Chapter 7. Igneous Rocks

Introduction

Learning Objectives

After carefully reading this chapter, completing the exercises within it, and answering the questions at the end, you should be able to:

• Explain partial melting and the geological processes that lead to melting
• Describe, in general terms, the range of chemical compositions of magmas
• Discuss the processes that take place during the cooling and crystallization of magma, and the typical order of crystallization according to the Bowen reaction series
• Explain how magma composition can be changed by fractional crystallization and partial melting of the surrounding rocks
• Apply the criteria for igneous rock classification based on mineral proportions
• Describe the origins of phaneritic, porphyritic, and pegmatitic textures
• Identify plutons on the basis of their morphology and their relationships to the surrounding rocks
• Explain the origin of a chilled margin

7.1 Magma and Magma Formation

Magmas can vary widely in composition, but in general they are made up of only eight elements. In order of importance these are oxygen, silicon, aluminum, iron, calcium, sodium, magnesium, and potassium (Figure 7.1). Of those elements, oxygen comprises a little less than half, and silicon is just over one-quarter.

The composition of magma depends on the rock it was formed from (by melting), and the conditions of that melting. Magmas derived from the mantle have higher levels of iron, magnesium, and calcium, but they are still likely to be dominated by oxygen and silicon. All magmas have varying proportions of elements such as hydrogen, carbon, and sulphur, which are converted into gases like water vapour, carbon dioxide, and hydrogen sulphide as the magma cools.

Virtually all of the igneous rocks that we see on Earth are derived from magmas that formed from partial melting of existing rock, either in the upper mantle or the crust. Partial melting is what happens when only some of the minerals within a rock melt. It takes place because different minerals have different melting temperatures.

To understand how this works, consider the mix of materials in Figure 7.2a. It contains blocks of candle wax (white), black plastic pipe, green beach glass, and pieces of aluminum wire. If you put the mixture into a warm
oven at 50 °C, the wax would begin to melt into a clear liquid (Figure 7.2b), but the other materials in the mix would stay solid. That’s partial melting. Next, imagine that the oven is heated up to 120 °C. The plastic would melt too and mix with the liquid wax, but the aluminum and glass would remain solid (Figure 7.2c). This is also partial melting. If at the end of the experiment the plastic and wax “magma” mixture were poured into a separate container and let cool, the result would be a solid with a very different composition from the original mixture (Figure 7.2d).

Of course partial melting in the real world isn’t as simple as in the example in Figure 7.2. One difference is that rocks are much more complex than the four-component system used here. Another difference is that the mineral components of most rocks have similar melting temperatures, so two or more minerals are likely to melt at the same time to varying degrees. Yet another important difference is that when rocks melt, the process takes thousands to millions of years, not the 90 minutes it took in the example.

**Adding Heat Does Not Melt Rocks, But Dropping Pressure and Adding Water Can**

Contrary to what one might expect, most partial melting does not involve adding heat to rock. The two main mechanisms through which rocks melt are decompression melting and flux melting.

**Decompression melting** takes place when a body of rock within the Earth is held at approximately the same temperature but the pressure is reduced. This can happen if the rock is being moved toward the surface, either at
a mantle plume (a hot spot), or in the upwelling part of a mantle convection cell.\textsuperscript{1} The mechanism of decompression melting is shown in Figure 7.3a. If a rock that is hot enough to be close to its melting point is moved toward the surface, the pressure is reduced, and the rock can pass to the liquid side of its melting curve. At this point, partial melting begins.

Flux melting occurs when a substance such as water is added to the system, causing the melting point of a rock to decrease. If a rock is already close to its melting point, the effect of adding water can be enough to trigger partial melting. In Figure 7.3b this is illustrated by the melting curve for dry rock (dotted line) shifting to the left.

Partial melting of rock happens in a wide range of situations, most of which are related to plate tectonics (Figure 7.4). Decompression melting happens at both mantle plumes and in the upward parts of convection systems. Rock is being moved toward the surface in these locations, so the pressure is dropping. At some point the rock crosses to the liquid side of its melting curve. At subduction zones, pressure and heat cause minerals within subducting ocean crust to release water into the overlying hot mantle. This provides the flux needed to lower the melting temperature.

