CHAPTER 3

The Moon

3.1 ON THE MOON

The Moon's Vital Statistics

Radius	1738 km
Surface area	3.79×10^7 km^2
Mass	7.35×10^{22} kg
Density	3.36 g/cm^3
Local gravity	1.62 m/s^2
Escape velocity	2.38 km/s
Albedo	0.07–0.24
Surface temperature	380 K (day) 120 K (night)
Sidereal period	27.32 days
Distance to Earth	384,401 km

What is it like to stand on another planet? The Moon is the one place, other than Earth, about which a human being can actually answer that question firsthand, because it is the one other planet where people actually have stood, picked up samples, and walked on the surface (Figure 3.2).

The pull of gravity one feels is one sixth that on Earth. Not only do you feel lighter, but things fall more slowly. Astronauts tended to travel across the lunar surface by hopping. Because the gravity is much less, it took less effort for them to get off the ground and they seemed to float for a longer time before they settled back down to the Moon.

The soil is white and powdery, a very fine sand, with the consistency of the scouring powder you might use to clean a kitchen sink. There are larger rocks on the surface as well, but many of them turn out to be a kind of rock called a **breccia**, a conglomeration of unrelated bits of stone and sand that have been packed together into boulder-sized chunks.

The terrain comes in two distinct types. Most of the Apollo landings took place on **maria** regions. These are remarkably flat plains that stretch for hundreds of kilometers, interrupted only by an occasional crater or by long cracks, called **rilles**, that can range from a few hundred meters to several kilometers wide and run for tens to hundreds of kilometers along the flat plains. The other kind of terrain is called the **highlands**, mountainous regions of bright white rock towering over the maria and covered with craters.

Because the Moon is smaller than Earth, the horizon appears closer. The eye, trained to Earth perspectives, has a hard time judging the size or distance of these soft, round-topped hills. Another factor confusing the perspective is the lack of atmosphere. There's no hazy air to obscure far-off mountains. The sky is black and stars are visible,

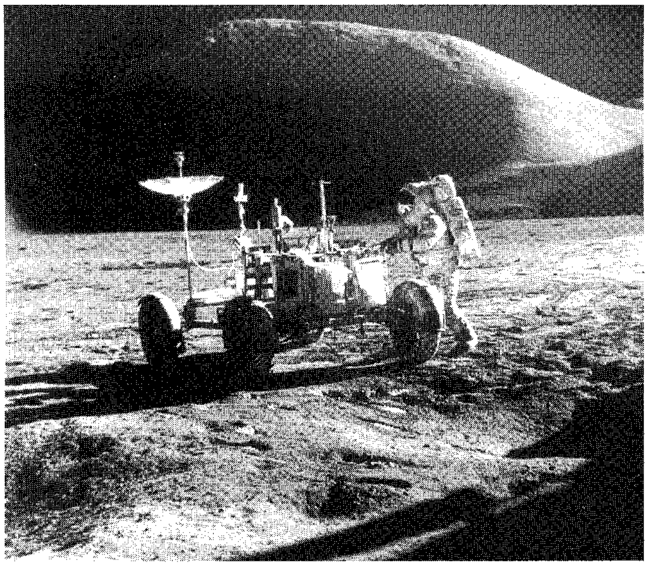

FIGURE 3.2 *Apollo 15* buggy on the Moon.

even during the daytime. At night, when you don't have to worry about the glare of the Sun reflecting off the grayish white rocks, you can see the stars more brightly and more clearly than anywhere on Earth. The familiar constellations of bright stars can get lost in a background of fainter stars many times more numerous than you are used to seeing on Earth.

Because there's no air, you'll have to be inside a spacesuit, of course. The suit also has to have a heating and cooling system, because the Sun is relentless during the day but its heat is quickly lost at night. Temperatures range from 380 K during the day (water boils at 373 K on Earth) to 120 K at night. Your cooling system has a special difficulty on the Moon. Like all refrigerators, it has to dump the excess heat it has removed from your spacesuit to some place outside the suit. On Earth, air conditioners just let air outside the space being cooled carry the heat away. But on the airless Moon the heat has to be radiated away as infrared photons, which is a much less efficient process.

This brief picture of the Moon's surface raises many major questions we would want to answer concerning the Moon. Why is its gravity so low? What is the soil made of, and why is it powdery? What scattered the rocks, and what compacted them back together into breccias? How and when were the highlands made? How were the maria made? Why are they low and flat and darker than the highlands?

Global Properties of the Moon

First, let's look at the general properties of the whole Moon: its size and shape, its gravity field, and its magnetic field.

To the naked eye, the Moon seems perfectly round; and indeed, it is rounder and smoother than any billiard ball. Its radius is 1738 km; in comparison, a mountain or crater wall 2 km high is less noticeable than a scratch a thousandth of an inch deep on a billiard ball.

But by carefully measuring the position of the Moon in its orbit, we can determine the position of the center of mass of the Moon, and the center of the mass is not located at the center of the 1738-km ball. There appears to be more mass in the half of the Moon facing Earth than in the far hemisphere; enough that the center of mass is 3 km closer to Earth than is the **center of figure** (the geometrical center).

The Apollo spacecraft landed seismometers on the Moon's surface. Using data they collected on "moonquakes," we can determine the structure of the lunar interior (Figure 3.3). What we find is that the Moon has a thick crust of low-density rocks, presumably similar in composition to the samples returned from the highlands regions, made of **feldspars** and **pyroxenes**. This crust is up to 120 km thick on the far side, but only 60 km thick on the side near Earth. The far side is nearly all highlands (Figure 3.4), while virtually all the mare regions are on the near side, and the rocks from these maria are basalts rich in **olivine**, much denser than highland rocks.

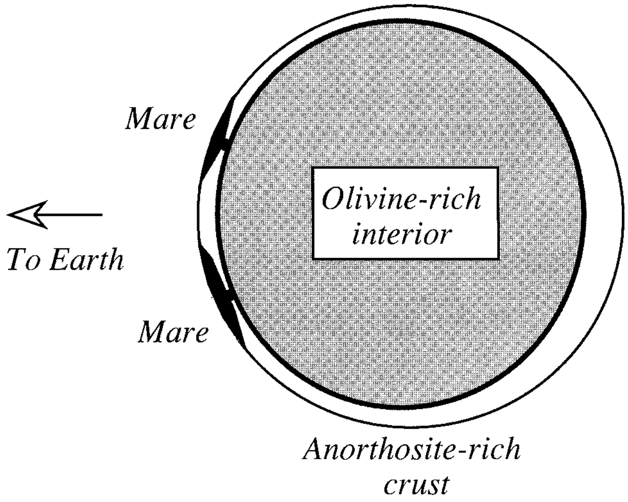

FIGURE 3.3 The distribution of material inside the Moon. The thickness of the crust is exaggerated.

One result is that over the mare basalts, the local force of gravity is found to be slightly stronger than the average gravity on the surface. This is caused by the extra concentration of mass that the thick, dense mare basalts have placed on the lunar surface. These **mass con**centrations are called **mascons**. By measuring the inferred extra mass of each mascon, one can determine that the mare basalts range from 0.5 km thick over the Oceanus Procellarum area to more than 5 km thick in the smaller round basins such as Imbrium.

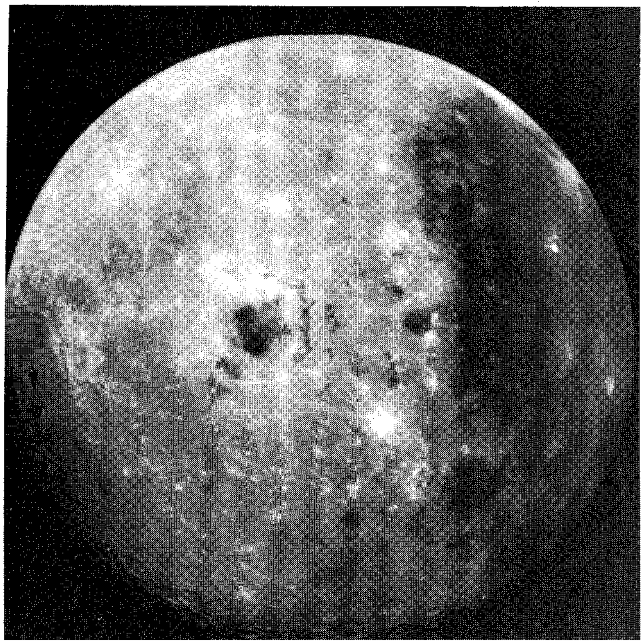

FIGURE 3.4 The far side of the Moon, photographed by the *Galileo* space probe.

