first, as textures provide the best evidence for rock origin and permit classification into the broadest genetic categories. The first step in igneous rock description should be to determine whether the rock falls into one of the following three categories:

Phaneritic. The crystals that compose the rock are readily visible with the naked eye.

Aphanitic. The crystals, if any, are too small to be seen readily with the naked eye.

Fragmental. The rock is composed of pieces of disaggregated igneous material, deposited and later amalgamated. The fragments themselves may include pieces of pre-existing (predominately igneous) rock, crystal fragments, or glass.

If a rock exhibits phaneritic texture it crystallized slowly beneath the surface of the earth, and is called **plu-tonic, or intrusive.** If it is aphanitic, it crystallized rapidly at the earth's surface, and is called **volcanic,** or **extrusive.** Fragmental rocks are collectively called **pyroclastic.** Some rocks classified as phaneritic and aphanitic are relatively *equigmnular* (of uniform grain size), whereas others exhibit a range of grain sizes because different minerals may experience somewhat different growth rates. The grain size usually varies over only a modest range, and does so somewhat gradually. If, on the other hand, the texture displays two dominant grain sizes that vary by a significant amount, the texture is called **por-phyritic.** The larger crystals, called **phenocrysts,** formed earlier during a period of slower cooling. The finer crystals are referred to as **groundmass.** Whether such rocks are considered plutonic or volcanic is based on the grain size of the *groundmass.* Because the grain size is generally determined by cooling rate, porphyritic rocks commonly result when a magma experiences two distinct phases of cooling. This is most common in, although not limited to, volcanics, in which the phenocrysts form in the slow-cooling magma chamber, and the finer ground-mass forms upon eruption.

2.2 COMPOSITIONAL TERMS

Nearly all igneous rocks are composed principally of silicate minerals, which are most commonly those included in Bowen's Series: quartz, plagioclase, alkali feldspar, muscovite, biotite, hornblende, pyroxene, and olivine. Of these the first four (and any feldspathoids present) are **felsic** minerals (from *feldspar + silica)* and the latter four are **mafic** (from magnesium + *ferric* iron). Generally, felsic refers to the light colored silicates whereas mafic refers to the darker ones. In addition to these principal minerals, there may also be a number of *accessory* minerals, present in very small quantities, usually consisting of apatite, zircon, sphene, epidote, an oxide or sulfide, or a silicate alteration product such as chlorite.

The composition of igneous rocks can be described in a variety of ways. Most geologists agree that the best way to form the compositional basis for a classification of igneous rocks is to use the exact mineral content, as will be described shortly. A number of general descriptive terms, however, are not meant to name specific rocks, but to emphasize some compositional aspect of a rock. Unfortunately many of these terms address similar, but not equivalent, compositional parameters, resulting at times in confusion. For example, the terms in the previous paragraph are commonly applied not only to minerals, but also to the rocks that they compose. **Felsic,** then,

describes a rock composed predominantly of felsic minerals, whereas **mafic** describes a rock with far more mafic minerals. The term **ultramafic** refers to a rock that consists of over 90% mafic minerals. Similar, but not equivalent, terms are **leucocratic,** meaning a light-colored rock, and **melanocratic,** meaning a dark-colored, rock. The first two terms are based on mineral content, whereas the latter ones are based on rock color, but the relationship between these two parameters is obvious. Rocks composed principally of light-colored minerals (felsic), will surely be light-colored themselves (leucocratic). Color, however, is not a very reliable measure of the composition of a rock. Thus terms such as mafic, that are defined by color, yet have clear chemical connotations can be confusing. For example, when plagioclase becomes more calcic than about An_{50} it is commonly dark gray or even black. Smoky quartz is also quite dark. Should these minerals be considered mafic? Most geologists would resist this, as the mnemonic roots of the felsic and mafic terms refer to their chemical composition, even though color may have replaced composition as the common distinguishing feature. A rock composed of 90% dark feldspar would thus be considered both felsic and melanocratic. The color of a rock has been quantified by a value known as the **color index,** which is simply the volume percentage of dark minerals.

Purely chemical terms such as **silicic, magneisan, alkaline,** or **aluminous,** etc. that refer, respectively, to the SiO_2, MgO, (Na_2O + K_2O), and $A1_2O_3$ content of a rock may also be used, particularly when the content of some particular component is unusually high. Silica content is of prime importance, and the term **acidic** is synonymous with silicic. Although based on the outdated concept that silicic acid is the form of silica in solution, even in melts, the term is still in use. The opposite of acidic is **basic,** and the spectrum of silica content in igneous rocks has been subdivided as follows:

acidic	> 66 wt. % SiO_2 52-66
intermediate	wt. % SiO_2 45-52% wt. %
basic	SiO_2 < 45 wt. % SiO_2
ultrabasic	

Because the concept behind "acidic" and "basic" is not accurate, many petrologists consider these terms outdated as well, whereas others consider them useful. Naturally, basic rocks are also mafic, so we see some of the unnecessary complexity involved in simply describing the most general compositional properties of igneous rocks.

Yet a further problem arises when we attempt to refer to the composition of a *melt.* How, for example, do we refer to the magma that, when crystallized, becomes a basalt? Because there are few, if any, minerals in a melt, the mineral-based terms *felsic* and *mafic* are technically not appropriate. Color, as mentioned above is an unreliable compositional measure, and it is notoriously poor for magmas and glasses, as the common occurrence of black obsidian, that is quite silicic, can testify. *Basic* would apply, and it is here that the term may best be used. Many North American geologists consider the terms *mafic* and *felsic* appropriate compositional terms, and prefer to call a dark magma *mafic* rather than basic, which they consider outmoded. British geologists prefer *basic* and *acidic* to refer to magmas, because they are appropriately based on composition and not mineralogy. Others escape the problem altogether by naming the magma after the rock equivalent. The literature is thus full of descriptions of "basaltic magmas," "mafic magmas," and "basic magmas." As a true-blue North American, I tend to use the term "mafic" for both magmas and rocks. We will find some other descriptive chemical terms that are useful in the classification of melts and igneous rocks in Chapter 8.

2.3 THE IUGS CLASSIFICATION

Over time, a number of classification schemes have been applied to igneous rocks, resulting in a plethora of equivalent or overlapping rock names (see Table 2-1 at the end of the chapter for a partial list). In the 1960s and 1970s, the International Union of Geological Sciences (IUGS) formed a Subcommission on the Systematics of Igneous Rocks to attempt to develop a standardized and workable system of igneous rock nomenclature.

2.3.1 Calculations and Plotting

The IUGS system requires that we determine three mineral components and plot the percentages of those components on a triangular diagram to determine the proper name. Figure 2-1 shows how triangular diagrams are used.

In Figure 2-1, the three components are labeled X, Y, and Z. The percentage of X (at the upper apex) is zero along the Y-Z base, and increases progressively to 100% at the X apex. Any horizontal line represents a variation in the Y/Z ratio at a constant value of X. Such lines (at 10% X increments) have been shown on the left diagram. Likewise, lines of constant Y and constant Z have been added. These lines can be used like graph paper to plot a point, and a few of these lines have been labeled. In order to plot a point on a triangular diagram using particular values of X, Y, and Z they must total 100%. If they do not, then they must be **normalized** to 100%. This is accomplished by multiplying each by 100/(X+Y+Z). As an example, point A has the components X = 9.0, Y = 2.6, Z = 1.3. We can normalize these values to 100 by multiplying each by 100/(9.0 + 2.6 + 1.3) = 7.75. That gives

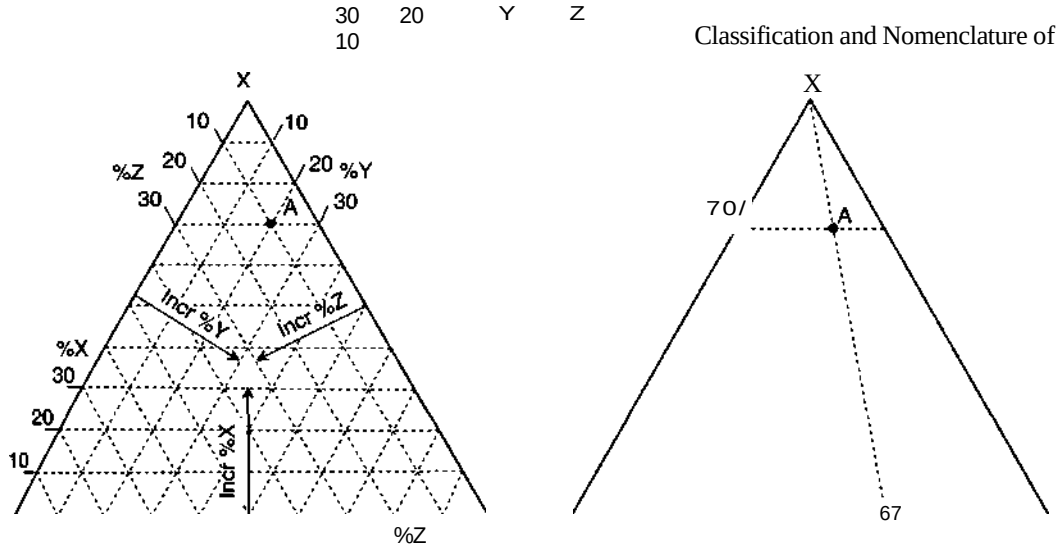


Figure 2-1 Two methods for plotting a point with the components: 70% *X*, 20% *Y*, and 10% Z on triangular diagrams.

the normalized values: X=70%, Y=2Q%, and Z=10%. If we count up 7 lines from the Y-Z base we get a line representing a constant 70% X. Next counting 1 line from the X-Y base toward Z, we get a line representing 10% Z. Their intersection (point A) is also intersected by the line representing 20% Y because the sum must be 100%.

If the proper grid lines are available this technique is simple and direct, however, on the diagrams in Figures 2-2 and 2-3 the lines are not supplied, and for these diagrams an alternative method is used to locate point A, as illustrated on the right diagram in Figure 2-1. Because Y=20% and Z=10%, the ratio 100Y/(Y+Z) = 2000/30 = 67. If we move 67% of the way along the Y-Z base from Z toward Y we have a point with the proper Y/Z ratio of our point A. Any point along a line from this point to the X apex will also have this same Y/Z ratio. If we proceed along such a line to the 70 %X position, we may plot point A at the same position as on the left diagram. Although this method is not as direct for locating a point exactly, it quickly determines the field in which a point falls in the figures that follow.

To classify a rock using the IUGS system, one should use the following procedure:

\. ^texvmne. the **mode** (the percentage of each mineral present, based on volume). The mode is estimated on the basis of the cumulative area of each mineral type as seen on the surface of a hand specimen or in a thin section under the microscope. A more accurate determination is performed by "point counting" a thin section. Point counting involves a mechanical apparatus that moves the section along a two-dimensional grid on the petrographic microscope stage. With each shift, the mineral at the cross-hair of the microscope is identified and counted. When several hundred such points are counted, the counts for each mineral is summed and the totals are normalized to 100% to determine the mode. All of these methods determine relative *areas* of the minerals, but these should correlate directly to volume in most cases.

2. From the mode determine the volume % of each of the following:

   Q' = % quartz

   P' = % plagioclase (An_5-An_{100}) The compositional restriction is meant to avoid confusing the nature of nearly pure albite, which should be considered an alkali feldspar

   A' = % alkali feldspar

   F' = total % feldspathoids ("folds")

   M' = total % mafics and accessories

3. The majority of igneous rocks found at the earth's surface have at least 10% Q'+A'+P' or F'+A'+P'. Because quartz is not compatible with felds pathoids, they will never occur in equilibrium to gether in the same rock. If a rock to be classified has at least this 10% of these constituents, **ignore M** and normalize the remaining 3 parameters to 100% (once again, by multiplying each by 100/(Q'+P'+A') (or 100/(F'+P'+A')). From this we get Q=100Q7(Q'+P'+A'), and similarly for P, A, and F (if appropriate) which sum to 100%. It may seem strange to ignore M, but this is the proce dure. As a result, a rock with 85% mafic minerals

can have the same name as a rock with 3% mafics, if the ratio of P:A:Q is the same.

4. Determine if the rock is phaneritic (plutonic) or aphanitic (volcanic). If it is phaneritic proceed to Figure 2-2. If it is aphanitic use Figure 2-3.
5. To find in which field the rock belongs, first de termine the ratio 100P/(P+A). Select a point along the horizontal P-A line (across the center of the diamond) on Figure 2-2a (or 2-3) that corre sponds to this ratio. Next proceed a distance cor responding to Q or F toward the appropriate apex. Because quartz and feldspathoids can't coexist, there should be no ambiguity as to which trian gular half of the diagram to select. The resulting point, representing the Q:A:P or F:A:P ratio should fall within one of the labeled sub-fields, which provides a name for the rock.
6. If the rock is phaneritic, and Q + A + P + F<10 see Section 2.3.4.

2.3.2 Phaneritic Rocks

Let's try an example. Upon examining a phaneritic rock, we determine that it has the following mode: 18% quartz, 32% plagioclase, 27% orthoclase, 12% biotite, 8% hornblende, and 3% opaques and other accessories. From this we get Q'=18, P'=32, and A'=27. Q'+P'+A' = 77, so, we multiply each by 100/77 to get the normalized values Q = 23, P = 42, and A = 35 that now sum to 100. Because the felsic minerals total over 10%, Figure 2-2a is appropriate. To determine in which field the rock plots we must calculate 100P/(P+A) which is 100(42/(42+35)) = 55. By counting along the P-A axis from A toward P in Figure 2-2a, we find that it plots between the 35 and 65 lines. Then we move upward directly toward point Q. Because 23 falls between 20 and 60, the appropriate name for this rock is "granite."

A rock with 9% nepheline, 70% orthoclase, and 2% plagioclase, with the rest mafics and accessories would be a "nepheline syenite." Try the calculation yourself. The term "foid" is a general term for any feldspathoid. ***Don't use the term "foid" in a rock name.*** Rather, substitute the name of the actual feldspathoid itself. The same applies for "alkali feldspar" in the fields for alkali feldspar granite and alkali feldspar syenite. Use the true feldspar name, if you can determine it, such as "orthoclase granite."

For **rocks that plot near P** a problem arises. Three relatively common rock types all plot near this corner: gabbro, diorite, and anorthosite, and cannot be distinguished on the basis of QAPF ratios alone. Anorthosite has greater than 90% plagioclase in the un-normalized mode, and is thus easily distinguished. Both diorite and gabbro, however, plot in the same field(s), and must be distinguished based on criteria other than QAPF ratios. Regrettably, they are distinguished on the basis of two different parameters. In hand sample, gabbro is defined as having greater than 35% mafic minerals in the mode (usually a pyroxene or olivine) whereas diorite has less than 35% mafics (usually hornblende, but possibly a pyroxene). In thin section, on the other hand, they are distinguished by plagioclase composition, which can be determined by optical properties. The plagioclase in gabbro is more anorthite rich than An_{50}, whereas the An-content of plagioclase in diorite is less than An_{50}. Of course, whenever there is more than one criterion, a rock could be a gabbro by one and a diorite by the other. When confusion results, the IUGS recommends that plagioclase composition has priority. Further complication may arise from compositionally zoned plagioclase (Section 3.1.3). Perhaps the most common method used to distinguish gabbro from diorite in hand specimen is based on color index (% dark minerals). As mentioned above, plagioclase more calcic than An_{50} is commonly black, whereas it is white when more sodic. Thus gabbros are commonly black and diorites usually have a salt-and-pep-per appearance. This method is *not* recommended by the IUGS. As mentioned above, the color index is far from reliable, so is at best a technique for first approximations in the field.

2.3.3 Modifying Terms

It is acceptable under the IUGS system to include min-eralogical, chemical, or textural features in a rock name. The goal here is to impart some descriptive information that you consider important enough to put in the name. This is a matter of judgement, and is flexible. If the rock is unusually light colored for its category, you may want to add the prefix "leuco-," as in "leuco-granite." If it's unusually dark, add the prefix "mela-," as in "mela-gran-ite." This would help distinguish the 85% mafic vs. 3% mafic rock mentioned in step 3 above. You may also use textural terms (see Chapter 3) such as "porphyritic granite," "rapakivi granite," "graphic granite," etc. When naming a rock, *always* try to find the correct name from the proper IUGS diagram. Names such as "pegmatite," "aplite," or "tuff" are incomplete. Rather use these *textural* terms to modify the rock name, as in "pegmatitic orthoclase granite," "aplitic granite," or "rhyolite tuff." If you want to convey some important mineralogical information, that may be added to the name as well. Naturally quartz, plagioclase, and alkali feldspar are already implicit in the name, so are redundant if mentioned specifically. However, you may want to describe a rock as a "riebeckite granite," or a "muscovite biotite granite." If more than one mineral is included, they are listed in the order of *increasing* modal concentration. In the previous example then, there should be more biotite than

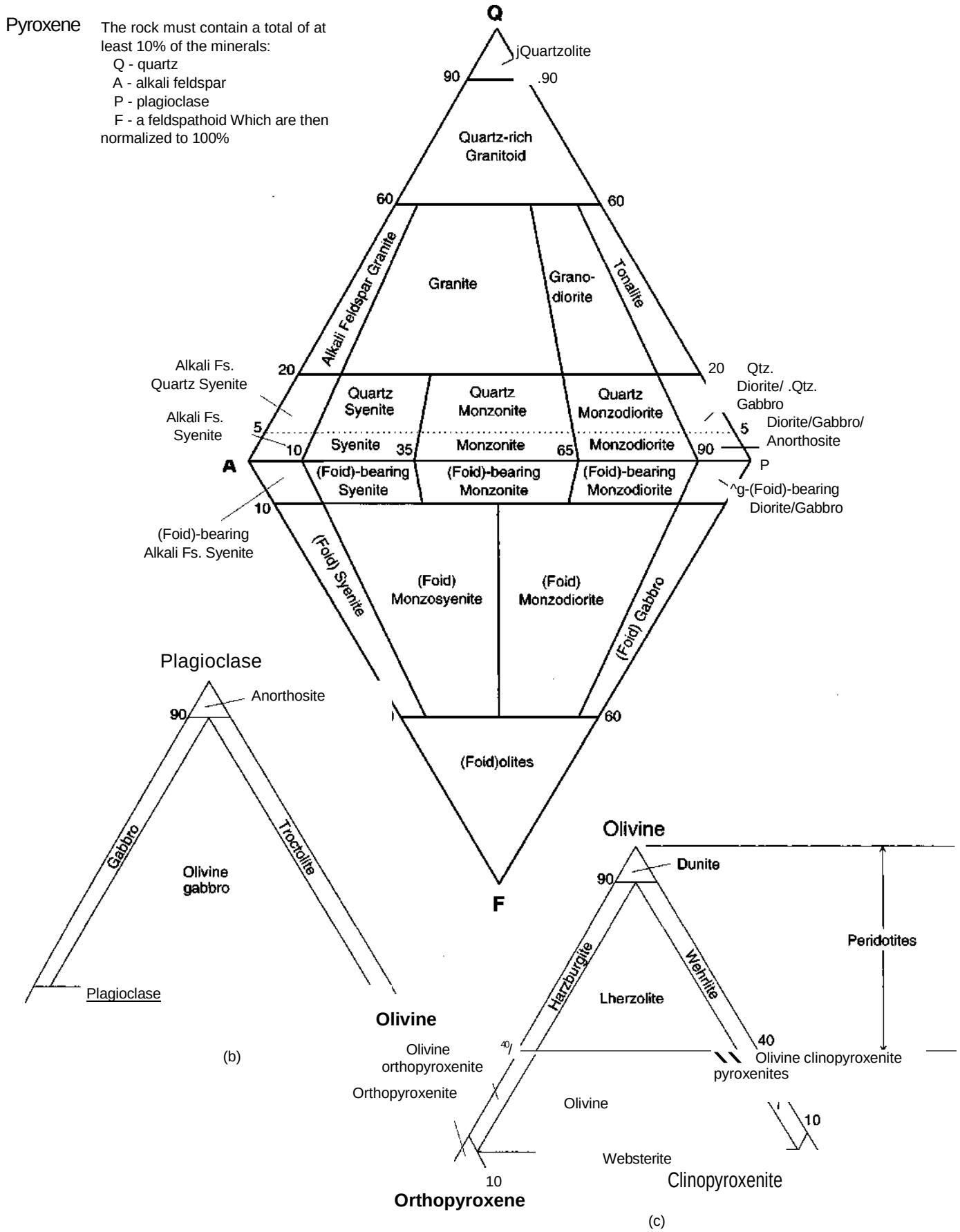


Figure 2-2 A classification of the phaneritic igneous rocks, **(a)** Phaneritic rocks with more than 10% (quartz + feldspar + feldspathoids). **(b)** Gabbroic rocks, (c) Ultramafic rocks. After IUGS (see references at the end of the chapter).

muscovite in the rock. At times it may also be desirable to add a chemical modifier, such as alkaline, calc-alka-line, aluminous, etc. A common example is the use of the prefix "alkali-." High contents of alkalis can stabilize an alkali amphibole, or an alkali pyroxene. We usually don't think of pyroxene-bearing granites, but some "alkali granites" may indeed contain a sodium-rich pyroxene. As we shall see in Chapter 8, some chemical characteristics are manifested throughout a whole series of coge-netic magmas in some igneous provinces. The chemical terms are thus more commonly applied to "suites" of igneous rocks (a group of rocks that are genetically related).

2.3.4 Mafic and Ultramafic Rocks

Gabbroic rocks (plagioclase + mafics) and ultramafic rocks (with over 90% mafics) are classified using separate diagrams (Figure 2-2b and c, respectively). As with any classification, the IUGS subcommittee has had to find a delicate balance between the tendencies for splitting and lumping. The same is true for us. Whereas the TUGS must serve the professional community, and guide terminology for professional communication, we must find a classification suitable for more common use in petrology laboratories.

Figure 2-2b for gabbroic rocks is simplified from the IUGS recommendations. When one can distinguish pyroxenes in a gabbro, there is more specific terminology (for example, an orthopyroxene gabbro is called a "norite"). Figure 2-2c is a more faithful to the IUGS recommendations. In hand specimen work it may be difficult to distinguish ortho- from clinopyroxene in black igneous rocks. Hence the terms "peridotite" and "py-roxenite" are commonly used for ultramafics, because they are independent of pyroxene type. When the distinction can be made, the more specific terms in Figure 2-2c are preferred. The presence of over 5% hornblende further complicates the nomenclature of both mafic and ultramafic rocks. I believe that the IUGS distinction between an "olivine-pyroxene hornblendite," an "olivine-hornblende pyroxenite," and a "pryoxene-hornblende peridotite" adds more detail than necessary at this point. The student is referred to the complete IUGS classification (Streckeisen, 1967,1974; LeMaitre, 1989) for proper names if it becomes important to make more detailed distinctions in nomenclature.

2.4 APHANITIC ROCKS

Volcanic rocks are treated in the same way as plutonics in the original IUGS classification. One determines the mode, normalizes to find P, A, and Q or F, and plots the result in Figure 2-3, in a manner identical to that described for Figure 2-2. Because the mode is commonly difficult to determine accurately for volcanics, Figure 2-3

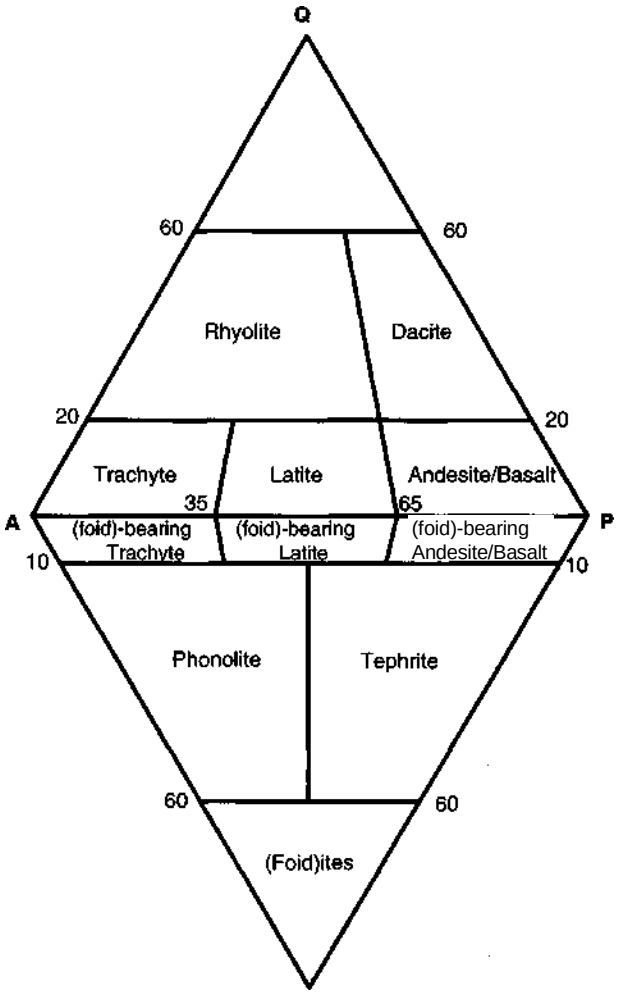


Figure 2-3 A classification and nomenclature of volcanic rocks. After IUGS.

is a simplification, modified from the more detailed diagram published by the IUGS (Streckeisen, 1979). The matrix of many volcanics is composed of minerals of extremely fine grain size, and may even consist of a considerable proportion of **vitreous** (glassy) or amorphous material. Thus it is commonly impossible, even in thin section, to determine a representative mineralogical mode. If it is impossible to recognize the mineralogy of the matrix, a mode must be based on phenocrysts. The IUGS recommends that rocks identified in such a manner be called **phenotypes,** and have the prefix "pheno-" inserted before the name (e.g. "pheno-latite"). As we shall soon see, minerals crystallize from a melt in a sequence (as indicated by, but certainly not restricted to, Bowen's Series), so the first minerals to crystallize do not necessarily represent the mineralogy of the rock as a whole. If based on phenocrysts, the position of a rock on Figure 2-3 will be biased toward the early-forming phases, and usually erroneous for the rock as a whole.

Again **rocks that plot near P** in Figure 2-3 present a problem in the volcanic classification, just as in the plu-tonic one. One cannot distinguish andesite from basalt using Figure 2-3. The IUGS recommends a distinction based on color index or silica content (see below), and not on plagioclase composition. An andesite is defined as a plagioclase-rich rock with either a color index below 35% or with greater than 52% SiO_2. Basalt has a color index greater than 35%, and has less than 52% SiO_2. Many andesites defined on color index or silica content have plagioclases of composition An_{65} or greater.

The most reliable way to avoid the matrix problem discussed above is to analyze the volcanic rock chemically, and use a classification scheme based on the analytical results (as is implicit using % SiO_2 in the IUGS distinction between andesite and basalt discussed above). The IUGS has subsequently recommended a classification of volcanics based on a simple diagram comparing the total alkalis with silica (Le Bas et al., 1986). The diagram (Figure 2-4) requires a chemical analysis and is divided into 15 fields. To use it, we normalize a chemical analysis of a volcanic to a 100% non-volatile basis, add $Na_2O + K_2O$, and plot them against SiO_2. Results are generally consistent with the QAPF diagram when a good mode is available.

The diagrams shown in Figures 2-2 to 2-4 should provide you with the names of most common igneous rocks, but a number of important rock types classified by the IUGS are not included in the figures. For instance, the classification shown does not cover any **hypabyssal** (shallow intrusive) rocks such as **diabase** (or **dolerite** in Britain), nor does it cover the less common rock types such as **carbonatites** (igneous carbonates), **lamproites/ lamprophyres,** (highly alkaline, volatile-rich mafic flow/dike rocks) **spilites** (sodic basalts), or **keratophyres** (sodic intermediate volcanics), etc.

Highly alkaline rocks, particularly those of continental origin, are varied, both mineralogically and chemically. The composition of highly alkaline rocks ranges to high concentrations of several elements present in only trace amounts in more common igneous rocks. The great variety results in a similarly complex nomenclature. Although these alkaline rocks comprise less than one percent of igneous rocks, fully half of the formal igneous rock names apply to them. Such an intricate nomenclature is far beyond the intended scope of this chapter. Chapter 19 will deal with several of the more common or interesting alkaline rock types.

I have tried to avoid the cumbersome detail that a comprehensive classification requires and have attempted to provide a useful compromise between completeness and practicality. Table 2-1 lists a much broader spectrum of igneous rock names that can be found in the literature. Those in **bold** are recommended by the IUGS and can be found either in Figures 2-2 to *2-A,* or in the IUGS publications referenced. The other terms are not recommended by the IUGS, as being too colloquial, too restrictive, inappropriate, or obsolete. I have attempted to provide a *very* brief definition of each term, including the lUGS-ap-proved term that most closely approximates it. It is impossible to do this with precision, however, as the chemical, mineralogical, and/or textural criteria seldom coincide perfectly. Table 2-1 is intended to provide you with a quick reference to rock terms that you may encounter in the literature, and not as a rigorous definition of each. For the latter you are referred to the *AGI*

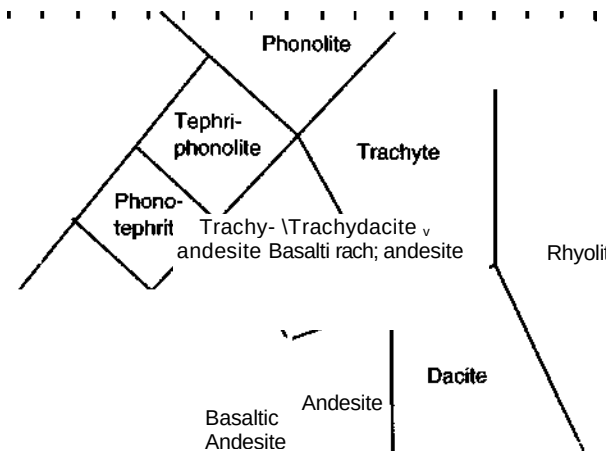


Figure 2-4 A chemical classification of volcanics based on total alkalis vs. silica. After Le Bas et al. (1986). Reprinted by permission of Oxford University Press.