In both of these cases, only partial melting takes place, typically melting only about 10% of the rock. It is always the most silica-rich components of the rock that melt, creating a magma that is more silica-rich than the rock from which it is derived. By analogy, the melt from the experiment in Figure 7.2 is richer in wax and

\textsuperscript{1} Mantle plumes and mantle convection are described in Chapters 3 and 4.
plastic than the mixture from which it was derived. The magma that is produced is less dense than the surrounding rock. It moves up through the mantle and eventually into the crust.

**Cooling Magma Becomes More Viscous**

As the magma moves toward the surface it interacts with the surrounding rock. This can lead to partial melting of the surrounding rock because the magma is often hotter than the melting temperature of rocks in the crust. The more silica-rich parts of the surrounding rock are preferentially melted, and this contributes to an increase in the silica content of the magma.

At very high temperatures (over 1300°C), most magma is entirely liquid. In a liquid state there is too much energy for the atoms to bond together, but this begins to change as the magma cools. Silicon and oxygen combine to form silica tetrahedra. As cooling continues, the tetrahedra link together to make chains (they **polymerize**). The presence of silica chains has the important effect of making the magma more viscous (less runny), and as we’ll see in Chapter 11, magma viscosity has significant implications for volcanic eruptions. As the magma continues to cool, crystals start to form. The presence of crystals increases magma viscosity even further.

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**Exercise 7.1 Making Magma Viscous**

This is a quick and easy experiment that you can do at home to help you understand the properties of magma. It will only take about 15 minutes, and all you need is half a cup of water and a few tablespoons of flour.

If you’ve ever made gravy, white sauce, or roux, you’ll know how this works.

Place about 1/2 cup (125 mL) of water in a saucepan over medium heat. Add 2 teaspoons (10 mL) of white flour (this represents silica) and stir while the mixture comes close to boiling. It should thicken like gravy because the gluten in the flour becomes polymerized into chains during this process.

Now you’re going to add more “silica” to see how this changes the viscosity of your magma. Take another 4 teaspoons (20 mL) of flour and mix it thoroughly with about 4 teaspoons (20 mL) of water in a cup and then add all of that mixture to the rest of the water and flour in the saucepan. Stir while bringing it back up to nearly boiling temperature, and then allow it to cool. This mixture should slowly become much thicker because there is more gluten and more chains have been formed (see the photo).

This is analogous to magma, of course. As we’ll see below, magmas have variable contents of silica and therefore have widely varying viscosities (thicknesses) during cooling.
7.2 Crystallization of Magma

The minerals that make up igneous rocks crystallize (solidify, freeze) at a range of different temperatures. This explains why a cooling magma can have some crystals within it and yet remain predominantly liquid. The sequence in which minerals crystallize from a magma is known as the Bowen reaction series (Figure 7.5).

**Bowen’s Reaction Series**

![Bowen's Reaction Series Diagram](image)

**How Did We Get Bowen’s Reaction Series?**

Understanding how the reaction series was derived is key to understanding what it means.

Norman Levi Bowen (Figure 7.6) was born in Kingston Ontario. He studied geology at Queen’s University and then at MIT in Boston. In 1912 he joined the Carnegie Institution in Washington, D.C., where he carried out ground-breaking experiments into how magma cools.

Working mostly with mafic magmas (magmas rich in iron and magnesium), he determined the order of crystallization of minerals as the temperature drops. The method, in brief, was to melt the rock to a magma in a specially made kiln, allow it to cool slowly to a specific temperature (allowing some minerals to form), and then quench it (cool it quickly) so that no new minerals form (only glass). The results were studied under the microscope and by chemical analysis. This was done over and over, each time allowing the magma to cool to a lower temperature before quenching.

