2.1 Carrollite has the formula Cu(Co,Ni)$_2$S$_4$.

2 Mineral Chemistry

KEY CONCEPTS

- All matter, including mineral matter, is made of elements.
- 8 to 10 elements account for most of Earth’s mass but the elements are distributed unevenly.
- The *Periodic Chart of the Elements* orders elements by their atomic number and groups elements with similar properties.
- Elements with similar properties may substitute for each other in minerals and many minerals have variable compositions.
- Ionic, covalent, or metallic bonding are most common in minerals.
- The kind of bonding affects mineral properties.
- We describe mineral compositions by giving weight %s of the elements or oxides present.
- Weight % values can be normalized to give chemical formulas.
- Although 1000s of minerals are known, only a small number are common.

### 2.1 Elements and Minerals

All matter is made of atoms of individual elements. For example, the matter of the Milky Way, our galaxy, is mostly hydrogen and helium – the same elements that make up >99% of our sun. Earth, however, is dominated by other elements, and 8 to 10 elements account for most of Earth’s mass. Although Earth’s crust and mantle contain the same major elements, the proportions are not the same (see histograms below). And, the core is completely different from the crust and mantle – it is mostly iron with perhaps 20% nickel and lesser amounts of other elements.
2.2 Elements in the crust and mantle

Most of the minerals and other geological materials we see derive from the crust, but some come from the uppermost part of the mantle. In both places, oxygen and silicon are the dominant elements (Figures 2.2 and 2.3). Together these two elements make up about three quarters of the crust, and two thirds of the mantle. Other quite abundant elements in both the crust and mantle include aluminum, iron, calcium, sodium, potassium, and magnesium. However, the mantle contains much more magnesium and iron, and less silicon, than the crust. Hydrogen (0.15 wt%) and carbon (0.18 wt%) are overall minor elements in the crust but are key components in some minerals. And other generally rare elements are sometimes concentrated by geological process to make exotic minerals.

The compositions of Earth’s outer layers vary somewhat laterally and vertically. For example, the oceanic crust is not the same composition as the continental crust. And, the composition of the shallow crust is somewhat different from the deep crust. Nonetheless, in most settings, we can expect common minerals to be made of the elements shown in the histograms above.
Some elements are common in many different minerals. Oxygen and silicon are perhaps the best examples. Many sedimentary rocks and nearly all igneous and metamorphic rocks are composed of multiple minerals containing these two elements. In contrast, because of their properties, some other elements tend to be found in only a few distinct minerals. For example, titanium (Ti) may occur as a minor component in biotite, amphibole, or other minerals. In many rocks, however, Ti is concentrated in Ti-rich minerals such as rutile (TiO$_2$), titanite (CaTiSiO$_5$), and ilmenite (FeTiO$_3$). Rocks rich in Ti always contain one of these latter three minerals. Similarly, rocks containing significant amounts of phosphorous usually contain apatite, Ca$_5$(PO$_4$)$_3$(OH,F,Cl), or monazite, (Ce,La,Th,Y)PO$_4$.

### 2.2 Elements and the Periodic Table

![Periodic Table of the Elements](image)

2.3 The Periodic Table of the Elements

Here (Figure 2.3) we see the Periodic Table of the Elements. Elements are ordered by increasing atomic number (the number of protons in their atom nucleus), which correlates with
atomic weight and size. The first element is hydrogen (atomic number 1) and the last is oganesson (atomic number 118). Oganesson is an obscure synthetic radioactive element that has only been produced a few times and in minute amounts. In this table, elements are in numbered periods (rows) and groups (columns) based on the configuration of their electron orbitals.

The chart contains 118 elements in all, but only about 90 occur naturally. The other 20 or so (with atomic numbers 104-118 and 99-103) elements are artificial – they are synthesized in nuclear reactors and some are radioactive with very short lives. The distinction between natural and artificial elements is, however, a bit uncertain because only a few atoms have (speculatively) been identified for some of the rarest natural elements. Although the Periodic Table of the Elements has appeared in many forms, the basic relationships are the same today as they were when Dmitri Mendeleyev (1834-1907) devised the first version in 1870. Some elements were unknown and omitted from the original table during Mendeleyev’s time, but were later discovered and added. Additionally, there has been a slight rearrangement so that the arrangement of the elements mirrors the order in which electrons occupy orbitals.

Chemists classify elements into different types that have related properties (shown by different colors in the table above). Hydrogen is a special element, but the other elements in Group 1 are alkali metals. The elements in group 2 are alkaline earth metals. Those elements in Groups 4 through 12 are transition metals. Group 17 elements are the halogens, and Group 18 elements are the noble gases. Elements between the transition metals and the halogens are called nonmetals or other metals (but different versions of the chart have them divided in slightly different ways). The asterisks in the chart show where the lanthanides (also called rare earth elements) and actinides (both considered transition metals)
were extracted from the main chart and put as separate rows at the bottom. If we did not extract them, the chart would be too wide to fit easily on this page. Scandium (Sc) and yttrium (Y) are closely related to the lanthanides and so are commonly classified with them.

2.2.1 Groups

![Perthite (exsolved alkali feldspar)](image)

Elements in the same groups (columns) have their outermost electrons in the same kind of orbitals. This is very significant because it means the elements have similar chemical properties, form similarly charged ions and, importantly, commonly substitute for each other in mineral crystals. For example, the mineral specimen shown in Figure 2.4 is an alkali feldspar. When it first crystallized from a magma, it was a homogeneous mixture of Na-feldspar and K-feldspar. Na and K, both alkali elements (Group 1), commonly substitute for each other in minerals. When this sample cooled, however, the two different feldspar components unmixed from each other, much the same way that chicken soup and fat separate on cooling. The result is called *exsolution* (which means unmixing) and the specimen now contains thin pinkish
veins of K-feldspar surrounded by Na-feldspar. We call feldspar that has exsolved like this, *perthite*.

### 2.2.2 Atomic Number and Mass

The variable $Z$ designates an element’s atomic number, the number of protons in its nucleus. $Z$ is also equal to the number of electrons orbiting the nucleus in neutral (non-ionized) atoms, and is close to the number of electrons in most ions. A neutral iron atom ($\text{Fe}_0$), for example, contains 26 protons ($Z = 26$) in its nucleus and 26 electrons in an electron cloud around the nucleus. Because the size of its electron cloud controls the diameter of an atom, elements with greater atomic number, with many protons and thus many electrons, are larger than those or lower atomic number.