The magnetic field is unusual in the vicinities of the mare basalts. The Moon as a whole does not have any magnetic field, but individual Moon rocks from the mare regions are slightly magnetic, typical of rocks that froze out of a lava in the presence of a magnetic field. The mare rocks are over 3 billion years old, and so it seems likely that 3 billion years ago the Moon did have a magnetic field, even though it does not have one now.

Planetary magnetic fields are thought to be generated in a region of molten metal, the **core**, that forms the central part of a planet. A planet without such a metal core would not have a magnetic field; or conversely, a planet with a magnetic field must have a metal core. The evidence for early lunar magnetism implies that the Moon must have a core, which was once molten but now is frozen.

If it does have such a core, it must be relatively small, less than a few hundred kilometers in radius, because the Moon's average density is low, lower than any other terrestrial planet. The core of Earth, by comparison, extends out to nearly half the distance from the center to the surface of Earth.

The Moon as a whole has the same density as an ordinary rock, 3.36 g/cm^3. But the average density of the other planets is far greater than the density of the rocks we see on their surfaces. This means that the other planets must have some dense material in their cores, covered over by ordinary rock, to raise up their average densities. From this one can conclude that some dense core component is present in other planets, but not in the Moon. The only material that is both dense and plentiful deep inside planets is metallic iron; thus it follows that the Moon does not have the metallic iron that other planets have and so cannot have a large metallic core.

The Moon is both smaller than the other terrestrial planets and made only of low-density material, and so it has less total mass than any other terrestrial planet. That is why the acceleration of gravity on the Moon's surface is so low: The less mass a planet has, the less its gravitational pull.

The Moon's Surface: Physical Aspects

The most numerous surface features on the Moon are craters. These craters come in all sizes, ranging from 100-km-wide basins like Orientale to tiny microscopic pits one millionth of a meter in diameter. The lunar craters are formed by meteorites hitting the surface of the Moon; the large basins are formed by the impact of huge rocks, kilometers in size, while the microscopic pits are made by grains of dust hitting a Moon rock.

Besides peppering the surface with round, bowl-shaped features, meteorite impacts have another effect. They are the primary way that the surface features of the Moon become eroded. The little impacts by tiny meteorites, which occur continuously, unimpeded by any atmosphere, slowly grind down the rough edges of the larger craters and break up the Moon rocks into ever smaller pieces. The result is that the surface of the Moon today is covered with a sandy "soil" tens of meters deep, called the **regolith**.

The continued impacts tend to mix up both dark mare rocks and light-colored highlands material and keep mixing the soil already on the surface constantly. This process, called **gardening**, means that over a few million years material from the bottom of the regolith is eventually uncovered and brought up to the surface by the force of an impacting meteorite; then, as more impacts occur nearby, other debris eventually buries this material back deep into the regolith. In this process, fragments of rock from originally far-flung parts of the Moon may be compressed together into the conglomerates called **breccias** (Figure 3.5).

The rate at which material impacted the Moon's surface was much greater in the first billion years of the Moon's existence than it is now, as debris left over from the formation of the solar system was slowly swept up by the newly formed planets. During this period the highland regions became heavily cratered and the large circular basins were formed. The energy of the impacts that made these basins was on the order of 10^{24} J, comparable to billions of nuclear explosions set off simultaneously. Debris blasted out from such impacts cut scars across the highlands mountains. These basins are often surrounded by rings, marking the collapse of material in the region of the basin after the impact, as shown in Figure 3.6.

After such a basin was made, often long afterwards, molten rock from the interior of the Moon erupted to flood

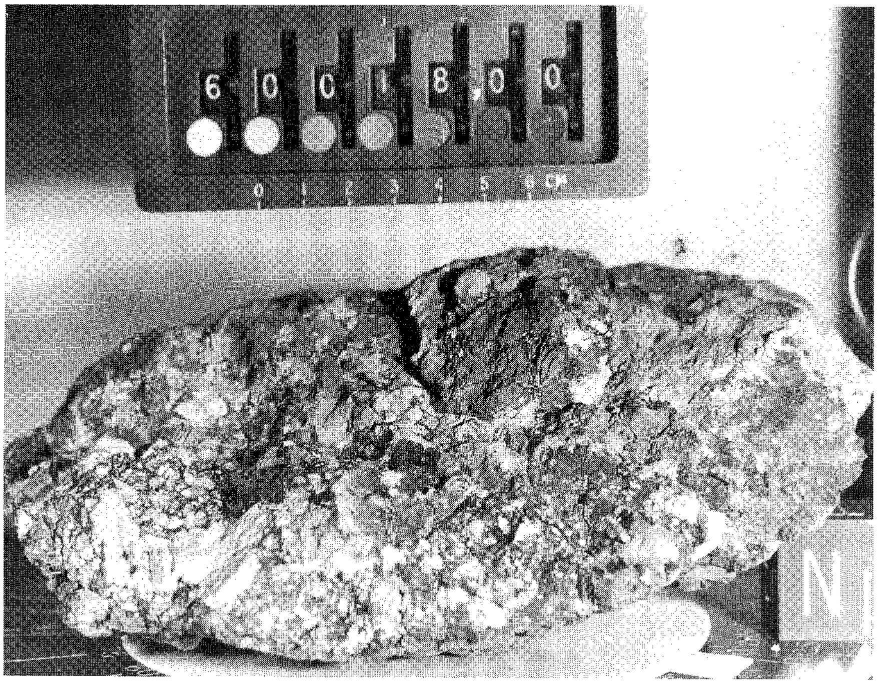

FIGURE 3.5 This conglomeration of rock was formed by bits of other rocks that were broken into shards when hit by meteorites and then glued together as further impacts piled up rocky debris on top of it. It is an example of a *breccia*.

the bottoms of these basins. This molten rock froze to make the dark mare areas of the Moon. We can tell that these areas are younger than the surrounding highlands because the lava lies on top of the heavily cratered material (which sometimes pokes up through the mare, as seen in the photographs) and because the mare regions are not nearly as heavily cratered as the highlands.

Even submicroscopic impacts can have an effect on the surface of the Moon. On airless bodies with no magnetic fields, high-energy particles from the solar wind or local planetary magnetospheres can impact directly onto the surface materials, a process called **sputtering**. These particles are usually protons, electrons, or the nuclei of atoms such as sodium or sulfur that are travelling at extremely high velocities.

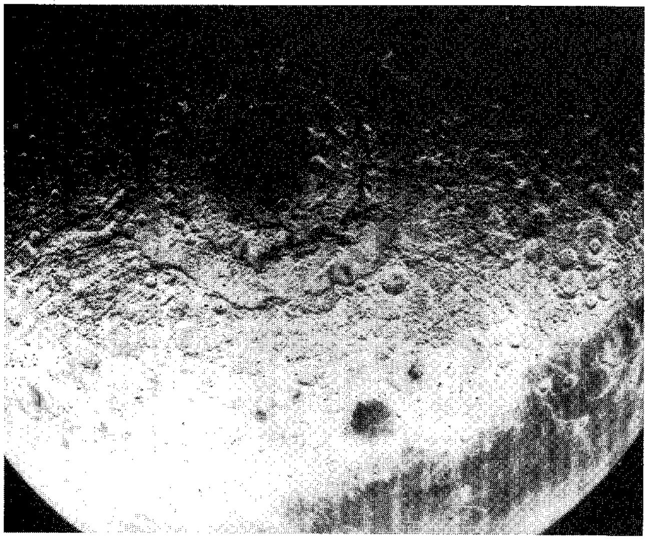

FIGURE 3.6 Mare Orientale. This **multi-ring basin** on the eastern side of the Moon was formed by the impact of a large asteroid-sized meteoroid early in the Moon's history.

Upon impact a number of different reactions can occur. The particles, although tiny, can break up chemical bonds as they plow into a rock. Cosmic rays, the most energetic (but rarest) of these particles, leave trails in the upper few millimeters of each rock or meteorite they hit, providing a convenient way of dating just how long that particular rock has been exposed to space. (The more trails, the longer the rock has been sitting on the surface of its parent planet.)

The surface of the Moon tends to get darkened, in a way not completely understood, as it sits exposed to the solar wind. A combination of mechanical effects and actual chemical reactions between the rocks and the impacting hydrogen nuclei may be responsible.

The Moon's Surface: Chemical Aspects

The energy of the impacts early in the formation of the Moon would have been sufficient to melt a considerable fraction of the Moon's surface. The chemistry of the Moon rocks reflects the outcome of such global melting.