The result of these experiments was the reaction series which, even a century later, is still an important basis for our understanding of igneous rocks.
Discontinuous and Continuous Series

Different Minerals Form in Sequence on the Discontinuous Side

Bowen’s reaction series has two pathways for minerals to form as magma cools. On the left of Figure 7.5 is the discontinuous series. This refers to the fact that one kind of mineral is being transformed into a different mineral through chemical reactions. For example, olivine begins to form at just below 1300°C, but as the temperature drops, olivine becomes unstable. The early-forming olivine crystals react with silica in the remaining liquid and are converted into pyroxene, something like this:

\[
\text{Mg}_2\text{SiO}_4 + \text{SiO}_2 \rightarrow 2\text{MgSiO}_3
\]

olivine
pyroxene

As long as there is silica remaining and the rate of cooling is slow, this process continues down the discontinuous branch: once all of the olivine has reacted to form pyroxene, the pyroxene will react and form amphibole. Under the right conditions amphibole will form to biotite. Finally, if the magma is quite silica-rich to begin with, there will still be some left at around 750 °C to 800 °C, and from this last magma, potassium feldspar, quartz, and maybe muscovite mica will form.

If the magma cools enough, the first minerals to form will be completely used up in later chemical reactions. This is why we never have an igneous rock made of both olivine (at the top of the series) and quartz (at the bottom).

Notice that the sequence of minerals that form goes from isolated tetrahedra (olivine) toward increasingly complex arrangements of silica tetrahedra. Pyroxene consists of single chains, amphibole has double chains, mica has sheets of tetrahedra, and potassium feldspar and quartz at the bottom of the series have tetrahedra connected to each other in three dimensions.

Plagioclase Feldspar Changes Composition on the Continuous Side

On the right of Figure 7.5 is the continuous series. At about the point where pyroxene begins to crystallize, plagioclase feldspar also begins to crystallize. At that temperature, the plagioclase is calcium-rich (anorthite). As the temperature drops, and providing that there is sodium left in the magma, the plagioclase that forms is a more sodium-rich variety (albite). The series is continuous because the mineral is always plagioclase feldspar, but the series involves a transition from calcium-rich to sodium-rich.

When cooling happens relatively quickly, instead of getting crystals which are of uniform composition, individual plagioclase crystals can be zoned from calcium-rich in the centre to more sodium-rich around the outside. This occurs when calcium-rich early-forming plagioclase crystals become coated with progressively more sodium-rich plagioclase as the magma cools. Figure 7.7 shows a zoned plagioclase crystal as

Figure 7.7 Microscopic view of a zoned plagioclase crystal. The central part is calcium-rich and the outside part is sodium-rich: [Sandra Johnstone, used with permission]
seen under a microscope.

**Magma Composition: Mafic, Intermediate, and Felsic**

The composition of the original magma determines how far the reaction process can continue before all of the silica is used up. In other words, it determines which minerals will form. The compositions of typical mafic, intermediate, and felsic magmas are shown in Figure 7.8. Notice that these compositions are expressed in terms of oxides (e.g., Al\textsubscript{2}O\textsubscript{3} rather than just Al). There are two reasons for this: one is that in the early analytical procedures, the results were always expressed that way, and the other is that all of these elements combine readily with oxygen to form oxides.

![Figure 7.8](https://physicalgeology.pressbooks.com) The chemical compositions of typical mafic, intermediate, and felsic magmas, and the types of rocks that form from them. [Steven Earle CC-BY 4.0]

There are three trends to notice in Figure 7.8:

1. Silica increases from mafic to felsic magmas. Although there is a range of composition for each magma type, mafic magmas are about half SiO\textsubscript{2}, and felsic magmas are about 75% SiO\textsubscript{2}.
2. In mafic magmas, FeO + MgO + CaO accounts for 25% of the composition, but this decreases to only about 5% in felsic magmas.
3. Mafic magmas have about 5% Na\textsubscript{2}O + K\textsubscript{2}O and in felsic magmas this increases to around 10%.

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2 "Mafic" combines the words MAGnesium and FerrIC (containing iron).
3 "Felsic" combines the words FELdspar and SILLiCa.
Exercise 7.2 Mafic, Felsic, or Intermediate?

The proportions of the main chemical components of felsic, intermediate, and mafic magmas are listed in the table below. (The values are similar to those shown in Figure 7.8.)

