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 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 Fe_2SiO_4 to $FeMgSiO_4$ to Mg_2SiO_4. Intervening compositions are written as $(Fe,Mg)_2SiO_4$ 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.

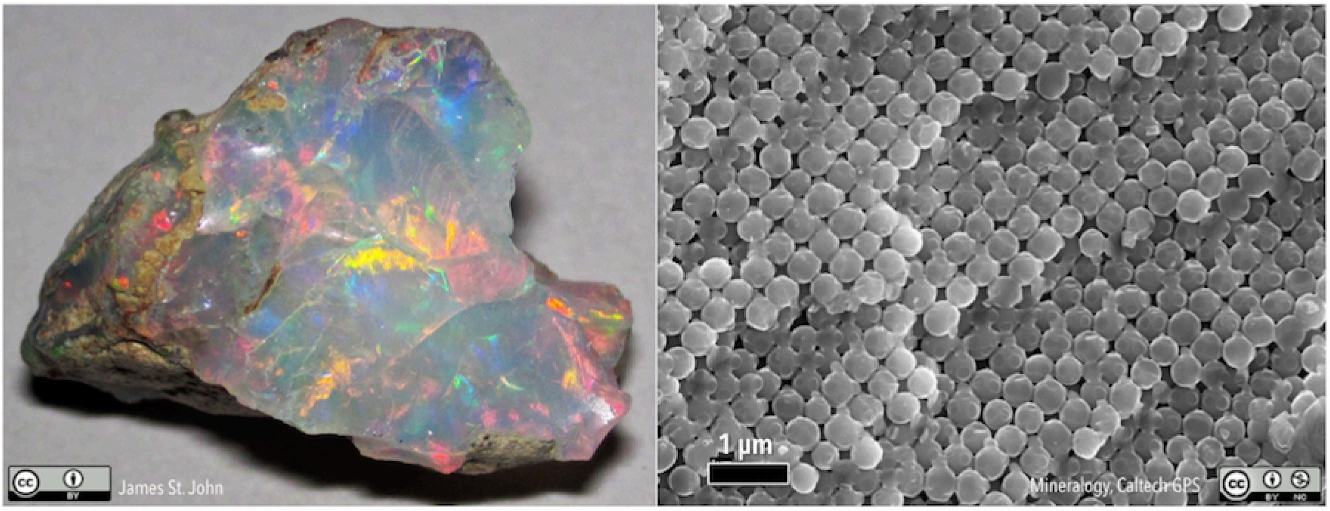


Figure 2.0.1 Opal (left) is like a mineral, but does not have a crystalline structure. Instead, it is made up of layers of closely packed silica spheres (right).

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	~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 (Å) 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.

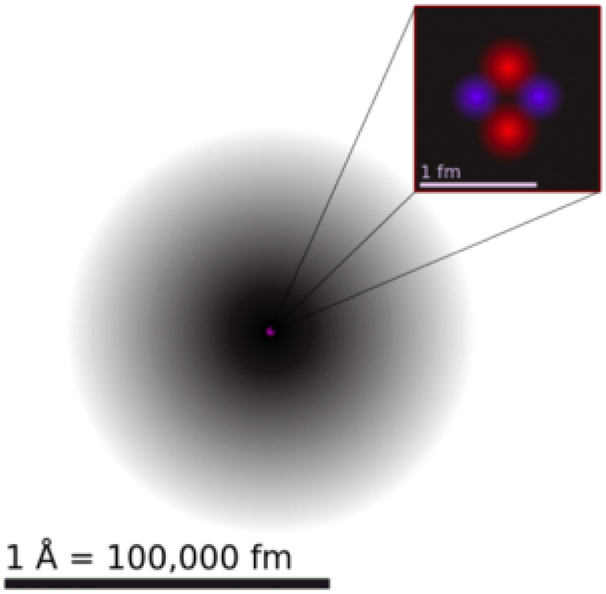


Figure 2.1.1 A depiction of a helium atom.

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.

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

[Skip Table]

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
Hydrogen	H	1	1	0	0	0
Helium *	He	2	2	0	0	0
Lithium	Li	3	2	1	0	0
Beryllium	Be	4	2	2	0	0
Boron	B	5	2	3	0	0
Carbon	C	6	2	4	0	0
Nitrogen	N	7	2	5	0	0
Oxygen	O	8	2	6	0	0
Fluorine	F	9	2	7	0	0
Neon *	Ne	10	2	8	0	0
Sodium	Na	11	2	8	1	0
Magnesium	Mg	12	2	8	2	0
Aluminum	Al	13	2	8	3	0
Silicon	Si	14	2	8	4	0
Phosphorus	P	15	2	8	5	0
Sulphur	S	16	2	8	6	0
Chlorine	Cl	17	2	8	7	0
Argon *	Ar	18	2	8	8	0
Potassium	K	19	2	8	8	1
Calcium	Ca	20	2	8	8	2
Scandium	Sc	21	2	8	9	2
Titanium	Ti	22	2	8	10	2
Vanadium	V	23	2	8	11	2
Chromium	Cr	24	2	8	13	1
Manganese	Mn	25	2	8	13	2

[Skip Table]

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).

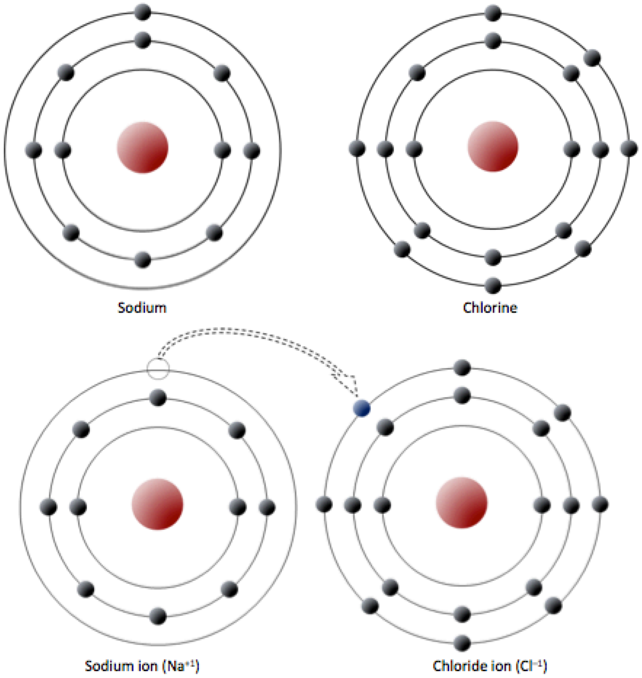


Figure 2.2.1 A very simplified electron configuration of sodium and chlorine atoms (top). Sodium gives up an electron to become a cation (bottom left) and chlorine accepts an electron to become an anion (bottom right). [Image Description]

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 (NaCl) is a mineral composed of chlorine and sodium linked together by ionic bonds (Figure 1.4.1). The mineral name for 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 (Cl_2) (Figure 2.2.2). Electrons are *shared* in a covalent bond.

