CHAPTER 3

Igneous Rocks, the Origin and Evolution of Magma, and Intrusive Activity

The distinctive cliffs and domes of Yosemite Valley, including Half Dome shown on the right of this image, are all formed of the igneous rock granite. *Photo (c) Parvinder Sethi*

LEARNING OBJECTIVES

- Use the rock cycle to understand the relationships between igneous, sedimentary, and metamorphic rocks.
- Classify the most common igneous rock types using their texture and chemical composition.
- Understand the conditions under which rocks in the Earth's interior melt to form magma.
- Describe the magmatic processes that produce igneous rocks of varying composition.
- Differentiate between the various types of intrusive igneous structures.
- Relate magmatic activity to plate tectonic theory.

In chapter 2 you learned all about minerals. In the next few chapters you will learn about the three rock types (igneous, metamorphic, and sedimentary), how they form, and what they can tell us about processes in the interior of the Earth and on its surface. In this chapter we will explore **igneous rocks**—those formed when magma cools and solidifies. The study of igneous rocks provides us with an insight into the composition of the Earth's mantle and the processes that cause it to melt. The wide variety of igneous rocks found on the Earth's surface tells us about the processes that occur within magma chambers in the crust. Magmatic processes are the cause of spectacular volcanic eruptions (covered in chapter 4), and magma that has cooled and crystallized deep in the Earth's crust forms the backbones of many large mountain chains. For example, the awe-inspiring formations of Yosemite Valley shown in the opening image of this chapter are formed of granite, a rock that forms from magma cooling slowly beneath the Earth's surface.

We begin this chapter by introducing the rock cycle, a conceptual device that shows the interrelationship between igneous, sedimentary, and metamorphic rocks. We then discuss how igneous rocks are described in terms of their texture and composition, and how these properties are used to classify igneous rocks. Following this, we will focus on how igneous rocks are formed by examining the conditions under which the mantle melts to form magma, and the magmatic processes that produce the wide variety of igneous rocks seen on Earth's surface. We will then focus on intrusive igneous structures and their relationship to other rocks in the Earth's crust. Finally, we will conclude by relating igneous activity to plate tectonic theory.

THE ROCK CYCLE

A **rock** is naturally formed, consolidated material usually composed of grains of one or more minerals. You will see in chapter 5 how some minerals break down chemically and form new minerals when a rock finds itself in a new physical setting. For instance, feldspars that may have formed at high temperatures deep within the Earth can react with surface waters to become clay minerals at the Earth's surface.

As mentioned in chapter 1, the Earth changes because of its internal and external heat engines. If the Earth's internal engine had died (and tectonic forces had therefore stopped operating), the external engine plus gravity would long ago have leveled the continents virtually at sea level. The resulting sediment would have been deposited on the sea floor. Solid Earth would not be changing (except when struck by a meteorite or other extraterrestrial body). The rocks would be at rest. The minerals, water, and atmosphere would be in *equilibrium* (and geology would be a dull subject). But this is not the case. The internal and external forces continue to interact, forcing substances out of equilibrium. Therefore, the Earth has a highly varied and ever-changing surface. And minerals and rocks change as well.

A useful aid in visualizing these changing relationships is the **rock cycle** shown in figure 3.1. The three major rock types—igneous, metamorphic, and sedimentary—are shown. As you see, each may form at the expense of another if it is forced out of equilibrium with its physical or climatic environment by either internal or surficial forces. It is important to be aware that rock moves from deep to shallow, and from high to low temperature and pressure in response to tectonic forces and isostasy (covered in chapter 1).

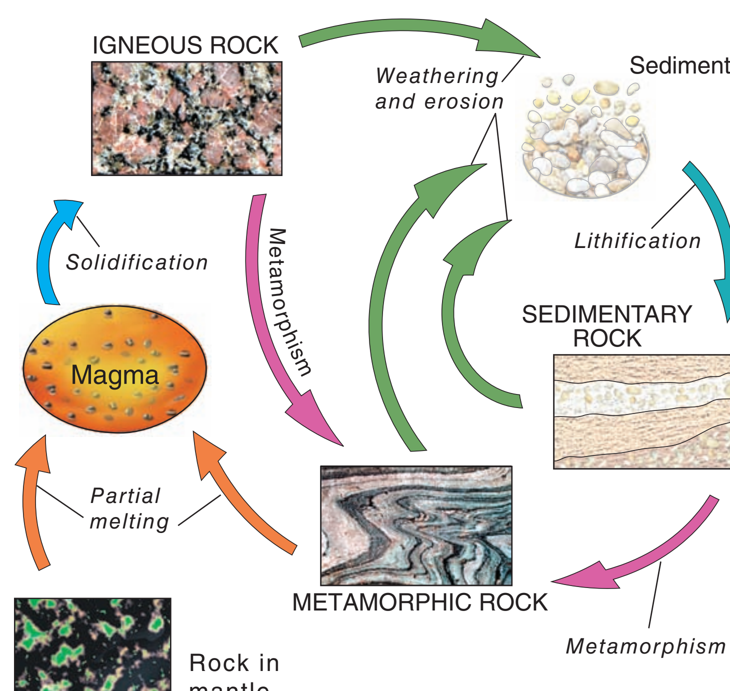


FIGURE 3.1

The rock cycle. The arrows indicate the processes whereby one kind of rock is changed to another. For clarity, arrows are not used to show that metamorphic rock can be re-metamorphosed to a different metamorphic rock or that igneous rock can be remelted to form new magma.

As described in chapter 1, *magma* is molten rock. (Magma may contain suspended solid crystals and gas.) *Igneous rocks* form when magma solidifies. If the magma is brought to the surface (where it is called lava) by a volcanic eruption, it will solidify into an *extrusive* igneous rock. Magma may also solidify very slowly beneath the surface. The resulting *intrusive* igneous rock may be exposed later after uplift and erosion remove the overlying rock (as shown in figure 1.14). The igneous rock, being out of equilibrium, may then undergo *weathering* and *erosion,* and the debris produced is transported and ultimately deposited (usually on a sea floor) as *sediment.* If the unconsolidated sediment becomes *lithified* (cemented or otherwise consolidated into a rock), it becomes a *sedimentary rock.* The rock is buried by additional layers of sediment and sedimentary rock. This process can only bury layered rock in the uppermost crust to a depth of several kilometers. Tectonic forces are required to transport sedimentary (and volcanic) rock to lower levels in the crust. Heat and pressure increase with increasing depth of burial. If the temperature and pressure become high enough, as occurs in the middle and lower levels of continental crust, the original sedimentary rock is no longer in equilibrium and recrystallizes. The new rock that forms is called a *metamorphic rock.* If the temperature gets very high, the rock partially melts, producing magma and completing the cycle.

The cycle can be repeated, as implied by the arrows in figure 3.1. However, there is no reason to expect all rocks to go through each step in the cycle. For instance, sedimentary rocks might be uplifted and exposed to weathering, creating new sediment.

We should emphasize that the rock cycle is a conceptual device to help students place the common rocks and how they form in perspective. As such, it is a simplification and does not encompass all geologic processes. For instance, most magma comes from partial melting of the mantle, rather than from recycled crustal rocks.

The Rock Cycle and Plate Tectonics

One way of relating the rock cycle to plate tectonics is illustrated by an example from what happens at a convergent plate boundary (figure 3.2). *Magma* is created in the zone of melting above the subduction zone. The magma, being less dense than adjacent rock, migrates upward toward the surface. A volcanic eruption takes place if magma reaches the surface. The magma solidifies

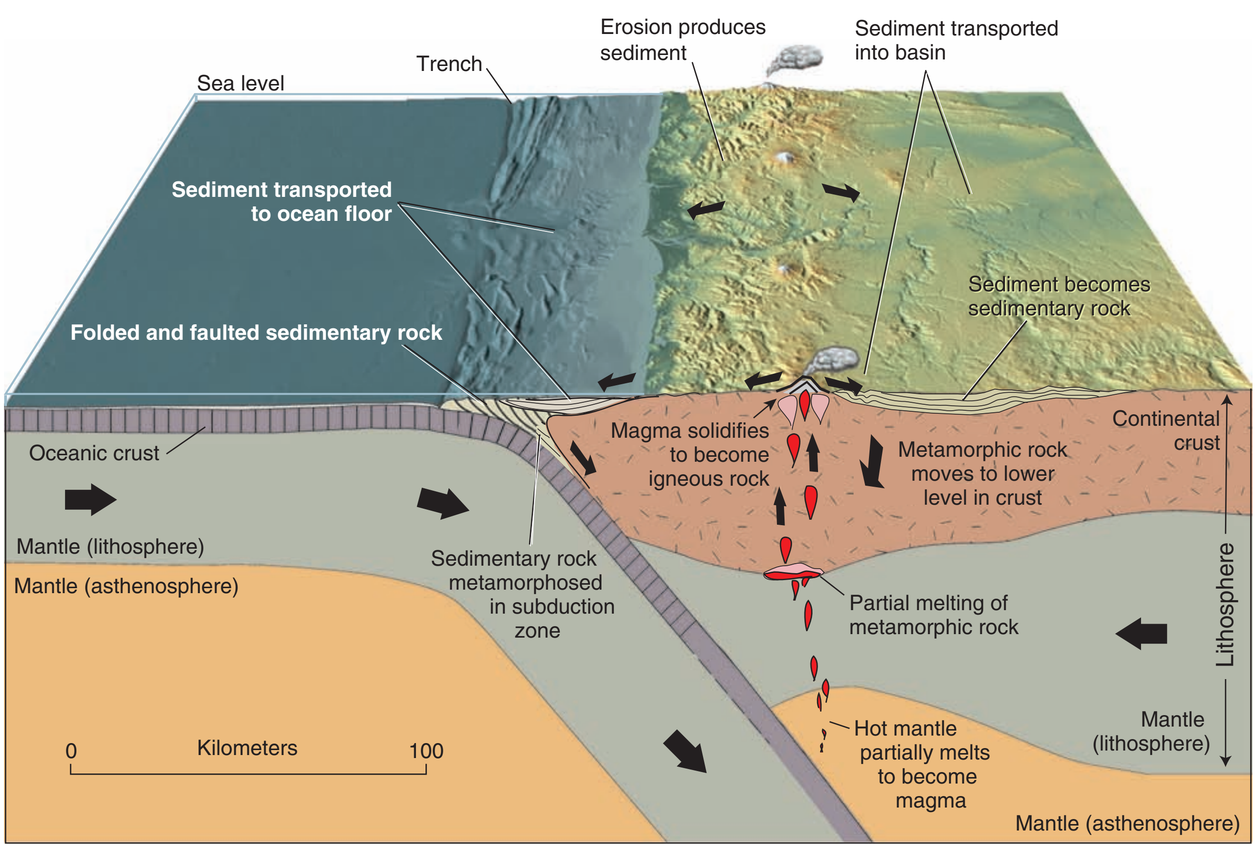


FIGURE 3.2

The rock cycle with respect to a convergent plate boundary. Magma formed within the mantle solidifies as igneous rock at the volcano. Sediment from the eroded volcano collects in the basin to the right of the diagram. Sediment converts to sedimentary rock as it is buried by more sediment. Deeply buried sedimentary rocks are metamorphosed. The most deeply buried metamorphic rocks partially melt, and the magma moves upward. An alternate way the rock cycle works is shown on the left of the diagram. Sediment from the continent (and volcano) becomes sedimentary rock, some of which is carried down the subduction zone. It is metamorphosed as it descends. It may contribute to the magma that forms in the mantle above the subduction zone.

into *igneous rock.* The igneous rock is exposed to the atmosphere and subjected to *weathering* and *erosion.* The resulting *sediment* is transported and then deposited in low-lying areas. In time, the buried layers of sediment lithify into *sedimentary rock.* The sedimentary rock becomes increasingly deeply buried as more sediment accumulates and tectonic forces push it deeper. After the sedimentary rock is buried to depths exceeding several kilometers, the heat and pressure become too great, and the rock recrystallizes into a *metamorphic rock.* As the depth of burial becomes even greater (several tens of kilometers), the metamorphic rock may find itself in a zone of melting. Temperatures are now high enough so that the metamorphic rock partially melts. Magma is created, thus completing the cycle.

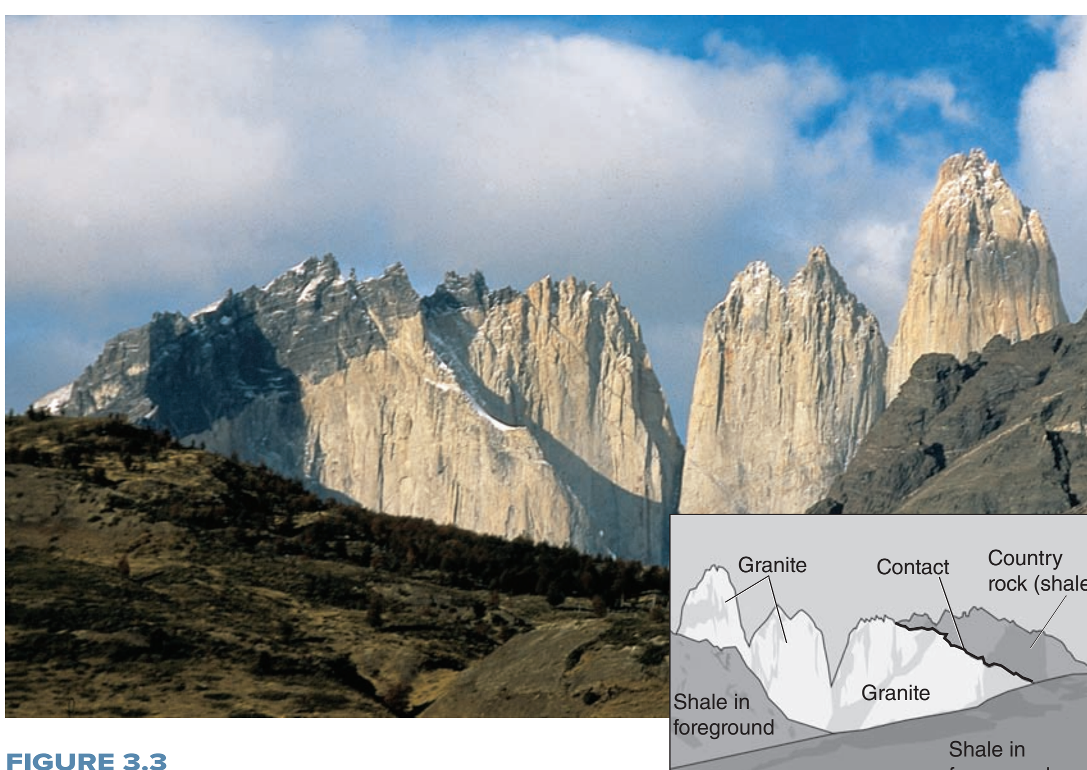
The rock cycle diagram reappears on the opening pages of chapters 4 through 7. The highlighted portion of the diagram will indicate where the material covered in each chapter fits into the rock cycle.

