Chapter 5. Minerals

Introduction

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 properties 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, and many more. And of course, everything made of metal is also derived from minerals.

The definition of mineral is very specific. A mineral is a naturally occurring combination of specific elements arranged in a particular repeating three-dimensional structure.

“Naturally occurring” implies that minerals are not artificially made, although many naturally occurring minerals (e.g., diamond) are also made in laboratories. That doesn’t disqualify them from being minerals.

“Specific elements” means that most minerals have a specific chemical formula or composition. The mineral pyrite, for example, is FeS\textsubscript{2} (two atoms of sulphur for each atom of iron), and any significant departure from that would make it a different mineral. But many minerals have variable compositions within a specific range. The mineral olivine, for example, can range all the way from Fe\textsubscript{2}SiO\textsubscript{4} to Mg\textsubscript{2}SiO\textsubscript{4}. Intervening compositions are written as (Fe,Mg)\textsubscript{2}SiO\textsubscript{4} meaning that Fe and Mg can be present in any proportion. 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. This regular structure means that all minerals are classified as crystals. The mineral halite is shown as an example in Figure 5.1. In halite 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

![Figure 5.1](https://physicalgeology.pressbooks.com/2/33/5/5.1.png)
composition and this structure.

The structure of halite happens to be one of the simplest mineral lattice of all. Most mineral lattices are much more complicated, as we’ll see.

Note: Element symbols such as Na and Cl are used extensively in this book. In Appendix 1 you can find a list of the symbols, the names of the elements common in minerals, and a copy of the periodic table of elements.

5.1 Electrons, Protons, Neutrons, and Atoms

Protons Are What Make Elements Distinct

All matter, including mineral crystals, is made up of atoms. All atoms are made up of three main particles: protons, neutrons, and electrons. Protons have a positive charge (+1), neutrons have no charge, and electrons have a negative (-1) charge. Protons and neutrons have the same mass, while electrons have a mass that is 10,000 times smaller.

The element hydrogen (H) has the simplest atoms, each with just one proton and one electron. The proton forms the nucleus (the centre of the atom), while the electron orbits around it. All other elements have neutrons as well as protons in their nucleus. Because protons all have a +1 charge, they repel each other. It is possible to have more than one proton in a nucleus because neutrons hold them together.

What makes an element distinct is the number of protons it has, so it is fitting that the number of protons is referred to as the atomic number. The total number of protons and neutrons is the atomic mass, which is fitting because the protons and neutrons are what account for nearly all of the mass of the atom. So for H, the atomic number is 1 and the atomic mass is 1 because there is one proton and no neutrons. For helium (He), which has 2 protons and 2 neutrons, the atomic number is 2 and the atomic mass is 4.

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. This is because these atoms have a large number of protons concentrated in a very small space, and those protons are repelling each other. Extra neutrons are needed to keep the nucleus together. An example is uranium. Its most common isotope has an atomic mass of 238. It has 92 protons, but requires 146 neutrons to keep them together. The neutrons are only partly successful. Uranium is radioactive, meaning that its nucleus tends to come apart and release energy.

Electrons Are What Control How Atoms Interact

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, but the next shell holds up to eight electrons. An atom can have many shells of electrons, but there are never more than 8 outermost electrons interacting with surrounding atoms.

The outermost electrons play an important role in bonding between atoms. Elements that have a full outer shell are inert because they do not react with other elements to form compounds. These are the noble gases which all appear in the far-right column of the periodic table (helium, neon, argon, etc). For elements that do not have a full outer shell, the outermost electrons can interact with the outermost electrons of nearby atoms to create chemical bonds. The electron shell configurations for the first 26 elements are listed in Table 2.1. Note
that some of the shells in the table below have more than 8 electrons. This is because they contain subshells. For example, the third shell can hold up to 18 electrons because it contains one subshell that can hold 2 electrons, and two subshells that can hold 8 electrons each.