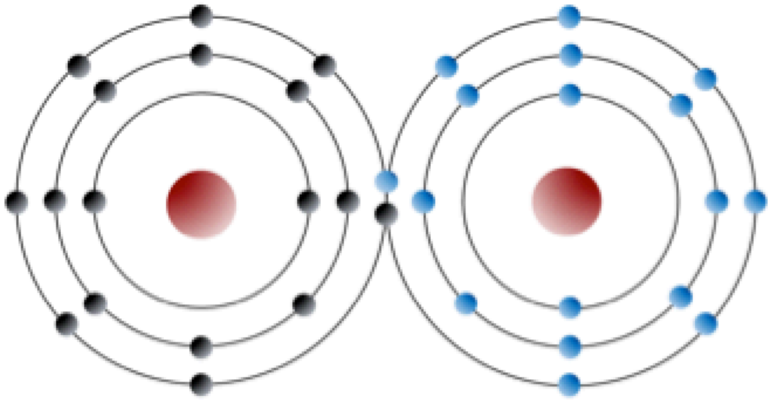


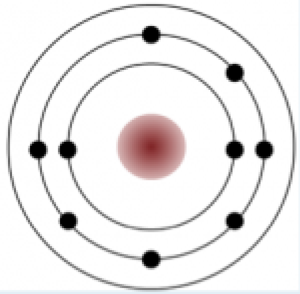
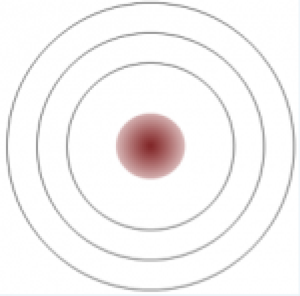
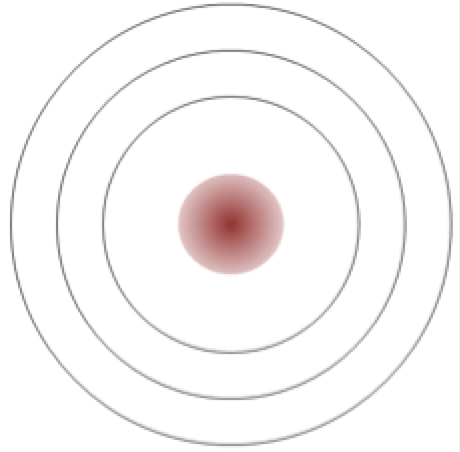
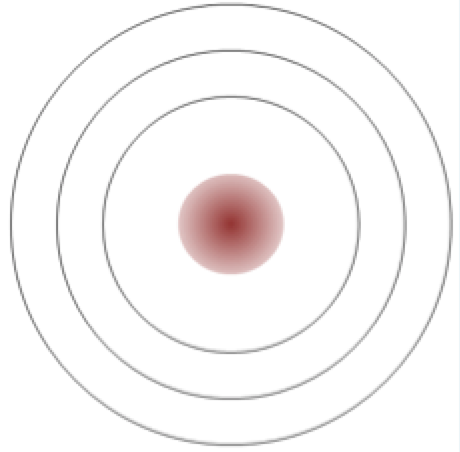
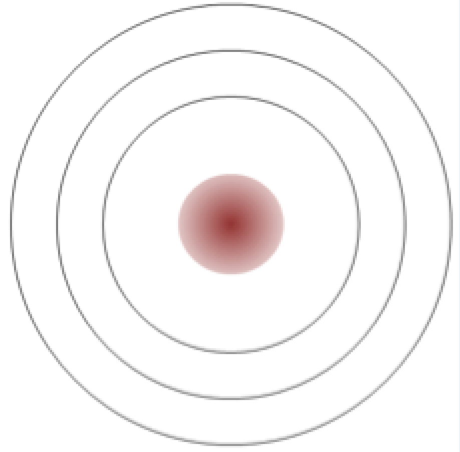
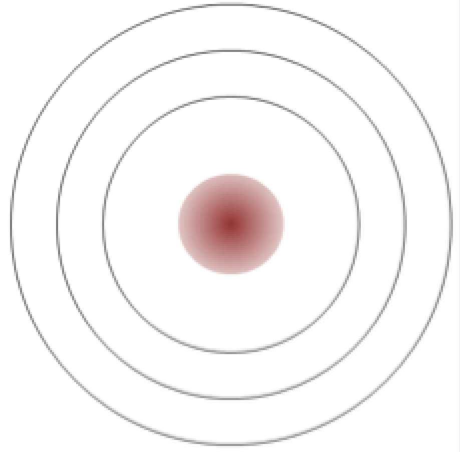
Figure 2.2.2 Depiction of a covalent bond between two chlorine atoms. The electrons are black in the left atom and blue in the right atom. Two electrons are shared (one black and one blue) so that each atom "appears" to have a full outer shell.

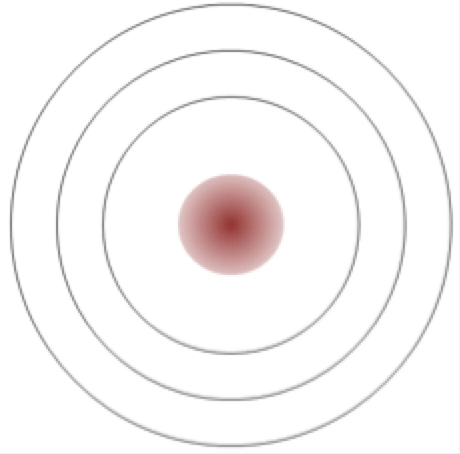
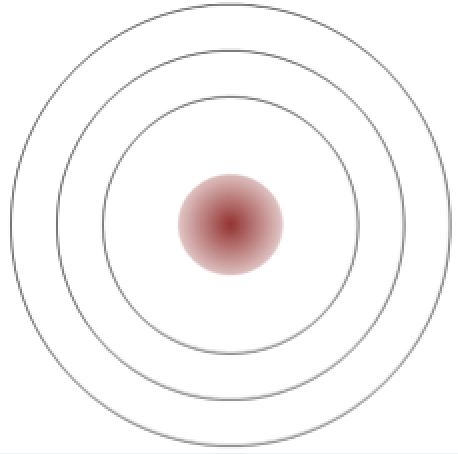
Exercise 2.1 Cations, anions, and ionic bonding

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 (3)**  ________

Oxygen (8)  ________	**Sodium (11)**  ________

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.

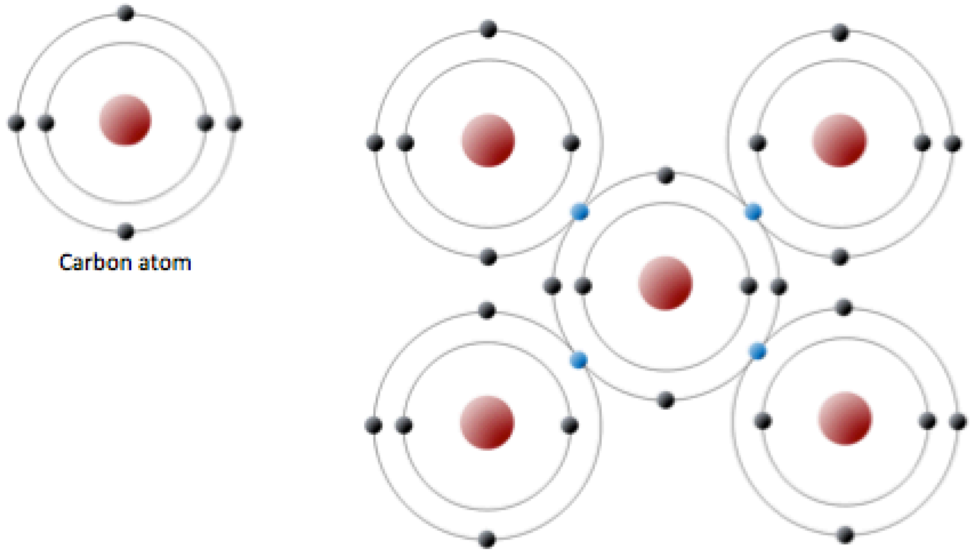


Figure 2.2.3 The electron configuration of carbon (left) and the sharing of electrons in covalent bonding of diamond (right). The electrons shown in blue are shared between adjacent Carbon atoms. Although shown here in only two dimensions, diamond has a three-dimensional structure as shown on Figure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

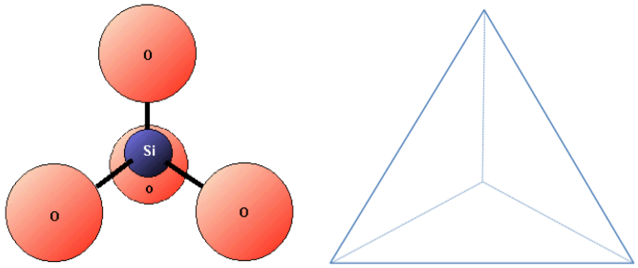


Figure 2.2.4 The silica tetrahedron, the building block of all silicate minerals. (Because the silicon ion has a charge of +4 and the four oxygen ions each have a charge of −2, the silica tetrahedron has a net charge of −4.)

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.