BASIC 52 INTERMEDIATE 63 ACIDIC

: **wt%SiO_2**

:

Table 2-1 Common Rock Names, with IUGS Recommended Terms in Bold Print

Name	Approximate Meaning
Adakite	Mg-rich andesite
Adamellite	Quartz Monzonite
Alaskite	Leuco-(alkali feldspar) granite
Alnoite	Melilite lamprophyre (Tbl. 19-6)
Alvekite	Med.- to fine-grained calcite carbonatite
Andesite	Figs. 2-3 and 2-4
Ankaramite	Olivine basalt
Anorthosite	Fig. 2-2
Aplite	Granitoid with fine sugary texture
Basalt	Figs. 2-3 and 2-4
Basanite	Olivine tephrite (Tbl. 19-1)
Beforsite	Dolomite carbonatite
Benmoreite	Trachyte (Fig. 14-2)
Boninite	Mg-rich andesite
Camptonite	Hornblende lamprophyre (Tbl. 19-6)
Cancalite	Enstatite sanidine phlogopite lamproite (Tbl. 19-5)
Carbonatite	> 50% carbonate
Cedricite	Diopside leucite lamproite (Tbl. 19-5)
Charnockite	Orthopyroxene granite
Comendite	Peralkaline rhyolite
Cortlandite	Pyroxene-olivine hornblendite
Dacite	Figs. 2-3 and 2-4
Diabase	Medium-grained basalt/gabbro
Diorite	Fig. 2-2
Dolerite	Medium-grained basalt/gabbro
Dunite	Fig. 2-2
Enderbite	Hypersthene tonalite
Essexite	Nepheline monzo-gabbro/diorite
Felsite	Microcrystalline granitoid
Fenite	Alkali feldspar-rich metasomatic rock associated with carbonatites
Fergusite	Pseudolucite (foid)ite
Fitzroyite	Leucite-phlog. lamproite (Tbl. 19-5)
(Foid)ite	Fig. 2-3 Replace "foid" with the actual
(Foid)olite	Fig. 2-2 feldspathoid name
Fourchite	Mafic analcime lamprophyre
Fortunite	Glassy olivine lamproite (Tbl. 19-5)
Foyaite	Nepheline syenite
Gabbro	Fig. 2-2
Glimmerite	Biotite-rich ultramafic
Granite	Fig. 2-2
Granitoid	Felsic Plutonic (granite-like) rock ~
Granodiorite	Fig. 2-2
Granophyre	Porphyritic granite with granophyric texture
Grazinite	Phonolitic nephelinite
Grennaite	Nepheline syenite
Harrisite	Troctolite
Harzburgite	**Fig. 2-2**
Hawaiite	Sodic trachybasalt (Fig. 14-2)
Hornblendite	Ultramafic rock >90% hornblende
Hyaloclastite	Pyroclastic rock = angular glass fragments
Icelandite	Al-poor, Fe-rich andesite
Ignimbrite	Applied to welded tuffs
Ijolite	Clinopyroxene nephelinite (Tbl. 19-1)
Italite	Glass-bearing leucititolite
Jacupirangite	Alkaline pyroxenite
Jotunite	Hypersthene monzonorite
Jumillite	Olivine madupitic lamproite (Tbl. 19-5)
Kalsilitite	Kalsilite-rich mafic volcanic (Tbl. 19-1)
Kamafugite	Collective term for the group: KAtungite-MA Furite-Ugandite
Katungite	K-rich olivine melilitite
Keratophyre	Albitized felsic volcanic
Kersantite	Biotite-plag. lamprophyre (Tbl. 19-6)
Kimberlite	Volatile-rich ultramafic (Tbl. 19-1)
Komatiite	Ultramafic volcanic (mostly Archean)
Kugdite	Olivine melilitolite
Ladogalite	Mafic alkali feldspar syenite.
Lamproite	A group of K-, Mg-, volatile-rich volcanics (Tbl. 19-1) A diverse group of dark,
Lamprophyre	porphyritic, mafic to ultramafic K-rich hypabyssal rocks (Tbl. 19-1)
Larvikite	Augite syenite-monzonite Fig. 2-3
Latite	Volcanic rock that is nearly all leucite Fig. 2-2
Leucitite	Volcanic = pyroxene + olivine + opaques in a glassy groundmass
Lherzolite	
Limburgite	Rhyolite
Liparite	Porphyritic granite with tourmaline Lamproite with poikilitic phlogopite ground-mass (Tbl.
Luxulianite	19-5) Alkaline ultramafic volcanic
Madupite	Lamprophyre
	Aegirine-augite nepheline syenite
Mafurite	Leucite-richterite lamproite (Tbl. 19-5)
Malchite	Hypersthene monzonite Mg-rich andesite
Malignite	Natrolite phonolite Picrite basalt
Mamilite	Ultramafic volcanic Ultramafic
Mangerite	melilite-clinopyroxene volcanic (Tbl. 19-1)
Marianite	
Marienbergite	Plutonic melilitite
Masafuerite	Mafic ijolite
Meimechite	Orbicular gabbro
Melilitite	Felsic biotite.-nepheline monzosyenite Biotitic lamprophyre (Tbl. 19-6) Mafic
Melilitolite	type of leucite (foid)ite Feldspar-free
Melteigite	lamprophyre (Tbl. 19-6) Fig. 2-2
Miagite	Orthoclase-bearing hypersthene gabbro
Miaskite	Mafic sodic trachyandesite (Fig. 14-2) Rare
Minette	sodic carbonatite volcanic Ilmenite-apatite
Missourite	dike rock Nepheline-bearing basalt (Tbl.
Monchiquite	19-1) Plutonic nephelinite (a foidite)
Monzonite	Quartz-bearing alkali feldspar syenite
Monzonorite	Orthopyroxene gabbro Volcanic glass
Mugearite	(typically silicic) Picrite
Natrocarbonatite	Lamprophyre Hypersthene
Nelsonite	granodiorite Lamprophyre
Nephelinite	Di-sanidine-phlog. lamproite (Tbl. 19-5)
Nephilinolite	Volcanic near the tephri-phonolite - phono-
Nordmarkite	tephrite border (Fig. 2-4) Ultramafic
Norite	lamprophyre Peralkaline rhyolite (Tbl.
Obsidian	19-1) Very coarse-grained igneous rock
Oceanite	Fig. 2-2 Volcanic glass that exhibite fine
Odinite	concentric
Opdalite	cracking
Orbite	
Orendite	
Orvietite	
Oachitite	
Pantellerite	
Pegmatite	
Peridotite	
Perlite	

Continued

Name	Approximate Meaning	Name	Approximate Meaning
Phonolite	Figs. 2-3 and 2-4	Tachylite	Basaltic glass
Phono-tephrite	**Fig. 2-3**	Tahitite	Haiiyne tephri-phonolite
Picrite	Olivine-rich basalt (Fig. 14-2)	**Tephrite**	Fig. 2-3 and Tbl. 19-1
Picrobasalt	Fig. 2-4	**Tephri-phonolite**	Fig. 2-3
Pitchstone	Hydrous volcanic glass	**Teschenite**	Analcime gabbro
Plagioclasite	Anorthosite	**Theralite**	Nepheline gabbro
Plagiogranite	Leucotonalite	**Tholeiite**	Tholeiitic basalt or magma series (Sec. 8.7)
Polzenite	Melilite lamprophyre (Tbl. 19-6)	**Tonalite**	Fig. 2-2
Pyroxenite	Fig. 2-2	**Trachyte**	**Figs. 2-3 and 2-4**
Quartzolite	Fig. 2-2	**Troctolite**	Olivine-rich gabbro (Fig. 2-2)
Rauhaugite	Coarse dolomitic carbonatite (Tbl. 19-2)	**Trohdhjemite**	Leuco-tonalite
Reticulite	Broken-up pumice	**Uncompahgrite**	Pyroxene melilitolite
Rhyolite	Figs. 2-3 and 2-4.	**Urtite**	Felsic nephilinolite (Tbl. 19-1)
Sannaite	Na-amph-augite lamprophyre (Tbl. 19-6)	**Verite**	Glassy ol-di-phlog. lamproite (Tbl. 19-5)
Sanukite	Mg-rich andesite	Vicolite	See orvietite
Scoria	Highly vesiculated basalt	**Vogesite**	Hbl-di-orthocl. lamprophyre (Tbl. 19-6)
Shonkinite	Alkaline plutonic with Kfs, foid, and augite	**Websterite**	Fig. 2-2
Shoshonite	K-rich basalt (Tbl. 19-1)	**Wehrlite**	Fig. 2-2
Sovite	Coarse calcite carbonatite (Tbl. 19-2)	Woldigite	Di-leucite-richterite madupidic lamproite (Table 19-5)
Spessartite	Hbl-di-plag. lamprophyre (Tbl. 19-6)		
Spilite	Altered albitized basalt	Wyomingite	Di-leucite-phlog. lamproite (Tbl. 19-5)
Syenite	Fig. 2-2		

Glossary or LeMaitre (1989). For the sake of brevity, I have included only the root IUGS terms, such as *granite, andesite,* or *trachyte,* and not compound terms such as *alkali-feldspar granite,* or *basaltic trachy-andesite,* etc.

2.5 PYROCLASTIC ROCKS

The initial IUGS classification did not cover pyroclastic rocks, but they were covered in a later installment (Schmid, 1981). As mentioned previously, if the chemical composition is available these rocks could be classified compositionally in the same manner as any volcanics, but they commonly contain significant impurities, and only those for which the foreign material is minimal can a name be applied with confidence. Pyroclastics are thus commonly classified on the basis of the *type* of fragmen-tal material (collectively called pyroclasts) or on the *size* of the fragments (in addition to a chemical or modal name if possible). If the type of material is to be the basis, the volume percent of glass, rock fragments, and crystal fragments is determined and the name derived from Fig-

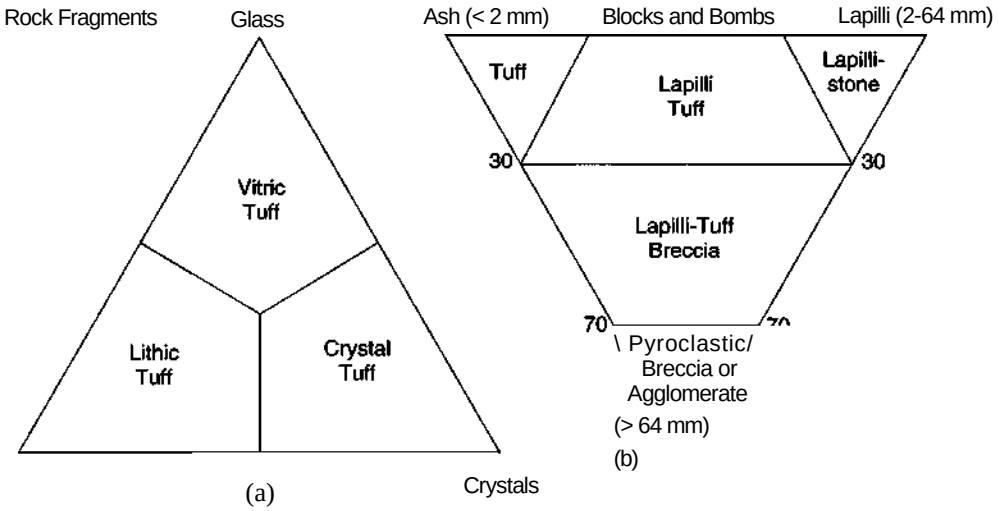


Figure 2-5 Classification of the pyroclastic rocks, (a) Based on type of material. After Pettijohn (1975) and Schmid (1981). (b) Based on the size of the material. After Fisher (1966). Copyright © with permission from Elsevier Science.

ure 2-5a. If the fragment size is the basis, determine the percentage of the fragments that fall into each of the following categories:

> 64 mm diameter
Bombs (if molten during fragmentation)
Blocks (if not)
2-64 mm Lapilli < 2 mm
Ash

The results are then plotted on Figure 2-5b to determine the name. Agglomerate in Figure 2-5b can be used for coherent or incoherent material, whereas most geologists restrict the term pyroclastic breccia to coherent material. Figures 2-5a and b are not necessarily mutually exclusive. It is acceptable to name a rock a "rhyolitic vitric lapilli tuff" if you want to emphasize that the lapilli are mostly composed of glassy fragments. An **aquagene tuff** is a waterborne accumulation of ash. It may result from a sub-aqueous eruption or it may be an airborne accumulation that has been reworked by water. **Hyalo-clastite** is an aquagene tuff that is created when magma is shattered when it comes into contact with water.

SUGGESTED FURTHER READINGS

The following are IUGS publications:

Le Bas, M. J., R. W. LeMaitre, A. L. Streckeisen, and B. Zanet-tin (1986). A chemical classification of volcanic rocks based on the total alkali-silica diagram. /. *Petrol*, 27, 745-750.

LeMaitre, R. W., ed. (1989). *A Classification of Igneous Rocks and a Glossary of Terms.* Blackwell. Oxford.

Streckeisen, A. L. (1967) Classification and nomenclature of igneous rocks. *Neues Jahrbuch fur Mineralogie Abhand-lungen.* **107,**144-240.

Streckeisen, A. L. (1973). Plutonic rocks, classification and nomenclature recommended by the IUGS subcommission on the systematics of igneous rocks. *Geotimes*, **18,**26-30.

Streckeisen, A. L. (1974). Classification and Nomenclature of Plutonic Rocks. *Geol. Rundschau,* **63,** 773-786.

Streckeisen, A. L. (1976). To each plutonic rock its proper name. *Earth-Sci. Rev.,* **12,**1-33.

Streckeisen, A. L. (1979). Classification and nomenclature of volcanic rocks, lamprophyres, carbonatites, and melilitic rocks: recommendations and suggestions of the IUGS sub-commission on the systematics of igneous rocks. *Geology*, 7, 331-335.

Schmid, R. (1981). Descriptive nomenclature and classification of pyroclastic deposits and fragments, recommendations and suggestions of the IUGS subcommission on the systematics of igneous rocks. Geology, 9, 4-43.

PROBLEM

1. Using the appropriate diagram, name the following rocks:

		Modal percentages								
Number	Volcanic or Plutonic	Quartz	Plagioclase	Orthoclase	Nepheline	Olivine	Orthopyroxene	Clinoyroxene	Hornblende	Others
1	P	20	30	30					10	10
2	P	10	15	15					40	20
3	P	18	55						12	15
4	P		32	37		6			25	
5	P	13	6	68				5	6	2
6	P		42	31		12			2	13
7	P	2	62	1			6	5		6
8	P					63		18		3
9	P					67		4		2
10	P		92				7			1
11	P					7		55		2
12	V	5	73	4					12	6
13	V	4	24	63				4	3	2
14	V		35	37		7			12	9
15	V		55			7		36		2

Chapter 3

Textures of Igneous Rocks

In the previous chapter, we learned how to categorize and name an igneous rock. It is also useful to include a rock's texture in a good description of hand samples or thin sections. But textures are much more important than mere descriptive aids. The texture of a rock is a result of various processes that controlled the rock's genesis and, along with mineralogy and chemical composition, provides the information that we use to interpret the rock's origin and history. It is thus important for us to be able to recognize and describe the textures of a rock, and to understand how they are developed. In Chapter 1, for example, I mentioned that **interlocking** texture (Figure 1-1) was produced by crystallization from a melt and can be used to determine the igneous origin of a rock. In this chapter, we will explore igneous textures in more detail, seeking to discover what controls those textures so that we can use textural criteria to aid us as we attempt to understand the crystallization history (and perhaps also some of the post-crystallization history) of a particular rock. I have supplied a glossary of textural terms at the end of the chapter where you will find the definition of foreign terms that you encounter. In some of the references listed at the end of the chapter you will also find excellent color photographs and line drawings (based mostly on thin sections) that further illustrate many of the textures.

The textures that you observe in an igneous rock result from a number of processes that can be grouped into two principal categories. **Primary** textures occur during igneous crystallization, and result from interactions between minerals and melt. **Secondary** textures are alterations that take place after the rock is completely solid. The following is a very general discourse on how a number of the more common textures develop. I will concentrate on thin section study, but many of the textures described can be recognized in hand specimens also. **Petrography** is the branch of petrology that deals with the description and classification of rocks, but most of us use the term to connote the study of rocks in thin section.

3.1 PRIMARY TEXTURES (CRYSTAL/MELT INTERACTIONS)

The formation and growth of crystals, either from a melt or in a solid medium (metamorphic mineral growth) involves three principal processes: 1) initial *nudeation* of the crystal, 2) subsequent crystal *growth,* and 3) *diffusion* of chemical species (and heat) through the surrounding medium to and from the surface of a growing crystal.

Nucleation is a critical initial step in the development of a crystal. Very tiny initial crystals have a high ratio of surface area to volume, and thus a large proportion of ions at the surface. Surface ions have unbalanced charges, because they lack the complete surrounding lattice that balances the charge of interior ions. The result is a high surface energy for the initial crystal, and therefore a low stability. The clustering of a few compatible ions in a cooling melt will thus tend to spontaneously separate, even at the saturation temperature when conditions are otherwise suitable for crystallization of a particular mineral. Under such conditions, crystallization would be possible, but the prerequisite nucleation isn't. Before crystallization can take place, a critical size of an "embryonic cluster" or "crystal nucleus" must form. This usually requires some degree of supersaturation or **undercooling** (cooling of a melt below the true crystallization temperature of a mineral) before enough ions to be stable can cluster together. Alternatively, a preexisting crystal surface may be present: either a "seed crystal" of the same mineral, or a different mineral with a similar structure on which the new mineral can easily nucleate and grow.

Several studies have indicated that crystals with simple structures tend to nucleate more easily than those with more complex structures. Oxides (magnetite or il-menite) or olivine, for example, generally nucleate more easily (with less undercooling) than does plagioclase with its complex Si-O polymerization.

Crystal growth involves the addition of ions onto existing crystals or crystal nuclei. In a simple structure with high symmetry, faces with a high density of lattice points ({100}, {110}) tend to form more prominent faces. In more complex silicates, this tendency may be superceded by preferred growth in directions with uninterrupted chains of strong bonds. Pyroxenes and amphiboles thus tend to be elongated in the direction of the Si-O-Si-O chains, and micas tend to elongate in the directions of the silicate sheets. Defects, such as screw dislocations, may also aid the addition of new ions to a growing face, and impurities may inhibit growth in some directions. In general, faces with low surface energy become more prevalent. When low-energy faces predominate over high-energy faces, the overall energy of the system is lower, and hence more stable. The surface energy on different faces may vary disproportionately with changing conditions, so the shape of a particular mineral may vary from one rock to another. Cashman (1990) also reminded petrologists of what metallurgists have long known: As the degree of undercooling increases, minerals change from well-faceted crystals to acicular habits, then to dendritic, and finally to spherulitic forms.

In most situations the composition of a growing crystal differs considerably from that of the melt. Only in simple chemical systems, such as water-ice, is this not true. In the general case, then, the growth of a mineral will gradually deplete the adjacent melt in the constituents that the mineral preferentially incorporates. For growth to proceed, new material must **diffuse** through the melt, cross the depleted zone, and reach the crystal surface. In addition, the formation of a crystal from a melt produces heat (the latent heat of crystallization, which is merely the opposite of the latent heat of fusion, see Section 6.3). This heat must also be able to diffuse away from the crystal, or else the temperature at the growing surface may become too high for crystallization to proceed.

3.1.1 Rates of Nucleation, Growth, and Diffusion

Because there are three main processes involved in mineral development, and not just one, their relative rates have considerable influence on the ultimate texture of the resulting rock. We shall see that, like the weak link in a chain, whichever rate is the *slowest* will be the overall rate-determining process, and exert the most control on crystallization. There is a further rate that we must also address: the **cooling rate** of the magma. If the cooling rate is very slow, equilibrium is maintained or closely approximated. If the cooling rate is high, significant undercooling can result, because there is seldom time for nucleation, growth, or diffusion to keep pace. The cooling rate is an important externally controlled variable, which influences the rates of the other crystal-forming processes. Much of the textural information that we observe is thus used to interpret the cooling rate of a rock.

The rates of both nucleation and crystal growth are strongly dependent on the extent of undercooling of the magma. Initially, undercooling enhances both rates, but further cooling decreases kinetics and increases viscosity, thus inhibiting the rates. As illustrated in Figure 3-1, the maximum growth rate is generally at a higher temperature than is the maximum nucleation rate, because it is easier to add an atom with high kinetic energy onto an existing crystal lattice than to have a chance encounter of several such atoms at once to form an embryonic cluster. Further undercooling inhibits growth, because atoms have to diffuse farther to add onto a few existing crystals, and it is easier for the slowed atoms to nucleate in local clusters than to move far.

We can use Figure 3-1 to understand why the rate of cooling so profoundly affects the grain size of a rock. As it relates to Figure 3-1, undercooling is the degree to which temperature falls below the melting point before crystallization occurs. For example, if the cooling rate is low, only slight undercooling will be possible (such as at temperature T_{fl} in Figure 3-1). At this temperature, the nucleation rate is very low and the growth rate is high. Fewer crystals thus form and they grow larger, resulting in the coarse-grained texture common among slow-cooled plutonic rocks. Quickly cooled rocks, on the other hand, may become significantly undercooled before crystallization begins. If rocks are undercooled to T_b in Fig-

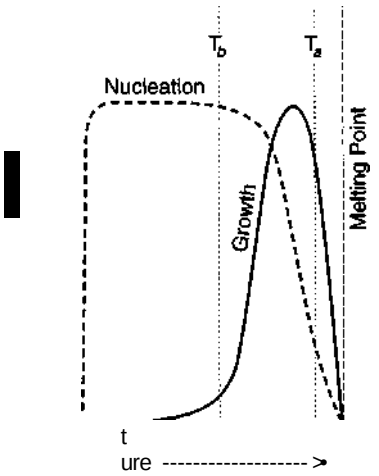


Figure 3-1 Idealized rates of crystal nucleation and growth as a function of temperature below the melting point. Slow cooling results in only minor undercooling (T_a), so that rapid growth and slow nucleation produce fewer coarse-grained crystals. Rapid cooling permits more undercooling *(T_b)*, so that slower growth and rapid nucleation produce many fine-grained crystals. Very rapid cooling involves little if any nucleation or growth(r_c) producing a glass.

ure 3-1, the nucleation rate exceeds the growth rate, and many small crystals are formed, resulting in the very fine-grained texture of volcanic rocks. Very high degrees of undercooling (T_c in Figure 3-1) may result in negligible rates of nucleation and growth, such that the liquid solidifies to a glass with very few crystals or none at all.

Two-stage cooling can create a bimodal distribution of grain sizes. Slow cooling followed by rapid is the only plausible sequence, and might occur when crystallization began in a magma chamber, followed by the opening of a conduit and migration of magma to the surface. Initially, the magma would be only slightly undercooled and a few coarse crystals would form, followed by volcanism and finer crystals. When there is a distinctly bimodal distribution in grain size, with one size considerably larger than the other, the texture is called **porphyritic.** The larger crystals are called **phenocrysts,** and the finer surrounding ones are called **matrix** or **groundmass.** A porphyritic rock is considered plutonic or volcanic on the basis of the *matrix* grain size. If the phenocrysts are set in a glassy groundmass, the texture is called **vitrophyric.** If the phenocrysts contain numerous inclusions of another mineral that they enveloped as they grew, the texture is called **poikilitic.** The host crystal may then be called an **oikocryst.**

The growth rate of a crystal depends upon the surface energy of the faces and the diffusion rate. For a constant cooling rate, the largest crystals will usually be those with the simplest structures (they also nucleated earliest) and/or the most plentiful or fastest diffusing components. The diffusion rate of a chemical species is faster at higher temperature and in material with low viscosity. Diffusion rate is thus low in highly polymerized viscous melts (such melts are generally silica-rich and also tend to be cooler than mafic melts). Small ions with low charges diffuse best, whereas large polymerized complexes diffuse poorly. In general, diffusion in a fluid is better than in a glass, and better in glass than in crystalline solids. H_2O dramatically lowers the degree of polymerization of magma (Chapter 7), thereby enhancing diffusion. Alkalis have a similar effect, although less extreme. The very coarse grain size of many pegmatites may be attributed more to the high mobility of species in the H_2O-rich melt from which they crystallize than to extremely slow cooling.

The rates of nucleation and growth vary with the surface energy of the minerals and faces involved, the degree of undercooling, and the crystal structure. These values can be different for different minerals, even in the same magma. Different minerals can be undercooled to differing extents because the melting point in Figure 3-1 is specific to each mineral. We shall see in Chapters 6 and 7 that minerals develop sequentially in a cooling magma as the melting point of each is progressively reached. The temperature may thus be under the melting point of one mineral (undercooled) and above that of another. Many stable nuclei of one mineral may thus form, while only a few of another, resulting in many small crystals of the former, and fewer, larger crystals of the latter. The popular notion that the large crystals in a porphyritic rock must have formed first or in a slower-cooling environment is thus not universally valid. The sudden loss of a H_2O-rich fluid phase will quickly raise the crystallization temperature (Section 7.5) and can also produce porphyritic texture in some plutonic rocks, as we shall see in Chapter 11.

When the diffusion rate is not the limiting (slowest) rate, crystals growing free and unencumbered in a melt will tend to be euhedral and nicely faceted. When the rate of diffusion is slower than the rate of growth (as in quickly cooled, or "quenched," lavas) the crystals take on an increasingly radiating form, or a tree-like, branching form termed **dendritic** (Figure 3-2). When diffusion is slower than growth, a zone of depleted liquid builds up at the crystal/liquid interface as described above. Some workers propose that the crystals reach out in thin tendrils beyond the zone to tap a supply of appropriate elements or cooler melt. Others suggest that the perturbations in the surface shape toward dendritic forms help to eliminate the local heat build-up that accompanies crystallization. Perhaps both processes contribute to dendritic or spherulitic growth. Ultramafic lavas, such as Pre-cambrian komatiites, when quenched, may develop spectacular elongated olivine crystals, in some cases up to a meter long, called **spinifex** texture. The unusual size may be caused by rapid growth of the simple olivine structure in a very low viscosity magma, not by slow cooling. Spinifex pyroxenes over 5 centimeters long have also been described.

Crystal corners and edges have a larger volume of nearby liquid to tap for components (or to dissipate the heat of crystallization) than do crystal faces (Figure 3-3). In addition, corners and edges have a higher proportion of unsatisfied bonds. Thus we might expect the corners and edges to grow more rapidly than the faces in such quench situations. The resulting forms are called **skeletal** crystals. In some cases, the extended corners may meet to enclose melt pockets at the recessed faces (Figure 3-4a). The corners of quenched plagioclase tend to grow straighter, creating a characteristic **swallow-tailed** shape (Figures 3.2a and 3.4b). Of course any *motion* of the liquid or crystals rehomogenizes it and reduces the limiting effects of slow diffusion.

3.1.2 Nucleation at Preferred Sites

Epitaxis is the general term used to describe the preferred nucleation of one mineral on another preexisting

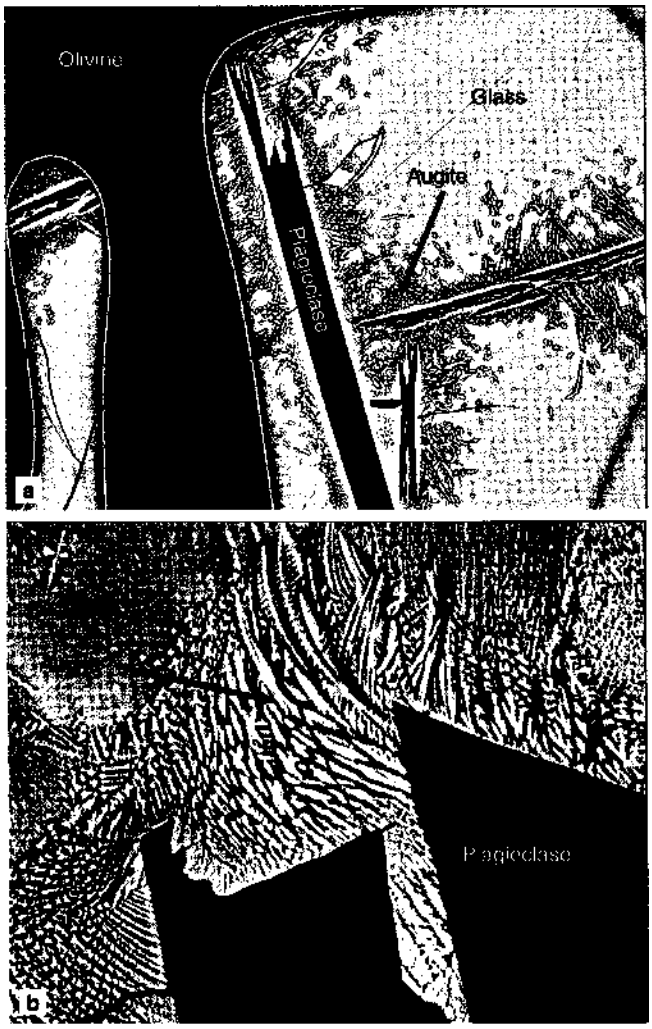


Figure 3-2 Backscattered electron image of quenched "blue glassy pahoehoe," 1996 Kalapana flow, Hawaii. Brightness is proportional to back-scattering ability, and hence to mean atomic number. Black minerals are felsic plagioclase and gray ones are mafics. **(a)** Large embayed olivine phenocryst with smaller plagioclase laths and clusters of feathery augite nucleating on plagioclase. Magnification ca. 400 x. (b) ca. 2000 x magnification of feathery quenched augite crystals nucleating on plagioclase (black) and growing in a dendritic form outward. Augite nucleates on plagioclase rather than pre-existing augite phenocrysts, perhaps due to local enrichment in mafic components as plagioclase depletes the adjacent liquid in Ca, Al, and Si. Photographs taken with the assistance of Jack Rice and Michael Schaeffer, University of Oregon. Photographs taken with the assistance of Jack Rice and Michael Schaeffer, University of Oregon.

mineral, thereby avoiding problems associated with slow nucleation. Similarity of the crystal structures of the mineral substrate and the new phase is a prerequisite for epitaxial growth. The atomic constituents of the new mineral thereby find favorable spots to accumulate and a stable nucleus forms. The growth of sillimanite on biotite or

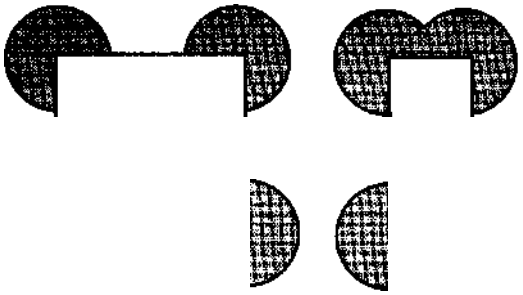


Figure 3-3 **(a)** Volume of liquid (shaded) available to an edge or corner of a crystal is greater than for a side, **(b)** Volume of liquid available to the narrow end of a slender crystal is even greater. After Shelley (1993). Copyright © by permission Kluwer Academic Publishers.

muscovite in metamorphic rocks, rather than as a direct replacement of available crystals of its polymorph kyan-ite, is a common example. The Si-Al-O structures in both sillimanite and mica are similar in geometry and bond lengths, so that sillimanite tends to form in areas of mica concentration. **Rapakivi** texture, involving albitic plagioclase overgrowths on orthoclase, occurs in some granites where the plagioclase preferentially forms on the structurally similar alkali feldspar, rather than nucleating on their own. A crystal nucleus may also form epitaxi-ally in a twin orientation on a preexisting grain of the same mineral, leading to the formation of growth twins. **Spherulitic** texture in silicic volcanics is a texture in which needles of quartz and alkali feldspar grow radially from a common center. This and **variolitic** texture of radiating plagioclase laths in some basalts are probably the result of nucleation of later crystals on the first nuclei to form. Both are considered to form during the devitrification of glass, which will be discussed below under sec-

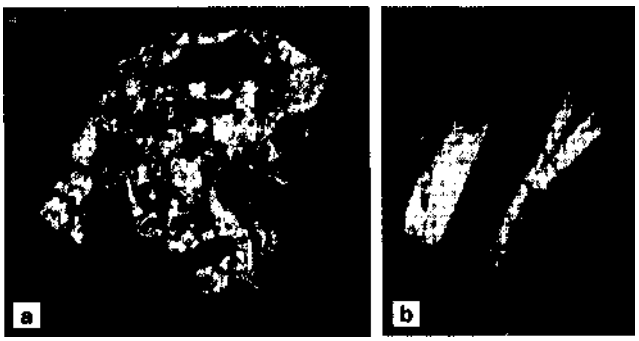


Figure 3-4 (a) Skeletal olivine phenocryst with rapid growth at edges enveloping melt at ends. Taupo, N.Z. **(b)** "Swallow-tail" plagioclase in trachyte, Remarkable Dike, N.Z. Length of both fields ca. 0.2 mm. From Shelley (1993). Copyright © by permission Kluwer Academic Publishers.

ondary textures. Nucleation of minerals on dike (or even vesicle) walls is also a common phenomenon. Growth of elongated crystals (generally quartz), with c-axes normal to vein walls, results in a structure called **comb structure,** because the parallel columns resemble the teeth of a comb. **Crescumulate** texture is similar, and describes the parallel growth of elongated, non-equilibrium arrangements of olivines, pyroxenes, feldspars, or quartz that appear to nucleate on a wall or layer and may grow up to several centimeters long. Crescumulate texture commonly occurs in layered mafic plutons (where it may appear in multiple layers) and in the margins of granites.

3.1.3 Compositional Zoning

Compositional **zoning** is a common phenomenon and occurs when a mineral changes composition as it grows during cooling. The composition of most solid-solution minerals in equilibrium with other minerals or liquid is temperature-dependent. The reasons for this will become clear in Chapters 5, 6, 7, and 27. Compositional zoning can only be observed petrographically when the color (Figure 3-5 a) or extinction position varies with composition. In the case of plagioclase, the extinction angle is highly composition-dependent, and the compositional variations show up as concentric bands of varying brightness in cross-polarized light (Figure 3-5b). If equilibrium between the crystal and the melt is maintained, the composition of the mineral will adjust to the lowering temperature, producing a compositionally homogeneous crystal. Chemical zoning, on the other hand, occurs when equilibrium is *not* maintained and a rim of the new composition is added around the old. Compositional re-equilibration in plagioclase requires Si-Al exchange, and this is difficult due to the strength of the Si-O and Al-O bonds. Diffusion of Al is also slow. Zoning in plagioclase is therefore very common.

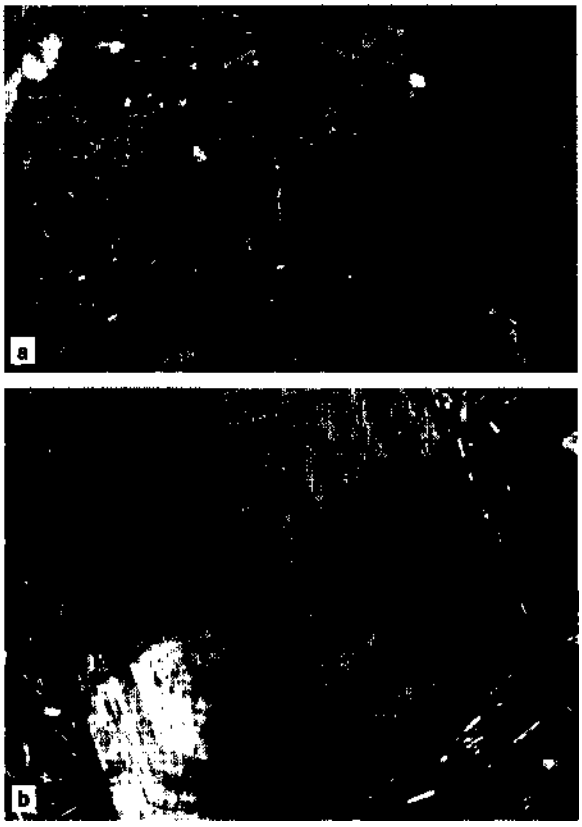


Figure 3-5 **(a)** Compositionally zoned hornblende phenocryst with pronounced color variation visible in plane-polarized light. Field width 1 mm. **(b)** Zoned plagioclase twinned on the carls-bad law. Andesite, Crater Lake, OR. Field width 0.3 mm.

As Figure 6-8 shows, the composition of plagioclase in equilibrium with a melt becomes more Na-rich as temperature drops. The expected zonation in cooling igneous plagioclase would thus be from a more anorthite-rich core to a more albite-rich rim. This type of zoning is called **normal zoning.** It is common in igneous rocks, although it is typically interrupted by reversals. **Reverse zoning** is the opposite of normal zoning, with more sodic inner and calcic outer zones. It is common in some meta-morphic plagioclase, where growth is accompanied by rising temperature. Reverse zoning is rarely a long-term trend in igneous plagioclase; rather it is typically a short-term event where it contributes to localized reversals as a component of oscillatory zoning. **Oscillatory zoning** is the most common type of zoning in plagioclase, because a regular decrease in An-content rarely dominates the full crystallization period. The sample illustrated in Figure 3-5b is oscillatory, and Figure 3-6 shows some typical oscillatory zoning profiles, extending from plagioclase cores to rims. Abrupt changes in zoning, such as the reversal in Figure 3-6a, require abrupt changes in the conditions of the magma chamber. Most petrologists favor the injection of hotter, more juvenile magma into a cooling and crystallizing chamber to effect this change. The common occurrence of corroded or remelted embay-ments of the crystal rim accompanying many reversals supports this conclusion. The more gradual oscillations illustrated in Figure 3-6b are more likely to result from diffusion-dependent depletion and re-enrichment of the liquid zone adjacent to the growing crystal in an undisturbed magma chamber. The depleted components could be the anorthite molecule, or a constituent such as H_2O that can lower the melting point and thus shift the equilibrium composition of the plagioclase (Loomis, 1982).

