Chapter 3 Intrusive Igneous Rocks

Learning Objectives

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

- Describe the rock cycle and the types of processes that lead to the formation of igneous, sedimentary, and metamorphic rocks, and explain why there is an active rock cycle on Earth.
- Explain the concept of partial melting and describe 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 rock textures.
- Identify plutons on the basis of their morphology and their relationships to the surrounding rocks.
- Explain the origin of a chilled margin.

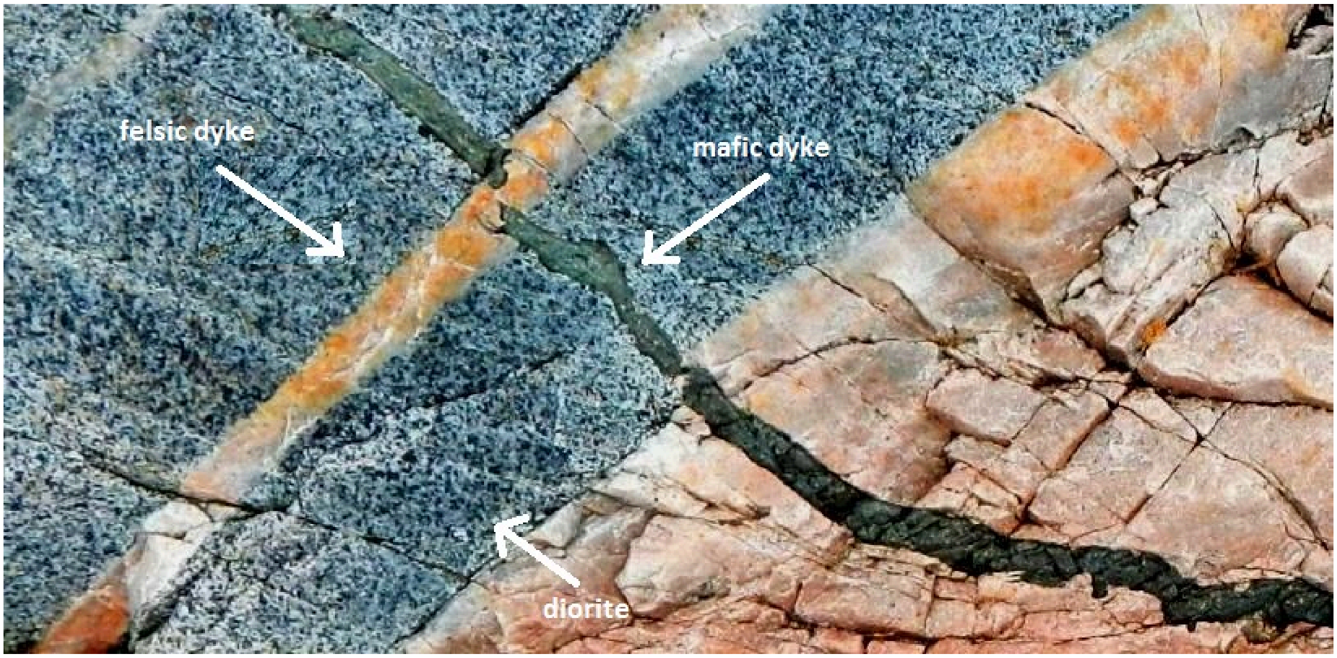

Figure 3.0.1 A fine-grained mafic dyke (dark green) intruded into a felsic dyke (pink) and into coarse diorite (grey), Quadra Island, B.C. All of these rocks are composed of more than one type of mineral. The mineral components are clearly visible in the diorite, but not in the other two rock types.

A rock is a consolidated mixture of minerals. By *consolidated*, we mean hard and strong; real rocks don’t fall apart in your hands! A *mixture of minerals* implies the presence of more than one mineral grain, but not necessarily more than one type of mineral (Figure 3.0.1). A rock can be composed of only one type of mineral (e.g., limestone is commonly made up of only calcite), but most rocks are composed of several different minerals. A rock can also include non-minerals, such as fossils or the organic matter within a coal bed or in some types of mudstone.

Rocks are grouped into three main categories based on how they form:

1. **Igneous:** formed from the cooling and crystallization of magma (molten rock)
2. **Sedimentary**: formed when weathered fragments of other rocks are buried, compressed, and cemented together, or when minerals precipitate directly from solution
3. **Metamorphic**: formed by alteration (due to heat, pressure, and/or chemical action) of a pre-existing igneous or sedimentary rock

Media Attributions

- Figure 3.0.1: © Steven Earle. Adapted by Josie Gray (labelled). CC BY.

3.1 The Rock Cycle

The rock components of the crust are slowly but constantly being changed from one form to another and the processes involved are summarized in the **rock cycle** (Figure 3.1.1). The rock cycle is driven by two forces: (1) Earth's internal heat engine, which moves material around in the core and the mantle and leads to slow but significant changes within the crust, and (2) the hydrological cycle, which is the movement of water, ice, and air at the surface, and is powered by the sun.

The rock cycle is still active on Earth because our core is hot enough to keep the mantle moving, our atmosphere is relatively thick, and we have liquid water. On some other planets or their satellites, such as the Moon, the rock cycle is virtually dead because the core is no longer hot enough to drive mantle convection and there is no atmosphere or liquid water.

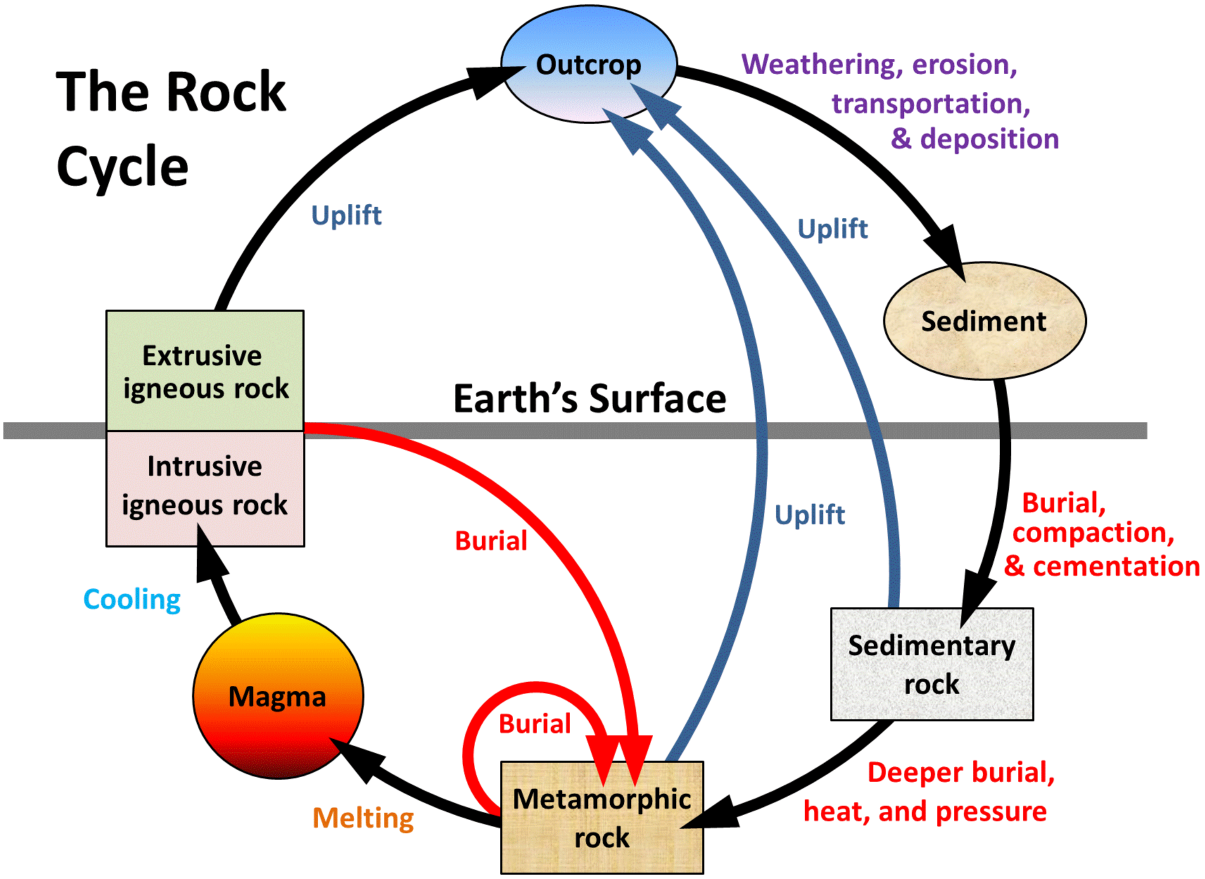

Figure 3.1.1 A schematic view of the rock cycle. [Image description]

In describing the rock cycle, we can start anywhere we like, although it's convenient to start with magma. As we'll see in more detail below, magma is rock that is hot to the point of being entirely molten, with a temperature of between about 800° and 1300°C, depending on the composition and the pressure.

Figure 3.1.2 Magma forming pahoehoe basalt at Kilauea Volcano, Hawaii.

Magma can either cool slowly within the crust (over centuries to millions of years)—forming **intrusive igneous rock**, or erupt onto the surface and cool quickly (within seconds to years)—forming **extrusive igneous rock** (volcanic rock) (Figure 3.1.2). Intrusive igneous rock typically crystallizes at depths of hundreds of metres to tens of kilometres below the surface. To change its position in the rock cycle, intrusive igneous rock has to be uplifted and then exposed by the erosion of the overlying rocks.

Through the various plate-tectonics-related processes of mountain building, all types of rocks are uplifted and exposed at the surface. Once exposed, they are weathered, both physically (by mechanical breaking of the rock) and chemically (by weathering of the minerals), and the weathering products—mostly small rock and mineral fragments—are eroded, transported, and then deposited as **sediments**. Transportation and deposition occur through the action of glaciers, streams, waves, wind, and other agents, and sediments are deposited in rivers, lakes, deserts, and the ocean.

Exercise 3.1 Rock around the rock-cycle clock

Referring to the rock cycle (Figure 3.1.1), list the steps that are necessary to cycle some geological material starting with a sedimentary rock, which then gets converted into a metamorphic rock, and eventually a new sedimentary rock.