### Table 2.1 Electron shell configurations of some elements

*Inert elements with filled outer shells are in shading.*

<table>
<thead>
<tr>
<th>Element</th>
<th>Symbol</th>
<th>Atomic Number</th>
<th>1st shell</th>
<th>2nd shell</th>
<th>3rd shell</th>
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<td>Beryllium</td>
<td>Be</td>
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</tr>
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<td>B</td>
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<td>2</td>
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<tr>
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<td>2</td>
<td>7</td>
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<td>1</td>
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<td>2</td>
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<td>8</td>
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<td>2</td>
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<td>26</td>
<td>2</td>
<td>8</td>
<td>14</td>
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</tr>
</tbody>
</table>
5.2 Bonding and Lattices

Atoms seek to have a full outer shell. That means there will be two electrons for hydrogen and helium, and eight electrons for the other elements. This is accomplished by transferring or sharing electrons with other atoms in chemical bonds. The type of chemical bond is important for the study of minerals because the type of bond will determine many of a mineral’s physical and chemical properties.

Ionic Bonds

Consider the example of halite again, which is made up of sodium (Na) and chlorine (Cl). Na has 11 electrons: two in the first shell, eight in the second, and one in the third (Figure 5.2, top). Na readily gives up the third shell electron so it can have the second shell with 8 electrons as its outermost shell. When it loses the electron, the total charge from the electrons is -10, but the total charge from the protons is +11, so it is left with a +1 charge over all.

Chlorine has 17 electrons: two in the first shell, eight in the second, and seven in the third. Cl readily accepts an eighth electron to fill its third shell, and therefore becomes negatively charged because it has a total charge of -18 from electrons, and a total charge of +17 from protons.

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 5.2, bottom). Since negative and positive charges attract, sodium and chlorine ions stick together, creating an ionic bond. Electrons can be thought of as being transferred from one atom to another in an ionic bond.

![Figure 5.2 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). [Steven Earle CC-BY 4.0]](https://physicalgeology.pressbooks.com)

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1 You can remember that a cation is positive by remembering that a cat has paws (paws sounds like "pos" in "positive"). You could also think of the "t" in "cation" as a plus sign.
Exercise 5.1 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 (−), and what charge it would have (e.g., +1, +2, −1). The first one is done for you.

Fluorine (9)  anion (−1)
Lithium (3)  
Magnesium (12)  
Argon (18)  
Chlorine (17)  
Beryllium (3)  
Oxygen (8)  
Sodium (11)
Covalent Bonds

An element like chlorine can also form bonds without forming ions. For example, two chlorine atoms can each complete their outer shells by sharing electrons. Chlorine gas (Cl₂, Figure 5.3) is formed when two chlorine atoms form a covalent bond.

Carbon is another atom that participates in covalent bonding. An uncharged carbon atom has six protons and six electrons. Two of the electrons are in the inner shell and four are in the outer shell (Figure 5.4, left). Carbon would need to gain or lose four electrons to have a filled outer shell, and this would create too great a charge imbalance. Instead, carbon atoms share electrons to create covalent bonds.

In the mineral diamond (Figure 5.5, left), the carbon atoms are linked together in a three-dimensional framework, where one carbon atom is bonded to four other carbon atoms and every bond is a very strong covalent bond.

Figure 5.3 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. [Steven Earle CC-BY 4.0]

Figure 5.4 The electron configuration of carbon (left) and the sharing of electrons in covalent C bonding (right). The electrons shown in blue are shared between adjacent C atoms. [Steven Earle CC-BY 4.0]

Figure 5.5 Covalently bonded structures. Left: Diamond with three-dimensional structure of covalently bonded carbon. Right: Graphite with covalently bonded sheets of carbon. [Materialscientist CC-BY-SA 3.0, http://bit.ly/1Pe6res]
Other Types of Bonds

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. Consider the mineral graphite (Figure 5.5, right): the carbon atoms are linked together in sheets or layers in which each carbon atom is covalently bonded to three others. Graphite-based compounds are strong because of the covalent bonding between carbon atoms within each layer, which is why they are used in high-end sports equipment such as ultralight racing bicycles. Graphite itself is soft, however, because the layers themselves are held together by relatively weak Van der Waals bonds. Van der Waals bonds, like hydrogen bonds (Figure 5.6), work because molecules or atoms can be electrostatically neutral, but still have an end that is more positive and an end that is more negative. The positive and negative ends attract each other.

Metallic bonding occurs in metallic elements because they have outer electrons which 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).

5.3 Mineral Groups

Minerals are organized according to the anion or anion group (a group of atoms with a net negative charge, e.g., SO\text{4}^{2-}) they contain, because the anion or anion group has the biggest effect on the properties of the mineral. Silicates, with the anion group SiO\text{4}^{4-}, are by far the most abundant group in the crust and mantle. They will be discussed in detail in section 5.4. The different anionic mineral groups along with examples of minerals in each group are summarized below.