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

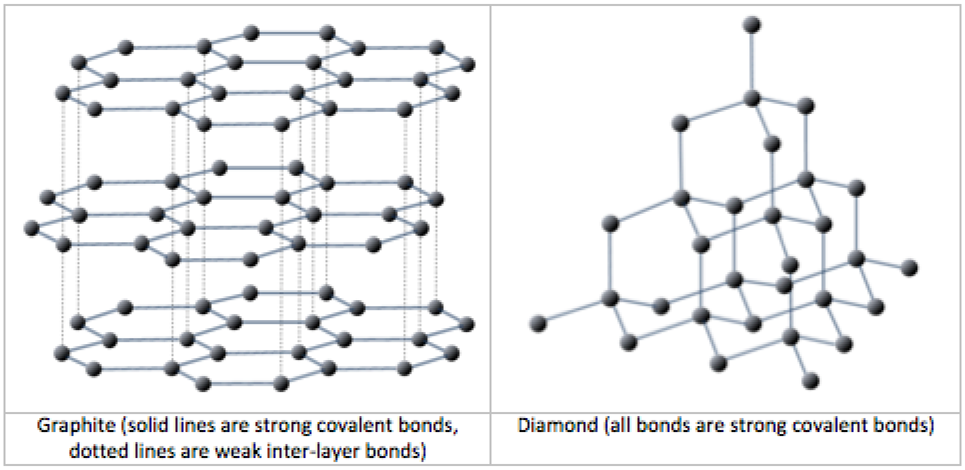
The element **silicon** is one of the most important geological elements and is the second-most abundant element in Earth's crust (after oxygen). Silicon bonds readily with oxygen to form a **silica** tetrahedron (Figure 2.2.4). Pure silicon crystals (created in a lab) are used to make semi-conductive 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_2. 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 caulking) 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 14th 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_2
Silica	A solid made out of SiO_2 (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.



Graphite (solid lines are strong covalent bonds, dotted lines are weak inter-layer bonds)

Diamond (all bonds are strong covalent bonds)

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 it 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.

[Skip Table]

Group	Examples
Oxides	Hematite (iron oxide Fe_2O_3), corundum (aluminum oxide Al_2O_3), water ice (H_2O)
Sulphides	Galena (lead sulphide PbS), pyrite (iron sulphide FeS_2), chalcopyrite (copper-iron sulphide $CuFeS_2$)
Sulphates	Gypsum (calcium sulphate $CaSO_4·H_2O$), barite (barium sulphate $BaSO_4$) *(Note that sulphates are different from sulphides. Sulphates have the SO_4^{-2} ion while sulphides have the S^{-2} ion)*
Halides	Fluorite (calcium flouride CaF_2), halite (sodium chloride NaCl) *(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 Appendix 1: List of Geologically Important Elements and the Periodic Table.)*
Carbonates	Calcite (calcium carbonate $CaCO_3$), dolomite (calcium-magnesium carbonate $(Ca,Mg)CO_3$)
Phosphates	Apatite ($Ca_5(PO_4)3(OH)$), Turquoise ($CuAl_6(PO_4)4(OH)_8·5H_2O$)
Silicates	Quartz (SiO_2), feldspar (sodium-aluminum silicate $NaAlSi_3O_8$), olivine (iron or magnesium silicate $(Mg,Fe)_2SiO_4$) *(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.)*
Native minerals	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 (H_2O) is a mineral (an oxide), but liquid water is not because it doesn't have a regular lattice.

Sulphides are minerals with the S^{-2} anion, and they 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_5FeS_4), stibnite (Sb_2S_3), and arsenopyrite (FeAsS).

Sulphates are minerals with the $SO_4{}^{-2}$ anion, and these include anhydrite ($CaSO_4$) and its cousin gypsum ($CaSO_4.2H_2O$) and 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.

The **halides** are so named because the anions include the **halogen** elements chlorine, fluorine, bromine, etc. Examples are halite (NaCl), cryolite (Na_3AlF_6), and fluorite (CaF_2).

The **carbonates** include minerals in which the anion is the $CO_3{}^{-2}$ complex. The carbonate combines with +2 cations to form minerals such as calcite ($CaCO_3$), magnesite ($MgCO_3$), dolomite ($(Ca,Mg)CO_3$)[1], and siderite ($FeCO_3$). The copper minerals malachite and azurite are also carbonates.

In **phosphate** minerals, the anion is the $PO_4{}^{-3}$ complex. An important phosphate mineral is apatite ($Ca_5(PO_4)_3(OH)$), 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 $Si : O_2$ to $Si : O_4$. 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 (FeS_2), Fe^{2+} is the cation and S^{-} is the anion. This helps us to know that it's a sulphide, but it is not always that obvious. Hematite (Fe_2O_3) is an oxide; that's easy, but anhydrite ($CaSO_4$) is a sulphate because $SO_4{}^{2-}$ is the anion, not O. Along the same lines, calcite ($CaCO_3$) is a carbonate, and olivine (Mg_2SiO_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:

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[1] The notations of two (or more) elements enclosed in parentheses with a comma between them: (Ca,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

[Skip Table]

Name	Formula	Group
sphalerite	ZnS	
magnetite	Fe_3O_4	
pyroxene	$MgSiO_3$	
anglesite	$PbSO_4$	
sylvite	KCl	
silver	Ag	
fluorite	CaF_2	
ilmenite	$FeTiO_3$	
siderite	$FeCO_3$	
feldspar	$KAlSi_3O_8$	
sulphur	S	
xenotime	YPO_4	

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.

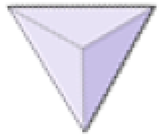
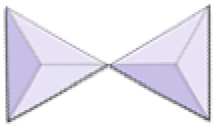
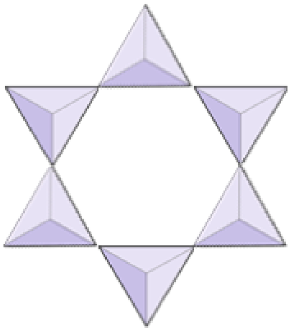
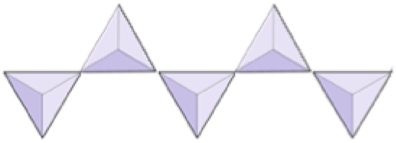
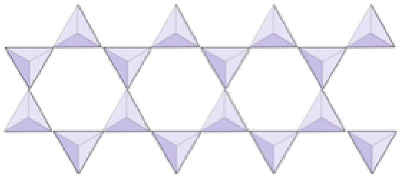
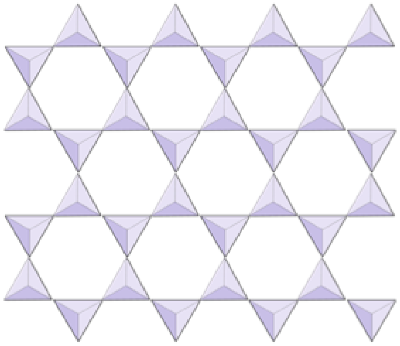
In silicate minerals, these tetrahedra are arranged and linked together in a variety of ways, from single units to complex frameworks (Table 2.6). The simplest silicate structure, that of the mineral **olivine**, is composed of isolated tetrahedra bonded to iron and/or magnesium ions. In olivine, the −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[1]). 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.

