

# Physical Geology - 2nd Edition

# Physical Geology - 2nd Edition

Steven Earle

Karla Panchuk

BCCAMPUS  
VICTORIA, B.C.



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*Mt. Meager—just right of centre—with Plinth Peak to its left and Pylon Peak to the left of that (partly obscured by cloud). Mt. Meager was the site of a major volcanic eruption about 2400 years ago (see Chapter 4). Plinth and Pylon Peaks are also volcanoes, but are now extinct. The bare patch extending down from the ridge to the right of Mt. Meager is the source area of the massive 2010 rock avalanche that is illustrated and described in Chapter 15. The stream in the foreground is flowing from the Magic Carpet Glacier. Capricorn Glacier can be seen to the right of Pylon Peak.*

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---

## Preface

This book was born out of a 2014 meeting of earth science educators representing most of the universities and colleges in British Columbia, and nurtured by a widely shared frustration that many students are not thriving in our courses because textbooks have become too expensive for them to buy. But the real inspiration comes from a fascination for the spectacular geology of western Canada and the many decades that I have spent exploring this region along with colleagues, students, family, and friends. My goal has been to provide an accessible and comprehensive guide to the important topics of geology, richly illustrated with examples from western Canada. Although this text is intended to complement a typical first-year course in physical geology, its contents could be applied to numerous other related courses.

As a teacher for many years, and as someone who is constantly striving to discover new things, I am well aware of that people learn in myriad ways, and that for most, simply reading the contents of a book is not one of the most effective ones. For that reason, this book includes numerous embedded exercises and activities that are designed to encourage readers to engage with the concepts presented, and to make meaning of the material under consideration. It is strongly recommended that you try the exercises as you progress through each chapter. You should also find it useful, whether or not assigned by your instructor, to complete the questions at the end of each chapter.

Over many years of teaching earth science I have received a lot of feedback from students. What gives me the most pleasure is to hear that someone, having completed my course, now sees Earth with new eyes, and has discovered both the thrill and the value of an enhanced understanding of how our planet works. I sincerely hope that this textbook will help you see Earth in a new way.

Steven Earle, Gabriola Island, 2015

# Chapter 1 Introduction to Geology

## 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 what geology is, how it incorporates the other sciences, and how it is different from the other sciences.
- Discuss why we study Earth and what type of work geologists do.
- Define some of the properties of a mineral and explain the differences between minerals and rocks.
- Describe the nature of Earth's interior and some of the processes that take place deep beneath our feet.
- Explain how those processes are related to plate tectonics and describe a few of the features that are characteristic of plate boundaries.
- Use the notation for geological time, gain an appreciation for the vastness of geological time, and describe how very slow geological processes can have enormous impacts over time.

## 1.1 What is Geology?

In its broadest sense, geology is the study of Earth—its interior and its exterior surface, the minerals, rocks and other materials that are around us, the processes that have resulted in the formation of those materials, the water that flows over the surface and through the ground, the changes that have taken place over the vastness of geological time, and the changes that we can anticipate will take place in the near future. Geology is a science, meaning that we use deductive reasoning and scientific methods to understand geological problems. It is, arguably, the most integrated of all of the sciences because it involves the understanding and application of all of the other sciences: physics, chemistry, biology, mathematics, astronomy, and others. But unlike most of the other sciences, geology has an extra dimension, that of time—deep time—billions of years of it. Geologists study the evidence that they see around them, but in most cases, they are observing the results of processes that happened thousands, millions, and even billions of years in the past. Those were processes that took place at incredibly slow rates—millimetres per year to centimetres per year—but because of the amount of time available, they produced massive results.

Geology is displayed on a grand scale in mountainous regions, perhaps nowhere better than the Rocky Mountains in Canada (Figure 1.1.1). The peak on the right is Rearguard Mountain, which is a few kilometres northeast of Mount Robson, the tallest peak in the Canadian Rockies (3,954 metres). The large glacier in the middle of the photo is the Robson Glacier. The river flowing from Robson Glacier drains into Berg Lake in the bottom right. There are many geological features portrayed here. The sedimentary rock that these mountains are made of formed in ocean water over 500 million years ago. A few hundred million years later, these beds were pushed east for tens to hundreds of kilometres by tectonic plate convergence and also pushed up to thousands of metres above sea level. Over the past two million years this area—like most of the rest of Canada—has been repeatedly glaciated, and the erosional effects of those glaciations are obvious.

The Robson Glacier is now only a small remnant of its size during the Little Ice Age of the 15th to 18th centuries, and even a lot smaller than it was just over a century ago in 1908. The distinctive line on the slope on the left side of both photos shows the elevation of the edge of the glacier a few hundred years ago. Like almost all other glaciers in the world, it receded after the 18th century because of natural climate change, is now receding even more rapidly because of human-caused climate change.



**Figure 1.1.1** Rearguard Mountain and Robson Glacier in Mount Robson Provincial Park, BC. Left: Robson Glacier in 2012. Right: Robson Glacier circa 1908.

Geology is also about understanding the evolution of life on Earth; about discovering resources such as water, metals and energy; about recognizing and minimizing the environmental implications of our use of those resources; and about learning how to mitigate the hazards related to earthquakes, volcanic eruptions, and slope failures. All of these aspects of geology, and many more, are covered in this textbook.

### What are scientific methods?

There is no single method of inquiry that is specifically the “scientific method”; furthermore, scientific inquiry is not necessarily different from serious research in other disciplines. The most important thing that those involved in any type of inquiry must do is to be skeptical. As the physicist Richard Feynman once said: the first principle of science is that “you must not fool yourself—and you are the easiest person to fool.” A key feature of serious inquiry is the creation of a hypothesis (a tentative explanation) that could explain the observations that have been made, and then the formulation and testing (by experimentation) of one or more predictions that follow from that hypothesis.

For example, we might observe that most of the cobbles in a stream bed are well rounded (see photo above), and then derive the hypothesis that the rocks are rounded by transportation along the stream bed. A prediction that follows from this hypothesis is that cobbles present in a stream will become increasingly rounded as they are transported downstream. An experiment to test this prediction would be to place some angular cobbles in a stream, label them so that we



**Figure 1.1.2**

can be sure to find them again later, and then return at various time intervals (over a period of years) to carefully measure their locations and roundness.

A critical feature of a good hypothesis and any resulting predictions is that they must be testable. For example, an alternative hypothesis to the one above is that an extraterrestrial organization creates rounded cobbles and places them in streams when nobody is looking. This may indeed be the case, but there is no practical way to test this hypothesis. Most importantly, there is no way to prove that it is false, because if we aren't able to catch the aliens at work, we still won't know if they did it!

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## 1.2 Why Study Earth?

The simple answer to this question is that Earth is our home—our only home for the foreseeable future—and in order to ensure that it continues to be a great place to live, we need to understand how it works. Another answer is that some of us can't help but study it because it's fascinating. But there is more to it than that:

- We rely on Earth for valuable resources such as soil, water, metals, industrial minerals, and energy, and we need to know how to find these resources and exploit them sustainably.
- We can study rocks and the fossils they contain to understand the evolution of our environment and the life within it.
- We can learn to minimize our risks from earthquakes, volcanoes, slope failures, and damaging storms.
- We can learn how and why Earth's climate has changed naturally in the past, and use that knowledge to understand both natural and human-caused climate change.
- We can recognize how our activities have altered the environment in many ways and the climate in increasingly serious ways, and how to avoid more severe changes in the future.
- We can use our knowledge of Earth to understand other planets in our solar system, as well as those around distant stars.

An example of the importance of geological studies for minimizing risks to the public is illustrated in Figure 1.2.1. This is a slope failure that took place in January 2005 in the Riverside Drive area of North Vancouver. The steep bank beneath the house shown gave way, and a slurry of mud and sand flowed down, destroying another house below and killing one person. This event took place following a heavy rainfall, which is a common occurrence in southwestern B.C. in the winter.



Figure 1.2.1 The aftermath of a deadly debris flow in the Riverside Drive area of North Vancouver in January 2005.

The irony of the 2005 slope failure is that the District of North Vancouver had been warned in a geological report written in 1980 that this area was prone to slope failure and that steps should be taken to minimize the risk to residents. Very little was done in the intervening 25 years, and the consequences of that were deadly.

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## 1.3 What Do Geologists Do?

Geologists are involved in a range of widely varying occupations with one thing in common: the privilege and responsibility of studying this fascinating planet. In Canada, many geologists work in the resource industries, including mineral exploration and mining and energy exploration and extraction. Other major areas where geologists work include hazard assessment and mitigation (e.g., assessment of risks from slope failures, earthquakes, and volcanic eruptions); water supply planning, development, and management; waste management; and assessment of geological issues in the forest industry, and on construction projects such as highways, tunnels, and bridges. Most geologists are employed in the private sector, but many work for government-funded geological organizations, such as the Geological Survey of Canada or one of the provincial geological surveys. And of course, many geologists are involved in education at the secondary and the post-secondary levels.

Some people are attracted to geology because they like to be outdoors, and it is true that many geological opportunities involve fieldwork in places that are as amazing to see as they are interesting to study. But a lot of geological work is also done in offices or laboratories. Geological work tends to be varied and challenging, and for these reasons and many others, geologists are among those who are the most satisfied with their employment.



Figure 1.3.1 Geologists examining ash-layer deposits at Kilauea Volcano, Hawaii.

In Canada, most working geologists are required to be registered with an association of professional geoscientists. This typically involves meeting specific post-secondary educational standards and gaining

several years of relevant professional experience under the supervision of a registered geoscientist. More information can be found at [Engineers and Geoscientists British Columbia](#).

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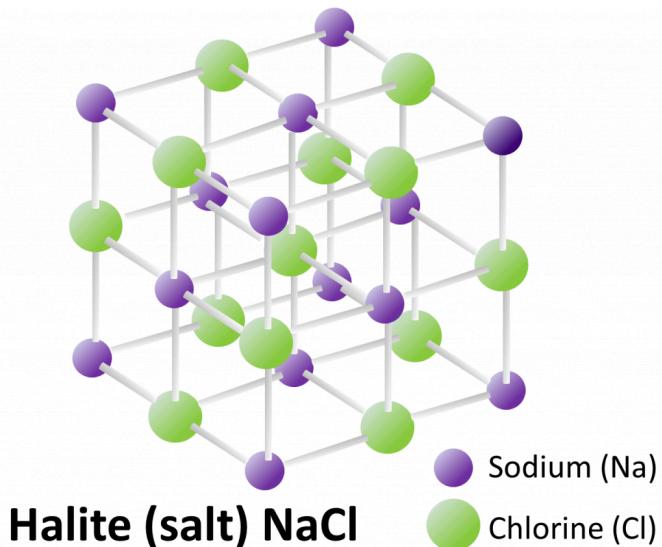
## 1.4 Minerals and Rocks

The rest of this chapter is devoted to a brief overview of a few of the important aspects of physical geology, starting with minerals and rocks. This is followed by a review of Earth's internal structure and the processes of plate tectonics, and an explanation of geological time.

The Earth is made up of varying proportions of the 90 naturally occurring elements—hydrogen, carbon, oxygen, magnesium, silicon, iron, and so on. In most geological materials, these combine in various ways to make minerals. Minerals will be covered in some detail in Chapter 2, but here we will briefly touch on what minerals are, and how they are related to rocks.

A mineral is a naturally occurring combination of specific elements that are arranged in a particular repeating three-dimensional structure or **lattice**.<sup>1</sup> The mineral **halite** is shown as an example in Figure 1.4.1.

In this case, atoms of sodium (Na: purple) alternate with atoms of chlorine (Cl: green) in all three dimensions, and the angles between the bonds are all 90°. Even in a tiny crystal, like the ones in your salt shaker, the lattices extend in all three directions for thousands of repetitions. Halite always has this composition and this structure.

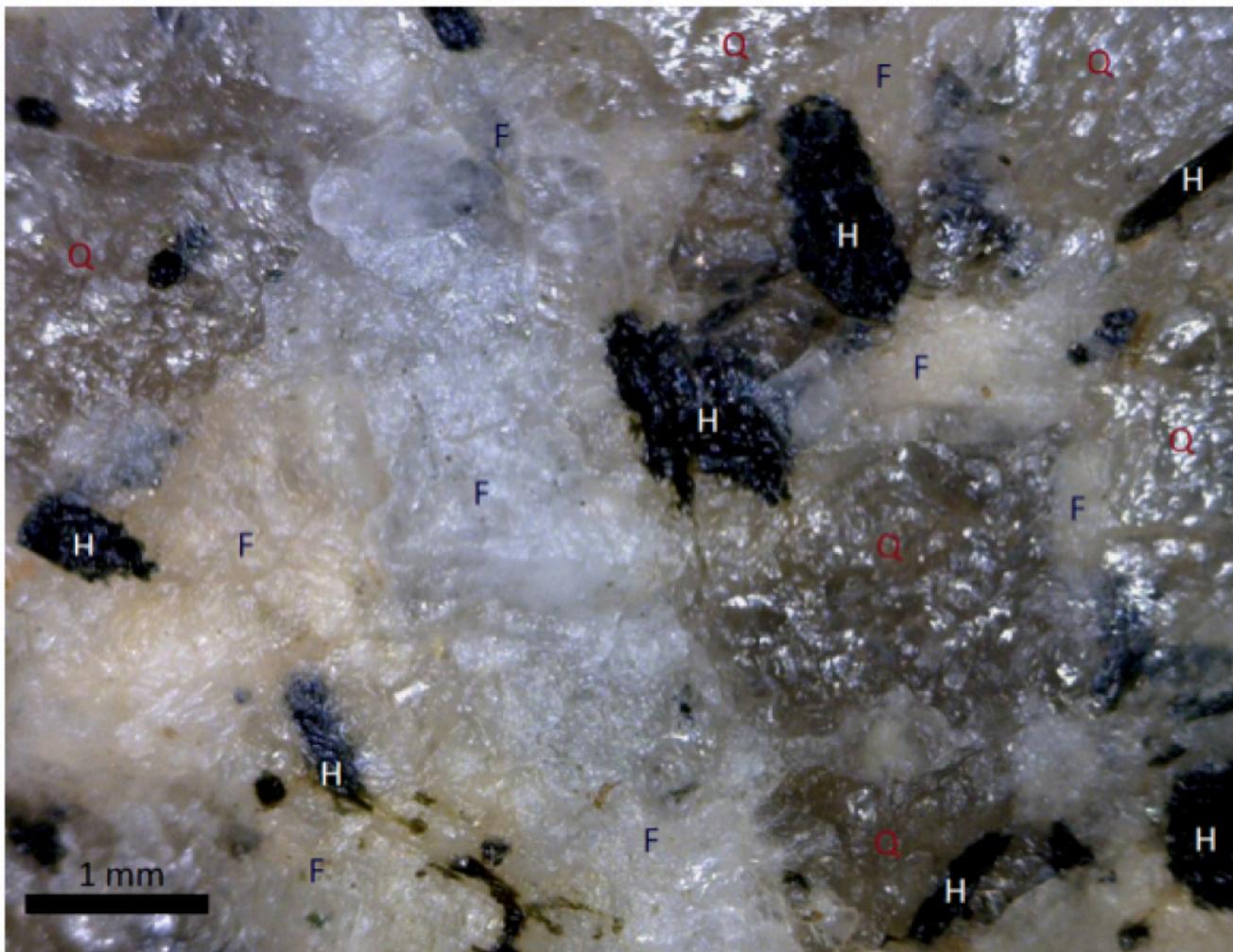


*Figure 1.4.1 The lattice structure and composition of the mineral halite (common table salt).*

**Note:** Element symbols (e.g., Na and Cl) are used extensively in this book. In [Appendix 1](#), you will find a list of the symbols and names of the elements common in minerals and a copy of the periodic table. Please use those resources if you are not familiar with the element symbols.

There are thousands of minerals, although only a few dozen are mentioned in this book. In nature, minerals are found in rocks, and the vast majority of rocks are composed of at least a few different minerals. A close-up view of **granite**, a common rock, is shown in Figure 1.4.2. Although a hand-sized piece of granite may have thousands of individual mineral crystals in it, there are typically only a few different minerals, as shown here.

1. Terms in bold are defined in the glossary at the end of the book.



Hornblende (amphibole)



Quartz



Plagioclase feldspar

Figure 1.4.2 A close-up view of the rock granite and some of the minerals that it typically contains (H = hornblende (amphibole), Q = quartz and F = feldspar). The crystals range from about 0.1 to 3 millimetres (mm) in diameter. Most are irregular in outline, but some are rectangular.

Rocks can form in a variety of ways. Igneous rocks form from **magma** (molten rock) that has either cooled slowly underground (e.g., to produce granite) or cooled quickly at the surface after a volcanic eruption (e.g., **basalt**). Sedimentary rocks, such as **sandstone**, form when the weathered products of other rocks accumulate at the surface and are then buried by other sediments. Metamorphic rocks form when either igneous or sedimentary rocks are heated and squeezed to the point where some of their minerals are unstable and new minerals form to create a different type of rock. An example is **schist**.

## 11 Chapter 1 Introduction to Geology

A critical point to remember is the difference between a mineral and a rock. A mineral is a pure substance with a specific composition and structure, while a rock is typically a mixture of several different minerals (although a few types of rock may include only one type of mineral). Examples of minerals are feldspar, quartz, mica, halite, calcite, and amphibole. Examples of rocks are granite, basalt, sandstone, limestone, and schist.

Key Takeaway: Know the difference between minerals and rocks!

If you are currently taking a geology course, you'll likely be asked more than once to name a mineral or a rock that has specific characteristics or composition, or was formed in a specific environment. Please make sure that if you're asked for a **rock name** that you don't respond with a **mineral name**, and *vice versa*. Confusing minerals and rocks is one of the most common mistakes that geology students make.

### Exercise 1.1 Find a piece of granite

The rock granite is very common in most parts of North America, and unless everything is currently covered with snow where you live, you should have no trouble finding a sample of it near you. The best places to look are pebbly ocean or lake beaches, a gravel bar of a creek or river, a gravel driveway, or somewhere where gravel has been used in landscaping. In Figure 1.4.3, taken on a beach, the granitic pebbles are the ones that are predominantly light-coloured with dark specks. The one in the very centre is a good example.



Figure 1.4.3

Select a sample of granite and, referring to Figure 1.4.2, see if you can identify some of the minerals in it. It may help to break it in half with a hammer to see a fresh surface, but be careful to protect your eyes if you do so. You should be able to see glassy-looking quartz, dull white plagioclase feldspar (and maybe pink potassium feldspar), and black hornblende or, in some cases, flaky black biotite mica (or both).

In addition to identifying the minerals in your granite, you might also try to describe the texture in terms of the range sizes of the mineral crystals (in millimetres) and the shapes of the crystals (some may be rectangular in outline, most will be irregular). Think about where your granite might have come from and how it got to where you found it.

See Appendix 3 for [Exercise 1.1 answers](#).

#### Media Attributions

- Figure 1.4.1, 1.4.2, 1.4.3: © Steven Earle. CC BY.

## 1.5 Fundamentals of Plate Tectonics

**Plate tectonics** is the model or theory that has been used for the past 60 years to understand and explain how the Earth works—more specifically the origins of continents and oceans, of folded rocks and mountain ranges, of earthquakes and volcanoes, and of continental drift. Plate tectonics is explained in some detail in Chapter 10, but is introduced here because it includes concepts that are important to many of the topics covered in the next few chapters.

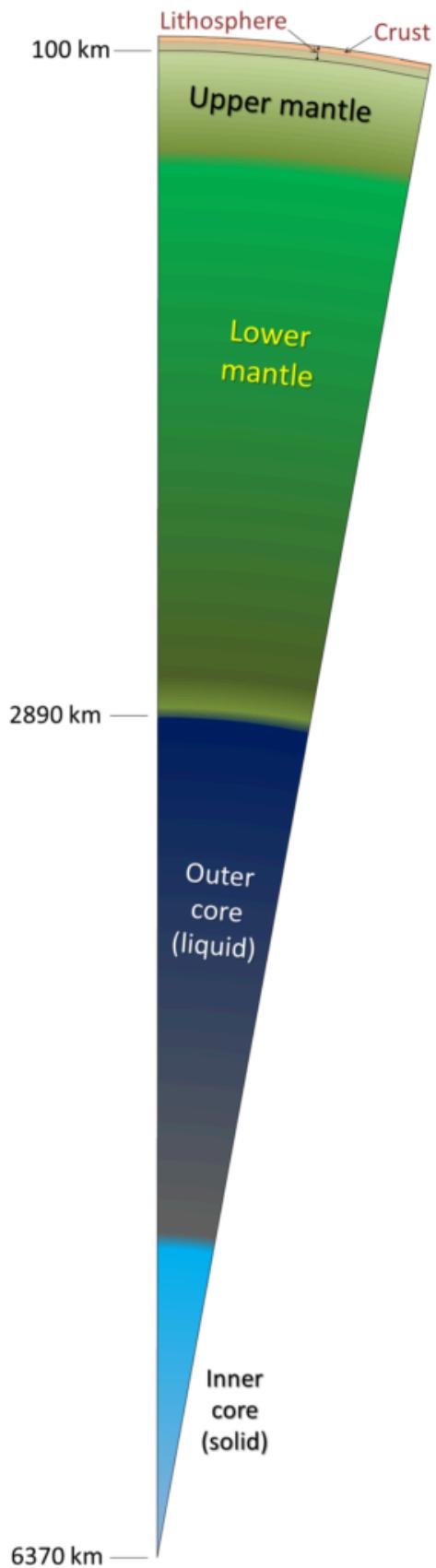
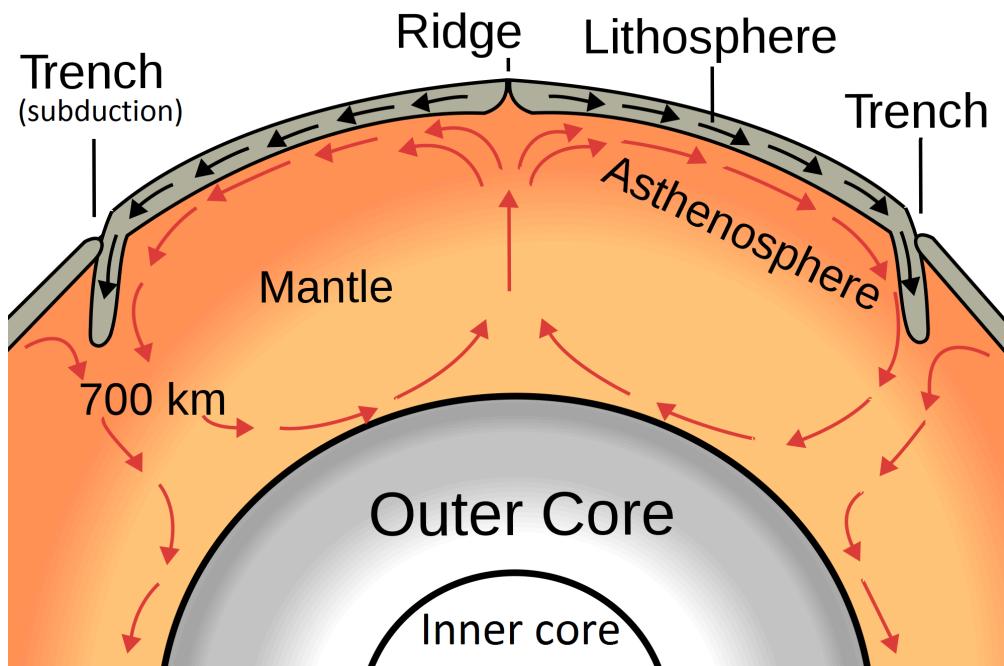


Figure 1.5.1 The components of the interior of the Earth (click on the image to see a

Key to understanding plate tectonics is an understanding of Earth's internal structure, which is illustrated in Figure 1.5.1. Earth's **core** consists mostly of iron. The outer core is hot enough for the iron to be liquid. The inner core—although even hotter—is under so much pressure that it is solid. The **mantle** is made up of iron and magnesium **silicate** minerals. The bulk of the mantle surrounding the outer core is solid rock, but is plastic enough to be able to flow slowly. The outermost part of the mantle is rigid. The **crust**—composed mostly of granite on the continents and mostly of basalt beneath the oceans—is also rigid. The crust and outermost rigid mantle together make up the **lithosphere**. The lithosphere is divided into about 20 **tectonic plates** that move in different directions on Earth's surface.