Oxides (O^{2-})

Oxide minerals (Figure 5.7) have oxygen (O^{2-}) as their anion. They don't include anion groups with other elements, such as the carbonate (CO\text{3}^{2-}), sulphate (SO\text{4}^{2-}), and silicate (SiO\text{4}^{4-}) anion groups. The iron oxides hematite (Fe\text{2}O\text{3}) and magnetite (Fe\text{3}O\text{4}) are two examples which are important ores of iron. Corundum (Al\text{2}O\text{3}) is 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\text{+}) the mineral is known as a hydroxide. Some important hydroxides are limonite and bauxite, which are ores of iron and aluminium, respectively. Frozen water (H\text{2}O) is an oxide mineral, but liquid water is not because it doesn't have a regular lattice.
Figure 5.7 Oxide minerals include metal ore minerals, industrial minerals, and gemstones. [Karla Panchuk with photos by R. Weller/Cochise College. Permission for non-commercial educational use.]

Figure 5.8 Sulphide minerals often have a metallic lustre and include metal ores. [Karla Panchuk with photos by R. Weller/Cochise College. Permission for non-commercial educational use.]
Sulphides ($S^{2-}$)

**Sulphide** minerals (Figure 5.8) include galena (PbS), sphalerite (ZnS), chalcopyrite (CuFeS$_2$), and molybdenite (MoS$_2$), which are the most important ores of lead, zinc, copper, and molybdenum, respectively. Some other sulphide minerals are pyrite (FeS$_2$), bornite (Cu$_9$FeS$_8$), stibnite (Sb$_2$S$_3$), and arsenopyrite (FeAsS). Sulphide minerals tend to have a metallic shine (lustre) to them.

Sulphates ($SO_4^{2-}$)

Many sulphate minerals (Figure 5.9) form when water evaporates which contains the sulphate anion group. Sometimes a deposit of sulphate minerals tells us that a lake or sea has dried up. Sulphates include anhydrite (CaSO$_4$) and its cousin gypsum (CaSO$_4$$\cdot$2H$_2$O), as well as the sulphates of barium and strontium: barite (BaSO$_4$) and celestite (SrSO$_4$). In all of these minerals, the cation has a +2 charge, which balances the −2 charge on the sulphate ion.

Halides (anions of the halogen group)

The halides (Figure 5.10) are so named because the anions include the halogen elements chlorine (Cl$^-$), fluorine (F$^-$), bromine (Br$^-$), etc. Examples are halite (NaCl), cryolite (Na$_3$AlF$_6$), and fluorite (CaF$_2$). Halide minerals are made of ionic bonds, and like the sulphates, come from the evaporation of mineral-rich water.
Carbonates (CO$_3^{2-}$)

The carbonate anion group combines with +2 cations to form minerals such as calcite (CaCO$_3$), magnesite (MgCO$_3$), dolomite ((Ca,Mg)CO$_3$), and siderite (FeCO$_3$). The copper minerals malachite and azurite are also carbonates. The carbonate mineral calcite is the main component of rocks formed by organisms such as corals and algae in ancient seas.

![Carbonate minerals](https://physicalgeology.pressbooks.com)

Phosphates (PO$_4^{3-}$)

The apatite group of phosphate minerals (Ca$_5$(PO$_4$)$_3$(Fl,Cl,OH)) includes hydroxyapatite, which makes up the enamel of your teeth. Turquoise is also a phosphate mineral (Figure 5.12).

![Phosphate minerals](https://physicalgeology.pressbooks.com)
Silicates ($\text{SiO}_4^{4-}$)

The silicate minerals include the elements silicon and oxygen in varying proportions. These are discussed at length in Section 5.4.

Native minerals

These are minerals made of a single element, such as gold, copper, silver, sulphur, and graphite (Figure 5.13).