Most other minerals are not as conspicuously zoned as plagioclase. This may be because the zoning is simply less obvious in thin section, because it may not affect the color or extinction. Most minerals apparently maintain equilibrium with the melt because the exchange of ions does not involve disruption of the strong Si-Al-O bonds. Fe-Mg exchange is also easier because these elements

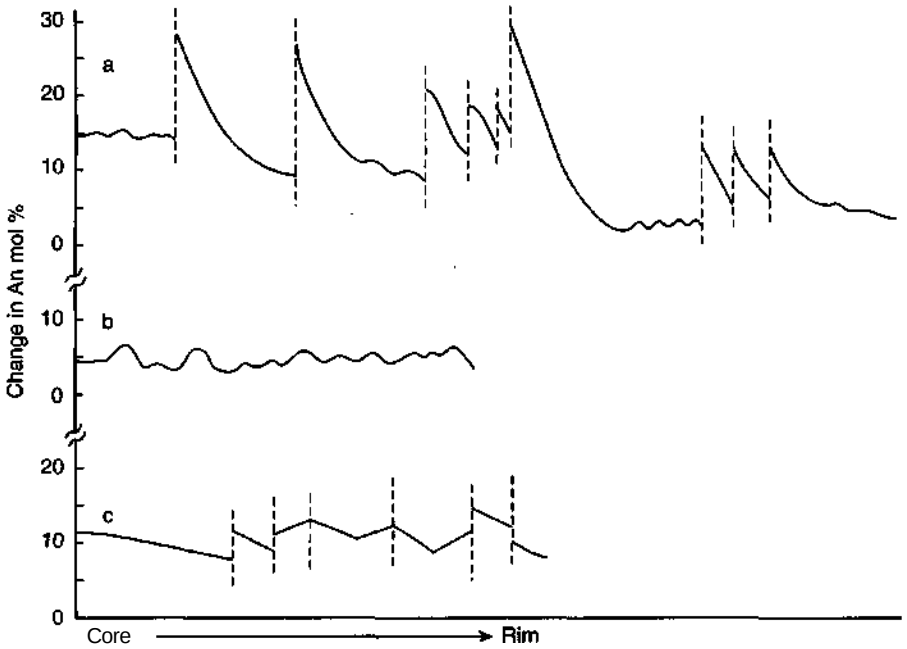


Figure 3-6 Schematic examples of plagioclase zoning profiles determined by microprobe point traverses, (a) Repeated sharp reversals attributed to magma mixing, followed by normal cooling increments, **(b)** Smaller and irregular oscillations caused by local disequilibrium crystallization, (c) Complex oscillations due to combinations of magma mixing and local disequilibrium. From Shelley (1993). Copyright © by permission Kluwer Academic Publishers.

diffuse more readily than Al-Si. Microprobe analysis, however, reveals chemical zonation in several igneous and metamorphic minerals. A color element map of garnet (courtesy of Jack Rice, University of Oregon) is shown on the back cover of the text.

3.1.4 Crystallization Sequence

As a rule, early-forming minerals in melts that are not significantly undercooled are surrounded completely by liquid and develop as **euhedral** crystals, bounded on all sides by crystal faces. As more crystals begin to form and fill the magma chamber, crystals will inevitably come into contact with one another. The resulting mutual interference impedes the development of crystal faces and **sub-hedral or anhedral** crystals form. In some cases, one can infer the sequence of mineral crystallization from these interferences. Early minerals tend to have better forms and the latest ones are **interstitial,** filling the spaces between the earlier ones (Figure 3-7). Euhedral minerals are commonly phenocrysts in an aphanitic groundmass, and thus clearly formed early in the sequence. Some com-positionally zoned minerals may show euhedral cores that formed when the crystals were suspended in melt and anhedral rims that formed later when the crystals were crowded together (see Figure 3-14c).

Unfortunately the simple principle that a crystal that molds itself to conform to the shape of another must have crystallized later is not as reliable as we might wish. Whether or not a crystal grows with well-developed faces depends largely upon the surface energy of the faces. Minerals with very low surface energy may form euhedral crystals even in metamorphic rocks where all growing crystals are necessarily in contact with neighboring grains. Garnet and staurolite, for example, are nearly always euhedral in pelitic schists. Igneous accessory minerals such as zircon, apatite and titanite likewise tend to be euhedral, even though they commonly form during the later stages of crystallization. Metamorphic petrolo-gists have noted a tendency toward euhedralism that diminishes in the order of increasing Si-O polymerization. Olivine and pyroxenes thus tend to be more euhedral than quartz or feldspar. Furthermore, Flood and Vernon (1988) noted that, although there is certainly a sequence of mineral formation, there is considerable overlap, and

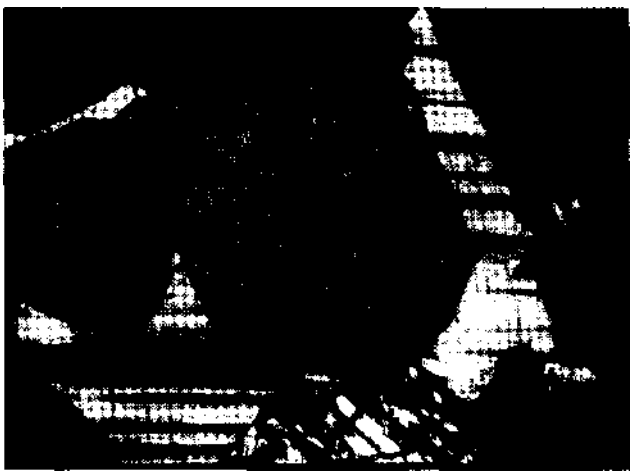


Figure 3-7 Euhedral early pyroxene with late interstitial plagioclase (horizontal twins). Still water complex, Montana. Field width 5 mm.

most crystallization takes place via simultaneous crystallization of several mineral types. We shall see why this is true in Chapters 6 and *1*. The molding relationships thus develop after most, if not all, of the minerals have begun to crystallize. When growth is simultaneous, the relative surface energy of mutually impinging minerals may have considerable influence on which mineral develops good crystal faces and which molds to the other. Hunter *(1981)* demonstrated that, although crystals suspended in melt tend to form euhedral grains, once they touch each other they are likely to dissolve at areas of high surface curvature (interfacial edges) and crystallize at areas of low curvature, thus becoming more rounded. Except for minerals with very low surface energy, then, euhedral crystals should be rare in cases of simultaneous crystallization.

Geologists have often appealed to grain size as another indicator of crystallization sequence. In porphyritic volcanics, the large phenocrysts are generally considered to have formed before the groundmass phases. Although this is commonly true, grain size is dependent on nucle-ation and growth rates, as discussed above, and some of the minerals in the groundmass may have formed early, with a faster nucleation and slower growth rate than the phenocrysts. The large, euhedral, K-feldspar megacrysts found in many granitic rocks, for example, are commonly believed to form *late* in the crystallization sequence of those rocks, not early.

Another sequence indicator is based on inclusion relationships. Igneous inclusions should have formed at an earlier stage than the host that enveloped them. One must be aware, however, that a thin section is a two-dimensional slice through a three-dimensional rock, and a mineral that may *appear to be* surrounded by another could be jutting into it from above or below the plane of the section. One should thus note whether a mineral is *consistently* included in another throughout the section before concluding that it is truly an inclusion. In the case of the K-feldspar megacrysts mentioned above, they are commonly poikilitic, and the numerous inclusions of other minerals in them are taken as important indicators of their late formation, overruling arguments for early formation based on grain size. But even when a mineral is consistently included in another, this is not always unequivocal evidence that the included phase ceased to crystallize before the host crystallization began. For example, **ophitic** texture (Figure 3-8) refers to the envelopment of plagioclase laths by larger clinopyroxenes, and is commonly interpreted to indicate that the clinopyrox-ene formed later. McBirney and Noyes *(1919),* however, noted a case in the Skaergard intrusion of Greenland in which the size of the plagioclase inclusions increases steadily from the clinopyroxene core to the rim. This suggests that plagioclase and clinopyroxene crystallized simultaneously. The clinopyroxene nucleated less readily,



Figure 3-8 Ophitic texture. A single pyroxene envelops several well-developed plagioclase laths. Width 1 mm.

so fewer crystals formed, and they enveloped the more numerous and smaller plagioclases. The later plagioclase grains that were included toward the host rims had longer to grow, and were therefore larger.

As Flood and Vernon (1988) concluded, none of the classical criteria for determining the sequence of crystallization are entirely satisfactory. Inclusions are perhaps the most reliable. Cases in which one mineral forms a rim around another, or where a mineral is included only in the core areas of another mineral, provide the strongest evidence for one mineral ceasing to crystallize before the other forms (or at least before the other ceases to crystallize). When one mineral occurs commonly as inclusions in another, and not vice versa, it strongly implies, but does not prove, that the included mineral crystallized earlier.

Although it may be difficult to unequivocally establish that one mineral formed entirely before another, some textures provide a clear testimony of the opposite: simultaneous mineral growth. The ophitic example of simultaneous plagioclase and clinopyroxene described by McBirney and Noyes (1979), cited above, is one example. In some shallow, H_2O-rich, granitic systems a single alkali feldspar might form (as discussed later in association with Figures 6-16 and 6-17). If the H_2O is suddenly lost, the melting point will rise quickly (Section 7.5.1), resulting in undercooling (even at a constant temperature) and rapid simultaneous crystallization of the alkali feldspar and quartz. Under these conditions, the two minerals do not have time to form independent crystals, but rather form an intergrowth of intricate skeletal shapes referred to as **granophyric** texture (Figure 3-9a). A rock dominated by such texture may be called a **granophyre.** The intergrowth may nucleate epitaxially on preexisting phenocrysts or dikelet walls. Granophyric texture looks like branching quartz rods set in a single crystal of

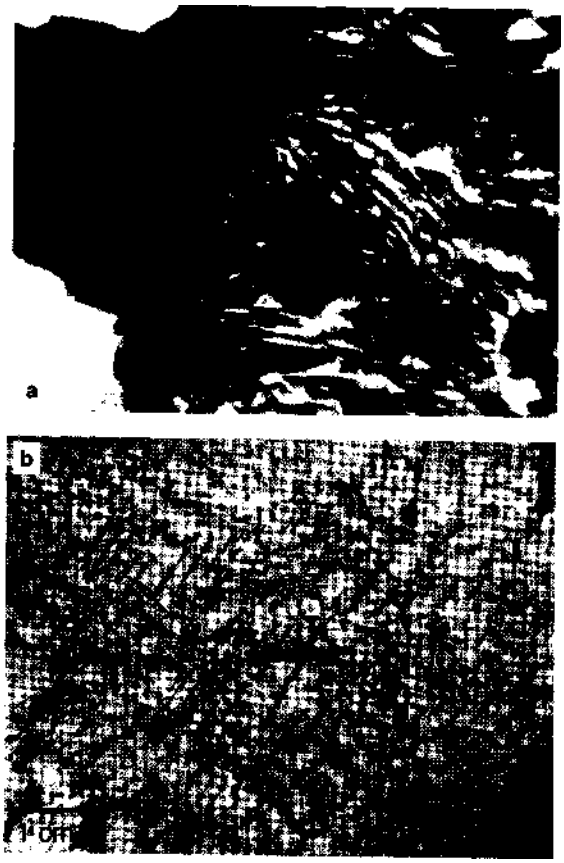


Figure 3-9 (a) Granophyric quartz-alkali feldspar intergrowth at the margin of a 1-cm dike. Golden Horn granite, WA. Width 1mm. **(b)** Graphic texture: a single crystal of cuneiform quartz (darker) intergrown with alkali feldspar (lighter). Laramie Range, WY. Scale grid in mm.

feldspar. The quartz rods go extinct at the same time indicating that they are all parts of the same larger crystal. A coarser variation of granophyric texture is called **graphic,** where the cuneiform nature of the quartz rods in the dominant feldspar host is readily seen hi hand specimen (Figure 3-9b).

3.1.5 Magmatic Reaction and Resorption

In some systems, early crystals react with the melt as crystallization proceeds. The reaction of olivine with melt to form pyroxene in the SiO_2-Mg_2SiO_4 system (seen later in Figures 6-12 and 6-14) is a prime example. Figure 3-10 illustrates an olivine phenocryst mantled by orthopy-roxene, produced at the olivine-melt interface by such a reaction. Other reactions may result from dropping pressure as a magma rapidly approaches the surface or from magma mixing or other compositional changes. Another common type of reaction occurs when a hydrous magma reaches the surface, where a sudden loss of pressure may release volatiles, and cause hydrous phenocrysts, such as hornblende or biotite, to dehydrate and oxidize, developing rims of fine iron oxides and pyroxenes (Figure 3-llc).

Resorption is the term applied to re-fusion or dissolution of a mineral back into a melt or solution from which it formed. Resorbed crystals commonly have rounded corners or are embayed. **Sieve texture** (Figure 3-1 la) or deep and irregular embayments (Figure 3-1 Ib) have been attributed by some to advanced resorption, but others claim that they are more likely to result from rapid growth enveloping melt due to undercooling (Figure

3.1 .6 Differential Movement of Crystals and Melt

Flow within a melt can result in alignment of elongated or tabular minerals producing **foliated** (planar) or **lin-eated** mineral textures. If lath-shaped microlites (typically plagioclase) in a volcanic rock are strongly aligned (commonly flowing around phenocrysts), the texture is called **trachitic** (Figure 3-12a). Random or non-aligned microlites are called **pilotaxitic or felty** (Figure 3-12b). Mingling of two magmatic liquids (either in a chamber or as flows) can create **flow banding** (alternating layers of different composition, Figure 3-13). Banding and mineral alignment can also result from flow near magma chamber walls, as discussed in the next chapter.

Suspended phenocrysts may cluster together and adhere by surface tension, a process called **synneusis** by many (Vance, 1969). Synneusis may be a prime mechanism for the production of growth twins, because the twin orientation may be an energetically favorable orientation for two crystals of the same mineral to adhere to each other. Multiple -grain clusters of adhering phenocrysts is called **cumulophyric** texture. If the clusters are essentially of a single mineral, the texture may be called **glomeroporphyritic** by some petrologists.

3.1 .7 Cumulate Textures

The development of igneous cumulates will be studied in conjunction with layered mafic intrusions in Chapter 12. Cumulate texture is a hallmark of these fascinating bodies (although it is not restricted to them). For the moment, we will avoid the complex question of how the crystals accumulate. Suffice it to say that historically the crystals were considered to accumulate by sinking or floating due to density contrasts with the liquid, but a recently proposed alternative is that they form in place near the roof, walls, or floor of the magma chamber.

In the ideal case, early-forming crystals of a single mineral accumulate (somehow) to the extent that they are

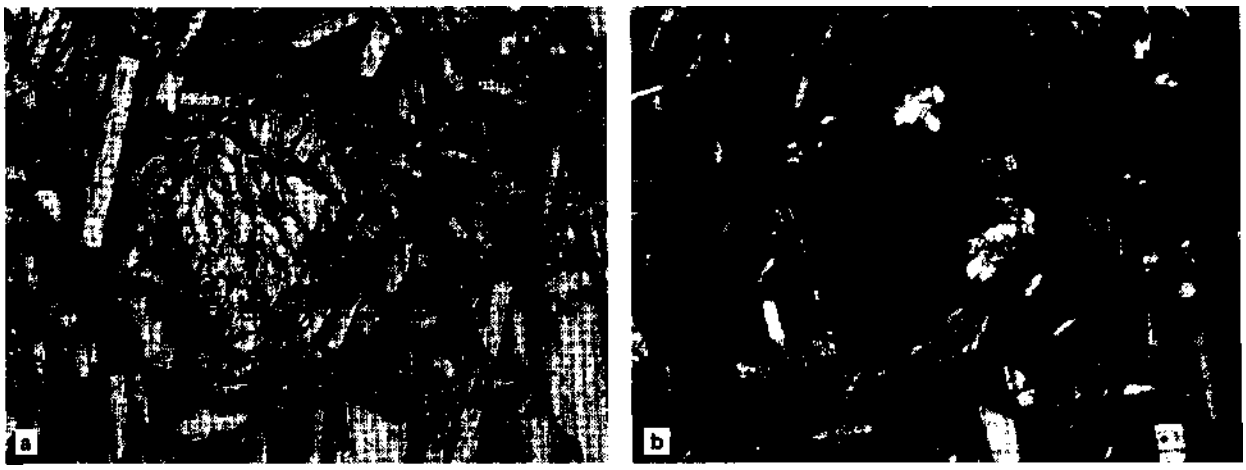


Figure 3-10 Olivine mantled by orthopyroxene in plane-polarized light **(a)** and crossed nicols **(b),** in which olivine is extinct and the pyroxenes stand out clearly. Basaltic andesite, Mt. McLaughlin, Oregon. Width ~ 5 mm.

in mutual contact, with the remaining liquid occupying the **interstitial** spaces between the crystals (Figures 3-7 and 3-14a). Mutual contact, however, is not a rigorous requirement for cumulate texture, and close approximation will suffice. The principal types of cumulates are distinguished on the basis of the extent to which the early-formed crystals, once accumulated, grow prior to ultimate solidification of the interstitial liquid. It would be unusual if the interstitial liquid had the same composition as the accumulated crystals, because most magmas are chemically more complex than any single mineral. So if the liquid crystallizes essentially in-place, without exchange with the larger magma reservoir in the interior of the chamber, it should produce some of the initial mineral (assumed to be plagioclase in Figure 3-14, but it could be olivine, pyroxene, chromite, etc.) plus any other minerals that together constitute the interstitial magma. There may thus be some modest additional growth of the early minerals, together with formation of other, later-forming minerals in the interstitial spaces. The result is called an **orthocumulate** texture (Figure 3-14b).

If the interstitial liquid can escape and exchange material (via diffusion and/or convection) with the liquid of the main chamber, the early-forming cumulate minerals may continue to grow as the extra components of the interstitial liquid escape. The result is **adcumulate** texture (Figure 3-14c): a nearly monomineralic cumulate with perhaps a few other minerals caught in the last interstitial points. Hunter (1987) concluded that compaction and expulsion of some of the intercumulus liquid *must* accompany the formation of adcumulates, because adcumulate textures can be observed in areas too far from the open melt for material to diffuse through the limited porosity associated with late-stage growth. He also concludes that textural equilibrium, resulting in **polygonal** texture (see polygonal mosaic below) may be approached or maintained as the grains change shape during late growth and compaction. If true, this readjustment of mineral shapes in contact with each other and melt has important bearing on the earlier discussion of crystallization sequence and molding of one crystal to accommodate another. Hunter (1987) further noted that dihedral angles (see Figure 11-1) between minerals in contact and the late stages of melt trapped at mineral edge boundaries are nearly constant in adcumulates. As crystals grow and impinge, he maintained, surface energy considerations (and hence dihedral angles) control the shape of the last liquid pockets and few euhedral terminations are preserved. Thus early euhedral minerals will not dominate the final textures of plutonic rocks.

If the later minerals have a slow nucleation rate, they may envelop the cumulus grains, as described above in Section 3.1.4. The result is **poikilitic** texture, but the host oikocryst may be so large and interstitial in some instances that it may be hard to recognize it as such in the small area of a thin section. A large oikocryst also requires exchange between the interstitial liquid and the main magma reservoir in order to provide enough of its components, and to dispose of excess components that would lead to the formation of other minerals. It is thus considered to be a type of adcumulate phenomenon, and is termed **heteradcumulate** (Figure 3-14d). Finally, **mesocumulate** is a term applied to cumulate textures that are intermediate between ortho-and adcumulates.

3.1.8 Primary Twinning

A twin is an intergrowth of two or more orientations of the same mineral with some special crystallographic relationship between them. Primary (or growth) twins are twins that form because of mistakes during crystalliza-

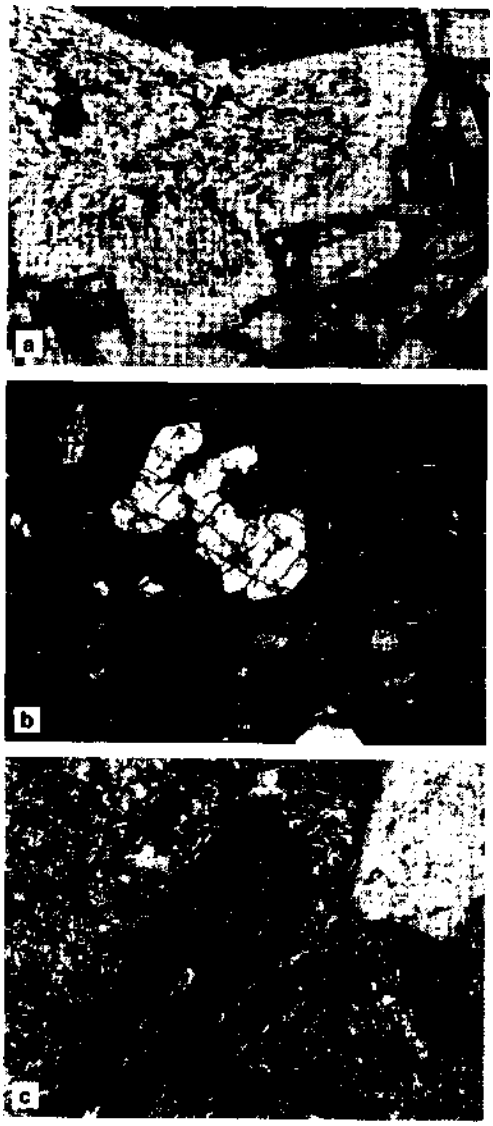


Figure 3-11 (a) Sieve texture in a cumulophyric cluster of pla-gioclase phenocrysts. Note the later non-sieve rim on the cluster. Andesite, Mt. McLoughlin, OR. Width 1 mm. **(b)** Resorbed and embayed olivine phenocryst. Width 0.3 mm. **(c)** Hornblende phenocryst dehydrating to Fe-oxides plus pyroxene due to pressure release upon eruption, andesite. Crater Lake, OR. Width 1 mm.

tion from a melt. An example is the simple (two-part) carlsbad twins in feldspars shown in Figures 3-5b and 3-18a). Nucleation error of twin-oriented domains is probably the predominant processes involved in primary twinning, but synneusis adsorption of two constituents in twin orientation may also occur. Nucleation error is most likely to occur during rapid growth, as might immediately follow nucleation, which should quickly reduce supersaturation or undercooling. For a detailed discussion of synneusis, see Vance (1969) and Dowty (1980). Repetitive albite twinning (Figure 3-18b) is also believed to result from nucleation errors during growth.

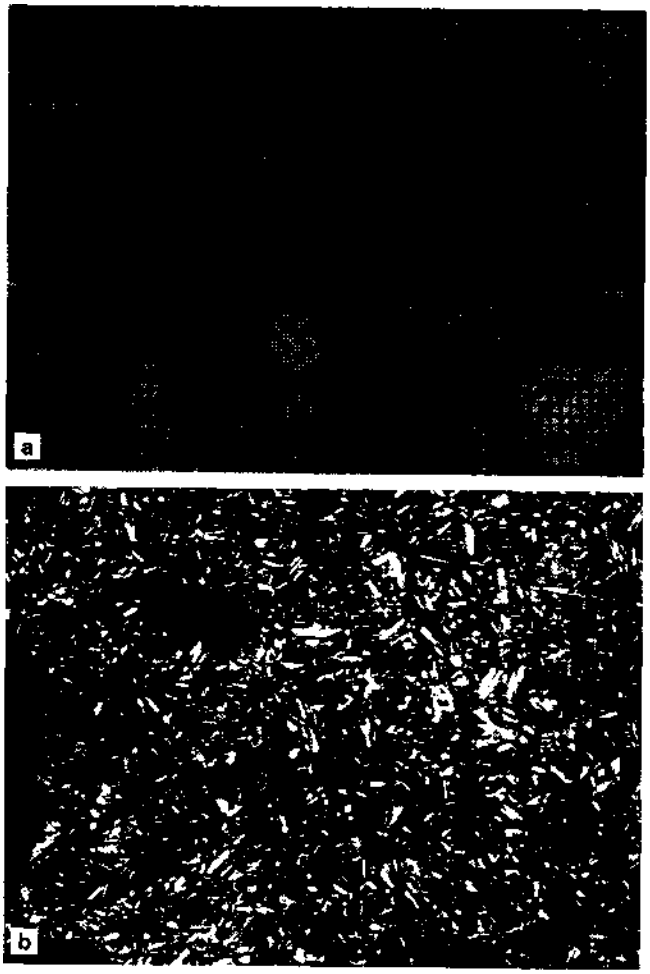


Figure 3-12 (a) Trachytic texture in which microphenocrysts of plagioclase are aligned due to flow. Note flow around phenocryst (P). Trachyte, Germany. Width 1 mm. From MacKen-zie et al. (1982). **(b)** Felty or pilotaxitic texture in which the microphenocrysts are randomly oriented. Basaltic andesite, Mt. McLaughlin, OR. Width 7 mm.

3.1.9 Volcanic Textures

Volcanic rocks cool quickly and tend to form numerous small crystals, as discussed above. Phenocrysts are an exception, and result from slower cooling beneath the surface prior to eruption. Upon eruption, the remaining liquid crystallizes to fine tabular or equaint crystals comprising the **ground m ass** The groundmass crystals are called **microlites** (if they are large enough to be birefrin-gent) or **crystallites** (if they are not). Microlites that are significantly larger than the groundmass, yet still microscopic, are called **microphenocrysts.** They are formed upon eruption and represent minerals with a higher ratio of growth rate to nucleation rate than the finer ground-mass phases.

Basalts crystallize readily, because they are very hot and are dominated by minerals with simple structures.



Figure 3-13 Flow banding in andesite. Mt. Rainier, WA.

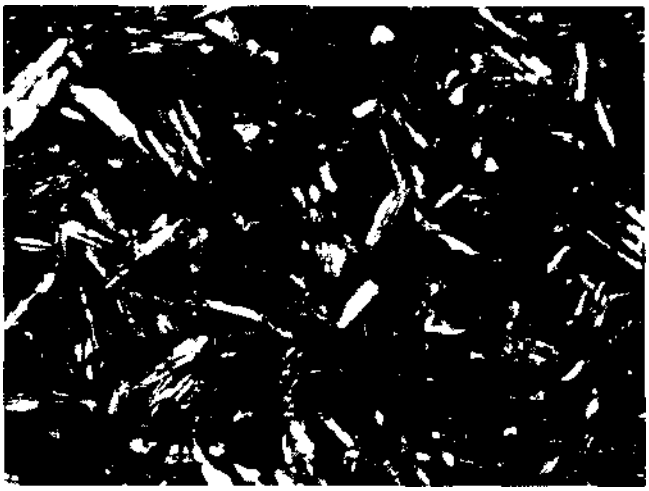


Figure 3-15 Intergranular texture in basalt. Columbia River Basalt Group, Washington. Width 1 mm.

The common result is a texture with a dense network of elongate plagioclase microphenocrysts and granular pyroxenes, with smaller magnetite crystals. Glass may solidify as a late interstitial material. The amount of glass in basaltic rocks is generally less than in more silicic vol-canics, but can vary considerably, from virtually none to highly glassy when basalt comes into contact with water. The lexicon of basaltic textures reflects the variation in glass content, which is crudely correlated with decreasing pyroxene size in a fretwork of intergrown plagioclase laths. **Ophitic** texture (Figure 3-8) refers to a dense network of lath-shaped plagioclase microphenocyrsts included in larger pyroxenes with little or no associated glass. This grades into **sub-ophitic** (smaller pyroxenes, that still partially envelop the plagioclase), and then into **intergranular** texture (Figure 3-15), in which the plagioclase and pyroxene crystals are subequal in size, and glass (or its alteration products) is still relatively minor. Intergranular texture grades into **intersertal** texture when interstitial glass or glass alteration is a significant component. When glass becomes sufficiently plentiful that it surrounds the microlites or microphenocrysts, the texture is called **hyalo-ophitic.** Hyalo-ophitic grades into **hyalopilitic** as the glass fraction becomes dominant, and both pyroxenes and plagioclases occur as tiny microlites. So don't these basalt petrographers have fun with nomenclature?

Holohyaline (glassy) texture is more common in silicic rhyolite and dacite flows. If a rock is > 80% glass it is called **obsidian.** Some workers prefer to restrict the term

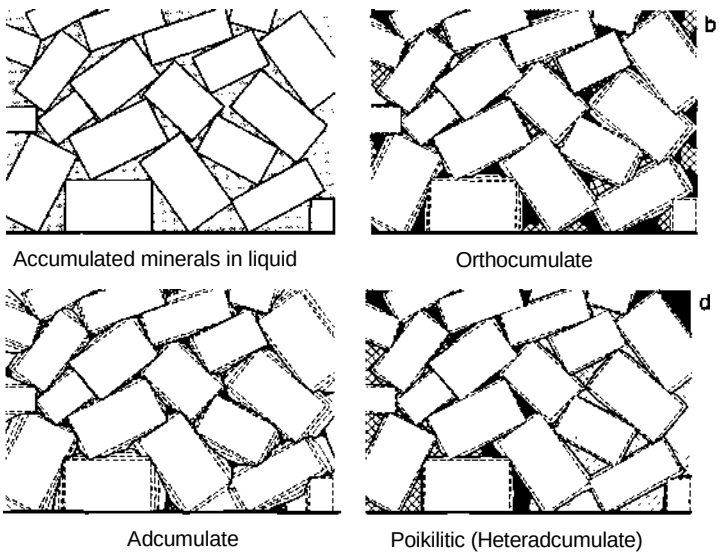


Figure 3-14 Development of cumulate textures, **(a)** Crystals accumulate by crystal settling or simply form in place near the margins of the magma chamber. In this case plagioclase crystals (white) accumulate in mutual contact, and an intercumulus liquid (gray) fills the interstices, **(b)** Orthocumulate: intercumulus liquid crystallizes to form additional plagioclase rims plus other phases in the interstitial volume (dark). There is little or no exchange between the intercumulus liquid and the main chamber, **(c)** Adcumulates: open-system exchange between the intercumulus liquid and the main chamber (plus compaction of the cumulate pile) allows components that would otherwise create additional intercumulus minerals to escape, and plagioclase fills most of the available space, **(d)** Heteradcumulate: intercumulus liquid crystallizes to additional plagioclase rims, plus other large minerals (hatched and shaded) that nucleate poorly and poikilitically envelop the plagioclases. After Wager and Brown (1967).

to more silica-rich glasses, and refer to basaltic varieties as **tachylite** or simply **basaltic glass.** Obsidian is very dark colored, in spite of its commonly silicic nature, because glass is quickly tinted by very minor amounts of impurity. The glass in silicic lavas is not necessarily caused by very rapid cooling, because some obsidian flows are too thick to have cooled so quickly. Motion and/or the characteristically slow diffusion and nucleation of highly polymerized and viscous silicic flows may impede crystallization and produce these highly glassy rocks. The textural terms described above are usually applied to randomly oriented crystals, but they may grade into **trachytic** texture as flow causes alignment of the microlites.

Trapped bubbles of escaping gas create sub-spherical voids in volcanics, called **vesicles.** Vesicles tend to rise in less viscous basaltic magmas, and thus concentrate near the surface of basaltic flows. There is a complete gradation from basalt to **vesicular basalt to scoria** with increasing vesicle content. Vesicles filled with later mineral growth, typically secondary zeolite, carbonate, or opal, are called **amygdules.** The silicic counterpart of scoria is **pumice.** Pumice is typically light and frothy, and fresh samples float in water. Frothy pumice is commonly light gray, even though its corresponding vesicle-free obsidian may be black. The reason for this contrast is that the bubbles expand the glass to a thin film between the bubbles, which refracts and diffuses light, just as breaking waves form whitecaps on the dark sea.

3.1.10 Pyroclastic Textures

Pyroclastic rocks are fragmental, and are produced by explosive volcanic activity. The classification of pyro-clastic rocks is based on the nature of the fragments **(py-roclasts, or tephra)** as discussed in Section 2.5. The eruptive and emplacement modes are discussed in Chapter 4. The ash component of pyroclasts is commonly a mixture of pulverized rock and primary glass (including shattered pumice and aerosol liquid). The vesicles in pumice expand rapidly upon explosive eruption and are usually destroyed. The interstitial glass then forms cus-pate- or spicule-shaped three-pointed shards (Figure 3-16a). Because these shards are commonly warm in a pyroclastic flow, they deform in a ductile fashion, and fold over to the shape indicated on the right of Figure 3-16a and in Figure 3-16b. This type of bending, and other structures caused by compression and deformation resulting from settling in hot ash accumulations, are collectively referred to as **eutaxitic** textures. Larger pieces of pumice may accumulate intact and have the gas squeezed from them, eliminating the bubbles. If all of the gas is expelled, the pumice returns to the black color of obsidian, and the squashed fragments are called **fi-amme.** In fluid lavas, such as basalts, bursting bubbles

Y



Figure 3-16 (a) The interstitial liquid (black) between bubbles f in pumice (left) become 3-pointed-star-shaped glass shards in I ash containing pulverized pumice. If they are sufficiently warm | (when pulverized or after accumulation of the ash) the shards I may deform and fold to contorted shapes, as seen on the right'. and in the photomicrograph **(b)** of the Rattlesnake ignimbrite, \ SE Oregon. Width 1 mm.

hurl fine spray aloft, and it falls as glassy pellets called' "Pele's tears" (after the Hawaiian volcano god, Pele), or[s]. the magma may be stretched to form delicate glass \ threads ("Pele's hair"). Ash falling through very moist j air may accumulate successive layers on a single ash mi- j cleus, forming spheroidal balls called **accretionary lapilli.** Consolidated deposits of such lapilli are called **pisolitic | tuffs.**

3.2 SECONDARY TEXTURES: POST-MAGMATIC CHANGES

Secondary textures are those that develop after the ig- i neous rock is entirely solid. These processes do not involve melt and are thus really *metamorphic* in nature, The process of crystallization does not necessarily cease when the magma becomes solid. As long as the temperature is high enough, recrystallization, and both chemical and textural re-equilibration, take place (otherwise metamorphic petrologists would have very little to do). Because large plutons may remain at temperatures

equivalent to high-grade metamorphism for thousands of years, there is ample opportunity for such processes to occur. Solid-state processes that occur as a result of igneous heat (even though waning) are called **autometa-morphic,** and will be covered in this section. Because solid-state crystallization processes are truly metamor-phic, they will be discussed more fully in Chapter 23.