Samples of highlands rocks returned by the Apollo astronauts showed several intriguing features. First, pristine highlands rocks are extremely rare; most are breccias. The ages of these rocks were difficult to measure because they had been so heavily battered by impacts, but all techniques indicated that they are at least 4 billion years old. But, finally, the very nature of the types of rocks they turned out to be was very surprising.

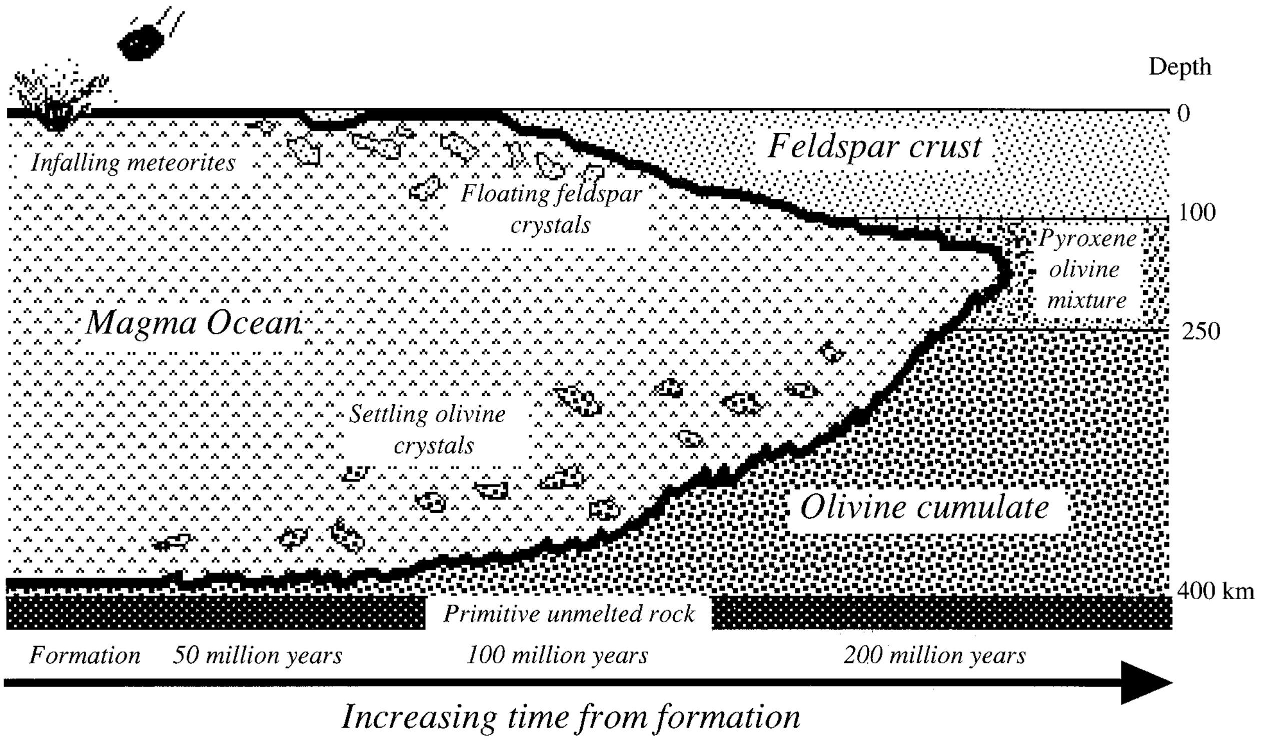

FIGURE 3.7 The Magma Ocean. Meteorites hit the early Moon with enough energy to melt the rocks, down to a depth of 400 km or so. In this ocean of magma, floating "rockbergs" rich in feldspar (lower density than the magma) gather on the surface to form an anorthosite crust, while higher-density crystals of olivine sink to the bottom. A mixture of different crystals, including all the incompatible elements, eventually freeze in a layer between the olivine and the anorthosite.

Igneous rocks made by lava flowing onto a planet's surface and then freezing have a specific chemical composition and mineralogic texture; basalts are rocks of this sort. Indeed, the rocks from the dark mare regions are classic examples of such basalts. But the highland rocks, for the most part, are not; rather, they belong to a class of rock called a **gabbro**. These are more typical of rocks crystallized slowly, deep inside a planet. So what are they doing on the surface of the Moon? Gabbros on Earth are brought to the surface with the mountain-building processes associated with plate tectonics, but there's no evidence for plate tectonics anywhere on the Moon.

One clue comes from the composition of these gabbros. They are rich in **plagioclase**, more specifically the mineral anorthite. This mineral is of a lower density than most other minerals, and indeed can be lower in density than the molten lava from which it was formed, if that melt also includes melted pyroxene or olivine. Thus, in a lava that was slowly cooling, the first mineral to crystallize would be plagioclase. Once formed, these crystals could tend to float upwards to the top of the lava, forming **anorthosites**, rocks made up almost completely of plagioclase.

Apparently the highlands rocks formed out of a large region of molten lunar material. How large might this region have been? The seismic detectors on the Moon indicated that the lunar crust, made up of gabbroic rocks, is 60 km to 120 km thick. Another sharp boundary in composition may exist about 400 km deep into the Moon. So one possibility is that the top 400 km of the Moon was completely molten at one time and the anorthosites crystallized out of this **magma ocean** to form the crust of the Moon (Figure 3.7).

This magma ocean was most likely melted by the energy of the large numbers of impacting rocks that hit the Moon as it was forming. It was probably completely frozen within 200 million years after the Moon was formed. As time went on and radioactive isotopes decaying in the lunar material caused the interior to warm up, temperatures eventually reached the the melting point of rock and basaltic lava could be formed. These basalts erupted to the surface and flooded the lowest areas of the Moon, creating the mare plains and their associated mascons. Judging from the ages of the mare rocks, this occurred between 3.9 and 3.0 billion years ago.

TABLE 3.1

Spacecraft That Have Orbited or Landed on the Moon

Name	Launch Date	Type
Luna 1	01/02/59	Unmanned flyby
Luna 2	09/12/59	Impact
Luna 3	10/04/59	Unmanned flyby
Ranger 3	01/26/62	Unmanned flyby
Ranger 4	04/23/62	Impact
Ranger 5	10/18/62	Unmanned flyby
Luna 4	04/02/63	Unmanned flyby
Ranger 6	01/30/64	Impact
Ranger 7	07/28/64	Impact
Ranger 8	02/17/65	Impact
Ranger 9	03/21/65	Impact
Luna 5	05/09/65	Impact
Luna 6	06/08/65	Unmanned flyby
Zond 3	07/18/65	Unmanned flyby
Luna 7	10/04/65	Impact
Luna 8	12/03/65	Impact
Luna 9	01/31/66	Unmanned lander
Luna 10	03/31/66	Unmanned orbiter
Surveyor 1	05/30/66	Unmanned lander
Lunar Orbiter 1	08/10/66	Unmanned orbiter
Luna 11	08/24/66	Unmanned orbiter
Luna 12	10/22/66	Unmanned orbiter
Lunar Orbiter 2	11/06/66	Unmanned orbiter
Luna 13	12/21/66	Unmanned lander
Lunar Orbiter 3	02/05/67	Unmanned orbiter
Surveyor 3	04/17/67	Unmanned lander
Lunar Orbiter 4	05/04/67	Unmanned orbiter
Explorer 35	07/19/67	Unmanned orbiter
Lunar Orbiter 5	08/01/67	Unmanned orbiter
Surveyor 5	09/08/67	Unmanned lander
Surveyor 6	11/07/67	Unmanned lander
Surveyor 7	01/07/68	Unmanned lander
Luna 14	04/07/68	Unmanned orbiter
Zond 5	09/14/68	Unmanned flyby
Zond 6	11/10/68	Unmanned flyby
Apollo 8	12/21/68	Manned orbiter
Apollo 10	05/18/69	Manned orbiter
Luna 15	07/13/69	Impact
Apollo 11	07/16/69	Manned lander
Zond 7	08/08/69	Unmanned flyby
Apollo 12	11/14/69	Manned lander
Apollo 13	04/11/70	Manned flyby
Luna 16	09/12/70	Unmanned lander
Zond 8	10/20/70	Unmanned flyby
Luna 17	11/10/70	Unmanned lander
Apollo 14	01/31/71	Manned lander
Apollo 15	07/26/71	Manned lander
Luna 18	09/02/71	Impact
Luna 19	09/28/71	Unmanned orbiter
Luna 20	02/14/72	Unmanned lander
Apollo 16	04/16/72	Manned lander
Apollo 17	12/07/72	Manned lander
Luna 21	01/08/73	Unmanned lander
Luna 22	05/29/74	Unmanned orbiter
Luna 23	10/28/74	Unmanned lander
Luna 24	08/09/76	Unmanned lander
Galileo	10/18/89	Unmanned flyby
Hiten	01/24/90	Unmanned flyby

The Origin of the Moon

Humans always have wondered where the Moon came from. For a long time, we believed that the Moon was formed much as other planets were, out of a condensing nebula about the Sun. But there have always been some difficulties with this hypothesis. The Moon is *extremely* dry, much drier than Earth, which one might think would be most like it in composition, being so close. Even though it has been differentiated, if it has a metallic core it must be very small. The uncompressed density is wrong; it also should be much like Earth's, if they were both formed simply by condensation from the solar nebula.