Exercise 5.2 Mineral Groups

We classify minerals according to the anion part of the mineral formula, and mineral formulas are always written with the anion part last. 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 $\text{O}$. 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 $\text{S}$) 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:

<table>
<thead>
<tr>
<th>Mineral and Formula</th>
<th>Group</th>
<th>Mineral and Formula</th>
<th>Group</th>
</tr>
</thead>
<tbody>
<tr>
<td>sphalerite ($\text{ZnS}$)</td>
<td></td>
<td>fluorite ($\text{CaF}_2$)</td>
<td></td>
</tr>
<tr>
<td>magnetite ($\text{Fe}_3\text{O}_4$)</td>
<td></td>
<td>ilmenite ($\text{FeTiO}_3$)</td>
<td></td>
</tr>
<tr>
<td>pyroxene ($\text{MgSiO}_3$)</td>
<td></td>
<td>siderite ($\text{FeCO}_3$)</td>
<td></td>
</tr>
<tr>
<td>anglesite ($\text{PbSO}_4$)</td>
<td></td>
<td>feldspar ($\text{KAISi}_3\text{O}_8$)</td>
<td></td>
</tr>
<tr>
<td>sylvite ($\text{KCl}$)</td>
<td></td>
<td>sulphur ($\text{S}$)</td>
<td></td>
</tr>
<tr>
<td>silver ($\text{Ag}$)</td>
<td></td>
<td>xenotime ($\text{YPO}_4$)</td>
<td></td>
</tr>
</tbody>
</table>
5.4 Silicate Minerals

Silicon and oxygen bond covalently to create a silica tetrahedron, which is a four-sided pyramid shape with O at each corner and Si in the middle (Figure 5.7). This structure is the building block of the many important silicate minerals. 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 (SiO$_4$) is –4.

In silicate minerals, these tetrahedra are arranged and linked together in a variety of ways, from single units to chains, rings, and more complex frameworks. In the rest of this section we will discuss the structures of the most common silicate minerals in Earth’s crust and mantle.

**Exercise 5.3 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. Either secure the tabs with glue or tape to the tetrahedron to hold it together, or else make a slice along the thin grey line and insert the pointed tab into the slit. If you’re feeling ambitious, make several tetrahedra and try joining them into pairs, rings, single and double chains, sheets, and even three-dimensional frameworks.

**Olivine Consists of Isolated Tetrahedra Linked by Cations**

The simplest silicate structure, that of the mineral olivine (Figure 5.15), is composed of isolated tetrahedra bonded to iron and/or magnesium ions (Figure 5.15 left). In olivine, the –4 charge of each silica tetrahedron is balanced by two iron or magnesium cations, each with a charge of +2. Olivine can be pure Mg$_2$SiO$_4$ or Fe$_2$SiO$_4$,
or some combination of the two, written as (Mg,Fe)$_2$SiO$_4$. The magnesium and iron cations are close in radius (0.73 Å versus 0.62 Å $^2$). Because of this size similarity, and because they both have a charge of +2, iron and magnesium can readily substitute for each other in olivine and in many other minerals.

![Olivine](https://physicalgeology.pressbooks.com/2e-chapter5/images/Olivine.png)

Figure 5.15  Olivine is a silicate mineral made of isolated silica tetrahedra bonded to Fe and Mg ions (left; modified after Steven Earle bit.ly/OlivineXtl). Olivine crystals (centre) can often be found in the volcanic igneous rock called basalt (right). [Karla Panchuk; R. Weller photos by permission for NC use]

Although the iron and magnesium ions are similar in size, allowing them to substitute for each other in some silicate minerals, the common ions in silicate minerals have a wide range of sizes (Figure 5.16). Ionic radii are critical to the composition of silicate minerals, so we’ll be referring to this diagram again.

![Ionic Radii](https://physicalgeology.pressbooks.com/2e-chapter5/images/IonicRadii.png)

Figure 5.16 The ionic radii (sizes in angstroms) of some of the common ions in silicate minerals shown to scale. Notice that iron appears twice with two different radii. This is because iron can exist as 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. [Karla Panchuk after Steven Earle]

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2 Å stands for Angstrom, a unit commonly used to express atomic-scale dimensions. One angstrom is $10^{-10}$ m or 0.0000000001 m.
Pyroxene and Amphibole Are Chain Silicates

Pyroxene (Figure 5.17 bottom left) is an example of a single-chain silicate. The structure of chain silicates is shown in Figure 5.17 (top). In pyroxene, silica tetrahedra form a chain because one oxygen from each tetrahedron is shared with the adjacent tetrahedron. This means there are fewer oxygens in the structure. This can be expressed as an oxygen-to-silicon ratio (O:Si). The O:Si is lower than in olivine (3:1 instead of 4:1), and the net charge per silicon atom is less (–2 instead of –4), since fewer cations are necessary to balance that charge. Pyroxene compositions have the silica tetrahedra represented as SiO$_3$, e.g., MgSiO$_3$, FeSiO$_3$, and CaSiO$_3$.) In other words, pyroxene has one cation for each silica tetrahedron (e.g., MgSiO$_3$) while olivine has two (e.g., Mg$_2$SiO$_4$). 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 calcium cations (radius 1.00 Å) substitute for iron (0.63 Å) and magnesium (0.72 Å).