Atomic nuclei (except one isotope of hydrogen) contain neutrons in addition to protons, and the number of neutrons, designated by $N$, may vary. This leads to isotopes of different mass numbers. Mass number, designated by the variable $A$, is equal to the number of protons and the number of neutrons combined: $A = Z + N$. Most chemical elements have several different naturally occurring isotopes. Some isotope varieties, however, are generally more common than others, and some only exist in minute amounts.

#### 2.5 Isotopes of Oxygen
Oxygen atoms, for example, may be any of three isotopes (shown in Figure 2.5). Oxygen may be $^{16}\text{O}$, $^{17}\text{O}$, or $^{18}\text{O}$, where the superscript number denotes $A$. Examination of the equation in the previous paragraph tells us that the three isotopes of oxygen must have 8, 9, and 10 neutrons, respectively, because all must have 8 protons if they are oxygen. $^{17}\text{O}$ and $^{18}\text{O}$ are very rare; $^{16}\text{O}$ is 99.8% of all natural oxygen.

### 2.2.3 Moles

A mole of an element (or of a compound) is defined as containing $6.022 \times 10^{23}$ atoms (or molecules). The number, $6.022 \times 10^{23}$, is known as Avogadro’s number. So, one mole of carbon is equivalent to $6.022 \times 10^{23}$ carbon atoms. The scale used to measure atomic mass has changed slightly over time. Today, it is standardized relative to carbon so that the mass of one mole of $^{12}\text{C}$ is exactly 12.0000. Consequently, all atomic masses are given in atomic mass units (amu), defined as one-twelfth the mass of $^{12}\text{C}$. Both protons and neutrons have equivalent mass, about one amu, and electrons have almost no mass (less than $1/1000^{th}$ the mass of protons and neutrons). So, we might expect the mass of an atom to be equal to the mass number ($A$, the total number of protons plus neutrons). But for several reasons, not worth going into here, the mass of an atom is generally close to, but not the same as, the mass number.

Elements are different from atoms. The atomic mass, also called the atomic weight, of an element is the sum of the masses of its naturally occurring isotopes weighted in accordance with their abundances. Atomic masses/weights of elements are molal quantities (and often given in units of grams/mole) but they are really dimensionless numbers because they are all calculated relative to the atomic mass/weight of a mole of carbon. Although isotope mass numbers are always
integers, atomic weights of elements are not. For example, many tables and charts give the atomic weight of oxygen as 15.999 and that of iron as 55.847. When elements combine to produce a compound, the atomic weight of the compound is the sum of the weights of the elements in the compound. FeO, for example, has atomic weight 71.846 (15.999 + 55.847) grams/mole.

Most elements have very small isotopic variation in nature, no matter where they are found. Thus, most quartz (SiO₂) contains about the same relative amounts of the three natural oxygen isotopes (mostly ¹⁶O) depicted in Figure 2.5 above. Furthermore, isotopic variations have extremely small effects on the properties of minerals. So, mineralogists generally do not worry much about isotopes. Small isotopic variations, however, may be significant to a geochemist trying to determine the genesis of a particular mineral or rock.

● Box 2-1 What Is a Mole of Quartz?
Quartz (SiO₂) is one of the most common and well-known minerals. A mole of quartz is 6.022 × 10²³ SiO₂ molecules, and that sounds like a lot. How much quartz is that?

To answer this question, we use the atomic weights of silicon and oxygen as well as some crystallographic data. Silicon and oxygen have atomic weights of 28.0855 and 15.9994, respectively. The atomic weight of quartz, SiO₂, is therefore 60.0843 (= 28.0855 + 15.9994 + 15.9994). This means that a mole of quartz, 6.022 × 10²³ SiO₂ molecules, weighs 60.0843 gm.

Crystallographers have determined that quartz crystals are made of fundamental unit cells shaped like trigonal prisms (discussed in detail in later chapters) containing three (Z = 3) SiO₂ molecules. Each unit cell has a volume of 112.985 Å³.

So we may calculate the volume of a mole of quartz as:

\[ V = N_A \times \frac{v}{Z} \]

(where \( V \), \( N_A \), \( v \), and \( Z \) are the molar volume, Avogadro’s number, the unit cell volume, and the number of molecules per unit cell, respectively.)

So, \( V = 6.022 \times 10^{23} \text{ SiO}_2/\text{mole} \times 112.986 \text{ Å}^3/\text{unit cell} + 3 \text{ SiO}_2/\text{unit cell} \)

\[ = 2.268 \times 10^{25} \text{ Å}^3 = 22.68 \text{ cm}^3 \]

which is slightly smaller than a golf-ball.

We can, if we wish, then calculate the density (\( \rho \)) of quartz from the molar data:

\[ \rho = \text{molar weight} \div \text{molar volume} = 60.0843 \text{ gm/mole} \div 22.68 \text{ cm}^3/\text{mole} = 2.649 \text{ gm/cm}^3 \]
2.2.4 Radioactive Minerals

Most common isotopes are stable isotopes. In 1896 Henri Becquerel discovered unstable isotopes and radioactivity when he unintentionally conducted an experiment. He put some radioactive samples in a drawer, along with a photographic plate, and subsequently found that the plate had recorded the image of a key that had been sitting on top of it. Although Becquerel didn’t know it, X-rays given off by uranium-rich minerals had caused the image. During the following decade, researchers including Marie and Pierre Curie and Ernest Rutherford determined that some elements, termed radioisotopes, emit alpha (α) and beta (β) particles, and gamma (γ) radiation as they decay to form daughter isotopes. So, minerals are radioactive if they contain radioisotopes, and the amount of radioactivity depends on isotope concentration. Most minerals that contain potassium, uranium, and thorium are radioactive, but many other elements besides these three can also contribute to radioactivity.

Radioactive minerals are quite rare but many of them, like those seen in Figures 2.6, 2.7 and 2.8 below, make spectacular mineral specimens. Torbernite (copper-uranium phosphate), autunite (calcium-uranium phosphate), and uraninite (uranium oxide) are three of the most common examples. Uraninite is the main ore for uranium; the specimen seen here comes from a mine in Topsham, Maine.
2.2.5 Periods

In atoms, electrons are in orbitals of different energies around atomic nuclei. Electrons are always moving, and orbitals are regions around a nucleus where electrons are likely to be found. We designate an orbital using a number and a letter, for example 1s or 1p – the number refers to a specific electron shell and the letter to a subshell. S subshells can hold up to 2 electrons, p subshells up to 6 electrons, d subshells up to 10 electrons, and f subshells up to 14 electrons. The table below lists, from lowest to highest energy, the orbitals and the maximum number of electrons they can hold. s orbitals are spherical, p orbitals are dumbbell shaped, d and f orbitals have more complicated shapes. The shape of an atom’s outermost occupied orbital is often important because it can affect mineral properties such as color. But, for simplicity, in the drawings below we show all orbitals as circular.