Ostwald ripening is a process of annealing (or textural maturing) of crystals in a static environment. As Hunter (1987) noted, differences in grain-boundary curvature will drive grain growth by Ostwald ripening until straight boundaries result (Figure 3-17). In such recrystallization, grain boundaries migrate toward their centers of curvature. Small grains with convex outward curvature are thus eliminated as the surfaces of neighboring larger grains with convex inward curvature encroach upon them. If the process attains textural equilibrium in a solid, there will be similarly sized grains having straight, approximately 120° triple-grain intersections (Figure 3-17). This equilibrium texture is most common in monomin-eralic metamorphic rocks (quartzite and marble), particularly if metamorphosed in a nearly static stress regime. We will discuss annealing further in Chapter 23. Most igneous rocks are not monomineralic, however, and rarely attain a good equilibrium texture. Relative differences in surface energy of contrasting mineral types and the coarse grain size of plutonics serve to establish and maintain interlocking textures in most cases. Ostwald ripening, however, may eliminate smaller grains in favor of larger neighbors at an early stage of growth, producing a more uniform distribution of grain sizes. Volcanic

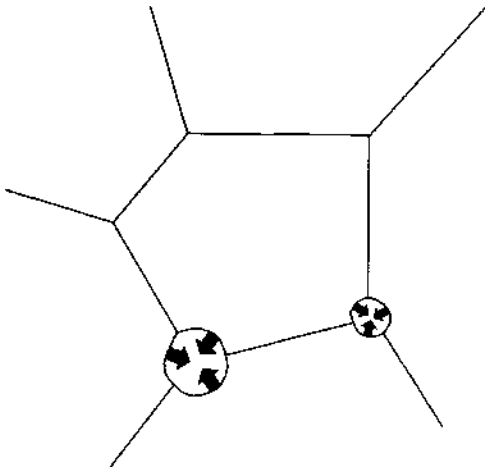


figure 3-17 "Ostwald ripening" in a monomineralic material. Grain boundaries with significant negative curvature (concave inward) migrate toward their center of curvature, thus eliminating smaller grains and establishing a uniformly coarse-grained equilibrium texture with 120° grain intersections (polygonal mosaic).

rocks with small initial grain size are far less stable than plutonics, and the groundmass recrystallizes readily. Glass is particularly unstable, and readily devitrifies to fine mineral replacements, as described below. But vol-canics cool quickly to low temperatures, and kinetic restrictions to recrystallization develop early. The retention of magmatic textures is thus surprisingly good in igneous rocks, but some types of solid-state recrystallization are well known and I shall review them here.

3.2.1 Polymorphic Transformation

As you undoubtedly learned in mineralogy, many natural substances have more than one crystal structure. Alternative structural forms of the same chemical substances are called **polymorphs.** Familiar polymorphs are graphite-diamond, calcite-aragonite, kyanite-andalusite-sillimanite, and the several polymorphs of SiO_2. As we shall see in Chapter 6, a given structure is most stable over a particular range of pressure and temperature conditions, so that one polymorph will transform to another when the conditions change from its stability range to that of an alternative structure. Cooling and depressurization of rising, partly or wholly crystalline igneous rocks may cross the stability field boundaries of some polymorphs, resulting in transformations. **Dis-placive transformations** involve only the shifting of atomic positions and bending of bond angles. A classic example is the high-quartz to low-quartz transition, in which the hexagonal high-quartz structure inverts to the trigonal structure of low-quartz upon cooling (Figure 6-6). **Reconstructive transformations,** such as graphite-diamond, or tridymite-high-quartz, involve the breaking and re-forming of bonds. Displacive transformations occur readily, so that one polymorph gives way to another as soon as its stability field is reached. Reconstructive polymorphism is less easily managed, and one polymorph may remain in the stability field of another. Polymorphic transformations are common in many minerals, including quartz and feldspars, but may be difficult to recognize texturally, because evidence for the initial phase might be completely lost, and only the replacement polymorph remains. If the crystal form of the earlier phase is distinctive, however, the replacement polymorph may form a **pseudomorph** of the original, rather than assume its own characteristic form. For example, high quartz may crystallize as early phenocrysts in some rhyolites. The crystal form of high quartz is characteristically a hexagonal dipyramid, without the predominant prism faces of low quartz. Because this is a displacive transformation, the high-quartz in such crystals must invert to low-quartz at 573°C (at atmospheric pressure). The original phenocryst shape is usually preserved, however, thus providing evidence for the initial form.

Another possible result of polymorphic transformations is the development of secondary twins, which can also provide a clue to the transformation process. Recognition of such features can provide some very useful information concerning the thermal history of a rock.

3.2.2 Secondary Twinning

In addition to the primary twins discussed above, twinning may occur by secondary processes in pre-existing minerals. Secondary twins can occur as a result of polymorphic transformation or deformation. **Transformation twins** are caused when a high-temperature crystal structure inverts to a low-temperature polymorph. As high-temperature structures have more vibrational energy, they generally exhibit a higher degree of symmetry than the low-temperature alternative. Because symmetry is lowered with cooling, the high-temperature form typically has a choice of two or more alternative low-symmetry orientations. If a whole crystal assumes one of the alternatives, no twinning will result. If, on the other hand, different portions of the same crystal are displaced into each of the alternative choices, the portions will generally be in twin relationship with the other portion. This is the origin of the "cross-hatched" or "tartan" twins of micro-cline (Figures 3-18c and d), which are produced when the high-temperature monoclinic form inverts to the low-temperature triclinic microcline structure. The familiar multiple "albite" twins of plagioclase (Figures 3-18b), are attributed to a similar monoclinic to triclinic transformation, but this does not occur in plagioclase of intermediate composition, and are commonly really primary twins resulting from nucleation error during growth. Cyclic twins in quartz and olivine are other examples of transformation twins.

Twins can also be generated by deformation of solidified igneous rocks. Twinning is an important aspect of a rock's response to deformation, because the shifting of a portion of a grain into the twin orientation is an easier response to effect than is grain rupture. **Deformation twins** in plagioclase can also occur on the albite twin law, but they usually lack the extremely straight lamellar form of their primary counterparts. Deformation twins in plagioclase are most easily recognized when they are wedge-shaped and bent (Figure 3-19). Calcite also readily develops deformation twins in response to shear. Deformation twins enhance the ductile response of rocks such as marble, and are uncommon in other minerals.

3.2.3 Exsolution

Exsolution, discussed further in Section 6.5.4, involves mixing that becomes increasingly limited in some solid solution minerals as they cool. Perhaps the most common example occurs in alkali-feldspars where unmixing results in separation of Na-rich and K-rich segregations. Because the unmixing in this case involves partitioning of K and Na ions, and not strongly-bonded Si and Al, it takes place relatively easily and the segregations appear as a coherent intergrowth of long wispy lamellae. "Co-

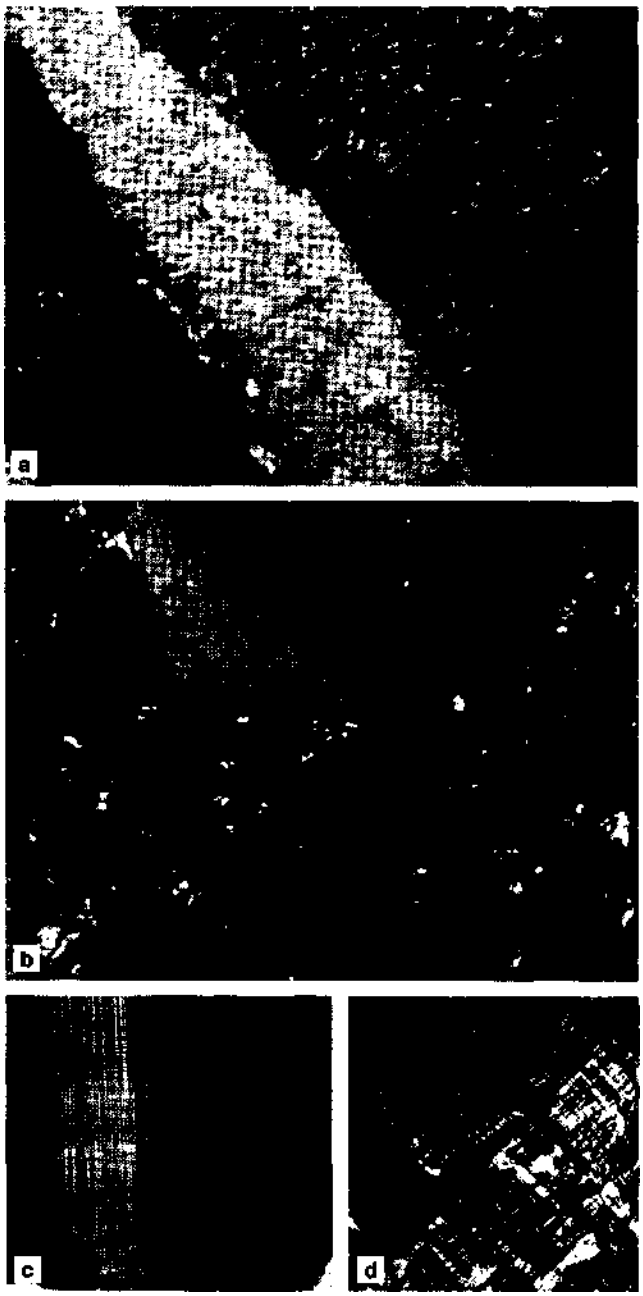


Figure 3-18 Feldspar twins, **(a)** Carlsbad twin in orthoclase. Wispy perthitic exsolution is also evident. Granite, St. Cloud MN. **(b)** Very straight multiple albite twins in plagioclase, set in felsitic groundmass. Rhyolite, Chaffee, CO. **(c-d)** Tartan twins in microcline. All field widths ~1 mm.



Figure 3-19 Polysynthetic deformation twins in plagioclase. Note how they concentrate in areas of deformation, such as at the maximum curvature of the bent cleavages, and taper away toward undeformed areas. Gabbro, Wollaston, Ontario. Width 1mm.

herent," in this sense, means that the lattices of the lamellae have a specific crystallographic relationship to the host, and are not randomly oriented. When the alkali feldspar is potassic, the result is exsolved albite lamellae in a K-feldspar host, and is called **perthite** (Figure 3-18a). When the alkali feldspar is sodic, the lamellae are K-feldspar in an albite host, and is referred to as an-tiperthite. Exsolution also occurs in plagioclase at times, but albite-anorthite unmixing involves Si-Al, as well as Na-Ca exchange, and the process, when it occurs at all, produces much finer lamellae.

Exsolution also occurs in pyroxenes, in which low-Ca orthopyroxene separates from high-Ca clinopyroxene. Thin lamellae of one of these in a host of the other are common. Pigeonite is an intermediate mixture, and is only found in volcanic rocks that cooled too quickly to allow unmixing to occur.

Unmixing does not have to be coherent, and irregular patches of exsolved phases are commonly found, as are instances of complete expulsion of the exsolved phase to form separate grains immediately outside the host. Ex-solution occurs in amphiboles and a few other common minerals. Some mafic silicates may even exsolve an Fe-Ti-oxide. In some anorthosites, high-temperature and high-pressure pyroxenes can dissolve a considerable amount of Al, and exsolve lamellae of plagioclase as they cool at lower pressures. The red or pink color common to some feldspars is caused by the exsolution of fine hematite.

Ocelli are spherical or ovoid bodies a few mm to a few cm across that occur in some igneous rocks. Some appear to result from *liquid* immiscibility, a primary exsolution phenomenon discussed in Chapters 6,11, and 19. Others are probably amygdule fillings, and yet others may be isolated blobs of mingled magmas.

3.2.4 Secondary Reactions and Replacement

Solid-solid and solid-vapor reactions are processes that dominate during metamorphism, and will be left largely to the second portion of this text. As stated above, however, igneous rocks cool through a temperature range appropriate to this realm, and plutonic rocks remain in that realm for a considerable length of time. Secondary mineral reactions that occur in igneous rocks as they cool, and that are not products of a distinct later metamorphic event, are commonly called **autometamorphic** processes rather than metamorphic ones, because they are a natural part of igneous cooling. Autometamorphic processes are more common in plutonic rocks than in volcanics, because they remain at elevated temperatures for a longer time. Diagenetic and weathering processes are not considered to be autometamorphic (a rather arbitrary distinction). Most, but not all, autometamorphic reactions involve minerals at moderate temperatures in an environment in which H_2O is either liberated from residual melt or externally introduced. Such alterations are a subset of autometamorphism which involves hydration, and they are called **deuteric** alterations. Some of the principal alteration processes follow.

Pyroxene is a common primary mafic mineral in a variety of igneous rocks. If H_2O penetrates at modest temperatures, a deuteric alteration of pyroxene to amphibole results, called **uralitization** (Figure 3-20a). Any gradation from amphibole rims on pyroxene cores to patchy multiple cores of pyroxene in an amphibole, to complete replacement, is possible. The amphibole may be a single crystal of hornblende, or a fibrous actinolite or hornblende aggregate. Either, when demonstrated to result from pyroxene alteration, may be called **uralite,** but the term is more commonly applied to the latter.

Biotitization is a similar process of hydration/deuteric alteration that produces biotite, either directly from pyroxene, or, more commonly, from hornblende. Because biotite contains little Ca, epidote may be produced as Ca is released during the alteration of hornblende to biotite.

Chloritization is the alteration of any mafic mineral to chlorite. Chlorite is a very hydrous phyllosilicate, and commonly replaces the less hydrous mafics at low temperatures when water is available. Pyroxenes, hornblendes, and biotites are commonly observed in thin section in various stages of alteration to chlorite. As in other deuteric alterations, hydration attacks the outer margin of a mineral, so that chlorite commonly replaces the initial mafic from the rim inward. In the case of biotite, H_2O may work its way along the prominent cleav-



Figure 3-20 (a) Pyroxene largely replaced by hornblende. Some pyroxene remains as light areas (Pyx) in the hornblende core. Width 1 mm. **(b)** Chlorite (light) replaces biotite (dark) at the rim and along cleavages. Tonalite. San Diego, CA. Width 0.3 mm.

ages, and chlorite can be seen to replace biotite margins as well as along cleavage planes (Figure 3-20b).

Sericite is a term applied to any very fine-grained white mica. **Seritization** is then the process by which fel-sic minerals (usually feldspars of feldspathoids in igneous rocks) are hydrated to produce sericite. Incipient stages can be recognized by a fine dusty appearance of feldspars in plane-polarized light. In more advanced stages of alteration, the feldspars appear speckled with fine micas having yellowish birefringence, and then larger clots with coarser crystals and higher birefringence. K^+ ions are required for plagioclase to be altered to sericite. Potassium may be released by chloritization of nearby biotite. K-feldspar requires no additional K^+, and may commonly be more sericitized than associated plagioclase.

Saussuritization is the alteration of plagioclase to an epidote mineral. More calcic plagioclase is stable at higher temperatures (Figure 6-8), and is less stable than its sodic counterpart at low temperatures. The calcium-rich types thus break down to more nearly pure albite, releasing Ca and Al to form an epidote mineral (± calcite and/or sericite). In zoned plagioclase, one can occasionally observe the products of saussuritization or seritiza-tion concentrated in the Ca-rich core or some of the more Ca-rich oscillatory bands in the original grain.

Olivine is easily altered in cooling mafic rocks, even in volcanics. It is commonly rimmed or replaced by serpentine or dark brown iddingsite.

Symplectite is a term applied to fine-grained inter-growths resulting from the combined growth of two or more minerals as they replace another mineral. As in any of the replacements described above, replacement may be partial or complete. Complete replacement and pseudomorphs are common. The fibrous actinolite-horn-blende "uralite" aggregates replacing pyroxene are an example. Biotite + epidote replacing hornblende is another. **Myrmekite** is an intergrowth of dendritic quartz in a single crystal of plagioclase (Figure 3-21). The quartz appears rod-like in thin section, and numerous adjacent

Figure 3-21 Myrmekite formed in plagioclase at the boundary with K-feldspar. Photographs courtesy L. Collins.

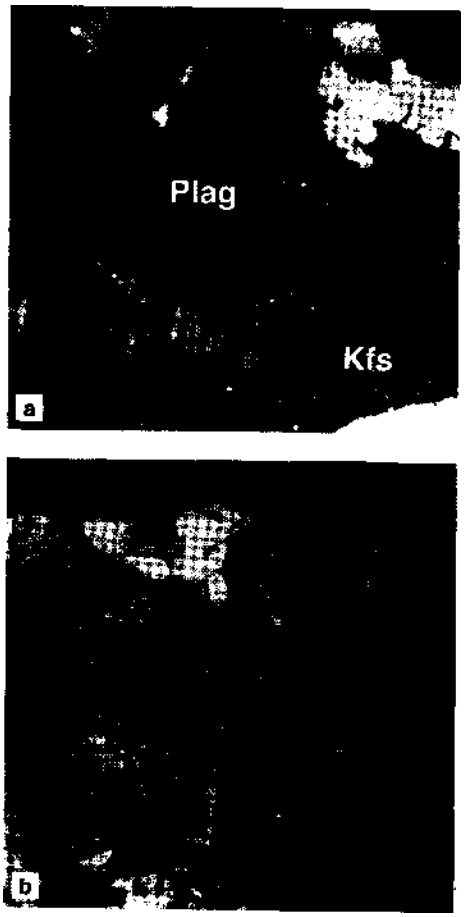

Plag
Kfs
a
b

Table 3-1 Common Igneous Textural Terms

Crystallinity	
Holocrystalline	Consisting entirely of crystals (default term, not commonly used).
Hypocrystalline	Containing both crystals and glass. Consisting entirely of glass.
Holohyaline, Vitric	
Grain Size Aphanitic	Most minerals are too fine-grained to see with the naked eye.
Phaneritic	The minerals are coarse enough to see with the naked eye.
Cryptocrystalline	Too fine-grained to distinguish the minerals microscopically.
Fine grained	Average crystal diameter is less than 1 mm.
Medium grained	Average crystal diameter is 1-5 mm.
Coarse grained	Average crystal diameter is greater than 5 mm.
Pegmatitic Aplitic	Very coarse grained.
Equigranular	Fine- to medium-grained xenomorphic and equigranular (looks sugary).
Inequigranular	Grains are all approximately the same size.
Porphyritic Textures	Grains vary considerably in size.
Porphyritic	
Megaporphyritic	Approximately bimodal size distribution (usually requires a great difference).
Microporphyritic	Porphyritic texture that can be seen in hand specimen (rarely used).
Phyric (-phyric)	Porphyritic texture that is visible only under the microscope.
Phenocryst	An adjective (or suffix) referring to porphyritic texture.
Microphenocryst	Large crystal set in a fine matrix.
Megacryst Poikilitic	Microscopic crystals that are still larger than the remainder of the groundmass.
Oikocryst	An unusually large crystal, either a phenocryst or a xenocryst.
Groundmass	The host phenocryst (oikocryst) contains many inclusions of other minerals.
Cumulophyric	The host phenocryst in poikilitic texture.
Glomeroporphyritic	The glassy or finer grained element in the porphyritic texture (matrix, mesostasis).
Hiatial porphyritic	Phenocrysts of the same or different minerals occur in clusters (Figure 3-1 la).
	Synonymous with cumulophyric (used by some to specify that only one mineral is involved).
	There is a pronounced difference in size between the phyric and groundmass phases (the default term: rarely used).
Seriate Aphyric **Form of Individual Grains**	
Euhedral (idiomorphic)	Completely bounded by crystal faces.
Subhedral (subidiomorphic)	Crystal faces only partially developed.
Anhedral (allotriomorphic)	Crystal faces are entirely absent.
Crystal habits: equaint, prismatic, columnar, accicular, fibrous, tabular, platy, micaceous, lath-shaped, etc. (see any mineralogy text).	
Corroded (embayed)	Subhedral or anhedral forms produced by partial melting (resorption) of phenocrysts by the melt.
Forms of Grains in the Rock as a Whole	
Panidiomorphic	There is a continuous gradation in size.
Hypidiomorphic	Non-porphyritic (no phenocrysts).
Allotriomorphic	Most of the grains are euhedral (rare).
Sutured	Consisting predominantly of subhedral grains (common in many granitic rocks).
	Most of the grains are anhedral. Common.
Mosaic	Characterized by articulation along highly irregular inter-penetrating boundaries. Common in recrystallized deformed rocks. A texture of polygonal equigranular crystals.
Intergrowths	
Host (Oikocryst)	The large mineral that includes others in poikilitic texture.
Guest	The included mineral in poikilitic texture.
Poikilitic	One large crystal contains several small discrete crystals of another mineral. Refers to growth phenomena, not exsolution or replacement.
Graphic	An intergrowth in which the guest shows angular wedge-like forms. Usually occurs with quartz in microcline (Figure 3-9b).
Micrographic	Graphic texture visible only under the microscope.
Granophyric	A texture of certain porphyritic rocks in which the groundmass minerals (usually quartz and alkali feldspar) penetrate each other as feathery irregular intergrowths. Resembles micrographic texture, but is more irregular.
Exsolution lamellae	Lamellar bands of a phase exsolved from a host phase (Figure 3-9a).
Perthitic	Irregular veins, patches, lenses, etc., of sodic plagioclase in an alkali feldspar host.Usually results from exsolution (Figure 3-18a).
Antiperthitic	Exsolution lamellae of alkali feldspar in a plagioclase host. Usually much thinner than perthite.
Symplectite	A replacement texture in which a mineral is replaced by an intergrowth of one or more minerals.
Myrmekite	A secondary texture consisting of irregular "wormy" blebs or rods of quartz in a plagioclase host adjacent to alkali feldspar grains (Figure 3-21).
Spherulitic	A radial intergrowth of fibrous minerals. Commonly alkali feldspar and quartz in devitrified silicic volcanics.
Axiolitic	Similar to spherulitic, but the fibers occur in a layer and are oriented normal to its walls.

(continued)

Table 3-1 *Continued*

Variolitic	Spherulitic structures consisting of divergent plagioclase fibers in certain basalts. Anhedral textures
Coalescent	developed by simultaneous growth of two mineral grains in contact.
Textures of Mafic Igneous Rocks	
Ophitic	Large pyroxene grains enclose small random plagioclase laths (Figure 3-8).
Subophitic	The plagioclase laths are larger and only partially enclosed by the pyroxene.
Nesophitic	Plagioclase is larger yet and the pyroxenes are interstitial.
Intergranular	Small discrete grains of pyroxene, olivine, etc., fill the interstices in a random network of larger plagioclase laths (Figure 3-15).
Intersertal	Glass, cryptocrystalline material, or alteration products occupy the spaces between plagioclase laths. An
Hyalo-ophitic	intersertal texture in which a larger amount of glass is present and less pyroxene. Still more glass is present and
Hyalopilitic	plagioclase occurs only as tiny random microlites.
Diktytaxitic	Texture of certain basalts containing abundant angular interstitial gas cavities between the plagioclase laths. Interstitial
Cumulate	growth of a mineral between earlier ones which are all in contact and give the distinct impression that they accumulated at the bottom of a magma chamber (Figure 3-14). Cumulate texture with other
Orthocumulate	minerals occupying the interstitial areas (Figure 3-14b). Cumulate texture in which the early cumulate
Adcumulate	minerals grow to fill the pore space (Figure 3-14c). Intermediate between ortho- and adcumulate.
Mesocumulate	
Replacement Textures	One or more minerals replace another, retaining the form of the original mineral.
Pseudomorph	A replacement texture in which a mineral is replaced by an intergrowth of one or more minerals.
Symplectite	**Other forms include replacement of rims, cores, or patches of one mineral by another. Common processes include:**
Uralitization	Replacement of pyroxene by amphibole (Figure 3-21 a).
Saussuritization	Replacement of plagioclase by epidote. Replacement of
Biotitization	pyroxene or amphibole by biotite. Replacement of any mafic
Chloritization	mineral by chlorite (Figure 3-20b). Replacement of feldspar or
Seritization	feldspathoids by fine white micas.
Miscellaneous Terms	
Interstitial	One mineral fills the interstices between earlier crystallized grains (Figure 3-7).
Crystallites	Minute, inchoate crystals in earliest stages of formation. They are isotropic and cannot be identified under the microscope.
Microlites	Tiny needle- or lath-like crystals of which at least some properties are microscopically determinable.
Felty	Consisting of random microlites (Figure 3-13b). Like felty.
Pilotaxitic	Consisting of (feldspar) microlites aligned due to flow (Figure 3-12a). Having embayments due to reaction with
Trachytic	the melt (resorption) (Figure 3-2). Crystals which grew as, or have been corroded to, a skeletal framework with a
Embayed	high proportion of internal voids
Skeletal	(Figure 3-4).
	Crystals are filled with channelways (appearing as holes) due to resorption (Figure 3-1 la).
Sieve	Oriented nucleation of one mineral on another of a different kind. Overgrowths of plagioclase
Epitactic	on alkali feldspar. Contains gas bubbles. Highly vesicular.
Rapakivi	Frothy vesicular structure characteristic of pumice. Containing
Vesicular	vesicles into which microphenocrysts protrude.
Scoriaceous	Gas cavities present in certain plutonic rocks into which euhedral minerals protrude.
Pumiceous	Tube-like elongate vesicles that result from rising gases. Cylindrical bodies highly
Diktytaxitic	charged with vesicles. Vesicles completely or partially filled with secondary minerals.
Miarolitic	Large ovoid structures representing gas bubbles in devitrified rhyolitic glass. A
Pipe vesicles	parallel structure developed as the result of flow. Planar parallelism. Alternating
Vesicular pipes	planar layers. Linear parallelism. An inclusion of country rock. A single-crystal
Amygdaloidal	foreign inclusion.
Lithophysae	Concentric fracture pattern resulting from contraction of some volcanic glasses upon cooling.
Flow	Comprised of fragments.
Foliation	Ovoid blobs created by liquid immiscibility, mingled magmas, or filled vesicles. Ovoid masses of
Banding	radiating crystals, commonly concentrically banded, found in some granites. Centimeter-scale
Lineation	sub-parallel to dendritic growth of olivine crystals in some quenched ultramafics.
Xenolith	
Xenocryst	
Perlitic	
Pyroclastic	Glassy lapilli.
Ocelli	Hair-like strands of glass.
Orbicules	Compressed pumice fragments in a tuff.
Spinifex	
Pyroclastic Terms	
See Section 2.5.	
Pyroclastic Glass textures	
Pele's tears	
Pele's hair	
Fiamme	

rods go extinct in unison, indicating that they are all parts of a single quartz crystal. Myrmekites are very common in granitic rocks, and occur preferentially where plagio-clase is in contact with K-feldspar. Myrmekites appear to have grown from the plagioclase/K-feldspar boundary into the K-feldspar. As the plagioclase replaces the K-feldspar, SiO_2 is released (the anorthite component of plagioclase contains less SiO_2 than the K-feldspar), thereby producing the quartz. Myrmekite commonly forms during cooling of granitic rocks, but may also occur in metamorphics. For details of the process see Shelley (1993, pp. 144-147) and Collins on his web site (http://www.csun.edu/~vcgeo005).

Devitrification is the secondary crystallization of glass to fine-grained mineral aggregates. Glass is an inherently unstable material, and is readily replaced by more stable minerals when kinetics permit. Water-quenched basaltic glass may be replaced by brown, optically isotropic oxidation-hydration products collectively known as **palagonite.** More silicic glassy rocks commonly devitrify to produce a microgranular mass of small, equidimensional grains of interlocking feldspar and silica minerals called **felsitic** texture (Figure 3-18b). Crystal form is entirely suppressed in felsitic rocks, and the texture looks very much like chert. Devitrification of glass may also produce radial aggregates of crystals (commonly cristobalite or tridymite plus feldspar) called **spherulites. Lithophysae** are large cavities bordered by spherulitic growth. Common in rhyolites, they probably represent late volatile releases that created a gas bubble in the glass. Spherulites may be found imbedded in a fel-site matrix.

3.2.5 Deformation

Deformation of solid rock can result in a variety of textures, most of which we will cover in the metamorphic portion of the text. For now, we will be concerned only with common types of post-crystallization deformation of igneous rocks. The result is bent, broken, and crushed crystals or fragments. Foliations are created in many igneous rocks that remain at depths and temperatures where they are sufficiently ductile to deform readily. Compaction of pyroclastic deposits creates a flattening effect and typically imparts an enhanced foliation. Deformation can also produce **undulose extinction,** a wavi-ness in the optical extinction pattern due to minor bending of the crystal lattice. It is not to be confused with compositional zoning, which results in a distinctly *concentric* pattern of varying extinction. Quartz is notoriously susceptible to the development of undulose extinction, and may exhibit it when other minerals do not. Undulose extinction is commonly used to distinguish quartz from non-undulatory feldspars when observing a thin section under low power.

3.3 A GLOSSARY OF IGNEOUS TEXTURAL TERMS

Table 3-1 is a brief glossary of igneous textural terms, designed to assist you in describing hand specimens and thin sections. The terms may seem strange to you now, but they serve to describe most of the common characteristics of igneous rocks. Terms are grouped into categories to help you find the one you're looking for. Petrography is an iterative process. It is not enough to look once and work through a checklist. Rather, at each step it is advantageous to reassess all previous information accumulated on a specimen in light of the latest observations. During this process, be sure to check for possible relations between pieces of information, as well as for consistency of the data.

A good approach to characterizing a rock is to describe the hand specimen in terms of the most general attributes such as color, crystallinity, felsic or mafic, and the most general textures (foliation, porphyritic, etc.). Then go on to determine the mode and describe individual minerals, their grain size (and variance in grain size), as well as shape, intergrowths and specific textures. Be *descriptive* here, but make deductions concerning origin of textures when you feel it is appropriate. Just be clear on the distinction between *observation* and *interpretation* when you do so. Then make any final general deductions you like and name the rock. Be sure you use a compositional term from Chapter 2 in the name. Textural features that you consider important or wish to emphasize features may be included (i.e., vesicular *basalt, rhyolitic* tuff, or porphyritic *monzonite).*

Avoid textural terms that make unwarranted assumptions regarding the genesis of a specimen. Some textures (such as foliations, for example) may occur in a number of ways. If you pick a genetic term (such as flow banding) be sure you can support your choice. It is always a good idea in science to separate your observations from your interpretations.

SUGGESTED FURTHER READINGS

Bard, J. P. (1986). *Microtextures of Igneous and Metamorphic Rocks.* Reidel. Dordrecht. Hibbard, M. J. (1995). *Petrography to Petrogenesis.* Prentice Hall. Englewood Cliffs, N.J. MacKenzie, W. S., C. H. Donaldson, and C. Guilford (1982). *Atlas of Igneous Rocks and Their Textures.* John Wiley & Sons. New York. McBirney, A. R. (1993). *Igneous Petrology.* Chapter 4: Igneous minerals and their textures. Jones and Bartlett. Boston. Shelley, D. (1993). *Igneous and Metamorphic Rocks Under the Microscope.* Chapman & Hall. London. Williams, H., F. J. Turner, and C. M. Gilbert (1982). *Petrography: An Introduction to the Study of Rocks in Thin Section.* W.H. Freeman. San Francisco.

Chapter 4

Igneous Structures and Field Relationships

Igneous rocks can be studied on scales from the microscopic to the global. Chapters 2 and 3 dealt with the hand specimen and microscopic scales. The present chapter explores the middle scale: that of the outcrop, landform, and deposit. We shall begin with the volcanic products, perhaps because they are the most dramatic and familiar. Then we shall proceed to the intrusive bodies and their structure, shape, and relationship to the country rocks. There is a tremendous accumulation of literature on these subjects, particularly on the volcanic side, and the present discussion can only be a condensed summary. For more detailed descriptions, turn to some of the excellent works listed at the end of the chapter.

4.1 EXTRUSIVE, OR VOLCANIC, PROCESSES, PRODUCTS, AND LANDFORMS

As I sit here in eastern Washington State, writing this textbook, I sit atop flows of basaltic lava that cover over 165.000 km^2, with a cumulative average thickness of more than a kilometer. The great bulk of these outpourings was extruded over the land surface 14 to 16 Ma ago, and must have been absolutely catastrophic. Mt. St. Helens erupted in 1980, darkening the skies and covering a quar ter of the state in ash. Many here remember the distant rumble and the ominous cloud that imposed an eerie darkness by mid-morning. In the American West, volcanism is much more than an academic subject. It cre ated much of our landscape and provides a great deal of recreation and scenic beauty. It also remains a constant threat.

4.1.1 Properties of Magma and Eruptive Styles

The style of volcanic eruption, and the resulting deposits, are determined by the physical properties of the magma, in particular, the viscosity and volatile (gas) content. Viscosity (the resistance to flow) is determined by the composition and temperature of the magma. The strong Si-O and Al-O bonds in silicate melts can link together (or **polymerize)** to form extensive networks. Recent work (see Section 7.5.1) has indicated that melts, at least near their melting point, have structures vaguely similar to the minerals that would crystallize from them (although less ordered and less polymerized). Remember that each Si atom is bonded to four oxygens in silicate minerals. Each oxygen can be bonded to another Si (called "bridging" oxygens, which create -Si-O-Si-O polymers) or, more weakly, to some other cation ("non-bridging" oxygens). Because melts approximate the structures of the corresponding minerals, basaltic melts tend to have more isolated Si-O with no bridging oxygens (as in olivine) and partially bridged -Si-O-Si-O- chains (as in pyroxene), than fully bridged three-dimensional -Si-O-Al-O- networks (as in plagioclase). Because they have much more feldspar, rhyolitic melts tend to be dominated by three-dimensional -Si-O-Al-O- networks. The more polymerized -Si-O-Al-O- networks means that there are more strong interconnected bonds in the rhyolitic melt, and thus higher viscosity. Greater viscosity thus generally correlates with higher silica content. Also, for a given SiO_2 content, polymerization is weaker at high temperatures and increases toward the melting point and with increasing crystallinity. Viscosities range from about 10 poise for anhydrous olivine basalt at 1400°C to about 10^5 poise for anhydrous rhyolite melt at the same temperature (Figure 4-1 a). Rhyolite viscosity increases to about 10^8 poise at 1000°C.