IGNEOUS ROCKS

If you go to the island of Hawaii, you might observe red-hot lava flowing over the land and, as it cools, solidifying into the fine-grained (the grains are too small to be seen with the naked eye), black rock we call basalt. Basalt is an igneous rock, rock that has solidified from magma. **Magma** is molten rock, usually rich in silica and containing dissolved gases. (**Lava** is magma on the Earth's surface.) Igneous rocks may be either **extrusive** if they form at the Earth's surface (e.g., basalt) or **intrusive** if magma solidifies underground. **Granite,** a coarse-grained (the crystals are large enough to be seen with the naked eye) rock composed predominantly of feldspar and quartz, is an intrusive rock. In fact, granite is the most abundant intrusive rock found in the continents.

Unlike the volcanic rock in Hawaii, nobody has ever seen magma solidify into intrusive rock. So what evidence suggests that bodies of granite (and other intrusive rocks) solidified underground from magma?

- Mineralogically and chemically, intrusive rocks are essentially identical to volcanic rocks.
- Volcanic rocks are fine-grained or glassy due to their rapid solidification; intrusive rocks are generally coarse-grained, which indicates that the magma crystallized slowly underground. Experiments show that the slower cooling of liquids results in larger crystals.
- Experiments have confirmed that most of the minerals in these rocks can form only at high temperatures. Other experiments indicate that some of the minerals could have formed only under high pressures, implying they were deeply buried. More evidence comes from examining *intrusive contacts,* such as those shown in figures 3.3 and 3.4. (A **contact** is a surface separating different rock types. Other types of contacts are described elsewhere in this book.)



Geologist's View

FIGURE 3.3

Granite (light-colored rock) solidified from magma that intruded dark-colored country rock in Torres del Paine, Chile. The dark-colored country rock is shale deposited in a marine environment. The spires are erosional remnants of rock that were once deep underground. *Photo by Kay Kepler.*

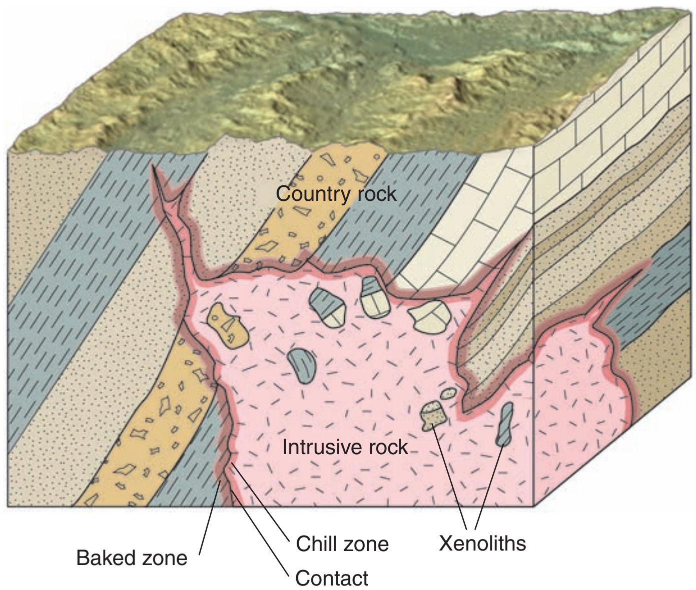


FIGURE 3.4

Igneous rock intruded preexisting rock (country rock) as a liquid. (Xenoliths are usually much smaller than indicated.)

- Preexisting solid rock, *country rock,* appears to have been forcibly broken by an intruding liquid, with the magma flowing into the fractures that developed. **Country rock,** incidentally, is an accepted term for any older rock into which an igneous body intruded.
- Close examination of the country rock immediately adjacent to the intrusive rock usually indicates that it appears "baked," or *metamorphosed,* close to the contact with the intrusive rock.
- Rock types of the country rock often match **xenoliths,** fragments of rock that are distinct from the body of igneous rocks in which they are enclosed.
- In the intrusive rock adjacent to contacts with country rock are **chill zones,** finer-grained rocks that indicate magma solidified more quickly here because of the rapid loss of heat to cooler rock.

Laboratory experiments have greatly increased our understanding of how igneous rocks form. However, geologists have not been able to make coarsely crystalline granite artificially. Only very fine-grained rocks containing the minerals of granite have been made from artificial magmas, or "melts." The temperature and pressure at which granite apparently forms can be duplicated in the laboratory—but not the time element. According to calculations, a large body of magma requires over a million years to solidify completely. This very gradual cooling causes the coarse-grained texture of most intrusive rocks. Chemical processes involving silicates may be exceedingly slow. Yet another problem in trying to apply experimental procedures to real rocks is determining the role of water and other gases in the crystallization of rocks such as granite. Only a small amount of gas is retained in rock crystallized underground from a magma, but large amounts of gas (especially water vapor) are released during volcanic eruptions. Laboratory experiments that involve melt solidification under gaseous conditions provide us with insight into the role that gases in underground magma might have played before they escaped. One example indicates the importance of gases. Laboratory studies have shown that the amount of water present in a granitic magma can control both the temperature range over which the granite crystallizes and the minerals that form from the magma as it cools. Later in this chapter, we will discuss how magma is formed and the processes that occur as it cools and crystallizes. But first, let's familiarize ourselves with igneous rocks and how they are described and named.

Classification of Igneous Rocks

Igneous rocks are named based on their texture and chemical composition. The texture of an igneous rock (its appearance in terms of grain size and presence of gas bubbles, for example) gives you information on where that rock formed, whether beneath the surface as an intrusive igneous rock or on the surface as an extrusive igneous rock. The chemistry of an igneous rock tells you about the origin of the magma and how it evolved before finally solidifying. Geologists infer the chemistry of an igneous rock by looking at the minerals it is composed of or, if the rock is too fine-grained to see individual minerals, by its color. Because magma can either cool slowly in a magma chamber, forming an intrusive rock, or cool rapidly on the surface, forming an extrusive rock, one name is given to the coarse-grained intrusive version and another to the fine-grained extrusive version. For example, earlier we described granite, a coarse-grained rock composed predominantly of feldspar and quartz. If a magma of the same composition as granite were to erupt onto the surface and cool rapidly, it would still form a rock composed predominantly of feldspar and quartz, but it would be fine-grained. This fine-grained rock would be called rhyolite. Some rocks have a very distinctive texture and are named for that texture. For example, some rhyolite cools so quickly it forms a black glass. This rock is known as obsidian.

Figure 3.5 shows the relationship between texture, chemistry, mineral content, and igneous rock names. Pictures of the most common igneous rocks are shown. You should examine this figure carefully while reading the following sections on igneous rock textures, the chemistry of igneous rocks, and identifying igneous rocks.

Igneous Rock Textures

Texture refers to a rock's appearance with respect to the size, shape, and arrangement of its grains or other constituents. Igneous textures can be divided into three groups: **crystalline rocks** are made up of interlocking crystals (of, for instance, the minerals quartz and feldspar); **glassy rocks** are composed primarily of glass and contain few, if any, crystals; and **fragmental rocks** are composed of fragments of igneous material. As you will see, the texture of an igneous rock can tell you a lot about where and how it formed.

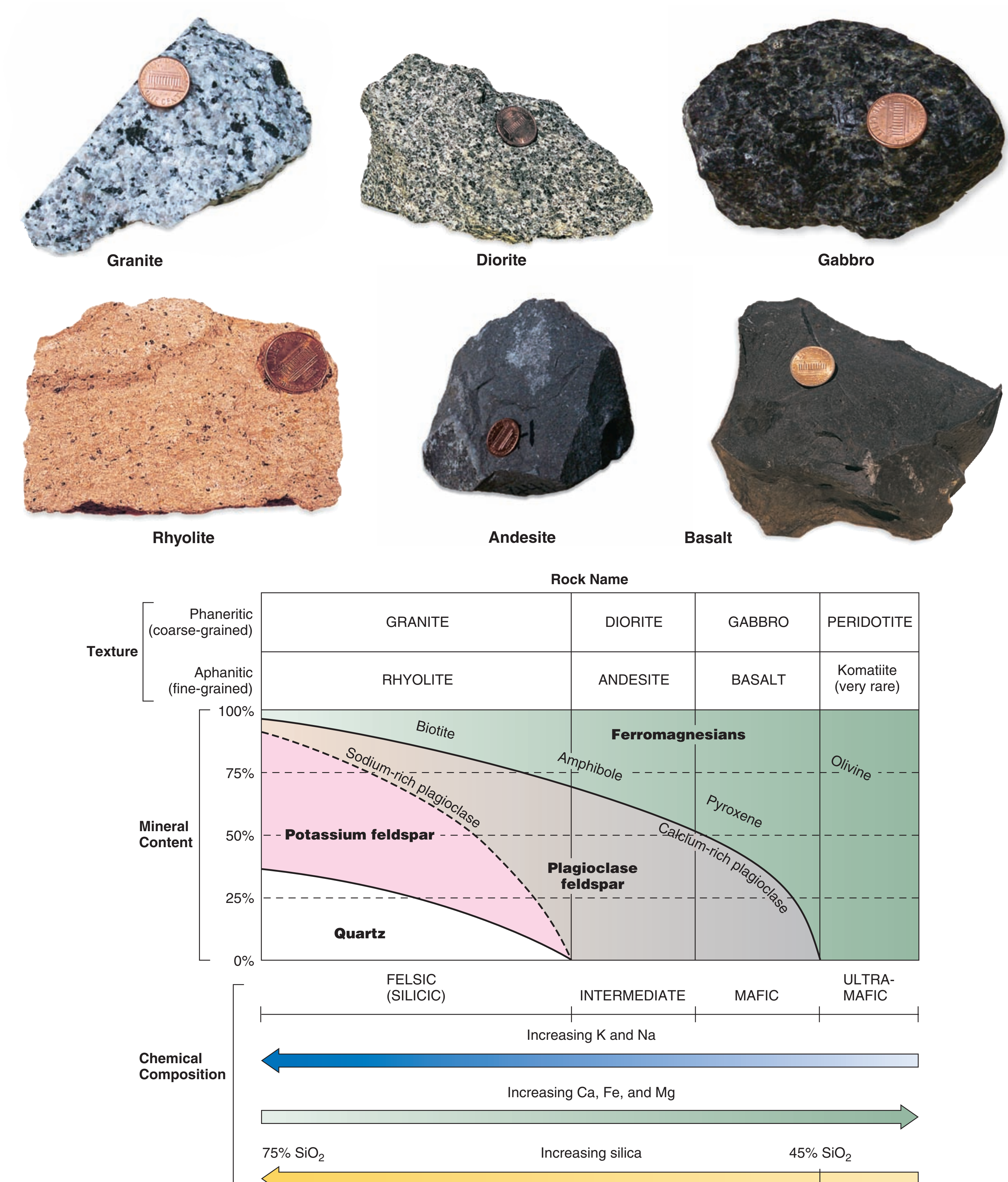


FIGURE 3.5

Classification chart for the most common igneous rocks. Rock names based on special textures are not shown. Sodium-rich plagioclase is associated with silicic rocks, whereas calcium-rich plagioclase is associated with mafic rocks. The names of the particular ferromagnesian minerals (biotite, etc.) are placed in the diagram at the approximate composition of the rocks in which they are most likely to be found. Samples of common igneous rocks are shown for reference. Do not try to identify real rocks by simply comparing them to photos—use the properties, such as identifying minerals and their amounts. *Photos by C. C. Plummer*

Crystalline Textures

Most igneous rocks form from the cooling and solidification of magma. Crystals form and grow as the melt cools, and the result is a crystalline texture. The most significant aspect of texture in crystalline igneous rocks is grain (or crystal) size. Two critical factors determine grain size during the solidification of igneous rocks: rate of cooling and viscosity. If magma cools rapidly, the atoms have time to move only a short distance; they bond with nearby atoms, forming only small crystals. Grain size is controlled to a lesser extent by the viscosity of the magma. Atoms in highly viscous lava cannot move as freely as those in a more fluid lava. Hence, a rock formed from viscous lava is more likely to be finer grained than one formed from more fluid lava.

Extrusive rocks are typically fine-grained or **aphanitic** (*a*, "not"; *phaner,* "visible"); that is, their crystals are too small to see easily with the naked eye. The grains, if they are crystals, are small because magma cools rapidly at the Earth's surface, so they have less time to form. Some intrusive rocks are also fine-grained; these occur as smaller bodies that apparently solidified near the surface upon intrusion into relatively cold country rock (probably within a couple kilometers of the Earth's surface). Igneous rocks that formed at considerable depth—usually more than several kilometers—are called **plutonic rocks** (after Pluto, the Roman god of the underworld). Characteristically, these rocks are coarse-grained, reflecting the slow cooling and solidification of magma.

For our purposes, coarse-grained or **phaneritic** (from *phaner,* meaning visible) rocks are defined as those in which the crystals are large enough to be seen easily with the naked eye. The crystals of plutonic rocks are commonly interlocked in a mosaic pattern (figure 3.6). An extremely coarse-grained (crystals ranging in size from a few centimeters to several meters in length) igneous rock is called a **pegmatite** (see box 3.1).

So, for practical purposes, if you can see the individual crystals in an igneous rock, you can regard it as phaneritic. If not, consider it aphanitic.

Some rocks contain **phenocrysts,** larger crystals that are enclosed in a *groundmass* of finer-grained crystals or glass. This is called a **porphyritic** texture. A milk chocolate bar containing whole almonds has the appearance of porphyritic texture where the chocolate represents the finer-grained groundmass and the almonds represent the phenocrysts. If the groundmass is aphanitic, the rock is considered to be extrusive. For instance, figure 3.7 shows an extrusive rock called *porphyritic andesite.* Porphyritic extrusive rocks are usually interpreted as having begun crystallizing slowly underground within a magma chamber followed by eruption and rapid solidification of the remaining magma at the Earth's surface. Some porphyritic rocks have a phaneritic groundmass. The larger phenocrysts enclosed in the groundmass are much bigger, usually 2 or more centimeters across. These rocks are considered to be intrusive. *Porphyritic granite* is an example.

Glassy Textures

With extremely rapid or almost instantaneous cooling, individual atoms in the lava are "frozen" in place, forming glass rather than crystals. Remember, you learned in chapter 2 that, by definition, minerals have a crystalline structure where the atoms are arranged in an orderly, repeating, three-dimensional pattern. In a glass, the atoms are not ordered, so minerals are not present. **Obsidian** (figure 3.8), which is a dark volcanic glass, is one of the few rocks that is not composed of minerals. Because obsidian breaks with a conchoidal fracture (see chapter 2), it can be broken in such a way as to produce very sharp edges. The manufacture of obsidian tools dates back to the Stone Age. Today, some surgeons use obsidian for scalpel blades as the cutting edge can be many times sharper than steel surgical scalpels.