<table>
<thead>
<tr>
<th>Electron Orbitals</th>
</tr>
</thead>
<tbody>
<tr>
<td>orbital</td>
</tr>
<tr>
<td>max # electrons</td>
</tr>
<tr>
<td>increasing energy</td>
</tr>
</tbody>
</table>

In a neutral (non-ionized) atom, the number of electrons is equal to the atom’s atomic number. So, a hydrogen atom (atomic number 1) contains one electron, a helium atom (atomic number 2) contains two electrons, a lithium atom (atomic number 3) contains three electrons, and so forth to high atomic number. Electrons populate orbitals sequentially, filling the lowest energy orbitals first and proceeding to higher energy orbitals. Thus, hydrogen has one electron in a 1s orbital. Helium has two electrons in a 1s orbital, lithium has two electrons in a 1s orbital and one electron in a 2s orbital, etc.
2.9 Electron orbits in a sodium atom

For an example of electron configuration in a heavier element, consider sodium. This schematic drawing in Figure 2.9 shows orbitals and electrons in a sodium atom. The electrons are orbiting a nucleus that contains protons and neutrons (but they are not all visible). Sodium has atomic number 11, so a neutral atom has 11 protons and 11 electrons. Natural sodium also has 12 neutrons. The electrons fill the 1s, 2s, and 2p orbitals. One additional electron is in the 3s orbital.

2.10 Electron orbitals of valence electrons

We call electrons that contribute to chemical bonding valence electrons. These electrons are generally in the outermost electron shells (but may be in an inner shell for transition
metals). This chart (Figure 2.10) shows the orbitals of valence electrons for different elements. Elements in the same period (row) of the periodic table have their valence electrons in the same s or p shell, or (for transition elements) in a d orbital of the next lower shell.

The first period contains only two elements (H and He), while the second and third have eight each. Electrons for second and third period elements are in 2s, 2p, 3s, or 3p orbitals. The fourth through seventh periods contain some elements with valence electrons in s or p orbitals, but also contain 20 elements having valence electrons in d-orbitals. The sixth and seventh periods contain elements with valence electrons in s, p, d, or f orbitals. The lanthanides and actinides have valence electrons in f orbitals. Thus, as seen in the chart above, elements with high atomic number (compared with elements of low atomic number) have valence electrons in higher-energy orbitals.

### 2.3 Ions

Atoms are somewhat unstable if valence electrons do not completely fill outer shells. They are very stable when the outer shells are fully occupied. So, atoms commonly give up, or borrow electrons, producing ions, to obtain this stability. As a result, they may become cations, which have a net positive charge because there are more protons than electrons, or anions, with a net negative charge because there are more electrons than protons.
2.11 Forming an oxygen anion

The schematic drawings in Figure 2.11 show how a neutral oxygen atom can gain two electrons to become an anion. The neutral atom has 8 protons in its nucleus and 8 electrons. It contains two 1s electrons and six electrons in its outer shell (2s and 2p orbitals). Thus, 2 additional electrons are needed to fill the outer shell. Commonly, two electrons move in to stabilize the atom – and the result is an oxygen anion. The ionic charge (-2) is the number of protons (still 8) less the number of electrons (now 10).

Different cations and anions may have different ionic charge. We call all ions with a charge of +1 or -1 monovalent. Those with a charge of +2 or -2 are divalent. Trivalent and tetravalent refer to charges of +3 or -3 and +4 or -4. The oxygen ion shown above is a diatomic anion.

Some elements ionize more easily than others. Those that ionize most easily to become cations are called metallic elements. The degree to which elements are metallic generally decreases from left to right in the Periodic Table. The elements on the right side of the table, which ionize to become anions, are nonmetallic.

2.3.1 Typical Valences
2.12 Valences of typical ions

The chart seen here shows the most common ions for different columns in the Periodic Table. Common cations are shaded blue and common anions are shaded red. This table is not to suggest that all the elements always form ions. But, when they do, for most elements, their likely ionic form is predictable as shown.

The red numbers below the columns are typical ionic charges. Elements in groups on the left side of the table have “extra” electrons in outer shells, and readily give up those electrons to become cations. Elements in group 1 (alkali elements) generally have an ionic charge of +1. Those in group 2 (alkaline earth elements) usually have an ionic charge of +2.

On the other side of the chart, helium and other elements in Group 18 of the Periodic Table have completely filled outer shells and, consequently, do not ionize. Elements in Groups 16 and 17 have room for a small number of additional electrons in their outer shells and consequently accept extra electrons to become anions. Elements in group 17 (halogens) are lacking a single electron to fill their outer shells completely. So, they typically become monovalent anions (charge of -1) because they acquire an extra electron to fill the shell. Oxygen and other elements in Group 16 typically form divalent anions.
Elements in Group 13, Group 14, and Group 15 typically ionized to form cations with charge of +3, +4, and +5, respectively, although other valences are possible.

Ionization of the transition metals in the central portion of the table (groups 3 through 12) is less predictable than ionization of elements near the table’s sides. When ionized, elements in groups 3, 4, and 5 commonly have valences of +3, +4, and +5, respectively. The rest of the transition metals exist in a number of ionic states, typically +2 or +3.

The three most abundant elements in the Earth’s crust (oxygen, silicon, and aluminum) are on the right side of the Periodic Table, in groups 13, 14, and 16. Oxygen typically has a charge of -2, silicon of +4, and aluminum of +3.

2.4 Bonding in Minerals

2.4.1 Ionic Bonds

Negatively and positively charged particles attract each other. Protons (positively charged) attract electrons (negatively charged) in atoms. Similarly, positively charged cations attract negatively charged anions, producing ionic bonds in minerals.

![Ionic bonding to form halite (NaCl)](image)

We call the process of losing electrons to become a cation oxidation. Gaining electrons, and becoming an anion, is
reduction. The formation of cations and anions go hand in hand. Metals becoming cations donate electrons to non-metals becoming anions. The drawing shown in Figure 2.13 is a schematic showing how sodium and chlorine can combine to make the mineral halite (NaCl). First sodium (Na) and chlorine (Cl) react to produce the ions Na\(^+\) and Cl\(^-\) (when a 3s electron from sodium is added to the 3p orbital in chlorine). Na is oxidized as Cl is reduced. These two ions then bond, producing NaCl.