A theory that is gaining in popularity is that a proto-Moon (perhaps originally more the size of the present Mars) and Earth underwent a giant collision with each other, after they were condensed but still very early in their history. This impact completely disrupted the proto-Moon, melting or even vaporizing much of it. Part of it was completely ejected from the system, but the rest formed a cloud of debris around Earth.

This collision had the effect of removing a lot of the volatile material originally present in the Moon, changing its density and chemical composition. The metallic material in the cores of both planets was combined into that of Earth. The material that had been in the mantle of the proto-Moon recondensed from the debris in orbit around Earth to form the present Moon.

SUMMARY

The Moon, a small, rocky, airless body orbiting Earth, is the closest, most visited, and best explored extraterrestrial body (Table 3.1). Its surface is covered with impact craters, scars left over from the accretion of the Moon 4.5 billion years ago. Continued impacts to this day tend to wear down the walls of these craters and churn up the soil to a depth of several tens of meters. These impacts also tend to mix up, and compress together, fragments of different rock to form the conglomerate type rock called breccia that is the most typical type of boulder seen on the Moon today.

Early in its history these impacts melted the Moon to a depth of at least 400 km. Out of this magma ocean, a crust roughly 100 km thick of anorthositic gabbros was formed. Roughly half a billion years later, internal heating caused basaltic lavas to form; this molten rock erupted onto the surface, filling the low-lying areas in the bottoms of the largest impact basins, forming the dark mare regions. These eruptions went on for a billion years. For the past 3 billion years, however, little more has happened than the occasional new crater being formed to change the surface of the Moon.

The center of mass of the Moon and most of the mare regions are in the hemisphere facing Earth. Many mare basins have mascons associated with them, gravity anomalies due to the extra mass of the dense mare basalts that fill the basins.

STUDY QUESTIONS

1. The Moon's surface is divided into two main types of terrain. What are the dark, flat areas called? What are the bright, rough areas called?
2. What is the most common feature seen on the Moon's surface?
3. For an astronaut standing on the Moon, it is hard to judge by sight just how far away features such as mountains are. There are several reasons for this difficulty. Name one.
4. One theory for the formation of the lunar crust proposes that it was frozen from a layer of molten rock 400 km thick. What is this molten layer commonly called?
5. The daytime temperature of the Moon is:
   (a) hot enough to boil water
   (b) hot enough to melt lead
   (c) not that different from Earth

3.2 MINERALS AND ROCKS

We have learned, by going there, that the surface of the Moon is covered with specific rock types. To understand better how the Moon was formed and how it produced the surface we see today, we have to know something about minerals and rocks.

There are three major issues we need to address. First, how are minerals and rocks made? Second, what can their composition tell us about the interior of the planet where they are found? And third, how can we discover how long ago the formation of these rocks took place?

Nucleosynthesis and Rocks

Nuclear fusion in the centers of stars combines protons and neutrons to make the nuclei of the elements. When a supernova explodes, the elements are scattered to space. Out of these elements, the planets are formed.

But places like the Moon and the other rocky planets are not simply chaotic blobs of chemical elements. Rather, the elements are in chemical compounds, and these compounds form themselves into ordered crystal structures, which we call **minerals** (Figure 3.8). Finally, in the processes that form a planet's surface, several different kinds of minerals may be formed together and mixed into a conglomerate we call a rock.

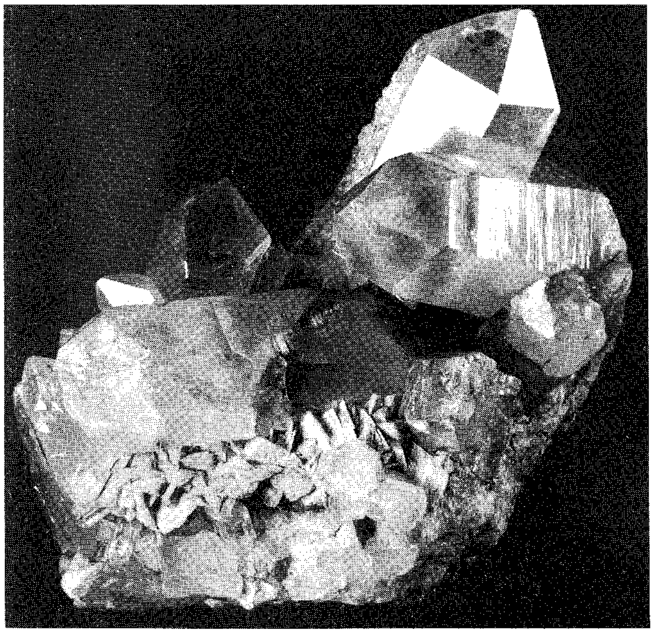

FIGURE 3.8 Minerals are chemical compounds with a regular crystal structure. The regular ordering of atoms in a crystal can create large, well-formed crystals, as seen in this sample of quartz (SiO_2).

Given the cosmic abundances of the elements, we should be able to predict which chemical compounds are the most plentiful. It turns out that virtually all the rocks in the solar system can be characterized as mixtures of just a few general types of minerals. Likewise, the most important types of rocks can be characterized by the way they are formed, which is limited to a few basic processes.

The Primary Rock-Forming Minerals

Of the 83 naturally occurring elements, only a handful are abundant enough to be important in most minerals. In Earth and most of the other planets, these elements react chem-

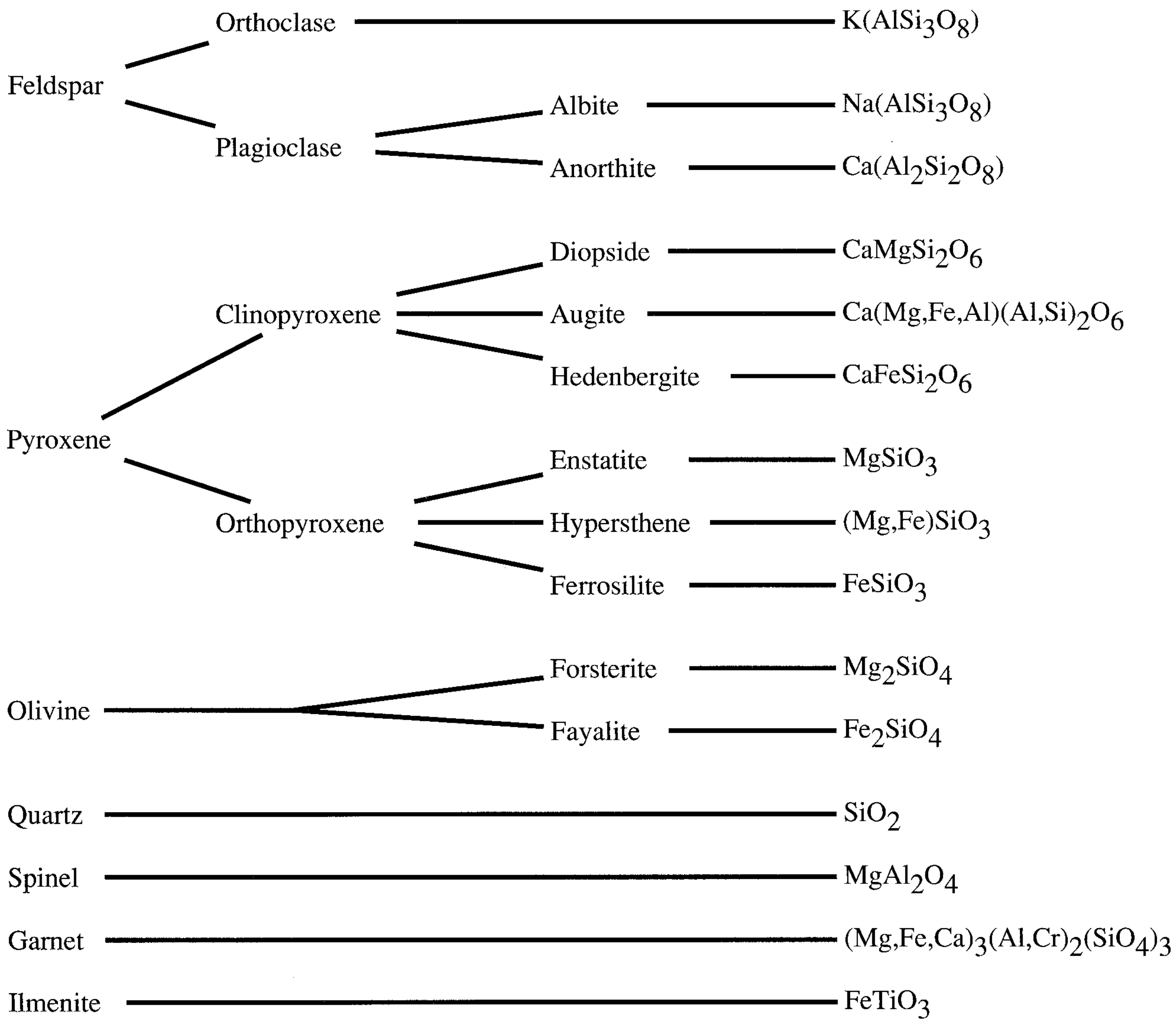

FIGURE 3.9 The major rock-forming minerals.

ically to form oxides. These oxides, in turn, combine to form compounds that are the basis of the minerals.