An important property of Earth (and other planets) is that the temperature increases with depth, from close to 0°C at the surface to about 7000°C at the centre of the core. In the crust, the rate of temperature increase is about 30°C every kilometre. This is known as the **geothermal gradient**.

Heat is continuously flowing outward from Earth's interior, and the transfer of heat from the core to the mantle causes convection in the mantle (Figure 1.5.2). This convection is the primary driving force for the movement of tectonic plates. At places where convection currents in the mantle are moving upward, new lithosphere forms (at ocean ridges), and the plates move apart (diverge). Where two plates are converging (and the convective flow is downward), one plate will be **subducted** (pushed down) into the mantle beneath the other. Many of Earth's major earthquakes and volcanoes are associated with convergent boundaries.



*Figure 1.5.2 Depiction of the convection in the mantle and its relationship to plate motion*

Earth's major tectonic plates and the directions and rates at which they are diverging at sea-floor ridges, are shown in Figure 1.5.3.

### Exercise 1.2 Plate

Using either a map of the tectonic plates from the Internet or Figure 1.5.3 determine which tectonic plate you are on right now, approximately how fast it is moving, and in what direction. How far has that plate moved relative to Earth's core since you were born?



Figure 1.5.3 A map showing 15 of the Earth's tectonic plates and the approximate rates and directions of plate motions.

See Appendix 3 for [Exercise 1.2 answers](#).

### Media Attributions

- Figure 1.5.1: © Steven Earle. CC BY.
- Figure 1.5.2: [Oceanic Spreading](#) by Surachit. Public domain.
- Figure 1.5.3: [Tectonic Plates](#) by USGS. Public domain. Adapted by Steven Earle.

## 1.6 Geological Time

In 1788, after many years of geological study, James Hutton, one of the great pioneers of geology, wrote the following about the age of Earth: *The result, therefore, of our present enquiry is, that we find no vestige of a beginning — no prospect of an end.*<sup>1</sup> Of course he wasn't exactly correct, there was a beginning and there will be an end to Earth, but what he was trying to express is that geological time is so vast that we humans, who typically live for less than a century, have no means of appreciating how much geological time there is. Hutton didn't even try to assign an age to Earth, but we now know that it is approximately 4,570 million years old. Using the scientific notation for geological time, that is 4,570 **Ma** (for *mega annum* or “millions of years”) or 4.57 **Ga** (for *giga annum* or billions of years). More recent dates can be expressed in **ka** (*kilo annum*); for example, the last cycle of glaciation ended at approximately 11.7 ka or 11,700 years ago. This notation will be used for geological dates throughout this book.

### Exercise 1.3 Using geological time notation

To help you understand the scientific notation for geological time—which is used extensively in this book—write the following out in numbers (for example, 3.23 Ma = 3,230,000 years).

1. 2.75 ka
2. 0.93 Ga
3. 14.2 Ma

We use this notation to describe geological events in the same way that we might say “they arrived at 2 pm.” For example, we can say “this rock formed at 45 Ma.” But this notation is not used to express elapsed time. We don’t say: “I studied for 4 pm for that test.” And we don’t say: “The dinosaurs lived for 160 Ma.” Instead, we could say: “The dinosaurs lived from 225 Ma to 65 Ma, which is 160 million years.”

See Appendix 3 for [Exercise 1.3 answers](#).

Unfortunately, knowing how to express geological time doesn’t really help us understand or appreciate its extent. A version of the geological time scale is included as Figure 1.6.1. Unlike time scales you’ll see in other places, or even later in this book, this time scale is linear throughout its length, meaning that 50 Ma during the **Cenozoic** is the same thickness as 50 Ma during the **Hadean**—in each case about the height of the “M” in Ma. The Pleistocene glacial epoch began at about 2.6 Ma, which is equivalent to half the thickness of the thin grey line at the top of the yellow bar marked “Cenozoic.” Most other time scales have earlier parts of Earth’s history compressed so that more detail can be shown for the more recent parts. That makes it difficult to appreciate the extent of geological time.

1. Hutton, J. 1788. Theory of the Earth; or an investigation of the laws observable in the composition, dissolution, and restoration of land upon the Globe. Transactions of the Royal Society of Edinburgh.

To create some context, the **Phanerozoic** Eon (the last 542 million years) is named for the time during which visible (*phaneros*) life (*zoi*) is present in the geological record. In fact, large organisms—those that leave fossils visible to the naked eye—have existed for a little longer than that, first appearing around 600 Ma, or a span of just over 13% of geological time. Animals have been on land for 360 million years, or 8% of geological time. Mammals have dominated since the demise of the dinosaurs around 65 Ma, or 1.5% of geological time, and the genus *Homo* has existed since approximately 2.8 Ma, or 0.06% (1/1,600th) of geological time.

Geologists (and geology students) need to understand geological time. That doesn't mean memorizing the geological time scale; instead, it means getting your mind around the concept that although most geological processes are extremely slow, very large and important things can happen if such processes continue for enough time.

For example, the Atlantic Ocean between Nova Scotia and northwestern Africa has been getting wider at a rate of about 2.5 centimetres (cm) per year. Imagine yourself taking a journey at that rate—it would be impossibly and ridiculously slow. And yet, since it started to form at around 200 Ma (just 4% of geological time), the Atlantic Ocean has grown to a width of over 5,000 kilometres (km)!

A useful mechanism for understanding geological time is to scale it all down into one year. The origin of the solar system and Earth at 4.57 Ga would be represented by January 1, and the present year would be represented by the last tiny fraction of a second on New Year's Eve. At this scale, each day of the year represents 12.5 million years; each hour represents about 500,000 years; each minute represents 8,694 years; and each second represents 145 years. Some significant events in Earth's history, as expressed on this time scale, are summarized on Table 1.1.

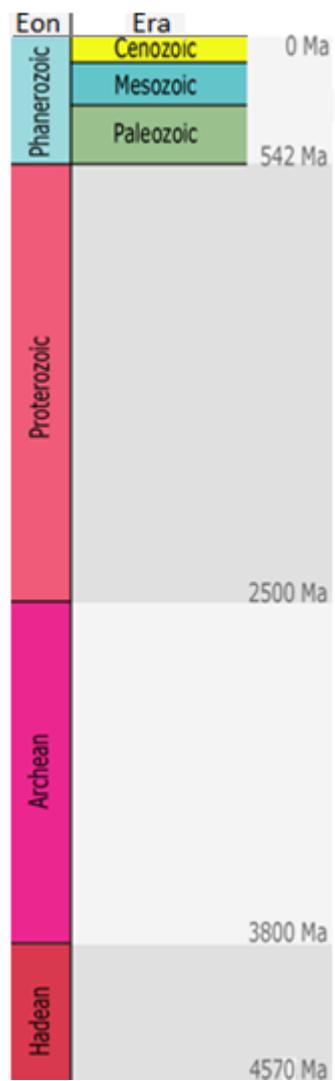


Figure 1.6.1 The geological time scale. [\[Image Description\]](#)

**Table 1.1 A summary of some important geological dates expressed as if all of geological time was condensed into one year.**

<a href="#">[Skip Table]</a>		
Event	Approximate Date	Calendar Equivalent
Formation of oceans and continents	4.5 to 4.4 Ga	January
Evolution of the first primitive life forms	3.8 Ga	early March
Formation of British Columbia's oldest rocks	2.0 Ga	July
Evolution of the first multi-celled animals	0.6 Ga or 600 Ma	November 15
Animals first crawled onto land	360 Ma	December 1
Vancouver Island reached North America and the Rocky Mountains were formed	90 Ma	December 25
Extinction of the non-avian dinosaurs	65 Ma	December 26
Beginning of the Pleistocene ice age	2 Ma or 2000 ka	8 p.m., December 31
Retreat of the most recent glacial ice from southern Canada	14 ka	11:58 p.m., December 31
Arrival of the first people in British Columbia	10 ka	11:59 p.m., December 31
Arrival of the first Europeans on the west coast of what is now Canada	250 years ago	2 seconds before midnight, December 31

#### Exercise 1.4 Take a trip through geological time

We're going on a road trip! Pack some snacks and grab some of your favourite music. We'll start in Tofino on Vancouver Island and head for the Royal Tyrrell Museum just outside of Drumheller, Alberta, 1,500 km away. Along the way, we'll talk about some important geological sites that we pass by, and we'll use the distance as a way of visualizing the extent of geological time. Of course it's just a "virtual" road trip, but it will be fun anyway. To join in, go to: [Virtual Road Trip](#).

Once you've had a chance to do the road trip, answer these questions:

1. We need oxygen to survive, and yet the first presence of free oxygen ( $O_2$  gas) in the atmosphere and the oceans was a "catastrophe" for some organisms. When did this happen and why was it a catastrophe?
2. Approximately how much time elapsed between the colonization of land by plants and animals?
3. Explain why the evolution of land plants was such a critical step in the evolution of life on

Earth.

See Appendix 3 for [Exercise 1.4 answers](#).

#### Image descriptions

**Figure 1.6.1 image description:** The Hadean eon (3800 Ma to 4570 Ma), Archean eon (2500 Ma to 3800 Ma), and Proterozoic eon (542 Ma to 2500 Ma) make up 88% of geological time. The Phanerozoic eon makes up the last 12% of geological time. The Phanerozoic eon (0 Ma to 542 Ma) contains the Paleozoic, Mesozoic, and Cenozoic eras. [\[Return to Figure 1.6.1\]](#)

#### Media Attributions

- Figure 1.6.1: © Steven Earle. CC BY.

## Summary

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

Section	Summary
<a href="#"><u>1.1 What is Geology?</u></a>	Geology is the study of Earth. It is an integrated science that involves the application of many of the other sciences, but geologists also have to consider geological time because most of the geological features that we see today formed thousands, millions, or even billions of years ago.
<a href="#"><u>1.2 Why Study Earth?</u></a>	Geologists study Earth out of curiosity and for other more practical reasons, including understanding the evolution of life on Earth, searching for resources, understanding risks from geological events such as earthquakes, volcanoes, and slope failures, and documenting past environmental and climate changes so that we can understand how human activities are affecting Earth.
<a href="#"><u>1.3 What Do Geologists Do?</u></a>	Geologists work in the resource industries and in efforts to protect our natural resources and the environment in general. They are involved in ensuring that risks from geological events (e.g., earthquakes) are minimized and that the public understands what the risks are. Geologists are also engaged in fundamental research about Earth and in teaching.
<a href="#"><u>1.4 Minerals and Rocks</u></a>	Minerals are naturally occurring, specific combinations of elements that have particular three-dimensional structures. Rocks are made up of mixtures of minerals and can form through igneous, sedimentary, or metamorphic processes.
<a href="#"><u>1.5 Fundamentals of Plate Tectonics</u></a>	The Earth's mantle is convecting because it is being heated from below by the hot core. Those convection currents contribute to the movement of tectonic plates (which are composed of the crust and the uppermost rigid mantle). Plates are formed at divergent boundaries and consumed (subducted) at convergent boundaries. Many important geological processes take place at plate boundaries.
<a href="#"><u>1.6 Geological Time</u></a>	Earth is approximately 4,570,000,000 years old; that is, 4.57 billion years or 4.57 Ga or 4,570 Ma. It's such a huge amount of time that even extremely slow geological processes can have an enormous impact.

### Questions for Review

Answers to Review Questions at the end of each chapter are provided in [Appendix 2](#).

1. In what way is geology different from the other sciences, such as chemistry and physics?
2. How would some familiarity with biology be helpful to a geologist?
3. List three ways in which geologists can contribute to society.
4. Describe the lattice structure and elemental composition of the mineral halite.

5. In what way is a mineral different from a rock?
6. What is the main component of Earth's core?
7. What process leads to convection in the mantle?
8. How does mantle convection contribute to plate tectonics?
9. What are some of the processes that take place at a divergent plate boundary?
10. Dinosaurs first appear in the geological record in rocks at 225 Ma and then disappear at 65 Ma.  
For what proportion (%) of geological time did dinosaurs exist?
11. If a typical rate for the accumulation of sediments is 1 mm/year, what thickness (metres) of sedimentary rock could accumulate over a period of 30 million years?

# Chapter 2 Minerals

## Learning Objectives

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

- Describe the nature of atoms and their constituents, particularly the behaviour of electrons and the formation of ions.
- Apply your understanding of atoms to explain bonding within minerals.
- Describe mineral lattices and explain how they influence mineral properties.
- Categorize minerals into groups based on their compositions.
- Describe a silica tetrahedron and the ways in which tetrahedra combine to make silicate minerals.
- Differentiate between ferromagnesian and other silicate minerals.
- Explain some of the mechanisms of mineral formation.
- Describe some of the important techniques for identifying minerals.

Minerals are all around us: the graphite in your pencil, the salt on your table, the plaster on your walls, and the trace amounts of gold in your computer. Minerals can be found in a wide variety of consumer products including paper, medicine, processed foods, cosmetics, electronic devices, and many more. And of course, everything made of metal is also derived from minerals.

As defined in Chapter 1, a mineral is a naturally occurring combination of specific elements arranged in a particular repeating three-dimensional structure (Figure 1.4.1).

“**Naturally occurring**” implies that minerals are not artificially made. Many minerals (e.g., diamond) can be made in laboratories, but if they can also occur naturally, they still qualify as minerals.

“**Specific elements**” means that most minerals have a specific chemical formula or composition. The mineral pyrite, for example, is  $\text{FeS}_2$  (two atoms of sulfur for each atom of iron), and any significant departure from that would make it a different mineral. But many minerals can have variable compositions within a specific range. The mineral olivine, for example, can range all the way from  $\text{Fe}_2\text{SiO}_4$  to  $\text{FeMgSiO}_4$  to  $\text{Mg}_2\text{SiO}_4$ . Intervening compositions are written as  $(\text{Fe}, \text{Mg})_2\text{SiO}_4$  meaning that Fe and Mg can be present in any proportion, and that there are two of them for each Si present. This type of substitution is known as **solid solution**.

Most important of all, a mineral has a specific “**repeating three-dimensional structure**” or “lattice,” which is the way in which the atoms are arranged. We’ve already seen in Chapter 1 how sodium and chlorine atoms in halite alternate in a regular pattern. That happens to be about the simplest mineral lattice of all; most mineral lattices are much more complicated, as we’ll see.

Some substances that we think must be minerals are not because they lack that repeating 3-dimensional structure of atoms. Volcanic glass is an example, as is pearl or opal. As shown in Figure 2.0.1, opal appears to have a regular structure, but it's not an atomic structure.

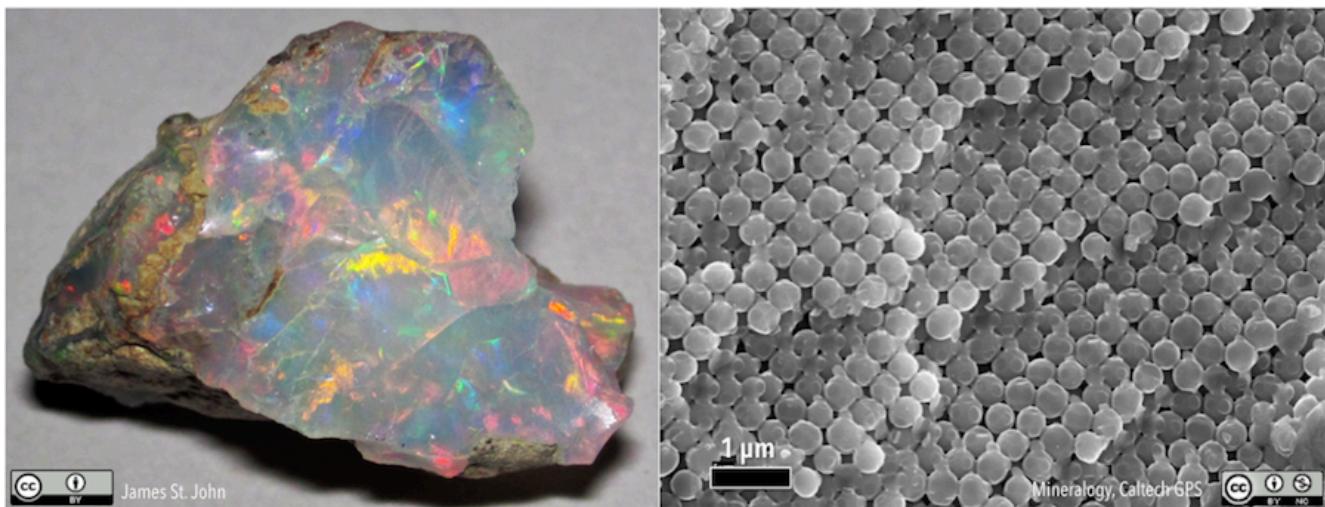


Figure 2.0.1 Opal (left) is like a mineral, but does not have a crystalline structure. Instead, it is made up of layers of closely packed silica spheres (right).

#### Media Attributions

- Figure 2.0.1 (left): [Precious opal](#). © [James St. John](#). CC BY.
- Figure 2.0.1 (right): [Opal beads](#). © [Mineralogy Division, Geological and Planetary Sciences, Caltech](#). CC BY-NC.

## 2.1 Electrons, Protons, Neutrons, and Atoms

All matter that we are familiar with, including mineral crystals, is made up of atoms, and all atoms are made up of three main particles: **protons**, **neutrons**, and **electrons**. As summarized in Table 2.1, protons are positively charged, neutrons are uncharged and electrons are negatively charged. The  $-1$  charge of one electron balances the  $+1$  charge of one proton. Both protons and neutrons have a mass of 1, while electrons have almost no mass.

**Table 2.1 Charges and masses of the particles within atoms**

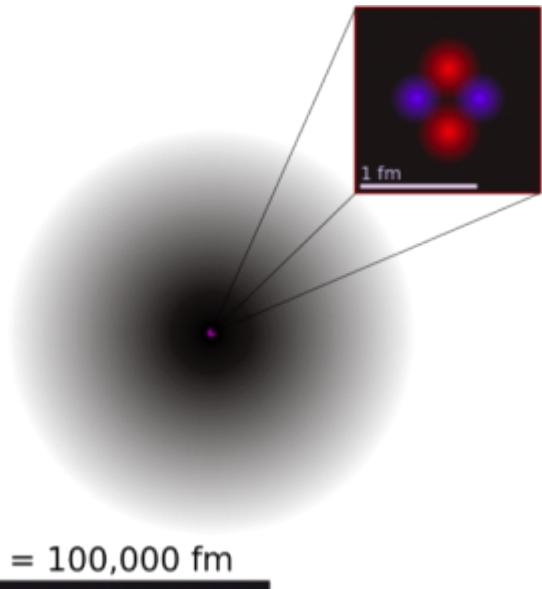
Elementary Particle	Charge	Mass
Proton	$+1$	1
Neutron	0	1
Electron	$-1$	$\sim 0$

The element hydrogen has the simplest atoms, each with just one proton and one electron. The proton forms the nucleus, while the electron orbits around it. All other elements have neutrons as well as protons in their nucleus, such as helium, which is depicted in Figure 2.1.1. The positively charged protons tend to repel each other, but the neutrons help to hold the nucleus together. The number of protons is the **atomic number**, and the number of protons plus neutrons is the **atomic mass**. For hydrogen, the atomic mass is 1 because there is one proton and no neutrons. For helium, it is 4: two protons and two neutrons.

For most of the 16 lightest elements (up to oxygen) the number of neutrons is equal to the number of protons. For most of the remaining elements there are more neutrons than protons because extra neutrons are needed to keep the nucleus together by overcoming the mutual repulsion of the increasing numbers of protons concentrated in a very small space. For example, silicon has 14 protons and 14 neutrons. Its atomic number is 14 and its atomic mass is 28. The most common isotope of uranium has 92 protons and 146 neutrons. Its atomic number is 92 and its atomic mass is 238 ( $92 + 146$ ).