Oxidation and reduction also occur when iron and oxygen combine to form hematite (Fe\(_2\)O\(_3\)). Two Fe\(^0\) atoms each donate 3 electrons that are picked up by three O\(^0\) atoms. So, the iron atoms become Fe\(^{3+}\) while the oxygen atoms become O\(^{2-}\). The ions may then bond ionically to produce (Fe\(_2\)O\(_3\)). Alternatively, iron sometimes oxidizes to Fe\(^{2+}\) and combines with oxygen to form the mineral wüstite (FeO). A third iron oxide mineral, magnetite (Fe\(_3\)O\(_4\)), contains both Fe\(^{2+}\) and Fe\(^{3+}\). Wüstite, hematite and magnetite are all minerals, but wüstite is not a common mineral on Earth (although it is found in meteorites) because under normal Earth surface conditions, iron easily oxidizes to become trivalent. The bonds in all three iron oxides, however, are not entirely ionic; they are partially covalent (discussed later in this chapter).

### 2.4.1.1 Common Cations and Anions

<table>
<thead>
<tr>
<th>Common Cations, Anions, and Anionic Groups</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>cations</strong></td>
</tr>
<tr>
<td>Na(^+)</td>
</tr>
</tbody>
</table>
The table seen here lists the most common cations and anions in Earth’s crust. For the most part, these are the same elements that we discussed at the beginning of this chapter when we considered the composition of the crust and mantle.

While thinking of individual atoms bonding together to form minerals seems straightforward, in reality atoms are seldom unbonded to others. Single atoms are very reactive. They tend, when possible, to bond to other atoms to form molecules and often compounds. Sometimes, they bond to other atoms of the same element. For example, N\textsubscript{2}, composed of molecules containing two nitrogen atoms, dominates the Earth’s atmosphere. Small atoms with several valence electrons, such as silicon or carbon, are especially reactive. They seldom exist by themselves, readily combining with oxygen, and sometimes other elements, to form strongly bonded anionic units called \textit{anionic groups}, also called \textit{molecular ions} or \textit{polyatomic ions}. Molecular ions are so strongly bonded that they behave like individual anionic units in many minerals. So when minerals

<table>
<thead>
<tr>
<th>Cation</th>
<th>Anion</th>
<th>Combination</th>
</tr>
</thead>
<tbody>
<tr>
<td>K\textsuperscript{+}</td>
<td>F\textsuperscript{-}</td>
<td>(OH\textsuperscript{-})</td>
</tr>
<tr>
<td>H\textsuperscript{+}</td>
<td></td>
<td></td>
</tr>
<tr>
<td>\textit{divalent}</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Fe\textsuperscript{2+}</td>
<td>O\textsuperscript{2-}</td>
<td>(SiO\textsubscript{4})\textsuperscript{4-}</td>
</tr>
<tr>
<td>Ca\textsuperscript{2+}</td>
<td></td>
<td>(SO\textsubscript{4})\textsuperscript{2-}</td>
</tr>
<tr>
<td>Mg\textsuperscript{2+}</td>
<td></td>
<td>(CO\textsubscript{3})\textsuperscript{2-}</td>
</tr>
<tr>
<td></td>
<td></td>
<td>(WO\textsubscript{4})\textsuperscript{2-}</td>
</tr>
<tr>
<td>\textit{other ions}</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Al\textsuperscript{3+}</td>
<td></td>
<td>(BO\textsubscript{4})\textsuperscript{3-}</td>
</tr>
<tr>
<td>Fe\textsuperscript{3+}</td>
<td></td>
<td>(PO\textsubscript{4})\textsuperscript{3-}</td>
</tr>
<tr>
<td>Si\textsuperscript{4+}</td>
<td></td>
<td></td>
</tr>
<tr>
<td>C\textsuperscript{4+}</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
dissolve, molecular ions such as \((\text{CO}_3)^{2-}\) will not dissociate in water.

The right-hand column of the table lists the most common and important anionic groups. They include, from top to bottom, the nitrate, hydroxyl, silicate, sulfate, carbonate, tungstate, borate and phosphate groups. As pointed out in the previous chapter, we generally classify minerals based on their anion species because the properties of minerals with the same anions or anionic groups are generally very similar. So, we often write mineral formulas with parentheses to emphasize any anions or anionic groups that are present.

Any of the cation species listed in the table can combine with the anions or molecular anions to produce a long list of different ionic compounds. For example, all these minerals contain \(\text{Ca}^{2+}\):

<table>
<thead>
<tr>
<th>lime</th>
<th>fluorite</th>
<th>larnite</th>
<th>anhydrite</th>
<th>calcite</th>
<th>scheelite</th>
</tr>
</thead>
<tbody>
<tr>
<td>CaO</td>
<td>CaF(_2)</td>
<td>Ca(_2)SiO(_4)</td>
<td>CaSO(_4)</td>
<td>CaCO(_3)</td>
<td>CaWO(_4)</td>
</tr>
</tbody>
</table>

Lime, fluorite, larnite, anhydrite, calcite, and scheelite belong, respectively to the oxide, halide, silicate, sulfate, carbonate, and tungstate mineral groups. Additional \(\text{Ca}^{2+}\) minerals with more complicated formulas include sinjarite \((\text{CaCl}_2 \cdot 2\text{H}_2\text{O})\), colemanite \((\text{Ca}_2\text{B}_6\text{O}_{11} \cdot 5\text{H}_2\text{O})\), nitrocalcite \((\text{Ca(NO}_3)_2 \cdot 4\text{H}_2\text{O})\), and apatite \((\text{Ca}_5(\text{PO}_4)_3(\text{OH}))\). Thus, \(\text{Ca}^{2+}\) can combine to make minerals with all except one of the anion and anionic species listed in the table above. The lone exception is nitrate, \((\text{NO}_3)^{-}\). But, synthetic calcium nitrate \((\text{CaNO}_3)\), called Norwegian saltpeter is sometimes manufactured for industrial use.

### 2.4.2 Covalent Bonds

Rather than giving up or gaining electrons to become ions,
atoms with incompletely filled energy levels may become stabilized by sharing electrons. The sharing of pairs of electrons between atoms produces a covalent bond. If the sharing is complete, the bond is 100% covalent. This is the case for diatomic gases such as N\textsubscript{2}, but does not occur in common minerals except for diamond.