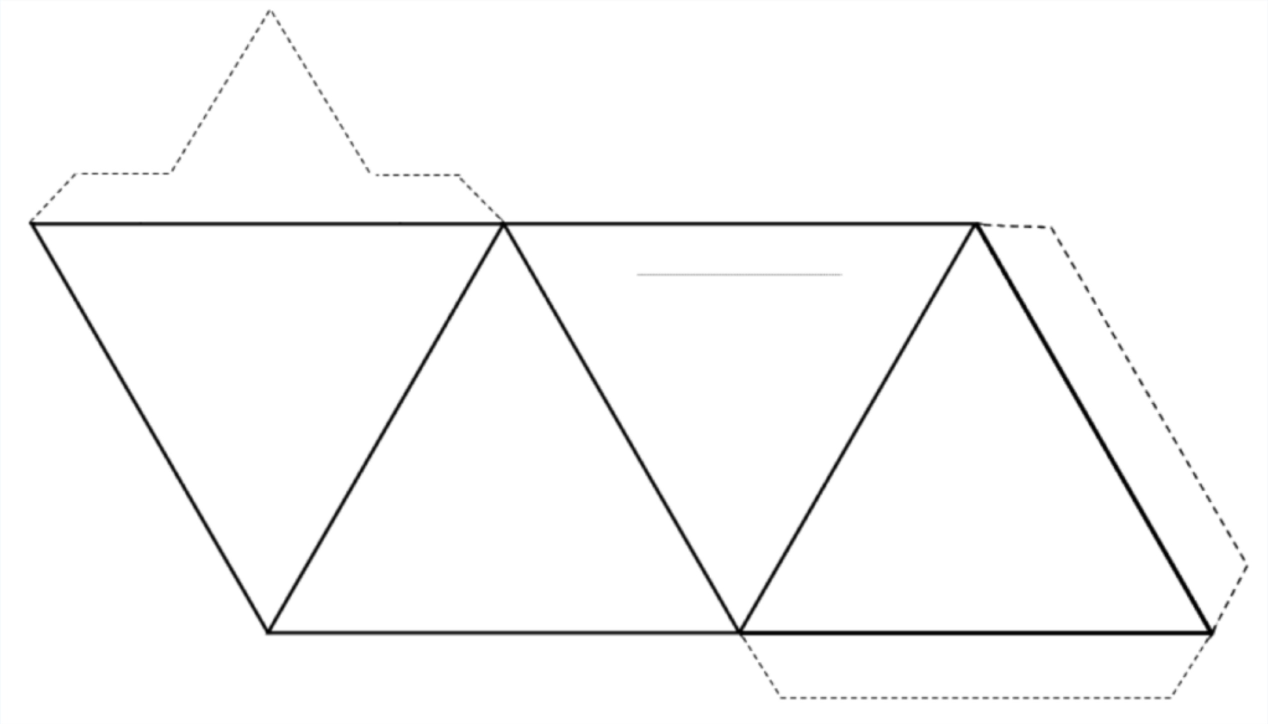
[Skip Table]

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

[Skip Table]

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.

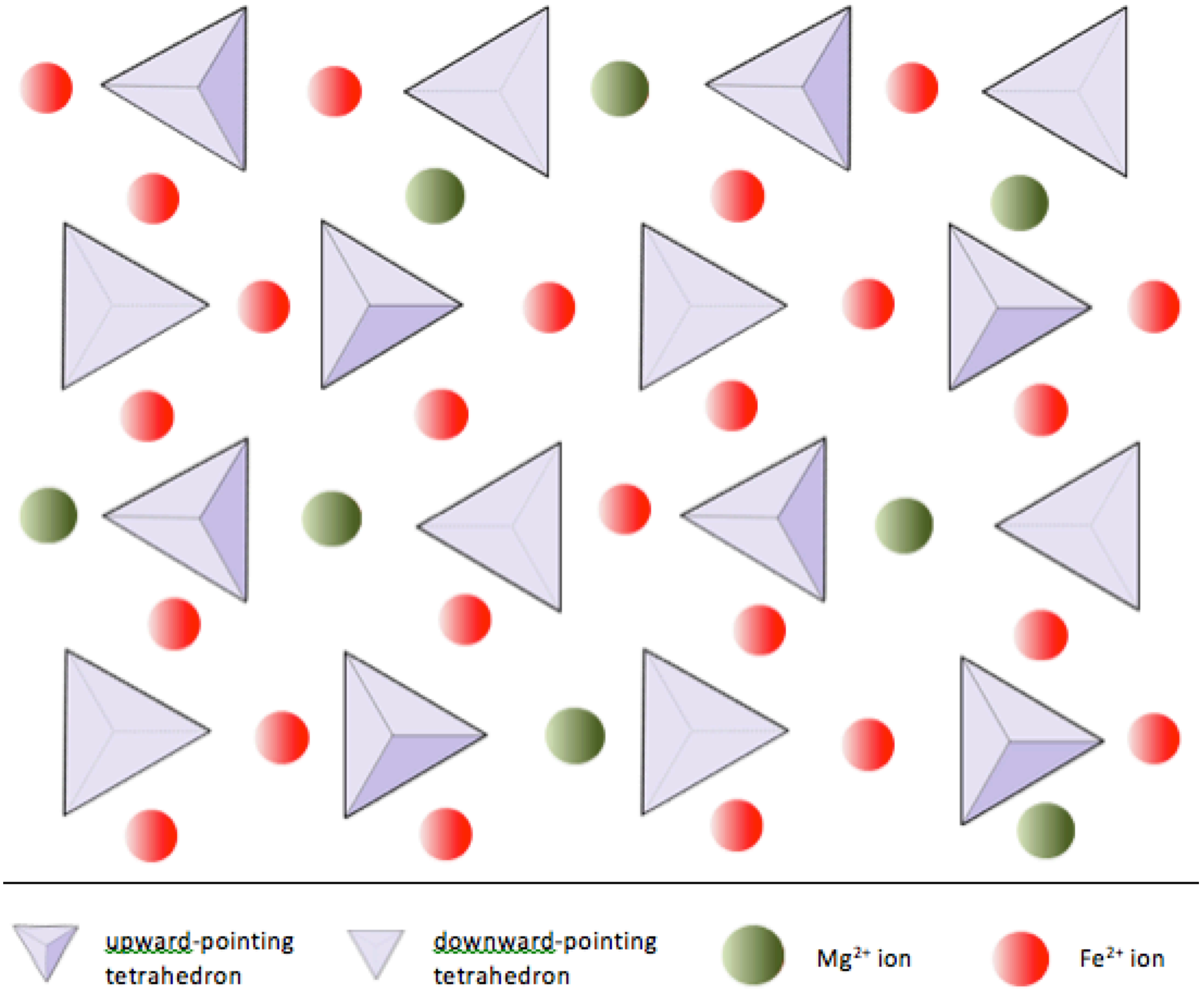


Figure 2.4.1 A depiction of the structure of olivine as seen from above. The formula for this particular olivine, which has three Fe ions for each Mg ion, could be written: $Mg_{0.5}Fe_{1.5}SiO_4$.

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.