When crystals form, the viscosity of resulting melt-crystal mixtures increases rapidly as they cool, owing perhaps to the rigid crystals or surface adsorption effects (Figure 4-1 b). When stationary, such crystal-melt mixtures also develop a certain resistance to induced flow. This resistance, called **yield strength,** must be overcome, before the material can deform and behave as a viscous fluid again.

H_2O and alkalis have the ability to break down some of the polymerized networks into shorter segments (Section 7.5.1), thereby reducing viscosity. Adding 2 weight

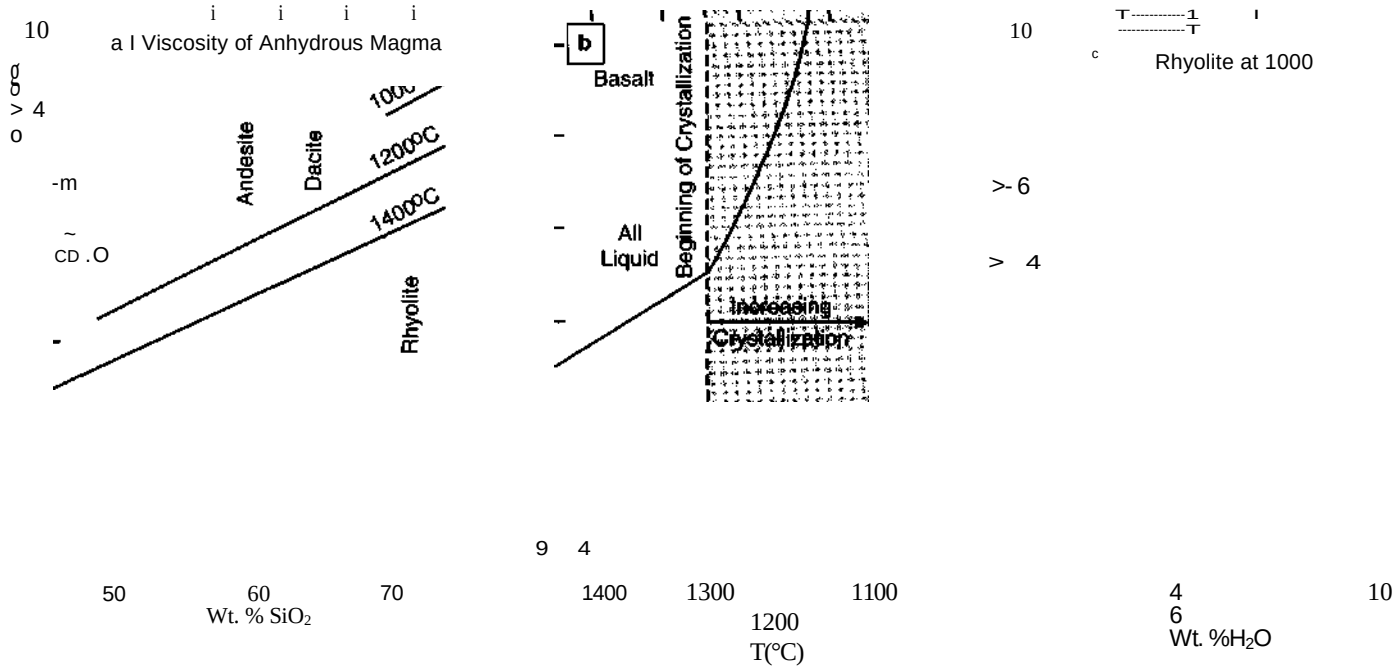


Figure 4-1 (a) Calculated viscosities of anhydrous silicate liquids at one atmosphere pressure, calculated by the method of Bottinga and Weill (1972) by Hess (1989). **(b)** Variation in the viscosity of basalt as it crystallizes (Murase and McBirney, 1973). Copyright © The Geological Society of America, Inc. **(c)** Variation in the viscosity of rhyolite at 1000°C with increasing H_2O content (Shaw, 1965). Reprinted by permission of the American Journal of Science.

% (wt. %) H_2O to rhyolite melt at 1000°C lowers the viscosity from 10^8 to 10^6 poise (Figure 4-lc). It requires about 8 wt. % H_2O to lower the viscosity to 10^4 poise, so the effect of adding H_2O gradually tapers off. Volatiles are an important constituent of magmas for yet another reason. As the magma rises, the pressure is reduced, and the volatile constituents escape from solution in the magma and expand. The gas pressures can become higher than the confining pressure of the surrounding rocks at shallow levels. This can contribute to very explosive eruptions. H_2O and CO_2 are the dominant volatile species, but there may also be SO_2, H_2, HC1, $C1_2$, F_2, and a number of other constituents. Volatile content may range from less than 0.5 wt.% in some basaltic magmas to over 5 wt.% in some rhyolitic ones, and considerably more in carbonatites and some other exotic melts. As explained more completely in the Section 7.5.1, 5% may not seem like much, but most volatiles have low molecular weights, so their *molecular* proportions may be several times the reported percentages by *weight.* Volatile pressure can thus be substantial.

The combination of viscosity and volatile content determines whether a volcanic eruption will be violent or quiescent. Volatiles diffuse fairly easily through silicate magmas, especially if polymerization is low. Because volatiles are also of low density, they concentrate at or near the top of shallow magma chambers. As a result, the initial stages of most eruptions are more violent than the later stages, because the volatile-rich upper portions are the first to be expelled. Pressure keeps the volatiles dissolved in the magma while at greater depths, but they are released when the magma approaches the surface and pressure is relieved, in a manner similar to the uncapping of carbonated soda. If the viscosity is low enough, the volatiles can escape easily. Basaltic erup-

tions in Hawaii commonly exhibit early volatile-driven lava fountaining, followed by a calmer phase in which a lava lake forms and latent gases escape gradually, or locally and suddenly as large bubbles or bursts, throwing incandescent blobs of **spatter** into the air. At the other extreme, the high viscosity of many rhyolitic and dacitic magmas resists volatile escape until the pressure of the expanding gases overcomes the high resistance, resulting in tremendous explosive eruptions. The 1883 explosion of Krakatau in Indonesia was heard near Mauritius, over 4800 kilometers away in the Indian Ocean, and spread a cloud of ash around the globe, lowering atmospheric temperatures everywhere for several years.

Although most of the dissolved volatiles are expelled during pressure release associated with the initial eruption, some remain in the magma. The soda pop analogy continues to be a useful illustration. If you shake the can and open it, the pressure release will make a rather messy initial eruptive episode. If you pour the remaining drink into a glass, however, you will be able to watch more carbonated bubbles form and rise to the surface for several minutes. In a similar fashion, any dissolved gases remaining in the lava after the first eruption will leave solution and form bubbles, called **vesicles,** which rise and concentrate toward the surface of lava flows. Highly vesicular basalt usually results from rapid vesiculation during explosive eruptions; it is called **scoria.** More viscous magmas, such as rhyolite (both because they generally contain more volatiles, and they more effectively trap the bubbles) can become so vesicular that blocks of the cooled rock actually float on water. This light-colored frothy glass is called **pumice,** and it generally forms as bits of magma included in explosive ejecta, but also at the top of some rhyolitic flows. Long after solidification of lava flows, vesicles may fill with later minerals (such

as opal, prehnite, calcite, or zeolites) deposited by hy-drothermal solutions. Such filled vesicles are called **amyg-dules** and are common in mafic flows.

4.1.2 Central Vent Landforms

Magma can issue from either a central vent or from a linear fissure. In **vent** eruptions, lava issues from a generally cylindrical, pipe-like conduit through a subcircular surface hole (the vent). There may be a bowl or funnel-shaped depression at the vent, called a **crater.** The rising magma follows weakened zones, fractures, or fracture intersections in the brittle shallow crust. Magma solidifies in the vent and plugs it, strengthening the old weakness and forcing later eruptive phases to seek other conduits nearby. There may thus be more than one vent associated with a single volcano, particularly if we consider successive eruptions over a sufficient time span. **Fissure** or **rift** eruptions occur when magma of low viscosity escapes from long fractures that tap a magma source. These eruptions are common in plateau or flood basalts (Chapter 15).

Figure 4-2 illustrates some examples of the major types of volcanic landforms associated with a single vent and their approximate relative sizes. **Shield** volcanoes range in size from a few kilometers across to the largest vent landforms (Figure 4-2a). The lavas that comprise a shield volcano are predominantly basalts. Because of the relatively low viscosity, flows predominate, and cover a large area, producing a landform with a convex upward profile and low slope (generally less than 10° and commonly closer to 2 to 3°). Over some "hot spots" (Chapters 14 and 15) basalts issue forth in vast quantities, producing the largest single landforms on Earth. The Hawaiian shield volcanoes, Mauna Loa and Mauna Kea, rise nearly 9 kilometers from the floor of the Pacific Ocean, Mauna Loa to an elevation of 4169 meters above sea-level. The total volume of basalt in these shields exceeds 40,000 km^3. Individual flows have traveled up to 50 kilometers from the source vent, and are typically only 5 to 10 meters thick. Although many flows in shield volcanoes come from the central vent, there may also be marginal **flank eruptions.** Fissure eruptions, in which lavas issue from cracks in the swelling shield, are also common. Several small satellite vents may occur along a fissure as local effusive concentrations.

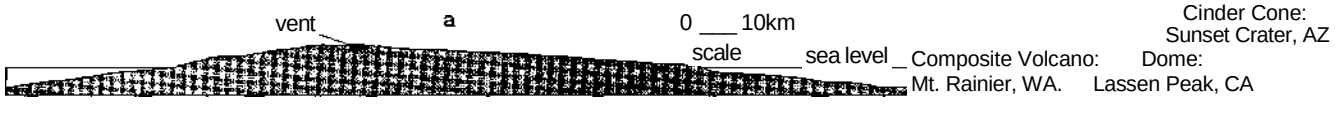
Another common volcanic landforn is the **composite volcano, or stratovolcano** (Figures 4-2b and 4-3). These steep-sided cones are usually slightly concave-upward and have slopes up to 36°. They are the volcanic land-form that we typically associate with volcanoes represented by such famous examples as Mt. Fuji, Mt. Rainier, Mt. Vesuvius, and Mt. St. Helens. They average about 2 kilometers in height, and are only V_{100} the volume of a large shield.

As we shall see in Chapters 16 and 17, stratovolcanoes are typically composed of a fairly large range of magma compositions, even at a single locality. They are generally more silicic than shields, with andesite as the most common magma type. Their eruptive histories are generally complex, having repeated eruptions of both flows and pyroclastics, the layering of which give the volcanic form its name. The ratio of flows to pyroclastic deposits varies considerably among volcanoes, and may also vary over time at a given volcano. Flows are generally associated with the more mafic and hotter magmas, whereas the more silicic and cooler types produce explosive eruptions. The 1980 eruption of Mt. St. Helens, for example, was dacitic and locally devastating. Whereas some volcanic centers may be rather monotonous, either in terms of composition, eruptive style, or both, others may show considerable variety. The underlying reasons for this difference is not clear.

Each eruptive phase will leave some solidified magma filling the central vent and conduit. Successive magmas must move through, or around, earlier plugs. Shallow magma chambers may send offshoots to feed flank eruptions, forming parasitic cones, as illustrated in Figure 4-3.

Although the volcano in Figure 4-3a looks symmetrical and is strictly aggradational (continuously built up, or added to), real composite volcanoes are far more complex. Mass wasting, glacial erosion, and explosive destruction of an earlier edifice are important elements in the history of these volcanoes (Figure 4-3b). Many present-day composite volcanoes are merely the most recent edifices in a series of cones that were constructed, and subsequently destroyed, at the same general site. At some localities the "site" may be somewhat distributed, and volcanism is spread out over a larger area with sporadic activity lasting thousands to hundreds of thousands of years. Such occurrences are commonly called **volcano complexes.** Figure 4-4 is a schematic cross section of the Lassen Peak area in northeastern California. Note the removed mass of the original andesitic Brokeoff Mountain, and the later occurrence of several dacitic and rhyolitic domes.

Smaller volcanic landforms are associated with more limited eruptive events, and are not really comparable



Shield Volcano: Mauna Loa, HI **Figure 4-2** Volcanic

landforms associated with a central vent (all at same scale).

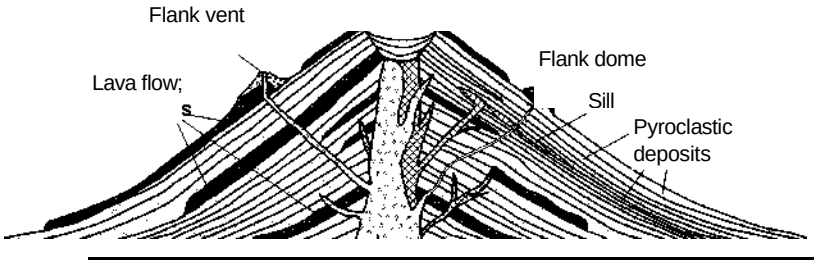




Figure 4-3 (a) Illustrative cross section of a stratovolcano. After Mac-donald (1972). **(b)** Deeply glaciated north wall of Mt. Rainier, WA, a stratovolcano, showing layers of pyro-clastics and lava flows.

to the large shields or composite cones which build up over time as a result of successive eruptive phases. Examples of some of these landf orms are illustrated in Figure 4-2 and in Figure 4-5 (where they are not at the same scale). **Pyroclastic cones,** such as **scoria cones or cinder** cones (Figures 4-2c, 4-5a, and 4-6c), result from the collection of airborne ash, lapilli, and blocks as they fall around a central vent in association with weak explosive activity. They are typically less than 200 to 300 meters high and 2 kilometers in diameter, and generally last only a few years to tens of years. Paricutin, for example, a scoria cone in central Mexico, grew in a cornfield to 410 meters in a few years beginning in 1943. Activity ceased in 1952. These small cones are usually basaltic, and straight-sided, with slopes at approximately 33°, the angle of repose of the loose scoria. These small cones are typically asymmetrical, either elongated along a fissure or larger on the side downwind at the time of the eruption. They have a central, bowl-shaped crater, which gradually fills and flattens as a result of mass wasting. A **maar** (Figures 4-5b and 4-6a) is typically lower than a scoria cone, and has a much larger central crater, relative to the deposited ring of debris. Maars result from the explosive interaction of hot magma with groundwater, which flashes to steam. Such explosions are called **hydromagmatic, or phreatic** eruptions. Note that the explosive power here is supplied by groundwater, not water contained in the melt. Geologists commonly use the terms "meteoric" and "juvenile" to make the distinction clear. **Meteoric** water refers to surface or groundwater, and **juvenile** refers to either water or other constituents that are products of the magma itself. A maar is primarily a negative feature, in

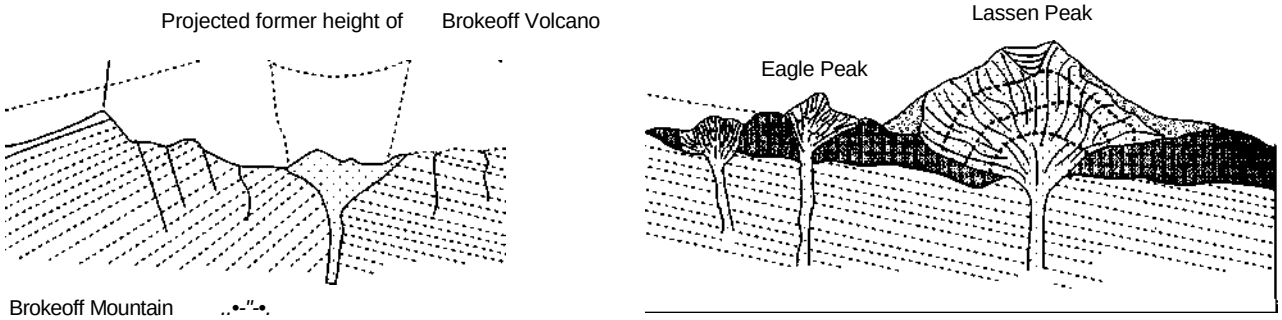


Figure 4-4 Schematic cross section of the Lassen Peak area. After Williams (1932).

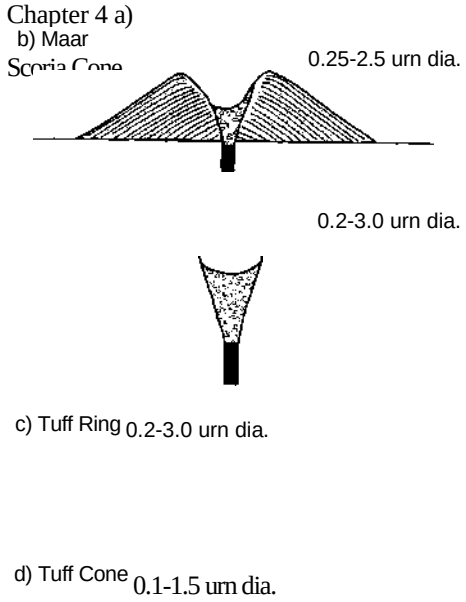


Figure 4-5 Cross-sectional structure and morphology of small explosive volcanic landforms with approximate scales. After Wohletz and Sheridan (1983). Reprinted by permission of the American Journal of Science.

that the phreatic eruption excavates a crater into the original substrate. Tuff rings and tuff cones also form as a result of magma-H_2O interactions, but neither have deeply excavated craters. **Tuff rings** (Figure 4-5c) form when rising magma (usually basaltic) comes closer to the surface than with a maar before interacting explosively with shallow ground or surface water. They also involve a higher ratio of magma to H_2O than maars, forming a subdued ring of scoria and ash that has a low rim, and layering of pyroclastic material that dips inward and outward at about the same angle. Perhaps the most famous tuff ring is Diamond Head, which stands above the southeast end of Waikiki Beach (Figure 4-6b). **Tuff** cones (Figure 4-5d) are smaller than tuff rings, with steeper sides and smaller central craters. They form where magma interacts with very shallow surface water. They seem to result from less violent and more prolonged eruptions than maars or tuff rings. They look like scoria cones, but have bedding that dips into the craters as well as outward.

Most classifications, including the one above, give the impression that the "pigeonholes" are distinct, reflecting discontinuities in the processes that generate the forms. Although there may be some clustering of forms into the types described above, there is a continuous spectrum of these small pyroclastic landforms reflecting variations in viscosity, composition, the ratio of magma to meteoric water, and the depth of the explosive interaction. Even shields and composite cones are not completely distinct.

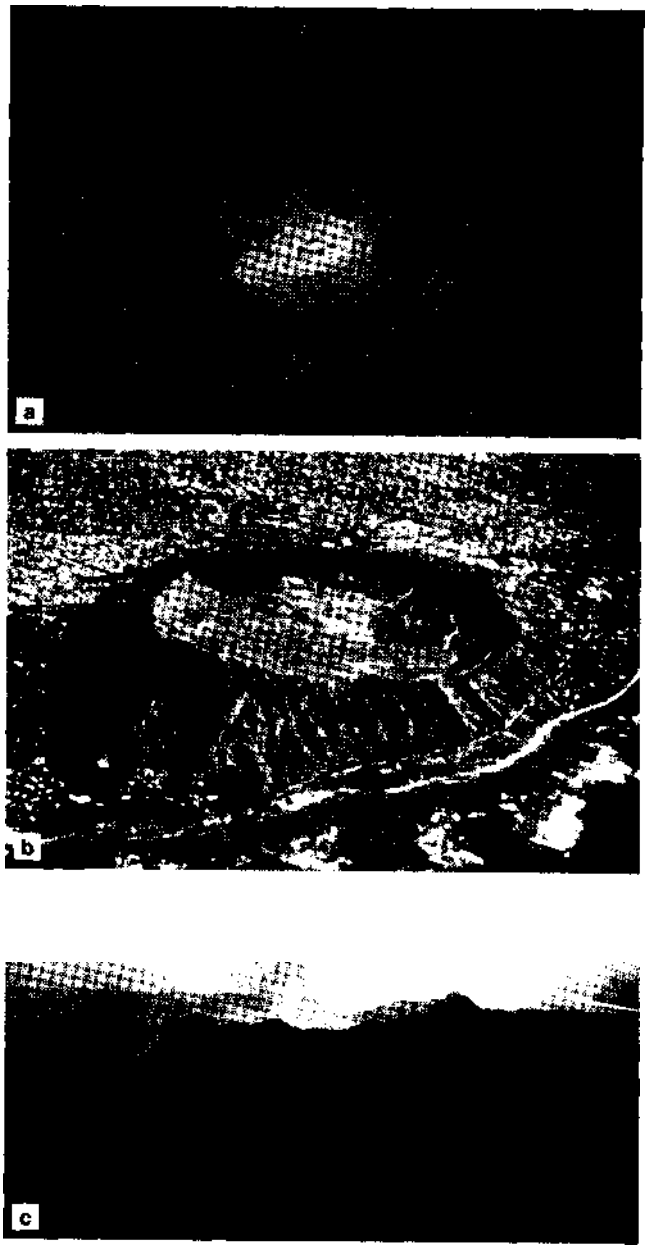


Figure 4-6 (a) Maar: Hole-in-the-Ground, Oregon (courtesy of USGS). (b) Tuff ring: Diamond Head, Oahu, Hawaii (courtesy of Michael Garcia), (c) Scoria cone, Surtsey, Iceland, 1996 (courtesy of Bob and Barbara Decker).

Domes (Figure 4-7) form when largely degassed, viscous, silicic magma, such as dacite or rhyolite (less commonly andesite), moves slowly and relatively quietly to the surface. Domes range in size from less than a hundred meters to several kilometers in diameter. Domes may form early or late in an eruptive cycle, but they are typically late. After an early phase of explosive activity, the final (gas-depleted) magma may inflate to a dome in the central crater. The process is usually **endogenous,** in that the dome inflates by the injection of magma from

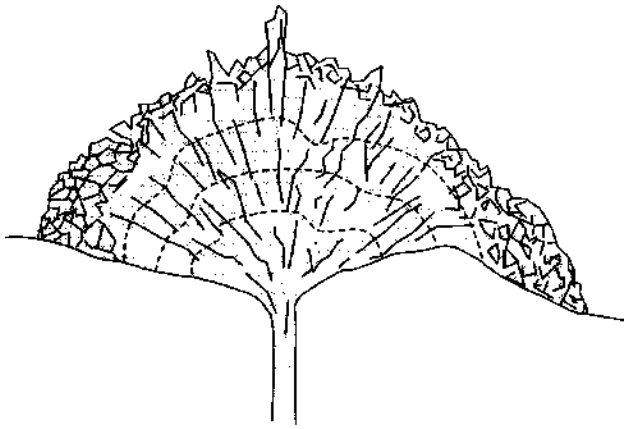


Figure 4-7 Schematic cross section through a lava dome.

within. **Exogenous** dome eruptions are events in which the later additions break through the crust and flow outward. The crust that forms on the dome brecciates as the dome inflates, giving the surface a very rough texture. Brecciated blocks fall off and accumulate at the base of the dome as an apron of talus. The inflation may fracture the outer zones in a radial fashion, and even push some sections outward or upward, forming a **spine,** as illustrated in Figure 4-7.

There is a gradation from steep-sided domes, to domes that flatten and flow downhill (called **coulees,** a name also applied to a type of canyon), to thick flows. Viscosity, slope, and rate of magma production are important factors in the final form. Coulees and thick rhyolite or obsidian flows are viscous, and don't travel far from the source. They typically have a surface that folds into "pressure ridges" that are usually convex toward the advancing front (Figure 4-8). How some domes, coulees, and thick flows can be glassy throughout is an interesting problem. Certainly they didn't cool that quickly. Perhaps



Figure 4-8 Pressure ridges on the surface of Big Obsidian Flow, Newberry Volcano, OR. Flow direction is toward the left.

the viscous and highly polymerized melt inhibits crystal nucleation, or the motion retards nucleation (as in taffy candy making). Some domes are inflated beneath the ground surface, and are called **cryptodomes.** The inflation that caused the spectacular "bulge" on the flank of Mt. St. Helens in 1980 was a cryptodome. It was a bit unusual in that it was certainly not late in the eruptive cycle, nor was it depleted in volatiles, as became clear when the bulge oversteepened, the lower portion slid away, and the unroofed and depressurized section of the dome released the gas in a very explosive event.

Calderas are large-scale collapse features that typically form at a central vent, fairly late in an eruptive episode. They form when the denser solid strata above a shallow magma chamber founder into the draining chamber. The magma may drain as a flank eruption, or it may move up the fractures separating the blocks of the collapsing roof. On basalt shields, the caldera may fill with magma from below to create a **lava lake.** In such an occurrence, the process is one in which the cool dense solid roof founders into the less dense magma chamber beneath, displacing the magma upwards.

In more silicic cases, fractures reaching down to the magma chamber reduce the pressure at the top of the chamber, inducing rapid vesiculation, and a substantial proportion of the magma may then escape along these fractures in the roof in the form of pyroclastic activity. Figure 4-9 illustrates the explosive eruption of Mt. Mazama to form Crater Lake 6850 years ago. According to Bacon (1983) and Bacon and Lanphere (1990), Mt. Mazama was a large composite volcano, consisting of several overlapping basaltic andesite shields and andesitic to rhyolitic composite cones that stood -3600 meters high. The first material from the large magma chamber that led to the climactic eruption was a rhyodicite pumice fall and lava flow dated at -7015 b.p. The climactic eruption began with a single-vent phase that produced a Plin-ian column, producing a widespread airfall tephra, then collapsed to a valley-hugging series of ignimbrites (now welded). Immediately thereafter, a ring-dike phase began, characterized by caldera collapse and fairly high pyroclastic columns issuing from the marginal faults. The columns produced lithic volcanic breccias near the vent and on nearby hilltops and poorly to non-welded pumiceous ignimbrites in the valleys. A later ring-dike phase was characterized by pyroclastic basaltic andesite and andesite scoria. The total volume of ejected material is estimated at -50 km^3. A few hundred years following the climactic eruption, small andesitic vents erupted sporadically within the caldera, producing small cones, such as Wizard Island. Whether they represent new magma or remnants from the older chamber is unknown.

The largest calderas, some over 100 km across, are associated with tremendous pyroclastic eruptions of rhyolite. The resulting calderas are too large to visualize at

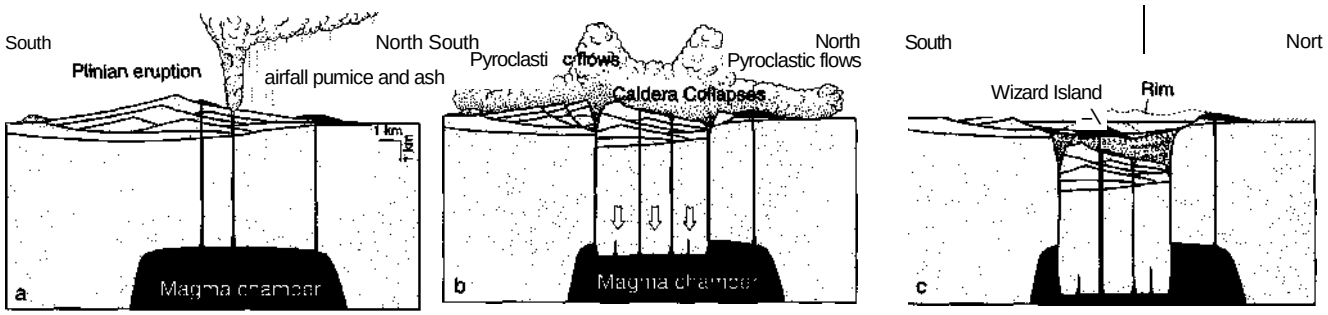


Figure 4-9 Development of the Crater Lake caldera. From Bacon (1988).

ground level, and some have only been recognized recently from satellite imagery. Examples of large calderas in the western United States are those at Yellowstone, Valles Caldera in New Mexico, and Long Valley in California. Yellowstone contains several partially overlapping calderas. Such occurrences are called **caldera complexes.** The first eruption was 2 Ma ago, and produced approximately 2500 km^3 of rhyolitic material in a single huge ash flow. The third eruption 600,000 years ago produced over 1,000 km^3 of material. Following the collapse of the third caldera, the center has risen again (perhaps as new magma refilled a chamber below) to become what is called a **resurgent caldera.**

Smith (1979) has shown a strong correlation between caldera area and volume of the expelled ash flow, which holds for caldera sizes spanning five orders of magnitude. He used caldera *area* because volume is less easy to determine in such collapsed, ash-covered, and eroded features. If this correlation reflects a similar correlation between the ash flow volume and the caldera volume, it suggests that calderas drop approximately the same distance (about 0.5 km), regardless of diameter, suggesting that some process limits the depth of collapse. Perhaps at greater depths, the lithostatic pressure is sufficient to impede volatile separation, which might disrupt the magma, lower the viscosity, and permit rapid escape.

4.1.3 Fissure Eruptions

In contrast to central vent volcanism, **fissure eruptions** occur as magma erupts to the surface along either a single fracture or a set of fractures. The planar conduits, when exposed by erosion, are filled with solidified magma, and are referred to as **feeder dikes.** As I mentioned above, some fissure eruptions occur on the flanks of central vents, where fractures typically accompany inflation of the edifice as the magma chamber fills. Such fractures may form singly, or in multiple sets (in a concentric or radial pattern about the vent, or as parallel sets). Fissure eruptions also occur in larger areas undergoing regional extension. Examples include the African Rift Valleys, the Basin and Range, Iceland, and exten-sional basins behind volcanic arcs. Figure 4-10 illustrates the location of some of the exhumed feeder dikes and vents of the Columbia River Basalts, a major plateau-type, or flood basalt province in eastern Washington. Although we think of these great flood basalts as fed exclusively by dikes, a number of localized vents also feed some flows or even form local cones. The amount of basalt in plateau or flood basalts is tremendous, and the displacement of such a large volume to the surface commonly results in sagging of the crust below, producing a structural basin. These great outpourings will be discussed in detail in Chapter 15.

The most common type of fissure eruption is never seen by most of us. These fissures typically occur beneath

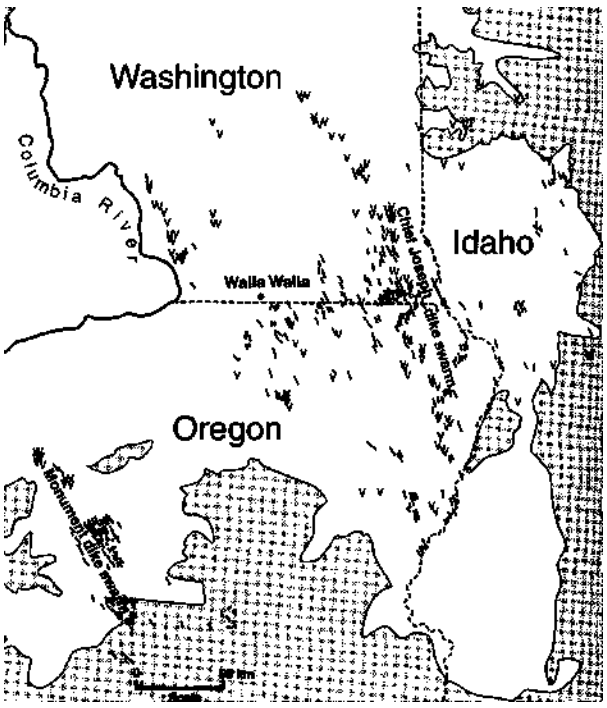


Figure 4-10 Location of the exposed feeder dikes (heavy lines) and vents (V's) of the southeastern portion of the Columbia River Basalts. Unshaded area covered by CRB. After Tolan et al. (1989). Copyright © The Geological Society of America, Inc.

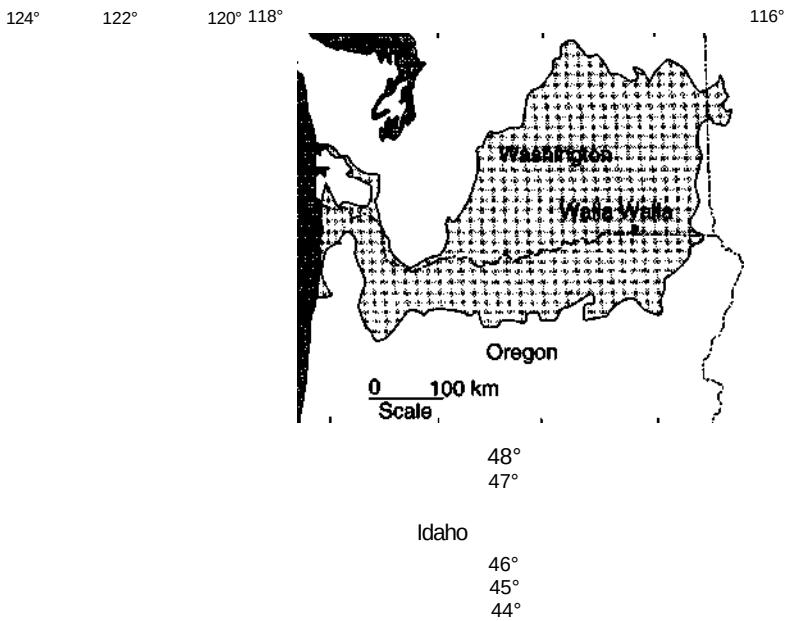
the sea, at mid-ocean ridges, and produce the most common and voluminous rock type on Earth: mid-ocean ridge basalts (MORBs), which constitute the oceanic crust. Here two plates diverge, resulting in normal faults and extensional tectonics, an ideal situation for fissure eruptions. Only at Iceland is the this type of fissure eruption exposed above the sea. We will examine MORBs and the processes which generate them further in Chapter 13.