Mineral refers not only to a chemical formula but also to a crystal structure. Two minerals can be very different, even if they have the same chemical composition, if those elements are arranged in different positions. On the other hand, two minerals might be very similar to each other, even if they're made of differing chemical elements, if they share the same crystal structure. The important classes of minerals are shown in Figure 3.9. The three most abundant types of minerals on the surfaces of planets are olivine, pyroxene, and feldspar. They are all examples of **silicate** minerals. This means that they are built from a crystal structure of silicon and oxygen atoms, with the other elements fitting in wherever their size and electric charge allows (Figure 3.10). The elements magnesium (Mg) and iron (Fe) both occur with a +2 charge and have similar sizes, so they can substitute freely for each other in many crystals. That's why there's a range of olivines, from the iron-rich olivine (called **fayalite**) to the magnesium-rich olivine (**forsterite**). Likewise, calcium (Ca), sodium (Na), and potassium (K) all substitute for each other in pyroxenes and feldspars. But calcium has a +2 charge, while sodium and potassium are +1, so to balance out the charge, the abundance of aluminum (Al) and silicon (Si) has to be juggled. (Aluminum in minerals normally carries a charge of +3, while silicon carries +4.)

Olivine is a dense, usually olive-green mineral (hence its name). It is often found in the interiors of planets or in surface rocks made from a lava formed deep within the planet. It consists of silicon (Si), magnesium (Mg), and iron (Fe) with two atoms of Mg or Fe for each silicon

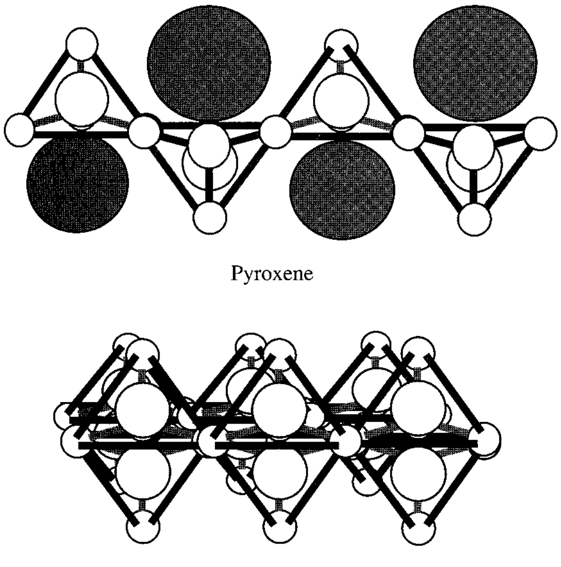

FIGURE 3.10 The positions of the atoms in a mineral determine its crystal structure; examples of pyroxene and olivine are shown.

atom. These are the three most abundant rock-forming elements, and so we expect olivine to be the most common mineral in planets. Rocks rich in olivine are of necessity rich in **ma**gnesium and iron (**F**e); these are often called **mafic** rocks.

Pyroxene is also generally greenish colored. Its chemical formula is quite similar to olivine, except that it has twice as much SiO_2, **silica**. Because the cosmic abundances of silicon and magnesium are roughly equal, a rock made with cosmic abundances of these elements but without iron oxide (FeO) would be mostly pyroxene instead of olivine. Pyroxene too, therefore, is a major constituent of planetary interiors, especially where iron is present only as metallic iron, and not iron oxide.

Feldspar is a general term given to the class of silicate minerals that are rich in aluminum and have an open, "framework" sort of crystal structure. Because of this open structure, there's room in feldspar crystals for larger-sized atoms, such as sodium (Na), calcium (Ca), and potassium (K). Many trace elements (any of 70 or so less-abundant elements that might be found in a rock) can also fit into this structure. Also, because the aluminum atoms are less massive than Fe or Mg, and because of the open structure of the crystal, feldspar minerals tend to be less dense than the other minerals and even less dense than most lava. As a result, when planets melt and refreeze, this lighter mineral tends to float to the crust of the planets. That is how we explained why the surface of the Moon is rich in feldspar.

The other types of minerals listed in Figure 3.9 can be important under certain circumstances. Deep inside planets high pressures force the crystals into densely packed structures such as **garnet** and **spinel**. Compared to Earth, the mare rocks from the Moon show a high abundance of titanium (which does not fit easily into the crystal structure of the other minerals); on the Moon, the titanium-bearing **ilmenite** is an important mineral.

The Formation of Rocks

Minerals are found combined together into rocks. Each rock type can have any number of minerals, usually two or three major minerals.

Traditionally, scientists have talked about three types of rock on Earth. These are **igneous** rocks, formed from molten lava that cools and crystallizes; **sedimentary** rocks, formed from the accumulation and conglomeration of weathered bits of other rocks; and **metamorphic** rocks, rocks that have been subjected to heat and pressure sufficient to alter their internal structure without melting them completely.

Not all these types are found on every planet. But it is clear that both sedimentary and metamorphic rocks are just altered forms of some other rock; and, because it appears that every terrestrial planet was melted at some time in its history, it is reasonable to assume that all rocky material on these planets started out as igneous rock. Thus it is the formation of igneous rocks that we will explore here.

The different types of igneous rock can be characterized by the process that made the rock. **Basalt**, an **extrusive** or **volcanic** rock, is made by lavas erupting onto the surface of a planet, then cooling; **granite**, **gabbro**, and **peridotite**, are formed from the slow freezing out of crystals from a melt below the surface of a planet (to form **intrusive**, or **plutonic**, rocks).

Basalts tend to have many tiny crystals, because the lava froze too quickly to organize itself (Figure 3.11, upper). One typically needs a microscope to easily see the individual crystals in such rocks. By contrast, the minerals in plutonic rocks had time to form large crystals, several millimeters across at least, as the lava slowly cooled (Figure 3.11, lower). Thus we call basalts "fine grained" while the plutonic rocks are "coarse grained."

Molten lava, from which igneous rocks are derived, is a part of a planet that has become hot enough to melt. A pure substance has a specific temperature at which it melts, called the melting point. But rocks are not made up of pure substances; they are mixtures of different minerals. A mixture of two (or more) substances often starts to melt at a temperature much lower than the melting point of either substance (Figure 3.12). This temperature is called the **eutectic temperature**.