A helium atom is depicted on Figure 2.1.1. The dot in the middle is the nucleus, and the surrounding cloud represents where the two electrons might be at any time. The darker the shade, the more likely that an electron will be there. The helium atom is about 1 angstrom across. An angstrom ( $\text{\AA}$ ) is  $10^{-10}$  metres (m). The helium nucleus is about 1 femtometre across. A femtometre (fm) is  $10^{-15}$  m. In other words, a helium atom's electron cloud is about 100,000 times bigger than its nucleus. Stanley Park in Vancouver is about 2 km across. If Stanley Park was a helium atom, the nucleus would be the size of a walnut.

Electrons orbiting around the nucleus of an atom are arranged in shells—also known as “energy levels.” The first shell can hold only two electrons, while the next shell holds up to eight electrons. Subsequent shells can hold more electrons, but the outermost shell of any atom holds no more than eight electrons. As we'll see, the electrons in the outermost shell play an important role in bonding between atoms. The electron shell configurations for 29 of the first 36 elements are listed in Table 2.2.



$$\underline{1 \text{ \AA} = 100,000 \text{ fm}}$$

*Figure 2.1.1 A depiction of a helium atom.*

**Table 2.2 Electron shell configurations up to element 36. (The inert elements, with filled outer shells, have a \*.)**

<a href="#">[Skip Table]</a>						
Element	Symbol	Atomic No.	Number of Electrons in the 1st Shell	Number of Electrons in the 2nd Shell	Number of Electrons in the 3rd Shell	Number of Electrons in the 4th Shell
<b>Hydrogen</b>	H	1	1	0	0	0
<b>Helium *</b>	He	2	2	0	0	0
<b>Lithium</b>	Li	3	2	1	0	0
<b>Beryllium</b>	Be	4	2	2	0	0
<b>Boron</b>	B	5	2	3	0	0
<b>Carbon</b>	C	6	2	4	0	0
<b>Nitrogen</b>	N	7	2	5	0	0
<b>Oxygen</b>	O	8	2	6	0	0
<b>Fluorine</b>	F	9	2	7	0	0
<b>Neon *</b>	Ne	10	2	8	0	0
<b>Sodium</b>	Na	11	2	8	1	0
<b>Magnesium</b>	Mg	12	2	8	2	0
<b>Aluminum</b>	Al	13	2	8	3	0
<b>Silicon</b>	Si	14	2	8	4	0
<b>Phosphorus</b>	P	15	2	8	5	0
<b>Sulphur</b>	S	16	2	8	6	0
<b>Chlorine</b>	Cl	17	2	8	7	0
<b>Argon *</b>	Ar	18	2	8	8	0
<b>Potassium</b>	K	19	2	8	8	1
<b>Calcium</b>	Ca	20	2	8	8	2
<b>Scandium</b>	Sc	21	2	8	9	2
<b>Titanium</b>	Ti	22	2	8	10	2
<b>Vanadium</b>	V	23	2	8	11	2
<b>Chromium</b>	Cr	24	2	8	13	1
<b>Manganese</b>	Mn	25	2	8	13	2

<a href="#">[Skip Table]</a>						
Element	Symbol	Atomic No.	Number of Electrons in the 1st Shell	Number of Electrons in the 2nd Shell	Number of Electrons in the 3rd Shell	Number of Electrons in the 4th Shell
Iron	Fe	26	2	8	14	2
.	.	.	.	.	.	.
Selenium	Se	34	2	8	18	6
Bromine	Br	35	2	8	18	7
Krypton *	Kr	36	2	8	18	8

## Media Attributions

- Figure 2.1.1: [Helium Atom QM](#). © [Yzmo](#). CC BY-SA.

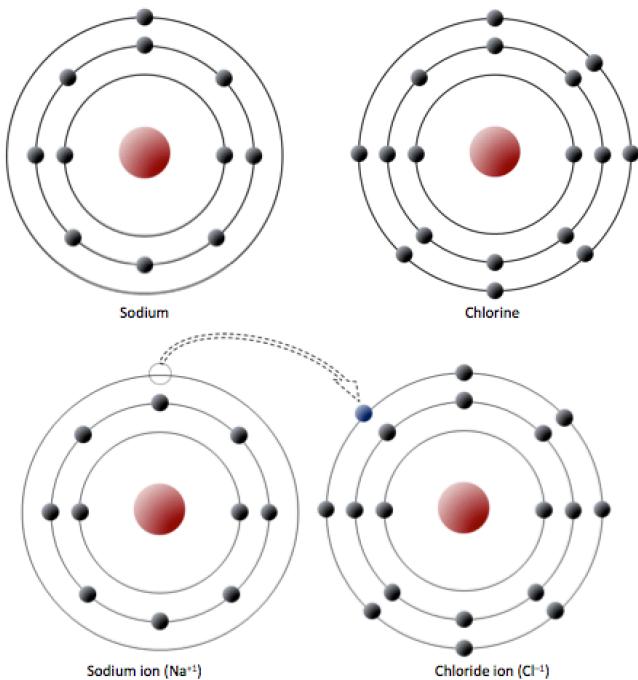
## 2.2 Bonding and Lattices

As we've just seen, an atom seeks to have a full outer shell (i.e., eight electrons for most elements, or two electrons for hydrogen and helium) to be atomically stable. This is accomplished by transferring or sharing electrons with other atoms.

Sodium has 11 electrons: two in the first shell, eight in the second, and one in the third (Figure 2.2.1). Sodium readily gives up that single third-shell electron, and when it loses this one negative charge, it becomes positively charged (because it now has 11 protons and only 10 electrons). By giving up its lone third-shell electron, sodium ends up with a full outer shell. Chlorine, on the other hand, has 17 electrons: two in the first shell, eight in the second, and seven in the third. Chlorine readily accepts an eighth electron to fill its third shell, and therefore becomes negatively charged because it has 17 protons and 18 electrons. In changing their number of electrons, these atoms become **ions**—the sodium loses an electron to become a positive ion or **cation**, and the chlorine gains an electron to become a negative ion or **anion** (Figure 2.2.1).

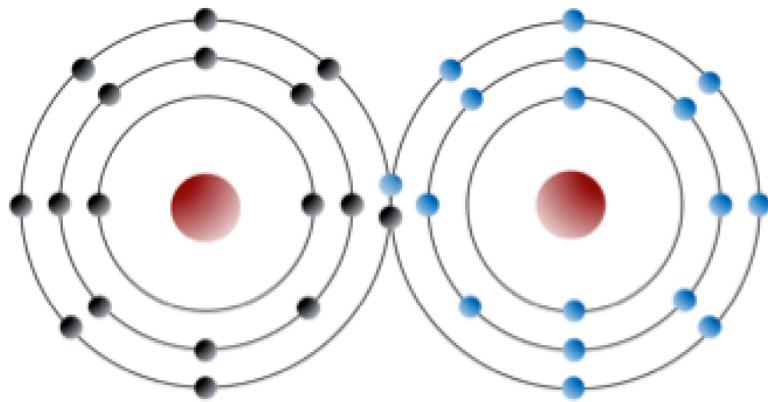
Since negative and positive charges attract, sodium and chlorine ions can stick together, creating an **ionic bond**. Electrons can be thought of as being *transferred* from one atom to another in an ionic bond. Common table salt ( $\text{NaCl}$ ) is a mineral composed of chlorine and sodium linked together by ionic bonds (Figure 1.4.1). The mineral name for  $\text{NaCl}$  is halite.

An element like chlorine can also form bonds without forming ions. For example, two chlorine atoms, which each seek an eighth electron in their outer shell, can share an electron in what is known as a **covalent bond** to form chlorine gas ( $\text{Cl}_2$ ) (Figure 2.2.2). Electrons are *shared* in a covalent bond.



*Figure 2.2.1 A very simplified electron configuration of sodium and chlorine atoms (top). Sodium gives up an electron to become a cation (bottom left) and chlorine accepts an electron to become an anion (bottom right).*

[\[Image Description\]](#)



*Figure 2.2.2 Depiction of a covalent bond between two chlorine atoms. The electrons are black in the left atom and blue in the right atom. Two electrons are shared (one black and one blue) so that each atom “appears” to have a full outer shell.*

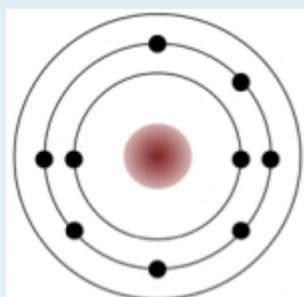
#### Exercise 2.1 Cations, anions, and ionic bonding

A number of elements are listed below along with their atomic numbers. Assuming that the first electron shell can hold two electrons and subsequent electron shells can hold eight electrons, sketch in the electron configurations for these elements. Predict whether the element is likely to form a cation (+) or an anion (−) when electron transfer takes place, and what charge it would have (e.g., +1, +2, −1).

The first one is done for you. Fluorine needed an extra electron to have 8 in its outermost shell, and in gaining that electron it became negatively charged.

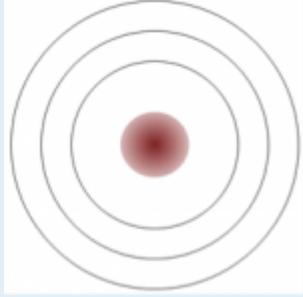
See Appendix 3 for [Exercise 2.1 answers](#).

**Fluorine (9)**

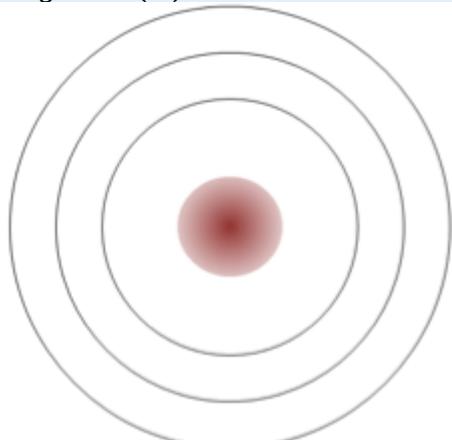


anion (-1)

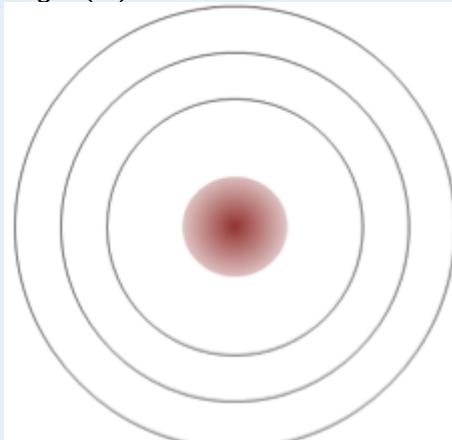
**Lithium (3)**



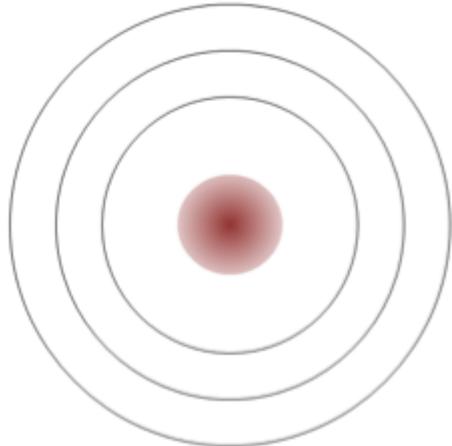
**Magnesium (12)**



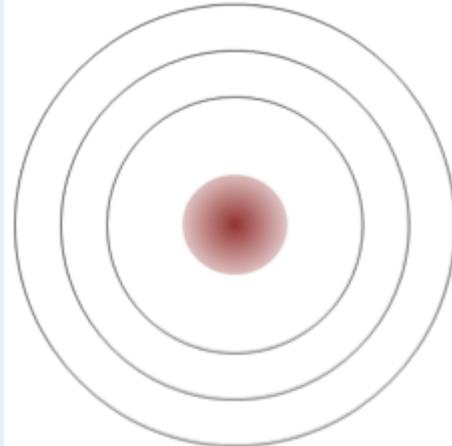
**Argon (18)**

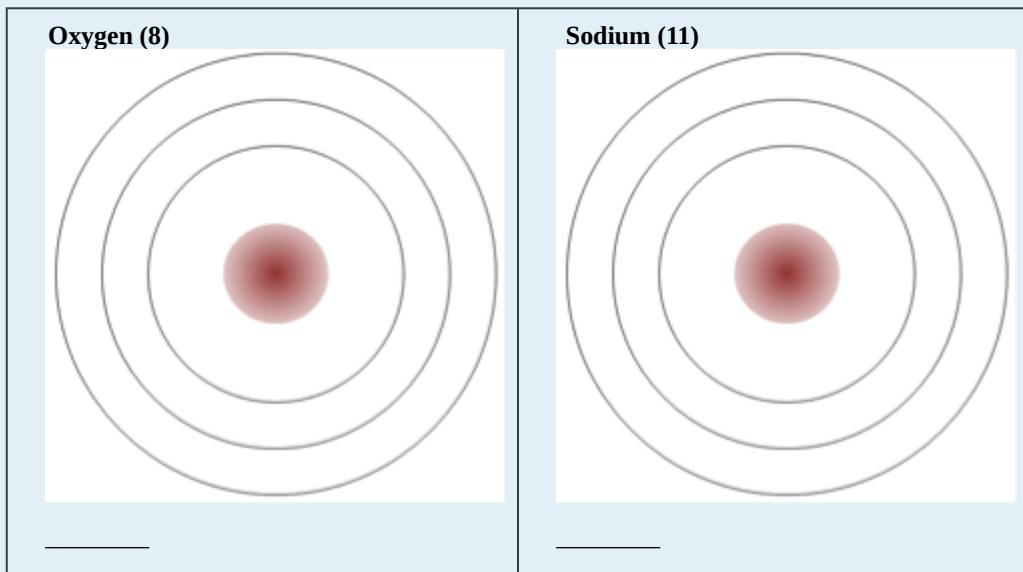


**Chlorine (17)**

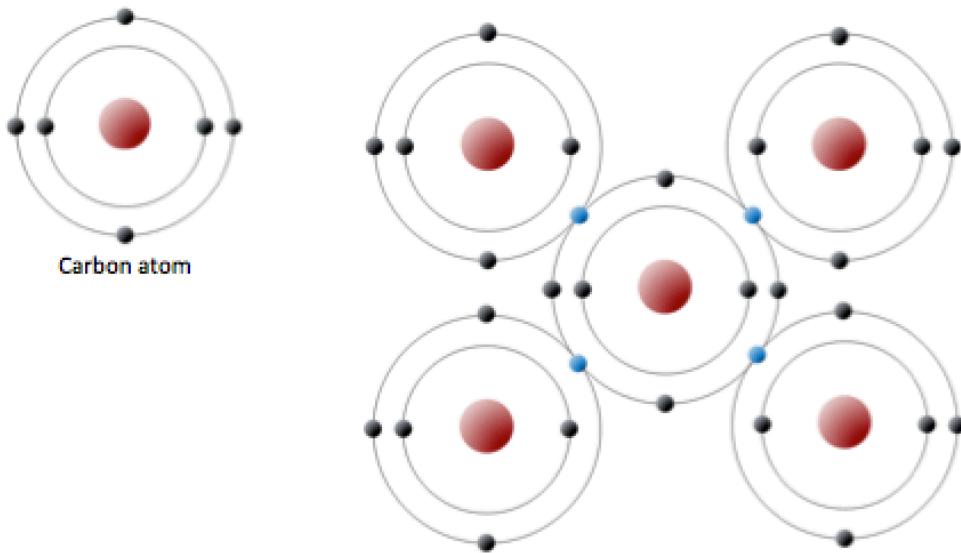


**Beryllium (4)**





A carbon atom has six protons and six electrons; two of the electrons are in the inner shell and four in the outer shell (Figure 2.2.3). Carbon would need to gain or lose four electrons to have a filled outer shell, and this would create too great a charge imbalance for the ion to be stable. On the other hand, carbon can share electrons to create covalent bonds. In the mineral diamond, the carbon atoms are linked together in a three-dimensional framework, where each carbon atom is bonded to four other carbon atoms and every bond is a very strong covalent bond. In the mineral graphite, the carbon atoms are linked together in sheets or layers (Figure 2.2.3), and each carbon atom is covalently bonded to three others. Graphite-based compounds, which are strong because of the strong intra-layer covalent bonding, are used in high-end sports equipment such as ultralight racing bicycles. Graphite itself is soft because the bonding between these layers is relatively weak, and it is used in a variety of applications, including lubricants and pencils.

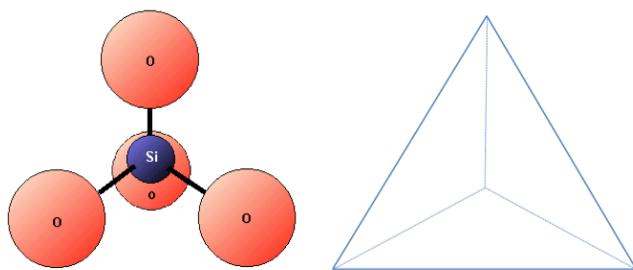


*Figure 2.2.3 The electron configuration of carbon (left) and the sharing of electrons in covalent bonding of diamond (right). The electrons shown in blue are shared between adjacent Carbon atoms. Although shown here in only two dimensions, diamond has a three-dimensional structure as shown on Figure 2.2.5. [\[Image description\]](#)*

Silicon and oxygen bond together to create a **silica tetrahedron**, which is a four-sided pyramid shape with O at each corner and Si in the middle (Figure 2.2.4). This structure is the building block of the silicate minerals (which are described in Section 2.4). The bonds in a silica tetrahedron have some of the properties of covalent bonds and some of the properties of ionic bonds. As a result of the ionic character, silicon becomes a cation (with a charge of +4) and oxygen becomes an anion (with a charge of -2). The net charge of a silica tetrahedron ( $\text{SiO}_4$ ) is:  $4 + 4(-2) = 4 - 8 = -4$ . As we will see later, silica tetrahedra (plural of *tetrahedron*) link together in a variety of ways to form most of the common minerals of the crust.

Most minerals are characterized by ionic bonds, covalent bonds, or a combination of the two, but there are other types of bonds that are important in minerals, including metallic bonds and weaker electrostatic forces (hydrogen or Van der Waals bonds). Metallic elements have outer electrons that are relatively loosely held. (The metals are highlighted on the periodic table in [Appendix 1](#).) When bonds between such atoms are formed, these electrons can move freely from one atom to another. A metal can thus be thought of as an array of positively charged atomic nuclei immersed in a sea of mobile electrons. This feature accounts for two very important properties of metals: their electrical conductivity and their malleability (they can be deformed and shaped).

Molecules that are bonded ionically or covalently can also have other weaker electrostatic forces holding them together. Examples of this are the force holding graphite sheets together and the attraction between water molecules.



*Figure 2.2.4 The silica tetrahedron, the building block of all silicate minerals. (Because the silicon ion has a charge of +4 and the four oxygen ions each have a charge of -2, the silica tetrahedron has a net charge of -4.)*

What's with all of these "sili" names?

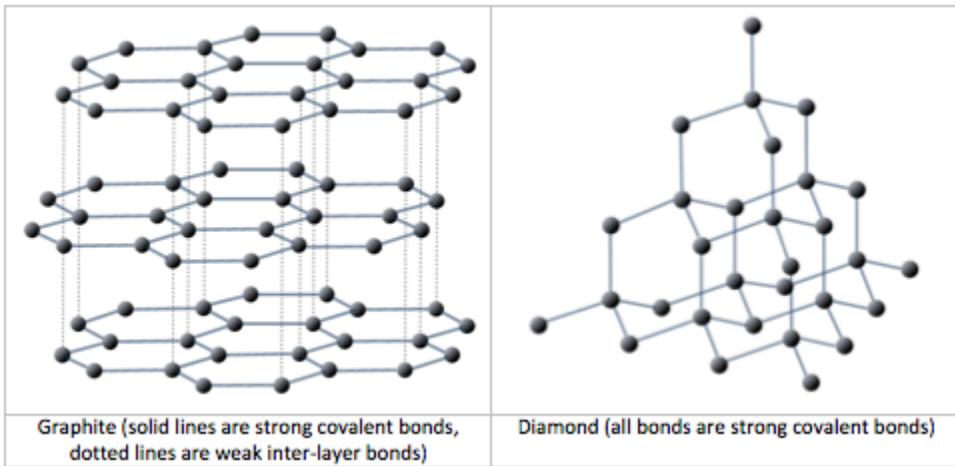
The element **silicon** is one of the most important geological elements and is the second-most abundant element in Earth's crust (after oxygen). Silicon bonds readily with oxygen to form a **silica** tetrahedron (Figure 2.2.4). Pure silicon crystals (created in a lab) are used to make semi-conductive media for electronic devices. A **silicate** mineral is one in which silicon and oxygen are present as silica tetrahedra. Silica also refers to a chemical component of a rock and is expressed as % SiO<sub>2</sub>. The mineral quartz is made up entirely of silica tetrahedra, and some forms of quartz are also known as "silica". **Silicone** is a synthetic product (e.g., silicone rubber, resin, or caulk) made from silicon-oxygen chains and various organic molecules. To help you keep the "sili" names straight, here is a summary table:

**Table 2.3 Summary of "Sili" names**

"Sili" name	Definition
Silicon	The 14 <sup>th</sup> element
Silicon wafer	A crystal of pure silicon sliced very thinly and used for electronics
Silica tetrahedron	A combination of one silicon atom and four oxygen atoms that form a tetrahedron
% silica	The proportion of a rock that is composed of the component SiO <sub>2</sub>
Silica	A solid made out of SiO <sub>2</sub> (but not necessarily a mineral – e.g., opal)
Silicate	A mineral that contains silica tetrahedra (e.g., quartz, feldspar, mica, olivine)
Silicone	A flexible synthetic material made up of Si–O chains with attached organic molecules

Elements that have a full outer shell are described as **inert** because they do not tend to react with other elements to form compounds. That's because they don't need to lose or gain any electrons to become stable, and so they don't become ions. They all appear in the far-right column of the periodic table. Examples are: helium, neon, argon, etc.