2.14 Covalent bonding to make hydrogen gas

The simplest example of a covalent bond involves two hydrogen atoms combining to produce H\textsubscript{2} gas, shown in Figure 2.14. Hydrogen atoms have one proton in their nucleus and one electron in orbit. If two hydrogen atoms share electrons, this effectively puts two electrons in the outer orbits (1s) of each atom, completely filling the orbits and making a very stable diatomic molecule, H\textsubscript{2}. Fluorine, chlorine, and other halogen elements form covalently bonded diatomic gasses (F\textsubscript{2}, Cl\textsubscript{2}, etc.) like hydrogen.

2.15 Covalent bonding to make oxygen gas

Sometimes covalent bonds involve more than a single pair of shared electrons. Figure 2.15 shows how two oxygen atoms can share four electrons to form covalent bonds, and produce O\textsubscript{2}.
gas. By sharing four electrons, each atom has a total of 8 electrons in its 2s and 2p orbitals, making the outer shells fully occupied and stable. Because there are two pairs of electrons shared, we call this kind of covalent bond a double bond.

2.16 Simplified view of bonding in quartz

For a mineralogical example of covalent bonding, let’s consider quartz (SiO$_2$). The drawing shown in Figure 2.16 is the atomic arrangement in quartz reduced to two dimensions. Four oxygen atoms surround each silicon atom, and each oxygen atom bonds to two silicon atoms. A lone silicon atom has 4 electrons in its outer shell (3s and 3p orbitals). But, each of the surrounding oxygen atoms can share an electron with the silicon atom, bringing the total number of electrons to 8, filling the outer shell, and making it stable. And, each silicon atom can share one electron with each of its four surrounding oxygen atoms. So, oxygen, which begins with 6 electrons in its outer level (2s and 2p orbitals), gains 2 electrons, bringing the total to 8, which stabilizes the atom.

Many elements form both covalent and ionic bonds. As mentioned
previously, Cl\(^-\) forms ionic bonds with Na\(^+\) to make the mineral halite. But, two Cl atoms can also bond covalently to form a stable gas, Cl\(_2\). The two types of bonds are not exclusive – ionic bonds and covalent bonds form a spectrum, and in nature, 100% ionic and covalent bonds do not exist. In nearly 100% ionic bonds, valence electrons are almost exclusively associated with one atom. (But, even the highly ionic NaCl bond has a slight covalent character.) In nearly 100% covalent bonds, electrons are shared more or less evenly between two atoms. Diatomic gases such as H\(_2\) and N\(_2\) are extremely close to being 100% covalent.

So, many minerals (and other substances) contain bonds that are partly ionic and partly covalent. For example, the preceding discussion of covalent bonding in quartz was an oversimplification. In actuality, silicon-oxygen bonds are not completely covalent – there is still some ionization and electrons are more localized around oxygen than silicon. The overall bonding is about half covalent and half ionic. And, although they behave as anions, (SiO\(_4\))\(^2-\), (SO\(_4\))\(^2-\), (CO\(_3\))\(^2-\), (OH)\(^-\) and all other molecular ions (listed in the table above in Section 2.4.1.1) are held together by partially covalent bonds. This explains why they act as individual ionic units in ionic crystals.

2.4.3 Metallic Bonds
2.17 Metallic bonding

In covalent bonding, pairs of atoms share electrons. In a third kind of bonding, \textit{metallic bonding}, many atoms share the same electrons. Individual atoms give up their valence electrons and the delocalized electrons are free to move and interact with all the positively charged ions in the structure (Figure 2.17). Metallic bonding, is especially common in minerals involving transition metals. Gold, silver, and copper are examples of minerals with metallic bonds. Because valence electrons move easily throughout the structure, metallically bonded compounds are good conductors of heat and electricity. Electrons are easily transferred along wires, for example. Minerals with metallic bonds may be malleable and have only low-to-moderate hardness, reflecting the loose nature of their metallic bonds.

2.4.4 Other Kinds of Bonds

Besides the three kinds of bonds just discussed, some minerals include other types of bonds such as \textit{van der Waals bonds} and \textit{hydrogen bonds}. These bonds do not involve valence electrons but instead result from weak electrostatic forces due to uneven charge distribution in a crystal structure. Very weak van der Waals bonds are important in graphite and some clay
minerals, for example. This explains why graphite is much softer than diamond, which has the same composition but covalent bonds. In the mineral *brucite*, combinations of hydrogen and van der Waals bonds hold sheets of MgO₆ molecules together. Clay minerals have excellent cleavage because covalent and ionic bonds create strongly bonded layers, but weak van der Waals and hydrogen bonds hold the layers together. So, the minerals split easily into sheets. Because most mineral properties are explained by ionic, covalent, or metallic bonds and we can ignore van der Waals and hydrogen bonds for most purposes.

### 2.4.5 Ionic, Covalent, and Metallic Crystals

As shown in the triangular diagram (Figure 2.18), bonding in most minerals is neither 100% ionic, 100% covalent, nor 100% metallic, but some come close. Most minerals contain combinations of ionic and covalent bonding. Metallic ore minerals such as pyrite (FeS₂), stibnite (Sb₂S₃), and copper (Cu) generally have little ionic character. Most of them, especially those in the sulfide and sulfosalts groups, contain
combinations of covalent and metallic bonding. Metallic and ionic bonds do not often combine, although galena (PbS) may be one example in which they do to a limited extent.

The degree to which a bond is ionic depends on both elements involved. For example, because alkali elements (group 1) have a very strong tendency to become cations, and halogens (group 17) have an equal tendency to become anions, halite (NaCl) and other alkali halides form crystals in which bonds are nearly 100% ionic. Fluorite (CaF$_2$), too, is nearly completely ionic. Alkaline earth oxides such as periclase (MgO) or lime (CaO), involving cations from group 2 and oxygen from group 16, are about 75% ionic. Many other oxides too, are mostly ionic. Silicates, the most common kind of mineral in Earth’s crust, are generally about 50% ionic and 50% covalent. And, bonds in diamond (C) are entirely covalent.

### 2.4.5.1 Bonding and Mineral Properties

The nature of chemical bonding in minerals controls many properties. The table below makes some comparisons. Covalent bonds are stronger than ionic bonds, which in turn are stronger than metallic bonds. Consequently, covalent minerals are hard and tenacious, while metallic ones are usually not, and ionic minerals fall between. Melting temperatures follow a similar pattern (covalent compounds generally melt at the highest temperatures due to their stronger bonds), while solubility in water is greatest for ionic crystals (because of their weak bonds and easy ionization).