A *conservative* estimate is that each of these steps would take approximately 20 million years (some may be less, others would be more, and some could be much more). How long might it take for this entire process to be completed?

See Appendix 3 for Exercise 3.1 Answers.

Figure 3.1.3 Cretaceous-aged marine sandstone overlying marine mudstone, Gabriola Island, B.C.

Unless they are re-eroded and moved along, sediments will eventually be buried by more sediments. At depths of hundreds of metres or more, they become compressed and cemented into **sedimentary rock** (See Figure 3.1.3 for example). Again through various means, largely resulting from plate-tectonic forces, different kinds of rocks are either uplifted, to be re-eroded, or buried deeper within the crust where they are heated up, squeezed, and changed into **metamorphic rock** (Figure 3.1.4)

Figure 3.1.4 Metamorphosed and folded Triassic-aged limestone, Quadra Island, B.C.

Image Descriptions

Figure 3.1.1 image description: The rock cycle takes place both above and below the Earth's surface. The rock deepest beneath the earth's surface, and under extreme heat and pressure, is metamorphic rock.

This metamorphic rock can melt and become magma. When magma cools below the earth's surface, it becomes "intrusive igneous rock." If magma cools above the earth's surface, it is "extrusive igneous rock" and becomes part of the outcrop. The outcrop is subject to weathering and erosion, and can be moved and redeposited around the earth by forces such as water and wind. As the outcrop is eroded, it becomes sediment which can be buried, compacted, and cemented beneath the Earth's surface to become sedimentary rock. As sedimentary rock gets buried deeper and comes under increased heat and pressure, it returns to its original state as metamorphic rock. Rocks in the rock cycle do not always make a complete loop. It is possible for sedimentary rock to be uplifted back above the Earth's surface and for intrusive and extrusive igneous rock to be reburied and become metamorphic rock. [Return to Figure 3.1.1]

Images Attributions

- Figure 3.1.1, 3.1.2, 3.1.3, 3.1.4: © Steven Earle. CC BY.

3.2 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: oxygen, silicon, aluminum, iron, calcium, sodium, magnesium, and potassium (Figure 3.2.1). Oxygen, the most abundant element in magma, comprises a little less than half the total, followed by silicon at just over one-quarter. The remaining elements make up the other one-quarter. Magmas derived from crustal material are dominated by oxygen, silicon, aluminum, sodium, and potassium.

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.

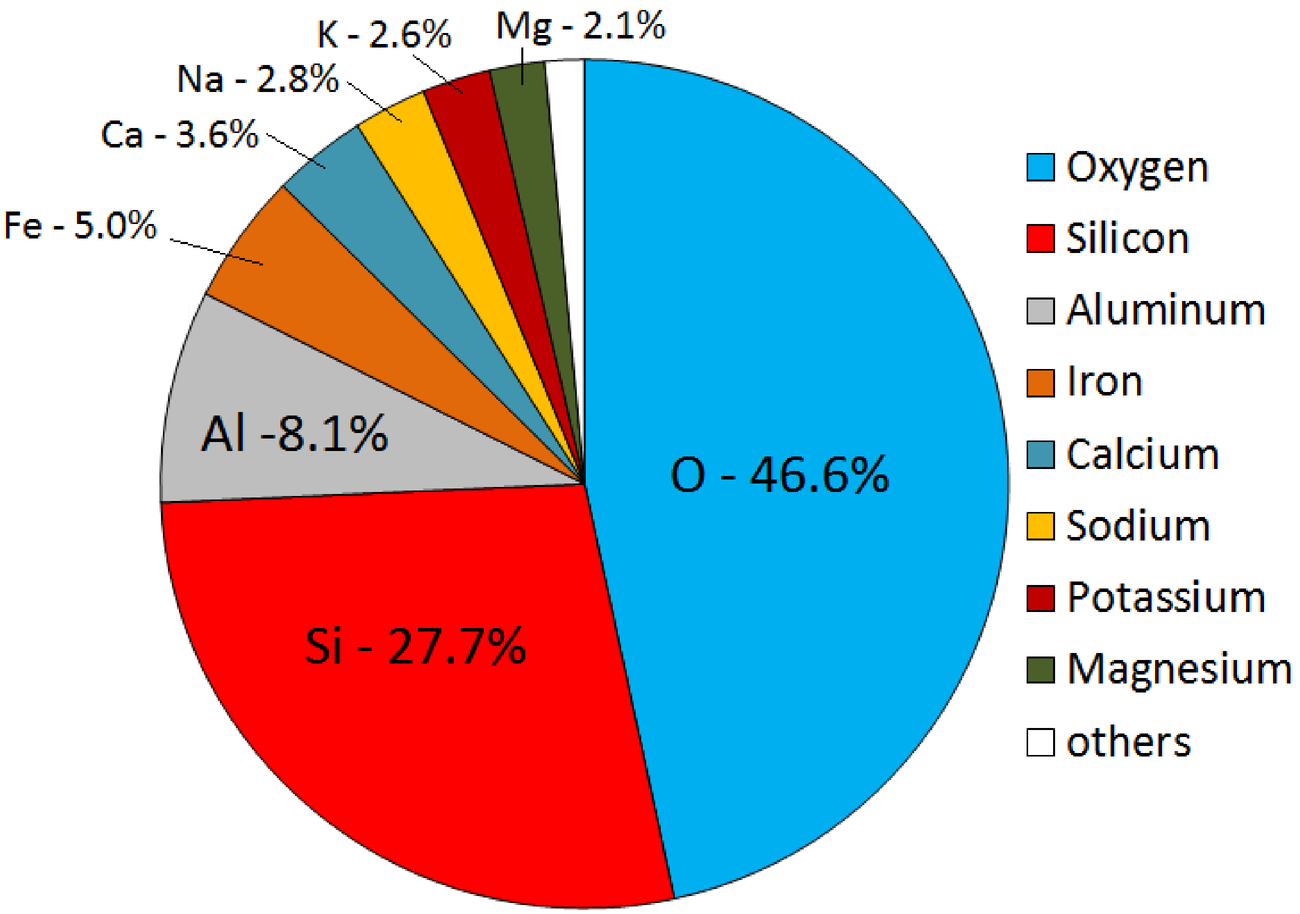

Figure 3.2.1 Average elemental proportions in Earth's crust, which is close to the average composition of magmas within the crust. [Image Description]

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 parts of a rock melt; it takes place because rocks are not pure materials. Most rocks are made up of several minerals, each of which has a different melting temperature. The wax in a candle is a pure material. If you put some wax into a warm oven (50°C will do as the melting temperature of most wax is about 40°C) and leave it there for a while, it will soon start to melt. That's complete melting, not partial melting. If instead you took a mixture of wax, plastic, aluminum, and glass and put it into the same warm oven, the wax would soon start to melt, but the plastic, aluminum, and glass would not melt (Figure 3.2.2a). That's partial melting and the result would be solid plastic, aluminum, and glass surrounded by liquid wax (Figure 3.2.2b). If we heat the oven up to around 120°C, the plastic would melt too and mix with the liquid wax, but the aluminum and glass would remain solid (Figure 3.2.2c). Again this is partial melting. If we separated the wax/plastic "magma" from the other components and let it cool, it would eventually harden. As you can see from Figure 3.2.2d, the liquid wax and plastic have mixed, and on cooling, have formed what looks like a single solid substance. It is most likely that this is a very fine-grained mixture of solid wax and solid plastic, but it could also be some other substance that has formed from the combination of the two.