As described in Chapter 1, all minerals are characterized by a specific three-dimensional pattern known as a lattice or crystal structure. These structures range from the simple cubic pattern of halite (NaCl) (Figure 1.4.1), to the very complex patterns of some silicate minerals. Two minerals may have the same composition, but very different crystal structures and properties. Graphite and diamond, for example, are both composed only of carbon, but while diamond is the hardest substance known, graphite is softer than paper. Their lattice structures are compared in Figure 2.2.5.



*Figure 2.2.5 A depiction of the lattices of graphite and diamond.*

Mineral lattices have important implications for mineral properties, as exemplified by the hardness of diamond and the softness of graphite. Lattices also determine the shape that mineral crystals grow in and how they break. For example, the right angles in the lattice of the mineral halite (Figure 1.4.1) influence both the shape of its crystals (cubic), and the way those crystals break (Figure 2.2.6).



*Figure 2.2.6 Cubic crystals (left) and right-angle cleavage planes (right) of the mineral halite. If you look closely at the cleavage fragment on the right, you can see where it would break again (cleave) along a plane parallel to an existing surface. In most minerals, cleavage planes do not align with crystal surfaces.*

#### Image Descriptions

**Figure 2.2.1 image description:** Sodium has one electron in its outer shell and chlorine has 7 electrons in its outer shell. Sodium's one outer electron goes to chlorine which makes Chlorine slightly negative

and Sodium slightly positive. They attract each other and together they form Sodium Chloride. [\[Return to Figure 2.2.1\]](#)

**Figure 2.2.3 image description:** (Left) A carbon atom has two electrons in its inner shell and four electrons in its outer shell. (Right) One Carbon atom shares electrons with four other carbon atoms to form a complete outer shell. [\[Return to Figure 2.2.3\]](#)

#### Media Attributions

- Figures 2.2.1, 2.2.2, 2.2.3, 2.2.4, 2.2.5 and 2.2.6 (right): © Steven Earle. CC BY.
- Figure 2.2.6 (left): [Halite](#). © [Rob Lavinsky, iRocks.com](#). CC BY-SA.

## 2.3 Mineral Groups

Most minerals are made up of a cation (a positively charged ion) or several cations, plus an anion (a negatively charged ion (e.g.,  $S^{2-}$ )) or an anion complex (e.g.,  $SO_4^{2-}$ ). For example, in the mineral hematite ( $Fe_2O_3$ ), the cation is  $Fe^{3+}$  (iron) and the anion is  $O^{2-}$  (oxygen). The two  $Fe^{3+}$  ions have an overall charge of +6 and that balances the overall charge of -6 from the three  $O^{2-}$  ions.

We group minerals into classes on the basis of their predominant anion or anion complex. These include oxides, sulphides, carbonates, silicates, and others. Silicates are by far the predominant group in terms of their abundance within the crust and mantle. (They will be discussed in Section 2.4). Some examples of minerals from the different mineral groups are given in Table 2.4.

**Table 2.4 The main mineral groups and some examples of minerals in each group.**

<a href="#">[Skip Table]</a>	
Group	Examples
<b>Oxides</b>	Hematite (iron oxide $Fe_2O_3$ ), corundum (aluminum oxide $Al_2O_3$ ), water ice ( $H_2O$ )
<b>Sulphides</b>	Galena (lead sulphide $PbS$ ), pyrite (iron sulphide $FeS_2$ ), chalcopyrite (copper-iron sulphide $CuFeS_2$ )
<b>Sulphates</b>	Gypsum (calcium sulphate $CaSO_4 \cdot H_2O$ ), barite (barium sulphate $BaSO_4$ ) ( <i>Note that sulphates are different from sulphides. Sulphates have the <math>SO_4^{2-}</math> ion while sulphides have the <math>S^{2-}</math> ion</i> )
<b>Halides</b>	Fluorite (calcium fluoride $CaF_2$ ), halite (sodium chloride $NaCl$ ) ( <i>Halide minerals have halogen elements as their anion — the minerals in the second last column on the right side of the periodic table, including F, Cl, Br, etc. — see the periodic table in <a href="#">Appendix 1: List of Geologically Important Elements and the Periodic Table</a>.</i> )
<b>Carbonates</b>	Calcite (calcium carbonate $CaCO_3$ ), dolomite (calcium-magnesium carbonate $(Ca,Mg)CO_3$ )
<b>Phosphates</b>	Apatite ( $Ca_5(PO_4)_3(OH)$ ), Turquoise ( $CuAl_6(PO_4)_4(OH)_8 \cdot 5H_2O$ )
<b>Silicates</b>	Quartz ( $SiO_2$ ), feldspar (sodium-aluminum silicate $NaAlSi_3O_8$ ), olivine (iron or magnesium silicate $(Mg,Fe)_2SiO_4$ ) ( <i>Note that in quartz the anion is oxygen, and while it could be argued, therefore, that quartz is an oxide, it is always classed with the silicates.</i> )
<b>Native minerals</b>	Gold ( $Au$ ), diamond ( $C$ ), graphite ( $C$ ), sulphur ( $S$ ), copper ( $Cu$ )

**Oxide** minerals have oxygen ( $O^{2-}$ ) as their anion, but they exclude those with oxygen complexes such as carbonate ( $CO_3^{2-}$ ), sulphate ( $SO_4^{2-}$ ), and silicate ( $SiO_4^{4-}$ ). The most important oxides are the iron oxides hematite and magnetite ( $Fe_2O_3$  and  $Fe_3O_4$ , respectively). Both of these are common ores of iron. Corundum ( $Al_2O_3$ ) is used as an abrasive, but can also be a gemstone in its ruby and sapphire varieties. If the oxygen is also combined with hydrogen to form the hydroxyl anion ( $OH^-$ ) the mineral is known as a **hydroxide**. Some important hydroxides are limonite and bauxite, which are ores of iron and aluminium

respectively. Frozen water ( $\text{H}_2\text{O}$ ) is a mineral (an oxide), but liquid water is not because it doesn't have a regular lattice.

**Sulphides** are minerals with the  $\text{S}^{2-}$  anion, and they include galena ( $\text{PbS}$ ), sphalerite ( $\text{ZnS}$ ), chalcopyrite ( $\text{CuFeS}_2$ ), and molybdenite ( $\text{MoS}_2$ ), which are the most important ores of lead, zinc, copper, and molybdenum respectively. Some other sulphide minerals are pyrite ( $\text{FeS}_2$ ), bornite ( $\text{Cu}_5\text{FeS}_4$ ), stibnite ( $\text{Sb}_2\text{S}_3$ ), and arsenopyrite ( $\text{FeAsS}$ ).

**Sulphates** are minerals with the  $\text{SO}_4^{2-}$  anion, and these include anhydrite ( $\text{CaSO}_4$ ) and its cousin gypsum ( $\text{CaSO}_4 \cdot 2\text{H}_2\text{O}$ ) and the sulphates of barium and strontium: barite ( $\text{BaSO}_4$ ) and celestite ( $\text{SrSO}_4$ ). In all of these minerals, the cation has a +2 charge, which balances the -2 charge on the sulphate ion.

The **halides** are so named because the anions include the **halogen** elements chlorine, fluorine, bromine, etc. Examples are halite ( $\text{NaCl}$ ), cryolite ( $\text{Na}_3\text{AlF}_6$ ), and fluorite ( $\text{CaF}_2$ ).

The **carbonates** include minerals in which the anion is the  $\text{CO}_3^{2-}$  complex. The carbonate combines with +2 cations to form minerals such as calcite ( $\text{CaCO}_3$ ), magnesite ( $\text{MgCO}_3$ ), dolomite ( $(\text{Ca},\text{Mg})\text{CO}_31), and siderite ( $\text{FeCO}_3$ ). The copper minerals malachite and azurite are also carbonates.$

In **phosphate** minerals, the anion is the  $\text{PO}_4^{3-}$  complex. An important phosphate mineral is apatite ( $\text{Ca}_5(\text{PO}_4)_3(\text{OH})$ ), which is what your teeth are made of. Note that it is called a phosphate, not a hydroxide, even though it has a hydroxyl ion.

The **silicate** minerals include the elements silicon and oxygen in varying proportions ranging from  $\text{Si} : \text{O}_2$  to  $\text{Si} : \text{O}_4$ . These are discussed at length in Section 2.4.

**Native minerals** are single-element minerals, such as gold, copper, sulphur, and graphite.

### Exercise 2.2 Mineral groups

We classify minerals according to the anion part of the mineral formula, and mineral formulas are always written with the anion part on the right. For example, for pyrite ( $\text{FeS}_2$ ),  $\text{Fe}^{2+}$  is the cation and  $\text{S}^-$  is the anion. This helps us to know that it's a sulphide, but it is not always that obvious. Hematite ( $\text{Fe}_2\text{O}_3$ ) is an oxide; that's easy, but anhydrite ( $\text{CaSO}_4$ ) is a sulphate because  $\text{SO}_4^{2-}$  is the anion, not O. Along the same lines, calcite ( $\text{CaCO}_3$ ) is a carbonate, and olivine ( $\text{Mg}_2\text{SiO}_4$ ) is a silicate. Minerals with only one element (such as S) are native minerals, while those with an anion from the halogen column of the periodic table (Cl, F, Br, etc.) are halides. Provide group names for the following minerals:

1. The notations of two (or more) elements enclosed in parentheses with a comma between them: ( $\text{Ca},\text{Mg}$ ), indicates that both can be present, in varying proportions, but that there is still only one of them for each anion present.

**Table 2.5 Provide group names for the following minerals**

<a href="#">[Skip Table]</a>		
Name	Formula	Group
sphalerite	ZnS	
magnetite	Fe <sub>3</sub> O <sub>4</sub>	
pyroxene	MgSiO <sub>3</sub>	
anglesite	PbSO <sub>4</sub>	
sylvite	KCl	
silver	Ag	
fluorite	CaF <sub>2</sub>	
ilmenite	FeTiO <sub>3</sub>	
siderite	FeCO <sub>3</sub>	
feldspar	KAlSi <sub>3</sub> O <sub>8</sub>	
sulphur	S	
xenotime	YPO <sub>4</sub>	

See Appendix 3 for [Exercise 2.2 answers](#).

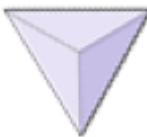
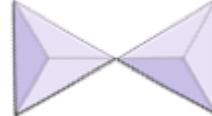
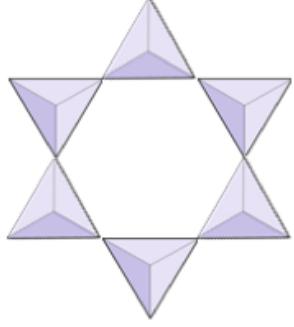
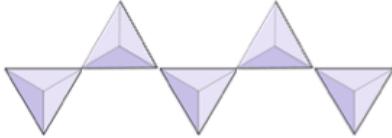
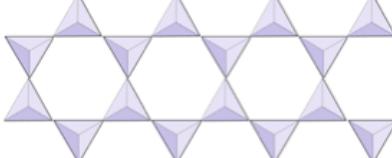
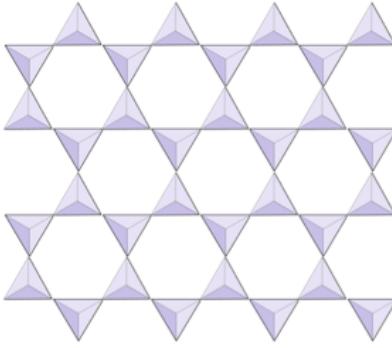
## 2.4 Silicate Minerals

The vast majority of the minerals that make up the rocks of Earth's crust are silicate minerals. These include minerals such as quartz, feldspar, mica, amphibole, pyroxene, olivine, and a variety of clay minerals. The building block of all of these minerals is the **silica tetrahedron**, a combination of four oxygen atoms and one silicon atom. As we've seen, it's called a tetrahedron because planes drawn through the oxygen atoms form a shape with 4 surfaces (Figure 2.2.4). Since the silicon ion has a charge of 4 and each of the four oxygen ions has a charge of -2, the silica tetrahedron has a net charge of -4.

In silicate minerals, these tetrahedra are arranged and linked together in a variety of ways, from single units to complex frameworks (Table 2.6). The simplest silicate structure, that of the mineral **olivine**, is composed of isolated tetrahedra bonded to iron and/or magnesium ions. In olivine, the -4 charge of each silica tetrahedron is balanced by two **divalent** (i.e., +2) iron or magnesium cations. Olivine can be either  $Mg_2SiO_4$  or  $Fe_2SiO_4$ , or some combination of the two  $(Mg,Fe)_2SiO_4$ . The divalent cations of magnesium and iron are quite close in radius (0.73 versus 0.62 angstroms<sup>1</sup>). Because of this size similarity, and because they are both divalent cations (both can have a charge of +2), iron and magnesium can readily substitute for each other in olivine and in many other minerals.

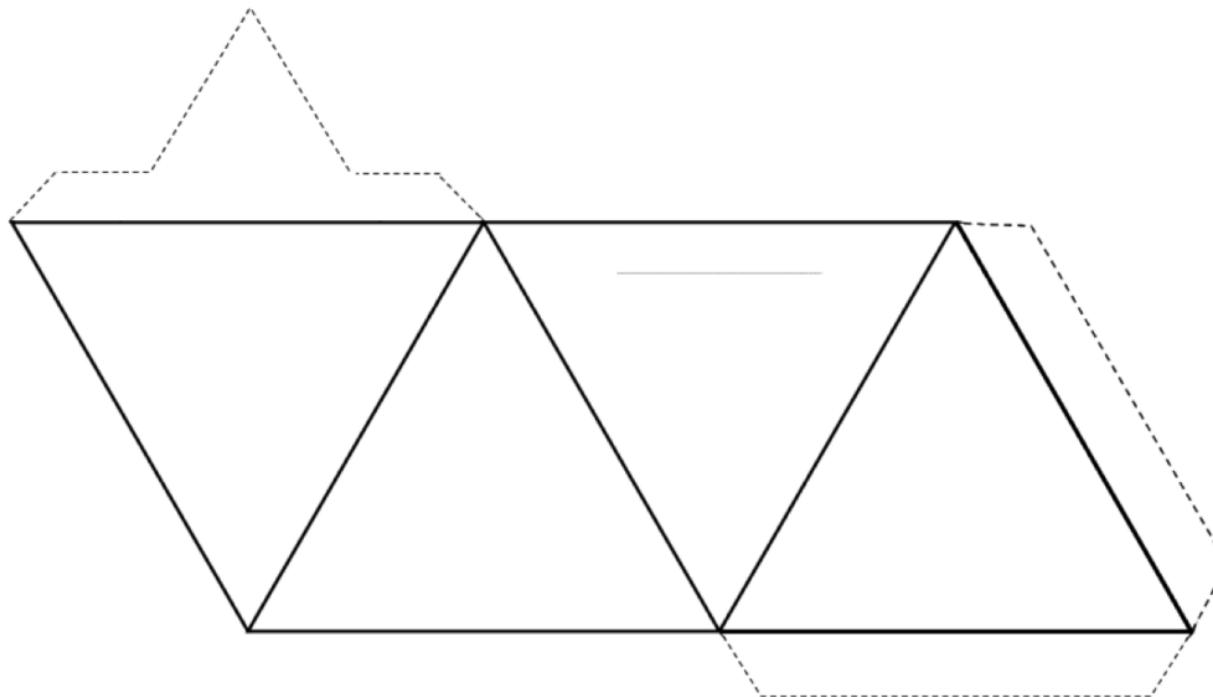
1. An angstrom is the unit commonly used for the expression of atomic-scale dimensions. One angstrom is  $10^{-10}$  metres or 0.0000000001 metres. The symbol for an angstrom is Å.

**Table 2.6 Silicate mineral configurations. The triangles represent silica tetrahedra.**

<a href="#">[Skip Table]</a>		
Tetrahedron Configuration Picture	Tetrahedron Configuration Name	Example Minerals
	Isolated (nesosilicates)	Olivine, garnet, zircon, kyanite
	Pairs (sorosilicates)	Epidote, zoisite
	Rings (cyclosilicates)	Tourmaline
	Single chains (inosilicates)	Pyroxenes, wollastonite
	Double chains (inosilicates)	Amphiboles
	Sheets (phyllosilicates)	Micas, clay minerals, serpentine, chlorite

<a href="#">[Skip Table]</a>		
Tetrahedron Configuration Picture	Tetrahedron Configuration Name	Example Minerals
3-dimensional structure	Framework (tectosilicates)	Feldspars, quartz, zeolite

## Exercise 2.3 Make a Tetrahedron



Cut around the outside of the shape (solid lines and dotted lines), and then fold along the solid lines to form a tetrahedron. If you have glue or tape, secure the tabs to the tetrahedron to hold it together. If you don't have glue or tape, make a slice along the thin grey line and insert the pointed tab into the slit.

If you are doing this in a classroom, try joining your tetrahedron with others into pairs, rings, single and double chains, sheets, and even three-dimensional frameworks.

See Appendix 3 for [Exercise 2.3 answers](#).

In olivine, unlike most other silicate minerals, the silica tetrahedra are not bonded to each other. Instead they are bonded to the iron and/or magnesium ions, in the configuration shown on Figure 2.4.1.

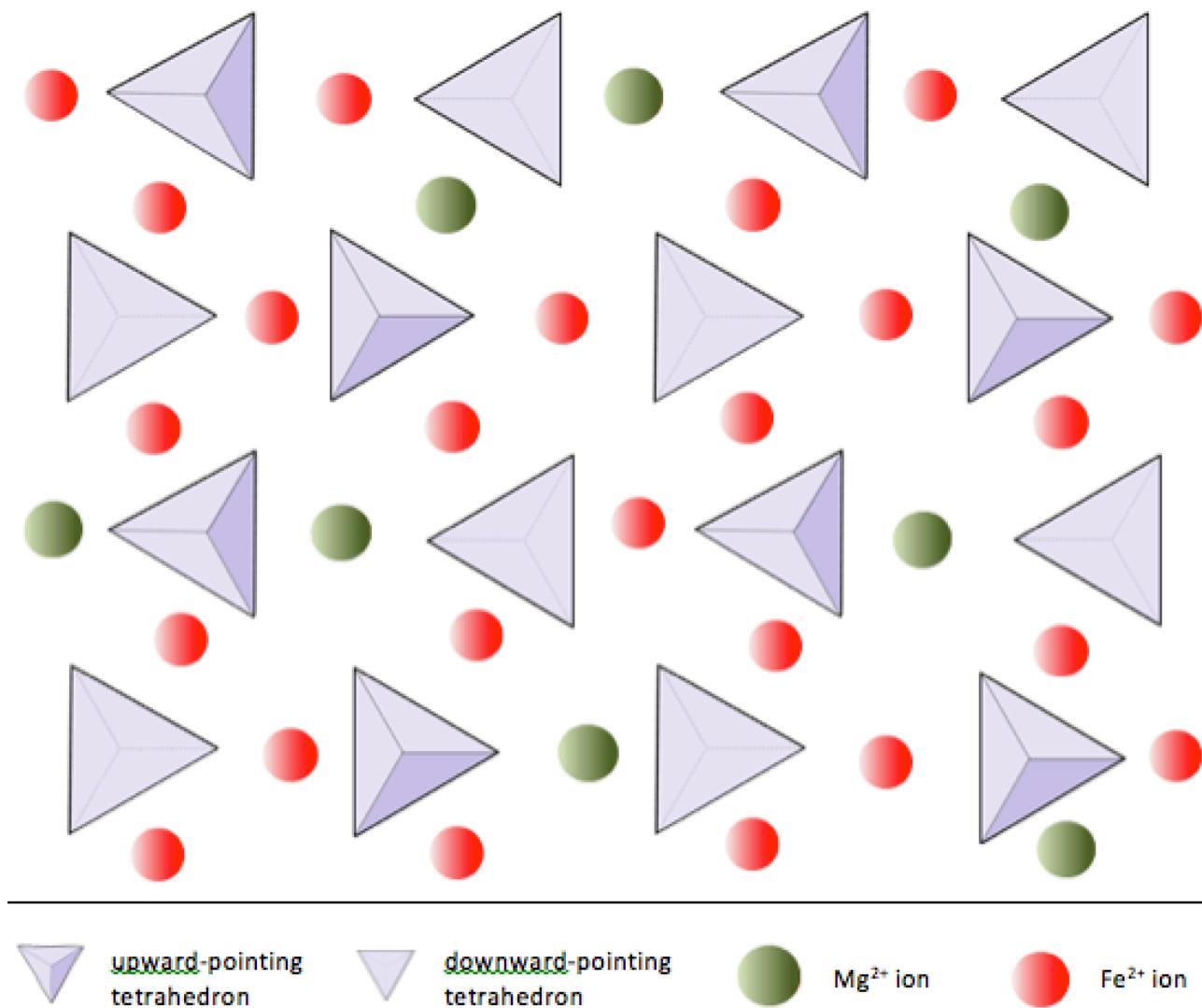


Figure 2.4.1 A depiction of the structure of olivine as seen from above. The formula for this particular olivine, which has three Fe ions for each Mg ion, could be written:  $Mg_0.5Fe_{1.5}SiO_4$ .

As already noted, the 2 ions of iron and magnesium are similar in size (although not quite the same). This allows them to substitute for each other in some silicate minerals. In fact, the ions that are common in silicate minerals have a wide range of sizes, as depicted in Figure 2.4.2. All of the ions shown are cations, except for oxygen. Note that iron can exist as both a +2 ion (if it loses two electrons during ionization) or a +3 ion (if it loses three).  $Fe^{2+}$  is known as **ferrous** iron.  $Fe^{3+}$  is known as **ferric** iron. Ionic radii are critical to the composition of silicate minerals, so we'll be referring to this diagram again.