3 The variation in composition can also be written as (Mg,Fe,Ca)SiO$_3$, where the elements in the brackets can be present in any proportion.
In **amphibole** (Figure 5.17 bottom right), the silica tetrahedra are linked in a double chain (Figure 5.17 top) 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 permisive than pyroxene and its compositions can be very complex, as shown by the formula for the amphibole called hornblende in Figure 5.17 (bottom right).

**Exercise 5.4 Silicon to Oxygen Ratios**

The diagram on the left 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. Confirm for yourself that the ratio of silicon to oxygen for the single chain is 3:1.

The diagram on the right represents a double chain in a silicate mineral. Again, count the number of tetrahedra versus the number of oxygen ions. What is the ratio of Si to O in double-chain silicates (e.g., amphibole)?

**Mica and Clay Minerals Are Sheet Silicates**

In **mica** structures the silica tetrahedra are arranged in continuous sheets (Figure 5.18 top), where each tetrahedron shares three oxygen anions with adjacent tetrahedra. Because more oxygens are shared between adjacent tetrahedra, fewer charge-balancing cations are needed for sheet silicate minerals. Bonding between sheets is relatively weak, and this accounts for the tendency of mica minerals to split apart in sheets (referred to as one-directional cleavage; Figure 5.18 bottom right). Two common micas in silicate rocks are **biotite** (Figure 5.18 bottom left), which contains iron and/or magnesium, making it a dark mineral; and **muscovite** (Figure 5.18 right) which contains aluminum and potassium, and is light in colour.

Apart from muscovite, biotite, and chlorite, there are many other **sheet silicates** (or **phyllosilicates**), which usually exist as clay-sized fragments (i.e., less than 0.004 mm). These include the clay minerals **kaolinite**, **illite**, and **smectite**, and although they are difficult to study because they are occur as very small particles, they are extremely important components of rocks and especially of soils. All of the sheet silicate minerals also have water in their structure. It occurs as the OH⁻ anion.

**Quartz and Feldspar Are Framework Silicates**

In framework silicates, three or four oxygen on a silica tetrahedron are shared with other silica tetrahedra. This means the tetrahedra are connected to each other in three dimensions rather than in two-dimensional chains and sheets.
Figure 5.18 Micas are sheet silicates (top) and split easily into thin layers along planes parallel to the sheets. Biotite mica (lower left) is dark because it contains Fe and Mg cations. Muscovite mica (lower right) does not have these cations and so is light in colour. It has Al and K instead. The muscovite mica shows how thin layers can split away in a sheet silicate. [Karla Panchuk; R. Weller photos by permission for NC use]

**Feldspars** (Figure 5.19) are a very important group of minerals with three-dimensional frameworks. The three main feldspar minerals are **potassium feldspar**, (also referred to as **K-feldspar** or K-spar; Figure 5.19 right) and two types of **plagioclase feldspar**: **albite** (NaAlSi$_3$O$_8$; Figure 5.19 left) and **anorthite** (CaAl$_2$Si$_3$O$_8$; Figure 5.19 centre). As is the case for iron and magnesium in olivine, there is a continuous range of compositions (referred to as a solid solution series) between albite and anorthite in plagioclase. This is because the calcium and sodium ions are almost identical in size (1.00 Å versus 0.99 Å). Any intermediate compositions between CaAl$_2$Si$_3$O$_8$ and NaAlSi$_3$O$_8$ can exist. This is a little bit surprising because calcium and sodium ions don’t have the same charge (Ca$^{2+}$ versus Na$^+$. This problem is accounted for substituting some Al$^{3+}$ for Si$^{4+}$. Albite has one Al and three Si, while anorthite has two Al and two Si. Plagioclase feldspars of intermediate composition have intermediate proportions of Al and Si.
K-feldspar (KAlSi₃O₈; Figure 5.19 right) has a slightly different structure than that of plagioclase, owing to the larger size of the potassium ion (1.37 Å). It is possible for potassium and sodium to substitute for each other, but this requires very high temperatures.