<table>
<thead>
<tr>
<th>Oxide</th>
<th>Felsic Magma</th>
<th>Intermediate Magma</th>
<th>Mafic Magma</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>65-75%</td>
<td>55-65%</td>
<td>45-55%</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>12-16%</td>
<td>14-18%</td>
<td>14-18%</td>
</tr>
<tr>
<td>FeO</td>
<td>2-4%</td>
<td>4-8%</td>
<td>8-12%</td>
</tr>
<tr>
<td>CaO</td>
<td>1-4%</td>
<td>4-7%</td>
<td>7-11%</td>
</tr>
<tr>
<td>MgO</td>
<td>0-3%</td>
<td>2-6%</td>
<td>5-9%</td>
</tr>
<tr>
<td>Na₂O</td>
<td>2-6%</td>
<td>3-7%</td>
<td>1-3%</td>
</tr>
<tr>
<td>K₂O</td>
<td>3-5%</td>
<td>2-4%</td>
<td>0.5-3%</td>
</tr>
</tbody>
</table>

Chemical data for four rock samples are shown in the following table. Compare these with those in the table above to determine whether each of these samples is felsic, intermediate, or mafic.

<table>
<thead>
<tr>
<th>SiO₂</th>
<th>Al₂O₃</th>
<th>FeO</th>
<th>CaO</th>
<th>MgO</th>
<th>Na₂O</th>
<th>K₂O</th>
<th>Type?</th>
</tr>
</thead>
<tbody>
<tr>
<td>55%</td>
<td>17%</td>
<td>5%</td>
<td>6%</td>
<td>3%</td>
<td>4%</td>
<td>3%</td>
<td></td>
</tr>
<tr>
<td>74%</td>
<td>14%</td>
<td>3%</td>
<td>3%</td>
<td>0.5%</td>
<td>5%</td>
<td>4%</td>
<td></td>
</tr>
<tr>
<td>47%</td>
<td>14%</td>
<td>8%</td>
<td>10%</td>
<td>8%</td>
<td>1%</td>
<td>2%</td>
<td></td>
</tr>
<tr>
<td>65%</td>
<td>14%</td>
<td>4%</td>
<td>5%</td>
<td>4%</td>
<td>3%</td>
<td>3%</td>
<td></td>
</tr>
</tbody>
</table>

How Do We Get Different Magma Compositions?

Why Is There No Ultramafic Magma?

Refer back to the Bowen’s reaction series diagram in Figure 7.5, and notice that on the far right-hand side of the diagram under “Rock Types” the magma compositions we’ve just discussed are listed. At the very top of the list is ultramafic, which we haven’t mentioned yet. Ultramafic rocks have higher MgO than mafic rocks, and even less SiO₂.

The vast majority of silicate rocks on Earth are ultramafic rocks, because that’s what the mantle is made of. Nevertheless, we don’t talk about ultramafic magma because we never encounter it… you could say it’s gone extinct. The reason is that although the Earth was once hot enough to have ultramafic magma, it is no longer hot enough to melt ultramafic rocks. There do exist ultramafic volcanic rocks, called komatiites, but with two notable exceptions, the youngest of these is 2 billion years old.

Partial Melting Makes Magma That Is Richer In Silica

In section 7.1 we discussed partial melting, where some components of a mixture melt before others do. In the case of mafic magma, ultramafic rocks undergo partial melting to produce mafic magma. One place this

4 The komatiites of the Song Da zone in northwestern Vietnam are 270 million years old, and those on Gorgona Island, Columbia are 89 million years old. Exactly how they formed is still a bit of a mystery. See Table 1 of arXiv:physics/0512118v2 [physics.geo-ph] for a compilation of komatiite ages with references.
happens is along ocean spreading centres where there is less pressure from the overlying mass of the lithosphere, and decompression is an important driver of melting.

In general, silicate minerals with more silica will melt before those with less silica. This means the melt will have more silica than the rock as a whole.

**Fractional Crystallization Also Makes Magma Richer In Silica**

A number of processes that take place within a magma chamber can affect the types of rocks produced in the end. If the magma has a low viscosity (i.e., it’s runny) — which is likely if it is mafic — the crystals that form early, such as olivine (Figure 7.9a), may slowly settle toward the bottom of the magma chamber (Figure 7.9b). This process is called *fractional crystallization*.