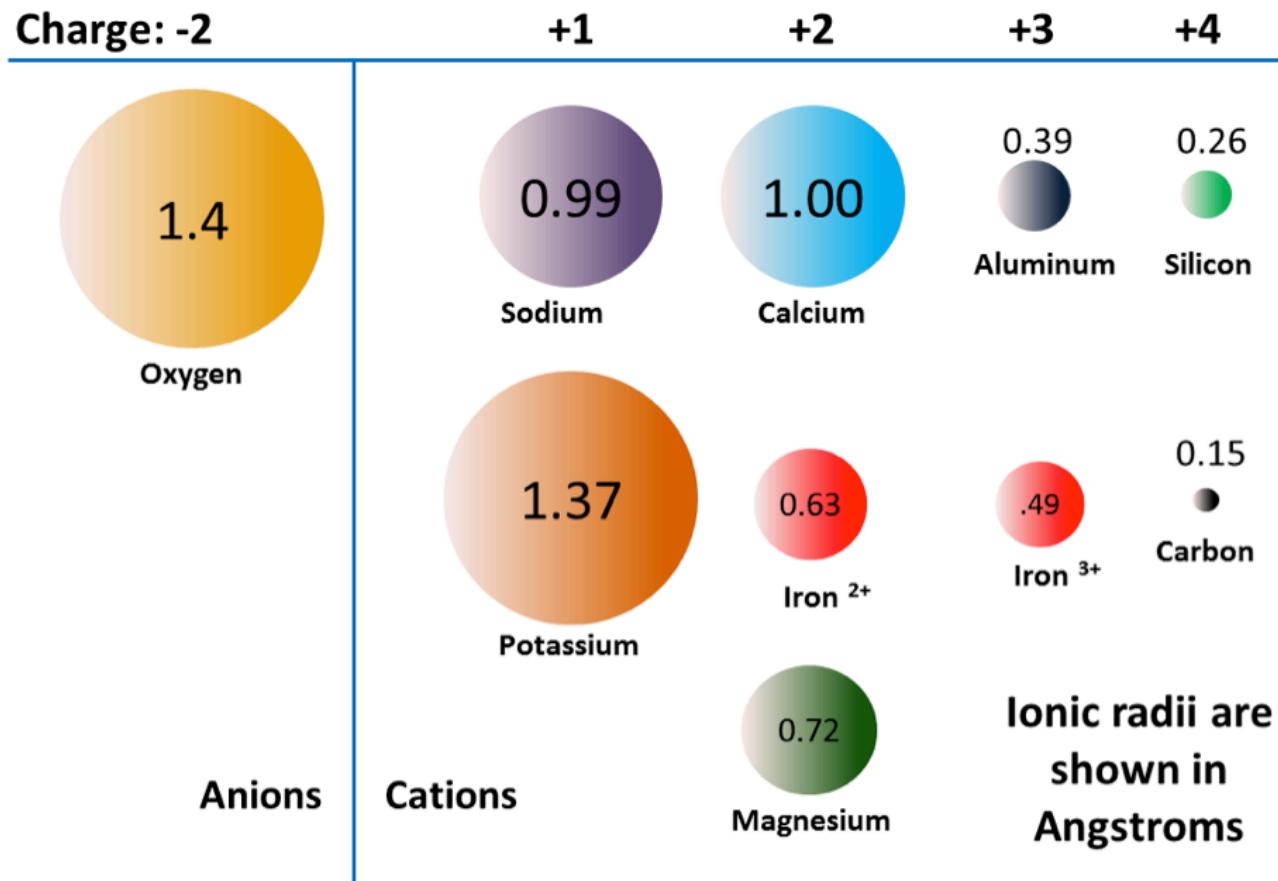
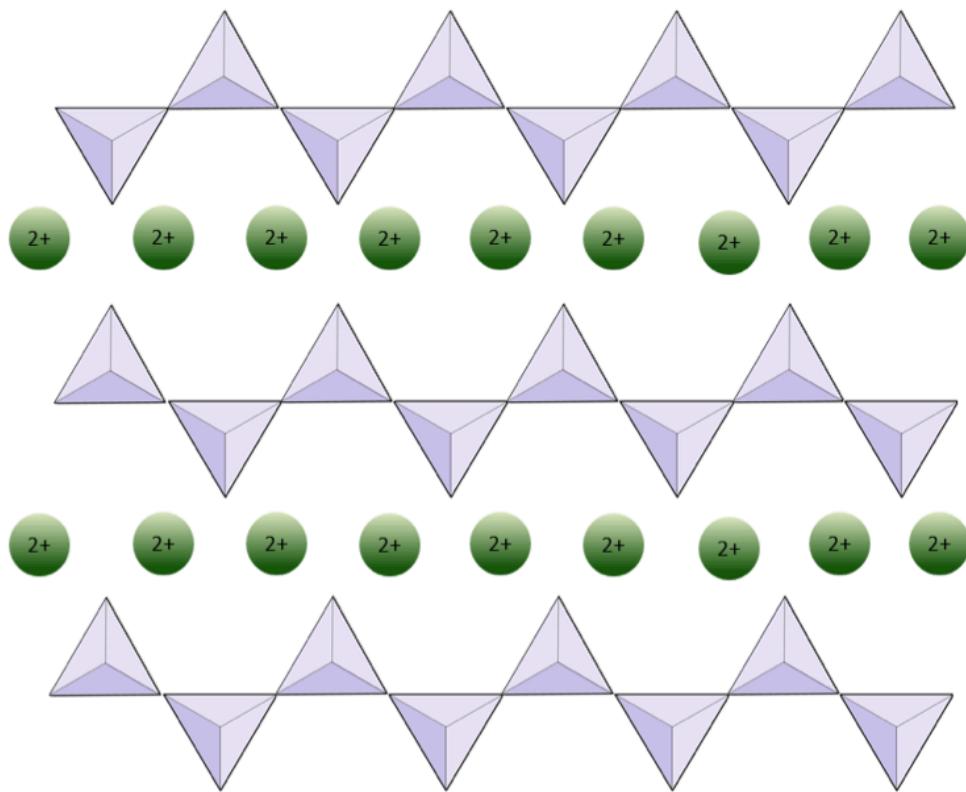
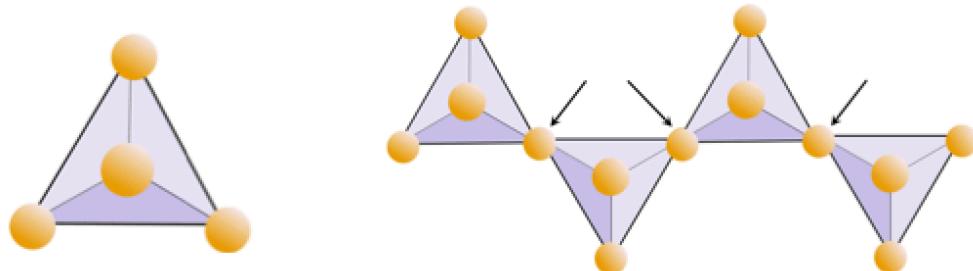


Figure 2.4.2 The ionic radii (effective sizes) in angstroms, of some of the common ions in silicate minerals. [\[Image Description\]](#)

The structure of the single-chain silicate pyroxene is shown on Figures 2.4.3 and 2.4.4. In **pyroxene**, silica tetrahedra are linked together in a single chain, where one oxygen ion from each tetrahedron is shared with the adjacent tetrahedron, hence there are fewer oxygens in the structure. The result is that the oxygen-to-silicon ratio is lower than in olivine (3:1 instead of 4:1), and the net charge per silicon atom is less (-2 instead of -4). Therefore, fewer cations are necessary to balance that charge. Pyroxene compositions are of the type  $\text{MgSiO}_3$ ,  $\text{FeSiO}_3$ , and  $\text{CaSiO}_3$ , or some combination of these. Pyroxene can also be written as  $(\text{Mg}, \text{Fe}, \text{Ca})\text{SiO}_3$ , where the elements in the brackets can be present in any proportion. In other words, pyroxene has one cation for each silica tetrahedron (e.g.,  $\text{MgSiO}_3$ ) while olivine has two (e.g.,  $\text{Mg}_2\text{SiO}_4$ ). Because each silicon ion is +4 and each oxygen ion is -2, the three oxygens (-6) and the one silicon (+4) give a net charge of -2 for the single chain of silica tetrahedra. In pyroxene, the one divalent cation (2) per tetrahedron balances that -2 charge. In olivine, it takes two divalent cations to balance the -4 charge of an isolated tetrahedron. The structure of pyroxene is more “permissive” than that of olivine—meaning that cations with a wider range of ionic radii can fit into it. That’s why pyroxenes can have iron (radius 0.63 Å) or magnesium (radius 0.72 Å) or calcium (radius 1.00 Å) cations (see Figure 2.4.2 above).



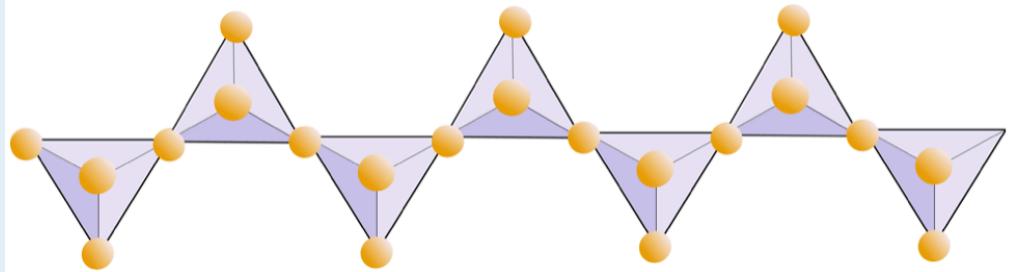
*Figure 2.4.3 A depiction of the structure of pyroxene. The tetrahedral chains continue to left and right and each is interspersed with a series of divalent cations. If these are Mg ions, then the formula is  $MgSiO_3$ .*



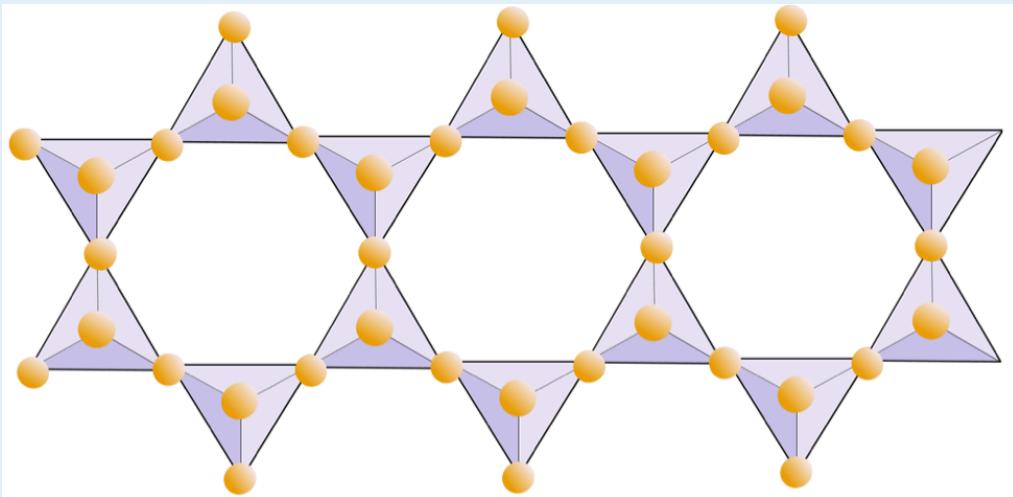
*Figure 2.4.4 A single silica tetrahedron (left) with four oxygen ions per silicon ion ( $SiO_4$ ). Part of a single chain of tetrahedra (right), where the oxygen atoms at the adjoining corners are shared between two tetrahedra (arrows). For a very long chain the resulting ratio of silicon to oxygen is 1 to 3 ( $SiO_3$ ).*

#### Exercise 2.4 Oxygen deprivation

The diagram below represents a single chain in a silicate mineral. Count the number of tetrahedra versus the number of oxygen ions (yellow spheres). Each tetrahedron has one silicon ion so this should give you the ratio of Si to O in single-chain silicates (e.g., pyroxene).



The diagram below represents a double chain in a silicate mineral. Again, count the number of tetrahedra versus the number of oxygen ions. This should give you the ratio of Si to O in double-chain silicates (e.g., amphibole).



See Appendix 3 for [Exercise 2.4 answers](#).

In **amphibole** structures, the silica tetrahedra are linked in a double chain that has an oxygen-to-silicon ratio lower than that of pyroxene, and hence still fewer cations are necessary to balance the charge. Amphibole is even more permissive than pyroxene and its compositions can be very complex. Hornblende, for example, can include sodium, potassium, calcium, magnesium, iron, aluminum, silicon, oxygen, fluorine, and the hydroxyl ion ( $\text{OH}^-$ ).

In **mica** structures, the silica tetrahedra are arranged in continuous sheets, where each tetrahedron shares three oxygen anions with adjacent tetrahedra. There is even more sharing of oxygens between adjacent tetrahedra and hence fewer cations are needed to balance the charge of the silica-tetrahedra structure in sheet silicate minerals. Bonding between sheets is relatively weak, and this accounts for the well-developed one-directional cleavage in micas (Figure 2.4.5). **Biotite** mica can have iron and/or magnesium in it and that makes it a **ferromagnesian** silicate mineral (like olivine, pyroxene, and amphibole). **Chlorite** is another similar mineral that commonly includes magnesium. In **muscovite** mica, the only cations present are aluminum and potassium; hence it is a non-ferromagnesian silicate mineral.

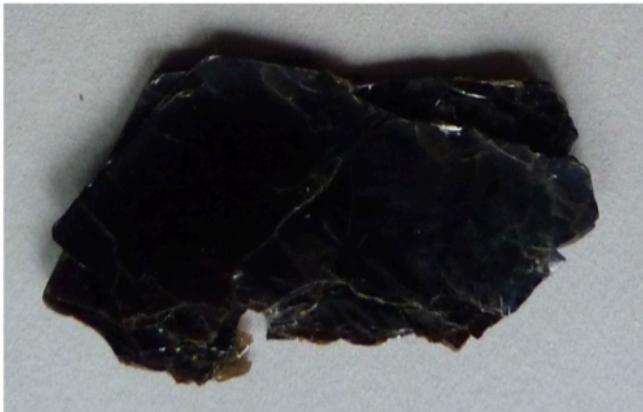


Figure 2.4.5 Biotite mica (left) and muscovite mica (right). Both are sheet silicates and split easily into thin layers along planes parallel to the sheets. Biotite is dark like the other iron- and/or magnesium-bearing silicates (e.g., olivine, pyroxene, and amphibole), while muscovite is light coloured. (Each sample is about 3 cm across.)

Apart from muscovite, biotite, and chlorite, there are many other **sheet silicates** (a.k.a. **phyllosilicates**), many of which exist as clay-sized fragments (i.e., less than 0.004 millimetres). These include the clay minerals **kaolinite**, **illite**, and **smectite**, and although they are difficult to study because of their very small size, they are extremely important components of rocks and especially of soils.

All of the sheet silicate minerals also have water molecules within their structure.

Silica tetrahedra are bonded in three-dimensional frameworks in both the **feldspars** and **quartz**. These are **non-ferromagnesian minerals**—they don't contain any iron or magnesium. In addition to silica tetrahedra, feldspars include the cations aluminum, potassium, sodium, and calcium in various combinations. Quartz contains only silica tetrahedra.

The three main **feldspar** minerals are **potassium feldspar**, (a.k.a. K-feldspar or K-spar) and two types of plagioclase feldspar: **albite** (sodium only) and **anorthite** (calcium only). As is the case for iron and magnesium in olivine, there is a continuous range of compositions (solid solution series) between albite and anorthite in plagioclase. Because the calcium and sodium ions are almost identical in size (1.00 Å versus 0.99 Å) any intermediate compositions between  $\text{CaAl}_2\text{Si}_3\text{O}_8$  and  $\text{NaAlSi}_3\text{O}_8$  can exist (Figure 2.4.6). This is a little bit surprising because, although they are very similar in size, calcium and sodium ions don't have the same charge ( $\text{Ca}^{2+}$  versus  $\text{Na}^+$ ). This problem is accounted for by the corresponding substitution of  $\text{Al}^{+3}$  for  $\text{Si}^{+4}$ . Therefore, albite is  $\text{NaAlSi}_3\text{O}_8$  (1 Al and 3 Si) while anorthite is  $\text{CaAl}_2\text{Si}_2\text{O}_8$  (2 Al and 2 Si), and plagioclase feldspars of intermediate composition have intermediate proportions of Al and Si. This is called a “coupled-substitution.”

The intermediate-composition plagioclase feldspars are oligoclase (10% to 30% Ca), andesine (30% to 50% Ca), labradorite (50% to 70% Ca), and bytownite (70% to 90% Ca). **K-feldspar** ( $\text{KAlSi}_3\text{O}_8$ ) has a slightly different structure than that of plagioclase, owing to the larger size of the potassium ion (1.37 Å) and because of this large size, potassium and sodium do not readily substitute for each other, except at high temperatures. These high-temperature feldspars are likely to be found only in volcanic rocks because intrusive igneous rocks cool slowly enough to low temperatures for the feldspars to change into one of the lower-temperature forms.

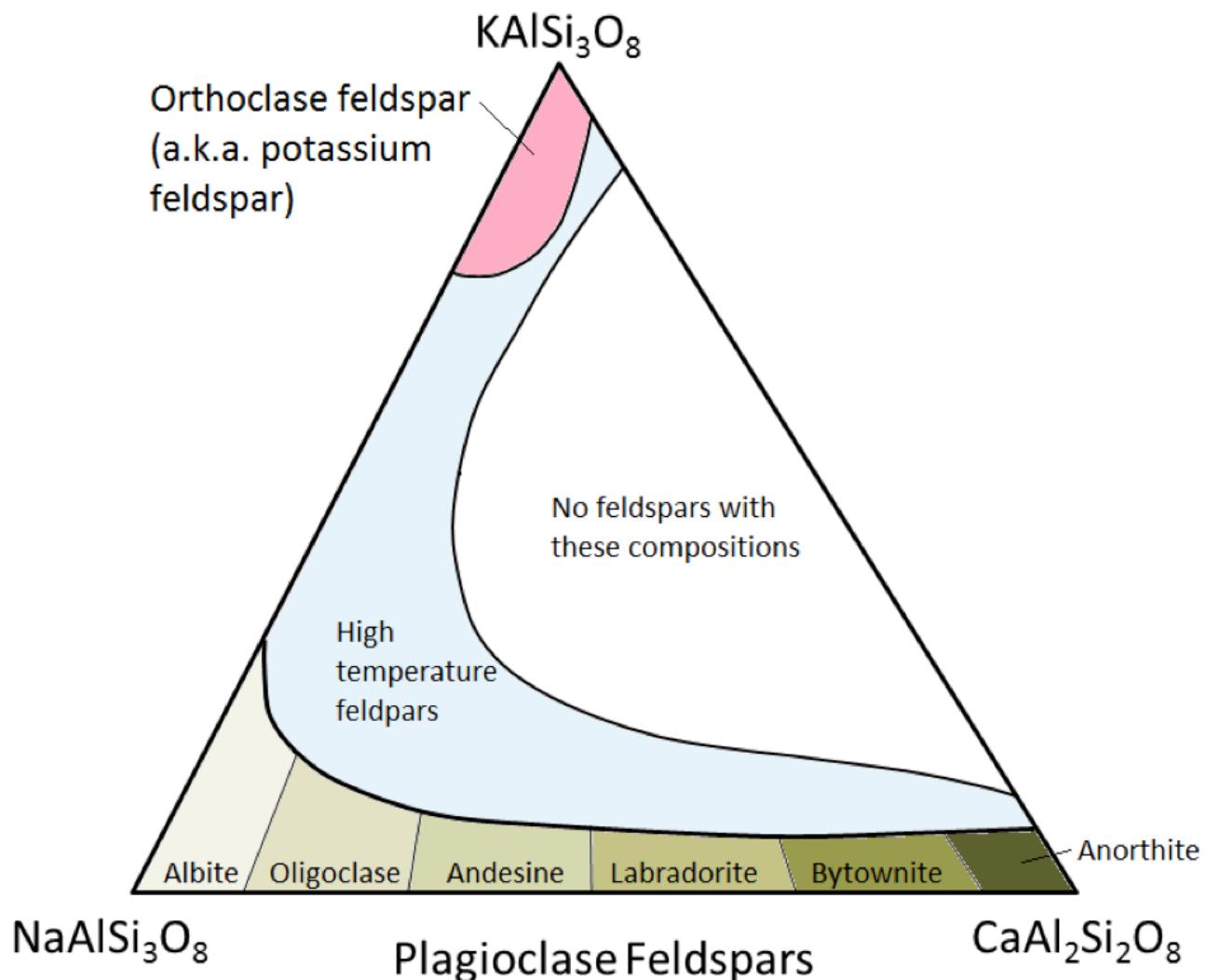


Figure 2.4.6 Compositions of the feldspar minerals.

In **quartz** ( $\text{SiO}_2$ ), the silica tetrahedra are bonded in a “perfect” three-dimensional framework. Each tetrahedron is bonded to four other tetrahedra (with an oxygen shared at every corner of each tetrahedron), and as a result, the ratio of silicon to oxygen is 1:2. Since the one silicon cation has a +4 charge and the two oxygen anions each have a -2 charge, the charge is balanced. There is no need for aluminum or any of the other cations such as sodium or potassium. The hardness and lack of cleavage in quartz result from the strong covalent/ionic bonds characteristic of the silica tetrahedron.

#### Exercise 2.5 Ferromagnesian silicates?

Silicate minerals are classified as being either ferromagnesian or non-ferromagnesian depending on whether or not they have iron (Fe) and/or magnesium (Mg) in their formula. A number of minerals and their formulas are listed below. For each one, indicate whether or not it is a *ferromagnesian silicate*.