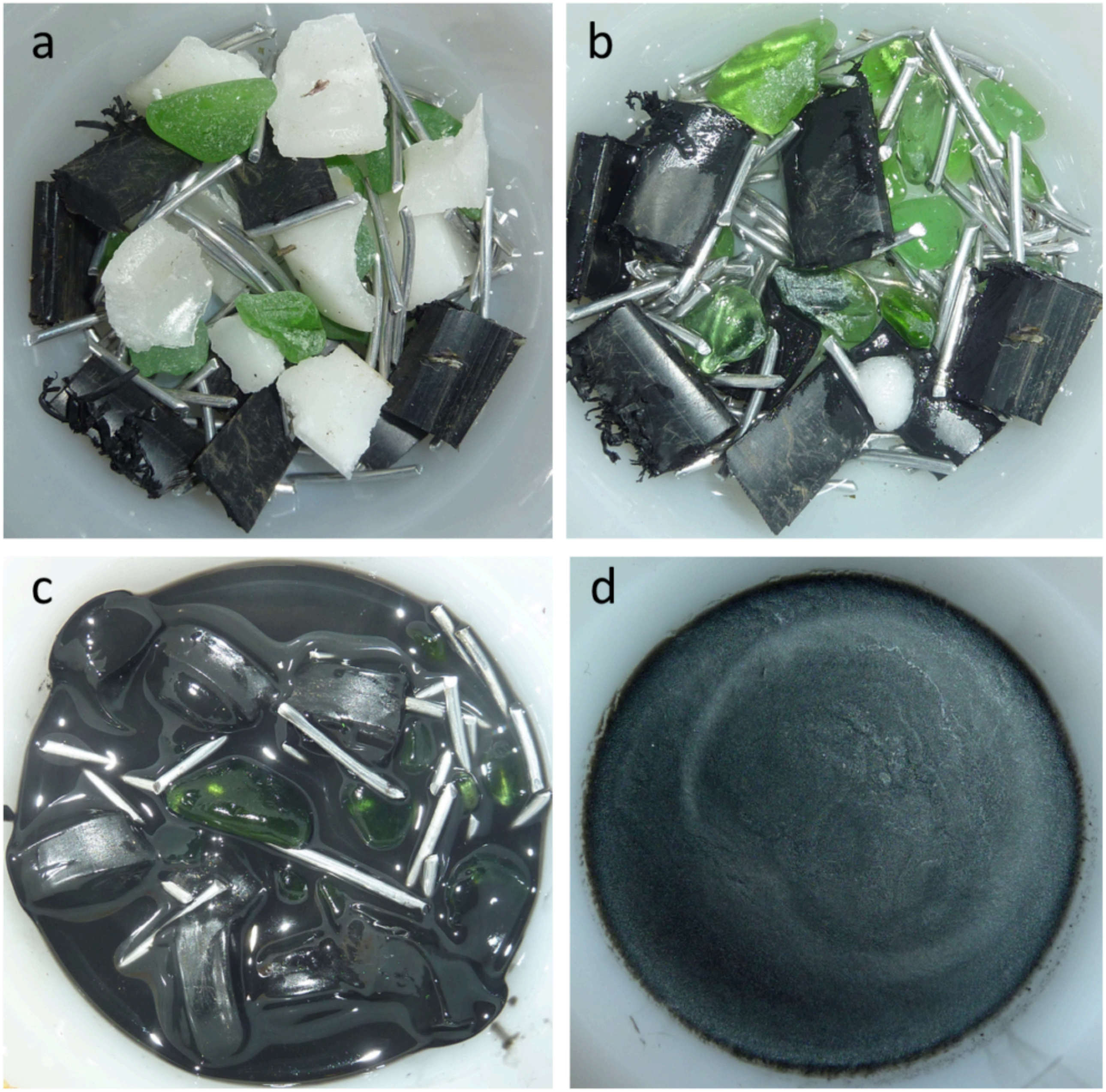

Figure 3.2.2 Partial melting of "pretend rock": (a) the original components of white candle wax, black plastic pipe, green beach glass, and aluminum wire, (b) after heating to 50˚C for 30 minutes only the wax has melted, (c) after heating to 120˚C for 60 minutes much of the plastic has melted and the two liquids have mixed, (d) the liquid has been separated from the solids and allowed to cool to make a "pretend rock" with a different overall composition.

In this example, we partially melted some pretend rock to create some pretend magma. We then separated the magma from the source and allowed it to cool to make a new pretend rock with a composition quite different from the original material (it lacks glass and aluminum).

Of course partial melting in the real world isn't exactly the same as in our pretend-rock example. The main differences are that rocks are much more complex than the four-component system we used, and the mineral components of most rocks have more similar melting temperatures, so two or more minerals are likely to melt at the same time to varying degrees. Another important difference is that when rocks melt, the process takes thousands to millions of years, not the 90 minutes it took in the pretend-rock example.

Contrary to what one might expect, and contrary to what we did to make our pretend rock, most partial melting of real rock does not involve heating the rock up. The two main mechanisms through which rocks melt are **decompression melting** and **flux melting**. Decompression melting takes place within Earth when a body of rock is held at approximately the same temperature but the pressure is reduced. This happens because the rock is being moved toward the surface, either at a **mantle plume** (a.k.a., hot spot), or in the upwelling part of a mantle convection cell.[1] The mechanism of decompression melting is shown in Figure 3.2.3a. 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 starts to take place. The process of flux melting is shown in Figure 3.2.3b. If a rock is close to its melting point and some water (a flux that promotes melting) is added to the rock, the melting temperature is reduced (solid line versus dotted line), and partial melting starts.

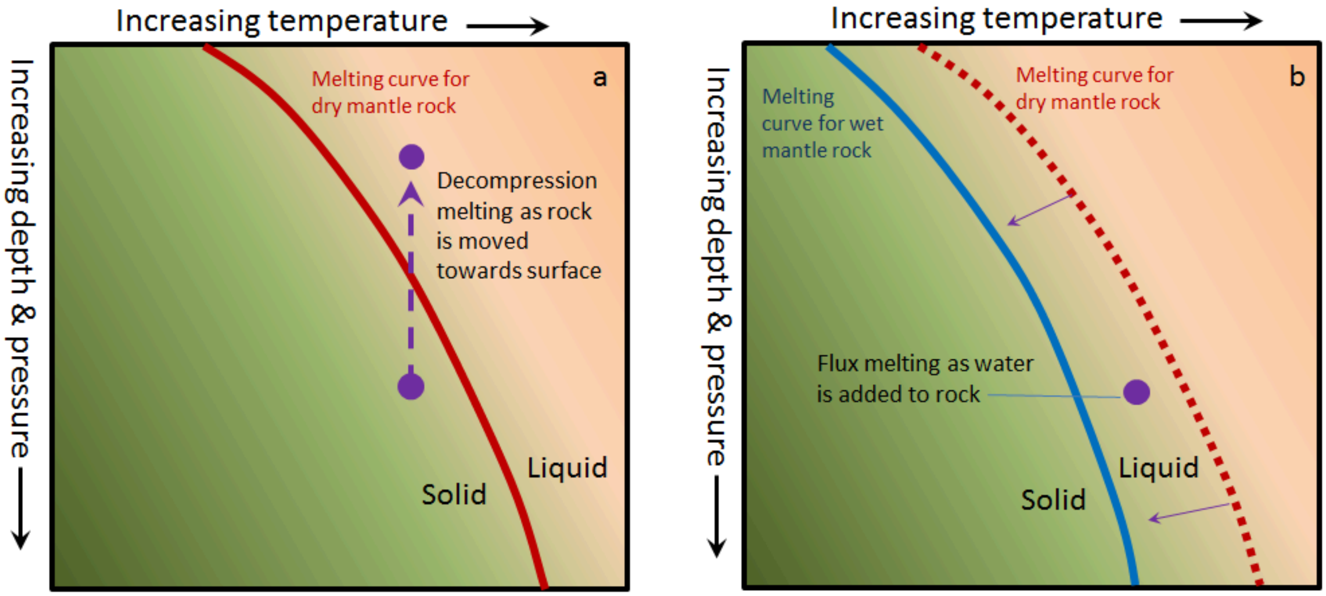

Figure 3.2.3 Mechanisms for (a) decompression melting (the rock is moved toward the surface) and (b) flux melting (water is added to the rock) and the melting curve is displaced. [Image Description]

The partial melting of rock happens in a wide range of situations, most of which are related to plate tectonics. The more important of these are shown in Figure 3.2.3. At both mantle plumes and in the upward parts of convection systems, rock is being moved toward the surface, the pressure is dropping, and at some point, the rock crosses to the liquid side of its melting curve. At subduction zones, water from the wet, subducting oceanic crust is transferred 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 only about 10% of the rock melts—and 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 our pretend rock is richer in wax and plastic than the “rock” from which it was derived.) The magma produced, being less dense than the surrounding rock, moves up through the mantle, and eventually into the crust.

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[1] Mantle plumes are described in Chapter 4 and mantle convection in Chapter 9.

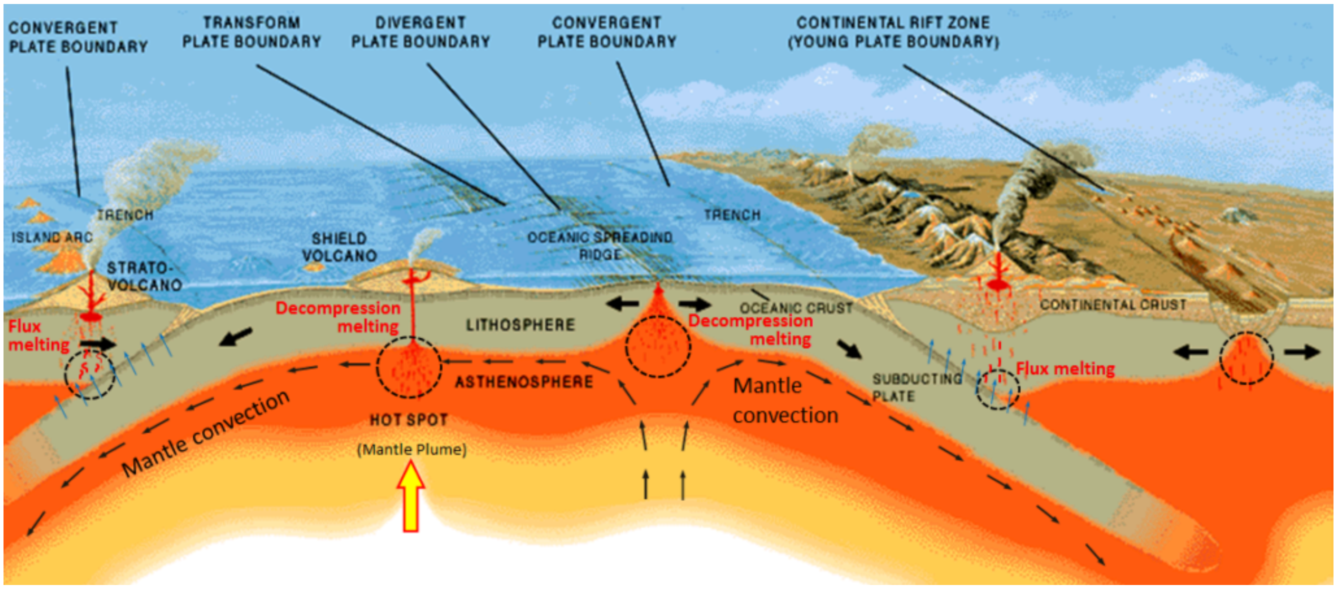

Figure 3.2.4 Common sites of magma formation in the upper mantle. The black circles are regions of partial melting. The blue arrows represent water being transferred from the subducting plates into the overlying mantle.

As it moves toward the surface, and especially when it moves from the mantle into the lower crust, the hot magma interacts with the surrounding rock. This typically leads to partial melting of the surrounding rock because most such magmas are hotter than the melting temperature of crustal rock. (In this case, melting is caused by an increase in temperature.) Again, 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 because there is too much energy for the atoms to bond together. As the temperature drops, usually because the magma is slowly moving upward, things start to change. Silicon and oxygen combine to form silica tetrahedra, and then, as cooling continues, the tetrahedra start to link together to make chains (**polymerize**). These silica chains have the important effect of making the magma more viscous (less runny), and as we'll see in Chapter 4, magma viscosity has significant implications for volcanic eruptions. As the magma continues to cool, crystals start to form.

Exercise 3.2 Making magma viscous

This is an 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 — something like porridge — because there is more gluten and more chains have been formed (see the photo).

Figure 3.2.5 Flour-and-water magma experiment.

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

See Appendix 3 for Exercise 3.2 answers.