For example, the mineral pyroxene melts at 1650 K and the mineral plagioclase melts at 1825 K, but if the two are mixed together, a molten lava starts to form when the

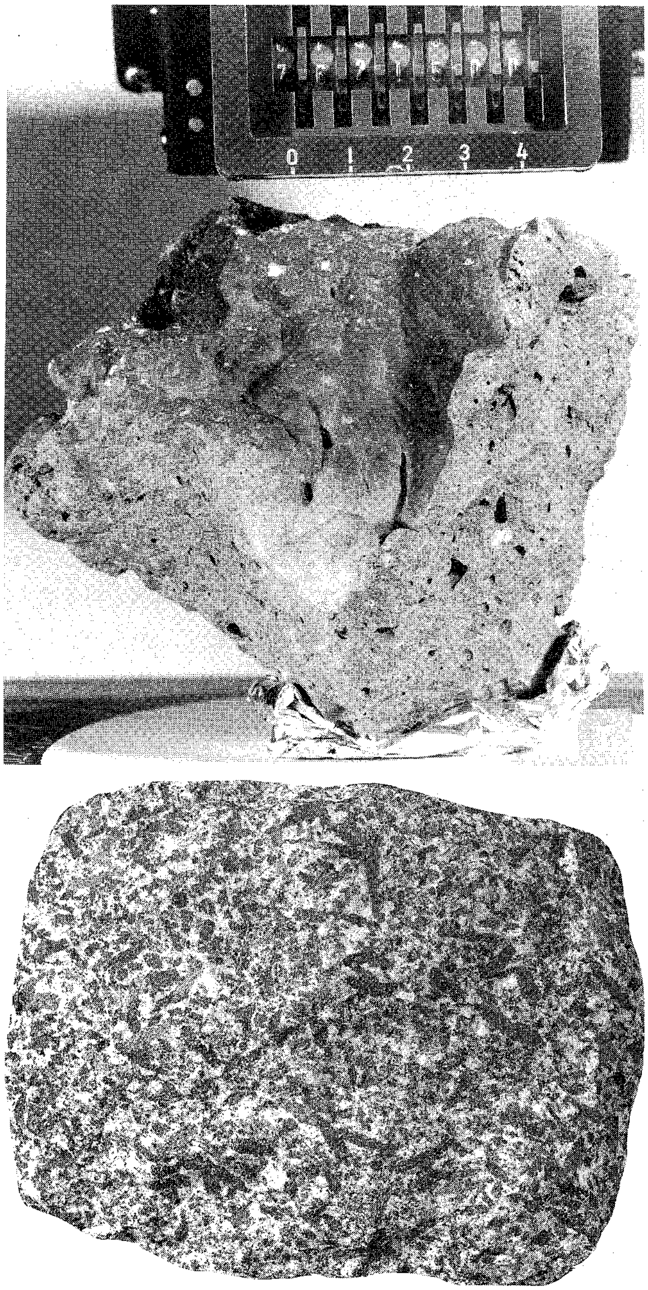

FIGURE 3.11 A lunar basalt (upper), and a terrestrial gabbro (lower).

temperature reaches only 1550 K. (The precise temperatures depend on the compositions of the minerals.)

The first lava to form when a rock starts to melt has a very specific composition, called the **eutectic composition**. Any mixture of these two minerals starts to melt with the same eutectic composition. You get the same composition whether you start out with almost all plagioclase and just a little pyroxene, or the other way around. In fact, substantial amounts of other minerals, such as olivine or metallic iron, can be present without changing the composition of the eutectic melt.

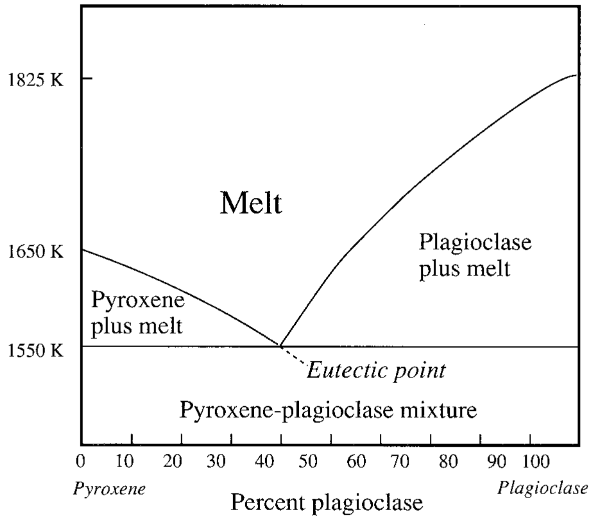

FIGURE 3.12 A **phase diagram** for the pyroxene–plagioclase system.

The composition of this eutectic melt is generally close to 60% pyroxene and 40% plagioclase. Although it does not depend on the proportions of plagioclase or pyroxene present, the precise composition does depend on the pressure. High pressures, such as might be seen deep inside a large planet, can cause a shift in the eutectic composition.

By precisely measuring the proportions of each mineral in a basalt, it is sometimes possible to determine how deep in the planet the lava was when it melted. This technique is called **geobarometry**. Naturally, nature is never quite this simple. Other complications, such as the iron content of the pyroxene, can also change the proportions of the minerals in a basalt.

The place where the lava is formed is called the **source region** of the lava. Using the techniques of geobarometry, we can infer that the basalts on the Moon came from several hundred kilometers below the surface of the Moon, while very similar-looking basalts seen in certain meteorites must have been formed inside much smaller planets where the pressure was very low.

Trace Elements in Igneous Rocks

As we saw above, only a few elements are abundant enough to be important constituents of planets. Common minerals are formed from the arrangement of silicon, iron, magnesium, aluminum, calcium, and oxygen atoms into only a few configurations, a few crystal structures. Some other atoms, such as potassium and titanium, are abundant enough to make their own minor minerals. But where do rare elements such as hafnium or praseodymium fit in?

The answer is, quite literally, they fit in wherever they can. Most of these rare elements are more massive and physically larger than the major elements we listed above.

FURTHER INFORMATION...

Ternary Diagrams

Understanding how mixtures of minerals behave is the key to understanding rocks. Minerals with similar structures, such as iron-rich olivine (fayalite) and magnesium-rich olivine (forsterite), can form a **solid solution** because the two minerals are intimately mixed and behave as if they were dissolved in each other. Other very different minerals, such as pyroxene and plagioclase, can form eutectic melts when heated. We saw (in Figure 3.12) how to represent a rock made of two different minerals. But how can we examine a system made of three minerals?

The answer is a three-sided figure called a **ternary phase diagram**. Assume that our rock is made of three minerals: quartz, olivine, and the calcium-rich plagioclase, anorthite (see Figure 3.13). The less anorthite there is in our rock, the farther away our composition is from the anorthite corner. If no anorthite is present at all, our composition lies on the line between olivine and quartz, and we are back to a two-mineral situation. Notice that the mineral pyroxene has a composition equivalent to one part olivine, one part quartz; thus it, too, can appear on our ternary diagram.

Look closely at the ternary phase diagram, Figure 3.13, for minerals under lunar conditions. The "v" markings along each axis mark off the composition scale, in intervals of 10%; each side of the "v" lines up with a "v" side on another axis. Thus, for example, the lunar basalt 60315 has a composition of about 45% olivine, 30% quartz, and 25% anorthite.

A melt of this composition, as it cooled, would first form olivine crystals (hence the point lies in a region marked "olivine"). The composition of the melt, once the olivine is removed, moves towards the plagioclase region. When the composition hits the boundary between the regions, both plagioclase and olivine crystallize together; the melt composition follows the arrow until it reaches the junction between the pyroxene, olivine, and plagioclase regions.

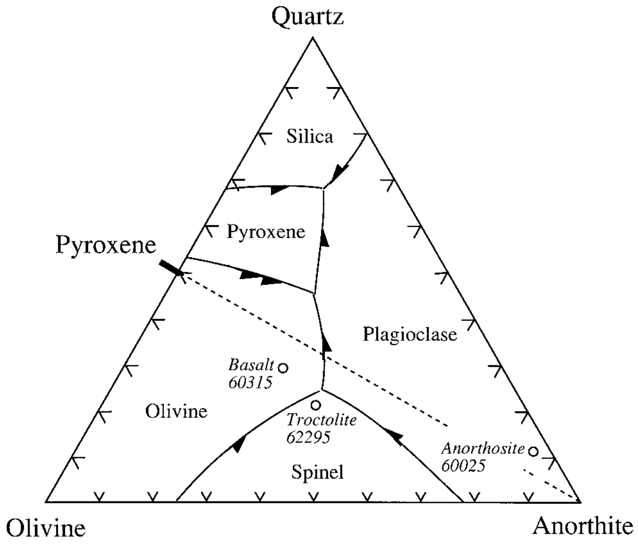

FIGURE 3.13 **ternary phase diagram** for the olivine–anorthite–quartz system.

Melts whose initial composition lie below the dotted line, like our rock, will not evolve any further. However, more quartz-rich melts (like the anorthosite 60025) could continue to evolve up to the junction between the pyroxene, plagioclase, and silica regions. The boundary between olivine and pyroxene has two arrows, signifying that chemical reactions between the melt and the quartz take place as the melt evolves along this line.

Rocks that melt evolve in the opposite way; the first melt has a composition matching one of these two junctions; which junction depends on the overall composition of the rock. In effect, they mark the eutectic melting points for this system.

For them to fit into a crystal, the whole crystal has to be distorted in shape just a bit to make room for them.

Thus it is not surprising that once a crystal starts to melt, among the first bonds to be broken are those distorted bonds holding the large, incompatible elements into the crystal framework. The result is that molten lava tends to be relatively enriched in these trace elements. Thus basalts, which are frozen lavas, are rich in these trace elements, while gabbros, which are formed of crystals that slowly settled out of molten rock, leave those elements behind in the melt and thus are poor in them.