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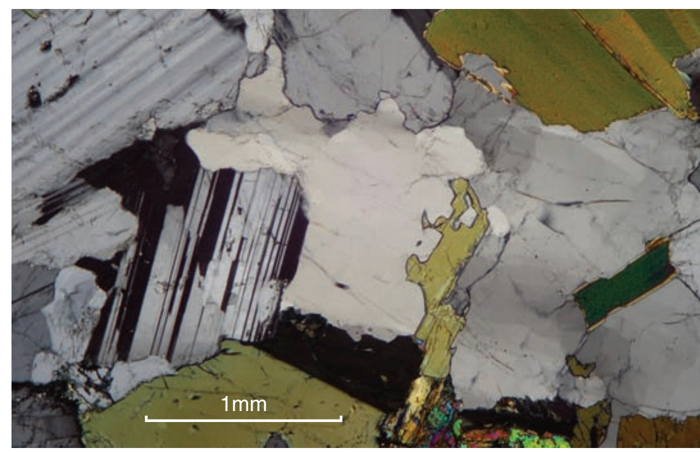

B

FIGURE 3.6

(*A*) Phaneritic (coarse-grained) texture characteristic of plutonic rock. Feldspars are white and pink. Although this quartz is transparent, it appears light gray. Biotite mica is black. A U.S. penny is used for a scale because the roof of the monument is 1 millimeter thick and 1 centimeter wide. (*B*) A similar rock seen through a polarizing microscope. Note the interlocking crystal grains of individual minerals. *Photo A by C.C. Plummer. Photo B by L. Hammersley.*

Textures Due to Trapped Gas

A magma deep underground is under high pressure, generally high enough to keep all its gases in a dissolved state. On eruption, the pressure is suddenly released and the gases come out of solution. This is analogous to what happens when a bottle

IN GREATER DEPTH 3.1

Pegmatite—A Rock Made of Giant Crystals

Pegmatites are extremely coarse-grained igneous rocks (see box figure 1). In some pegmatites, crystals are as large as 10 meters across. Strictly speaking, a pegmatite can be of diorite, gabbro, or granite. However, the vast majority of pegmatites are silica-rich, with very large crystals of potassium feldspar, sodium-rich plagioclase feldspars, and quartz. Hence, the term *pegmatite* generally refers to a rock of granitic composition (if otherwise, a term such as *gabbroic pegmatite* is used). Pegmatites are interesting as geological phenomena and important as minable resources.

The extremely coarse texture of pegmatites is attributed to both slow cooling and the low viscosity (resistance to flow) of the fluid from which they form. Lava solidifying to rhyolite is very viscous. Magma solidifying to granite, being chemically similar, should be equally viscous.

Pegmatites, however, probably crystallize from a fluid composed largely of water under high pressure. Water molecules and ions from the parent, granitic magma make up a residual magma. Geologists believe the following sequence of events accounts for most pegmatites.

As a granite pluton cools, increasing amounts of the magma solidify into the minerals of a granite. By the time the pluton is well over 90% solid, the residual magma contains a very high amount of silica and ions of elements that will crystallize into potassium and sodium feldspars. Also present are elements that could not be accommodated into the crystal structures of the common minerals that formed during the normal solidification phase of the pluton. Fluids, notably water, that were in the original magma are left over as well. If no fracture above the pluton permits the fluids to escape, they are sealed in, as in a pressure cooker. The watery residual magma has a low viscosity, which allows appropriate atoms to migrate easily toward growing crystals. The crystals add more and more atoms and grow very large.

Pegmatite bodies are generally quite small. Many are pod-like structures, located either within the upper portion of a granite pluton or within the overlying country rock near the contact with granite, the fluid body evidently having squeezed into the country rock before solidifying. Pegmatite dikes are fairly common, especially within granite plutons, where they apparently filled cracks that developed in the already solid granite. Some pegmatites form small dikes along contacts between granite and country rock, filling cracks that developed as the cooling granite pluton contracted.

Most pegmatites contain only quartz, feldspar, and perhaps mica. Minerals of considerable commercial value are found in a few pegmatites. Large crystals of muscovite mica are mined from pegmatites. These crystals are called "books" because the cleavage flakes (tens of centimeters across) look like pages. Because muscovite is an excellent insulator, the cleavage sheets are used in electrical devices, such as toasters, to separate uninsulated electrical wires. Even the large feldspar crystals in pegmatites are mined for various industrial uses, notably the manufacture of ceramics.

BOX 3.1 ■ FIGURE 1

Pegmatite in northern Victoria Land, Antarctica. The knife is 8 centimeters long. The black crystals are tourmaline. Quartz and feldspar are light colored. *Photo by C. C. Plummer*

Many rare elements are mined from pegmatites. These elements were not absorbed by the minerals of the main pluton and so were concentrated in the residual pegmatitic magma, where they crystallized as constituents of unusual minerals. Minerals containing the element lithium are mined from pegmatites. Lithium becomes part of a sheet silicate structure to form a pink or purple variety of mica (called lepidolite). Uranium ores, similarly concentrated in the residual melt of magmas, are also extracted from pegmatites.

Some pegmatites are mined for gemstones. Emerald and aquamarine, varieties of the mineral beryl, occur in pegmatites that crystallized from a solution containing the element beryllium. A large number of the world's very rare minerals are found only in pegmatites, many of these in only one known pegmatite body. These rare minerals are mainly of interest to collectors and museums.

Hydrothermal veins (described in chapter 7) are closely related to pegmatites. Veins of quartz are common in country rock near granite. Many of these are believed to be caused by water that escapes from the magma. Silica dissolved in the very hot water cakes on the walls of cracks as the water cools while traveling surfaceward. Sometimes valuable metals such as gold, silver, lead, zinc, and copper are deposited with the quartz in veins. (See chapter 7 for more on veins.)

of beer or soda is opened. Because the drink was bottled under pressure, the gas (carbon dioxide) is in solution. Uncapping the drink relieves the pressure, and the carbon dioxide separates from the liquid as gas bubbles. If you freeze the newly opened drink very quickly, you have a piece of ice with small, bubble-shaped holes. Similarly, when a lava solidifies while gas is bubbling through it, holes are trapped in the rock, creating a distinctive vesicular texture. **Vesicles** are cavities in extrusive

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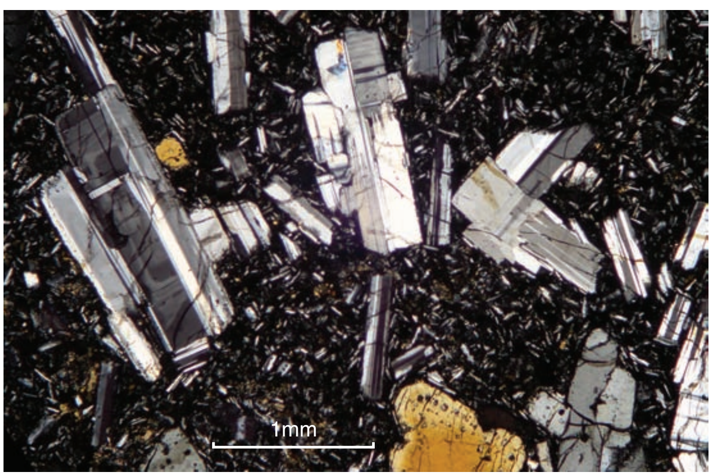


B

FIGURE 3.7

Porphyritic andesite. A few large crystals (phenocrysts) are surrounded by a great number of fine grains. (*A*) Hand specimen. Grains in groundmass are too fine to see. (*B*) Photomicrograph (using polarized light) of a similar rock. The black-and-white striped phenocrysts are plagioclase, and the yellow ones are pyroxene. *Photo A © Parvinder Sethi. Photo B by L. Hammersley.*

FIGURE 3.8

Obsidian. Notice the curved surfaces formed by conchoidal fracture. *Photo by C. C. Plummer*

rock resulting from gas bubbles that were in lava, and the texture is called *vesicular.* A vesicular rock has the appearance of Swiss cheese (whose texture is caused by trapped carbon dioxide gas). *Vesicular basalt* is quite common (figure 3.9). **Scoria,** a highly vesicular basalt, actually contains more gas space than rock.

In more viscous lavas, where the gas cannot escape as easily, the lava is churned into a froth (like the head in a glass of beer). When cooled quickly, it forms **pumice** (figure 3.10), a frothy glass with so much void space that it floats in water. Powdered pumice is used as an abrasive because it can scratch metal or glass. For example, powdered pumice is used in the production of pencil erasers. You may also have a piece of pumice in your shower that you use to buff the bottom of your feet.

FIGURE 3.9

Vesicular basalt. *Photo © Dr. Parvinder Sethi*

Fragmental Texture

When trapped gases are released during an eruption, it can lead to lava being blasted out of the vent as fragments of rock known as **pyroclasts** (from the Greek *pyro,* "fire," and *clast,* "broken"). *Pyroclastic debris* is also known as *tephra.* When pyroclastic

A

B

FIGURE 3.10

(*A*) A boulder of pumice can be easily carried because it is mostly air. (*B*) Seen close up, pumice is a froth of volcanic glass. *Photo A by Diane Carlson; photo B by C. C. Plummer*

material (ash, pumice, or crystalline rock fragments) accumulates and is cemented or otherwise consolidated, the new rock is called **tuff,** or **volcanic breccia,** depending on the size of the fragments. A tuff (figure 3.11) is a rock composed of fine-grained pyroclastic particles (dust and ash). A volcanic breccia is a rock that includes larger pieces of volcanic rock.

Chemistry of Igneous Rocks

The chemical composition of magma determines which minerals and how much of each will crystallize when an igneous rock forms. Because magma contains quite a lot of silicon and oxygen, igneous rocks are composed primarily of the silicate minerals quartz, plagioclase feldspar, potassium feldspar, amphibole, pyroxene, biotite, and olivine. The lower part of figure 3.5 shows the relationship of chemical composition to mineral composition. A magma that is rich in silica, aluminum, potassium, and sodium will crystallize minerals that contain those elements (feldspar and quartz). A magma that is rich in iron and magnesium and calcium will contain a lot of the dark-colored ferromagnesian minerals (pyroxene, amphibole, olivine, and biotite).

Chemical analyses of rocks are reported as weight percentages of oxides (e.g., SiO_2, MgO, Na_2O, etc.) rather than as separate elements (e.g., Si, O, Mg, Na). For virtually all igneous rocks, SiO_2 (silica) is the most abundant component. The amount of SiO_2 varies from about 45% to 75% of the total weight of common igneous rocks. Based on the amount of silica, igneous rocks are classified into four groups, which are, in order of decreasing silica content, felsic, intermediate, mafic, and ultramafic. Figure 3.12 shows the average chemical composition for felsic, intermediate, and mafic rocks. It is also useful to refer to figure 3.5 as you read the following sections because it shows the minerals associated with each of the four groups of igneous rocks as well as pictures of the major igneous rock types described next.

FIGURE 3.11

Volcanic tuff with a quarter for scale. *Photo © Dr. Parvinder Sethi*

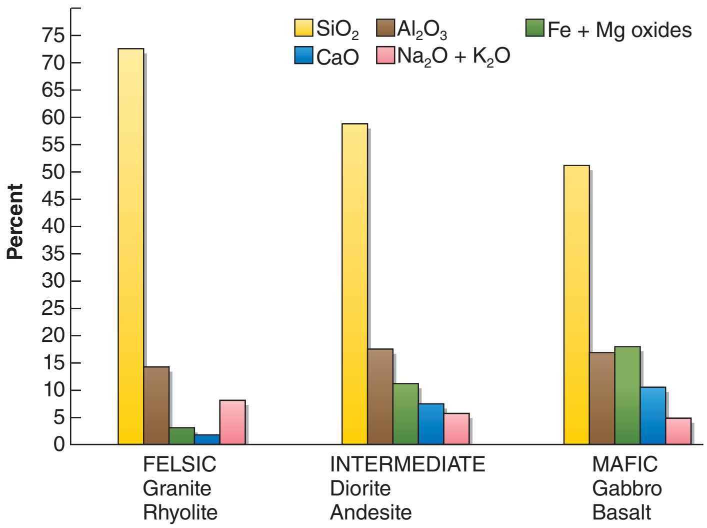


FIGURE 3.12

The average chemical composition of felsic, intermediate, and mafic rocks. Composition is given in weight percent of oxides. Note that as the amount of silica decreases, the oxides of Na and K decrease, and the oxides of Ca, Fe, and Mg increase. Al oxide does not vary significantly.

Felsic Rocks

Rocks with a silica content of 65% or more (by weight) are considered to be silica-rich (figure 3.12). The remaining 25% to 35% of these rocks is mostly aluminum oxide (Al_2O_3) and oxides of sodium (Na_2O) and potassium (K_2O), and they tend to contain only very small amounts of the oxides of calcium (CaO), magnesium (MgO), and iron (FeO and Fe_2O_3). These are called **felsic** rocks—silica-rich igneous rocks with a relatively high content of potassium and sodium (the *fel* part of the name comes from *feldspar,* which crystallizes from the potassium, sodium, aluminum, and silicon oxides; *si* in *felsic* is for silica). **Rhyolite** is a fine-grained, extrusive felsic rock. Granite is the coarse-grained intrusive equivalent of rhyolite.

Felsic rocks tend to be dominated by the light-colored minerals quartz, potassium feldspar, and plagioclase feldspar, with only small amounts of the dark ferromagnesian minerals. Because of this, felsic rocks are commonly light in color. A notable exception to this rule is the black glassy volcanic rock *obsidian.* So why is obsidian black—a color we usually associate with mafic rocks, such as basalt? If you look at a very thin edge of obsidian, it is transparent. Obsidian is, indeed, a form of stained glass. The black overall color is due to dispersion of extremely tiny magnetite crystals throughout the rock. Collectively, they act like pigment in ink or paint and give an otherwise clear substance color. For red obsidian, the magnetite (Fe_3O_4) has been exposed to air and has been oxidized to hematite (Fe_2O_3), which is red or red-brown.

Intermediate Rocks

Rocks with a silica content of between 55% and 65% are classified as **intermediate rocks.** Intermediate rocks contain significant amounts (30–50%) of dark ferromagnesian minerals like pyroxene and amphibole as well as light-colored plagioclase feldspar and small amounts of quartz. These can easily be discerned in **diorite,** the coarse-grained intermediate rock. Examine the picture of diorite in figure 3.5. You can see that about half the mineral grains are dark in color and the other half are white. **Andesite,** the fine-grained intermediate rock, is typically medium-gray or greenish-gray in color.