Image Descriptions

Figure 3.2.1 image description: The average elemental proportions in the Earth's crust from the largest amount to the smallest amount. Oxygen (46.6%), Silicon (27.7%), Aluminum (8.1%), Iron (5.0%), Calcium (3.6%), Sodium (2.8%), Potassium (2.6%), Magnesium (2.1%), Others (1.5%). [Return to Figure 3.2.1]

Figure 3.2.3a image description: Dry mantle rock is predominately solid. However, its melting point is dependent on the temperature and pressure the rock is under. The higher the pressure (meaning the farther the rock is from the Earth's surface), the more likely dry mantle rock is going to be solid. Dry mantle rock under extreme pressure requires a much higher temperature to melt than dry mantle rock under less pressure. As pressure drops (meaning as the rock rises towards the Earth's surface), the required temperature to melt the mantle rock drops as well.

Figure 3.2.3b image description: In comparison to dry mantle rock, wet mantle rock under the same amount of pressure (at the same distance from the earth's surface) requires a lower temperature to melt.

When liquid is added to dry mantel rock at a pressure and temperature point in which wet mantle rock would be melted, flux melting occurs. [Return to Figure 3.2.3]

Media Attributions

- Figure 3.2.1, 3.2.2, 3.2.3, 3.2.5: © Steven Earle. CC BY.
- Figure 3.2.4: "Cross section" by José F. Vigil from *This Dynamic Planet* — a wall map produced jointly by the U.S. Geological Survey, the Smithsonian Institution, and the U.S. Naval Research Laboratory. Adapted by Steven Earle. Public domain.

3.3 Crystallization of Magma

The minerals that make up igneous rocks crystallize 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 3.3.1 and Figure 3.3.3).

Of the common silicate minerals, olivine normally crystallizes first, at between 1200° and 1300°C. As the temperature drops, and assuming that some silica remains in the magma, the olivine crystals will react (combine) with some of the silica in the magma to form pyroxene. As long as there is silica remaining and the rate of cooling is slow, this process continues down the discontinuous branch: olivine to pyroxene, pyroxene to amphibole, and (under the right conditions) amphibole to biotite.

At about the point where pyroxene begins to crystallize, plagioclase feldspar also begins to crystallize. At that temperature, the plagioclase is calcium-rich (anorthite) (see Figure 2.6.1). As the temperature drops, and providing that there is sodium left in the magma, the plagioclase that forms is a more sodium-rich variety.

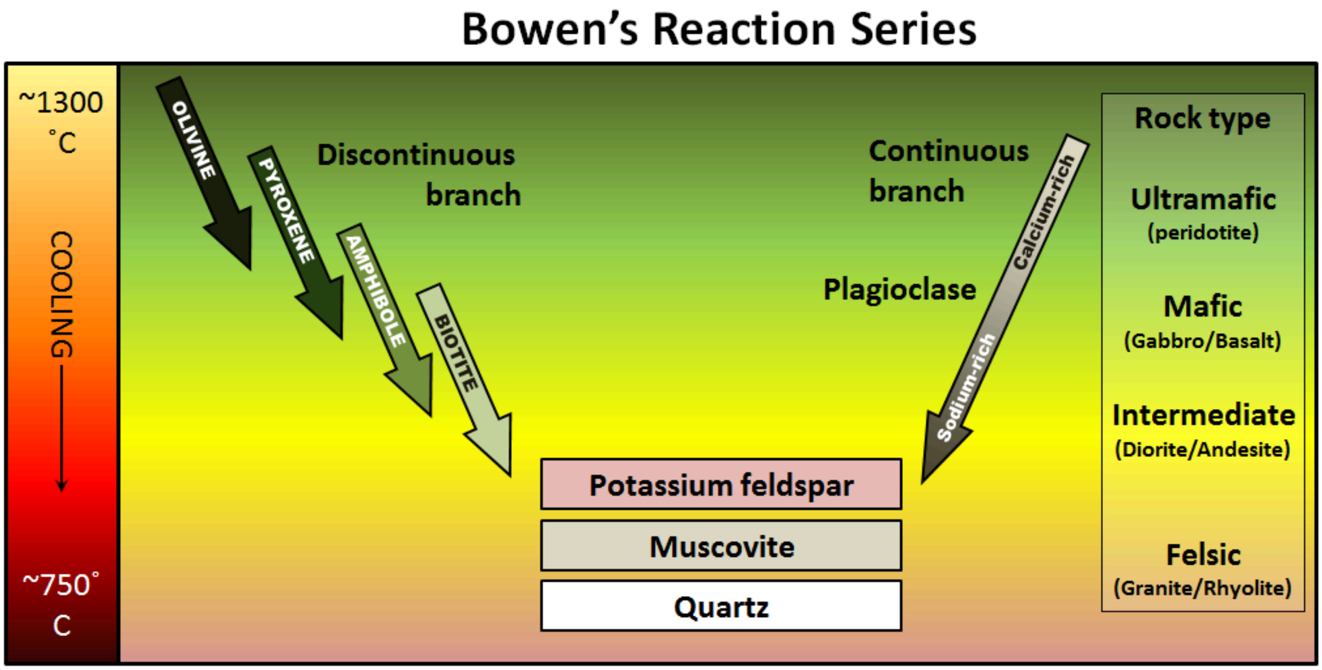

Figure 3.3.1 The Bowen reaction series describes the process of magma crystallization.

In some cases, 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 3.3.2 shows a zoned plagioclase under a microscope.

Finally, if the magma is quite silica-rich to begin with, there will still be some left at around 750° to 800°C, and from this last magma, potassium feldspar, quartz, and maybe muscovite mica will form.

Figure 3.3.2 A zoned plagioclase crystal. The central part is calcium-rich and the darker outside part is sodium-rich.

Who was Bowen, and what is a reaction series?

Figure 3.3.3

Norman Levi Bowen, born in Kingston Ontario, studied geology at Queen's University and then at MIT in Boston. In 1912, Norman Levi Bowenhe joined the Carnegie Institution in Washington, D.C., where he carried out groundbreaking experimental research into the processes of cooling magmas. Working mostly with basaltic magmas, 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 Bowen reaction series is one of the results of his work, and even a century later, it is an important basis for our understanding of igneous rocks. The word *reaction* is critical. In the discontinuous branch, olivine is typically the first mineral to form (at just below 1300°C). As the temperature continues to drop, olivine becomes unstable while pyroxene becomes stable. The early-forming olivine crystals *react* with silica in the remaining liquid magma and are converted into pyroxene, something like this:

Mg_2SiO_4 + SiO_2 (olivine) becomes $2MgSiO_3$ (proxene)

This continues down the chain, as long as there is still silica left in the liquid.

The composition of the original magma is critical to magma crystallization because it determines how far the reaction process can continue before all of the silica is used up. The compositions of typical **mafic**, intermediate, and **felsic** magmas are shown in Figure 3.3.4. Note that, unlike Figure 3.2.1, these compositions are expressed in terms of "oxides" (e.g., Al_2O_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.

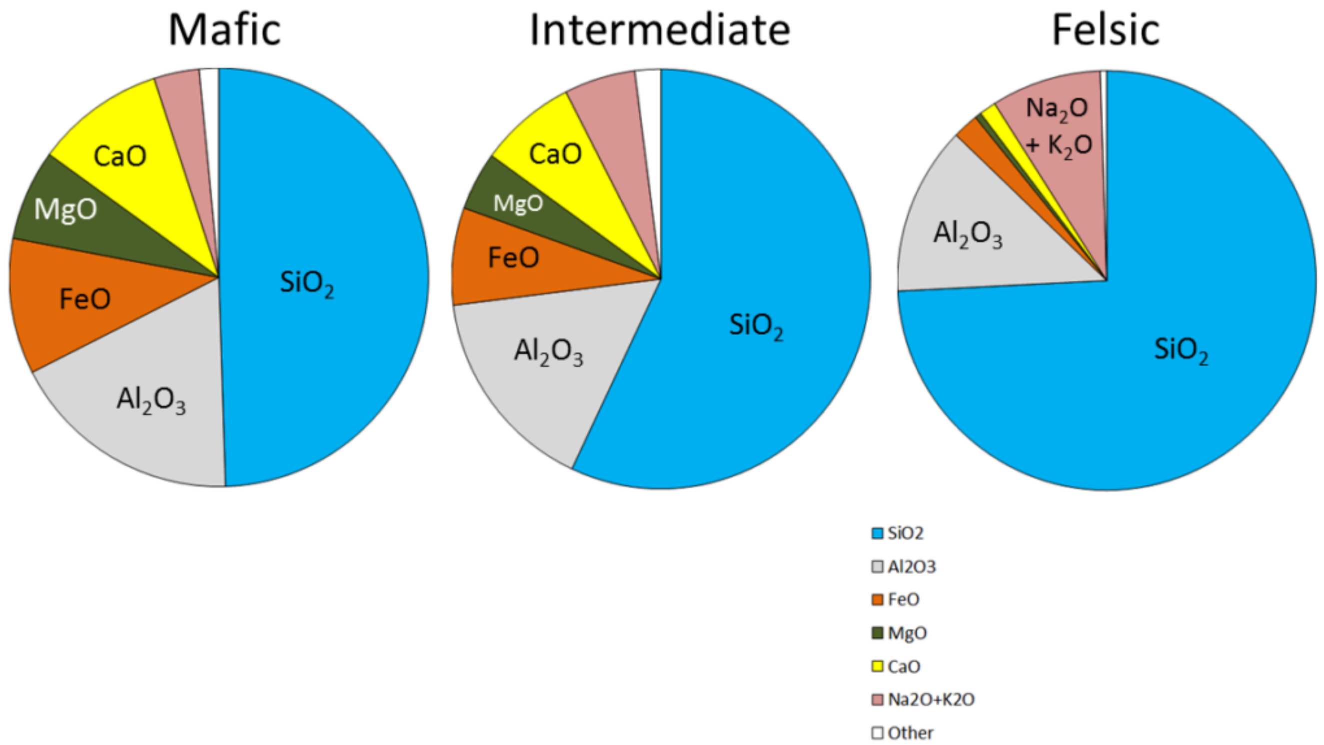

Figure 3.3.4 The chemical compositions of typical mafic, intermediate, and felsic magmas and the types of rocks that form from them.