That is the general pattern. Getting to specific cases, we find that each mineral has its own affinity, or disaffinity, for any given trace element. The **rare-earth** elements, the series from lanthanum (La) to lutetium (Lu), are a good case in point. (These are the elements that are usually found off by themselves in the periodic table.) These elements are almost chemically indistinguishable from one another. Experiments have shown, however, that plagioclase rejects lighter rare-earth elements, like lanthanum and cerium (Ce), more than the heavier rare earths, like ytterbium (Yb) and lutetium. But the opposite is true for pyroxene, which rejects the heavier ones more. Olivine and metal tend to reject all rare earths more uniformly. These trends are shown in Figure 3.14.

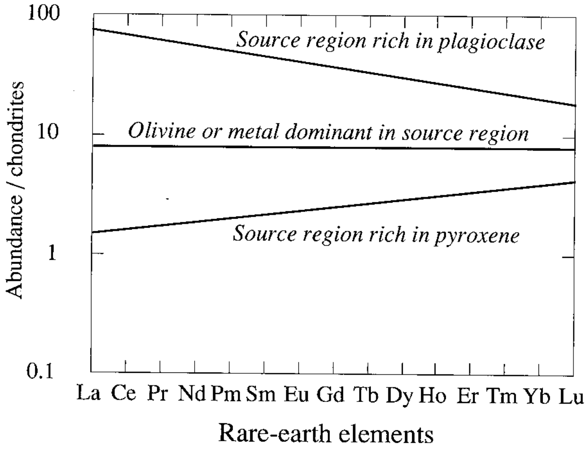

FIGURE 3.14 Comparison between three simple types of rare-earth patterns. All values are normalized to chondrite composition.

Thus we can take a basalt and analyze the rare-earth content. The abundance pattern of the high-titanium (Hi-Ti) basalt in Figure 3.15 is mostly flat but slightly richer in lutetium than in lanthanum. The flatness suggests that olivine or metal was the dominant material in the source region, even though neither is present in the basalt itself, and the slight tilt towards lutetium suggests that the source region was richer in pyroxene than in plagioclase. The dip at europium is also significant, as we shall see.

The troctolite and anorthosite rocks shown in Figure 3.15 are not basalts, and so modelling their origin is not quite so straightforward. But it is still possible using their rare-earth element abundances and arguments similar to those described for basalts to learn much about the magma from which these rocks crystallized. For instance, the large superabundance of europium is characteristic of plagioclase (an important constituent of anorthosite in highlands rocks), whose crystal structure can readily accept this particular rare-earth element. The negative europium anomaly in the Hi-Ti basalt probably occurred because this rock was melted out of a region that crystallized after most of the plagioclase in the original magma ocean had formed the highlands anorthosite. On the other hand, apparently there was very little plagioclase crystallized when the troctolite was formed, because it does not have a large europium anomaly.

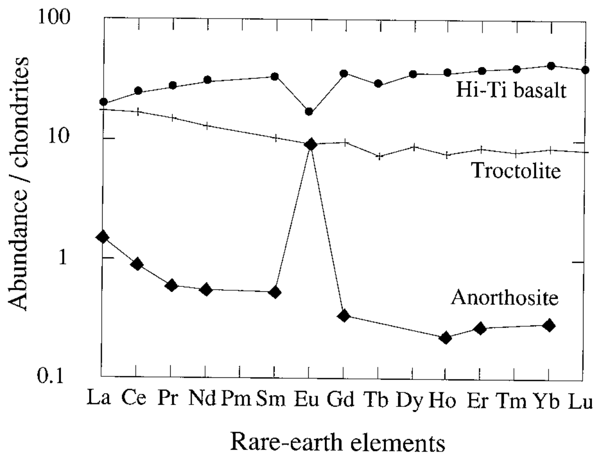

FIGURE 3.15 Rare-earth element abundances for three lunar rocks. Trace element abundances give us information on the environment where a rock was formed.

Dating Rocks

Is it possible to measure how old a rock is? First we must agree as to what we mean by a rock's age. Generally, the age of a rock is defined as the time that a rock has existed without being chemically disturbed. But there are different degrees of disturbance, so there may well be different ages for the same rock.

For instance, a lava on the Moon may have frozen into rock 4 billion years ago; but that rock may have been hit by an incoming meteorite some 500 million years later, which would significantly rearrange some of the atoms inside that rock. Such a rock could show two different ages, depending on how we measured the age.

The usual way to date a rock is to look at some radioactive element and measure how much of its **daughter** (decay product) is present in the rock. An example is an isotope of rubidium, ^{87}Rb, which decays into strontium, ^{87}Sr. The problem is how to sort out the daughter ^{87}Sr from the ^{87}Sr that was present in the rock originally ($^{87}\mathrm{Sr}_{orig}$).

The amount of ^{87}Sr present today ($^{87}\mathrm{Sr}_{now}$) depends on three unknown quantities: the amount of ^{87}Sr originally in the rock, the amount of original Rb ($^{87}\mathrm{Rb}_{orig}$), and the time, t. (The age also depends on the rate at which Rb decays in Sr. However, this can be measured in the laboratory.) The amount of original Rb can be found by measuring the amount of Rb present today ($^{87}\mathrm{Rb}_{now}$) and adding to it the excess Sr that's been made from the original time until now, the difference between $^{87}\mathrm{Sr}_{now}$ and $^{87}\mathrm{Sr}_{orig}$. We can also measure $^{87}\mathrm{Sr}_{now}$ today. That still leaves us with

$^{87}\mathrm{Sr}_{orig}$ and t to find. In terms a mathematician would use, we have two unknowns; to solve for them, we must find two equations.

To do this, we look at two different minerals in the same rock that were presumably made at the same time, but which started off with different Sr and Rb contents. By comparing the two, we can solve for both the different initial Sr contents and for the time. (See Figure 3.16. The steeper the slope of the final abundance line, the older the rock is, while the point where this line intercepts the ^{87}Sr axis denotes the original abundance of ^{87}Sr.)

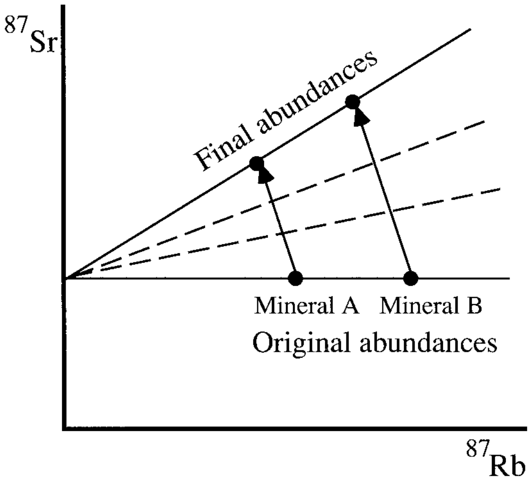

FIGURE 3.16 Dating a rock by looking at two minerals with different initial amounts of radioactive Rb. The more Rb the mineral started with, the more Sr it will have later on. The older the rock is, the steeper the slope of the line is.

Another atomic clock is provided by an isotope of potassium, ^{40}K, which decays into argon, ^{40}Ar. We can use a similar technique to Rb–Sr dating to find an age for the rock, but sometimes the two clocks arrive at different results. This happens because the daughter isotope of potassium decay, ^{40}Ar, is a gas. If the rock is hit very hard—for example, if a Moon rock were hit by a meteorite forming a crater on the surface of the Moon—then some of the ^{40}Ar could get shaken loose and escape. In this way, the K–Ar clock would be set back to zero.

Argon is a noble gas and does not chemically combine with any atom in the mineral, but merely sits in the place where its parent ^{40}K used to sit (Figure 3.17). ^{87}Sr, by contrast, does make a chemical bond with the same surrounding atoms that had been bound to ^{87}Rb, and so it won't be dislodged by a shock. Thus the same rock can tell us both the age when it was frozen from a lava and also the last time when it was struck by a meteorite.