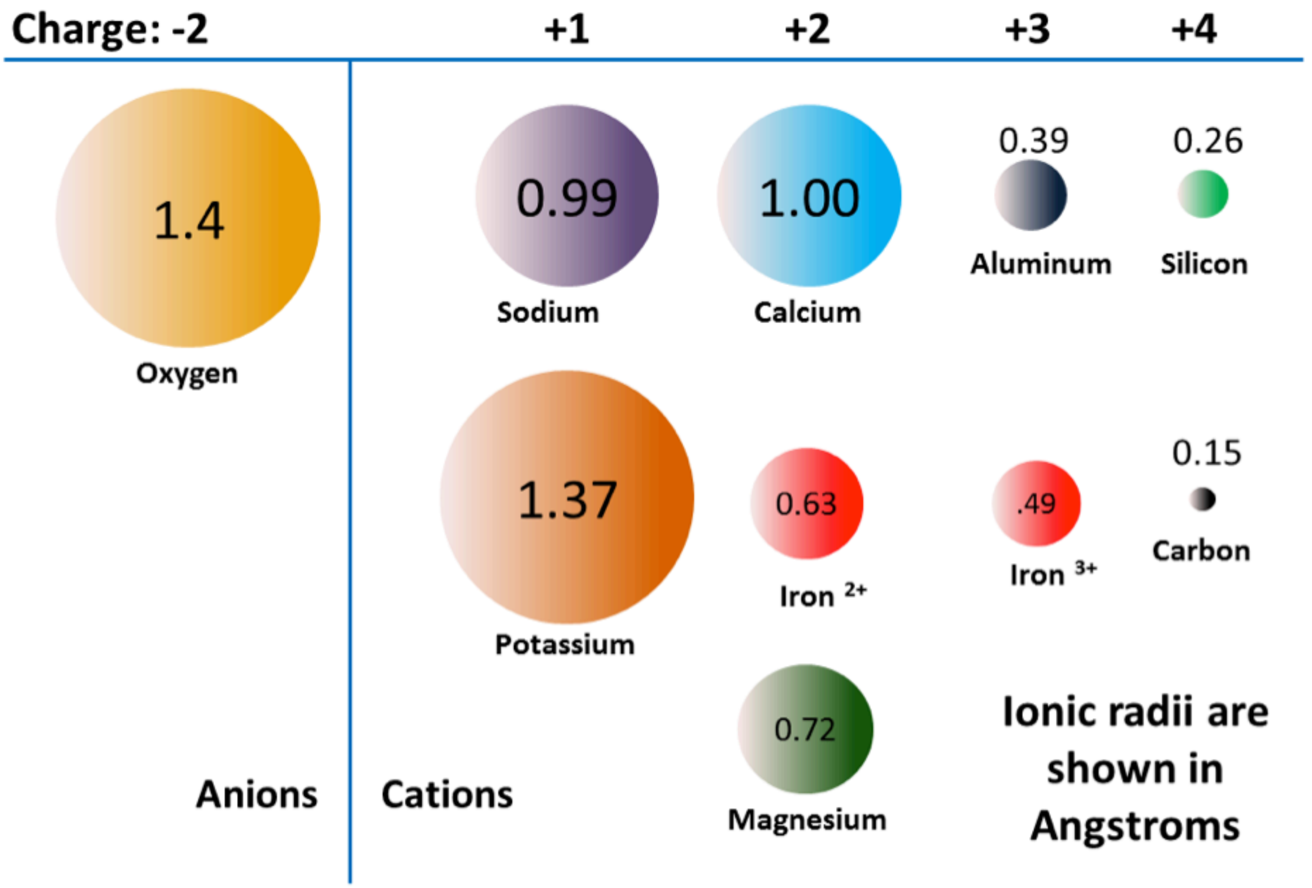


Figure 2.4.2 The ionic radii (effective sizes) in angstroms, of some of the common ions in silicate minerals. [Image Description]

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 $MgSiO_3$, $FeSiO_3$, and $CaSiO_3$, or some combination of these. Pyroxene can also be written as $(Mg,Fe,Ca)SiO_3$, 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., $MgSiO_3$) while olivine has two (e.g., Mg_2SiO_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).

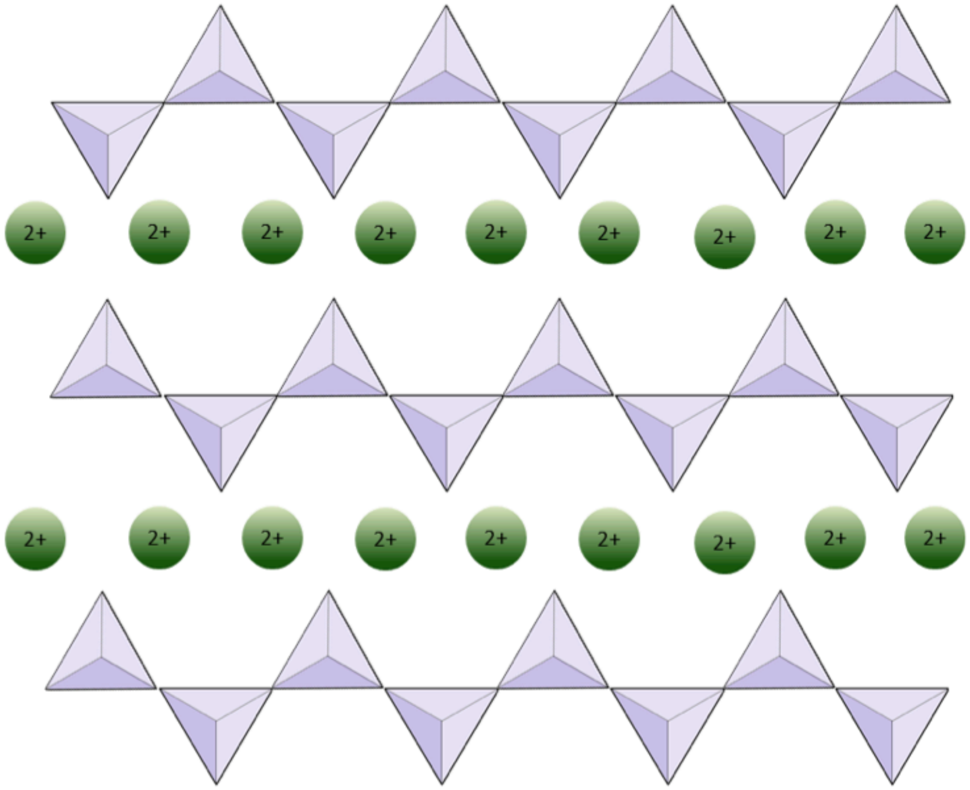


Figure 2.4.3 A depiction of the structure of pyroxene. The tetrahedral chains continue to left and right and each is interspersed with a series of divalent cations. If these are Mg ions, then the formula is $MgSiO_3$.

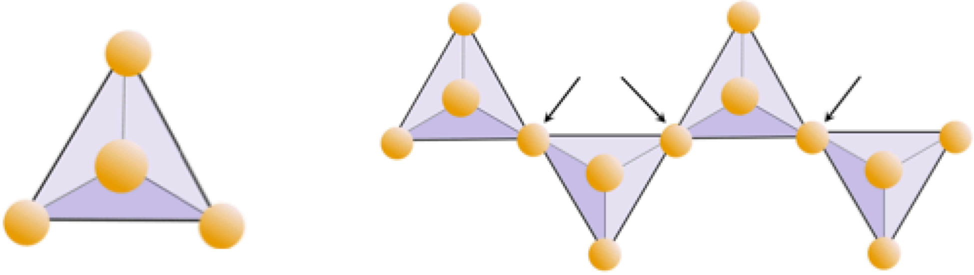
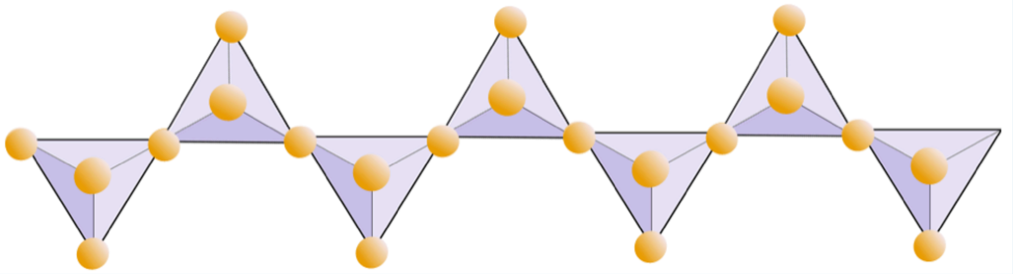


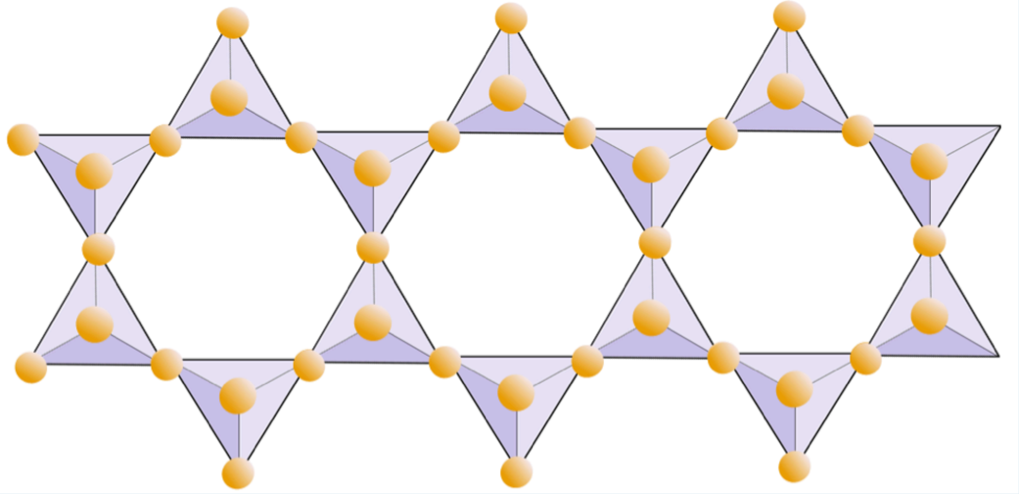
Figure 2.4.4 A single silica tetrahedron (left) with four oxygen ions per silicon ion (SiO_4). Part of a single chain of tetrahedra (right), where the oxygen atoms at the adjoining corners are shared between two tetrahedra (arrows). For a very long chain the resulting ratio of silicon to oxygen is 1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