Mineral	Formula	Ferromagnesian silicate?
olivine	$(\text{Mg}, \text{Fe})_2\text{SiO}_4$	.
pyrite	$\text{FeS}_2$	.
plagioclase feldspar	$\text{CaAl}_2\text{Si}_2\text{O}_8$	.
pyroxene	$\text{MgSiO}_3$	.
hematite	$\text{Fe}_2\text{O}_3$	.
orthoclase feldspar	$\text{KAlSi}_3\text{O}_8$	.
quartz	$\text{SiO}_2$	.
amphibole	$\text{Fe}_7\text{Si}_8\text{O}_{22}(\text{OH})_2$	.
muscovite	$\text{K}_2\text{Al}_4\text{Si}_6\text{Al}_2\text{O}_{20}(\text{OH})_4$	.
magnetite	$\text{Fe}_3\text{O}_4$	.
biotite	$\text{K}_2\text{Fe}_4\text{Al}_2\text{Si}_6\text{Al}_4\text{O}_{20}(\text{OH})_4$	.
dolomite	$(\text{Ca}, \text{Mg})\text{CO}_3$	.
garnet	$\text{Fe}_2\text{Al}_2\text{Si}_3\text{O}_{12}$	.
serpentine	$\text{Mg}_3\text{Si}_2\text{O}_5(\text{OH})_4$	.

See Appendix 3 for [Exercise 2.5 answers](#). \*Some of the formulas, especially the more complicated ones, have been simplified.

## Image Descriptions

**Figure 2.4.2 image description: The ionic radii of elements in angstroms and their charges.**

Element	Ionic Radii (in angstroms)	Charge
Oxygen	1.4	-2 (Anion)
Potassium	1.37	1 (Cation)
Calcium	1.00	2 (Cation)
Sodium	0.99	1 (Cation)
Magnesium	0.72	2 (Cation)
Iron	0.63	2 (Cation)
	0.49	3 (Cation)
Aluminum	0.39	3 (Cation)
Silicon	0.26	4 (Cation)
Carbon	0.15	4 (Cation)

[\[Return to Figure 2.4.2\]](#)

## 2.5 Formation of Minerals

In order for a mineral crystal to grow, the elements needed to make it must be present in the appropriate proportions, the physical and chemical conditions must be favourable, and there must be sufficient time for the atoms to become arranged.

Physical and chemical conditions include factors such as temperature, pressure, presence of water, pH, and amount of oxygen available. Time is one of the most important factors because it takes time for atoms to become ordered. If time is limited, the mineral grains will remain very small. The presence of water enhances the mobility of ions and can lead to the formation of larger crystals over shorter time periods.

Most of the minerals that make up the rocks around us form through the cooling of molten rock, known as **magma**. At the high temperatures that exist deep within Earth, some geological materials are liquid. As magma rises up through the crust, either by volcanic eruption or by more gradual processes, it cools and minerals crystallize. If the cooling process is rapid (minutes, hours, days, or years), the components of the minerals will not have time to become ordered and only small crystals can form before the rock becomes solid. The resulting rock will be fine-grained (i.e., with crystals less than 1 mm). If the cooling is slow (from decades to millions of years), the degree of ordering will be higher and relatively large crystals will form. In some cases, the cooling will be so fast (seconds) that the texture will be glassy, which means that no crystals at all form. **Volcanic glass** is not composed of minerals because the magma has cooled too rapidly for crystals to grow, although over time (millions of years) the volcanic glass may crystallize into various silicate minerals.

Minerals can also form in several other ways:

- Precipitation from aqueous solution (i.e., from hot water flowing underground, from evaporation of a lake or inland sea, or in some cases, directly from seawater) (see Exercise 2.7 below),
- Precipitation from gaseous emanations,
- Metamorphism: formation of new minerals directly from the elements within existing minerals under conditions of elevated temperature and pressure,
- Weathering: during which minerals unstable at Earth's surface may be altered to other minerals,
- Organic formation: formation of minerals within shells (primarily calcite) and teeth and bones (primarily apatite) by organisms (these organically formed minerals are still called minerals because they can also form inorganically).

Opal is a mineraloid (i.e., not an actual mineral) because although it has all of the other properties of a mineral, it does not have a specific structure. Pearl is not a mineral because it can *only* be produced by organic processes.

Exercise 2.7 Making crystals from solution

Place about  $\frac{1}{2}$  teaspoon ( $\sim 2.5 \text{ cm}^3$ ) of any kind of table salt into a small bowl. Add about 2 teaspoons ( $\sim 10 \text{ mL}$ ) of very hot water and swirl it around for a few minutes until all or almost all of the salt has dissolved. (Be careful not to splash yourself with the hot water.)

Place the bowl in a safe place (windowsill, bookshelf), and check back every 24 hours to see what has happened. Depending on the level of humidity in the room, you should see crystals forming within 24 hours, and all of the water should be gone, with reasonably large crystals formed, within about 3 days. They should look a little like those shown here. In other words, they should be cubes.

Now try the same experiment again, but this time put the salt and water into a small sauce pan on the stove top at the lowest heat possible. Within 10 to 20 minutes all of the water should be gone and you should be left with some very small salt crystals—too small to even see their shapes. It takes time for mineral crystals to form.



Figure 2.5.1

### Where does lithium come from?

The global demand for lithium has increased dramatically in the past decade, and will increase even more in the future as long as there is increasing demand for lithium-ion batteries in electronic devices, electric vehicles and for storage of solar- and wind-generated energy. Most of the world's lithium supply comes from salt lakes (*salars* in Spanish) like the one shown below in southwestern Bolivia.



Figure 2.5.2

The salty water of this and other lakes in the region has enough lithium in it to make it a viable source of the metal, especially because, in the dry climate, that concentration can be increased by more evaporation. When this water is evaporated lithium crystallizes out as the mineral lithium carbonate ( $\text{Li}_2\text{CO}_3$ ). For use in batteries the lithium is converted to other mineral forms—such as lithium cobalt oxide or lithium iron phosphate.

#### Media Attributions

- Figure 2.5.1: © Steven Earle. CC BY.
- Figure 2.5.2: © Isaac Earle. CC BY.

## 2.6 Mineral Properties

Minerals are universal. A crystal of hematite on Mars will have the same properties as one on Earth, and the same as one on a planet orbiting another star. That's good news for geology students who are planning interplanetary travel since we can use those properties to help us identify minerals anywhere. That doesn't mean that it's easy, however; identification of minerals takes a lot of practice. Some of the mineral properties that are useful for identification are as follows: colour, streak, lustre, hardness, crystal habit, cleavage/fracture, density and a few others.

### Colour

For most of us, colour is one of our key ways of identifying objects. While some minerals have particularly distinctive colours that make good diagnostic properties, many do not, and for many, colour is simply unreliable. The mineral sulphur (2.6.1 left) is always a distinctive and unique yellow. Hematite, on the other hand, is an example of a mineral for which colour is not diagnostic. In some forms hematite is deep dull red, but in others it is black and shiny metallic (Figure 2.6.2). Many other minerals can have a wide range of colours (e.g., quartz, feldspar, amphibole, fluorite, and calcite). In most cases, the variations in colours are a result of varying proportions of trace elements within the mineral. In the case of quartz, for example, yellow quartz (citrine) has trace amounts of ferric iron ( $\text{Fe}^{3+}$ ), rose quartz has trace amounts of manganese, purple quartz (amethyst) has trace amounts of iron, and milky quartz, which is very common, has millions of fluid inclusions (tiny cavities, each filled with water).



Sulphur



Hematite (earthy)



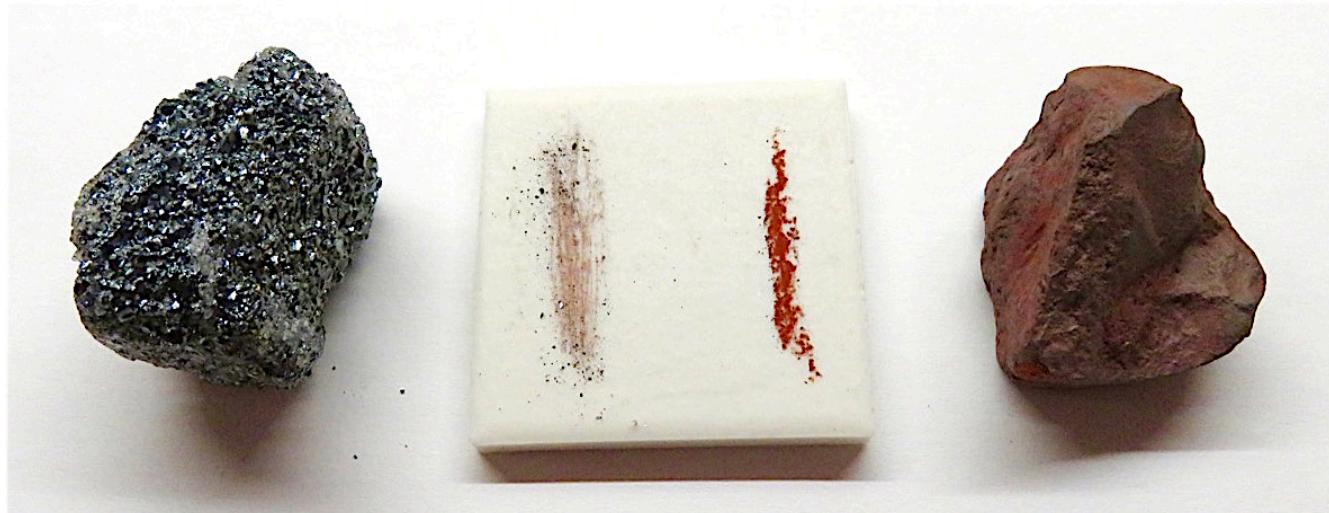
Hematite (specular)

*Figure 2.6.1 Examples of the colours of the minerals sulphur and hematite.*

### Streak

In the context of minerals, “colour” is what you see when light reflects off the surface of the sample.

One reason that colour can be so variable is that the type of surface is variable. It may be a crystal face or a fracture surface or a cleavage plane, and the crystals may be large or small depending on the nature of the rock. If we grind a small amount of the sample to a powder we get a much better indication of its actual colour. This can easily be done by scraping a corner of the sample across a streak plate (a piece of unglazed porcelain) to make a **streak**. The result is that some of the mineral gets ground to a powder and we can get a better impression of its “true” colour (Figure 2.6.2).



*Figure 2.6.2 The streak colours of specular (metallic) hematite (left) and earthy hematite (right). Hematite leaves a distinctive reddish-brown streak whether the sample is metallic or earthy.*

## Lustre

Lustre is the way light reflects off the surface of a mineral, and the degree to which it penetrates into the interior. The key distinction is between **metallic** and **non-metallic lustres**. Light does not pass through metals, and that is the main reason they look “metallic.” Even a thin sheet of metal—such as aluminum foil—will not allow light to pass through it. Many non-metallic minerals may look as if light will not pass through them, but if you take a closer look at a thin edge of the mineral you can see that it does. If a non-metallic mineral has a shiny, reflective surface, then it is called “glassy.” If it is dull and non-reflective, it is “earthy.” Other types of non-metallic lustres are “silky,” “pearly,” and “resinous.” Lustre is a good diagnostic property since most minerals will always appear either metallic or non-metallic. There are a few exceptions to this (e.g., hematite in Figure 2.6.1).

## Hardness

One of the most important diagnostic properties of a mineral is its hardness. In 1812 German mineralogist Friedrich Mohs came up with a list of 10 reasonably common minerals that had a wide range of hardnesses. These minerals are shown in Figure 2.6.3, with the Mohs scale of hardness along the bottom axis. In fact, while each mineral on the list is harder than the one before it, the relative measured hardnesses (vertical axis) are not linear. For example apatite is about three times harder than fluorite and diamond is three times harder than corundum. Some commonly available reference materials are also shown on this diagram, including a typical fingernail (2.5), a piece of copper wire (3.5), a knife

blade or a piece of window glass (5.5), a hardened steel file (6.5), and a porcelain streak plate (7). These are tools that a geologist can use to measure the hardness of unknown minerals. For example, if you have a mineral that you can't scratch with your fingernail, but you can scratch with a copper wire, then its hardness is between 2.5 and 3.5. And of course the minerals themselves can be used to test other minerals.

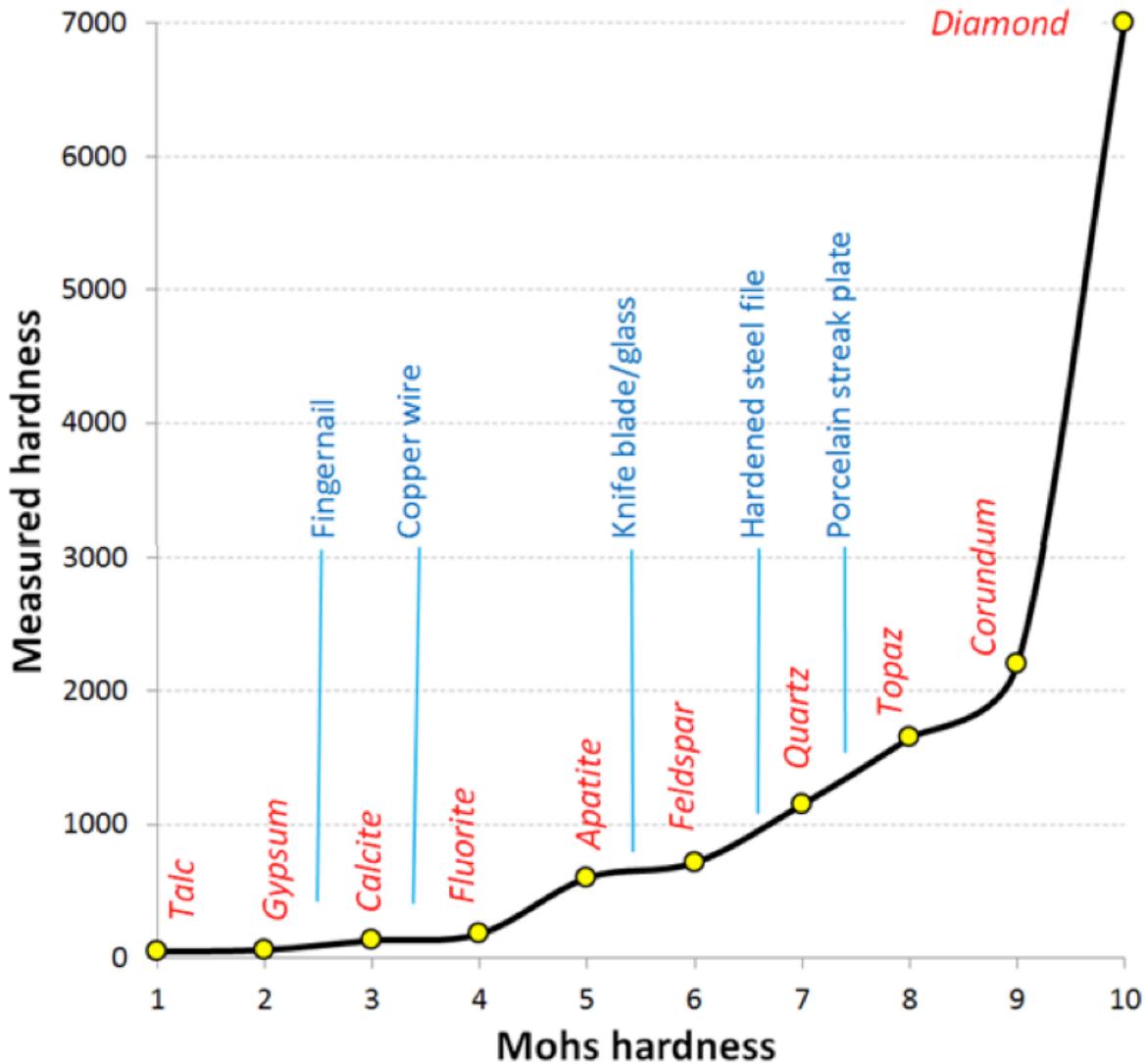


Figure 2.6.3 Minerals and reference materials in the Mohs scale of hardness. The “measured hardness” values are Vickers Hardness numbers. [\[Image Description\]](#)

## Crystal Habit

When minerals form within rocks, there is a possibility that they will form in distinctive crystal shapes if they formed slowly and if they are not crowded out by other pre-existing minerals. Every mineral has one or more distinctive crystal **habits**, but it is not that common, in ordinary rocks, for the shapes to be obvious. Quartz, for example, will form six-sided prisms with pointed ends (Figure 2.6.4a), but

this typically happens only when it crystallizes from a hot water solution within a cavity in an existing rock. Pyrite can form cubic crystals (Figure 2.6.4b), but can also form crystals with 12 faces, known as **dodecahedra** (“dodeca” means 12). The mineral garnet also forms dodecahedral crystals (Figure 2.6.4c).



*Figure 2.6.4a Hexagonal prisms of quartz.*



Figure 2.6.4b Cubic crystals of pyrite.



Figure 2.6.4c A dodecahedral crystal of garnet.

Because well-formed crystals are rare in ordinary rocks, habit isn't as useful a diagnostic feature as one might think. However, there are several minerals for which it is important. One is garnet, which is common in some metamorphic rocks and typically displays the dodecahedral shape. Another is amphibole, which forms long thin crystals, and is common in igneous rocks like granite (Figure 1.4.2).

Mineral habit is often related to the regular arrangement of the molecules that make up the mineral.

Some of the terms that are used to describe habit include bladed, botryoidal (grape-like), dendritic (branched), drusy (an encrustation of minerals), equant (similar in all dimensions), fibrous, platy, prismatic (long and thin), and stubby.

## Cleavage and Fracture

Crystal habit is a reflection of how a mineral grows, while cleavage and fracture describe how it breaks. Cleavage and fracture are the most important diagnostic features of many minerals, and often the most difficult to understand and identify. **Cleavage** is what we see when a mineral breaks along a specific plane or planes, while **fracture** is an irregular break. Some minerals tend to cleave along planes at various fixed orientations, some do not cleave at all (they only fracture). Minerals that have cleavage can also fracture along surfaces that are not parallel to their cleavage planes.

As we've already discussed, the way that minerals break is determined by their atomic arrangement and specifically by the orientation of weaknesses within the lattice. Graphite and the micas, for example, have cleavage planes parallel to their sheets (Figures 2.2.5 and 2.4.5), and halite has three cleavage planes parallel to the lattice directions (Figure 2.2.6).

Quartz has no cleavage because it has equally strong Si–O bonds in all directions, and feldspar has two cleavages at  $90^\circ$  to each other (Figure 2.6.5).