Mafic Rocks

Rocks with a silica content of between 45% and 55% (by weight) are considered *silica-poor,* even though SiO_2 is, by far, the most abundant constituent (figure 3.12). The remainder is composed mostly of the oxides of aluminum, calcium, magnesium, and iron. Rocks in this group are called **mafic**—silica-deficient igneous rocks with a relatively high content of magnesium, iron, and calcium. (The term *mafic* comes from *magnesium* and *ferrum,* the Latin word for iron.) Mafic rocks are made up predominantly of gray plagioclase feldspar and the ferromagnesian minerals pyroxene and olivine and tend to be dark in color. Mafic magma that cools slowly beneath the surface forms the coarse-grained, intrusive rock **gabbro.** If mafic magma erupts on the surface, it forms the dark, fine-grained, extrusive rock **basalt.**

Ultramafic Rocks

An **ultramafic rock** is one that contains less than 45% silica and is rich in iron, magnesium, and calcium. Ultramafic rocks are typically composed almost entirely of the ferromagnesian minerals olivine and pyroxene. No feldspars are present and, of course, no quartz. Note from the chart in figure 3.5 that **komatiite,** the volcanic ultramafic rock, is very rare. Ultramafic extrusive rocks are mostly restricted to the very early history of the Earth. For our purposes, they need not be discussed further. **Peridotite,** the coarse-grained intrusive rock, is composed of olivine and pyroxene and is the most abundant ultramafic rock.

Most ultramafic rocks come from the mantle rather than from the Earth's crust. Where we find large bodies of ultramafic rocks, the usual interpretation is that a part of the mantle has traveled upward as solid rock. In some cases, ultramafic rocks can form when ferromagnesian minerals crystallize in a mafic magma and settle down to the base of the magma chamber where they accumulate. This process, called crystal settling, is discussed later in this chapter.

Identifying Igneous Rocks

In order to identify an igneous rock, you must consider both its texture and the minerals it contains. Because of their larger mineral grains, plutonic rocks are the easiest to identify. The physical properties of each mineral in a plutonic rock can be determined more readily. And, of course, knowing what minerals are present makes rock identification a simpler task. For instance, gabbro is formed of coarse-grained ferromagnesian minerals and gray plagioclase feldspar. One can positively identify the feldspar on the basis of cleavage and, with practice, verify that no quartz is present. Gabbro's aphanitic counterpart

TABLE 3.1 Identification of Most Common Igneous Rocks

	Felsic	Intermediate	Mafic	Ultramafic
Phaneritic (coarse-grained)	Granite	Diorite	Gabbro	Peridotite
Aphanitic (fine-grained)	Rhyolite	Andesite	Basalt	—
Mineral Content	Quartz, feldspars (white, light gray, or pink). Minor ferromagnesian minerals.	Feldspars (white or gray) and about 35–50% ferromagnesian minerals. No quartz.	Predominance of ferro-magnesian minerals. Rest of rock is plagioclase feldspar (medium to dark gray).	Entirely ferromagne-sian minerals (olivine and pyroxene).
Color of Rock (most commonly)	Light-colored	Medium-gray or medium-green	Dark gray to black	Green to black

is basalt, which is also composed of ferromagnesian minerals and plagioclase. The individual minerals cannot be identified by the naked eye, however, and one must use the less reliable attribute of color—basalt is usually dark gray to black.

As you can see from figure 3.5, *granite* and *rhyolite* are composed predominantly of feldspars (usually white or pink) and quartz. Granite, being phaneritic, can be positively identified by verifying that quartz is present. Rhyolite is usually cream-colored, tan, or pink. Its light color indicates that ferromagnesian minerals are not abundant. *Diorite* and *andesite* are composed of feldspars and significant amounts of ferromagnesian minerals (30–50%). The minerals can be identified and their percentages estimated to indicate diorite. Andesite, being aphanitic, can tentatively be identified by its medium-gray or medium-green color. Its appearance is intermediate between light-colored rhyolite and dark basalt.

Use the chart in figure 3.5 along with table 3.1 to identify the most common igneous rocks. You may also find it helpful to turn to appendix B, which includes a key for identifying common igneous rocks.

HOW MAGMA FORMS

Now that you've learned about the common igneous rocks and how they vary in terms of texture and composition, it is time to explore how magma forms and what happens to it as it moves through the Earth's crust. Let's begin with the question of how magma forms.

A common misconception is that the lava erupted from volcanoes comes from an "ocean of magma" beneath the crust. However, as you have already learned in chapter 1, the mantle is not molten but solid rock. In order to understand how magma forms, we must consider the source of heat for melting rock, the conditions at depth beneath the Earth's surface, and the conditions under which rocks in the mantle and lower crust will melt.

Heat for Melting Rock

Most of the heat that contributes to the generation of magma comes from the very hot Earth's core (where temperatures are estimated to be greater than 5,000°C). Heat is conducted toward the Earth's surface through the mantle and crust. This is comparable to the way heat is conducted through the metal of a frying pan. Heat is also brought from the lower mantle when part of the mantle flows upward, either through convection (described in chapters 1 and 19) or by hot mantle plumes.

The Geothermal Gradient and Partial Melting

A miner descending a mine shaft notices a rise in temperature. This is due to the **geothermal gradient,** the rate at which temperature increases with increasing depth beneath the surface. Data show the geothermal gradient, on the average, to be about 3°C for each 100 meters (30°C/km) of depth in the upper part of the crust, decreasing in the mantle. Figure 3.13 shows the geothermal gradient for the crust and upper mantle.

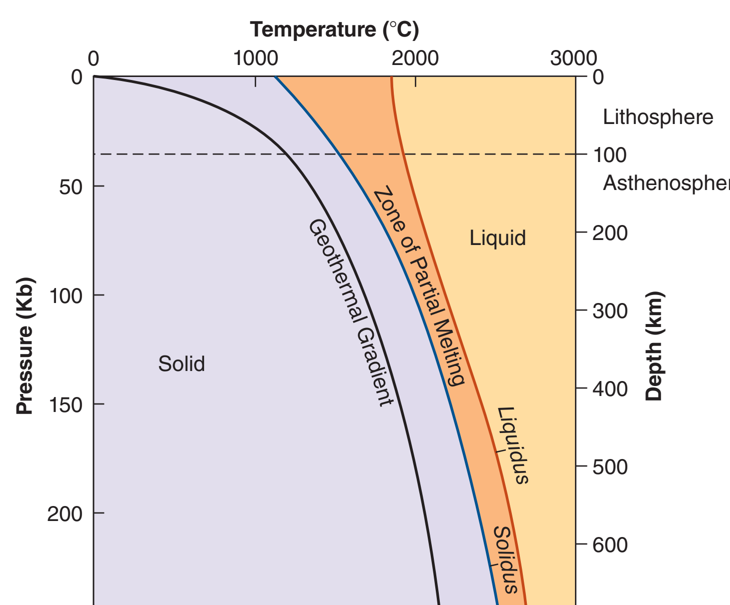


FIGURE 3.13

Geothermal gradient and zone of partial melting for mantle peridotite.

Unlike ice, which has a single melting point, rocks melt over a range of temperatures. This is because they are made up of more than one mineral, and each mineral has its own melting point. Figure 3.13 shows the range of temperatures over which rocks will melt below the Earth's surface. The dark blue line labeled *solidus* shows the temperature at any given depth, below which the rocks are completely solid. The dark orange line labeled *liquidus* shows the temperature above which the rocks are completely liquid. Between the solidus and the liquidus lies the zone of partial melting, within which the rocks are partly solid and partly molten.

It is important to notice that in figure 3.13 the geothermal gradient does not intersect the zone of partial melting. At all depths, the temperature is not high enough to allow the rock to melt, and no magma is forming. This is typical of the mantle in most locations. In order for magma to form, conditions must change so that the geothermal gradient can intersect the zone of partial melting. The two most common mechanisms believed to create these conditions are decompression melting and the addition of water.

Decompression Melting

The melting point of a mineral generally increases with increasing pressure. Pressure increases with depth in the Earth's crust, just as temperature does. So a rock that melts at a given temperature at the surface of the Earth requires a higher temperature to melt deep underground. **Decompression melting** takes place when a body of hot mantle rock moves upward and the pressure is reduced. Figure 3.14*A* shows the effect of decompression melting. Consider point *a*. At this pressure and temperature, the rock is below the zone of partial melting and will not melt. If, however, the pressure decreases, the geothermal gradient will move up as the pressure at any given temperature decreases. The rock at point *a* will now be at point *a′*, which is within the zone of partial melting, and magma will begin to form. As you will learn at the end of this chapter, decompression melting is an important process at divergent plate boundaries, where mantle material is rising and melting beneath mid-oceanic ridges.

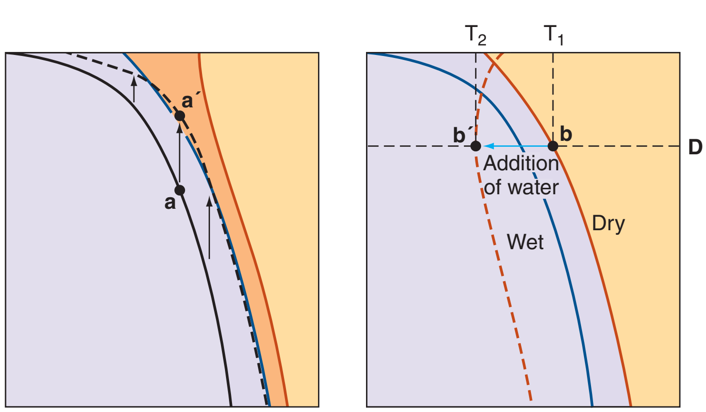


FIGURE 3.14

Mechanisms for melting rocks in the mantle. (*A*) Decompression melting. (*B*) Flux melting.

Addition of Water (Flux Melting)

If enough water is present and under high pressure, a dramatic change occurs in the melting process. Water sealed in under high pressure helps break the silicon-oxygen bonds in minerals, causing the crystals to liquify. A rock's melting temperature is significantly lowered by water under high pressure. This process is known as **flux melting.** Figure 3.14*B* shows the effect of water on the melting point of rocks in the mantle. The "dry" curve shows the temperature needed to melt rock that contains no water. The "wet" curve shows the temperature needed to melt rock that contains water. Consider "dry" mantle rock at point *b*. At this depth, the mantle needs to be above temperature T_1 in order to melt. Point *b* lies to the right of the geothermal gradient, so the temperature in the mantle is not high enough for melting to occur. Addition of water to the mantle moves the melting curve to the left. Point *b′* represents the new melting point of the mantle (T_2), which lies to the left of the geothermal gradient. "Wet" mantle at this depth will therefore undergo melting. As you will learn at the end of this chapter, flux melting is an important process at convergent plate boundaries, where subduction carries water down into the mantle.

HOW MAGMAS OF DIFFERENT COMPOSITIONS EVOLVE

A major topic of investigation for geologists is why igneous rocks are so varied in composition. On a global scale, magma composition is clearly controlled by geologic setting. But why? Why are basaltic magmas associated with oceanic crust, whereas granitic magmas are common in the continental crust? On a local scale, igneous bodies often show considerable variation in rock type. For instance, individual plutons typically display a considerable range of compositions, mostly varieties of granite, but many also will contain minor amounts of gabbro or diorite. In this section, we describe processes that result in differences in composition of magmas. The final section of this chapter relates these processes to plate tectonics for the larger view of igneous activity.

Sequence of Crystallization and Melting

Early in the twentieth century, N. L. Bowen conducted a series of experiments that determined the sequence in which minerals crystallize in a cooling magma. Bowen's experiments involved melting powdered rock, cooling the melt to a given temperature, and then observing the mineral present in the cooled rock. By repeating this process to different temperatures, he was able to observe that as you cool a basaltic melt, minerals tend to crystallize in a sequence determined by their melting temperatures. This sequence became known as **Bowen's reaction series** and is shown in figure 3.15. A simplified explanation of the series and its importance to igneous rocks is presented next.

Bowen's experiments showed that in a cooling magma, certain minerals are stable at higher melting temperatures and

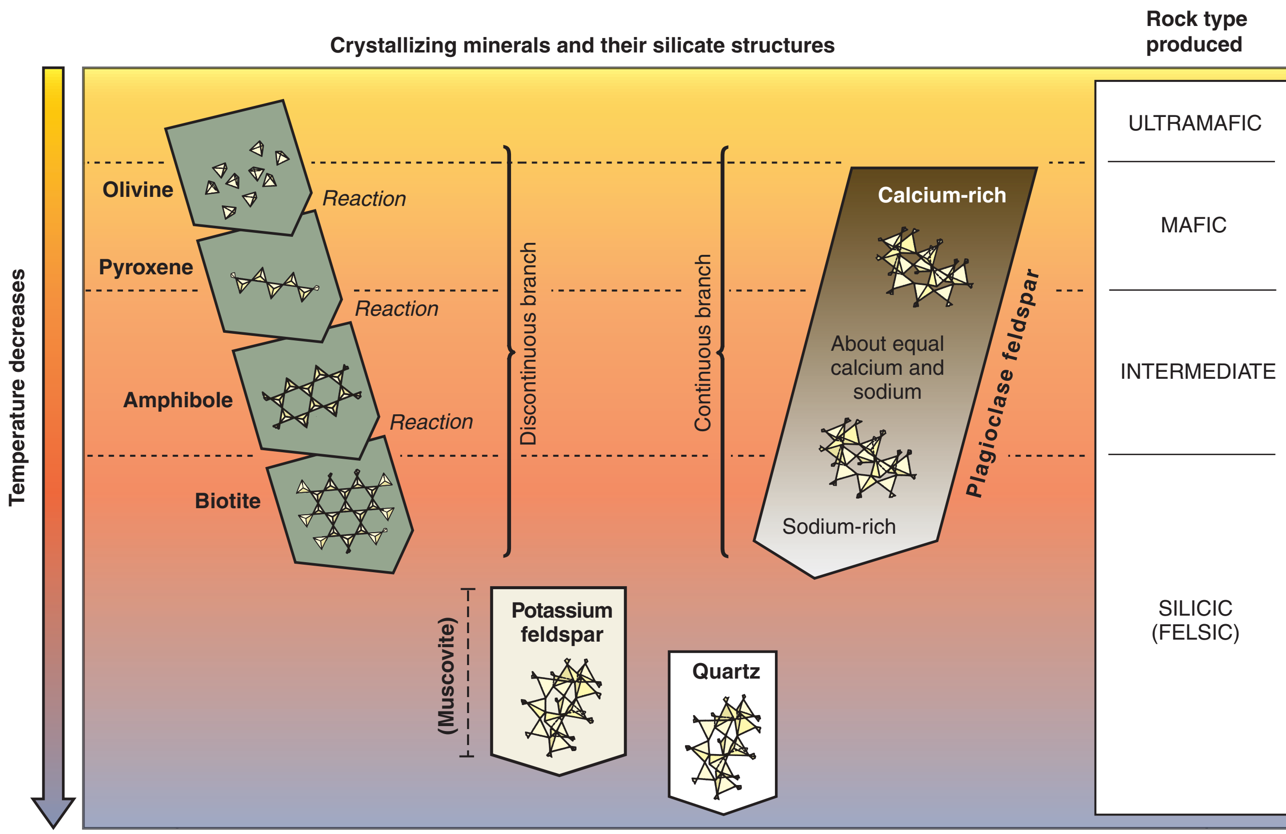


FIGURE 3.15

Bowen's reaction series. The reaction series as shown is very generalized. Moreover, it represents Bowen's experiments that involved melting a relatively silica-rich variety of basalt.

crystallize before those that are stable at lower temperatures. At higher temperatures, the sequence is broken into two branches. The *discontinuous branch* on the left-hand side of the diagram describes the formation of the ferromagnesian minerals olivine, pyroxene, amphibole, and biotite. This series is called the discontinuous series because these minerals have very different crystalline structures. The *continuous branch* on the right-hand side of the diagram contains only plagioclase feldspar. Plagioclase is a *solid solution* mineral (discussed in chapter 2 on minerals) in which either sodium or calcium atoms can be accommodated in its crystal structure, along with aluminum, silicon, and oxygen. The continuous branch describes the evolution of plagioclase from more calcium-rich compositions to more sodium-rich compositions with decreasing temperature.