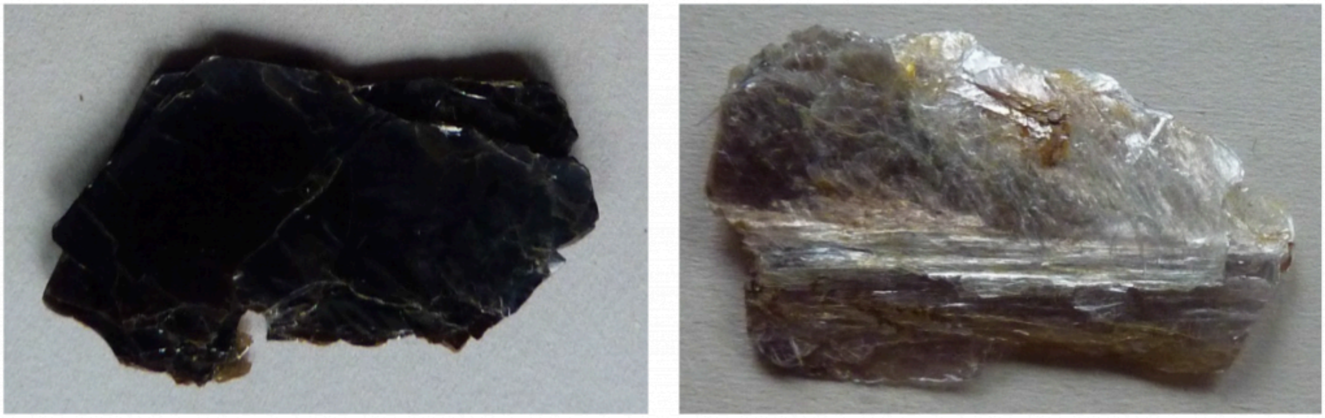


Figure 2.4.5 Biotite mica (left) and muscovite mica (right). Both are sheet silicates and split easily into thin layers along planes parallel to the sheets. Biotite is dark like the other iron- and/or magnesium-bearing silicates (e.g., olivine, pyroxene, and amphibole), while muscovite is light coloured. (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 $CaAl_2Si_3O_8$ and $NaAlSi_3O_8$ 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 (Ca^{2+} versus Na^+). This problem is accounted for by the corresponding substitution of Al^{+3} for Si^{+4}. Therefore, albite is $NaAlSi_3O_8$ (1 Al and 3 Si) while anorthite is $CaAl_2Si_2O_8$ (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** ($KAlSi_3O_8$) 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.

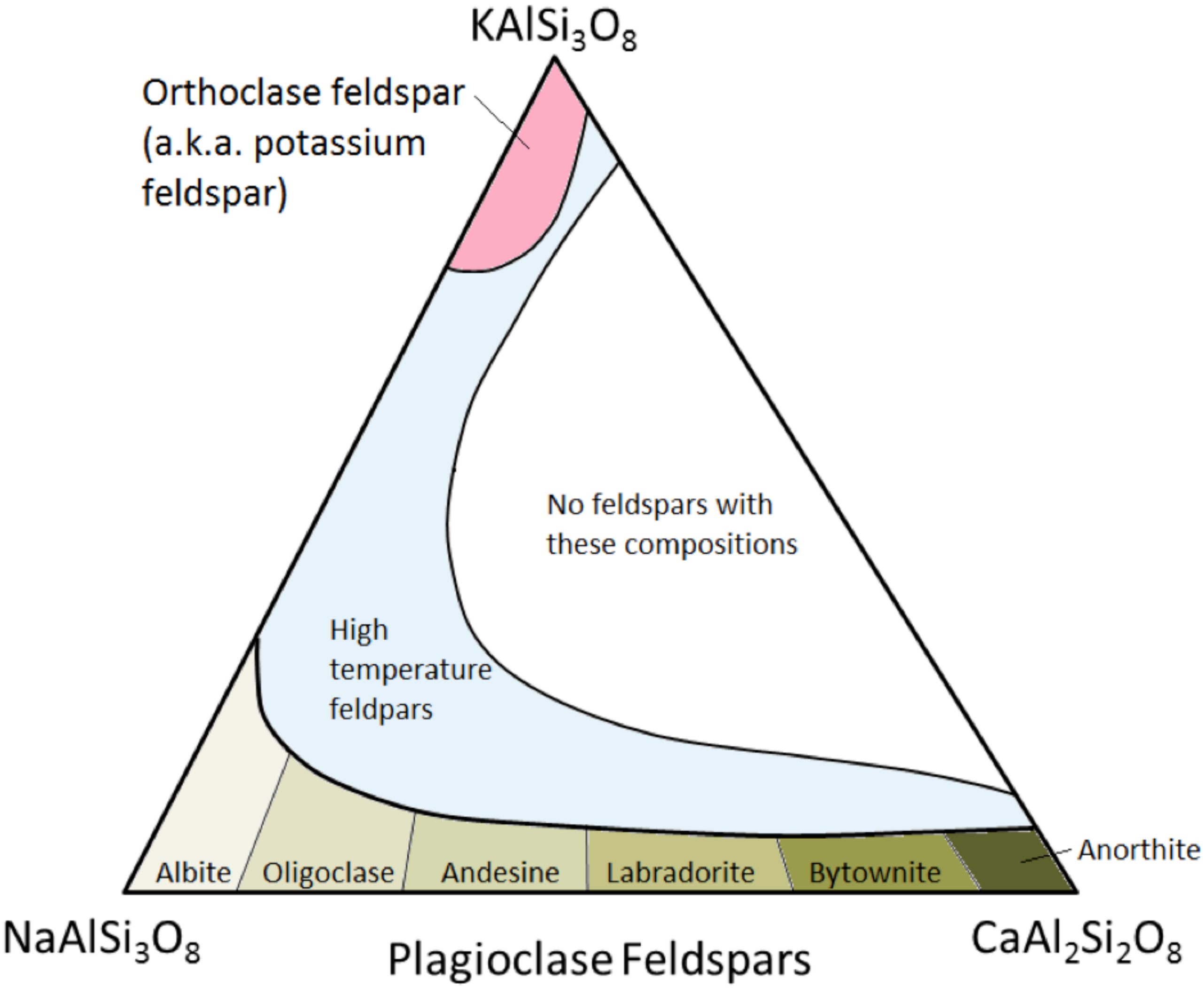


Figure 2.4.6 Compositions of the feldspar minerals.

In **quartz** (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	$(Mg,Fe)_2SiO_4$	.
pyrite	FeS_2	.
plagioclase feldspar	$CaAl_2Si_2O_8$	.
pyroxene	$MgSiO_3$	.
hematite	Fe_2O_3	.
orthoclase feldspar	$KAlSi_3O_8$	.
quartz	SiO_2	.
amphibole	$Fe_7Si_8O_{22}(OH)_2$	.
muscovite	$K_2Al_4Si_6Al_2O_{20}(OH)_4$	.
magnetite	Fe_3O_4	.
biotite	$K_2Fe_4Al_2Si_6Al_4O_{20}(OH)_4$	.
dolomite	$(Ca,Mg)CO_3$	.
garnet	$Fe_2Al_2Si_3O_{12}$	.
serpentine	$Mg_3Si_2O_5(OH)_4$	.

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 tehe rocks around us 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., 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 ½ teaspoon (~2.5 cm^3) of any kind of table salt into a small bowl. Add about 2 teaspoons (~10 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.

Figure 2.5.1

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.

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 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 (Li_2CO_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 (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).