Mafic magmas have 45% to 55% SiO_2, about 25% total of FeO and MgO plus CaO, and about 5% Na_2O + K_2O. Felsic magmas, on the other hand, have much more SiO_2 (65% to 75%) and Na_2O + K_2O (around 10%) and much less FeO and MgO plus CaO (about 5%).

Exercise 3.3 Determining rock types based on magma composition

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 3.3.4.)

Table 3.1 Proportions of the main chemical components in felsic, intermediate, and mafic magma

[Skip Table]

Oxide	Felsic Magma	Intermediate Magma	Mafic Magma
SiO_2	65% to 75%	55% to 65%	45% to 55%
Al_2O_3	12% to 16%	14% to 18%	14% to 18%
FeO	2% to 4%	4% to 8%	8% to 12%
CaO	1% to 4%	4% to 7%	7% to 11%
MgO	0% to 3%	2% to 6%	5% to 9%
Na_2O	2% to 6%	3% to 7%	1% to 3%
K_2O	3% to 5%	2% to 4%	0.5% to 3%

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 3.2 Chemical Data for Four Unidentified Rock Samples

[Skip Table]

Rock Sample	SiO_2	Al_2O_3	FeO	CaO	MgO	Na_2O	K_2O	What type of magma is it?
Rock 1	55%	17%	5%	6%	3%	4%	3%	
Rock 2	74%	14%	3%	3%	0.5%	5%	4%	
Rock 3	47%	14%	8%	10%	8%	1%	2%	
Rock 4	65%	14%	4%	5%	4%	3%	3%	

See Appendix 3 for Exercise 3.3 answers.

As a *mafic* magma starts to cool, some of the silica combines with iron and magnesium to make olivine. As it cools further, much of the remaining silica goes into calcium-rich plagioclase, and any silica left may be used to convert some of the olivine to pyroxene. Soon after that, all of the magma is used up and no further changes takes place. The minerals present will be olivine, pyroxene, and calcium-rich plagioclase. If the magma cools slowly underground, the product will be **gabbro**; if it cools quickly at the surface, the product will be **basalt** (Figure 3.3.5).

Felsic magmas tend to be cooler than mafic magmas when crystallization begins (because they don’t have to be as hot to remain liquid), and so they may start out crystallizing pyroxene (not olivine) and plagioclase. As cooling continues, the various reactions on the discontinuous branch will proceed because silica is abundant, the plagioclase will become increasingly sodium-rich, and eventually potassium feldspar and quartz will form. Commonly even very felsic rocks will not have biotite or muscovite because they may not have enough aluminum or enough hydrogen to make the OH complexes that are necessary for mica minerals. Typical felsic rocks are **granite** and **rhyolite** (Figure 3.3.5).

The cooling behaviour of intermediate magmas lie somewhere between those of mafic and felsic magmas. Typical mafic rocks are gabbro (intrusive) and basalt (extrusive). Typical intermediate rocks are **diorite** and **andesite**. Typical felsic rocks are granite and rhyolite (Figure 3.3.5).

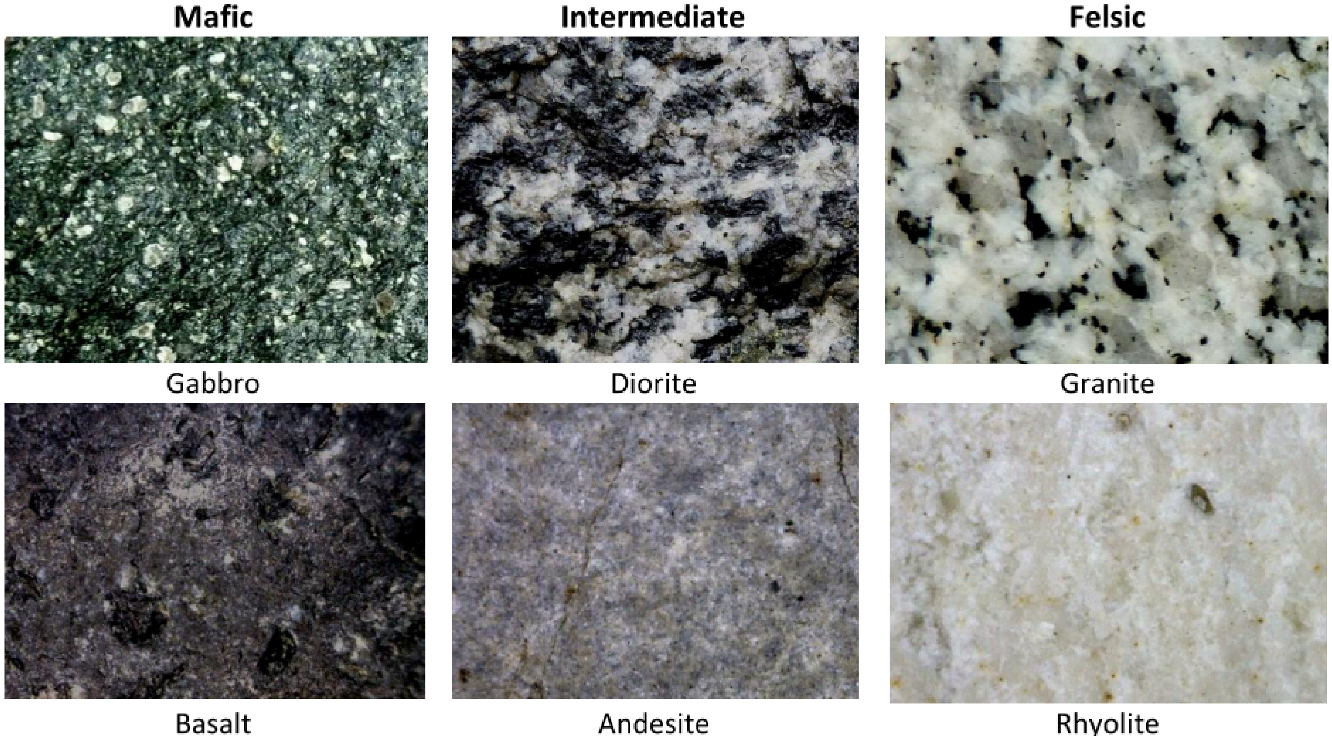

Figure 3.3.5 Examples of the igneous rocks that form from mafic, intermediate, and felsic magmas.

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 3.3.6a), may slowly settle toward the bottom of the magma chamber (Figure 3.3.6b). The means that the overall composition of the magma near the top of the magma chamber will become more felsic, as it is losing some iron- and magnesium-rich components. This process is known as **fractional crystallization**. The crystals that settle might either form an olivine-rich layer near the bottom of the magma chamber, or they might remelt because the lower part is likely to be hotter than the upper part (remember, from Chapter 1, that temperatures increase steadily with depth in Earth because of the geothermal gradient). 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 3.3.6c).

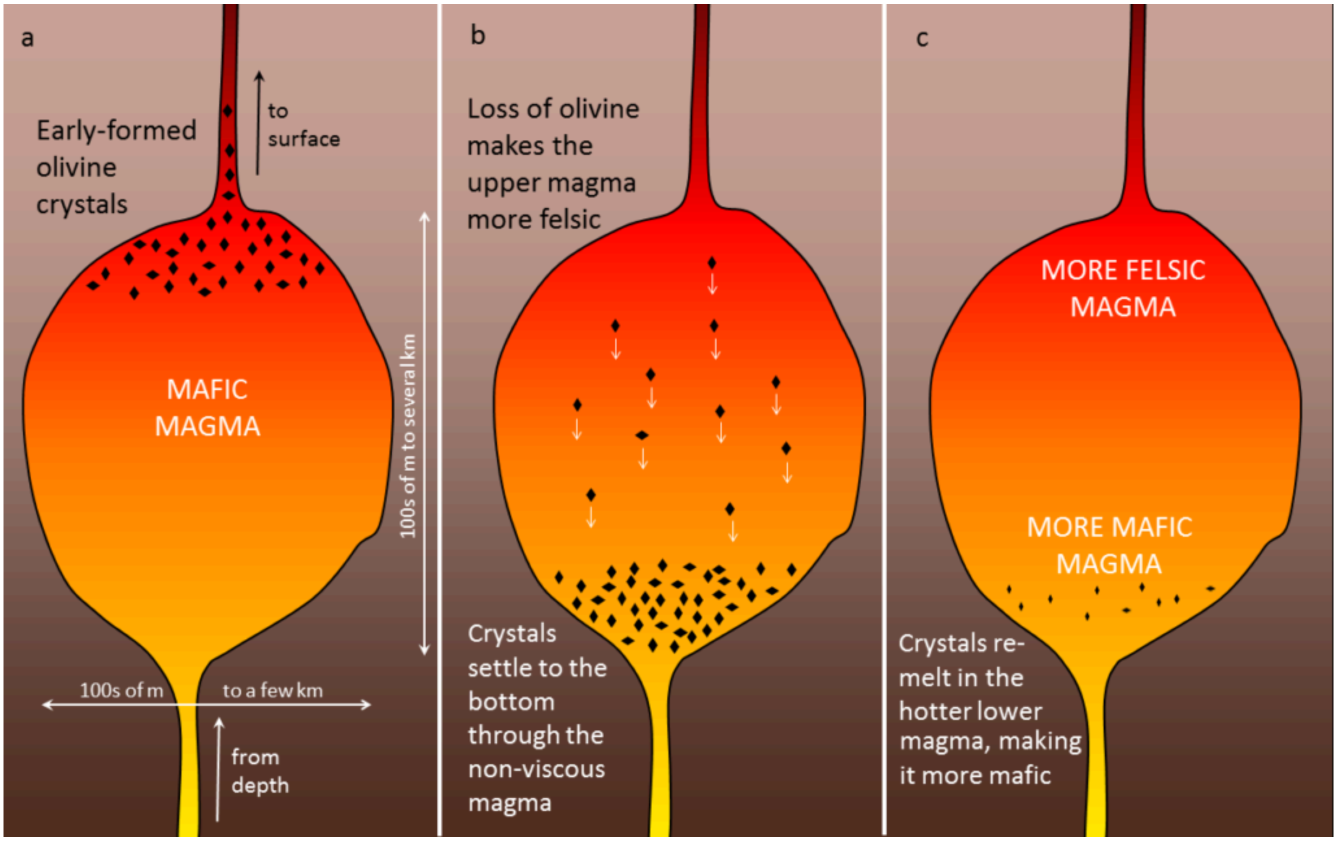

Figure 3.3.6 An example of crystal settling and the formation of a zoned magma chamber.