Imagine a mafic magma as it cools and crystallizes. In figure 3.15, we can see that olivine crystallizes first, followed by calcium-rich plagioclase feldspar. Further cooling leads to the crystallization of pyroxene, then amphibole. The composition of plagioclase changes as magma is cooled and earlier-formed crystals react with the melt. The first plagioclase crystals to form as a hot melt cools contain calcium but little or no sodium. As cooling continues, the early-formed crystals grow and incorporate progressively more sodium into their crystal structures. As we progress through the reaction series, the formation of minerals rich in magnesium, iron, and calcium causes the remaining magma to become depleted in those elements and, at the same time, enriched in silicon, aluminum, and potassium. Any magma left at the point where the two branches meet is richer in silicon than the original magma; it also contains abundant potassium and aluminum. The potassium and aluminum combine with silicon to form *potassium feldspar.* (If the water pressure is high, *muscovite* may also form at this stage.) Excess SiO_2 crystallizes as *quartz.*

A complication is that early-formed crystals *react* with the remaining melt and recrystallize as cooling proceeds. For instance, early-formed olivine crystals react with the melt and recrystallize to pyroxene when pyroxene's temperature of crystallization is reached. Upon further cooling, pyroxene continues to crystallize until all of the melt is used up or the melting temperature of amphibole is reached. At this point, pyroxene reacts with the remaining melt, and amphibole forms at its expense. If all of the iron and magnesium in the melt is used up before all of the pyroxene recrystallizes to amphibole, then the ferromagnesian minerals in the solid rock will be amphibole and pyroxene. (The rock will not contain olivine or biotite.) If, however, minerals are separated from a magma, the remaining magma is more felsic than the original magma. For example, if olivine and calcium-rich plagioclase are removed, the residual melt will be richer in

ENVIRONMENTAL GEOLOGY 3.2

Harnessing Magmatic Energy

Buried magma chambers indirectly contribute the heat for today's geothermal electric generating plants. As explained in chapter 11 (Groundwater), water becomes heated in hot rocks. The heat source is usually presumed to be an underlying magma chamber. The rocks containing the hot water are penetrated by drilling. Steam exiting the hole is used to generate electricity.

Why not drill into and tap magma itself for energy? The amount of energy stored in a body of magma is enormous. The U.S. Geological Survey estimates that magma chambers in the United States within 10 kilometers of Earth's surface contain about 5,000 times as much energy as the country consumes each year. Our energy problems could largely be solved if significant amounts of this energy were harnessed.

There are some formidable technical difficulties in drilling into a magma chamber and converting the heat into useful energy. Despite these difficulties, the United States has considered developing magmatic energy. Experimental drilling has been carried out in Hawaii through the basalt crust of a lava lake that formed in 1960.

As drill bits approach a magma chamber, they must penetrate increasingly hot rock. The drill bit must be made of special alloys to prevent it from becoming too soft to cut rock. The rock immediately adjacent to a basaltic magma chamber is around 1,000°C, even though that rock is solid. Drilling into the magma would require a special technique. One that was experimented with is a jet-augmented drill. As the drill enters the magma chamber, it simultaneously cools and solidifies the magma in front of the drill bit. Thus, the drill bit creates a column of rock that extends downward into the magma chamber and simultaneously bores a hole down the center of this column. Once the hollow column is deep enough within the magma chamber, a boiler is placed in the hole. The boiler is protected from the magma by the jacket of the column of rock. Water would be pumped down the hole and turned to water vapor in the boiler by heat from the magma. Steam emerging from the hole would be used to generate electricity.

In principle, the idea is fairly simple, but there are serious technical problems. For one thing, high pressures would have to be maintained on the drill bit during drilling and while the boiler system was being installed; otherwise, gases within the magma might blast the magma out of the drill hole and create a human-made volcano. (The closest thing to a human-made volcano occurred in Iceland when a small amount of magma broke into a geothermal steam well and erupted briefly at the well head, showering the area with a few tons of volcanic debris.)

silicon and sodium and poorer in iron and magnesium. This process, known as *differentiation,* which can produce magmas of different composition from an initially mafic magma, is discussed further in the following section.

Bowen's reaction series can also be used to consider the formation of magma by melting rock. If you heat a rock, the minerals will melt in reverse order. In other words, you would be going up the series as diagrammed in figure 3.15. Any quartz and potassium feldspar in the rock would melt first. If the temperature were raised further, biotite and sodium-rich plagioclase would contribute to the melt. Any minerals higher in the series would remain solid unless the temperature were raised further.

Differentiation

The process by which different ingredients separate from an originally homogeneous mixture is **differentiation.** An example is the separation of whole milk into cream and nonfat milk. Differentiation in magmas takes place through a number of processes. As previously described, crystals that remain in contact with the melt as it cools will react to form minerals lower in Bowen's reaction series. If, however, crystals are no longer in contact with the melt, they will not react with it, and the elements they contain are effectively removed from the melt. For example, if olivine, pyroxene, and calcium-rich plagioclase crystallize from a mafic magma, they take up magnesium, iron, and calcium from the melt. If these early-formed crystals are removed from the melt, the magma becomes depleted in those elements and enriched in silicon, aluminum, sodium, and potassium. In other words, it becomes more felsic in composition. This differentiation process, often called *fractional crystallization,* can occur in a number of ways. One example is **crystal settling,** the downward movement of minerals that are denser (heavier) than the magma from which they crystallized.

If crystal settling takes place in a mafic magma chamber, olivine and, perhaps, pyroxene crystallize and settle to the bottom of the magma chamber (figure 3.16). This makes the remaining magma more felsic. Calcium-rich plagioclase also separates as it forms. The remaining magma is, therefore, depleted of calcium, iron, and magnesium. Because these minerals were economical in using the relatively abundant silica, the remaining magma becomes richer in silica as well as in sodium and potassium.

It is possible that by removing enough mafic components, the residual magma would be felsic enough to solidify into granite (or rhyolite). But it is more likely only enough mafic components would be removed to allow an intermediate residual magma, which would solidify into diorite or andesite. The lowermost portions of some intrusions are composed

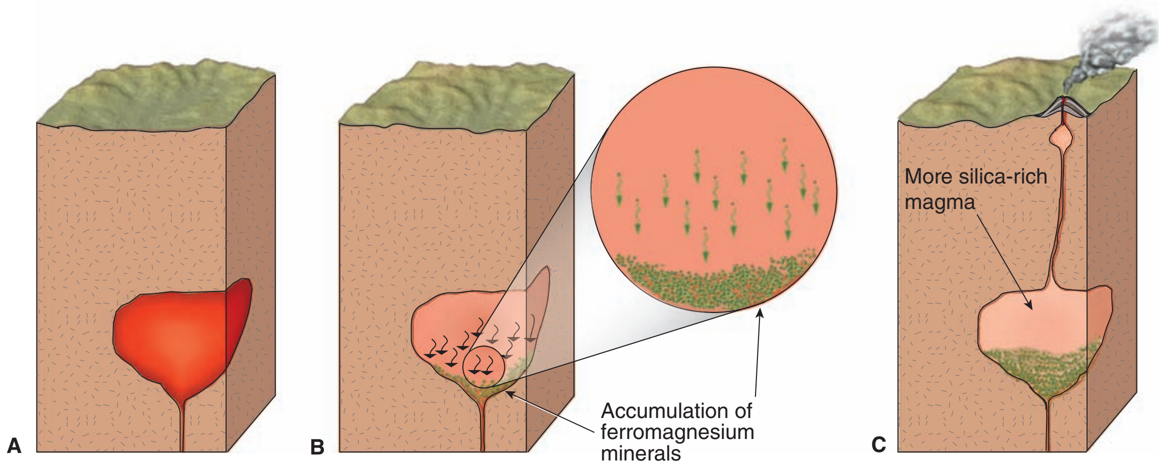


FIGURE 3.16
Differentiation of a magma body. (*A*) Recently intruded mafic magma is completely liquid. (*B*) Upon slow cooling, ferromagnesian minerals, such as olivine, crystallize and sink to the bottom of the magma chamber. The remaining liquid is now an intermediate magma. (*C*) Some of the intermediate magma moves upward to form a smaller magma chamber at a higher level that feeds a volcano.

predominantly of olivine and pyroxene, whereas upper levels are considerably less mafic. Even in large intrusions, however, differentiation has rarely progressed far enough to produce granite within the sill.

Partial Melting

As mentioned earlier, progressing upward through Bowen's reaction series (going from cool to hot) gives us the sequence in which minerals in a rock melt. As might be expected, the first portion of a rock to melt as temperatures rise forms a liquid with the chemical composition of quartz and potassium feldspar. The oxides of silicon plus potassium and aluminum "sweated out" of the solid rock could accumulate into a pocket of felsic magma. If higher temperatures prevailed, more mafic magmas would be created. Small pockets of magma could merge and form a large enough mass to rise buoyantly through the lower crust and toward the surface. In nature, temperatures rarely rise high enough to melt a rock entirely.

Partial melting of the lower continental crust likely produces felsic magma. The magma rises and eventually solidifies at a higher level in the crust into granite, or rhyolite if it reaches Earth's surface.

Geologists generally regard basaltic magma (Hawaiian lava, for example) as the product of partial melting of ultramafic rock in the mantle, at temperatures hotter than those in the crust. The solid residue left behind in the mantle when the basaltic magma is removed is an even more silica-deficient ultramafic rock.

Assimilation

A very hot magma may melt some of the country rock and *assimilate* the newly molten material into the magma (figure 3.17). This is like putting a few ice cubes into a cup of hot coffee. The ice melts and the coffee cools as it becomes diluted. Similarly, if a hot basaltic magma, perhaps generated from the mantle, melts portions of the continental crust, the magma simultaneously becomes richer in silica and cooler. Possibly intermediate magmas, such as those associated with circum-Pacific andesite volcanoes, may derive from assimilation of some crustal rocks by a basaltic magma.

Magma Mixing

Some of our igneous rocks may be "cocktails" of different magmas. The concept is quite simple. If two magmas meet and merge within the crust, the combined magma should be compositionally intermediate (figure 3.18). If you had approximately equal amounts of a granitic magma mixing with a basaltic magma, one would think that the resulting magma would crystallize underground as diorite or erupt on the surface to solidify as andesite. But you are unlikely to get a homogeneous magma or rock. This is because of the profound differences in the properties of felsic and mafic magmas, most notably their respective temperature differences. The mafic magma likely has a temperature of over 1,100°C, whereas a felsic magma would likely be several hundred degrees cooler. The mafic magma would be quickly cooled, and most of it would solidify when the two magmas meet. Some of the mafic minerals would react with the felsic magma and be absorbed in it, but most of the mafic magma would become blobs of basalt or gabbro included in the more felsic magma. Overall, the pluton would have an average chemical composition that is intermediate, but the rock that forms would not be a homogeneous intermediate rock. Because of this, *magma mingling* might be a better term for the process.

INTRUSIVE BODIES

Intrusions, or **intrusive structures,** are bodies of intrusive rock whose names are based on their size and shape as well as their relationship to surrounding rocks. They are important aspects of the architecture, or *structure,* of the Earth's crust. The various intrusions are named and classified on the basis of the following considerations: (1) Is the body large or small? (2) Does it have a particular geometric shape? (3) Did the rock form at a considerable depth, or was it a shallow intrusion? (4) Does it follow layering in the country rock?

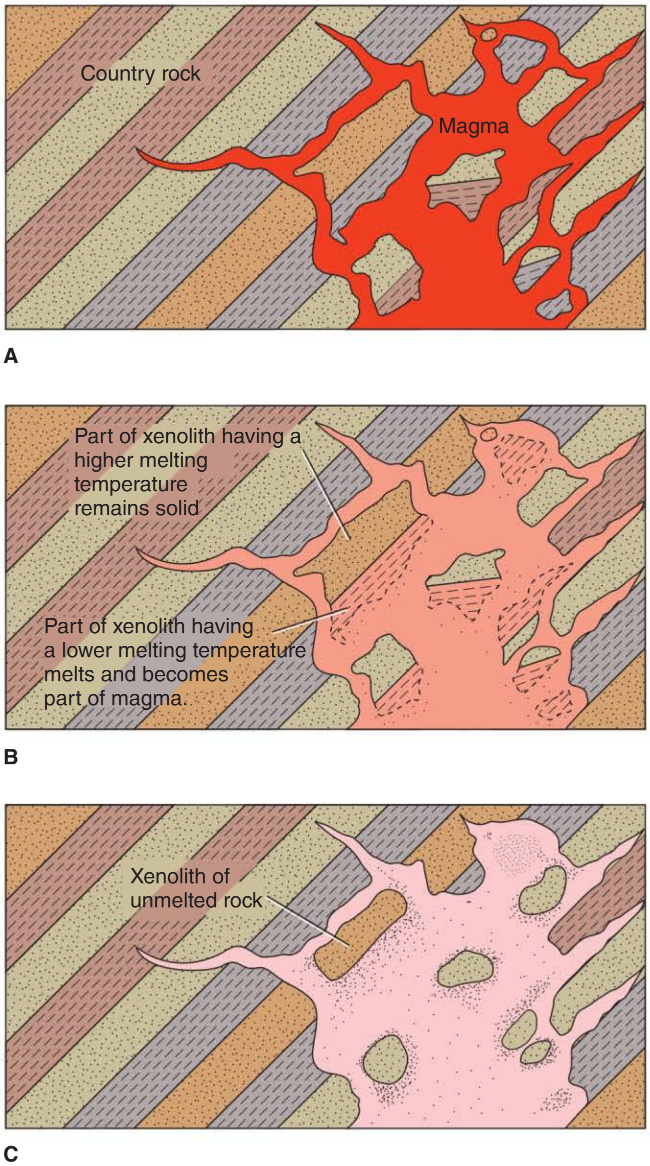


FIGURE 3.17

Assimilation. Magma formed is intermediate in composition between the original magma and the absorbed country rock. (*A*) Ascending magma breaks off blocks of country rock (the process called *stoping*). (*B*) Xenoliths of country rock with melting temperatures lower than the magma melt. (*C*) The molten country rock blends with the original magma, leaving unmelted portions as inclusions.