Bond type also affects crystal symmetry. Both ionic and metallic bonds are nondirectional so bonding can occur equally in all directions. In contrast, covalent bonds involve pairs of atoms and are linear. Consequently, metallic and ionic minerals generally have high symmetry compared with covalent ones.
### Characteristics of Minerals Dominated by Different Kinds of Bonds

<table>
<thead>
<tr>
<th>property</th>
<th>ionic bonds</th>
<th>covalent bonds</th>
<th>metallic bonds</th>
</tr>
</thead>
<tbody>
<tr>
<td>common elements involved</td>
<td>from opposite sides of the Periodic Chart</td>
<td>close together in the Periodic Chart</td>
<td>transition metals</td>
</tr>
<tr>
<td>electrical conductivity</td>
<td>low</td>
<td>low</td>
<td>high</td>
</tr>
<tr>
<td>thermal conductivity</td>
<td>low</td>
<td>low</td>
<td>high</td>
</tr>
<tr>
<td>solubility in water</td>
<td>high</td>
<td>low</td>
<td>very low</td>
</tr>
<tr>
<td>melting temperature</td>
<td>moderate to high</td>
<td>high</td>
<td>variable</td>
</tr>
<tr>
<td>hardness</td>
<td>medium to hard</td>
<td>very hard</td>
<td>often malleable</td>
</tr>
<tr>
<td>ability to break</td>
<td>brittle, good cleavage</td>
<td>brittle, common fractures</td>
<td>variable</td>
</tr>
<tr>
<td>crystal symmetry</td>
<td>high symmetry</td>
<td>low symmetry</td>
<td>very high symmetry</td>
</tr>
<tr>
<td>transparency</td>
<td>fully to partially transparent</td>
<td>partially transparent or opaque</td>
<td>opaque</td>
</tr>
<tr>
<td>how common?</td>
<td>most nonmetallic minerals</td>
<td>some minerals</td>
<td>most metallic minerals</td>
</tr>
<tr>
<td>examples</td>
<td>halite (NaCl); calcite (CaCO₃)</td>
<td>diamond (C); sphalerite (ZnS)</td>
<td>copper (Cu); silver (Ag)</td>
</tr>
</tbody>
</table>

#### 2.5 Compositional Variation in Minerals
Most natural minerals are formed from combinations of many elements. Unlike mechanical mixtures (such as two things being ground up together) the elements are ordered and intimately bonded together. The resulting crystalline solutions are solid solutions. Common plagioclase (a feldspar), for example, is mostly a solution of $\text{CaAl}_2\text{Si}_2\text{O}_8$ (anorthite) and $\text{NaAlSi}_3\text{O}_8$ (albite). It may also contain a small amount of orthoclase ($\text{KAlSi}_3\text{O}_8$). Anorthite, albite, and orthoclase are plagioclase end-members. End-member formulas describe ideal components that make up solid-solution minerals. The plagioclase sample seen in Figure 2.19 contains blades of a variety of plagioclase called cleavelandite accompanied by a few small flakes of silver-gray muscovite. Cleavelandite has composition close to end-member albite.

Some minerals, such as hornblende, contain many elements, and are complex solid solutions that have long and complicated formulas. Other minerals, including plagioclase and other feldspars, have limited compositional variations. Still others, such as fluorite ($\text{CaF}_2$) or quartz ($\text{SiO}_2$), have relatively simple formulas and vary little from their ideal compositions.
We may broadly classify the elements that comprise a mineral in one of three categories: major elements, minor elements, and trace elements. Major elements are those fundamental to a mineral; they control a mineral’s basic atomic structure and gross properties. They must be in the mineral or the mineral has been misidentified. Minor elements are those present in small amounts, usually as replacements for a major element. Such elements, perhaps in amounts up to a several weight %, may affect color and a few other properties, but the basic atomic arrangement of a mineral is controlled by its major element chemistry. Minerals also contain extremely small amounts of elements called trace elements. Trace elements are in all minerals and provide valuable information for geologists attempting to determine how, when, and where specific minerals formed. They have little effect on most mineral properties. A notable exception to this is sometimes color; even trace amounts of some elements can have major effects on a mineral’s color.

Mineralogists can acquire mineral analyses in many ways. In the past, most chemical analyses were determined by titration and other “wet chemical” techniques. Today we use sophisticated analytical instruments, including atomic absorption spectrophotometers and electron microprobes. We normally report analytical results by listing oxide weight percentages. We must convert weight percent values to mole value if we wish to have mineral formulas.
As an example of chemical variability in minerals, let’s consider olivine. An example of olivine is shown in Figure 2.20, above, and the table below lists analyses of six olivines from different geologic environments. The top part of the table presents the analyses in terms of element weight %; the middle part presents the same information in terms of oxide weight %; the bottom part presents the composition in terms of the number of atoms in an olivine formula. The three parts of the table are redundant; values from one can be converted to the others by an arithmetical process called normalization (described later in this chapter). Most mineralogists and petrologists prefer to consider oxide weight % (for reasons explained in a later chapter), or numbers of atoms (because numbers of atoms directly translate into mineral formulas), instead of element weight %.

<table>
<thead>
<tr>
<th>Source:</th>
<th>Burma</th>
<th>Finland</th>
<th>New Zealand</th>
<th>Minnesota</th>
<th>East Greenland</th>
<th>Germany</th>
</tr>
</thead>
<tbody>
<tr>
<td>Element Weight %</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
As the analyses above show, Mg, Fe, and Si are major elements in olivines. Olivine may also contain Ca, Mn, Ti, and sometimes Ni or Cr, but except in extremely rare
circumstances, these elements are minor or trace elements.

The compositions of these six olivines are all different. Silicon, magnesium, and iron are especially variable. How can we make better sense of it all? The answer is to use the atom numbers in the bottom third of the table above to write formulas for each olivine. The ideal stoichiometry (ratios of elements) of olivine is \((R)_{2}SiO_{4}\), where \(R\) is usually Fe, Mg, Mn, or Ca. Thus, a general formula is \((Fe,Mg,Mn,Ca)_{2}SiO_{4}\) and a good olivine analysis should produce a formula close to this. Formulas for each of the six olivines – seen in the table below – come out to be very close to ideal. Apparent deviations from ideal stoichiometry (for example, the numbers of Si in the formulas are not exactly equal to 1) can mostly be attributed to inclusions of other minerals within the olivine, or to analytical error.