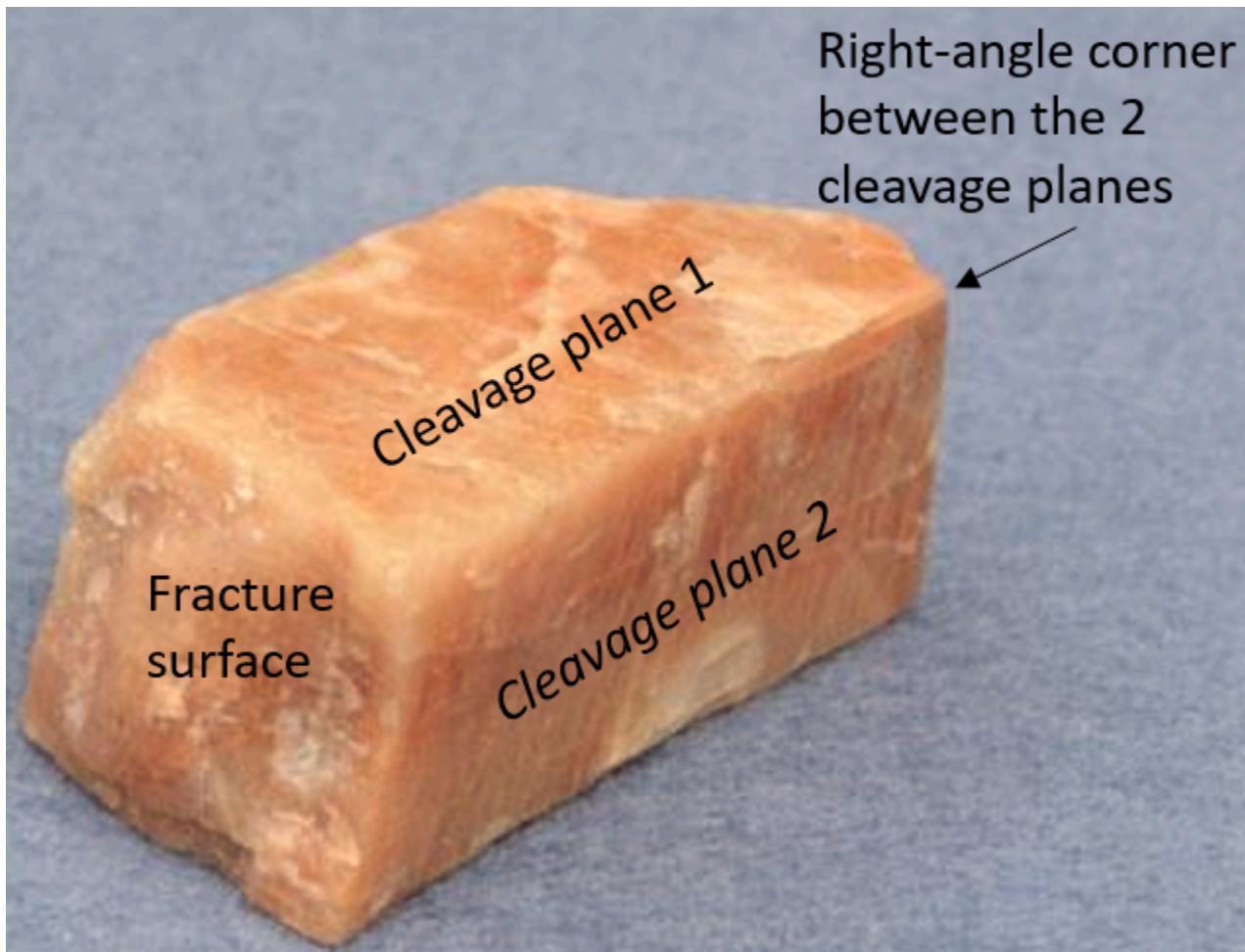
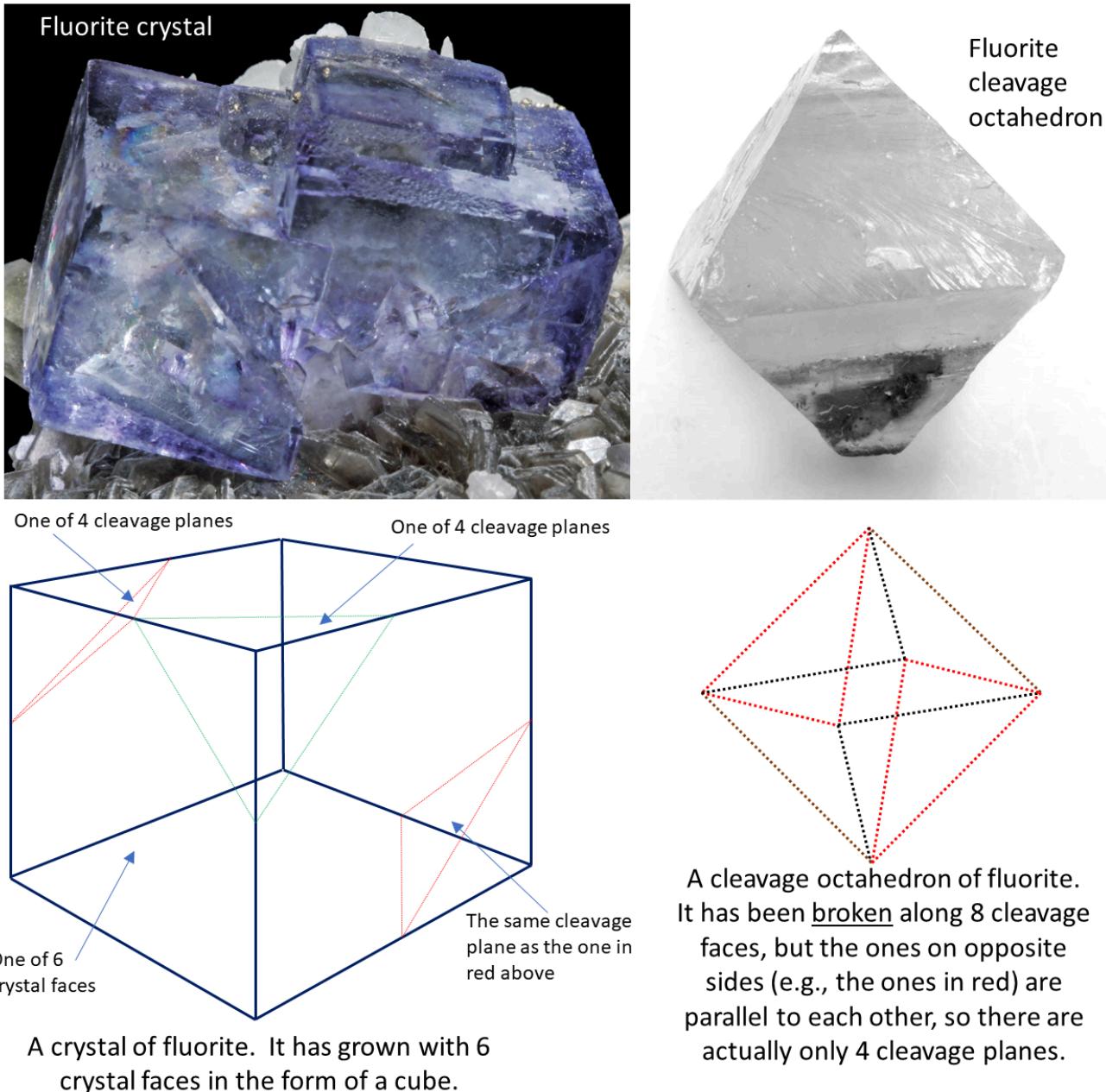


Figure 2.6.5 Cleavage and fracture in potassium feldspar

One of the main difficulties with recognizing and describing cleavage is that it is visible only in individual crystals. Most rocks have small crystals and it's very difficult to see the cleavage within those crystals. Geology students have to work hard to understand and recognize cleavage, but it's worth the effort since it is a reliable diagnostic property for most minerals.

One last thing: it is important to recognize the difference between cleavage planes and crystal surfaces. As already noted, crystal surfaces are related to how a mineral grows while cleavage planes are related to how it breaks. In most minerals cleavage planes and crystal surfaces do not align with one-another. An exception is halite, which grows in cubic crystals and has cleavage along those same planes (Figure 1.4.1 and 2.2.6). But this doesn't hold for most minerals. Quartz has crystal surfaces but no cleavage at all. Fluorite forms cubic crystals like those of halite, but it cleaves along planes that differ in orientation from the crystal surfaces. This is illustrated in Figure 2.6.6.



*Figure 2.6.6 Crystal faces and cleavage planes in the mineral fluorite. The top-left photo shows a natural crystal of fluorite. It has crystal surfaces but you can see some future cleavage planes inside the crystal. The top-right photo shows what you can create if you take a crystal like the one on the left and carefully break it along its cleavage planes.*

## Density

**Density** is a measure of the mass of a mineral per unit volume, and it is a useful diagnostic tool in some cases. Most common minerals, such as quartz, feldspar, calcite, amphibole, and mica, have what we call “average density” (2.6 to 3.0 grams per cubic centimetre ( $\text{g}/\text{cm}^3$ )), and it would be difficult to tell them apart on the basis of their density. On the other hand, many of the metallic minerals, such as pyrite, hematite, and magnetite, have densities over  $5 \text{ g}/\text{cm}^3$ . They can easily be distinguished from the lighter

minerals on the basis of density, but not necessarily from each other. A limitation of using density as a diagnostic tool is that one cannot assess it in minerals that are a small part of a rock that is mostly made up of other minerals.

## Other Properties

Several other properties are also useful for identification of some minerals. For example, calcite is soluble in dilute acid and will give off bubbles of carbon dioxide. Magnetite is magnetic, so will affect a magnet. A few other minerals are weakly magnetic.

Image Descriptions

**Figure 2.6.3 image description**

	Talc	Gypsum	Calcite	Fluorine	Apatite	Feldspar	Quartz	Topaz	Corundum	Diamond
Measured Hardness	50	60	105	200	659	700	1100	1648	2085	7000
Mohs Hardness	1	2	3	4	5	6	7	8	9	10

[\[Return to Figure 2.6.3\]](#)

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- Figure 2.6.2: © Karla Panchuk. CC BY.
- Figure 2.6.4a: [Quartz Bresil](#) © [Didier Descouens](#). CC BY.
- Figure 2.6.4b: [Pyrite cubic crystals on marlstone](#) © Carles Millan. CC BY-SA.
- Figure 2.6.4c: [Almandine garnet](#) © [Eurico Zimbres \(FGEL/UERJ\) and Tom Epaminondas \(mineral collector\)](#). CC BY-SA.

## Summary

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

Section	Summary
<a href="#"><u>2.1 Electrons, Protons, Neutrons, and Atoms</u></a>	An atom is made up of protons and neutrons in the nucleus and electrons arranged in energy shells around the nucleus. The first shell holds two electrons, and outer shells hold more, but atoms strive to have eight electrons in their outermost shell (or two for H and He). They either gain or lose electrons (or share) to achieve this, and in so doing become either cations (if they lose electrons) or anions (if they gain them).
<a href="#"><u>2.2 Bonding and Lattices</u></a>	The main types of bonding in minerals are ionic bonding (electrons transferred) and covalent bonding (electrons shared). Some minerals have metallic bonding or other forms of weak bonding. Minerals form in specific three-dimensional lattices, and the nature of the lattices and the type of bonding within them have important implications for mineral properties.
<a href="#"><u>2.3 Mineral Groups</u></a>	Minerals are grouped according to the anion part of their formula, with some common types being oxides, sulphides, sulphates, halides, carbonates, phosphates, silicates, and native minerals.
<a href="#"><u>2.4 Silicate Minerals</u></a>	Silicate minerals are, by far, the most important minerals in Earth's crust. They all include silica tetrahedra (four oxygens surrounding a single silicon atom) arranged in different structures (chains, sheets, etc.). Some silicate minerals include iron or magnesium and are called ferromagnesian silicates.
<a href="#"><u>2.5 Formation of Minerals</u></a>	Most minerals in the crust form from the cooling and crystallization of magma. Some form from hot water solutions, during metamorphism or weathering, or through organic processes.
<a href="#"><u>2.6 Mineral Properties</u></a>	Some of the important properties for mineral identification include hardness, cleavage/fracture, density, lustre, colour, and streak colour. It's critical to be able to recognize these properties in order to be able to identify minerals.

### Questions for Review

[Answers to Review Questions](#) can be found in Appendix 2.

1. What is the electrical charge on a proton? A neutron? An electron? What are their relative masses?
2. Explain how the need for an atom's outer shell to be filled with electrons contributes to bonding.

3. Why are helium and neon non-reactive?
4. What is the difference in the role of electrons in an ionic bond compared to a covalent bond?
5. What is the electrical charge on an anion? A cation?
6. What chemical feature is used in the classification of minerals into groups?
7. Name the mineral group for the following minerals:
  - calcite
  - gypsum
  - hematite
  - quartz
  - biotite
  - galena
  - graphite
  - fluorite
  - pyrite
  - orthoclase
  - magnetite
  - olivine
8. What is the net charge on an unbonded silica tetrahedron?
9. What allows magnesium to substitute freely for iron in olivine?
10. How are the silica tetrahedra structured differently in pyroxene and amphibole?
11. Why is biotite called a ferromagnesian mineral, while muscovite is not?
12. What are the names and compositions of the two end-members of the plagioclase series?
13. Why does quartz have no additional cations (other than  $\text{Si}^{+4}$ )?
14. Why is colour not necessarily a useful guide to mineral identification?
15. You have an unknown mineral that can scratch glass but cannot scratch a porcelain streak plate. What is its approximate hardness?

# 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.

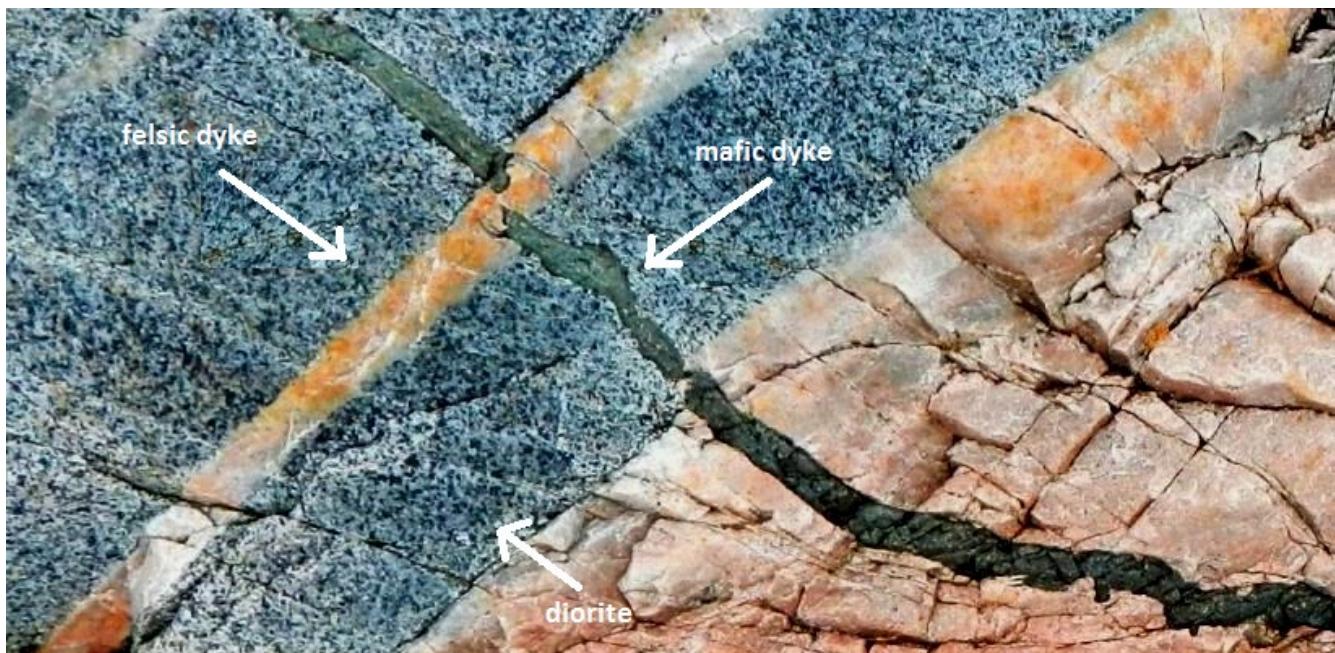


Figure 3.0.1 A fine-grained mafic dyke (dark green) intruded into a felsic dyke (pink) and into coarse diorite (grey), Quadra Island, B.C. All of these rocks are composed of more than one type of mineral. The mineral components are clearly visible in the diorite, but not in the other two rock types.

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.

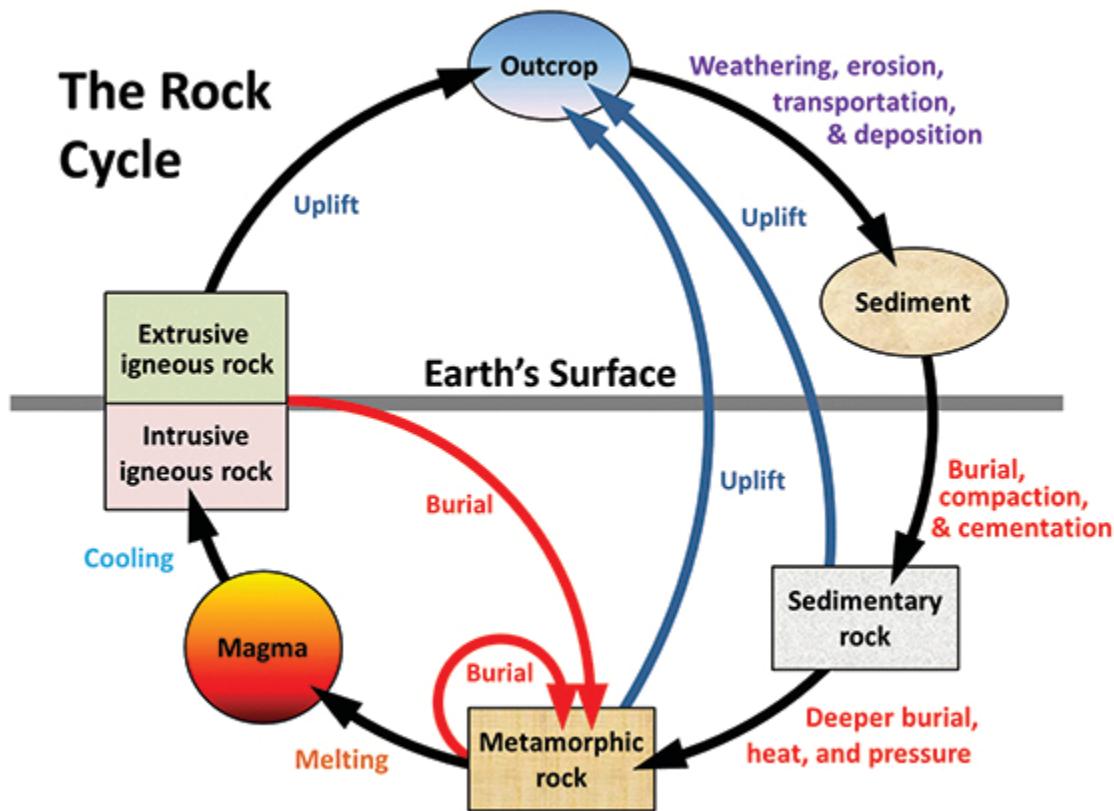


Figure 3.3 A schematic view of the rock cycle. [\[Image description\]](#)

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.

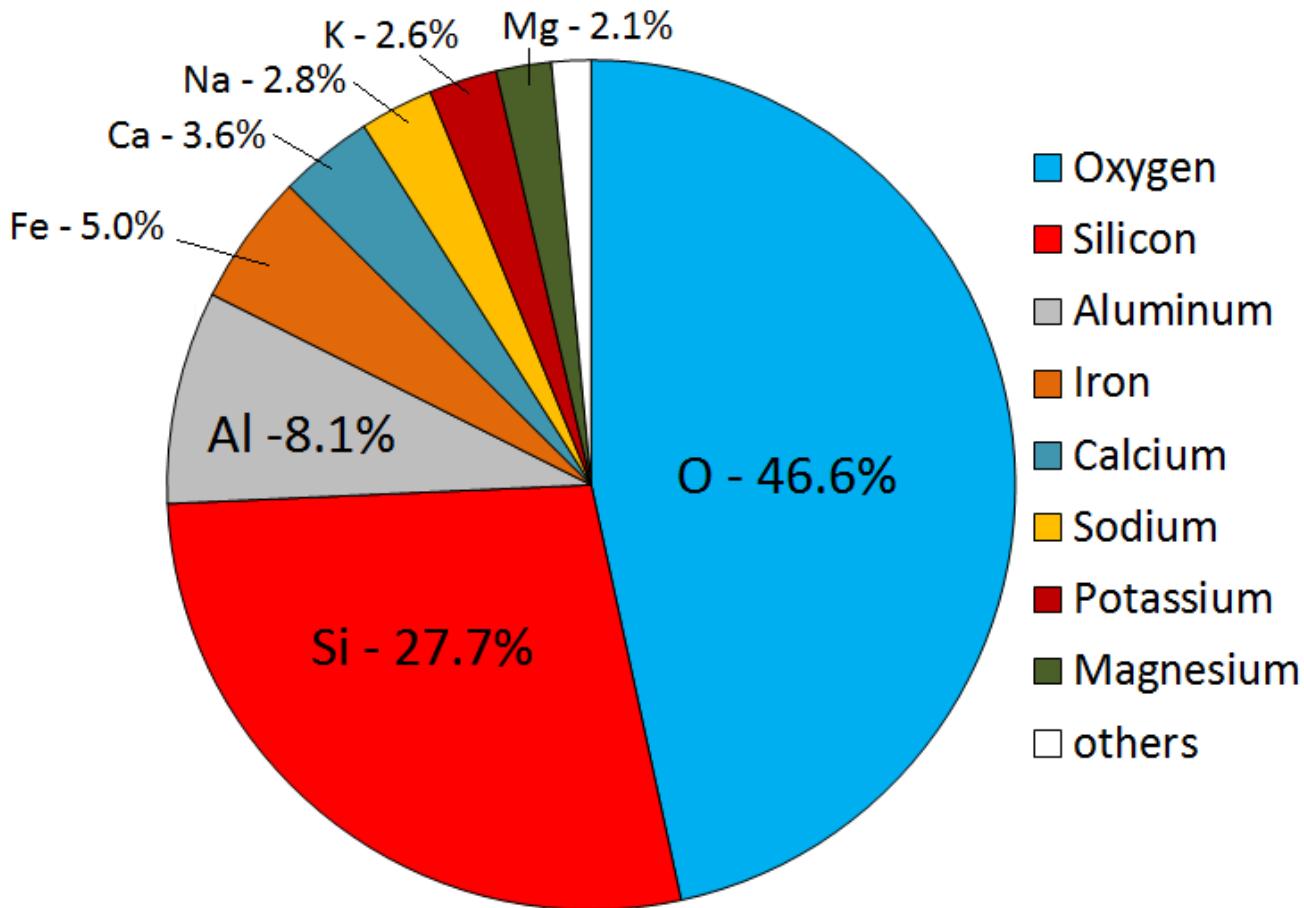
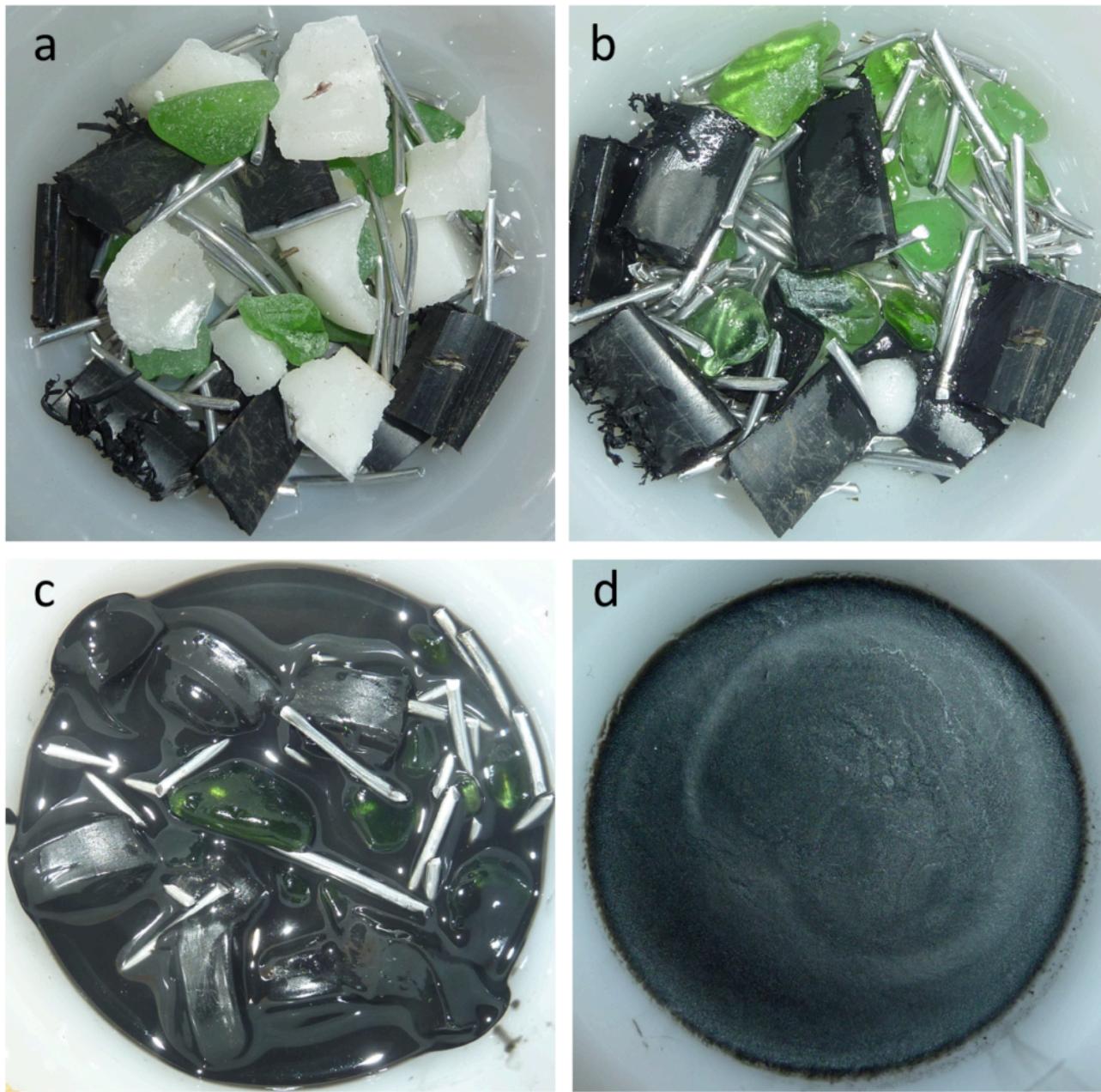


Figure 3.2.1 Average elemental proportions in Earth's crust, which is close to the average composition of magmas within the crust. [\[Image Description\]](#)

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 ( $50^{\circ}\text{C}$  will do as the melting temperature of most wax is about  $40^{\circ}\text{C}$ ) 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  $120^{\circ}\text{C}$ , 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.<sup>1</sup> 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.