Quartz (SiO₂; Figure 5.20) contains only silica tetrahedra. In quartz, each silica tetrahedron is bonded to four other tetrahedra (with an oxygen shared at every corner of each tetrahedron), making a three-dimensional framework. 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 to add cations to balance the charge. The hardness of quartz and the fact that it breaks irregularly (notice the bottom of the crystal in Figure 5.20 right) and not along smooth planes result from the strong covalent/ionic bonds characteristic of the silica tetrahedron.
5.5 Formation of Minerals

In order for a mineral crystal to grow the following are required:

- The elements needed to make it must be present in the appropriate proportions.
- The physical and chemical conditions must be favourable.
- There must be sufficient time for the atoms to become arranged.

Physical and chemical conditions include factors such as temperature, pressure, amount of oxygen available, pH, and the presence of water. The presence of water makes it easier for ions to move to where there are needed, and can lead to the formation of larger crystals over shorter time periods. Time is one of the most important factors because it takes time for atoms to line themselves up into an orderly structure. If time is limited, the mineral grains will remain very small.

Most of the minerals that make up the rocks in the crust and mantle formed 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., 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 (e.g., from hot water flowing underground, from evaporation of a lake or inland sea, or in some cases, directly from seawater)
- Precipitation from gas (e.g., from vents releasing volcanic gases)
- 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.

5.6 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, habit, cleavage/fracture, and density.
Colour

For most of us, colour is a key way to identify an object. While some minerals have particularly distinctive colours that make good diagnostic properties, many do not. For some minerals colour is simply unreliable. The mineral sulphur (Figure 5.21 left) always has the same characteristic yellow colour. Hematite (Figure 5.21 right), on the other hand, is an example of a mineral for which colour is not necessarily diagnostic. In some forms hematite is a distinctive deep dull red. But there are also silvery metallic forms which could be mistaken for other minerals with a similarly silvery metallic appearance.

For other minerals, the problem is that a single mineral can have a wide range of colours. In most cases, the variations in colours are a result of varying proportions of trace elements within the mineral. In the case of quartz (Figure 5.22), yellow quartz (citrine) has trace amounts of ferric iron (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). Smoky quartz gets its colour by being exposed to natural radiation.

![Figure 5.21 Colour is a useful diagnostic property for sulphur (left) and for some types of hematite (right) because the yellow and dark red colours are unique to those minerals. In contrast, silvery metallic forms of hematite are similar in appearance to many other minerals. [Karla Panchuk with photos by R. Weller/Cochise College (permission for non-commercial educational use).]

![Figure 5.22 The many colours of quartz. [Karla Panchuk; photos by R. Weller/ Cochise College with permission for NC use]
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. A way to get around this problem is to grind a small amount of the sample to a powder and observe the *streak*—the colour of the powder. This is done by scraping the sample across a streak plate (a piece of unglazed porcelain; Figure 5.23). Figure 5.23 shows the same reddish-brown streak for two samples of hematite, even though one sample is metallic and the other is deep red.

![Figure 5.23 Hematite leaves a distinctive reddish-brown streak whether the sample is specular (metallic) or earthy. [Karla Panchuk]](image)

Streak is an especially helpful property when minerals look similar. In Figure 5.24 all of the minerals are dark in colour, with varying degrees of metallic lustre. The streaks of the minerals are much more distinctive.

![Figure 5.24 Similar-looking metallic and sub-metallic minerals leave different colour streaks. The minerals are from upper left clockwise: hematite, magnetite, sphalerite, and galena. [Karla Panchuk]](image)

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 (e.g., the hematite on left of Figure 5.23). Even a thin sheet of
metal — such as aluminum foil — will prevent light from passing 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, it is said to have a **glassy** lustre. The quartz crystals in Figure 5.22 are examples of minerals with glassy lustre. If the mineral surface is dull and non-reflective, it has an **earthy** lustre (like the hematite on the right of Figure 5.23). Other types of non-metallic lustres are silky, pearly, and resinous. Lustre is a good diagnostic property because most minerals will always appear either metallic or non-metallic, although as Figure 5.23 shows, there are exceptions.

## Hardness

One of the most important diagnostic properties of a mineral is its hardness. In practical terms, hardness determines whether or not a mineral can be scratched by a particular material.