The formation of olivine removes iron- and magnesium-rich components, leaving the overall composition of the magma near the top of the magma chamber more felsic. The crystals that settle might either form an olivine-rich layer near the bottom of the magma chamber, or they might re-melt because the lower part is likely to be hotter than the upper part. If any melting takes place, crystal settling will make the magma at the bottom of the chamber more mafic than it was to begin with (Figure 7.9c).

![Fractional Crystallization Diagram](https://physicalgeology.pressbooks.com/page/79)
Magma Composition Also Changes When Other Rocks Are Melted And Mixed In

Magma chambers aren’t isolated from their surroundings. The rock in which the magma chamber is located (called the country rock) can melt, adding to the magma already in the magma chamber (Figure 7.10). Sometimes magma carries fragments of unmelted rock, called xenoliths, with it. Melting of xenoliths can also alter the composition of magma, as can re-melting of crystals that have settled out of the magma.

7.3 Classification of Igneous Rocks

Igneous Rocks Are Classified By Mineral Abundance

In the last section you learned that igneous rocks are classified into four categories based on their chemical composition: felsic, intermediate, mafic, and ultramafic. From the diagram of Bowen’s reaction series in Figure 7.5, it is clear that differences in chemical composition correspond to differences in the types of minerals within an igneous rock. Igneous rocks are given names based on the proportion of different minerals which they contain. Figure 7.11 is a diagram with the minerals from Bowen’s reaction series, and is used to decide which name to give an igneous rock.

To see how Figure 7.11 works, first notice the scale in percent along the vertical axis. The interval between each tick mark represents 10% of the minerals within a rock. An igneous rock can be represented as a vertical line drawn through the diagram, and the vertical scale used to break down the proportion of each mineral it contains. For example, the arrows in the ultramafic field of the diagram represent a rock containing 40% olivine and 60% pyroxene. An igneous rock at the boundary between the mafic and ultramafic fields (marked with the vertical dashed line) would have approximately 15% olivine, 75% pyroxene, and 10% Ca-rich plagioclase feldspar.
Igneous Rocks Are Also Classified By Grain Size

The name an igneous rock gets also depends on whether it cools within the Earth (an intrusive or plutonic igneous rock), or whether it cools on the Earth’s surface after erupting from a volcano (an extrusive or volcanic igneous rock). For example, a felsic intrusive rock is called granite, whereas a felsic extrusive rock is called rhyolite.

A key difference between intrusive and extrusive igneous rocks is the size of crystals making them up. The longer magma has to cool, the larger the crystals within it can become. Magma cools much more slowly within the Earth than on Earth’s surface because magma within the Earth is insulated by surrounding rock. Notice that in Figure 7.11, the intrusive rocks in the first row have crystals large enough that you can identify individual crystals. This is referred to as phaneritic texture. The extrusive rocks in the second row have much smaller crystals. The crystals are so small that the rock looks like a dull mass, and it isn’t possible to see individual crystals. This texture is referred to as aphanitic. Table 7.1 summarizes the key differences between intrusive and extrusive igneous rocks.
Table 7.1 Comparison of intrusive and extrusive igneous rocks

<table>
<thead>
<tr>
<th></th>
<th>Magma cools within the Earth</th>
<th>Lava cools on Earth's surface</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Terminology</strong></td>
<td>intrusive/plutonic</td>
<td>extrusive/volcanic</td>
</tr>
<tr>
<td><strong>Cooling rate</strong></td>
<td>Slow: surrounding rocks insulate the magma chamber</td>
<td>Rapid: heat is exchanged with the atmosphere</td>
</tr>
<tr>
<td><strong>Texture</strong></td>
<td>Phaneritic: crystals are large enough to see without significant magnification (coarse grained)</td>
<td>Aphanitic: crystals are too small to see without significant magnification (fine grained)</td>
</tr>
</tbody>
</table>

What this means is that two igneous rocks with exactly the same minerals making them up, and in the same proportions, can have different names. If the rock is of intermediate composition it is diorite if it is course-grained, and andesite if it is fine-grained. If the rock is mafic it is gabbro if it is course-grained, and basalt if fine-grained. If the rock is ultramafic, the course-grained version is peridotite, and the fine-grained version is komatite. It makes sense to use different names because rocks of different grain sizes form in different ways and in different geologic settings.