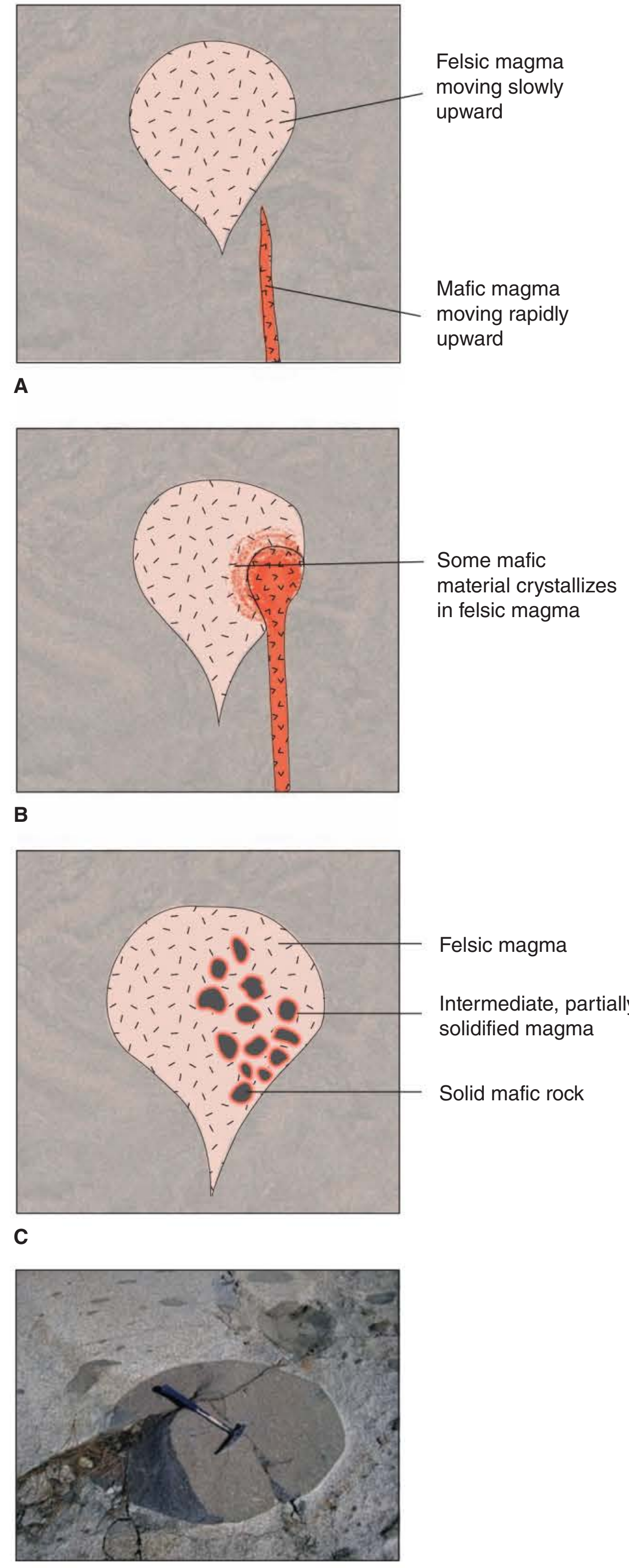


FIGURE 3.18

Mixing of magmas. (*A*) Two bodies of magma moving surfaceward. (*B*) The mafic magma catches up with the felsic magma. (*C*) An inhomogeneous mixture of felsic, intermediate, and mafic material. (*D*) "Blobs" of mafic magma in granite at Loch Leven Lakes, CA. *Photo by Jonathan Meurer.*

Shallow Intrusive Structures

Some igneous bodies apparently solidified near the surface of the Earth (probably at depths of less than 2 kilometers). These bodies appear to have solidified in the subsurface "plumbing systems" of volcanoes or lava flows. Shallow intrusive structures tend to be relatively small compared with those that formed at considerable depth. Because the country rock near the Earth's surface generally is cool, intruded magma tends to chill and solidify relatively rapidly. Also, smaller magma bodies will cool faster than larger bodies, regardless of depth. For both of these reasons, shallow intrusive bodies are likely to be fine-grained.

A **volcanic neck** is an intrusive structure apparently formed from magma that solidified within the throat of a volcano. One of the best examples is Ship Rock in New Mexico (figure 3.19). Here is how geologists interpret the history of this feature. A volcano formed above what is now Ship Rock. The magma for the volcano moved upward through a more or less cylindrical conduit. Eruptions ceased and the magma underground solidified into what is now Ship Rock. In time, the volcano and its underlying rock—the country rock around Ship Rock—eroded away. The more resistant igneous body eroded more slowly into its present shape. Weathering and erosion are continuing. (Falling rock has been a serious hazard to rock climbers.)

Dikes and Sills

Another, and far more common, intrusive structure can also be seen at Ship Rock. The low, wall-like ridge extending outward from Ship Rock is an eroded dike. A **dike** is a tabular (shaped like a tabletop), discordant, intrusive structure (figure 3.20). *Discordant* means that the body is not parallel to any layering in the country rock. (Think of a dike as cutting across layers of country rock.) Dikes may form at shallow depths and be

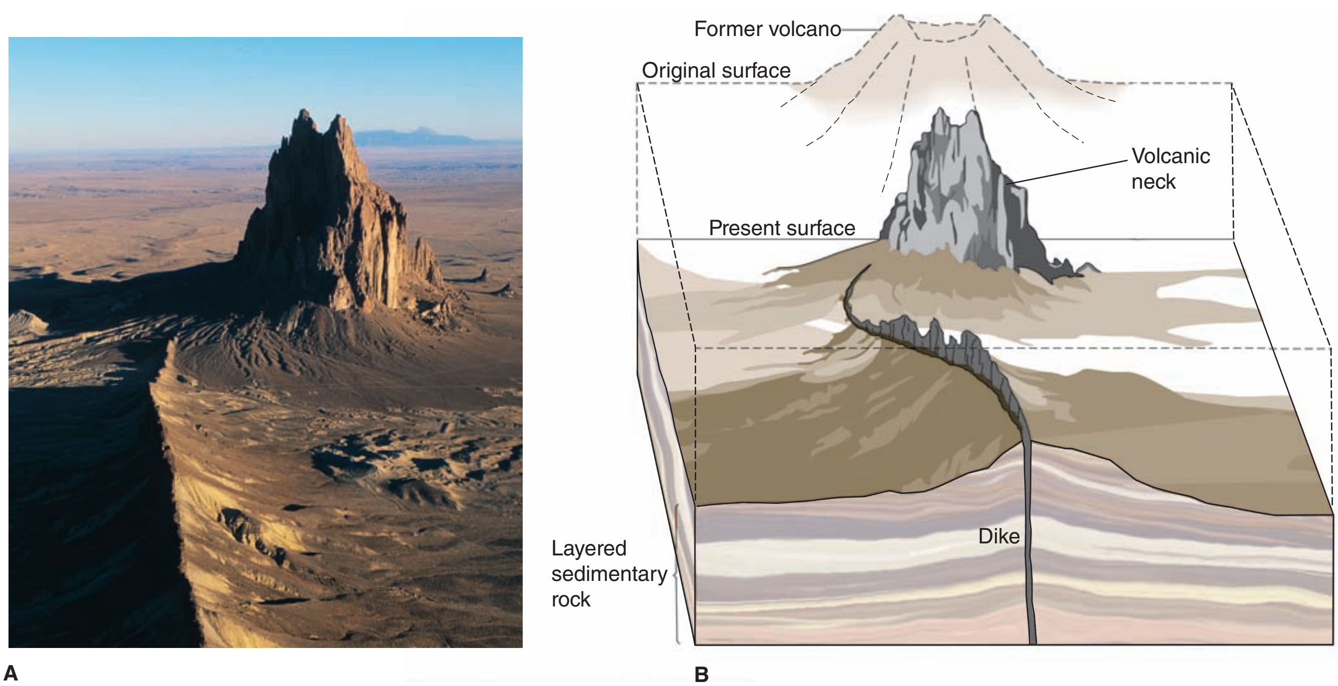


FIGURE 3.19

(*A*) Ship Rock in New Mexico, which rises 420 meters (1,400 feet) above the desert floor. *Photo © Bill Hatcher/National Geographic/Getty Images (B) Relationship to the former volcano.*

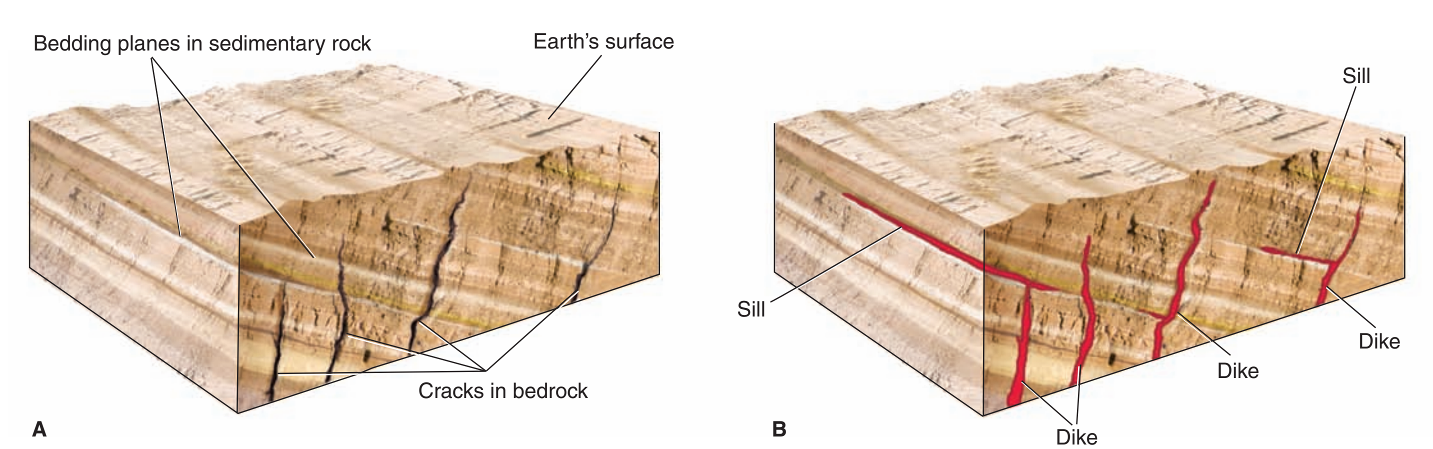


FIGURE 3.20

(*A*) Cracks and bedding planes are planes of weakness. (*B*) Concordant intrusions where magma has intruded between sedimentary layers are sills; discordant intrusions are dikes.

fine-grained, such as those at Ship Rock, or form at greater depths and be coarser-grained. Dikes need not appear as walls protruding from the ground (figure 3.21). The ones at Ship Rock do so only because they are more resistant to weathering and erosion than the country rock.

A **sill** is also a tabular intrusive structure, but it is *concordant.* That is, sills, unlike dikes, are parallel to any planes or layering in the country rock (figures 3.20 and 3.22). Typically, the country rock bounding a sill is layered sedimentary rock. As magma squeezes into a crack between two layers, it solidifies into a sill.

If the country rock is not layered, a tabular intrusion is regarded as a dike.

Intrusives that Crystallize at Depth

A **pluton** is a body of magma or igneous rock that crystallized at considerable depth within the crust. Where plutons are exposed at the Earth's surface, they are arbitrarily distinguished by size. If the area of surface exposure of plutonic rock is indicated on a map to be greater than 100 square kilometers, the body is called a **batholith** (figure 3.23). Most batholiths extend over areas vastly greater than the minimum 100 square kilometers. A smaller pluton that has a surface exposure of less than 100 square kilometers is called a **stock.** As can be seen in figure 3.23, however, a stock may be the exposed portion of a larger batholith.

Although batholiths may contain mafic and intermediate rocks, they almost always are predominantly composed of granite. Detailed studies of batholiths indicate that they are formed of numerous, coalesced plutons. Apparently, large blobs of magma worked their way upward through the lower crust and collected 5 to 30 kilometers below the surface, where they solidified (figure 3.24). These blobs of magma, known as **diapirs,** are less dense than the surrounding rock that is pliable and shouldered aside as the magma rises. When viscous magma intrudes between two layers of sedimentary rock, it may generate enough pressure to lift the overlying layer into an arch. The magma then cools to form a concordant, mushroom-shaped or domed pluton known as a *laccolith.*

Batholiths occupy large portions of North America, particularly in the west. Over half of California's Sierra Nevada (figure 3.25) is a batholith whose individual plutons were emplaced during a period of over 100 million years. An even larger batholith extends almost the entire length of the mountain ranges of Canada's west coast and southeastern Alaska—a distance of 1,800 kilometers. Smaller batholiths are also found in eastern North America in the Piedmont east of the Appalachian Mountains and in New England and the coastal provinces of Canada. (The extent and location of North American batholiths are shown on the geologic map on the inside back cover.) Laccoliths are common in the United States, often forming distinctive landforms. Pine Valley Mountain, near St. George, Utah, is one of the largest laccoliths in the United States.