### Normalized Olivine Formulas

<table>
<thead>
<tr>
<th>specimen</th>
<th>source</th>
<th>formula</th>
<th>%</th>
<th>%</th>
</tr>
</thead>
<tbody>
<tr>
<td>Burma</td>
<td></td>
<td>((Mg_{2.022}Fe_{0.022})Si_{0.978}O_{4})</td>
<td>98</td>
<td>1</td>
</tr>
<tr>
<td>Finland</td>
<td></td>
<td>((Mg_{1.899}Fe_{0.086}Mn_{0.004}Al_{0.015}Ti_{0.001})Si_{0.981}O_{4})</td>
<td>95</td>
<td>4</td>
</tr>
<tr>
<td>New Zealand</td>
<td></td>
<td>((Mg_{1.848}Fe_{0.157}Mn_{0.063}Ca_{0.046}Al_{0.006})Si_{0.979}O_{4})</td>
<td>92</td>
<td>7</td>
</tr>
<tr>
<td>Minnesota</td>
<td></td>
<td>((Mg_{0.383}Fe_{1.516}Ca_{0.044}Al_{0.018}Ti_{0.002})Si_{0.957}O_{4})</td>
<td>20</td>
<td>8</td>
</tr>
<tr>
<td>East Greenland</td>
<td></td>
<td>((Mg_{0.65}Fe_{1.817}Mn_{0.023}Ca_{0.078}Al_{0.003}Ti_{0.005})Si_{1.001}O_{4})</td>
<td>2</td>
<td>97</td>
</tr>
<tr>
<td>Germany</td>
<td></td>
<td>((Fe_{1.978}Mn_{0.088})Si_{1.011}O_{4})</td>
<td>0</td>
<td>100</td>
</tr>
</tbody>
</table>

Chemical substitutions in some minerals involve many elements and are complex. In olivine the substitutions are relatively simple. We can describe the compositions of most natural olivines as combinations of the end members (represented by ideal formulas) called forsterite \((Mg_{2}SiO_{4})\), fayalite \((Fe_{2}SiO_{4})\), tephroite \((Mn_{2}SiO_{4})\), and calcio-olivine \((Ca_{2}SiO_{4})\). In contrast with olivine, many minerals (for example amphiboles and micas)
have lengthy formulas and have many elements substituting in their structures. For such minerals, choosing useful end members is difficult and arbitrary.

The six olivines considered above contain very little besides forsterite and fayalite. This is typical for most olivines and, if we ignore the other components, we get the percent values in the two columns on the right in the table. The % forsterite and % fayalite were calculated, respectively, from the ratio of Mg/(Mg+Fe) and the ratio of Fe/(Mg+Fe) in each olivine. The binary (2-component) diagram below (Figure 2.21) shows the compositions plotted on a line anchored by pure forsterite on the left, and pure fayalite on the right. The sample from Burma is nearly 100% forsterite (Mg$_2$SiO$_4$), while that from Germany is 100% fayalite (Fe$_2$SiO$_4$). The other olivines fall between.

![Figure 2.21 Comparing olivine compositions](image)

The degree to which elements may substitute for each other depends on the elements and on the mineral. In olivine, Fe and Mg mix freely, so any composition between fayalite and forsterite is possible. Olivines can incorporate only minor amounts of Ca, however, so no compositions midway between calcio-olivine and forsterite are found in nature. In contrast, Ca, Mg, and Fe mix freely in some other minerals, garnets for example. Natural garnets can have any composition between end members grossular (Ca$_3$Al$_2$Si$_3$O$_{12}$), almandine (Fe$_3$Al$_2$Si$_3$O$_{12}$), and pyrope (Mg$_3$Al$_2$Si$_3$O$_{12}$).
2.5.1 How to Normalize a Mineral Analysis

We normally report mineral analyses in values of oxide weight %, but mineral formulas are written in terms of numbers of atoms. **Normalization** is the process of converting an analysis into a formula. Normalization is a tedious but straightforward arithmetical operation when done by hand. Fortunately there are Excel spreadsheets and programs to do it for you.

The table below gives the chemical analysis of a feldspar from Grorud, Norway. Column A lists the oxides and column B gives the results of the analysis. Columns C, D, E, and F give chemical data for the oxides.

<table>
<thead>
<tr>
<th>oxide</th>
<th>oxide wt% in mineral</th>
<th>atomic wt of oxide</th>
<th>cation</th>
<th># of cations in oxide</th>
<th># of oxygen in oxide</th>
<th># moles oxide value</th>
<th># moles cation in mineral</th>
<th># moles oxygen from each oxide</th>
<th># moles oxygen from each oxygen</th>
<th># of cations normalized to 8 oxygen</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>65.90</td>
<td>60.085</td>
<td>Si⁺⁺</td>
<td>1</td>
<td>2</td>
<td>0.19076</td>
<td>0.382</td>
<td>0.572</td>
<td>1.025</td>
<td></td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>19.45</td>
<td>191.961</td>
<td>Al³⁺</td>
<td>2</td>
<td>3</td>
<td>0.00645</td>
<td>0.013</td>
<td>0.019</td>
<td>0.005</td>
<td></td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>1.03</td>
<td>159.692</td>
<td>Fe²⁺</td>
<td>2</td>
<td>3</td>
<td>0.01088</td>
<td>0.011</td>
<td>0.011</td>
<td>0.029</td>
<td></td>
</tr>
<tr>
<td>CaO</td>
<td>0.61</td>
<td>56.079</td>
<td>Ca²⁺</td>
<td>1</td>
<td>1</td>
<td>0.11488</td>
<td>0.230</td>
<td>0.115</td>
<td>0.617</td>
<td></td>
</tr>
<tr>
<td>Na₂O</td>
<td>7.12</td>
<td>61.979</td>
<td>Na⁺⁺</td>
<td>2</td>
<td>1</td>
<td>0.06582</td>
<td>0.132</td>
<td>0.066</td>
<td>0.354</td>
<td></td>
</tr>
<tr>
<td>K₂O</td>
<td>6.20</td>
<td>94.195</td>
<td>K⁺⁺</td>
<td>2</td>
<td>1</td>
<td>1.48556</td>
<td>1.863</td>
<td>2.977</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Total</td>
<td>100.31</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

The Grorud feldspar contains 65.90 wt % SiO₂, 19.45 wt % Al₂O₃, 1.03 wt % Fe₂O₃, 0.61 wt % CaO, 7.12 wt % Na₂O, and 6.20 wt % K₂O. To convert from % grams of oxides to moles of oxides, we divide the oxide weight % values (column B) by oxide atomic weight (column C); column G shows the results. The values in column G are relative values only; they total to 1.486, which has no scientific meaning.