If crystal settling does not take place, because the magma is too viscous, then the process of cooling will continue as predicted by the Bowen reaction series. In some cases, however, partially cooled but still liquid magma, with crystals in it, will either move farther up into a cooler part of the crust, or all the way to the surface during a volcanic eruption. In either of these situations, the magma that has moved toward the surface is likely to cool much faster than it did within the magma chamber, and the rest of the rock will have a finer crystalline texture. An igneous rock with large crystals embedded in a matrix of much finer crystals is indicative of a two-stage cooling process, and the texture is **porphyritic** (Figure 3.3.7). For the rock to be called "porphyritic" there has to be a significant difference in crystal size, where the larger crystals are at least 10 times larger than the average size of the smaller crystals.

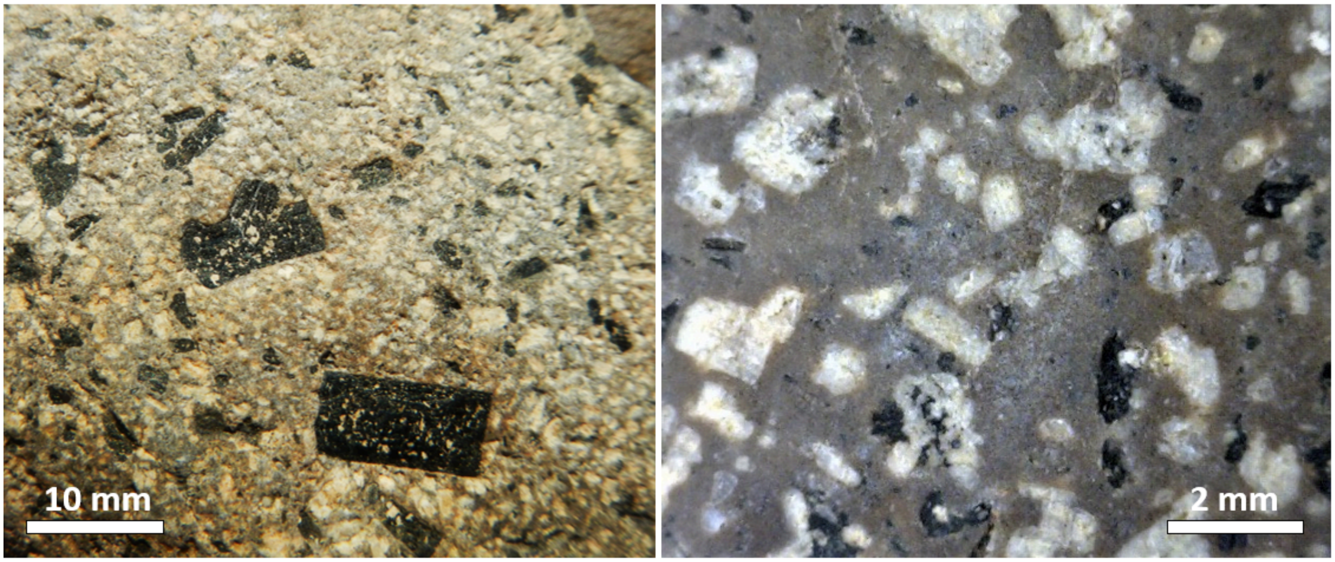

Figure 3.3.7 Porphyritic textures, left: 1.3 cm long amphibole crystals in an intrusive igneous rock in which most of the crystals are less than 1 mm, right: 1 to 2 mm long feldspar crystals and 1 mm long amphibole crystals in a volcanic rock where most of the crystals are less than 0.1 mm.

Exercise 3.4 Porphyritic minerals

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 Figure 3.3.8, predict what phenocrysts might be present where the magma cooled as far as line **a** in one case, and line **b** in another.

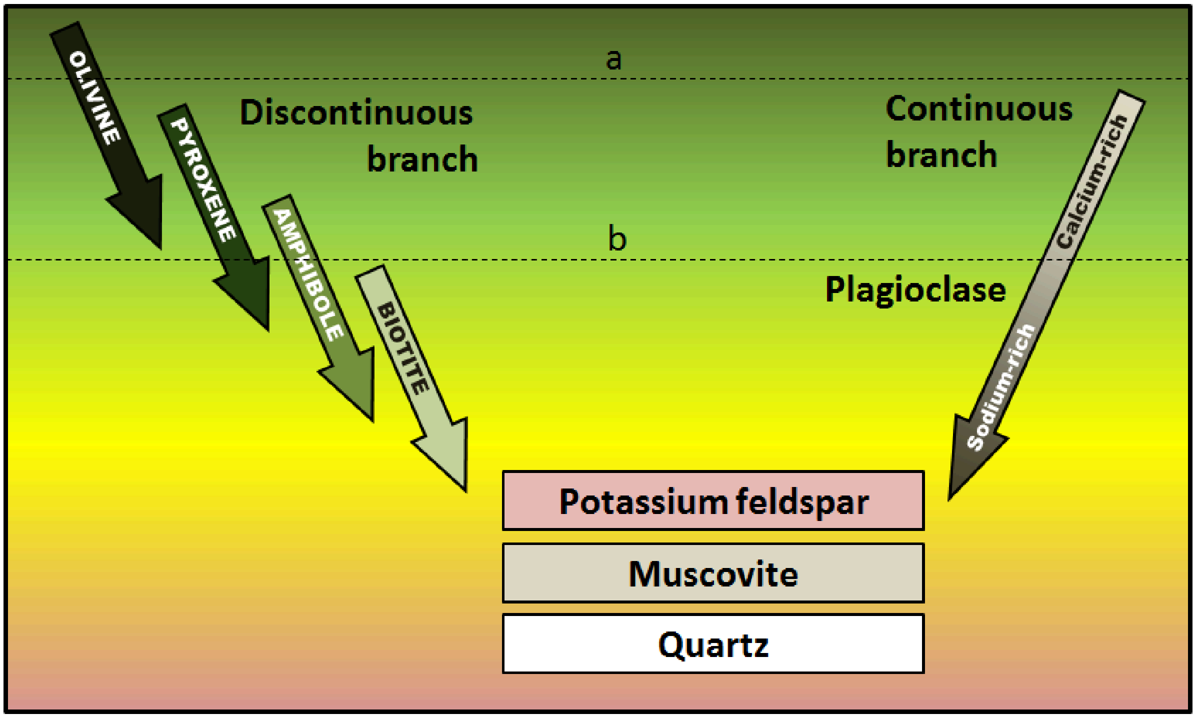

Figure 3.3.8 Bowen reaction series. Line a – at high temperature – intersects olivine, Line b – at a lower temperature – intersects pryroxene and amphibole on the left, and plagioclase feldspar on the right

See Appendix 3 for Exercise 3.4 answers.

Media Attributions

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3.4 Classification of Igneous Rock

As has already been described, igneous rocks are classified into four categories: felsic, intermediate, mafic, and ultramafic, based on either their chemistry or their mineral composition. The diagram in Figure 3.4.1 can be used to help classify igneous rocks by their mineral composition. An important feature to note on this diagram is the red line separating the non-ferromagnesian silicates in the lower left (K-feldspar, quartz, and plagioclase feldspar) from the ferromagnesian silicates in the upper right (biotite, amphibole, pyroxene, and olivine). In classifying intrusive igneous rocks, the first thing to consider is the percentage of ferromagnesian silicates. In most igneous rocks the ferromagnesian silicate minerals are clearly darker than the others, but it is still quite difficult to estimate the proportions of minerals in a rock.

Based on the position of the red line in Figure 3.4.1, it is evident that felsic rocks can have between 1% and 20% ferromagnesian silicates (the red line intersects the left side of the felsic zone 1% of the distance from the top of the diagram, and it intersects the right side of the felsic zone 20% of the distance from the top). Intermediate rocks have between 20% and 50% ferromagnesian silicates, and mafic rocks have 50% to 100% ferromagnesian silicates. To be more specific, felsic rocks typically have biotite and/or amphibole; intermediate rocks have amphibole and, in some cases, pyroxene; and mafic rocks have pyroxene and, in some cases, olivine.

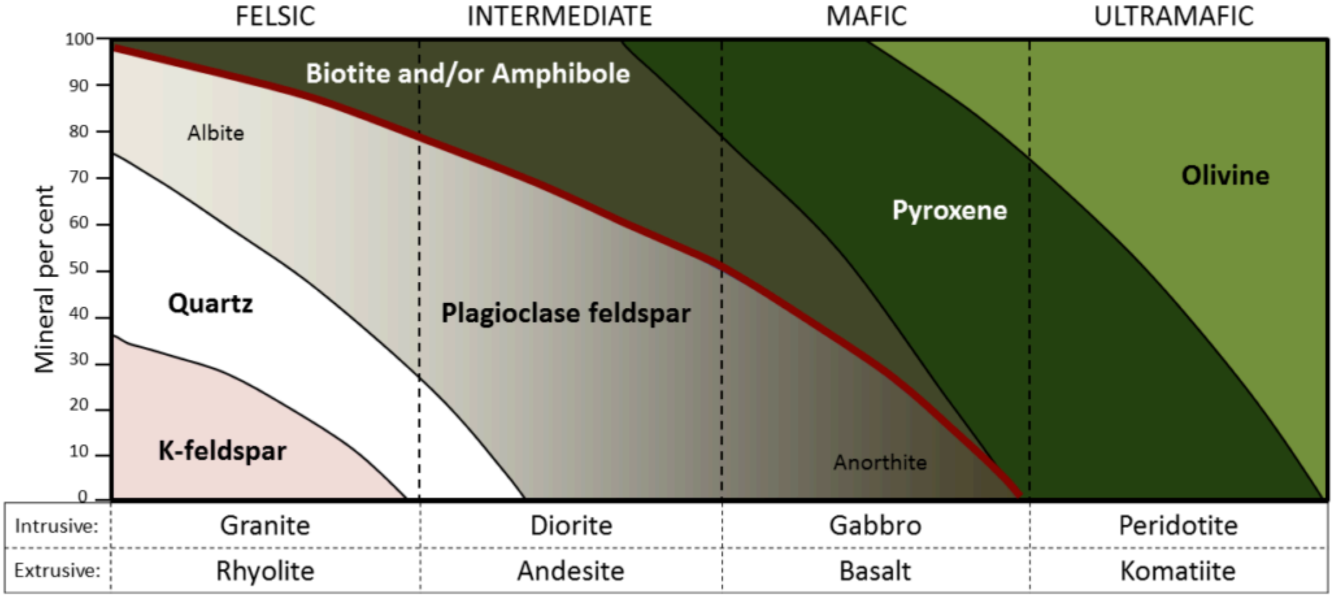


Figure 3.4.1 A simplified classification diagram for igneous rocks based on their mineral compositions. [Image Description]

If we focus on the non-ferromagnesian silicates, it is evident that felsic rocks can have from 0% to 35% K-feldspar, from 25% to 35% quartz (the vertical thickness of the quartz field varies from 25% to 35%), and from 25% to 50% plagioclase (and that plagioclase will be sodium-rich, or albitic). Intermediate rocks can have up to 25% quartz and 50% to 75% plagioclase. Mafic rocks only have plagioclase (up to 50%), and that plagioclase will be calcium-rich, or anorthitic.