**Does That Mean an Igneous Rock Will Only Ever Have Crystals of One Grain Size?**

No. Something interesting happens when there is a change in the rate at which melted rock is cooling. If magma is cooling in a magma chamber, some minerals will begin to crystallize before others do. If cooling is slow enough, those crystals can become quite large. Now imagine the magma is suddenly heaved out of the magma chamber and erupted from a volcano. The larger crystals will flow out with the lava. The lava will then cool very rapidly, and the larger crystals will be surrounded by much smaller ones. An igneous rock with crystals of distinctly different size is said to have a porphyritic texture, or might be referred to as a porphyry. The larger crystals are called phenocrysts, and the smaller ones are referred to as the groundmass or matrix. Figure 7.12 shows a porphyritic rhyolite with quartz and potassium feldspar phenocrysts within a darker groundmass.
Exercise 7.3 Which Mineral Will the Phenocryst Be?

As a magma cools below 1300°C, minerals start to crystallize within it. If that magma is then involved in a volcanic eruption, the rest of the liquid will cool quickly to form a porphyritic texture. The rock will have some relatively large crystals (phenocrysts) of the minerals that crystallized early, and the rest will be very fine grained or even glassy. Using the diagram shown here, predict what phenocrysts might be present where the magma cooled as far as line a in one case, and line b in another.

Bowen’s reactions series. [Steven Earle CC-BY 4.0]

Classifying Igneous Rocks According to the Proportion of Dark Minerals

If you aren’t exactly sure which minerals are present in an intrusive igneous rock that you’re studying, there is a quick way to approximate the composition of that rock. In general, igneous rocks have an increasing proportion of dark minerals as they become more mafic. The dark minerals are those higher in iron and magnesium (e.g., olivine, pyroxene), and for that reason they are sometimes referred to collectively as ferromagnesian minerals. Figure 7.13 is the result of simplifying Figure 7.11 by grouping minerals as either light or dark.

Figure 7.13 Simplified igneous rock classification according to the proportion of dark (ferromagnesian) minerals. [Karla Panchuk CC-BY-ND 4.0]
By estimating the proportion of light minerals to dark minerals in a sample, it is possible to place that sample in Figure 7.13. Guides like the ones in Figure 7.14 are used to help visualize the proportions of light and dark.

![Figure 7.14 A guide for estimating the proportion of dark minerals in an igneous rock. [Steven Earle CC-BY 4.0]](image)

It is important to note that the method of estimating the proportion of dark minerals is approximate as a means for identifying igneous rocks. At issue is the fact that Na-rich plagioclase feldspar is light in colour, whereas Ca-rich plagioclase feldspar can appear darker (Figure 7.15), especially when surrounded by darker minerals. Nevertheless, plagioclase feldspar does not qualify as ferromagnesian, so it falls in the light-coloured minerals field in Figure 7.13.

![Figure 7.15 Na-rich and Ca-rich plagioclase feldspars differ in colour. [Karla Panchuk CC-BY-ND 4.0]](image)

**Exercise 7.4 Classifying Igneous Rocks By Proportion of Ferromagnesian Minerals**

The four igneous rocks shown below have differing proportions of ferromagnesian silicates (dark minerals). Estimate those proportions using the diagrams in Figure 7.14, and then use Figure 7.13 to determine the likely rock name for each one.

![Exercise 7.4 images](image)
Classifying Igneous Rocks When No Crystals Are Visible

The method of estimating the percentage of minerals works well for phaneritic igneous rocks, where crystals are visible with the naked eye or with the use of a magnifying lens. If an igneous rock is porphyritic but otherwise aphanitic as in Figure 7.12, the minerals present as phenocrysts can be used to identify the rock (e.g., Exercise 7.3). However, there are two cases where mineral composition cannot be determined by looking at visible crystals. These include volcanic rocks without phenocrysts, and glassy igneous rocks.

Volcanic Rocks Without Phenocrysts

Without the aid of visible crystals or phenocrysts, volcanic rocks can be classified on the basis of colour and sometimes on other textural features. As you may have noticed in Figure 7.11, the colour of volcanic rocks moves from light to dark as the composition goes from felsic to mafic. Figure 7.16 shows this progression with rhyolite (often a tan or pinkish colour), andesite (often grey), and basalt (ranging from brown to dark green to black).