Granite is considerably more common than rhyolite, its volcanic counterpart. Why is this? Silicic magma is much more *viscous* (that is, more resistant to flow) than mafic magma. Therefore, a silicic magma body will travel upward through the

FIGURE 3.21

Dikes (light-colored rocks) in northern Victoria Land, Antarctica. *Photo by C. C. Plummer*

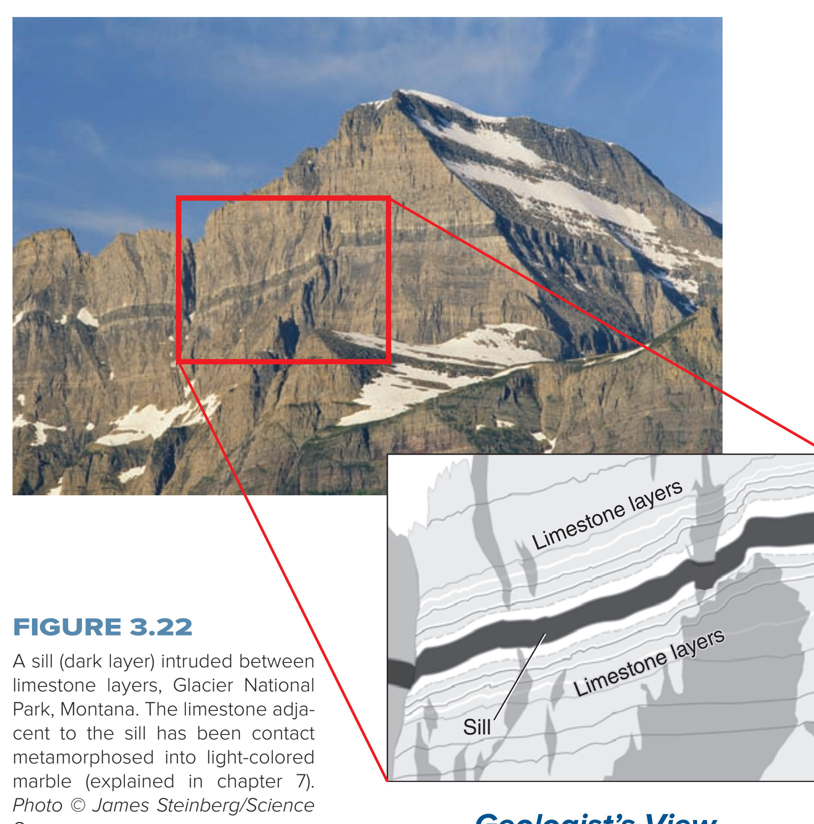


FIGURE 3.22

A sill (dark layer) intruded between limestone layers, Glacier National Park, Montana. The limestone adjacent to the sill has been contact metamorphosed into light-colored marble (explained in chapter 7). *Photo © James Steinberg/Science Source*

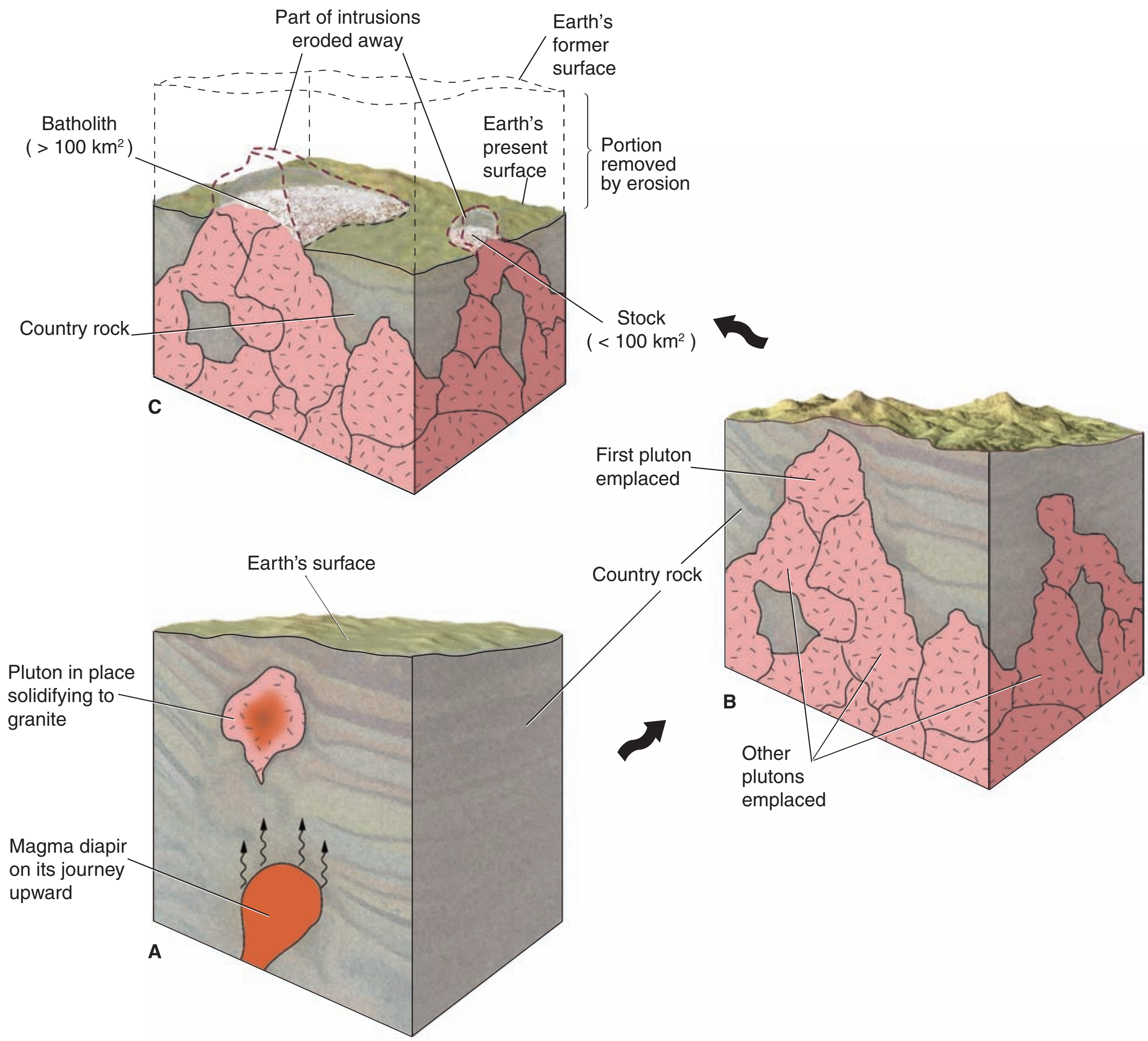


FIGURE 3.23

(*A*) The first of numerous magma diapirs has worked its way upward and is emplaced in the country rock. (*B*) Other magma diapirs have intruded, coalesced, and solidified into a solid mass of plutonic rock. (*C*) After uplift and erosion, surface exposures of plutonic rock are a batholith and a stock.

crust more slowly and with more difficulty than mafic or intermediate magma. Unless it is exceptionally hot, a silicic magma will not be able to work its way through the relatively cool and rigid rocks of the upper few kilometers of crust. Instead, it is much more likely to solidify slowly into a pluton.

ABUNDANCE AND DISTRIBUTION OF PLUTONIC ROCKS

Granite is the most abundant igneous rock in mountain ranges. It is also the most commonly found igneous rock in the interior lowlands of continents. Throughout the lowlands of much of Canada, very old plutons have intruded even older metamorphic rock. As explained in chapter 20 on mountains and the continental crust, very old mountain ranges have, over time, eroded and become the stable interior of a continent. Metamorphic and plutonic rocks similar in age and complexity to those in Canada are found in the Great Plains of the United States. Here, however, they are mostly covered by a veneer (a kilometer or so) of younger, sedimentary rock. These basement rocks are exposed to us in only a few places. In Grand Canyon, Arizona, the Colorado River has eroded through the layers of sedimentary rock to expose the ancient plutonic and metamorphic basement. In the Black Hills of South Dakota, local uplift and subsequent erosion have exposed similar rocks.

Granite, then, is the predominant igneous rock of the continents. As described in chapter 4, basalt and gabbro are the

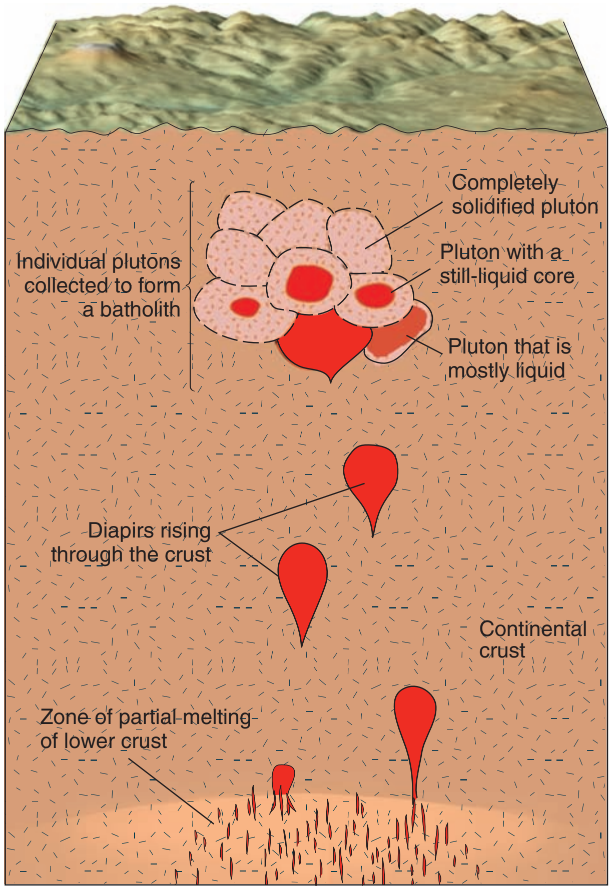


FIGURE 3.24

Diapirs of magma travel upward from the lower crust and solidify in the upper crust. (Not drawn to scale.)

predominant rocks underlying the oceans. Andesite (usually along continental margins) is the building material of most young volcanic mountains. Underneath the crust, ultramafic rocks make up the upper mantle.

EXPLAINING IGNEOUS ACTIVITY BY PLATE TECTONICS

One of the appealing aspects of the theory of plate tectonics is that it accounts reasonably well for the variety of igneous rocks and their distribution patterns. (Chapter 1 has an overview of plate tectonics.) Divergent boundaries are associated with the creation of basalt and gabbro of the oceanic crust. Andesite and granite are associated with convergent boundaries. Table 3.2 summarizes the relationships.

Igneous Processes at Divergent Boundaries

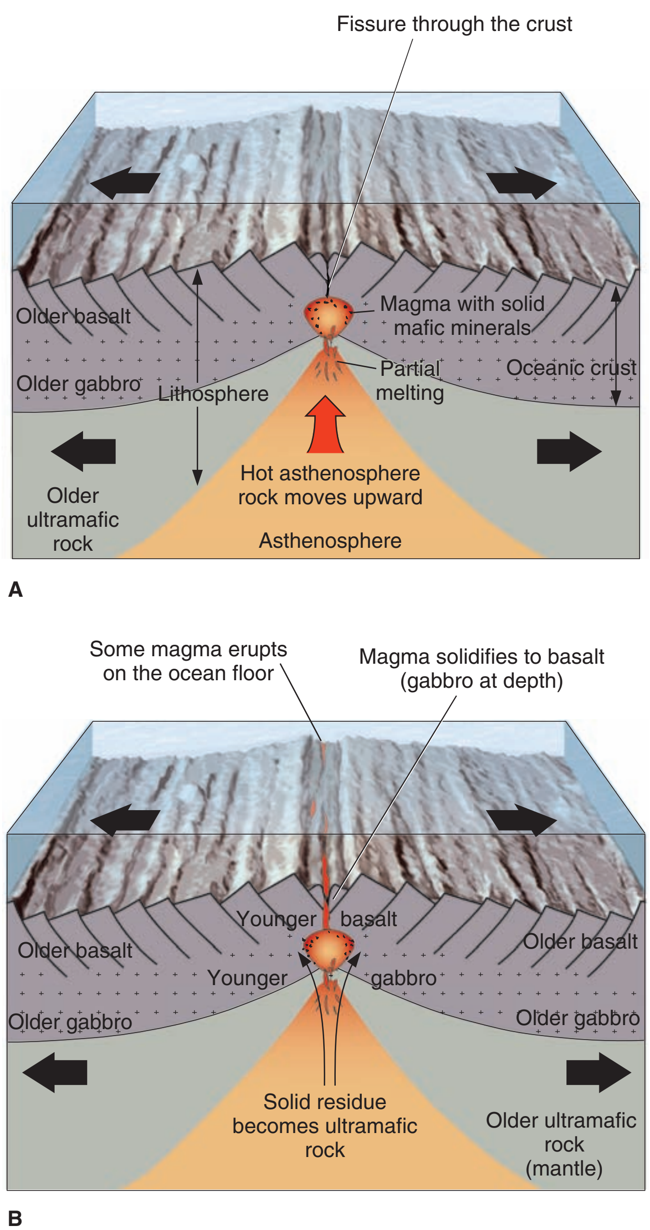
The crust beneath the world's oceans (over 70% of Earth's surface) is mafic volcanic and intrusive rock, covered to a varying extent by sediment and sedimentary rock. Most of this basalt and gabbro was created at mid-oceanic ridges, which also are divergent plate boundaries. Geologists agree that the mafic magma produced at divergent boundaries is due to partial melting of the asthenosphere. The *asthenosphere,* as described in chapter 1, is the plastic zone of the mantle beneath the rigid *lithosphere* (the upper mantle and crust that make up a plate). Along divergent boundaries, the asthenosphere is relatively close (5 to 10 kilometers) to the surface (figure 3.26).

FIGURE 3.25

Part of the Sierra Nevada batholith. All light-colored rock shown here (including that under the distant snow-covered mountains) is granite. The extent of the Sierra Nevada batholith is shown in the inset. *Photo by C. C. Plummer*

TABLE 3.2 Relationships between Rock Types and Their Usual Plate-Tectonic Setting

Rock	Original Magma	Final Magma	Processes	Plate-Tectonic Setting
Basalt and gabbro	Mafic	Mafic	Partial melting of mantle (asthenosphere)	1. Divergent boundary—oceanic crust created 2. Intraplate • plateau basalt • volcanic island chains (e.g., Hawaii)
Andesite and diorite	Mafic (usually)	Intermediate	Partial melting of mantle (asthenosphere) followed by: • differentiation or • assimilation or • magma mixing	Convergent boundary
Granite and rhyolite	Felsic	Felsic	Partial melting of lower crust	1. Convergent boundary 2. Intraplate • over mantle plume



The probable reason the asthenosphere is plastic or "soft" is that temperatures there are only slightly lower than the temperatures required for partial melting of mantle rock.

If extra heat is added, or pressure is reduced, partial melting should take place. The asthenosphere beneath divergent boundaries probably is mantle material that has welled upward from deeper levels of the mantle. As the hot asthenosphere gets close to the surface, decrease in pressure results in partial melting. In other words, *decompression melting* takes place. The magma that forms is mafic and will solidify as basalt or gabbro. The portion that did not melt remains behind as a silica-depleted, iron-and-magnesium-enriched ultramafic rock.

Some of the basaltic magma erupts along a submarine ridge to form pillow basalts (described in chapter 4), while some fills near-surface fissures to create dikes. Deeper down, magma solidifies more slowly into gabbro. The newly solidified rock is pulled apart by spreading plates; more magma fills the new fracture and some erupts on the sea floor. The process is repeated, resulting in a continuous production of mafic crust.

The basalt magma that builds the oceanic crust is removed from the underlying mantle, depleting the mantle beneath the ridge of much of its calcium, aluminum, and silicon oxides. The unmelted residue (olivine and pyroxene) becomes depleted mantle, but it is still a variety of ultramafic rock. The rigid ultramafic rock, the overlying gabbro and basalt, and any sediment that may have deposited on the basalt collectively are the lithosphere of an oceanic plate, which moves away from a spreading center over the asthenosphere. (The nature of the oceanic crust is described in more detail in chapter 18.)