To calculate relative numbers of cations, we multiply the values in column G (moles of oxides) by the number of cations in each oxide (column E). To calculate moles of oxygen, we multiply the values in column G by the number of O in each oxide (column F). Columns H and I give the results of these
calculations.

Using the numbers in column H and the oxygen total of 2.977, we could write the formula of the feldspar as $\text{Ca}_{0.011}\text{Na}_{0.230}\text{K}_{0.132}\text{Fe}_{0.013}\text{Al}_{0.382}\text{Si}_{1.097}\text{O}_{2.977}$, but normal feldspar analyses are written with 8 atoms of oxygen. So, in the last column (J) we have multiplied all the atom numbers in columns H and I by a constant factor so that the number of atoms of oxygen is exactly 8. The values in column J give us the normalized formula for the feldspar $(\text{Ca}_{0.029}\text{Na}_{0.617}\text{K}_{0.354})(\text{Fe}_{0.035}\text{Al}_{1.025}\text{Si}_{2.948})\text{O}_{8}$. The parentheses remind us which elements substitute for each other.

As pointed out earlier in this chapter, most feldspars are primarily solutions of three end members: anorthite ($\text{CaAl}_{2}\text{Si}_{2}\text{O}_{8}$), albite ($\text{NaAlSi}_{3}\text{O}_{8}$), and orthoclase ($\text{KAlSi}_{3}\text{O}_{8}$). So, ideal feldspar formulas have stoichiometry described by $(\text{Ca,Na,K})(\text{Al,Si})_{4}\text{O}_{8}$; the Grorud feldspar comes very close to this. Any discrepancy is due to analytical imprecision.

If we ignore the small amount of Fe that is present, the Grurud feldspar composition can be described as:

- $X_{\text{An}} = 0.029 \text{ CaAl}_{2}\text{Si}_{2}\text{O}_{8} (= 2.9\% \text{ anorthite})$
- $X_{\text{Ab}} = 0.617 \text{ NaAlSi}_{3}\text{O}_{8} (= 61.7\% \text{ albite})$
- $X_{\text{Or}} = 0.354 \text{ KAlSi}_{3}\text{O}_{8} (= 35.4\% \text{ orthoclase})$

The numbers 0.029, 0.617, and 0.354 are called mole fractions; they are equivalent to percent divided by 100.

Above we plotted olivine compositions on a line – on a binary diagram. Lines work for depicting solid solutions with two end members but not for solutions, like feldspars, that have three. We use triangular (ternary) diagrams for that purpose. And, for solutions with more than three end members, we are out of luck because we cannot easily draft diagrams in more than two dimensions.
The diagram seen in Figure 2.22 is a ternary feldspar diagram with end members anorthite, albite, and orthoclase at the corners. The red numbers label lines of constant anorthite content, the purple numbers label lines of constant orthoclase content, and the green numbers label lines of constant albite content. Thus, any feldspar made of combinations of albite, anorthite, and orthoclase can be plotted on diagrams like this.

The yellow region shows a range of compositions that includes most common feldspars. Except in very high temperature igneous rocks, feldspars are either alkali feldspars (solutions dominated by orthoclase and albite) or plagioclase (solutions dominated by albite and anorthite). Solutions between anorthite and orthoclase do not exist in nature. The white dot shows where the composition of the Grorud feldspar plots; it is an alkali feldspar that is 62% albite.
2.6 Common Minerals

We classify rock-forming minerals in many ways. Often, geologists contrast primary minerals and secondary minerals. Primary minerals are those that are present from the time a rock first forms. Secondary minerals form later by chemical or physical reaction within the rock. Often, such secondary reactions involve H₂O or CO₂ and occur during weathering, diagenesis, or other low-temperature alteration of a preexisting rock. The distinction between primary and secondary minerals is not exact. A primary mineral in one rock may be a secondary mineral in another.

We further divide primary minerals into essential minerals and accessory minerals. Essential minerals are those whose presence is implied by the name of the rock. All limestones, for example, contain calcite or dolomite, and all granites contain quartz and K-feldspar. Essential minerals, for the most part, control rock properties.

Accessory minerals are generally present in minor amounts and do not affect most rock properties. These minerals may be made of common elements such as iron in magnetite (which is a common accessory mineral). Accessory minerals also commonly contain incompatible elements, elements that are not easily incorporated into essential minerals. Zirconium (Zr), for example, often concentrates in zircon, ZrSiO₄, a minor accessory mineral in many rocks. Phosphorus (P) may lead to formation of phosphate minerals such as apatite, Ca₅(PO₄)₃(OH,F,Cl), also a common accessory mineral. In most igneous and metamorphic rocks, silicates make up the essential minerals, while oxides, sulfides, and other kinds of minerals make up the accessory minerals.

2.6.1 Common Elements Make Up the Common
Minerals

Although more than 90 natural elements exist on Earth, many exist in very small amounts. Occasionally they may be concentrated in accessory minerals, but more often they are just minor diluents for major elements. It is estimated that 14 elements (O, Si, Al, Fe, Ca, Na, K, Mg, Ti, H, P, Mn, Ba, C) make up more than 99.7 wt % of Earth’s crust. We can expect the common minerals to be made from these elements. And, we can expect minerals made of other elements to be rare.

2.23 Common minerals in the crust

Mineralogists have identified more than 5,000 minerals. Of these, fewer than 200 are common, and fewer than 50 are common enough to be considered essential minerals. Abundant oxygen and silicon, together making up nearly three quarters of Earth’s crust, naturally lead to an abundance of silicate minerals. It is no wonder that other types of minerals, or minerals that include rarer elements as key components, are less common. Of the known minerals, about 92% are silicates. Thus, all the nonsilicates together only make up about 8% of the total.
Plagioclase, a feldspar

The pie diagram above in Figure 2.23 shows estimated abundances of different minerals in the crust. Plagioclase, like the plagioclase shown in this photo (Figure 2.24), is the most abundant mineral (40% of the total) in Earth’s crust because it comprises more than half of the basalt that makes up the oceanic crust. Alkali feldspar and quartz, the major components in granite, make up about another 25%. The other minerals, combined, account for 37% of the total.

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