4.1.4 Lava Flow Features

Lava flows are more quiescent than the dramatic explosive volcanic eruptions, but they are the dominant form of volcanism on Earth, and perhaps throughout the solar system. Flows occur most typically in lavas with low viscosity and low volatile content. They are thus most common in basalts, but some flows may be as silicic as rhyolite. The Hawaiian flows you may have seen generally travel fairly slowly and over broad, but limited, areas in any single event. Flows rarely kill people, but they have engulfed a lot of property, including the occasional town. Fortunately, they are usually slow enough that people can evacuate their homes before they are lost to the advancing flow.

Some flows, however, particularly those associated with flood basalts, are of enormous size. The lack of phe-nocryst alignment in many suggests that these flows typically travel in a turbulent fashion, and come to rest before they have cooled enough to be viscous and preserve any internal flow features. Some individual flow units of the Columbia River Basalt Group cover nearly 120,000 km^2 (Figure 4-11), and approach 3000 km^3 in volume (Tolan et al, 1989). Self et al. (1997) concluded that large individual flows required 5 to 50 months for emplacement, and averaged ~4000m^3/s of lava effusion. These are formidable flows, and if one were to occur in

Figure 4-11 Aerial extent of the N2 Grande Ronde flow unit (approximately 21 flows). After Tolan et al. (1989). Copyright © The Geological Society of America, Inc.



modern times it would have considerable impact on property, lives, and even the global climate.

In the early stages of a basaltic eruption, such as in Hawaii, magma emerges as incandescent lava at about 1200°C. This lava has a very low viscosity, and runs down slope in rivers with initial velocities as high as 60 km/hr. This runny lava cools and forms a smooth black surface, which may develop a corrugated, or ropy, appearance (Figure 4-12a and the left side of Figure 4-12b). The corrugations are usually less than 2 cm high, with axes perpendicular to, or convex to, the flow direction. Such lavas are called **pahoehoe.** As the lava cools further, and the viscosity increases, the flows begin to move more slowly, and develop a thicker scoriaceous crust. As the fluid interior continues to move, the crust breaks up into blocks of clinkery scoria, which ride passively on the top. Pieces also tumble down the advancing front. The motion is like a conveyor belt, in which the surface slides beneath the front of the advancing flow. Thus the blocks are found both at the top and the base of the flow. The rubble-like lava flows that result are called aa (the right side of Figure 4-12b). Occasionally the fluid core material escapes and flows as tongues through the carapace. **Lava tubes** also form as conduits within many basalt flows as an efficient means of conserving heat and delivering lava to the advancing front. Lava typically drains from the conduit, leaving a tube-like tunnel. Aa and pahoehoe are end-members of a continuous series of flow-top characteristics. Pahoehoe is restricted to *basalts* of low viscosity, but aa can occur in flows spanning a range of compositions. Numerous basalt flows begin as pahoehoe, and turn to aa further from the vent as they cool and slow down.

Holcomb (1987) recently described the **inflated** flow, applying the term to some Hawaiian basaltic flows (Figure 4-12c-e). These flows begin as a thin pahoehoe flow, perhaps as thin as 20 to 30 cm. After the crust sets, continued lava is added beneath it, and inflates the flow internally to thicknesses as great as 18 m. As a result, the crust rises, cracks, and tilts in complex patterns. Inflated flows have been described in several places, including Hawaii, Oregon, the Juan de Fuca Ridge, and the Columbia River Basalts. Inflation of the Columbia River Basalts would avoid the incredibly rapid flow emplacement rates of about a week, estimated by some on the basis of cooling rate, because the initial crust would insulate the flow, slow the cooling rate, and allow the basalt to flow for greater distances (Self et al., 1997).

Andesite flows are more viscous than basalts, and, although they may occur as flows, they more typically form either as aa or **block lavas.** Whereas aa is composed of sharp cindery and scoreaceous rubble, block lavas have larger, smooth-sided blocks. Block lava flow fronts are very steep, and the block piles may be over 100 m high. To my knowledge, no one has ever observed a block lava eruption, but they appear to result from a process in

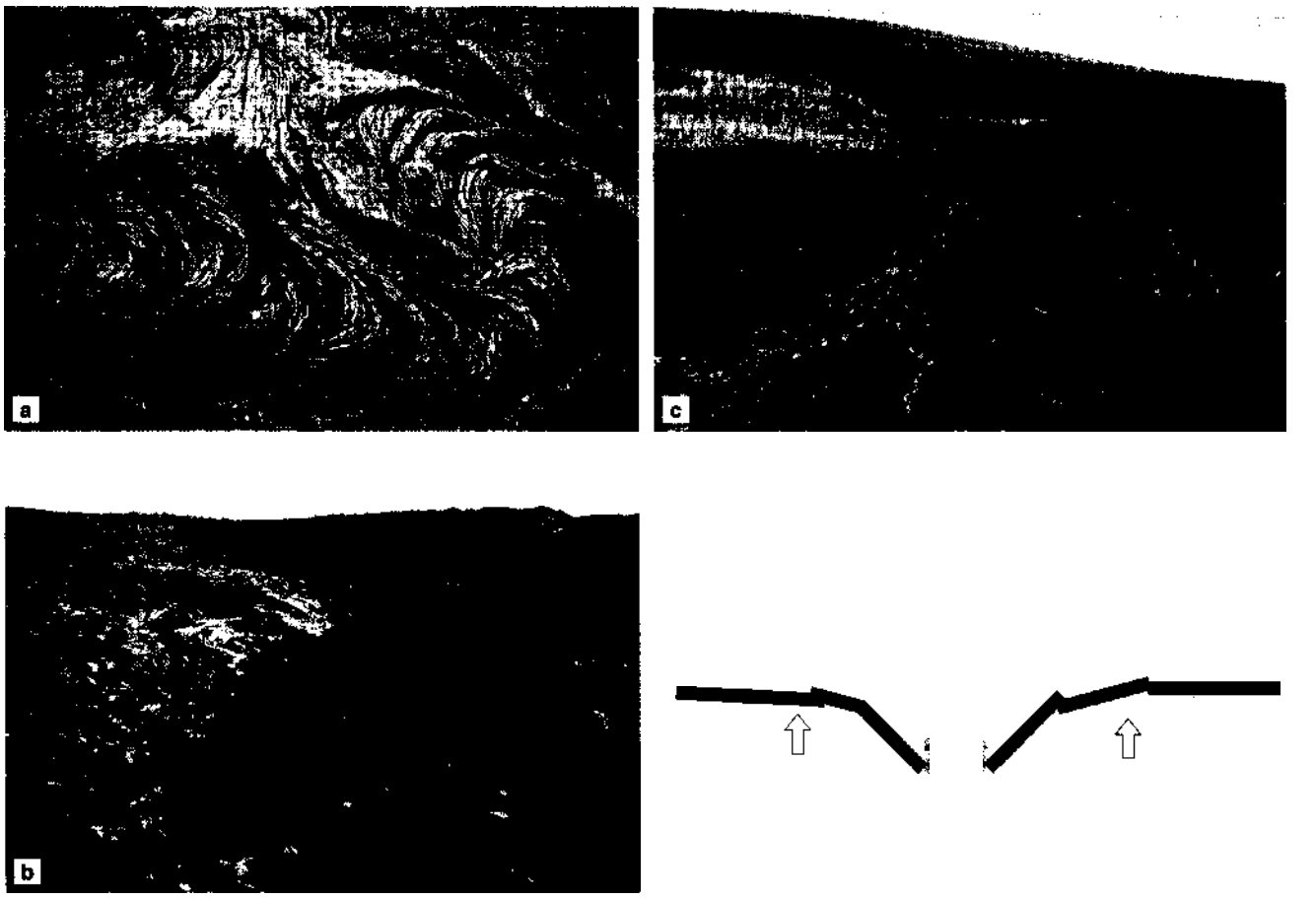


Figure 4-12 (a) Ropy surface of a pahoehoe flow, 1996 flows, Kalapana area, Hawaii, (b) Pahoehoe (left) and aa (right) meet in the 1974 flows from Mauna Ulu, Hawaii, c. Inflated lava flow north of Kilauea (in background), **(d-e)** Illustration of the development of an inflated flow. In (d), a thin flow spreads around a rock wall. In (e), the flow is inflated by the addition of more lava beneath the earlier crust. A old stone wall anchors the crust, keeping it from lifting. The wall can be seen in the low area in part (c).

which the lava is nearly solid, and the blocks are literally pushed out a vent. In larger flows, a more massive flow may form behind a blocky front.

Dacitic and rhyolitic lava flows are less common, as these magmas are typically more explosive and produce pyroclastic deposits. When silicic flows do occur, they form domes, coulees, or thick flows, as discussed earlier. Rhyolite flows are commonly composed of obsidian and are **aphyric** (lacking in phenocrysts), indicating that they were hot (and hence less viscous).

Thick intermediate to silicic lavas typically exhibit **flow foliation,** which may consist of aligned phenocrysts, bands of different color (Figure 3-13), or pumice bands. These layers could have been different batches of mingled magmas or portions of the same magma with a different temperature, composition, or content of crystals, H_2O, or oxygen. The layers were then stretched, sheared, and/or folded during flow. In the case of pumice layers in obsidian, these layers may have been frothy surface layers incorporated back into the flow, but are more likely zones of localized shear that induced vesiculation.

Subaerial lava flows (those that flow on land), and some shallow sheet-like intrusions, may develop a characteristic jointing pattern called **columnar joints.** Figure 4-13 illustrates the structures within a typical basalt flow (in this instance, one of the Columbia River Basalt flows). In the ideal case, there are four subdivisions: a thin vesiculated and brecciated flow top, an upper and lower **colonnade** with fairly regular straight columns, and a central **entablature** that has more irregular columns that are typically curved and skewed. The three main subdivisions are not uniformly developed, in that each may vary greatly in thickness, be absent entirely, or occur repeatedly within a single flow. Although models have been proposed for the development of columnar joints based on convection currents (Sosman, 1916; Lafeber, 1956), or diffusion (Kantha, 1981; Hsui, 1982), the most widely accepted mechanism is based on contraction of the flow as it cools (Tomkeieff, 1940; Spry, 1962; Long and Wood, 1986; Budkewitsch and Robin, 1994). Because the top and bottom of the flow cool before the central layer, these outer areas contract whereas the center

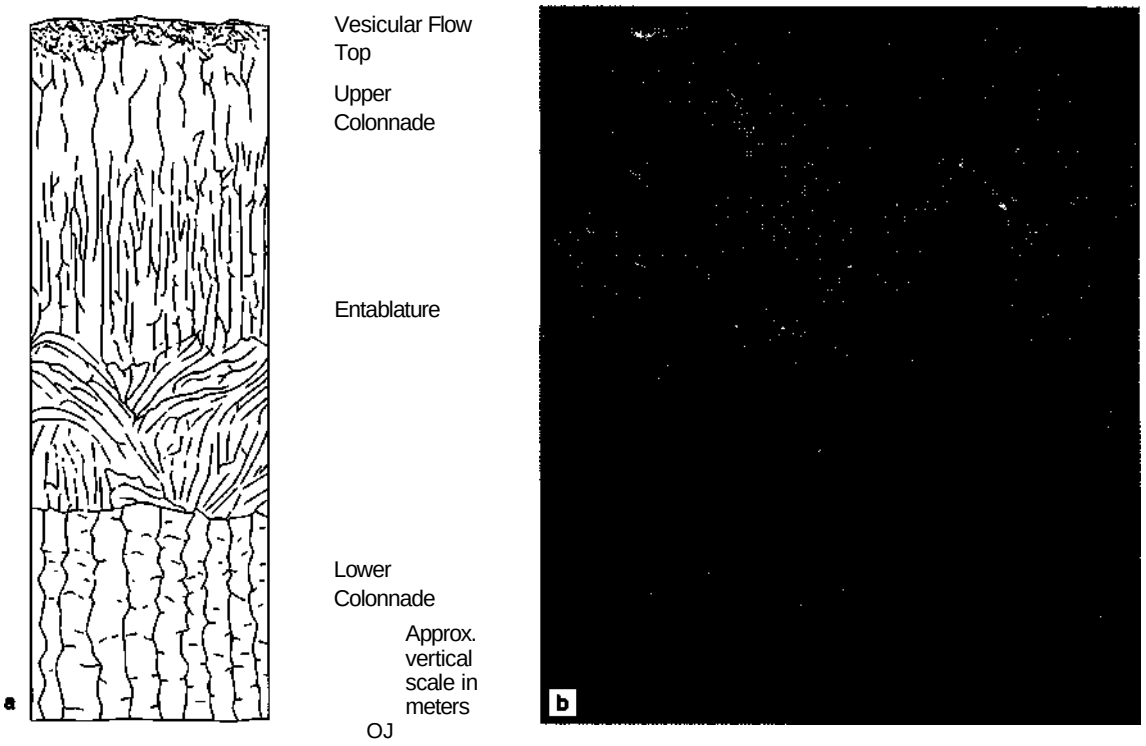


Figure 4-13 (a) Schematic drawing of columnar joints in a basalt flow, showing the four common subdivisions of a typical flow. The column widths in (a) are exaggerated about 4 x. After Long and Wood (1986). (b) Colonnade-entablature-colonnade in a basalt flow, Crooked River Gorge, OR.

doesn't. This results in tensional stresses that create regular joint sets as blocks pull away from one another to create polygons separated by joints. The joints propagate down from the top and up from the bottom as cooling progresses toward the center, forming the column-like structures in the colonnades. Columns form perpendicular to the surfaces of constant temperature, which are usually parallel to the surfaces of the flow (i.e. horizontal). Columns may be 4-, 5-, 6-, or 7-sided, but 5-and 6-sided ones dominate.

The irregular entablature is more difficult to explain, with ideas generally relying on disturbed cooling surfaces or post-joint deformation of the still ductile flow center. Petrographic textural differences between the colonnade and the entablature (Swanson, 1967; Long, 1978, Long and Wood, 1986) suggest that the difference results from primary crystallization effects. The greater proportion of glassy mesostasis, and the feathery Fe-Ti oxides in the entablature suggested to Long and Wood (1986) that the entablature cooled more rapidly than the colonnades, contrary to what one would expect for the more insulated flow interior. They proposed a model in which water infiltrated along the joints (perhaps associated with lakes or floods) to cool the entablature portion by con-vective water circulation through the upper joint system. Variation in the development of the subdivisions might then be explained by the erratic nature of flooding, and the fanning of columns in the entablature may be related to the preferential percolation of water down large, early-formed joints.

When basaltic lava flows enter standing water, they form either tongues or more equidimensional blob-like structures, both of which are called **pillows.** The resulting **pillow lavas** (Figure 4-14) contain numerous packed pillows, usually with glassy rinds, concentric bands, and



Figure 4-14 Sub-aqueous pillow basalts (hammer handle for scale in center foreground). Olympic Peninsula, Washington.

radial fractures within. Later pillows commonly flatten and conform to the shape of the pillows upon which they settle, providing geologists with a method of determining the original top of the deposit in folded terranes. The pillows may be embedded in fine glassy debris called **hyalo-clastite,** produced by fragmentation of the hot lava as it contacts the water. The hyaloclastite commonly devitri-fies quickly to an orangish-brown material called **palag-onite.**

4.1.5 Pyroclastic Deposits

The term **volcaniclastic** refers to any fragmental aggregate of volcanic material, regardless of how it formed. **Autoclastic** refers to volcanics that suffer a quiescent self-imposed breakup, such as aa and block flows, dome talus aprons, or gravitational collapse features. **Pyroclastic** deposits are a subset of volcaniclastics that consist of fragmented material formed from explosive volcanic activity or aerial expulsion from a volcanic vent. Other volcaniclastic deposits include volcanic mudflows, or **lahars,** which form when volcanic debris mixes with sufficient water, either rain or melted snow/ice, to mobilize. Lahars may be eruption-associated and hot, or occur later and be cold debris flows off volcanoes. Hot dome collapse features are generally considered pyroclastic as well, though they do not technically qualify as such. Pyroclastic deposits may be deposited in water also. They may then mix with the water to become lahars or water-laid deposits (and are thus no longer pyroclastic by our definition). Collectively, the pyroclastic *particles* that comprise the deposits are called **pyroclasts,** and a collective term for the deposited *material* is **tephra.** Further classification of pyroclasts is based on particle size and is discussed in Chapter 2.

Pyroclastic deposits are classified on the basis of the mode of transport and deposition. They are divided into **falls and flows. Surges** are distinguished by some geologists as a third category of pyroclastic deposit, but they are really a type of flow.

1) Pyroclastic fall deposits comprise fallout from a vertical eruption. Pyroclasts may be propelled forcefully upward during an explosive eruption, or they may be carried aloft by convection and the buoyancy of the hot gases above the vent. **Plinian** eruptions (Figures 4-9a, 4-15a and 4-18a) are a combination of the two. Particles are forced from the vent by powerful jetting of gases, and are then carried aloft by hot convection. The simplest falls accompany the small basaltic eruptions that produce scoria cones and tuff rings/cones. The more dramatic and larger eruptions produce a high ash plume that may reach heights of 50 km. As the plume expands, the driving force dissipates, and the pyroclasts fall back to Earth under the influence of gravity. Larger and denser particles fall more quickly, and the finer ash remains aloft longer. As a result, fall deposits are well sorted, with grain size decreasing both vertically toward the top of a deposit and laterally away from the source vent. The size, thickness, and shape of the deposit, and the distribution of particle size, depends on the rate of expulsion, the volume erupted, the force of the explosion, and the direction and velocity of the prevailing winds at the time of eruption.

Fall deposits *mantle* the land surface. Thickness decreases gradually away from the source, but is independent of topography (much like snow). Fall deposits have a better chance to cool while suspended, so they are rarely welded by their own heat after deposition, except, perhaps, near the vent. Figure 4-15b illustrates the time-sequence development of the 1980 Mt. St. Helens ash plume. In only 13 minutes (8:32 to 8:45 A.M.), the ash cloud rose to nearly 25 km, and spread 50 km laterally, as monitored from ground stations and by satellite. The thickness of the resulting ash fall deposit is shown in Figure 4-15c, with lines of constant thickness, or **isopachs,** expressed in centimeters. Over a centimeter of ash was deposited as far as 500 km away. The ash blanket from the 6950 year-old eruption of Mt. Mazama (now Crater Lake) covered portions of seven states and part of Canada (Figure 4-16). Ash deposits 50-cm thick are found as far as 100 km from Crater Lake.

The most astonishing ash fall deposits are associated with huge rhyolitic eruptions, such as those associated with the Yellowstone (Wyoming), Long Valley (California), or Valles (New Mexico) calderas. For example, the area covered with ash from the Long Valley eruption, some 700,000 years ago, covered much of the western United States (Figure 4-17). Ash was a meter thick over a 75-km radius and 1-cm thick in Kansas. It is estimated that devastation was total over a 120-km radius (the circle in Figure 4-17). Ash from three Yellowstone eruptions over the last 2.2 Ma produced 3800 km^3 of ash and pumice, some of which reached Minnesota and Louisiana (2200 km away).

As awful as these ash deposits are, the thinner extensions, which cover the largest area, are commonly transient. For example, the ash deposits of the May 18th, 1980 Mt. St. Helens eruption created quite a mess in eastern Washington, but it is very difficult to find remnants of the ash blanket over here today. The thick deposits near the mountain are still there, but the blanket that covered much of eastern Washington is practically gone. It would appear that the deposits in the subaerial environment are quickly eroded and accumulate in sheltered lowlands, particularly those with standing water. Thus, although they originally mantle the ground surface, the deposits may eventually concentrate in depressions.

2) Pyroclastic flow deposits are left by dense ground-hugging clouds of gas-suspended pyroclastic debris (mostly pumice and ash, with variable amounts of lithic and crystal fragments). They are generated in several ways (Figure 4-18). The largest flows result from the col-

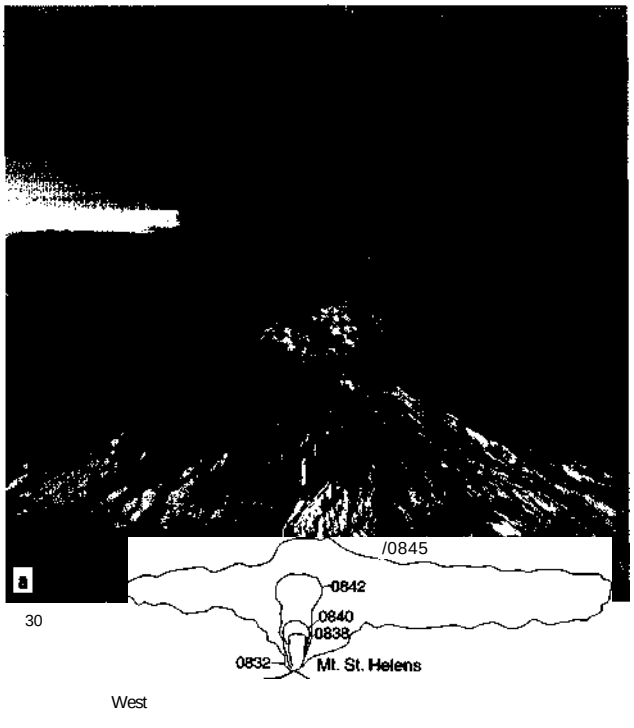


Figure 4-15 Ash cloud and deposits of the 1980 eruption of Mt. St. Helens, **(a)** Photo of Mt. St. Helens vertical ash column, May 18,1980 (courtesy USGS). **(b)** Vertical section of the ash cloud showing temporal development during first 13 minutes, (c) Map view of the ash deposit. Thickness is in centimeter. After Sarna-Wojcicki et al. (1981).

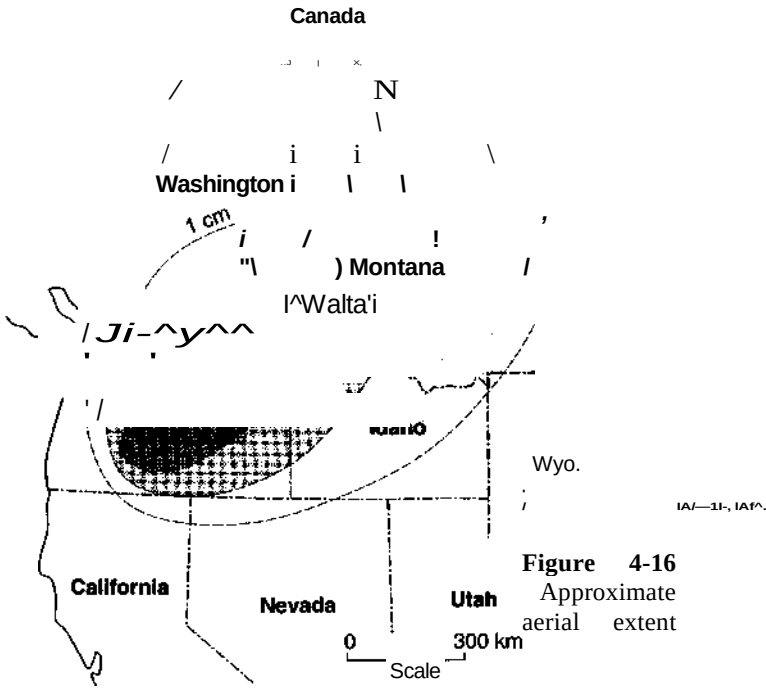


Figure 4-16 Approximate aerial extent and thickness of Mt. Mazama (Crater Lake) ash fall, erupted 6950 years ago. After Young (1990).

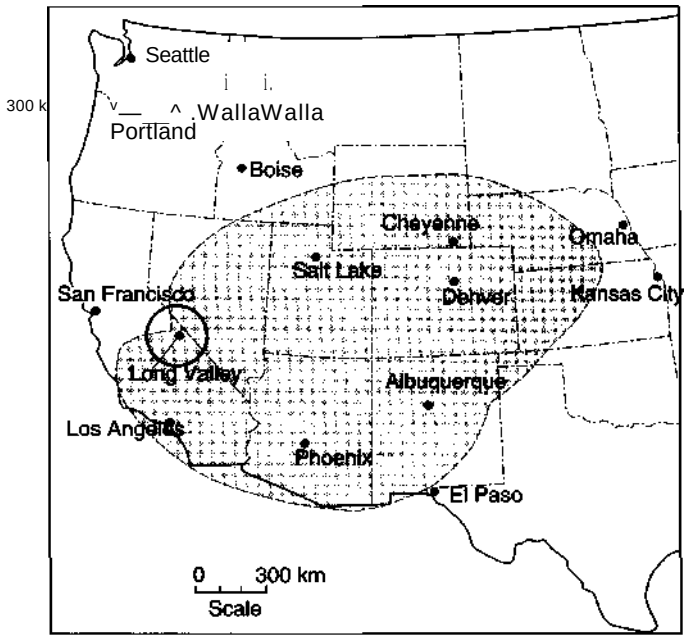


Figure 4-17 Maximum aerial extent of the Bishop ash fall deposit erupted at Long Valley 700,000 years ago. After Miller et al. (1982).

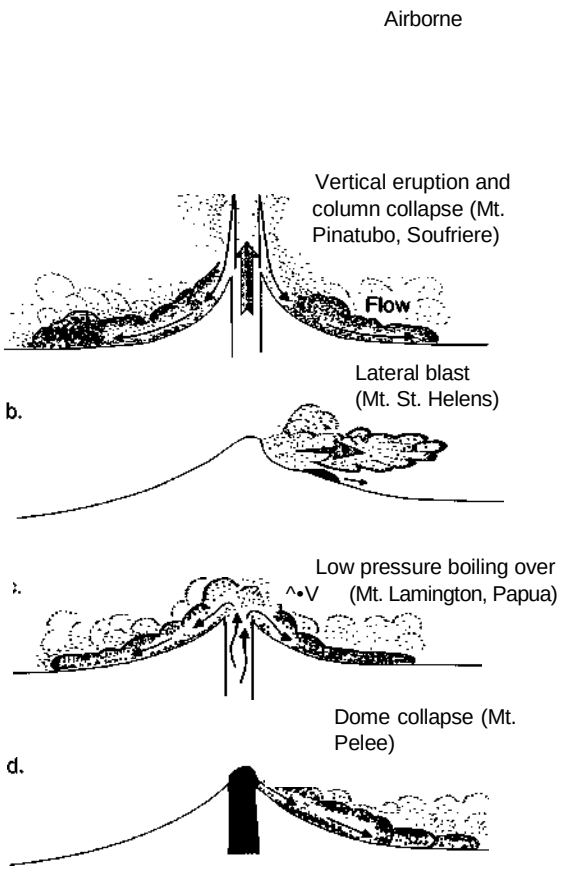


Figure 4-18 Types of pyroclastic flow deposits. After Mac-Donald (1972),'Fisher and Schminke (1984).

lapse of a vertical explosive or plinian column that falls back to earth, and continues to travel along the ground surface (Figure 4-18a). An alternative explosive mechanism is a lateral blast, such as occurred at Mt. St. Helens in 1980 (Figure 14-18b). The flows are **fluidized** because the magmatic gases and air trapped within and beneath the advancing flow cannot easily escape from the high concentration of suspended particles (see Reynolds, 1954; or Wilson, 1984 for discussions). Other pyroclastic flows result from the "boiling-over" of a highly gas-charged magma from a vent (Figure 4-18c), or the gravitational collapse of a hot dome (Figure 4-18d). In all cases, the hot particle-laden cloud is (or becomes) denser than the surrounding atmosphere, so it flows downward and deposits the pyroclasts as it loses momentum.

Pyroclastic flows are controlled by topography, and the deposits concentrate in valleys and depressions, unlike the uniform mantling of ash falls. Pyroclastic flows are hot (400-800°C). Velocities vary from 50 to over 200 km/hr (the Mt. St. Helens blast was estimated at a maximum of 540 km/hr). They can also travel long distances (typically a few kilometers, but some deposits are found over 100 km from the source). The temperature and velocity make them very dangerous phenomena. Pyroclas-tic flow terminology can be a bit confusing. The moving flow itself is called either a pyroclastic flow or an ash flow. The deposits are called **ignimbrites** (Greek for "fire cloud material"). Some authors reserve the term "ign-imbrite" for the pumice-rich varieties, using the term **block-and-ash deposits** for the unvesiculated-clast-bear-ing types. The rock name for a sample taken from the deposit is called a **tuff** (see Figure 2-5 for more specific rock names).

Pyroclastic flow deposits are generally unstratified and poorly sorted, probably because of the high particle concentration and energy of the flow. Occasionally there may be some sorting, with denser lithic blocks settling toward the bottom of the deposit, and the lighter pumice blocks floating toward the top (Figure 4-19). The deposits are emplaced at high temperatures, and the lower portions typically become **welded** by the internal heat in the settled pile to a very hard rock called a **welded tuff.** Such welded portions of a deposit are ductile for a period, and are typically compressed by the weight of the overlying mass, becoming dense and foliated. Pumice shards commonly have the vesicles squeezed out, and take on the black color of obsidian. The light gray color of pumice results from the vesicles, much like whitecaps on the sea. The flattened black bits of compressed pumice are called **fiamme.**

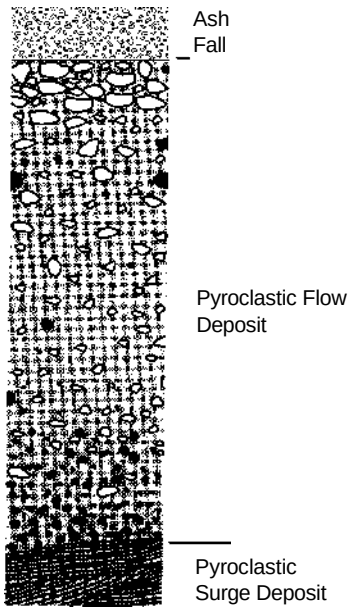


Figure 4-19 Section through a typical ignimbrite, showing basal surge deposit, middle flow, and upper ash fall cover. White blocks represent pumice, and black represent denser lithic fragments. After Sparks et al. (1973). Copyright © The Geological Society of America, Inc.

Although pyroclastic flow deposits generally cover a smaller area than fall deposits, owing to their velocity and heat they can be absolutely devastating. Several such eruptions have occurred in historic times, but, for obvious reasons, eyewitness accounts are rare. On August 25,79 A.D., ash falls and several hot pyroclastic flows descended from Mt. Vesuvius and buried the cities of Pompeii and Herculaneum, near Naples, Italy. The inhabitants were either burned or suffocated, and their bodies were buried in the soft ash, preserved to this day as molds. On May 18,1902, an incandescent pyroclastic flow from Mt. Pelee descended suddenly on the city of St. Pierre, on the island of Martinique, and instantly killed approximately 28,000 people. Only two survived, one a prisoner. The pyroclastic flows associated with the Long Valley eruption of Figure 4-17 traveled at speeds over 200 km/hr, and are estimated to have covered over 1500 km^2 to depths of tens to hundreds of meters (Miller et al., 1982). The origin of ignimbrites was somewhat mysterious until the Mt. Pelee glowing cloud was observed in 1902. Still, no eruption of the truly huge ignimbrites, such as the Bishop Tuff (Long Valley) or Yellowstone Tuff has been observed. Nor do I expect any volunteers. These deposits are the only volcanic deposits that approach the magnitude of flood basalts, and their origin is still speculative.

3) Pyroclastic surge deposits are a type of pyroclastic flow. The common occurrence of dunes and anti-dunes indicate that surges result from more turbulent flow, with a lower concentration of particulates than the other flows. Surges hug the ground surface, but, because of their low density and high velocity, they are not as topographically constrained as other flows. The resulting deposits thus both mantle topography, and concentrate in low areas. Because of their low density, surges lose momentum rather quickly, so they tend to accumulate near the vent. The deposits are usually stratified and may show a number of current-bedding features.

Once again, the preceding three classifications are not totally distinct. There are gradations between them in terms of eruptive style and resulting deposit. Most single pyroclastic deposits have features associated with a combination of two mechanisms, or even of all three. As illustrated in Figure 4-19, an ignimbrite may have a stratified basal surge layer, followed by an unstratified flow deposit, which is then covered by an ash fall mantle. In Figure 4-19, the lithic fragments (black) are dense enough to be concentrated at the base of the flow unit. The more buoyant pumice fragments (white) are either not concentrated, or concentrated nearer the top. The sequence of events that such a deposit represents might involve a lateral surge laying down the initial deposit, followed by a flow caused by the collapse of the ash plume. The flow settled, and was covered by a blanket of the slow-settling ash fall. Lateral variations will also be apparent, because the surge will be restricted to the area near the vent, and the ash fall deposit may be the only layer found a great distance away.

Descriptions of pyroclastic deposits, and generalized sections such as Figure 4-19 give a simplified account of pyroclastic flows, which is quickly dispelled when one encounters the deposits in the field. Most pyroclastic deposits reflect an eruptive event that is episodic, and that changes character as it proceeds. The composition, viscosity, and volatile content of the magma are likely to vary as progressively deeper levels are tapped in the emptying magma chamber. If so, the deposit will be a complex accumulation of varying materials, textures, and structures that cools as a single unit.

4.2 INTRUSIVE, OR PLUTONIC, PROCESSES AND BODIES

The generic term for an intrusive igneous body is a plu-**ton,** and the rocks outside the pluton are called the **country rocks.** The size and shape of plutons is generally somewhat speculative, because erosion exposes only a small portion of most bodies. Nonetheless, we have managed to accumulate considerable data from more deeply eroded plutons, geophysical studies, and mining works. From these we have classified plutonic bodies into a few common forms. These forms are grouped into **tabular,** or sheet-like, bodies and **non-tabular** ones. Further classification is based on specific shapes, and whether a body cuts across the fabric (usually bedding) of the country rocks or whether it follows the external structures. Cross-cutting bodies are called **discordant,** and those that are intruded parallel to the country rock structure are called **concordant.**

4.2.1 Tabular Intrusive Bodies

Tabular intrusive bodies are simply magma that has filled a fracture. A concordant tabular body is called a **sill,** and a discordant one is called a **dike.** A sill occurs when magma exploits the planar weaknesses between sedimentary beds or other foliations, and is injected along these zones (Figure 4-20). A dike is a magma-filled fracture that cuts across bedding or other country rock structure. A fracture is an ideal conduit for magma, because fractures can penetrate deeply and form easily, particularly in areas affected by extension or the force of a rising magma diapir. Clearly, magma could not have been generated between two bedding layers, so a sill must be fed by a dike somewhere along its length (unless the bedding dips steeply, and the sill is nearly vertical). Dikes and sills are typically shallow and thin, occurring where the rocks are sufficiently brittle to fracture.