Exercise 3.5 Mineral proportions in igneous rocks

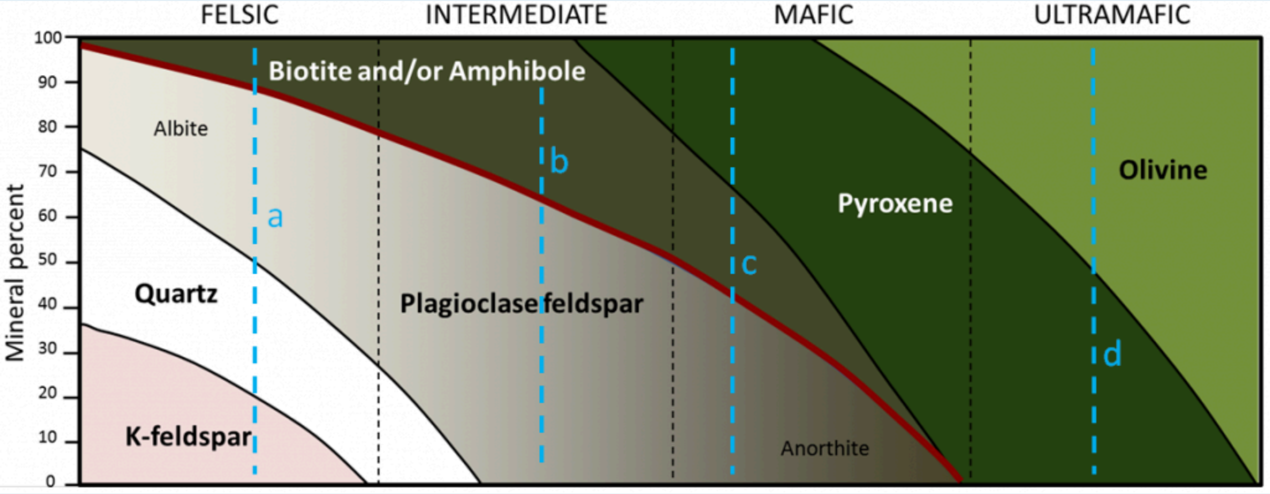


Figure 3.4.2

The dashed blue lines (labelled a, b, c, d) in Figure 3.4.2 represent four igneous rocks. Complete the table by estimating the mineral proportions (percent) of the four rocks (to the nearest 10%).

Hint: Rocks **b** and **d** are the easiest; start with those.

Rock	Biotite/amphibole	Pyroxene	Olivine	Plagioclase	Quartz	K-feldspar
a						
b						
c						
d						

See Appendix 3 for Exercise 3.5 answers.

Figure 3.4.3 provides a diagrammatic representation of the proportions of dark minerals in light-coloured rocks. You can use that when trying to estimate the ferromagnesian mineral content of actual rocks, and you can get some practice doing that by completing Exercise 3.6. Be warned! Geology students almost universally over-estimate the proportion of dark minerals.

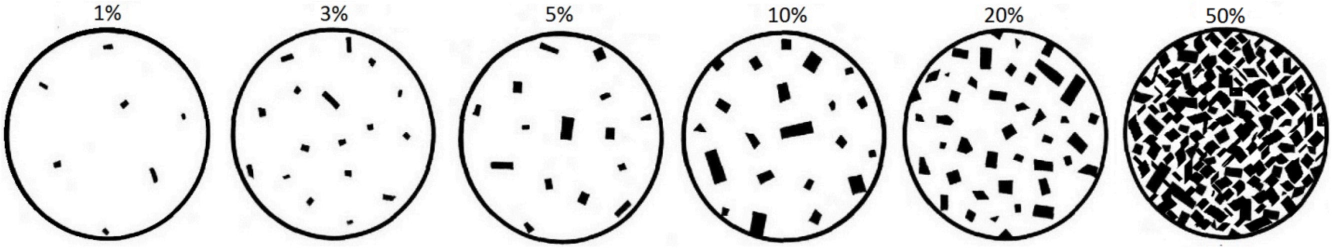
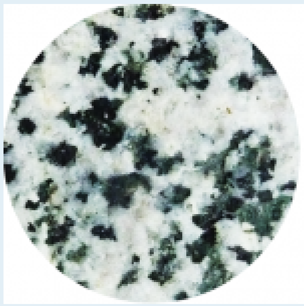
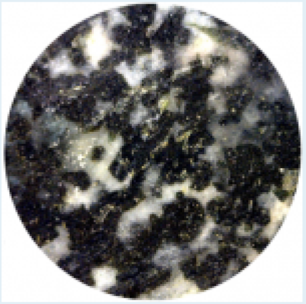


Figure 3.4.3 A guide to estimating the proportions of dark minerals in light-coloured rocks.

Exercise 3.6 Proportions of ferromagnesian silicates

The four igneous rocks shown below have differing proportions of ferromagnesian silicates. Estimate those proportions using the diagrams in Figure 3.4.3, and then use Figure 3.4.1 to determine the likely rock name for each one.

			
___%	___%	___%	___%
__________	__________	__________	__________

See Appendix 3 for Exercise 3.6 answers.

Igneous rocks are also classified according to their textures. The textures of volcanic rocks will be discussed in Chapter 4, so here we'll only look at the different textures of intrusive igneous rocks. Almost all intrusive igneous rocks have crystals that are large enough to see with the naked eye, and we use the term **phaneritic** (from the Greek word phaneros meaning visible) to describe that. Typically that means they are larger than about 0.5 millitmeres (mm) — the thickness of a strong line made with a ballpoint pen. (If the crystals are too small to distinguish, which is typical of most volcanic rocks, we use the term **aphanitic** (from the Greek word *aphanos* – unseen) The intrusive rocks shown in Figure 3.3.5 are all phaneritic, as are those shown in Exercise 3.6.

In general, the size of crystals is proportional to the rate of cooling. The longer it takes for a body of magma to cool, the larger the crystals can grow. It is not uncommon to see an intrusive igneous rock with crystals up to 1 centimetre (cm) long. In some situations, especially toward the end of the cooling stage, the magma can become water rich. The presence of liquid water (still liquid at high temperatures because it is under pressure) promotes the relatively easy movement of ions, and this allows crystals to grow large, sometimes to several centimetres (Figure 3.4.4). Finally, as already described, if an igneous rock goes through a two-stage cooling process, its texture will be porphyritic (Figure 3.3.7).

Figure 3.4.4 A pegmatitic rock with large crystals

Image Descriptions

Figure 3.4.1 image description: Mineral composition of igneous rocks

Igneous Rocks	**Felsic**	**Intermediate**	**Mafic**	**Ultramafic**
K-feldspar	0 to 35%	0%	0%	0%
Quartz	25 to 35%	0 to 25%	0%	0%
Plagioclase feldspar	25 to 50%	50 to 70%	0 to 50%	0%
Biotite and/or Amphibole	0 to 20%	20 to 40%	0 to 30%	0%
Pyroxene	0%	0 to 20%	20 to 75%	0% to 75%
Olivine	0%	0%	0 to 25 %	25% to 100%
Intrusive	Granite	Diorite	Gabbro	Peridotite
Extrusive	Rhyolite	Andesite	Basalt	Komatiite

[Return to Figure 3.4.1]

Attributions

- Figure 3.4.1, 3.4.2, 3.4.3: © Steven Earle. CC BY.
- Figure 3.4.4: Pegmatite. Public domain.

3.5 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 move very slowly up toward the surface. It does so in a few different ways, including filling and widening existing cracks, melting the surrounding rock (called **country rock**[1]**)**, pushing the rock aside (where it is somewhat plastic), and breaking the rock. Where some of the country rock is broken off, it may fall into the magma, a process called **stoping**. The resulting fragments, illustrated in Figure 3.5.1, are known as **xenoliths** (Greek for "strange rocks").

Figure 3.5.1 Xenoliths of mafic rock in granite, Victoria, B.C. The fragments of dark rock have been broken off and incorporated into the light-coloured granite.

Some upward-moving magma reaches the surface, resulting in volcanic eruptions, but most cools within the crust. The resulting body of rock is known as a **pluton**. Plutons can have various different shapes and relationships to the surrounding country rock as shown in Figure 3.5.2.

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[1] "Country rock" is not necessarily music to a geologist's ears. The term refers to the original "rock of the country" or region, and hence the rock into which the magma intruded to form a pluton.