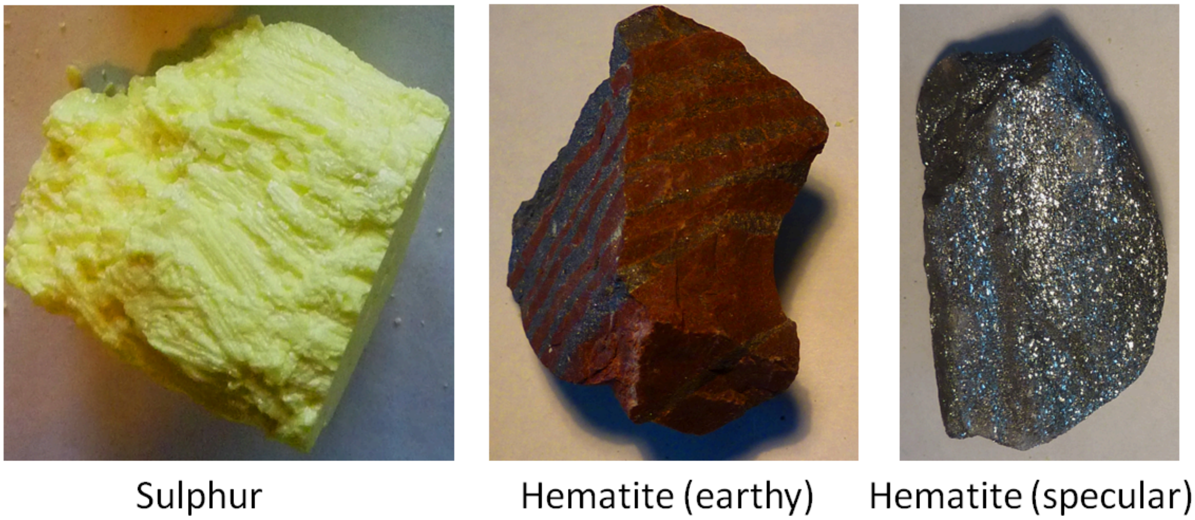


Figure 2.6.1 Examples of the colours of the minerals sulphur and hematite.

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).

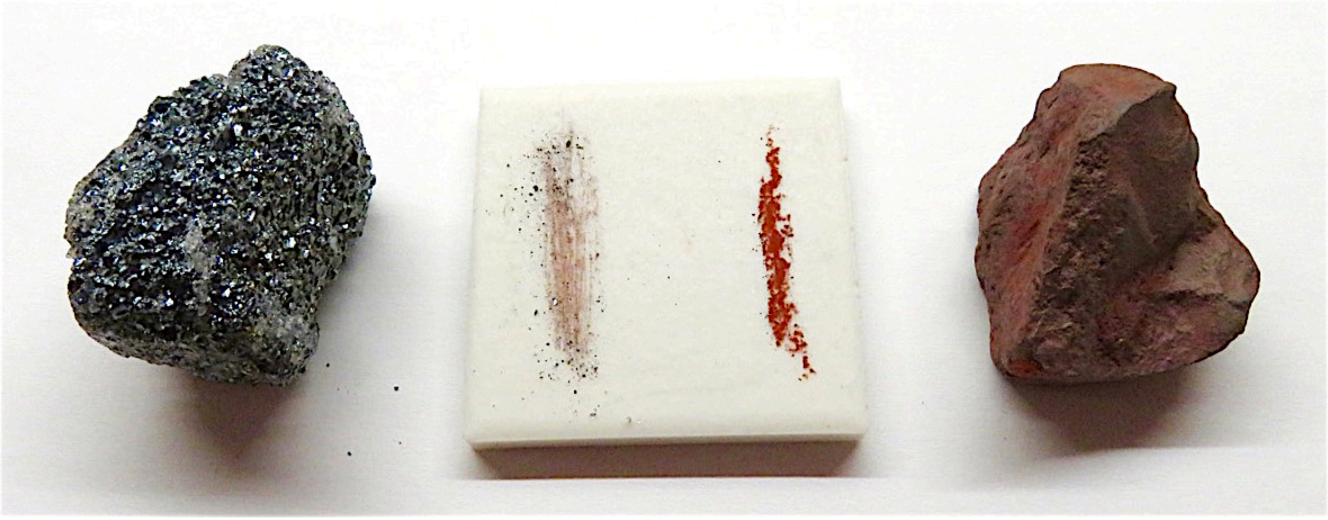


Figure 2.6.2 The streak colours of specular (metallic) hematite (left) and earthy hematite (right). Hematite leaves a distinctive reddish-brown streak whether the sample is metallic or earthy.

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.

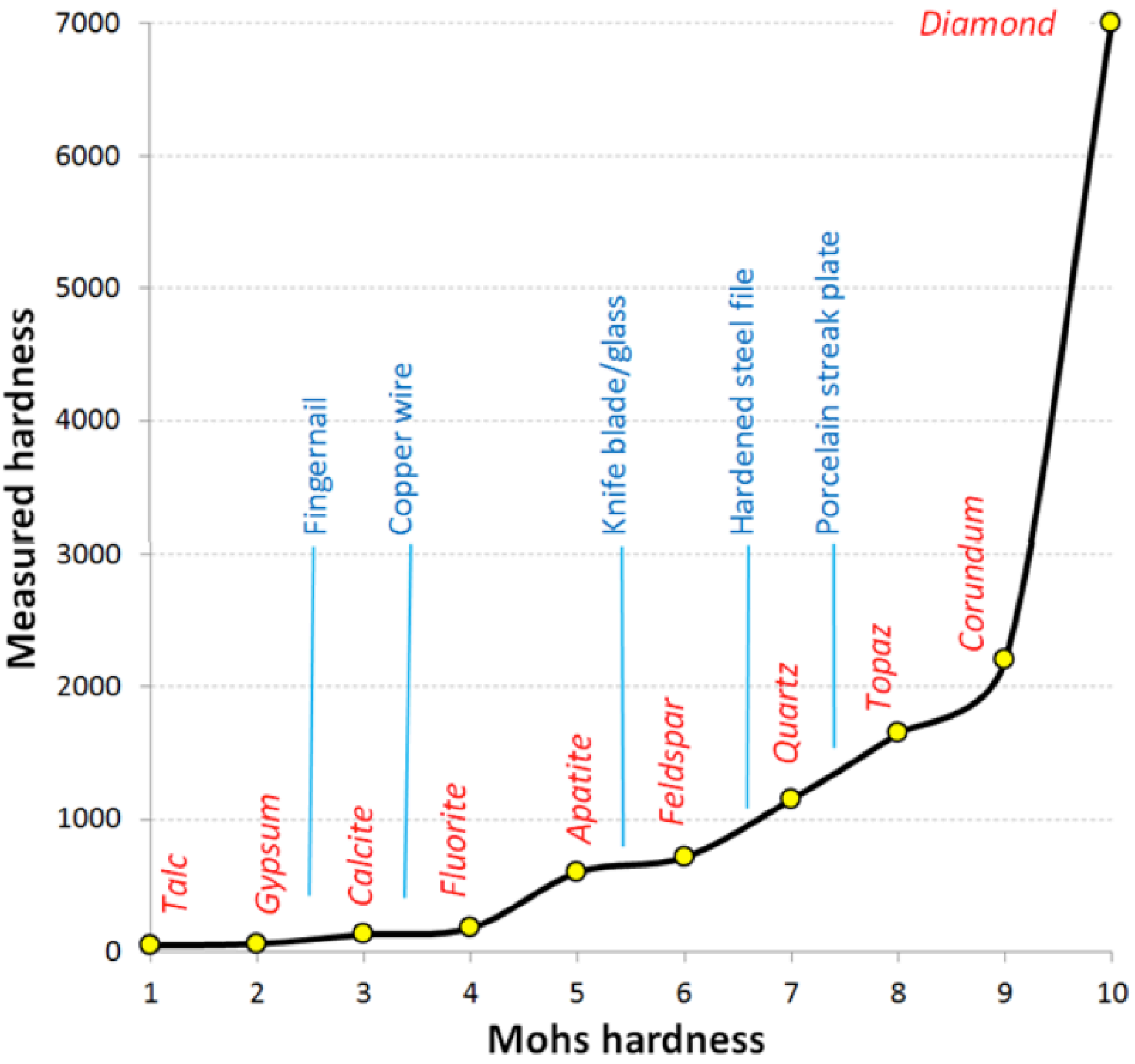


Figure 2.6.3 Minerals and reference materials in the Mohs scale of hardness. The "measured hardness" values are Vickers Hardness numbers. [Image Description]

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° to each other (Figure 2.6.5).