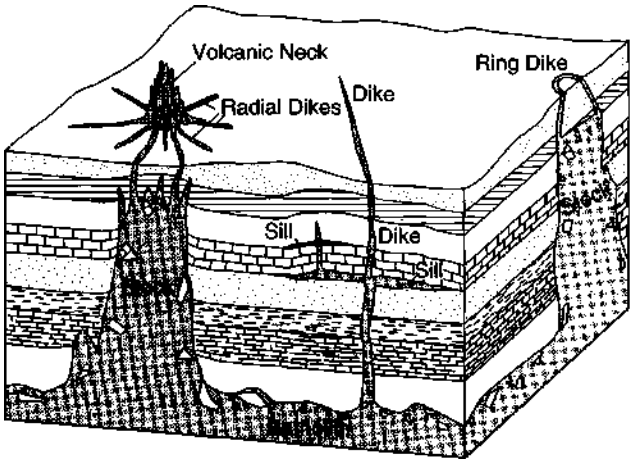


Figure 4-20 Schematic block diagram of some intrusive bodies.

Although most dikes and sills are emplaced in a single event, some may have a history of multiple injections. More than one stage of injection may occur because the dike or sill contracts as it cools, leaving a weakened zone for later magmas. Alternatively, the ductility contrast between a dike or sill and the country rock might make the contacts susceptible to localized deformation and later magmatic injection. A body is described as **multiple,** if all phases of injection are of the same composition, and **composite,** if more than one rock type is represented.

Dikes and sills can occur as solitary bodies, but dikes, at least, more typically come in sets, reflecting the tendency for fractures to form in sets as a brittle response to imposed stresses over an area. Genetically-related sets of numerous dikes or sills are called **swarms.** Note the Chief Joseph and Monument basaltic **dike swarms** in Figure 4—10. These swarms consist of sub-parallel dikes filling a set of fractures oriented NW-SE in response to SW-NE extension in the area during the Miocene. Dike swarms can consist of very large numbers of individual dikes. Figure 4-21 is a map showing two early Protero-zoic dike swarms in the S0ndre Str0mfjord region of SE Greenland. Both swarms are mafic, but are now highly metamorphosed. The principal NNE-striking swarm is the 1.95 Ga Kangamiut swarm, and the E-W swarm is somewhat older. These dikes represent up to 15% of the crust over large areas, indicating 15% crustal extension. The Kangamiut dikes are dramatic in their own right, but have also proved useful to delineate the margin of the area affected by the late phases of the Nagssugtoqidian mobile belt (as young as 1.8 Ga). In the southern and eastern portions of Figure 4-21 the dikes are unde-formed, but become highly deformed upon entering the mobile belt. Over a billion years of erosion has removed the great mountains that would have been produced by this deformational event, so we must rely upon structures within the rocks themselves to characterize it.

Dike swarms need not be parallel. A *radial* dike swarm about a volcanic neck is illustrated in the block diagram in Figure 4-20. Such radial fractures are a common response to the stresses imposed on the rocks above rising magma bodies. Figure 4-22a is a famous example of such a radial swarm of dikes in the Spanish Peaks area of Colorado. Figure 4-22b shows two radiating dikes associated with an eroded volcanic neck at Ship Rock, NM.

In addition to radial dikes over plutons, *concentric* ones may also form. There are two principal types of concentric dikes: ring dikes and cone sheets, illustrated in Figure 4-23. **Ring dikes** occur when the pressure exerted by the magma is less than the weight of the overlying rocks. In this case, cylindrical fractures will form as in Figure 4—23a. If the overlying rock is slightly more dense than the magma, cylinders of the roof will drop and the magma will intrude along the opening fracture, as in Figure 4-23b. This is more likely to occur with less dense, silicic magmas. Erosion to the level X-Y will result in a ring dike exposed at the surface (similar to Figure 4-23c). More than one concentric fracture may form, resulting in a series of dikes. A fracture may also penetrate to the surface, in which case a ring dike would feed a volcanic event, as in caldera collapse (Figure 4-9). The block that drops into the chamber may consist of country rock, an earlier phase of the pluton itself, or associated volcanics. Ring dikes are either vertical or dip steeply away from the central axis. Figure 4-24 illustrates the classic Tertiary ring dikes and cone sheets on the Island of Mull in western Scotland. Note that there are two centers, and several intrusive phases with rings of different compositions. Most of the blocks within a ring consist of earlier intrusives.

Cone sheets form when the pressure of the magma is greater than the confining pressure of the overlying rocks. In this case, inwardly dipping fractures form, as in Figure 4-23d, usually as a nest of concentric cones. Intrusion along these fractures produces a set of rings at the surface which, in contrast to ring dikes, dip *inwards.* Cone sheets and ring dikes can occur together, where they result from different phases of a single intrusion. For a more detailed discussion of the fracture orientations and stress trajectories associated with ring dikes and cone sheets above magma chambers, see Anderson (1936) and Roberts (1970).

Illustrations of dikes and sills, such as Figure 4-20, generally leave the impression that dikes are close to vertical and sills nearly horizontal. Although commonly true, this is not necessarily the case. Dikes, by definition, are discordant with the country rock structures, and sills are concordant. If the bedding is vertical, sills will be vertical and dikes may be horizontal.

The term **vein** refers to a small tabular body, whether or not it is discordant or concordant. It is typically used

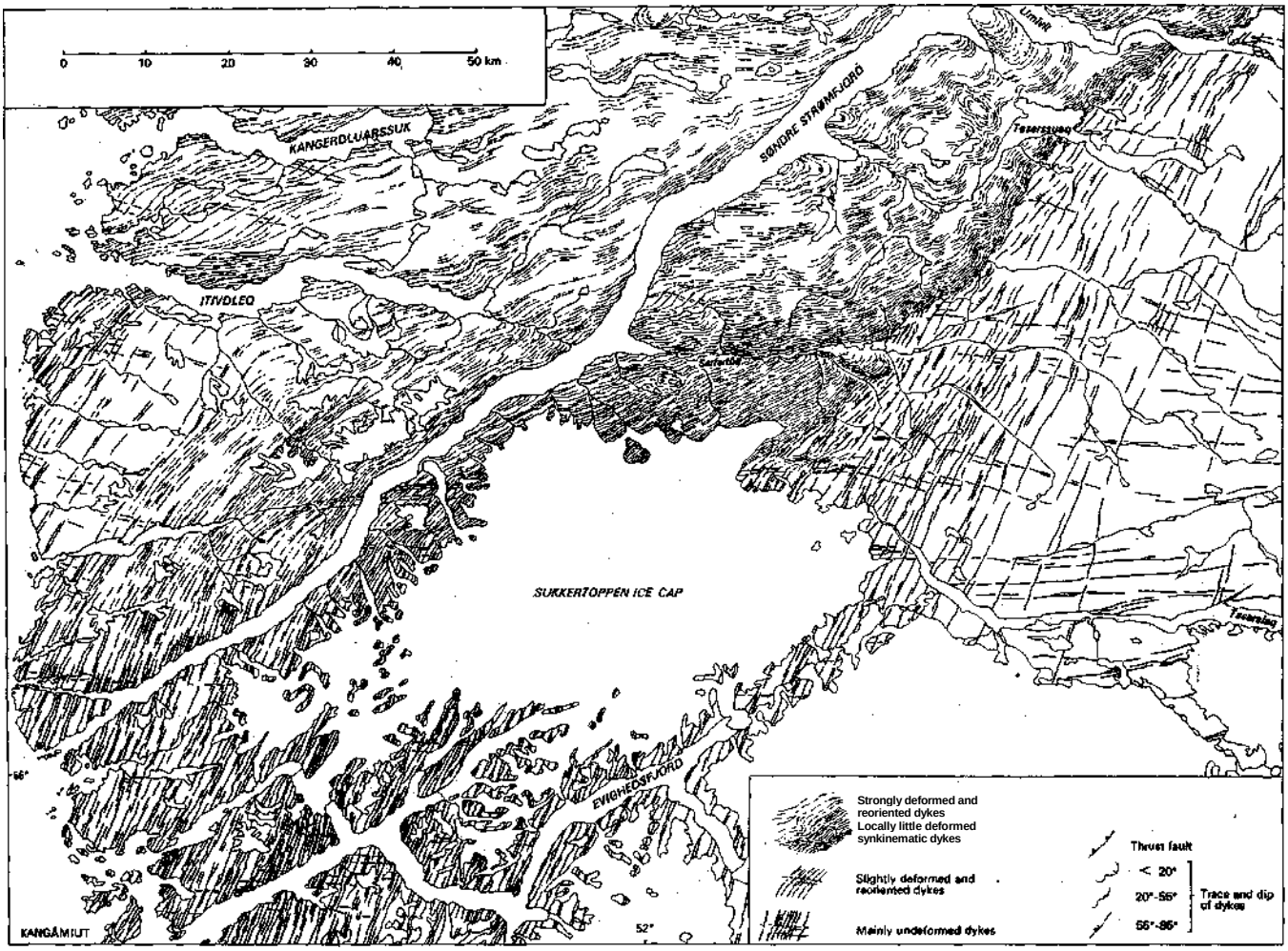
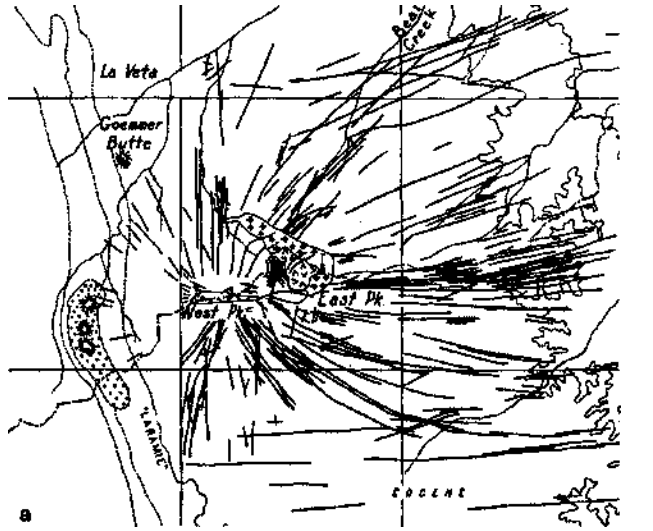


Figure 4-21 Kangamiut dike swarm in the S0ndre Str0mfjord region of SE Greenland. From Escher et al. (1976). Copyright Geological Survey of Denmark and Greenland.

in association with ore bodies, where numerous small offshoots from a pluton penetrate the adjacent country rocks. These offshoots (veins) are typically quartz-rich and may contain ore minerals. The term is *not* recommended for other igneous bodies. Rather, the terms dike or sill are preferable, no matter how small. Dikes and sills can thus range in thickness from less than a millimeter to over a kilometer, although most are in the 1 to 20 m range.

Tabular bodies are generally emplaced by **injection,** associated with dilation of the dike walls, as shown in Figure 4-25 a. In this case the walls of the NE-SW dike open, usually in a direction normal to the margins as the dike is injected. All features across the dike match up if we mentally remove the dike and close the walls in a direction normal to the margins. An alternative geometry is illustrated in Figure 4-25b. In this case, the features cross the dike as though the dike were a passive feature, requiring no movement of the dike walls. In fact, any movement of the walls in an effort to close the dike will fail to match both the bedding and the smaller dike. The distinction between these two types is strongest when there are at least two non-parallel structures in the country rocks, such as bedding and the small dike. This is a rare occurrence. The type of geometry in Figure 4-25b implies that the dike **replaced** the rock volume that it now occupies. This is most readily accomplished by chemical permeation and replacement of the country rock. Sloping of the country rock and injection of a dike is an alternative interpretation, but this requires a fortuitous lack of any motion whatsoever across the stoped block. If the dike is a product of replacement, it is no longer technically a dike. Such replacements are necessarily small, and perhaps the term **vein** would be preferable for such features. There may be a symmetrical mineral zonation parallel to the vein axis and some traces of the original structures of the replaced country rocks within the vein. There may also be a central joint, which acted as a con-



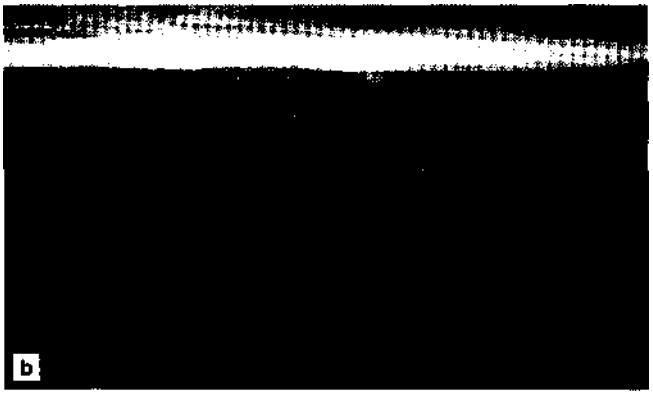


Figure 4-22 (a) Radial dike swarm around Spanish Peaks, Colorado. After Knopf (1936). Copyright © The Geological Society of America, Inc. **(b)** Eroded remnant of a volcanic neck with radial dikes. Ship Rock, New Mexico. From Shelton (1966).

duit for the fluids that carried the replacing elements. We will learn more about replacement when we discuss metasomatism in Chapter 30. Field evidence indicates that the vast majority of dikes and sills form by injection and dilation. Dikes and sills cool progressively from the walls inward, and in some cases develop columnar joints in the process. Sub-vertical dikes develop columns oriented nearly horizontally, which helps to identify them in places where the dikes intrude volcanic rocks of a similar composition, such as the feeder dikes to flood basalts.

4.2.2 Non-Tabular Intrusive Bodies

The two most common types of non-tabular plutons are illustrated in Figure 4-20; they are called stocks and batholiths. The shape of these bodies is irregular, and depends upon the depth of emplacement, the density and ductility of the magma and country rocks, and any structures existing in the country rocks at the time of emplacement. A **stock** is a pluton with an exposed area less than 100 km^2, and a **batholith** is a pluton with an exposed area larger than 100 km^2. Distinguishing the two on the basis of exposed area is very simple and useful in the field, but it is also unfortunate in that it is not based on the size of the body itself, but in large part on the extent of erosion. It is possible that a small pluton that is deeply eroded would have an exposed area greater than 100 km^2, and be a batholith, whereas a much larger pluton, only barely exposed, would be a stock. For that matter, the "batholith" illustrated in Figure 4-20 is not exposed, except as a volcanic neck and a ring dike, so would not technically be a batholith at all by the common definition. Also, several smaller stocks in an area may be so similar in age and composition that they are believed to be connected at depth, forming parts of a larger body. These separate extensions are called **cupolas** (see Figure 4-30). When geophysical evidence or mapping suggests a larger pluton beneath the surface, it would be better to

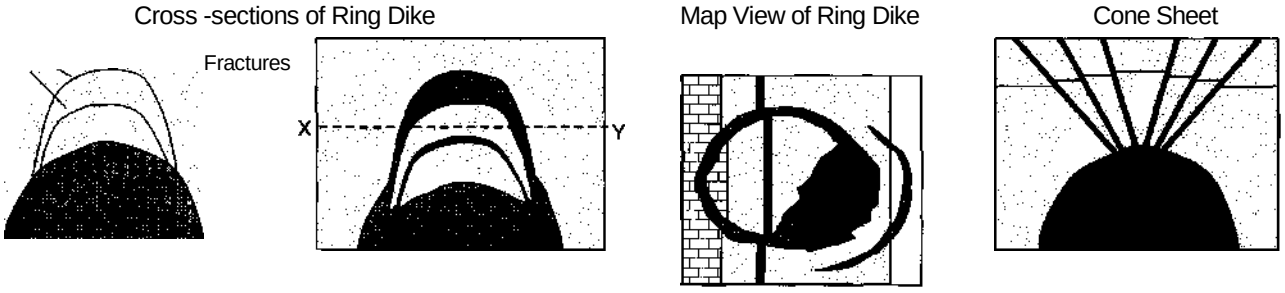
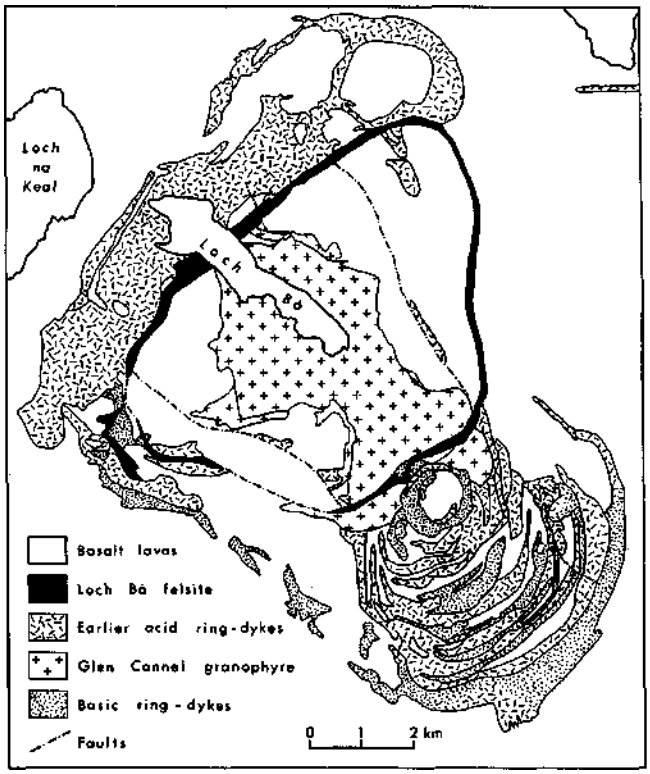


Figure 4-23 The formation of ring dikes and cone sheets, **(a)** Cross section of a rising pluton causing fracture and sloping of roof blocks, **(b)** Cylindrical blocks drop into less dense magma below, resulting in ring dikes, **(c)** Hypothetical map view of a ring dike with N-S striking country rock strata as might result from erosion to a level approximating X-Y in **(b). (d)** Upward pressure of a pluton lifts the roof as conical blocks in this cross-section. Magma follows the fractures, producing cone sheets. Original horizontal bedding plane shows offsets in the conical blocks, **(a), (b),** and **(d)** after Billings (1972), **(c)** after Compton (1985). Copyright ©, reprinted by permission of John Wiley & Sons, Inc.



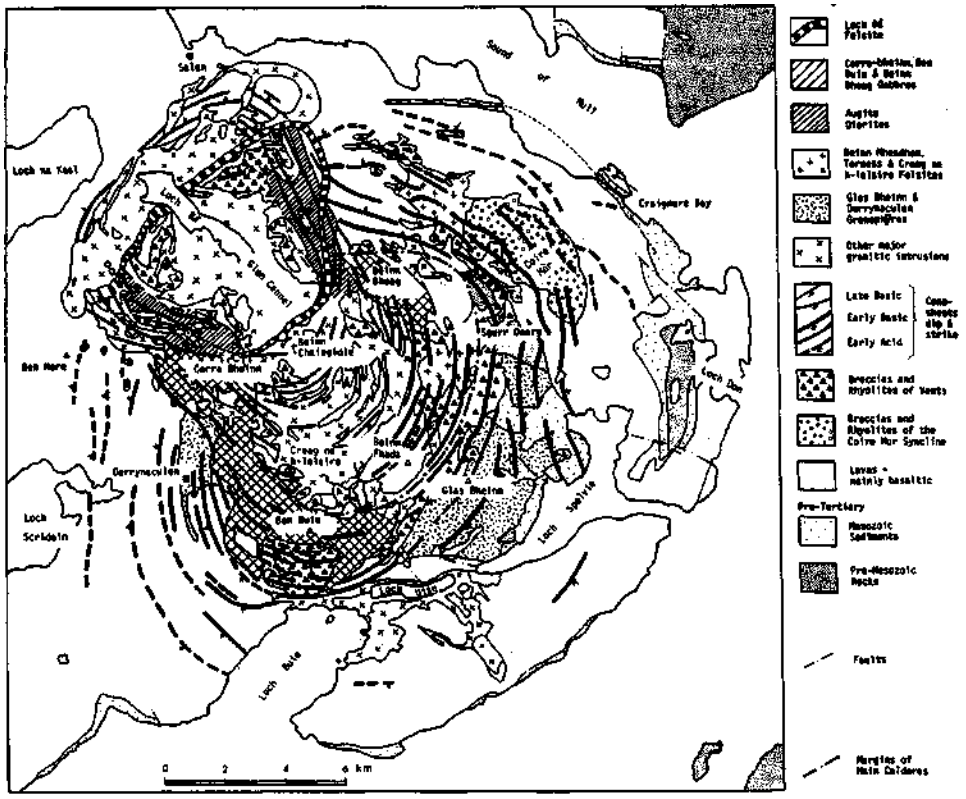


Figure 4-24 (a) Map of ring dikes, Island of Mull, Scotland. After Bailey et al. (1924). (b) Cone sheets in the same area of Mull, after Ritchey (1961). Note that the black felsite ring dike in part **(a)** is shown as the striped ring in the NW of part **(b).** Copyright © by permission British Geological Survey.

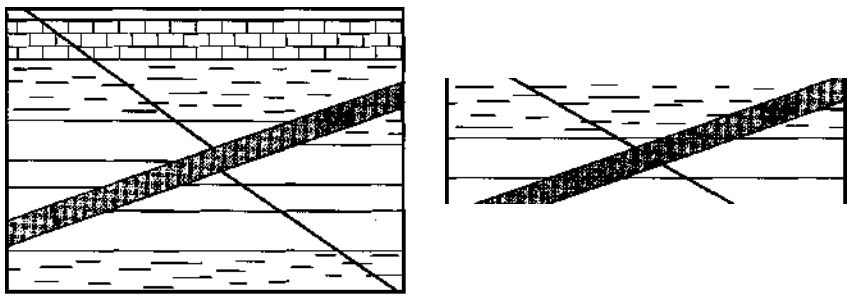


Figure 4-25 Types of tabular igneous bodies in bedded strata based on method of emplacement, **(a)** Simple dilation (arrows) associated with injection, **(b)** No dilation associated with replacement or sloping.

base the type of pluton on the estimated volume of the body, and not on the exposed area. With the exception of drill data, however, sub-surface interpretations are still somewhat speculative, and the classification based on exposed area does have the advantage of being definite. One way out of this dilemma was proposed by Pitcher (1993) who used the general term "pluton" for any large non-tabular body, and reserves the term "batholith" for the large composite arrays of multiple plutons developed in orogenic belts. Further complication stems from the Central European (and, until recently, French) practice of calling smaller bodies batholiths and larger ones plutons.

Some types of stocks are genuinely smaller bodies, and not simply limited exposures of larger batholiths. Some stocks represent the cylindrical conduit and magma chamber beneath volcanoes. This type of stock is called a **plug.** The exposed portion of a plug, commonly remaining after the more easily eroded volcanics of the cone have been removed, is called a **volcanic neck.** Alkaline igneous rocks, which form in mid-continent regions (typically associated with rifts), tend to rise along vertical conduits of limited cross-sectional area. Thus they typically form small plugs.

In addition to the two "generic" types of plutons, stocks and batholiths, there are a number of special types, based on shape rather than size. A **laccolith** is a concordant stock with a flat floor and an arched roof (Figure 4-26a). A **lopolith** is another concordant type of pluton intruded into a structural basin (Figure 4-26b). Both are essentially sills. A laccolith is sufficiently viscous (and silicic) to limit magma flow along the horizontal plane, and shallow enough to physically lift the roof rocks. Lopoliths are usually mafic, and characteristically much larger than laccoliths. The Duluth gabbro complex, for example, is a lopolith over 300 km across. The basin structure associated with lopoliths probably results from sagging due to the weight of the intrusion, but evacuation of a deeper magma chamber may also be a contributing factor. A fair number of structural geometries are possible for concordant plutons in folded rocks, and a correspondingly complex nomenclature developed earlier in the 20th century. We will forego indulging in this somewhat arcane classification. Perhaps the following excerpt, defining *cactolith,* as described by Hunt et al. (1953), from the AGI *Glossary of Geology* would serve to illustrate the degree of specialization attained:

> Cactolith: a quasi-horizontal chonolith composed of anastomosing ductoliths whose distal ends curl like a harpolith, thin like a sphenolith, or bulge discordantly like an akmolith or ethmolith.

4.2.3 Contact Relationships of Plutons

The emplacement of a pluton involves the juxtaposition of a hot, viscous, commonly fluid-saturated liquid in motion against a cooler, stationary, solid, generally of a much different composition. Such contrasts in properties and relative motion are certain to result in chemical and mechanical interactions that will impart to the contact zone some diagnostic structures and textures, but may also obscure the simple relationship of igneous material abutting country rock. The structures and textures,

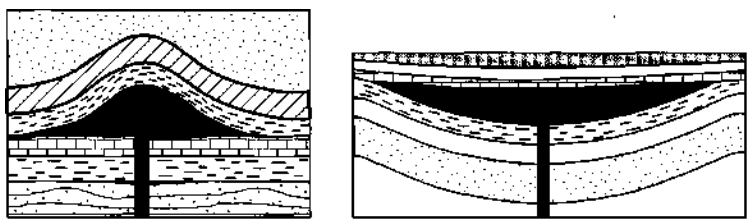


Figure 4-26 Shapes of two concordant plutons. (a) Laccolith with flat floor and arched roof, **(b)** Lopolith intruded into a structural basin. The scale is not the same for these two plutons, a lopolith is generally much larger.

both within the pluton and in the adjacent country rocks, can provide important clues to the processes involved during emplacement. The contact itself may be sharp, with igneous rock against relatively unaltered country rock, or it may have a gradational border zone.

The border zone may be strictly mechanical (injected), and, when broad, exhibit a gradation from undisturbed country rock through a zone of increasing dikes, veins, or tongues (collectively called **apophyses)** extending from the pluton. As shown in Figure 4-27a, the ratio of igneous to wall rock may gradually increase across such a zone, beginning with country rock containing a spaced network of small injected dikes through an **agmatite** zone (a rock with a high concentration of country rock fragments, or **xenoliths,** in an igneous matrix), to spaced xenoliths of country rock floating in igneous material, until even the xenoliths become rare. Any degree of gradation is possible.

Because many of the more silicic plutons are fluid-saturated and chemically distinct from the country rocks, fluids emanating from a pluton may permeate the surroundings, altering, or even partially melting, the country rocks and crystallizing igneous-type minerals in the exterior matrix. The result is a border zone across which the igneous rock gradually passes into the country rock without a distinct boundary (Figure 4-27b). We will return to this type of alteration in Chapter 30.

A third possible gradational border zone would combine the two processes of injection and permeation, as in Figure 4-27c. The various wall rocks, injected, and/or locally melted components can become so intimately associated as to result in a hybrid rock of mixed character.

Along sharper contacts at shallow depths, an intrusion may have thermal and chemical effects on the country rock. A gradational zone may result from strictly thermal processes, particularly if the intrusion is both hot and dry. In such cases, unmetamorphosed country rock is heated and recrystallized by the pluton across a narrow interval at the contact. The grain size and degree of re-crystallization would decrease rapidly away from the contact, and the contact itself should still be distinct. The thermal effects are more commonly combined with a chemical gradient, established by the silica-saturated fluids released from a fluid-saturated pluton. The result is a **contact metamorphic aureole,** which is usually gradational, but affects the country rocks more than the pluton itself, leaving the original contact distinct. Contact metamorphism will be discussed in more detail in the metamorphic section of this text.

The dynamics at the contact with the wall rocks may impart some lasting features *within* the pluton as well. In addition to xenoliths, there may be a **chill zone.** A chill zone is a zone of finer grain-size resulting from the rapid solidification of the pluton where it comes in contact with the cooler country rock. Because the igneous material being intruded is in motion against a stationary country rock, there may also be a zone of distributed shear in the viscous magma. The more viscous the magma, the more pronounced the shear and thicker the sheared zone. The result of the shear will be the development of an imparted lineation and/or foliation parallel to the contact in the marginal portion of the pluton (Figure 4-28). This will be most evident in any elongate or platy minerals, such as hornblendes, pyroxenes, and micas. There may also be some elongated or flattened mineral aggregates or ductile heated xenoliths that produce disc-shaped masses called **schlieren.** The intensity of magmatic foliations typically increases toward the margins where shear is greatest (Paterson et al., 1991). Some magmatic foliations also form at an angle to the contact (Berger and Pitcher, 1970; Courrioux, 1987).

If the magma is sufficiently viscous, and the country rock hot enough, the contact shear may also affect the country rocks, rotating the foliation *outside* the pluton into parallelism with the contact as in Figure 4-28 (Castro, 1986). Such transposition of the exterior foliation



(a) Injected (b) Permeated (c) Combination

Figure 4-27 Gradational border zones between homogeneous igneous rock (light) and country rock (dark). After Compton (1962).



Figure 4-28 Marginal foliations developed within a pluton as a result of differential motion across the contact. From Lahee (1961). Copyright ©, with permission of the McGraw Hill Companies.

makes an otherwise discordant contact become concordant.

The overall result of the shearing process is the loss of a clear, sharp, discordant igneous contact and the loss of the characteristically isotropic igneous texture. The igneous rock looks more like a metamorphic gneiss, and painstaking field and petrographic work may be required to interpret the contact correctly.

Because of the great difference in mechanical properties that typically exists between many crystallized plu-tons and the country rock, later deformation is commonly concentrated along the pluton margins. As a result, many igneous contacts suffer post-emplacement shearing, superimposing a tectonic imprint over the original igneous contact. This shear may have only minor effects, or it may obliterate any original igneous contact textures. If the shear is great enough, the pluton may be faulted into juxtaposition with a completely different rock, and the contact becomes strictly tectonic. It may be impossible to distinguish a extensively sheared igneous contact from a strictly tectonic one, or even from a contact that is sheared during emplacement of the pluton, as described above. As pointed out by Paterson et al. (1991), the structural patterns within and around plutons reflect the superposition of deformation related to emplacement and regional stresses. A variety of structures are possible, ranging from ductile to brittle.

4.2.4 Timing of Intrusion

Most batholiths are emplaced in mountain belts as part of the overall subduction/orogenic process, and they typically play an important and active role in the evolution of those belts (see Chapters 18 and 19, and Paterson et al., 1991). Any attempt to relate deformation with emplacement must demonstrate that deformation (and metamorphism) is closely related to intrusion, not only spatially, but also *temporally.* Textures within plutons and in the country rocks reflect the timing of emplacement with respect to tectonic activity.

Post-tectonic plutons are emplaced after an oro-genic/metamorphic episode, and the igneous rocks thus lack any deformational features, such as foliations (other than those related to intrusion). The regional deformational fabrics and structures of the country rocks will either be cut discordantly by the pluton, or curve into parallelism with the contact. They will also be overprinted by any emplacement-related structures.

Syn-tectonic plutons are emplaced during the orogenic episode. Any regional foliation will be continuous with emplacement-related foliations in the pluton. The pluton will also be affected by any later continuing orogeny.

Pre-tectonic plutons are emplaced prior to the orogenic episode. Both pre- and syn-tectonic plutons suffer the imprint of the deformational and metamorphic processes associated with the orogeny. They therefore have an internal foliation that parallels, and is continuous with, that of the country rocks (Figure 4-29). The regional foliation may curve around a non-foliated pluton because of ductility contrasts. It is commonly difficult or impossible to distinguish whether a pluton is pre- or syn-tectonic, since both share this attribute. Syn-tectonic intrusions are usually more ductile at the time of deformation, so they may be more elongated in the foliation direction, with more concordant contacts. Pre-tectonic plutons are cooler, and likely to be more resistant to deformation, which would be concentrated at the pluton margins. Because plutonism and orogeny are related in most orogenic belts, purely pre-tectonic plutons are rare. Most plutons that can be characterized as pre-tectonic are commonly syn- or post-tectonically associated with an earlier orogenic event in a multiply deformed belt.

4.2.5 Depth of Intrusion

In addition to timing, the depth of emplacement affects many structural and textural features of plutons. Many of these characteristics were summarized by Buddington (1959) on the basis of the level of emplacement with respect to the three **depth zones** first distinguished by Grubenmann (1904). These zones, the epizone, meso-zone, and catazone, are based on the characteristics of the country rocks, and the actual depth limits are only approximate (and variable as well) owing to differences in the geothermal gradient between deformed belts.

The epizone is characterized by relatively cool (less than 300°C) country rocks of low ductility, at depths less

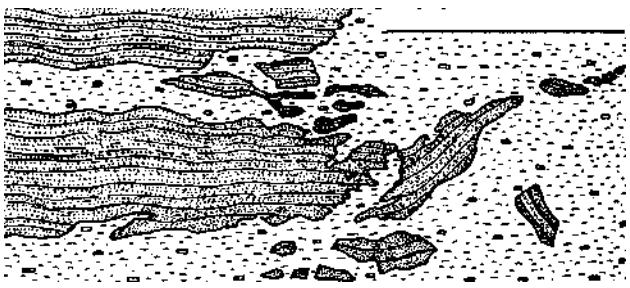


Figure 4-29 Continuity of foliation across an igneous contact for a pre- or syn-tectonic pluton. From Compton

than 10 km. Some plutons, for example the Boulder Batholith, Montana (Hamilton and Myers, 1967), may be so shallow as to intrude their own volcanic carapace, and thus be emplaced above the previous ground surface (Figure 4-36). Plutons of the epizone are usually post-tectonic, and have sharp, discordant contacts. The wall rocks are typically brecciated, and there may be numerous dikes and offshoots from the main igneous body. The top of a pluton commonly penetrates the roof rocks in an irregular manner, with several lobes extending upward and outward.