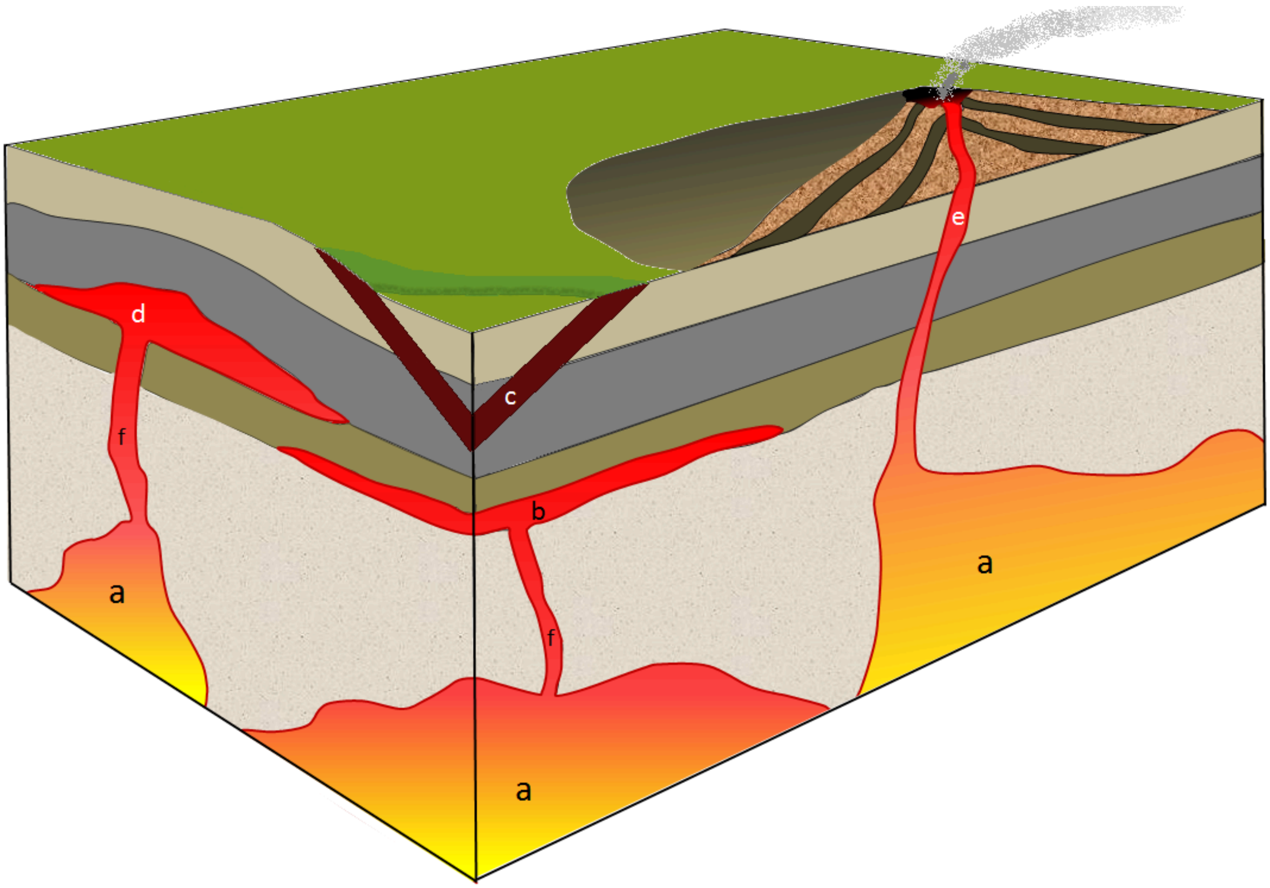

Figure 3.5.2 Depiction of some of the types of plutons. a: stocks (if they coalesce at depth then they might become large enough to be called a batholith), b: sill (a tabular body, in this case parallel to bedding), c: dyke (cross-cuts bedding), d: laccolith (a sill that has pushed up the overlying rock layers), e: pipe (a cylindrical conduit feeding a volcano). The two features labelled f could be pipes or dykes, but from this perspective it's not possible to determine if they are cylindrical or tabular.

Large irregular-shaped plutons are called either **stocks** or **batholiths**. The distinction between the two is made on the basis of the area that is exposed at the surface: if the body has an exposed surface area greater than 100 square kilometres (km^2), then it's a batholith; smaller than 100 km^2 and it's a stock. Batholiths are typically formed only 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, which extends all the way from the Vancouver region to southeastern Alaska (Figure 3.5.3). More accurately, it's many batholiths.

Tabular (sheet-like) plutons are distinguished on the basis of whether or not they are **concordant** with (i.e., parallel to) existing layering (e.g., sedimentary bedding or metamorphic foliation) in the country rock. A **sill** is concordant with existing layering, and a **dyke** is **discordant**. If the country rock has no bedding or foliation, then any tabular body within it is a dyke. Note that the sill-versus-dyke designation is not determined simply by the orientation of the feature. A dyke can be horizontal and a sill can be vertical (if the bedding is vertical). A large dyke can be seen in Figure 3.5.3.

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

Finally, a **pipe** is a cylindrical body (with a circular, ellipitical, or even irregular cross-section) that served as a conduit for the movement of magma from one location to another. Most known pipes fed volcanoes, although pipes can also connect plutons. It is also possible for a dyke to feed a volcano.

Figure 3.5.3 The Stawamus Chief, part of the Coast Range Plutonic Complex, near to Squamish, B.C. The cliff is about 600 metres (m) high. Most of the dark stripes are a result of algae and lichen growth where the surface is frequently wet, but there is a large (about 10 m across) vertical dyke that extends from bottom to top.

As discussed already, plutons can interact with the rocks into which they are intruded, sometimes leading to partial melting of the country rock or to stoping and formation of xenoliths. And, as we'll see in Chapter 7, the heat of a body of magma can lead to metamorphism of the country rock. The country rock can also have an effect on the magma within a pluton. The most obvious such effect is the formation of a chilled margin along the edges of the pluton, where it came in contact with country rock that was significantly colder than the magma. Within the chilled margin, the magma cooled more quickly than in the centre of the dyke, so the texture is finer and the colour may be different. An example is shown in Figure 3.5.4.

Figure 3.5.4 A mafic dyke with chilled margins within basalt at Nanoose, B.C. The coin is 24 mm in diameter. The dyke is about 25 centimetres (cm) across and the chilled margins are 2 cm wide.

Exercise 3.7 Pluton Problems

Figure 3.5.5 shows a cross-section through part of the crust showing a variety of intrusive igneous rocks. Except for the granite (a), all of these rocks are mafic in composition. Indicate whether each of the plutons labelled **a** to **e** on the diagram below is a **dyke**, a **sill,** a **stock**, or a **batholith**.

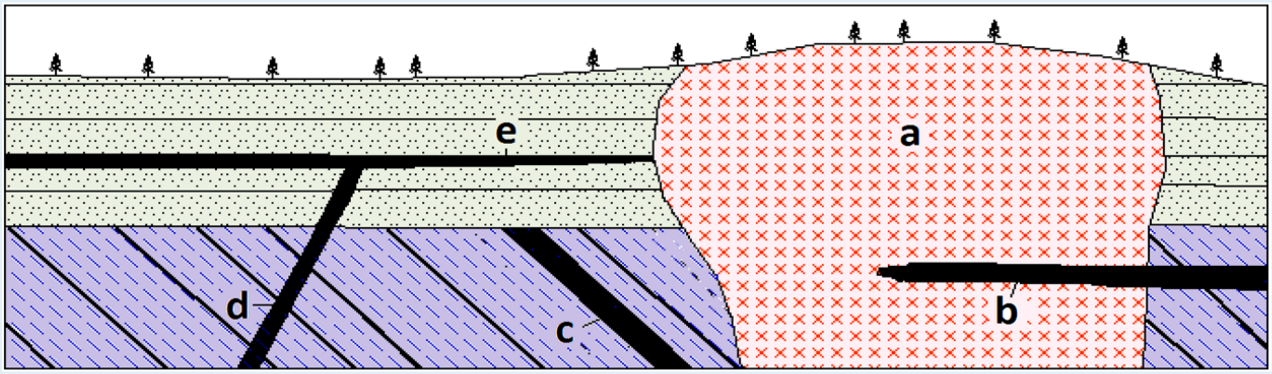

Figure 3.5.5

See Appendix 3 for Exercise 3.7 answers.

Media Attributions

- Figures 3.5.1, 3.5.2, 3.5.3, 3.5.4, 3.5.5: © Steven Earle. CC BY.

Summary

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

Section	**Summary**
3.1 The Rock Cycle	The three types of rocks are igneous: formed from magma; sedimentary: formed from fragments of other rocks or precipitation from solution; and metamorphic: formed when existing rocks are altered by heat, pressure, and/or chemical action. The rock cycle summarizes the processes that contribute to cycling of rock material among these three types. The rock cycle is driven by Earth’s internal heat, and by processes happening at the surface, which are driven by solar energy.
3.2 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. Intermediate rocks have compositions between felsic and mafic.
3.3 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.
3.4 Classification of Igneous Rock	Igneous rocks are classified based on their mineral composition and texture. Felsic igneous rocks have less than 20% ferromagnesian silicates (amphibole and/or biotite) plus varying amounts of quartz and both potassium and plagioclase feldspars. Mafic igneous rocks have more than 50% ferromagnesian silicates (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). If the cooling was extremely slow, or if water was present during cooling, the texture may be pegmatitic (very large crystals).
3.5 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 (dykes 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

Answers to Review Questions at the end of each chapter are provided in Appendix 2.

1. What processes must take place to transform rocks into sediment?
2. What processes normally take place in the transformation of sediments to sedimentary rock?
3. What are the processes that lead to the formation of a metamorphic rock?
4. What is the significance of the term *reaction* in the name of the Bowen reaction series?
5. Why is it common for plagioclase crystals to be zoned from relatively calcium-rich in the middle to more sodium-rich on the outside?
6. What must happen within a magma chamber for fractional crystallization to take place?
7. Explain the difference between aphanitic and phaneritic textures.
8. Explain the difference between porphyritic and pegmatitic textures.
9. Name the following rocks:
    1. An extrusive rock with 40% Ca-rich plagioclase and 60% pyroxene
    2. An intrusive rock with 65% plagioclase, 25% amphibole, and 10% pyroxene
    3. An intrusive rock with 25% quartz, 20% orthoclase, 50% feldspar, and minor amounts of biotit
10. With respect to tabular intrusive bodies, what is the difference between a concordant body and a discordant body?
11. Why does a dyke commonly have a fine-grained margin?
12. What is the difference between a batholith and a stock?
13. Describe two ways in which batholiths intrude into existing rock.
14. Why is compositional layering a common feature of mafic plutons but not of felsic plutons?