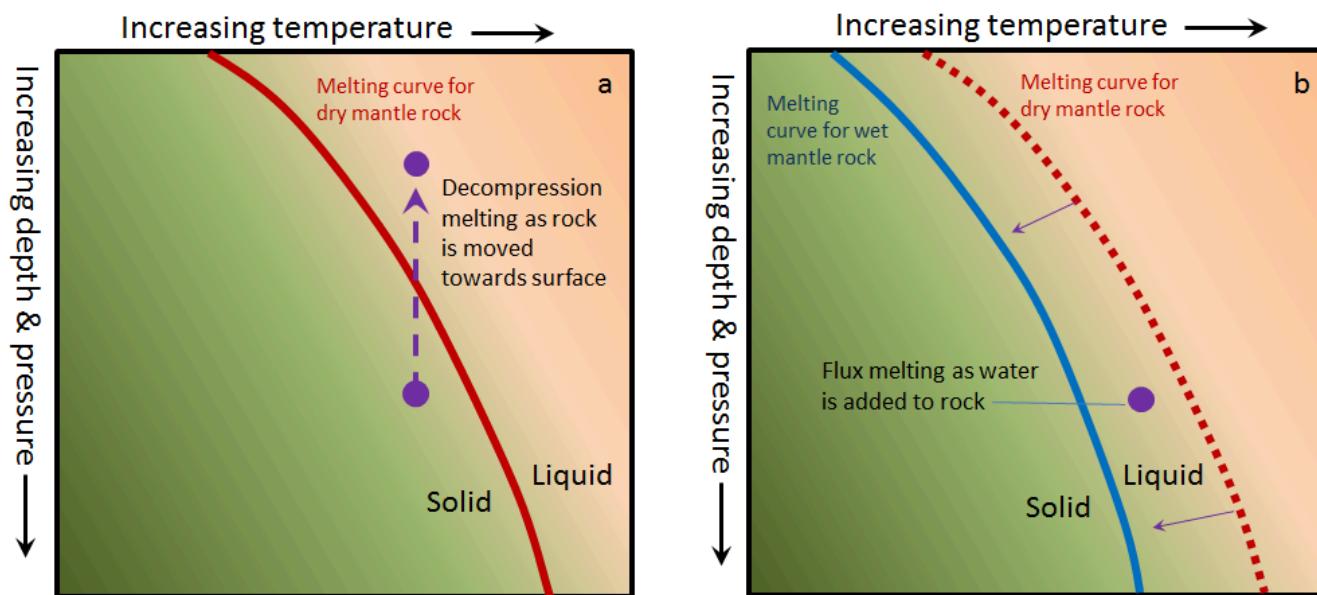
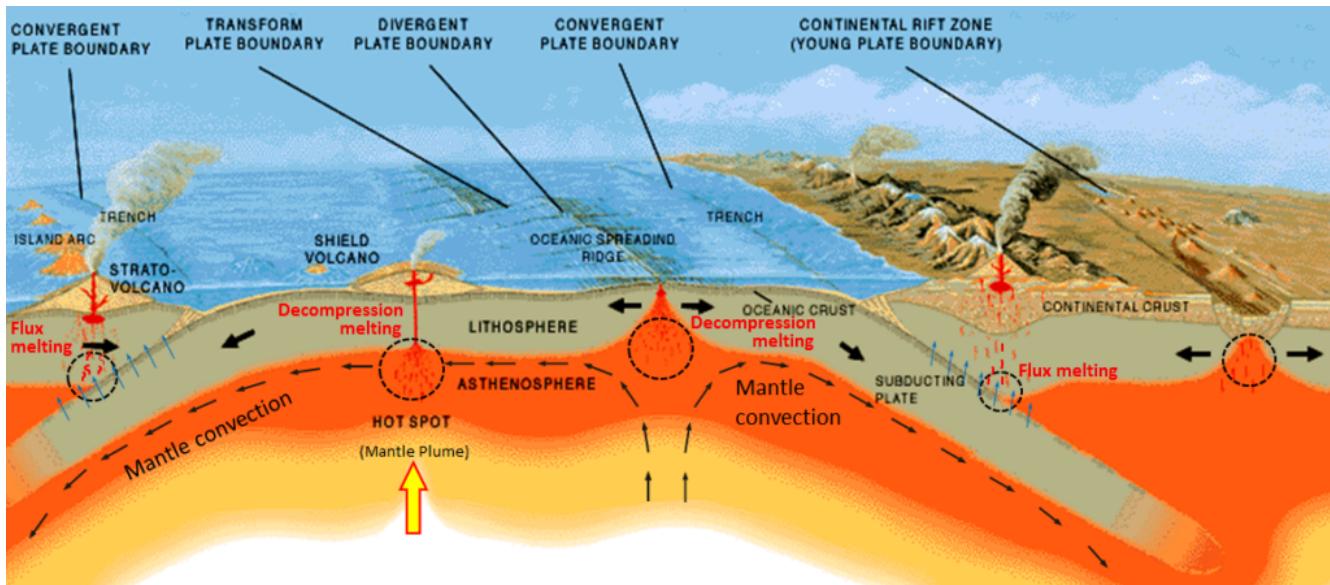


Figure 3.2.3 Mechanisms for (a) decompression melting (the rock is moved toward the surface) and (b) flux melting (water is added to the rock) and the melting curve is displaced. [\[Image Description\]](#)

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.

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  $1300^{\circ}\text{C}$ ), 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 mantle 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.

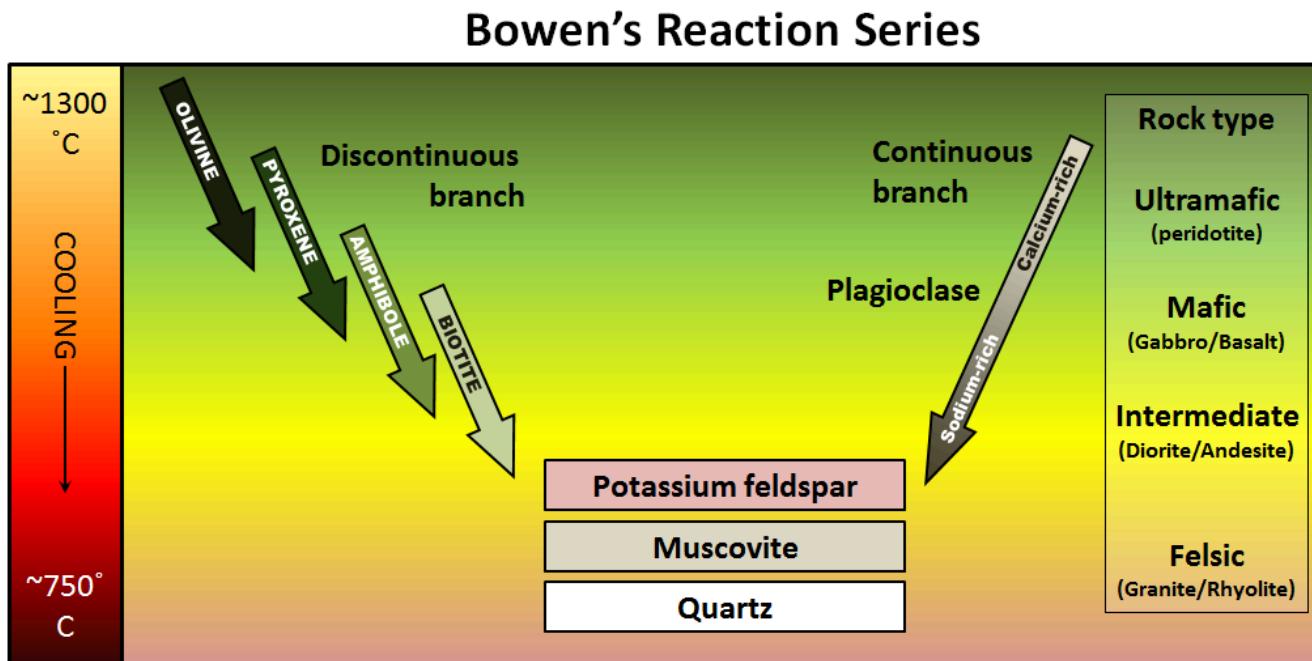


Figure 3.3.1 The Bowen reaction series describes the process of magma crystallization.

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.

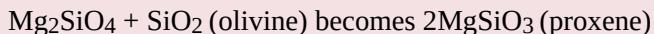
Who was Bowen, and what is a reaction series?



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 Bowen 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:



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.,  $\text{Al}_2\text{O}_3$  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.

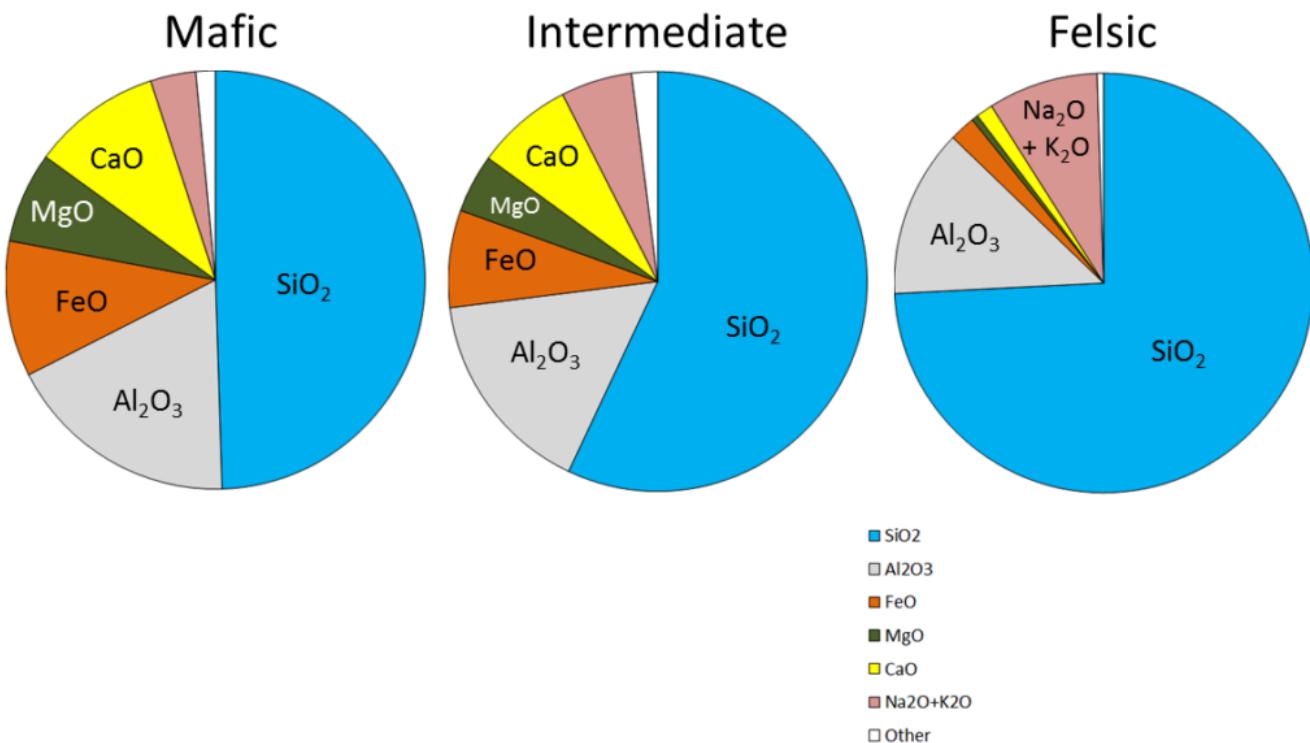


Figure 3.3.4 The chemical compositions of typical mafic, intermediate, and felsic magmas and the types of rocks that form from them.

Mafic magmas have 45% to 55%  $\text{SiO}_2$ , about 25% total of  $\text{FeO}$  and  $\text{MgO}$  plus  $\text{CaO}$ , and about 5%  $\text{Na}_2\text{O} + \text{K}_2\text{O}$ . Felsic magmas, on the other hand, have much more  $\text{SiO}_2$  (65% to 75%) and  $\text{Na}_2\text{O} + \text{K}_2\text{O}$  (around 10%) and much less  $\text{FeO}$  and  $\text{MgO}$  plus  $\text{CaO}$  (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**

<a href="#">[Skip Table]</a>			
Oxide	Felsic Magma	Intermediate Magma	Mafic Magma
SiO <sub>2</sub>	65% to 75%	55% to 65%	45% to 55%
Al <sub>2</sub> O <sub>3</sub>	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 <sub>2</sub> O	2% to 6%	3% to 7%	1% to 3%
K <sub>2</sub> O	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**

<a href="#">[Skip Table]</a>								
Rock Sample	SiO <sub>2</sub>	Al <sub>2</sub> O <sub>3</sub>	FeO	CaO	MgO	Na <sub>2</sub> O	K <sub>2</sub> O	What type of magma is it?
<b>Rock 1</b>	55%	17%	5%	6%	3%	4%	3%	
<b>Rock 2</b>	74%	14%	3%	3%	0.5%	5%	4%	
<b>Rock 3</b>	47%	14%	8%	10%	8%	1%	2%	
<b>Rock 4</b>	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).

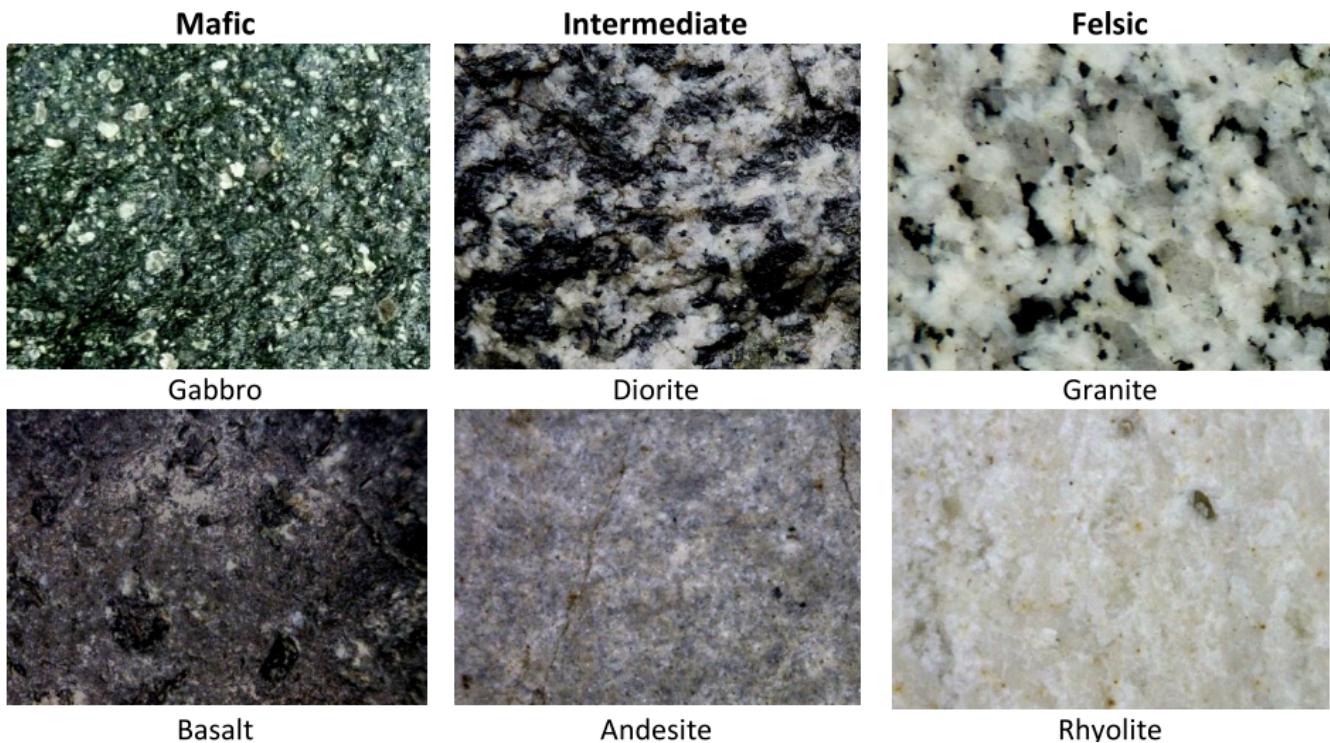


Figure 3.3.5 Examples of the igneous rocks that form from mafic, intermediate, and felsic magmas.

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). This 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).

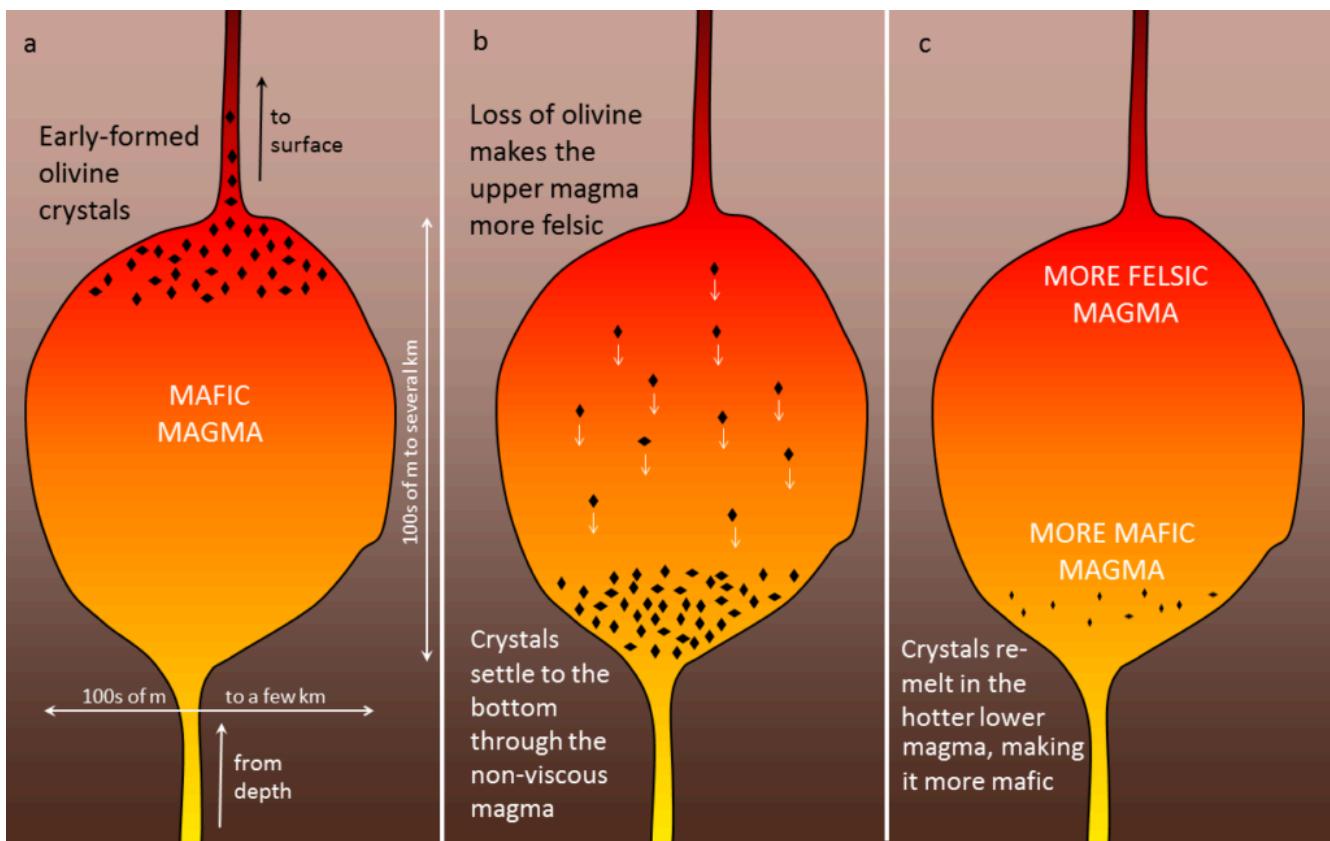


Figure 3.3.6 An example of crystal settling and the formation of a zoned magma chamber.

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.