Figure 4-30 illustrates several of the relationships between a barely exposed epizonal pluton and the country rocks composing the roof of the body. Contacts are generally sharp and discordant, crosscutting the structures in the country rocks. An **offshoot** is a general term for any lobe of the main plutonic body that protrudes into the country rock. A **cupola** is a non-tabular offshoot that is isolated from the main body, as observed in map view. A **septum** is a peninsula-like projection of the country rock into the pluton so that it separates two igneous lobes. A **roof pendant** is a projection of the roof rocks into the pluton that has become isolated *by erosion.* The structures of the roof pendant are parallel to those of the country rocks outside the pluton, signifying that the pendant was connected to the country rocks, and prevented from rotating. If the structures of the isolated block were at a different angle than those of the country rocks, the block would be free-floating and rotated, and thus a xenolith. Large xenoliths are commonly called **rafts.**

Although some large shallow batholiths are found in the epizone, most epizonal plutons are fairly small. This is to be expected, because erosion is rather limited if the country rocks are epizonal. Many of the small stocks may themselves be cupolas of a larger body at depth, such that progressive erosion would expose larger bodies and deeper country rocks.

Emplacement of an intrusive body may fracture the country rocks as it forces its way toward the surface. Hy-fooftieimal alteration and ore mineralization are corn-on, and concentrated along such fractures. A onvective hydrothermal system may be established above the pluton, driven by the heat of the intrusion, making the alteration intense and pervasive (see Section 4.3). Contact metamorphism may be dramatic where the country rocks are previously unmetamorphosed, or poorly metamorphosed. The principal limitation on the size and development of a contact aureole is the rapid cooling of small plutonic bodies, and the loss of fluids along fracture channelways.

Fabrics in an epizonal pluton are typically isotropic, and this commonly includes the areas near the contacts that have experienced some shear against the stationary wall rocks. **Miarolitic cavities,** representing "bubbles" of fluid (presumably liberated at low pressure) with euhe-dral mineral projections inward, are common. The epizone is the typical environment for laccoliths, lopoliths, plugs, ring dikes, and cone sheets.

The **mesozone** is the depth interval of about 5 to 20 km. The country rocks are low-grade regional metamor-phic rocks at temperatures of 300-500°C. Plutons in this zone have characteristics that are transitional between those of the epizone and the catazone (Figure 4-3la). They may be either syn- or post-tectonic. The contacts may be sharp or gradational, and discordant or concordant; because the country rocks are more ductile than in the epizone. A contact metamorphic aureole is typically well developed, because the plutons are larger and cool more slowly; and the country rocks are metamorphosed to only a low grade prior to emplacmennt, and thus are susceptible to modification by the thermal and compositional contrasts imposed by the pluton. The rocks of the contact aureole commonly have a foliated fabric, because they experience regional as well as contact metamorphism. "Spotted" slates and phyllites are common, in which the spots are larger minerals that form during, or after, the regional foliation. A chill zone is minor or absent. Fabrics in the pluton may be directionless (isotropic), but are commonly foliated or lineated near the contact.

The deepest zone, the **catazone,** is deeper than 10 km. Here the country rocks are undergoing medium to high-grade regional metamorphism in the 450 to 600°C range. Plutons are usually syntectonic with gradational contacts

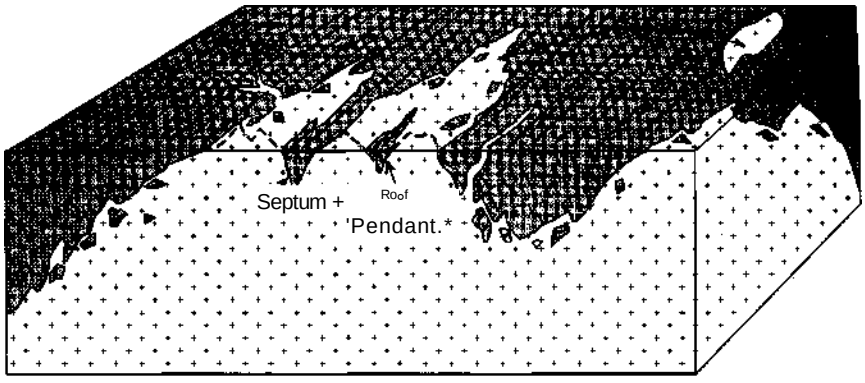
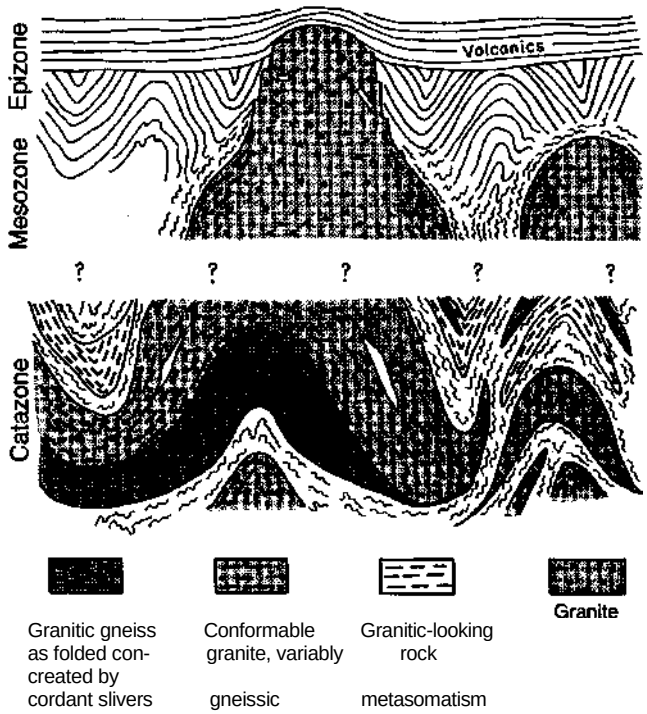


Figure 4-30 Block diagram several kilometers across, illustrating some relationships with the country rock near the top of a barely exposed pluton in the epizone. The original upper contact above the surface is approximated by the dashed line on the front plane. After Lahee (1961). Copyright ©, with permission of the McGraw Hill Companies.



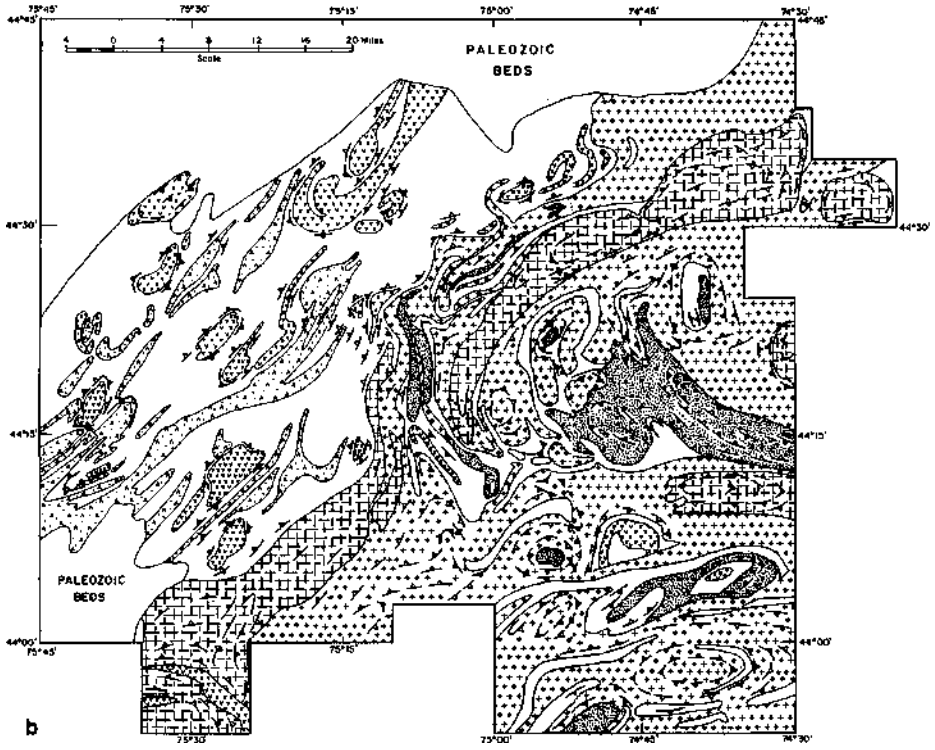


Figure 4-31 (a) General characteristics of plutons in the epizone, mesozone, and catazone. **(b)** Shape of concordant igneous bodies typical of the catazone, northwest Adirondacks, NY. The blank areas are metamorphic country rocks, and the patterned areas are various plutons. From Buddington (1959). Copyright © The Geological Society of America, Inc.

and no chill effects. The viscosity contrast between the country rocks and the magma is relatively low, so contacts are generally concordant, in the sense that the foliation of the fairly ductile country rocks has been sheared and rotated to become parallel to the contact. There are seldom noticeable contact metamorphic effects, because the country rocks are already at fairly high metamorphic grades. The plutons commonly occur as domes, or as sheets or pods, with a foliated and/or lineated internal fabric that passes directly into the fabric of the metamorphic country rocks (see Figure 4-31). In addition to any flow foliation, the igneous body may have been imprinted with a metamorphic foliation during or after final solidification. It is difficult to recognize such foliated plu-tonic rocks as igneous, because they have the characteristics of high-grade metamorphic gneisses. In the deep catazone and lower crustal levels the conditions for partial melting of the crustal rocks may be attained, creating localized melts from the country rocks that further blur the distinction between the igneous and metamorphic components.

Because of the similarities between some igneous and high grade metamorphic components and the gradational relationships between them in the deep crust, it is common in exposed high-grade metamorphic terranes to be able to traverse from rock that is metamorphic, through a mixed zone, into igneous-looking rocks in a completely gradational manner without ever encountering a distinct contact.

The boundaries between the depth zones are neither sharp, nor static. Rocks in any of these zones can enter the next deeper zone by burial or thrust stacking, or the next shallower zone by uplift and erosion. Thus neighboring plutons in a given exposed terrane can be em-placed at different times in a different zone.

4.2.6 **Multiple Injection and Zoned Plutons**

As with volcanic centers, many plutons show complex histories of multiple intrusion of magmas of varying composition. Large batholith belts (Chapter 17) are composed of numerous compound intrusions. For example, although early descriptions of the Sierra Nevada Batholith in California refer to it as a single batholith, it has long been recognized as a broad zone of repeated intrusion of numerous plutons over a considerable time span. The number of separate intrusions is now known to be in the hundreds. Even smaller intrusive bodies, if mapped carefully, commonly show multiple intrusive events.

One well-documented example of multiple emplacement is the Tuolumne Intrusive Series in Yosemite National Park, a pluton that is part of the Sierra Nevada Batholith (Bateman and Chappell, 1979). In this case (Figure 4-32) a small batholith has been shown to be a series of related intrusions, ranging in composition from an outer diorite through a granodiorite, and finally a por-phyritic granite at the core. A general concentric zona-tion, with later magmas toward the center, suggest that fresh magma surges took place within a solidifying outer shell. This concentric pattern is not perfect, however, because some portions of the later intrusives either abraded, assimilated, or fractured their carapace to crosscut the older crystallized margins. A number of chemical and mineralogical trends associated with this sequence of magmas are generally consistent with the evolution of a deeper magma chamber undergoing some combination of crystal fractionation, assimilation of wall rocks, and/or mixing of several magma injections. We will return to the Tuolumne Intrusive Series in Chapters 8 and 11, looking at these trends in more detail to assess the development of magma chambers.

The Tuolumne series is but a single example of a common phenomenon. Many plutons of various sizes display multiple injections and chemical variations across zones. Even where distinct magma types are not apparent there is commonly a systematic variation in the mineralogy or composition of some minerals. For example, the amount of quartz or potassium feldspar may increase, or the pla-gioclase composition may become progressively more sodic toward the center of a pluton. These trends are also compatible with a progressively cooling and evolving magma, as will be addressed in Chapter 8.

4.2.7 The Process of Magma Rise and Emplacement, and the "Room Problem"

Intrusive igneous rocks are simply magmas that have not reached the surface. Plutonism lacks the drama of vol-canism, because no one has ever died from pluton emplacement, and there has never been a pluton alert. The volume of igneous rock in plutonic bodies, however, is considerable, and there is a pluton of some sort beneath every volcano. In a sense we can think of these bodies as liquid that crystallized, or "froze," in the plumbing system on its way toward the surface.

When magma forms by some melting process at depth, it segregates from the residual unmelted solids to form a discrete liquid mass. This mass is less dense than the surrounding solid (see Chapter 5), so it becomes *buoyant.* The buoyant magma body tends to rise, and, if the surrounding material is sufficiently ductile, it is generally considered to do so as a **diapir.** A diapir is a mobile mass that rises and pierces the layers above it as it does so. Whenever a sufficiently ductile rock mass is covered by denser rocks that are also ductile, diapirs are capable of forming.

A well-documented example of diapirism is the formation of salt diapirs in areas such as northern Germany, Iran, and the American Gulf Coast. In these cases, the salt beds (evaporite deposits) are covered by subsequent sediments that compact and become denser than the salt

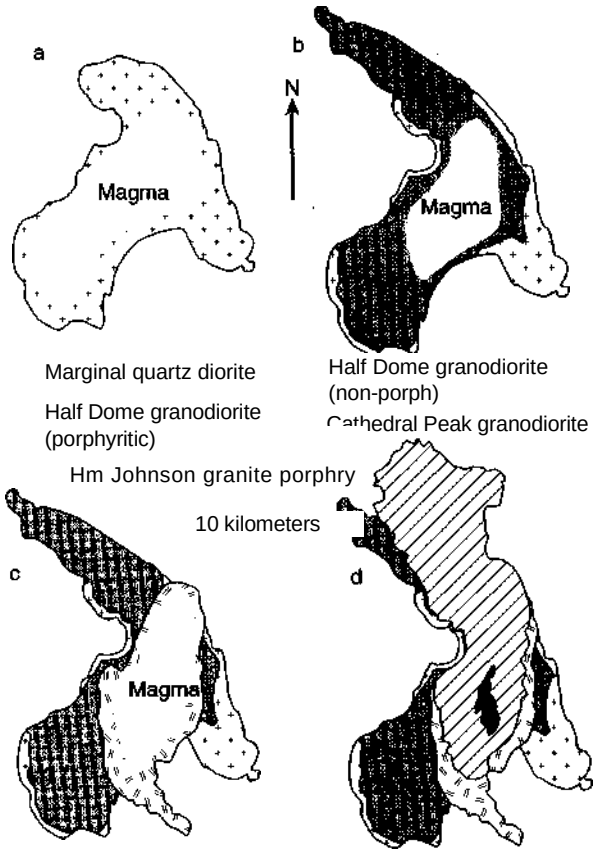


Figure 4-32 Developmental sequence of intrusions composing the Tuolumne Intrusive Series (after Bateman and Chap-pell, 1979). **(a)** Original intrusion and solidification of marginal quartz diorite. **(b)** Surge of magma followed by solidification of Half Dome Granodiorite. **(c)** Second surge of magma followed by solidification of porphyritic facies of Half Dome Granodiorite. **(d)** Third surge of magma followed by solidification of Cathedral Peak Granodiorite and final emplacement of Johnson Granite Porphry. Copyright © The Geological Society of America, Inc.

beneath. Halite (salt) not only has a low density, but it will begin to behave in a ductile fashion and flow under conditions of low confining pressure. After 5 to 10 km of overburden develops, the salt beds begin to flow, and the upper surface develops irregular swells (Figure 4-33). Salt flows upward into these swells and they form diapirs that rise toward the surface. If the diapir rises sufficiently above its source, and is not supplied by enough salt flow from the evaporite bed, the column feeding the diapir stretches and constricts until the body separates from the source bed. The inverted raindrop shape, with bulbous leading edge and tapered tail, is characteristic of rising diapirs. Because buoyancy is the driving force that causes a diapir to rise, once it reaches a level where the density of the surrounding rock is the same as that of the salt, it will stop rising and spread out laterally.

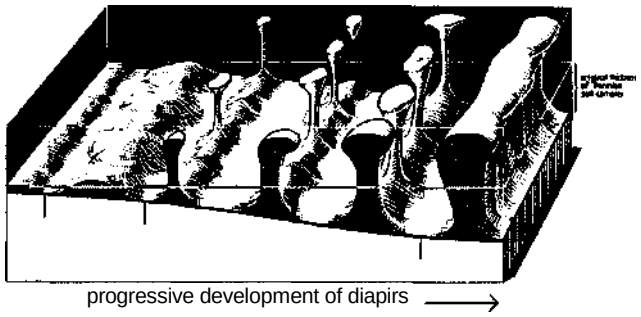


Figure 4-33 Block diagram of subsurface salt diapirs in Northern Germany. After Trusheim (1960). AAPG ©, reprinted by permission of the AAPG.

Magma diapirs are believed to behave in a similar fashion: Those forming in the mantle rise through the ductile mantle, just as the salt does through the overlying sediments. As long as the viscosity of the magma and country rock are fairly similar, they rise by radially distending the overlying rock as they force their way upward. In the less ductile areas of the upper mantle and crust, rising magma can no longer rise in diapiric fashion. Rather, it must exploit fractures or other weaknesses in the rocks through which it rises.

The method by which a large intrusive body moves upward through the crust and creates sufficient room for itself is far from clear. This **"room problem"** has been the subject of considerable debate for decades, and is still an enigmatic problem. For excellent reviews of the room problem, see Newell and Rast (1970) and Paterson et al. (1991).

Open fractures and voids are limited to the very shallow near-surface environment (a few tens of meters). Below this, the rise of magma by simply filling such open voids is not an option. Rising magma can follow preexisting (closed) fractures, however, by forcibly displacing the rocks that form the fracture walls and following these planar conduits. At depth, the ability of magma to force a fracture open is limited, because the injecting pressure of the magma is seldom great enough to significantly displace rigid rock walls that are forced together by litho-static pressure below a few kilometers depth. Of course, if an area is undergoing regional extension, the walls of any fractures will not be under compression, and magma can force them apart. The number and width of dikes or sills in such filled conduits depends upon the rate of extension, and should be commensurate with typical plate tectonic rates of less than ~ 3 cm/yr.

The room problem becomes more troublesome for larger intrusive bodies, which occupy a significant volume, and thus much more rock must be displaced for them to move upward. Figure 4-34 summarizes the proposed mechanisms by which a pluton could make room and rise. Plutons, such as laccoliths, can **lift the** roof

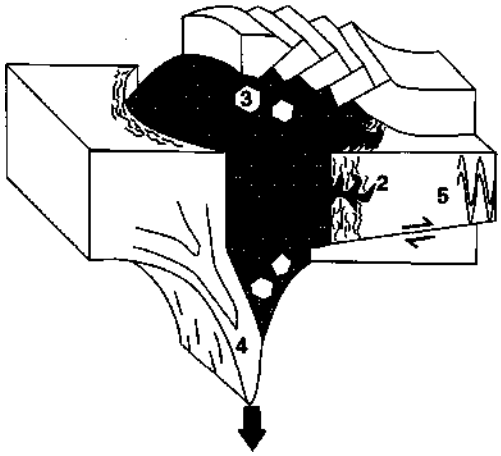


Figure 4-34 Diagrammatic illustration of proposed pluton emplacement mechanisms. **(1)** doming of roof; **(2)** wall rock assimilation, partial melting, zone melting; **(3)** stoping; **(4)** ductile wall rock deformation and wall rock return flow; **(5)** lateral wall rock displacement by faulting or folding; **(6-)** (and 1)- emplacement into extensional environment. After Pater-son et al. (1991). Copyright © the Mineralogical Society of America.

(#1 in Figure 4-34) by either folding or block elevation along faults. Controversy exists over whether the lifting force of plutons is restricted to magma buoyancy, which limits the ability of a pluton to lift the roof when it reaches the level at which its density equals that of the country rocks. Lifting may be facilitated in such cases by **magmatic overpressure,** which might supply additional depth-derived pressure. Roof doming to form laccoliths is limited to depths less than about 2 to 3 km, where the magmatic pressures can exceed the strength of the overburden (Corry, 1988). Similar restrictions apply to block uplift along faults.

Alternatively, magma may melt its way upward (a process called **assimilation:** number 2 in Figure 4-34). The ability to melt the walls using magmatic heat is limited by the available heat of the magma (and the amount of magma that can penetrate in the first place). As we shall see in Chapter 10, the intruded magmas are themselves most typically generated *by partial* melting, so they are seldom appreciably "superheated" (heated above the temperature at which the melt coexists with solids). Thus the heat available for melting the country rocks does not exist in excess, and must be supplied by the latent heat of crystallization of some portion of the magma, making it at least partially solid, and therefore less mobile. This type of heat is necessarily limited.

If the country rocks are sufficiently brittle, blocks of the roof over a rising pluton could become dislodged, fall, and sink through the magma (number 3 in Figure 4-34). This process is called **stoping,** after the mining practice of quarrying the roof of an underground working. Considerable evidence for stoping is to be found in the upper portions of many plutons, where blocks of country rock are suspended in the crystallized igneous rock as rafts or xenoliths. The process is evident where agmatites form by injection along a fracture network (see Figure 4-27a). Stoping requires that the country rock be denser than the magma. Also, for stoping to be effective, the stoped blocks must be large enough to sink rapidly enough in a viscous magma, which can require blocks as large as tens of meters in granitic melts. Stoping may be effective only in the shallow crust where the rocks can be fractured. Cauldron subsidence (Figure 17-16) and caldera formation (Figure 4-9) are examples of large-scale stoping at shallow depths. Of course the pluton must get to this level of the crust before this method can be possible, so stoping may be an emplacement mechanism at shallow depths, but cannot make the original room for the pluton where there was none before. The lack of field evidence for significant quantities of stoped xenoliths calls into question the effectiveness of stoping as a major process in magma emplacement.

A combination of assimilation and stoping, called **solution stoping** or **zone melting,** may operate at depths where the country rock is near the melting point. In this process, the minerals of the roof rock melt, and an equivalent amount of magma at the floor crystallizes, as proposed by Ahren et al. (1980). This process mitigates the heat loss from the intrusive magma, a major impediment to assimilation. Such a process may be effective in the mantle and deep crust, where the magma rises as a di-apir anyway, but if the country rocks are well below their melting point, more minerals would have to crystallize than melt, and the pluton would quickly solidify. This process will be discussed further in Chapter 11.

Ductile deformation and return downward flow (number 4 in Figure 4-34) are the mechanisms associated with the rise of diapirs at greater depths, and are probably efficient where the viscosity of the country rocks is low. At any depth, **ballooning,** or radial expansion of the magma chamber by addition of magma from below may also take place. The mechanism of roof yielding to this stretching would be ductile at depth, and more brittle near the surface. Ballooning may also physically compress the wall rocks at the sides of the diapir, forcing the walls apart with accompanying aureole deformation (number 5 in Figure 4-34). Theoretical and experimental studies suggest that for diapirism and ballooning to be effective, the viscosities of the pluton and wall rock must be similar (Marsh, 1982; Arzi, 1978; Van der Molen and Paterson, 1979). The lack of textural evidence for wall rock softening around most mid- and upper-level plutons argues against diapirism as an emplacement mechanism at these

crustal levels (Paterson et al., 1991). As in the case for dikes mentioned above, extensional environments (number 6 in Figure 4-34) would facilitate this process, but, as mentioned above, the rate of extension places limits on the degree to which it can contribute.

Although magma is believed to rise by some or all of the above processes, they all have limits to the extent and the circumstances under which they may operate. The room problem is still a problem to this day, and is particularly vexing when we consider the giant batholiths, which occupy vast volumes in the crust. Our concept of these great igneous bodies, however, is based on exposure at the surface, and the shape of a batholith at depth is not certain. Because these bodies are intruded from below, there is a natural tendency to think of them as extending outwards with depth, and occupying a considerable volume beneath the surface that we can observe (as in Figure 4-20). This is the way batholiths have been drawn for decades in most geologic cross sections involving them. They simply get larger downward, which certainly leaves the impression of much more igneous material below.

The shape of experimental diapirs, on the other hand, is different, at least as they near the surface. Figure 4-35 is a drawing of some results of the experiments by Ram-berg (1970), in which he modeled diapirs using soft materials of low density covered by layers of higher density strata, and placed them in a centrifuge to drive the di-apiric motion. The low-density material rises as diapirs due to its buoyancy and the ductility of all the material in the centrifuge. Upon reaching a level where the surroundings are less dense, however, the buoyancy of the diapir is reduced or lost entirely, at which point they tend to spread laterally and become more broad and thin. If this behavior is typical, even in the more brittle epizone, batholiths may be much thinner than we have imagined, and the amount of room required for their emplacement is greatly reduced. Based on gravity models and field-work in eroded batholiths, several are now suspected of being shallow-floored intrusions. Figure 4-36 is a proposed cross section of the Boulder Batholith in Montana, which suggests that it is less than 10 km thick. Other plu-tons are now suspected of being this thin or even thinner. The Lilesville Granite, North Carolina (Waskom and Butler, 1971), the Katahdin Pluton, New Hampshire (Hodge et al., 1982), and the Flamanville Granite, NW France (Brun et al., 1990) have gravity-modeled floors at depths less than 3 km. If this is true for most batholiths, it certainly alleviates the need for great amounts of room, and lessens the room problem considerably.

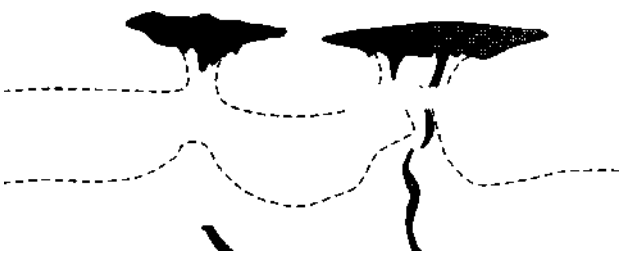


Figure 4-35 Sketches of diapirs in soft putty models created in a centrifuge by Ramberg (1970).

4.3 HYDROTHERMAL SYSTEMS

The fractured and permeable sedimentary and volcanic rocks overlying many hot, shallow intrusions are ideal sites for the development of extensive hydrothermal systems. A typical hydrothermal system associated with a silicic volcanic terrane is shown in Figure 4-37. Stable isotope studies have shown that meteoric water predominates over juvenile in most hydrothermal systems (Mazor, 1975), but the ratio is variable from one locality to another. The heat of the shallow magma chamber typically associated with recent volcanism heats the ground-water (plus some added juvenile component) so that it expands and rises through the permeable material above, commonly resulting in fumaroles and hot springs at the surface. The water then cools, moves laterally as more hot water rises from below, and descends again as it becomes denser in a typical convective system. The result is the recirculation of the groundwater system above the magma body, and perhaps in the upper solidified portion of the body itself, which may also be extensively fractured (Henley and Ellis, 1983; Hildreth, 1981).

Hydrothermal systems over shallow batholiths can affect a considerable volume of rock. Aqueous fluid flow is controlled by the permeability of the overlying rocks. Extensive fracture systems are common above shallow intrusions, and these act as particularly effective conduits for hydrothermal fluids. Caldera structures are therefore common centers for hydrothermal activity. Drill holes in some systems extend to depths of up to 3 km, where they have encountered near-neutral pH saline waters (brines) at temperatures up to 350°C. Below this, the character of the solutions is imprecisely known, but higher temperatures are certainly to be expected. If boiling occurs in the shallow portion of the system, CO_2 and H_2S are typically concentrated in the steam, which may reach the surface as fumarolic activity, or condense and oxidize to form a distinctive acidic sulfate/bicarbonate solution common in many geothermal fields.

The hydrothermal fluids evolve through chemical (and isotopic) exchange with the silicate melt, and/or the solidified portion of the pluton, whether the majority of



Figure 4-36 Diagrammatic cross section of the Boulder Batholith, Montana, prior to exposure. After Hamilton and Myers (1967).

the water is juvenile or not. It will thus contain a number of dissolved igneous constituents. As this fluid interacts with the surrounding rocks, it can cause a number of chemical, mineralogical, and textural changes to take place, depending on temperature, permeability, chemical composition and nature of the fluid and the rocks, the fluid/rock ratio, and the longevity of the hydrothermal system. Fluid/rock ratios range from 0.001 to about 4, based on oxygen isotope exchange (Taylor, 1974). The great variation in the physical and chemical nature of the hydrothermal system results in a similar variety of alteration products, including quartz, feldspars, clay minerals, chlorite, calcite, epidote, a number of sulfides and ores, zeolites, and biotite, actinolite, diopside, and/or garnet at higher temperature. Volcanic glass and mafic minerals in the country rocks are particularly susceptible to alteration. The alteration and ore mineralogy is typically zoned, and reflects a temperature gradient as well as chemical gradients and kinetic effects (Henley and Ellis, 1983). Such zonation may occur on a small scale, perhaps about a fracture, and/or on a large scale, as concentric zones that span the entire cap of the pluton. Considerable proportions of economic ore deposits are associated with present and ancient hydrothermal systems. These deposits are an important source of gold, silver, copper, lead, zinc, molybdenum, etc.

Analogous sub-ocean floor systems have been recognized at ocean ridge spreading centers in which ocean and juvenile water are convectively circulated in the thin oceanic crust. Submersible research vessels have even permitted scientists to see and document sea floor hot springs ("smokers") actively depositing metal sulfides and supporting a new and unique biological community (Francheteau et al., 1979; Hekinian et al., 1980).

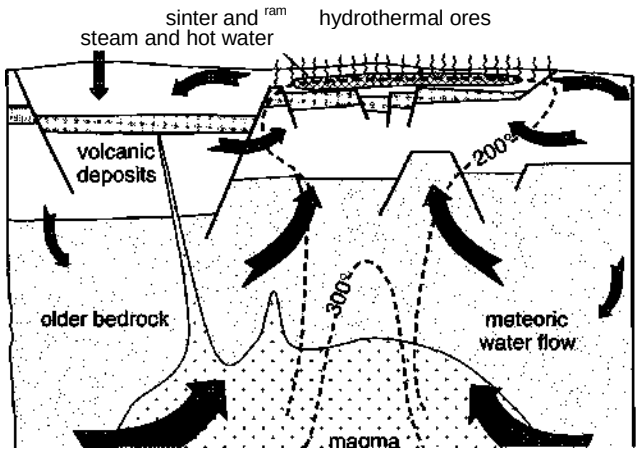


Figure 4-37 Schematic section through a hydrothermal system developed above a magma chamber in a silicic volcanic terrane. .After Henley and Ellis (1983). Copyright © with permission from Elsevier Science. Oxygen isotopic studies have shown that most of the water flow (dark arrows) is recirculated meteoric water. Juvenile magmatic water is typically of minor importance.

SUGGESTED FURTHER READINGS

General Field Relationship of Igneous Rocks

Compton, R. R. (1985). *Geology in the Field.* Wiley. New York.

Lahee, F. H. (1961). *Field Geology.* McGraw Hill. New York.

Thorpe, R. S. and G. C. Brown (1985). *The Field Description of Igneous Rocks.* Halsted. New York.

Volcanism and Volcanic Land forms

Bullard, F. M. (1976). *Volcanoes of the Earth.* Univ. of Queensland Press. St. Lucia.

Decker, R. W. and B. Decker (1998). *Volcanoes.* W.H. Freeman. New York.

Decker, R.W., T. L. Wright, and P. H. Stauffer (1987). *Volcanism in Hawaii.* USGS Prof. Paper 1350.

Francis, P. (1993). *Volcanoes, a Planetary Perspective.* Oxford Clarendon Press. Oxford.

Sigurdsson, H., B. Houghton, S. R. McNutty, H. Rymer, and S. Stix, eds. (2000). *Encyclopedia of Volcanoes.* Academic Press, San Diego.

Simkin, T. and R. S. Fiske (1983). *Krakatau 1883: The Volcanic Eruption and Its Effects.* Smithsonian Inst. Washington, D.C.

Williams, H. and A. R. McBirney (1979.) *Volcanology.* Freeman, Cooper. San Francisco.

Pyroclastics

Cas, R. A. F. and J. V. Wright (1988). *Volcanic Successions, Modern and Ancient.* Unwin Hyman. London.

Fisher, R. V. and H. -U. Schminke (1984). *Pyroclastic Rocks.* Springer-Verlag. Berlin.

Intrusions

Bergantz, G. W. (1991). Chemical and Physical Characterization of Plutons. In Kerrick, D. M. (ed). *Contact Metamor-phism.* Reviews in Mineralogy v. 26. Mineralogical Society of America.

Buddington, A. F. (1959). Granite emplacement with special reference to North America. *Geol. Soc. Am. Bull.* 70: 671-747.

Clarke, D. B. (1992). *Granitoid Rocks.* Chapman & Hall, London. Chapter 2: Field Relations.

Crittenden, M. D. and P. J. Coney (1980). *Cordilleran Meta-morphic Core Complexes.* Geol. Soc. Amer., Memoir **153.**

Didier, J. (1973). *Granites and Their Enclaves.* Elsevier. New York.

Marsh, B. D. (1982). On the mechanisms of igneous diapirism, sloping, and zone melting. *Am. J. Sci.* 282: 808-855.

Newell, G. and N. Rast (eds.). (1970). *Mechanisms of Igneous Intrusion.* Gallery Press. Liverpool.

Paterson, S. R., R. H. Vernon, and T. K. Fowler (1991). Aureole Tectonics. In Kerrick, D. M. (ed). *Contact Metamorphism.* Reviews in Mineralogy v. 26. Mineralogical Society of America.

Paterson, S. R. and T. K. Fowler (1993). Reexamining pluton emplacement mechanisms. 7. *Struct. Geol.* **15:**191-206.

Hydrothermal Systems

Burnham, C. W. (1967). Hydrothermal fluids at the magmatic stage. In: *Geochemistry of Hydrothermal Ore Deposits.* Holt, Reinhart and Winston. New York. pp. 38-76.

Henley, R. W. and A. J. Ellis (1983). Geothermal systems ancient and modern: A geochemical review. *Earth Sci. Rev.*, **19,** 1-50.