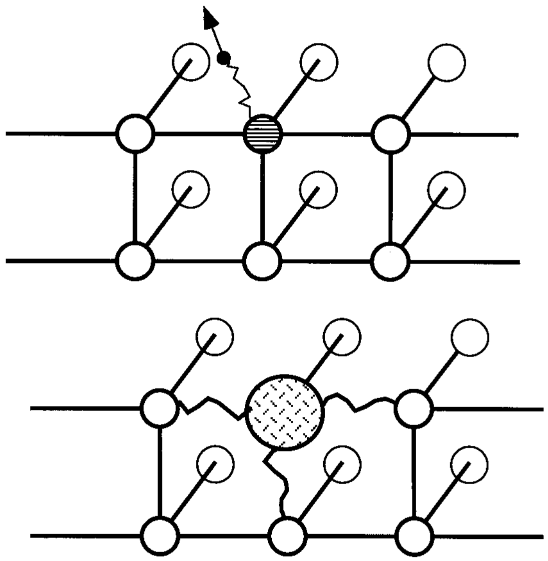

FIGURE 3.17 A radioactive element decaying while sitting in a crystal. By emitting a subatomic particle, the striped element transforms itself into a new element, with a different size and charge. The new element does not necessarily fit well into the old element's spot in the crystal structure; when ^{40}K decays into ^{40}Ar, for instance, the ^{40}Ar is not chemically bound at all to its site in the crystal.

SUMMARY

Rocks are collections of minerals. Minerals are specific chemical compounds with specific crystal structures. On Earth and most other planets, the large abundance of oxygen ensures that most minerals are compounds of oxygen and other elements; each mineral can be considered as a collection of the oxides of various elements. Thus, by understanding the chemistry of the mineral classes olivine (two parts magnesium or iron oxide, one part silica), pyroxene (one part magnesium or iron oxide, with an occasional calcium oxide, and one part silica), and feldspar (one part calcium or sodium oxide, one part aluminum oxide, and one part silica), one has mastered the most overwhelmingly abundant of the minerals found in planetary rocks.

Likewise, the most common types of rock themselves are basalts (pyroxene and feldspar, made from lavas erupted on the surface of a planet), gabbro (pyroxene and feldspar, slowly cooled and crystallized inside a planet), peridotites (pyroxene and olivine, cooled inside a planet), and granite

FURTHER INFORMATION...

The Math Behind Dating Rocks

In this chapter we have discussed the physical ideas behind how the measured abundances of radioactive trace elements in a rock can be used to determine its age. The mathematics behind this theory is a straightforward application of Newton's calculus.

Assume that we have an isotope with an abundance of N, which decays at a rate λ decays per second. By the definition of a derivative, we can write the rate at which the abundance changes as

$$\frac{dN}{dt} = -N\lambda \tag{3.1}$$

(the minus sign indicating that the abundance is decreasing with time). This can be integrated to show that after a time t,

$$N = N_{orig} e^{-\lambda t} \tag{3.2}$$

where N_{orig} is the original abundance of the isotope. The age of the rock, t, is ultimately what we're attempting to find, but to use this equation we also need to know N_{orig}. Furthermore, in a practical sense it is much easier to measure ratios of isotope abundances than to count the exact number of atoms of a given isotope in a particular rock.

In the Rb/Sr system, all abundances are measured relative to the stable isotope ^{86}Sr. We measure the abundance of ^{87}Rb/^{86}Sr; call that N. As the amount of ^{87}Rb decreases, the amount of ^{87}Sr increases, and we can also measure the amount of the daughter isotope, ^{87}Sr/^{86}Sr. Call this ratio D.

The amount of D present in a mineral depends on the initial abundance, D_{orig}, the amount of N present originally, and the amount of time that the rubidium has had to decay into strontium. Initially, when $t = 0$, we know that $D = D_{orig}$; after an infinite amount of time, all the initial N_{orig} is converted into D and so $D_{max} = D_{orig} + N_{orig}$. In between, the N that is present at time t represents material that is not yet converted into D, and so we can write

$$\begin{aligned} D &= D_{orig} + N_{orig}\left(1 - e^{-\lambda t}\right) \\ &= D_{orig} + N\left(e^{+\lambda t} - 1\right) \end{aligned} \tag{3.3}$$

Consider two minerals, A and B. If they formed at the same time, when strontium isotopes were well mixed, then even though the amount of strontium in one mineral may be very different from that in the other, the ratio of strontium isotopes, D, starts out exactly the same in both minerals. Hence D_{orig} of mineral A equals D_{orig} of mineral B. After formation, both minerals evolved for a time t (again, t is the same for both minerals). Thus we can write

$$D_A = D_{orig} + N_A\left(e^{+\lambda t} - 1\right) \tag{3.4}$$

$$D_B = D_{orig} + N_B\left(e^{+\lambda t} - 1\right) \tag{3.5}$$

Solve the second equation for D_{orig} and substitute into the first equation:

$$\begin{aligned} D_A = &\left[D_B - (N_B)\left(e^{+\lambda t} - 1\right)\right] \\ &+ (N_A)\left(e^{+\lambda t} - 1\right) \end{aligned} \tag{3.6}$$

This can be solved for the age of the rock, t:

$$t = \frac{1}{\lambda} \ln\left(\frac{D_A - D_B}{N_A - N_B} + 1\right) \tag{3.7}$$

(mostly feldspar and quartz, pure silica, cooled beneath the surface of a planet). By studying the trace element abundances of a rock, one can deduce the conditions under which its parent lava was formed. Likewise, if the trace element is the daughter of a radioactive decay process, one can estimate the time at which the lava froze and thus find the age of the rock.

STUDY QUESTIONS

1. For each term, state whether the material named is an element, a mineral, or a rock:

titanium	plagioclase	granite
gabbro	pyroxene	basalt
silicon	olivine	peridotite

FURTHER INFORMATION...

Modelling the Origin of a Rock

One can mathematically model, in detail, the evolution of the trace elements in a geochemically evolving system as melting goes on inside the planet. From such a model, one can determine precisely how much plagioclase, pyroxene, or other minerals were present in the original source region (which may be hundreds of kilometers below the surface and thus otherwise impossible to sample directly). Furthermore, from these models one can determine how much the rock had melted when the basaltic lava was erupted, which tells us how much heat was present at the time and how thorough was the evolution of the interior.

We noted that every mineral has a certain affinity for each trace element. When a mineral begins to melt, the ratio of the concentration of a trace element in the molten, liquid phase to that in the solid phase is constant; every time we start to melt that mineral we get the same ratio of trace element concentrations. We call this constant the **partition coefficient**, and it is often given the symbol $D_{\ell/s}$.

But the ratio of the concentration of trace element in the liquid phase to the total initial concentration changes continually as the melting proceeds; so does the solid phase/initial concentration ratio. When melting just begins, most of the trace element is in the solid; when melting is almost complete, most of it is in the liquid. It is obvious that if the system is closed, that is, if the lava hasn't started to erupt yet but is still sitting with the unmelted portions of the rock, then the total amount of trace element is not changing. The total amount of trace element present is equal to however much is in the lava plus however much is in the rock. Mathematically, we can write

$$c_o = c_\ell F + c_s (1 - F) \tag{3.8}$$

where c_o is the original concentration of the trace element, c_ℓ is the concentration of that trace element in the lava, and c_s is its concentration in the solid rock; F is the fraction of the original rock that has melted into lava. But our definition of the partition coefficient, $D_{\ell/s}$, is the ratio of the liquid to solid concentrations, so $D_{\ell/s} = c_\ell / c_s$. So we can write the equation above as

$$c_o = c_\ell \left[F + D_{\ell/s} (1 - F)\right] \tag{3.9}$$

and solve for the ratio

$$\frac{c_\ell}{c_o} = \frac{1}{\left[F + D_{\ell/s} (1 - F)\right]} \tag{3.10}$$

In a typical lunar rock model, c_ℓ may be the observed concentration of the different rare-earth elements measured in a certain lunar sample, c_o is the abundance of that trace element in the source region, and $D_{\ell/s}$ is the average value of the $D_{\ell/s}$ for the various minerals suspected of being in the source region. We measure c_ℓ in the lab; if we have a good guess for the composition of the source region, we can solve for amount of fractionation, F.

Alternatively, if we can estimate the original rare-earth element abundances, c_o (for instance, by assuming that the different trace elements originally occurred in the same relative proportions as are seen in our table of cosmic abundances), and the amount of fractionation, then we can solve for $D_{\ell/s}$ and determine what combination of minerals would provide an average $D_{\ell/s}$ value consistent with the trace element abundances we observe in the rock.

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2. We can measure precisely the amounts of rubidium and strontium isotopes in the minerals of a rock. What can this information tell us about the rock?
3. Argon is a gas. Do we ever find it in rocks?
4. True or False: If two rocks have identical amounts of the same elements present, then we can say that they are made of the same minerals.
5. The following is a list of qualities about an igneous rock. Name the one that cannot be determined from studying the geochemistry of that rock:
   - **(a)** age of the rock
   - **(b)** temperature at which it was formed
   - **(c)** pressure at which it was formed
   - **(d)** other minerals (not in the rock now) present when it was formed
   - **(e)** the depth inside the planet where it was formed
   - **(f)** none of the above