FIGURE 3.26

Schematic representation of how basaltic oceanic crust and the underlying ultramafic mantle rock form at a divergent boundary. The process is more continuous than the two-step diagram implies. (*A*) Partial melting of asthenosphere takes place beneath a mid-oceanic ridge, and magma rises into a magma chamber. (*B*) The magma squeezes into the fissure system. Solid mafic minerals are left behind as ultramafic rock.

Intraplate Igneous Activity

Igneous activity within a plate, a long distance from a plate boundary, is unusual. These hot spots have been hypothesized to be due to hot **mantle plumes,** which are narrow upwellings of hot material within the mantle. Examples include the long-lasting volcanic activity that built the Hawaiian Islands and the eruptions at Yellowstone National Park in Wyoming. The ongoing eruptions in Hawaii take place on oceanic crust, whereas eruptions at Yellowstone represent continental intraplate activity. The silicic eruptions at Yellowstone that took place some 600,000 years ago were much larger and more violent than any eruptions that have occurred in historical time.

The huge volume of mafic magma that erupted to form the Columbia plateau basalts of Washington and Oregon (described in chapter 4) is attributed to a past hot mantle plume, according to a recent hypothesis (figure 3.27). In this case, the large volume of basalt is due to the arrival beneath the lithosphere and decompression melting of a mantle plume with a large head on it.

Igneous Processes at Convergent Boundaries

Intermediate and felsic magmas are clearly related to the convergence of two plates and subduction. However, exactly what takes place is debated by geologists. Compared to divergent

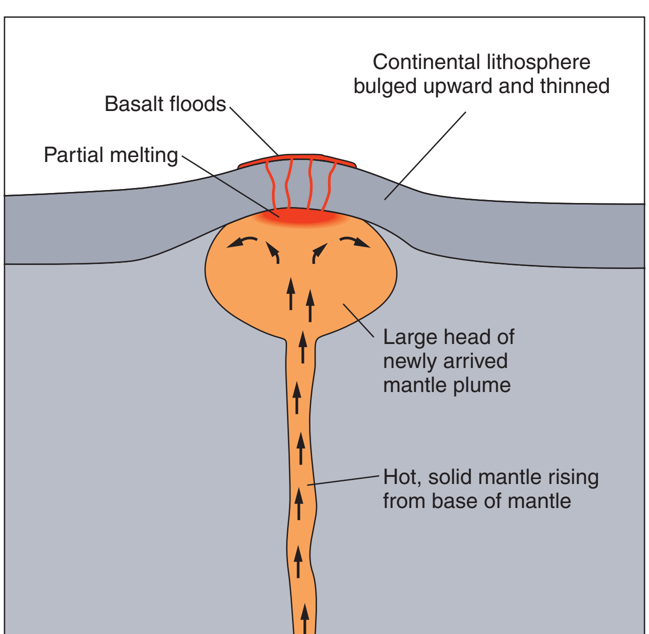


FIGURE 3.27

A hot mantle plume with a large head rises from the lower mantle. When it reaches the base of the lithosphere, it uplifts and stretches the overlying lithosphere. The reduced pressure results in decompression melting, producing basaltic magma. Large volumes of magma travel through fissures and flood the Earth's surface.

boundaries, there is less agreement about how magmas are generated at convergent boundaries. The scenarios that follow are currently regarded by geologists to be the best explanations of the data.

The Origin of Andesite

Magma for most of our andesitic composite volcanoes (such as those found along the west coast of the Americas) seems to originate from a depth of about 100 kilometers. This coincides with the depth at which the subducted oceanic plate is sliding under the asthenosphere (figure 3.28). Partial melting of the asthenosphere takes place, resulting in a mafic magma. In most cases, melting occurs because the subducted oceanic crust releases water into the asthenosphere. The water collected in the oceanic crust when it was beneath the ocean and is driven out as the descending plate is heated. The water lowers the melting temperature of the ultramafic rocks in this part of the mantle. In other words *flux melting* takes place. Partial melting produces a mafic magma.

But how can we keep producing magma from ultramafic rock after those rocks have been depleted of the constituents of the mafic magma? The answer is that hot asthenospheric rock continues to flow into the zone of partial melting. As shown in figure 3.28, asthenospheric ultramafic rock is dragged downward by the descending lithospheric slab. More ultramafic rock flows laterally to replace the descending material. A continuous flow of hot, fertile (containing the constituents of basalt) ultramafic rock is brought into the zone where water, moving upward from the descending slab, lowers the melting temperature. After being depleted of basaltic magma, the solid, residual, ultramafic rock continues to sink deeper into the mantle.

On its slow journey through the crust, the mafic magma evolves into an intermediate magma by differentiation and by assimilation of silicic crustal rocks.

Under special circumstances, basalt of the descending oceanic crust can partially melt to yield an intermediate magma. In most subduction zones, the basalt remains too cool to melt, even at a depth of over 100 kilometers. But geologists believe that partial melting of the subducted crust produces the magma for andesitic volcanoes in South America. Here, the oceanic crust is much younger and considerably hotter than normal. The spreading axis where it was created is not far from the trench. Because the lithosphere has not traveled far before being subducted, it is still relatively hot.

The reason that partial melting of subducted basalt is unusual is that this kind of subduction and magma generation is, geologically speaking, short-lived. Subduction will end when the overriding plate crashes into the mid-oceanic ridge. Most subduction zones are a long distance from the divergent boundaries of their plates, so steep subduction and magma production from the asthenosphere are the norm.

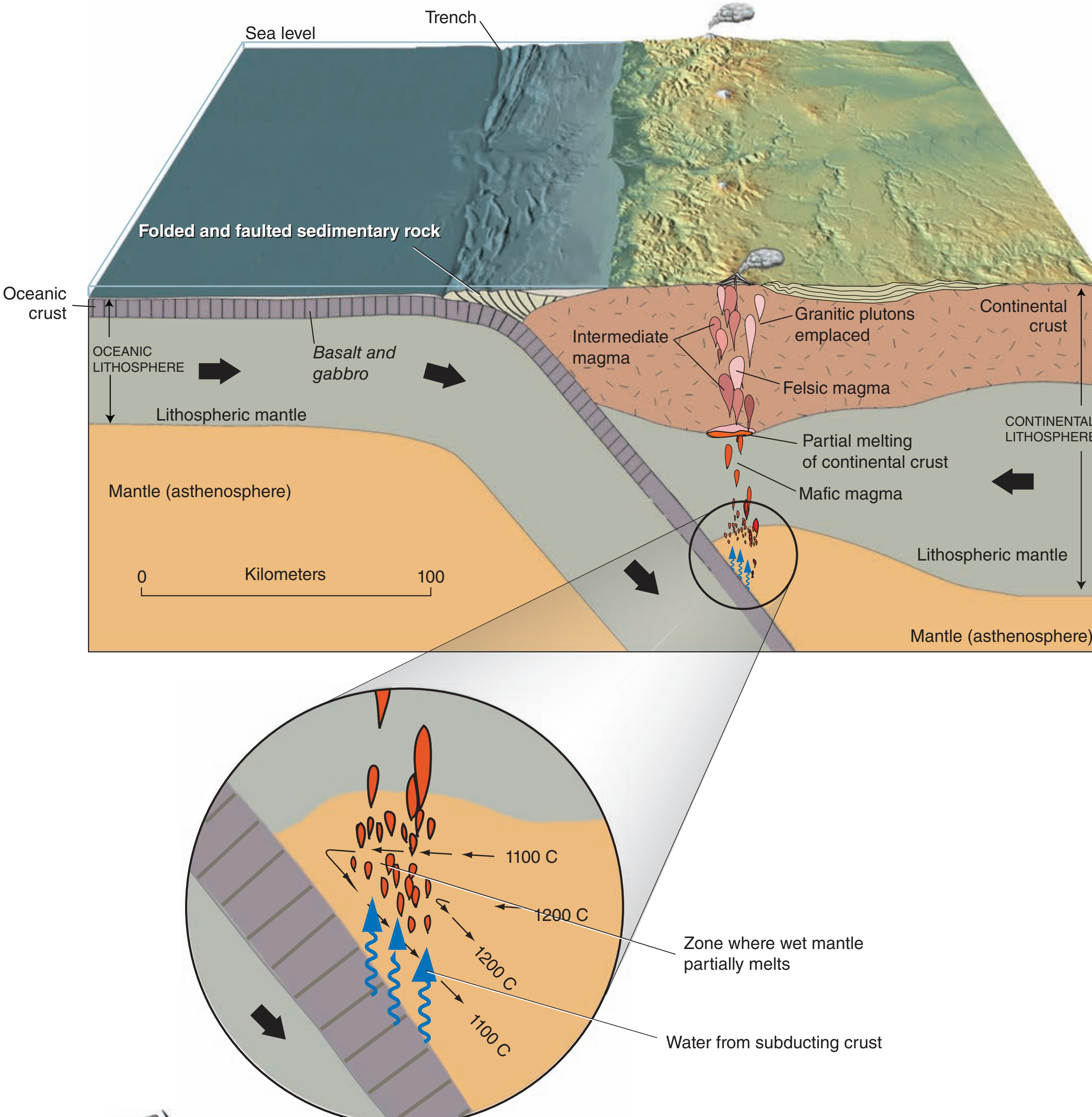


FIGURE 3.28

Generation of magma at a convergent boundary. Mafic magma is generated in the asthenosphere above the subducting oceanic lithosphere, and felsic magma is created in the lower crust. The inset shows the circulation of asthenosphere and lines of equal temperature (isotherms). Partial melting of wet ultramafic rock takes place in the zone where it is between 1100 and 1200°C.

The Origin of Granite

To explain the great volumes of granitic plutonic rocks, many geologists think that partial melting of the lower continental crust must take place. The continental crust contains the high amount of silica needed for a silicic magma. As the silicic rocks of the continental crust have relatively low melting temperatures (especially if water is present), partial melting of the lower continental crust is likely. However, calculations indicate that the temperatures we would expect from a normal geothermal gradient are too low for melting to take place. Therefore, we need an additional heat source.

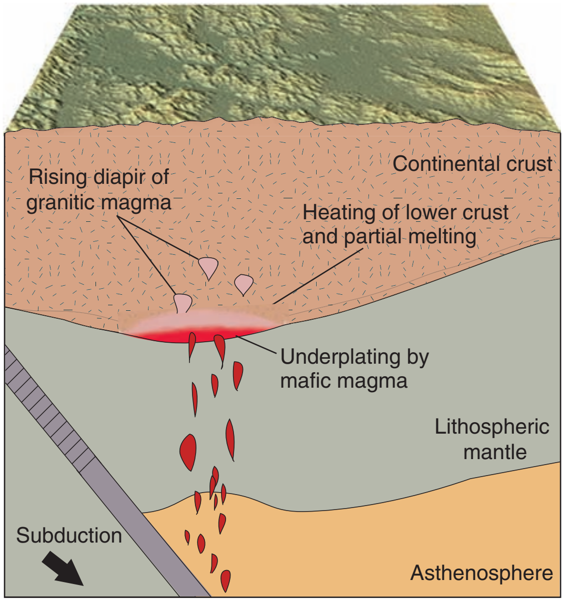


FIGURE 3.29

How mafic magma could add heat to the lower crust and result in partial melting to form a granitic magma. Mafic magma from the asthenosphere rises to underplate the continental crust.

Currently, geologists think that the additional heat is provided by mafic magma that was generated in the asthenosphere and moved upward. The process of *magmatic underplating* involves mafic magma pooling at the base of the continental crust, supplying the extra heat necessary to partially melt the overlying, silica-rich crustal rocks (figure 3.29). Mafic magma generated in the asthenosphere rises to the base of the crust. The mafic magma is denser than the overlying silica-rich crust; therefore, it collects as a liquid mass that is much hotter than the crust. The continental crust becomes heated (as if by a giant hotplate). When the temperature of the lower crust rises sufficiently, partial melting takes place, creating felsic magma. The felsic magma collects and forms diapirs, which rise to a higher level in the crust and solidify as granitic plutons (or, on occasion, reach the surface and erupt violently).

Summary

The interaction between the internal and external forces of the Earth is illustrated by the rock cycle, a conceptual device relating igneous, sedimentary, and metamorphic rocks to each other, to surficial processes such as weathering and erosion, and to internal processes such as tectonic forces. Changes take place when one or more processes force Earth's material out of equilibrium.

Igneous rocks form from solidification of magma. If the rock forms at the Earth's surface, it is *extrusive. Intrusive rocks* are igneous rocks that formed underground. Igneous rocks are named based upon their texture and composition. Intrusive rocks have a *phaneritic* (coarse-grained) texture, and extrusive rocks have an *aphanitic* (fine-grained) texture. Felsic (or silicic) rocks are rich in silica, whereas mafic rocks are silica-poor. The mineral content of igneous rocks reflects the chemical composition of the magmas from which they formed. *Granite, diorite,* and *gabbro* are the coarse-grained equivalents of *rhyolite, andesite,* and *basalt,* respectively. *Peridotite* is an *ultramafic* rock made entirely of ferromagnesian minerals and is mostly associated with the mantle.

Basalt and gabbro are predominant in the oceanic crust. Granite predominates in the continental crust. Younger granite batholiths occur mostly within younger mountain belts. Andesite is largely restricted to narrow zones along convergent plate boundaries.

The *geothermal gradient* is the increase in temperature with increase in depth. Melting of the mantle occurs through *decompression melting* and *flux melting. Partial melting* of the mantle usually produces basaltic magma, whereas granitic magma is most likely produced by partial melting of the lower continental crust.

No single process can satisfactorily account for all igneous rocks. In the process of *differentiation,* based on *Bowen's reaction series,* a residual magma more silicic than the original mafic magma is created when the early-forming minerals separate out of the magma. In *assimilation,* a hot, original magma is contaminated by picking up and absorbing rock of a different composition. Rocks of intermediate composition can form when *magma mixing* occurs between a felsic magma and a mafic magma.

Some intrusive rocks have solidified near the surface as a direct result of volcanic activity. Volcanic *necks* solidified within volcanoes. Fine-grained *dikes* and *sills* may also have formed in cracks during local extrusive activity. A sill is *concordant*—parallel to the planes within the country rock. A dike is *discordant*—not parallel to planes in the country rock. Both are tabular bodies. Coarser grains in either a dike or a sill indicate that it probably formed at considerable depth. Most intrusive rock is *plutonic*—that is, coarse-grained rock that solidified slowly at considerable depth. Most plutonic rock exposed at the Earth's surface is in *batholiths*—large plutonic bodies. A smaller body is called a *stock.*

The theory of *plate tectonics* incorporates the preceding concepts. Basalt is generated where hot mantle rock partially melts, most notably along divergent boundaries. The fluid magma rises easily through fissures, if present. The ferromagnesian portion that stays solid remains in the mantle as ultramafic rock. Granite and andesite are associated with subduction. Differentiation, assimilation, and partial melting may each play a part in creating the observed variety of rocks.