Basalt is often accompanied by two textural features, vesicles and amygdules. As magma travels to the surface and erupts as lava, the decrease in pressure allows gases dissolved in the lava to be released and form bubbles. When the lava freezes around the bubbles, vesicles are formed (circular inset in 7.16). If minerals later form within the vesicles, the filled vesicles are called amygdules (box inset in Figure 7.16).

Glassy Volcanic Rocks

Crystal size is a function of cooling rate. The faster magma or lava cools, the smaller the crystals it contains. It is possible for lava to cool so rapidly that no crystals can form. The result is referred to as volcanic glass. Volcanic glass can be smooth like obsidian or vesicular like scoria and pumice (Figure 7.17). The combination of a low-density felsic composition and enclosed vesicles mean that pumice can float on water.
7.4 Intrusive Igneous Bodies

In most cases, a body of hot magma is less dense than the rock surrounding it, so it has a tendency to creep upward toward the surface. It does so in a few different ways:

- Filling and widening existing cracks
- Melting the surrounding rock
- Pushing the rock aside (where the rock is hot enough and under enough pressure to deform without breaking)
- Breaking the rock

When magma forces itself into cracks, breaks off pieces of rock, and then envelops them, this is called *stoping*. The resulting fragments are called *xenoliths*. In Figure 7.18, the dark patches are xenoliths.

Some of the magma reaches the surface, resulting in volcanic eruptions, but most cools within the crust. The resulting body of rock is called a *pluton*. Plutons can have different shapes and different relationships with the surrounding country rock (Figure 7.19). These characteristics determine what name the pluton is given.

Large, irregularly shaped plutons are called *stocks* or *batholiths*, depending on size. Tabular plutons are called *dikes* if they cut across existing structures, and *sills* if they do not. *Laccoliths* are like sills, except they have caused the overlying rocks to bulge upward. *Pipes* are cylindrical conduits.

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5 From the Greek words xenos, meaning "foreigner" or "stranger," and lithos for "stone."
6 After Pluto was demoted from planet status, astronomers tried to come up with a name for objects like Pluto. For a while they considered "pluton" however geologists rightly objected that they had first claim on the word. In the end the International Astronomical Union settled on "dwarf planet" instead.
Types of Plutons

Stocks and Batholiths

Large irregular-shaped plutons are called either stocks or batholiths, depending on the area exposed at the surface. If the body has an exposed surface area greater than 100 km$^2$, then it’s a batholith, otherwise it’s a stock. Batholiths are typically formed when a number of stocks coalesce beneath the surface to create one large body. One of the largest batholiths in the world is the Coast Range Plutonic Complex$^7$, which extends all the way from the Vancouver region to southeastern Alaska (Figure 7.20). Because this naming scheme depends on the area of rock that is exposed, it is still possible that a stock is a small part of a larger batholith-sized complex deeper within the Earth.

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$^7$ Also referred to as the Coast Range Batholith
Dikes, Sills, and Laccoliths

Tabular (sheet-like) plutons are classified according to whether or not they are concordant with (parallel to) existing layering (e.g., sedimentary bedding or metamorphic foliation) in the country rock. A sill is concordant with existing layering, and a dike is discordant. If the country rock has no bedding or foliation, then any tabular body within it is a dike. Note that the sill-versus-dike designation is not determined simply by the orientation of the feature. A dike could be horizontal and a sill could be vertical— it all depends on the orientation of features in the surrounding rocks.

A laccolith is a sill-like body which has expanded upward by deforming the overlying rock.

Pipes

A pipe, as the name suggests, is a cylindrical body with a circular, elliptical, or even irregular cross-section, which serves as a conduit (or pipeline) for the movement of magma from one location to another. Pipes may feed volcanoes, but pipes can also connect plutons. It is also possible for a dike to feed a volcano.

Chilled Margins

Plutons can interact with the rocks into which they are intruded. This might lead to partial melting of the country rock, or to stoping and formation of xenoliths. And, as we’ll see in Chapter 10, the heat of a body of magma can lead to metamorphism of the country rock, causing mineralogical and textural changes. However, it is also the case that the country rock can affect the magma.