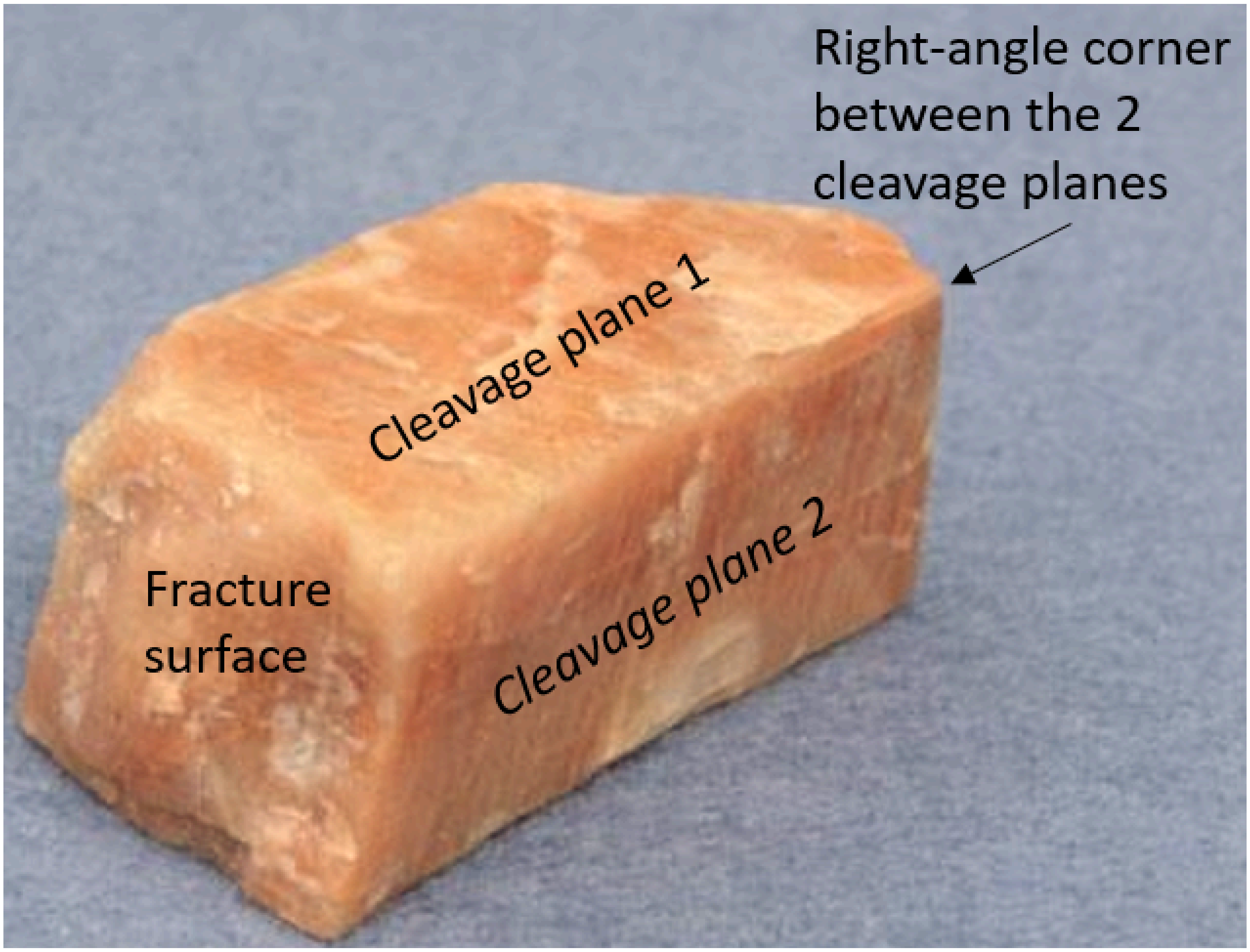
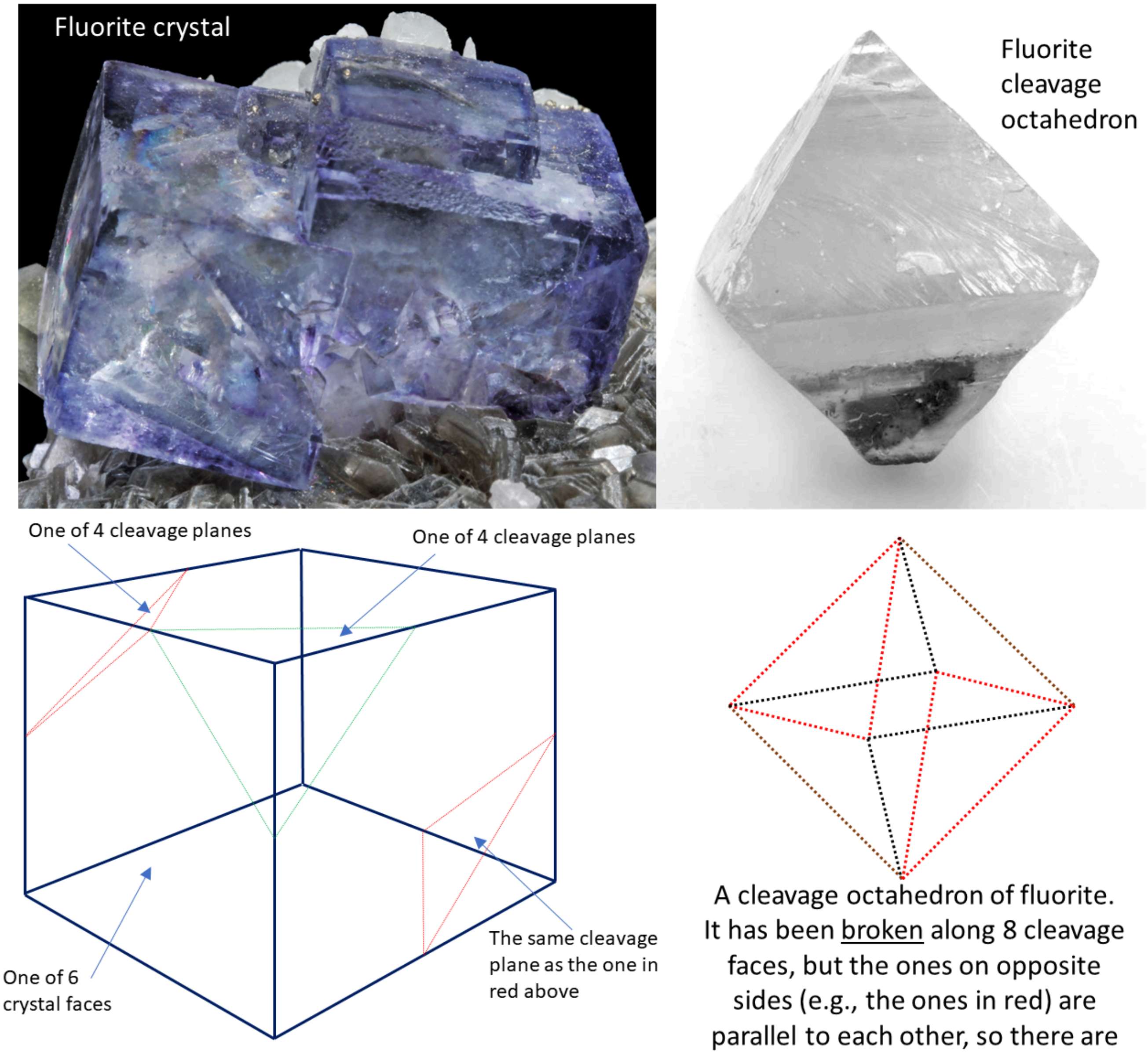


Figure 2.6.5 Cleavage and fracture in potassium feldspar

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.



A crystal of fluorite. It has grown with 6 crystal faces in the form of a cube.

A cleavage octahedron of fluorite. It has been broken along 8 cleavage faces, but the ones on opposite sides (e.g., the ones in red) are parallel to each other, so there are actually only 4 cleavage planes.

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 (g/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 g/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]

Media Attributions

- 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
2.1 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).
2.2 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.
2.3 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.
2.4 Silicate Minerals	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.
2.5 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.
2.6 Mineral Properties	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 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?