In 1812 German mineralogist Friedrich Mohs came up with a list of 10 minerals representing a wide range of hardnesses, and numbered them 1 through 10 in order of increasing hardness. These minerals are shown in Figure 5.25 with the Mohs scale of hardness along the bottom axis. 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. The minerals themselves can be used to test other minerals.

## Crystal Habit

When minerals form within rocks, there is a possibility that they will form in distinctive crystal shapes if they are not crowded out by other pre-existing minerals. Every mineral has one or more distinctive **crystal habits** determined by their atomic structure, although 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 5.26 left), 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 5.26 centre), but can also form crystals with 12 faces, known as dodecahedra. The mineral garnet also forms dodecahedral crystals (Figure 5.26 right).

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

**Cleavage and Fracture**

Cleavage and fracture describe how a mineral breaks. These characteristics 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, like quartz, 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.

The way that minerals break is determined by the arrangement of atoms within them, and more specifically by the orientation of weaknesses within the lattice. Graphite and the micas, for example, have cleavage planes parallel to their sheets (Figure 5.27).
Other minerals have two directions of cleavage, classified as two planes at 90° (Figure 5.28 upper left) and two planes not at 90° (Figure 5.28 lower left). The minerals in Figure 5.28 both have two planes of cleavage that are very close to 90°. The white dashed lines mark the edges of the planes, as with Figure 5.28. See if you can find the planes repeated in the images. The images are very close-up views of the minerals, only a few cm across. Sometimes you must look very carefully to find cleavage planes.

![Figure 5.28 Two planes of cleavage. Upper left: two planes at 90 degrees. Lower left: two planes not at 90 degrees. Upper right: pyroxene (diopside) with the edges of planes marked with dashed lines. Lower right: plagioclase feldspar with the edges of planes marked. [Karla Panchuk]](image)

Some minerals have many directions of cleavage. Figure 5.29 shows examples of minerals with three directions of cleavage. Halite (Figure 5.29 top) has three planes at 90° and calcite (Figure 5.29 bottom) has three planes not at 90°.

There are a few common difficulties that occur when students are learning to recognize and describe cleavage. One is being confused about whether a flat surface on a crystal is a cleavage plane, a crystal face, or simply a surface that happens to be flat. Another is that it might be necessary to look very closely at a sample to see mineral cleavage. The key features in Figure 5.28, for example, are only cm or mm in scale. If crystals are very small, it may not be possible to see cleavage at all. Sometimes cleavage is present, but it is poor, meaning the cleavage surface isn’t perfectly flat. The best way to overcome all of these problems is to look at lots of examples. It’s worth it to be able to identify cleavage and fracture, because cleavage is a reliable diagnostic property for most minerals.

**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, are of average density (2.6 to 3.0 g/cm³), 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 g/cm³. If you picked up a sample of one of these minerals, they would feel much heavier compared to a similarly sized sample of a mineral with average density. 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 with other minerals in it.
Other properties

Several other properties are useful for identification of some minerals. For example:

- Calcite dissolves in dilute acid and will give off bubbles of carbon dioxide.
- Magnetite is strongly magnetic, and other minerals are weakly magnetic.
- Sphalerite ((Zn,Fe)S) gives off a smell of sulphur when drawn across a streak plate.
- Halite tastes salty.
- Talc feels soapy to the touch.
- Plagioclase feldspar has striations (parallel razor-thin lines etched on the surface) and some varieties show a play of colours when light hits them at the right angle.

Chapter 5 Summary

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

5.1 Electrons, Protons, Neutrons, and Atoms

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).
5.2 Bonding and Lattices

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.

5.3 Mineral Groups

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.

5.4 Silicate Minerals

Silicate minerals are the most important minerals in Earth’s crust and mantle. They all have 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.

5.5 Formation of Minerals

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.

5.6 Mineral Properties

Some of the important properties for mineral identification include hardness, cleavage/fracture, density, lustre, colour, and streak colour.

Questions for Review

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:

   a. Calcite  
   b. Gypsum  
   c. Hematite  
   d. Quartz  
   e. Biotite  
   f. Galena  
   g. Graphite  
   h. Fluorite  
   i. Pyrite  
   j. Orthoclase  
   k. Magnetite  
   l. 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 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?

See Appendix 2 for solutions to review questions