The most obvious effect that country rock has on magma is a chilled margin along the edges of the pluton. The country rock is much cooler than the magma, so magma that comes into contact with the country rock cools much faster than magma toward the interior of the pluton. Rapid cooling leads to smaller crystals, so the texture along the edges of the pluton is different from that of the interior of the pluton, and the colour may be different. An example is shown in Figure 7.21.

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8 Sedimentary bedding refers to the layers in which sedimentary rocks form. Metamorphic foliation refers to the way minerals or other elements in a rock are aligned as a result of being deformed by heat and pressure. Bedding and foliation will be discussed in more detail in later chapters.

9 Also spelled dyke.
Exercise 7.5 Pluton Problems

The diagram below is a cross-section through part of the crust showing a variety of intrusive igneous rocks. Indicate whether each of the plutons labelled a to e on the diagram below is a dike, a sill, a stock, or a batholith. (Note the trees for scale.)

Chapter Summary

The topics covered in this chapter can be summarized as follows:

7.1 Magma and Magma Formation

Magma is molten rock, and in most cases, it forms from partial melting of existing rock. The two main processes of magma formation are decompression melting and flux melting. Magmas range in composition from ultramafic to felsic. Mafic rocks are rich in iron, magnesium, and calcium and have around 50% silica. Felsic rocks are rich in silica (~75%) and have lower levels of iron, magnesium, and calcium and higher levels of sodium and potassium than mafic rocks.

7.2 Crystallization of Magma

As a body of magma starts to cool, the first process to take place is the polymerization of silica tetrahedra into chains. This increases the magma’s viscosity (makes it thicker) and because felsic magmas have more silica than mafic magmas, they tend to be more viscous. The Bowen reaction series allows us to predict the order of crystallization of magma as it cools. Magma can be modified by fractional crystallization (separation of early-forming crystals) and by incorporation of material from the surrounding rocks by partial melting.

7.3 Classification of Igneous Rocks

Igneous rocks are classified based on their mineral composition and texture. Felsic igneous rocks have less than 20% of dark minerals (ferromagnesian silicates, amphibole and/or biotite) plus varying amounts of quartz, and both potassium and plagioclase feldspars. Mafic igneous rocks have more than 50% dark minerals (primarily pyroxene) plus plagioclase feldspar. Most intrusive igneous rocks are phaneritic (crystals are visible to the naked eye). If there were two stages of cooling (slow then fast), the texture may be porphyritic (large crystals in a matrix of smaller crystals). Glassy rocks have cooled too fast to allow crystals to form.
7.4 Intrusive Igneous Bodies

Magma intrudes into country rock by pushing it aside or melting through it. Intrusive igneous bodies tend to be either irregular (stocks and batholiths), tabular (dikes and sills), or pipe-like. Batholiths have exposed areas of greater than 100 km$^2$, while stocks are smaller. Sills are parallel to existing layering in the country rock, while dykes cut across layering. A pluton that intruded into cold rock it is likely to have a chilled margin.

Questions for Review

1. What is the significance of the term reaction in the name of the Bowen reaction series?

2. Why is it common for plagioclase crystals to be zoned from relatively calcium-rich in the middle to more sodium-rich on the outside?

3. What must happen within a magma chamber for fractional crystallization to take place?

4. Explain the difference between aphanitic and phaneritic textures.

5. Explain the difference between porphyritic and pegmatitic textures.

6. Name the following rocks:
   (a) An extrusive rock with 40% Ca-rich plagioclase and 60% pyroxene
   (b) An intrusive rock with 65% plagioclase, 25% amphibole, and 10% pyroxene
   (c) An intrusive rock with 25% quartz, 20% orthoclase, 50% feldspar, and minor amounts of biotite

7. With respect to tabular intrusive bodies, what is the difference between a concordant body and a discordant body?

8. Why does a dike commonly have a fine-grained margin?

9. What is the difference between a batholith and a stock?

10. Describe two ways in which batholiths intrude into existing rock.

11. Why is compositional layering a common feature of mafic plutons but not of felsic plutons?