

6

The evolution of atmospheres

We have shown how the behavior of an atmosphere depends on its composition and its mass. It is now time to consider the processes governing the composition and mass of an atmosphere and to describe how the atmospheres of the solar system could have developed the properties that they have today. This aspect of atmospheric science has many unanswered questions, and it is possible that the theories we shall describe will in the future turn out to be wrong. However, we cannot understand the workings of the world we live in without understanding its history. Some account of this history is therefore necessary, even if the account is highly speculative.

We must first realise that our atmosphere is not an isolated and static layer of gas. Various chemical constituents are continually being added to the atmosphere while other constituents are continually being removed. One example is carbon dioxide.

The Cycles of Carbon Dioxide

The ocean is the largest source and also the largest sink for carbon dioxide in the Earth's atmosphere. Atmospheric carbon dioxide dissolves in the ocean and is then converted to bicarbonate ion. At the same time, oceanic bicarbonate ion is converted by the reverse process into carbon dioxide, which subsequently escapes to the atmosphere. This cycle, which converts atmospheric carbon dioxide to oceanic bicarbonate ion and back again, is shown on the right in Fig. 6-1. Values are given for the rate at which the gas dissolves and for the rate at which the gas comes out of solution. The two processes occur about equally fast, which indicates that

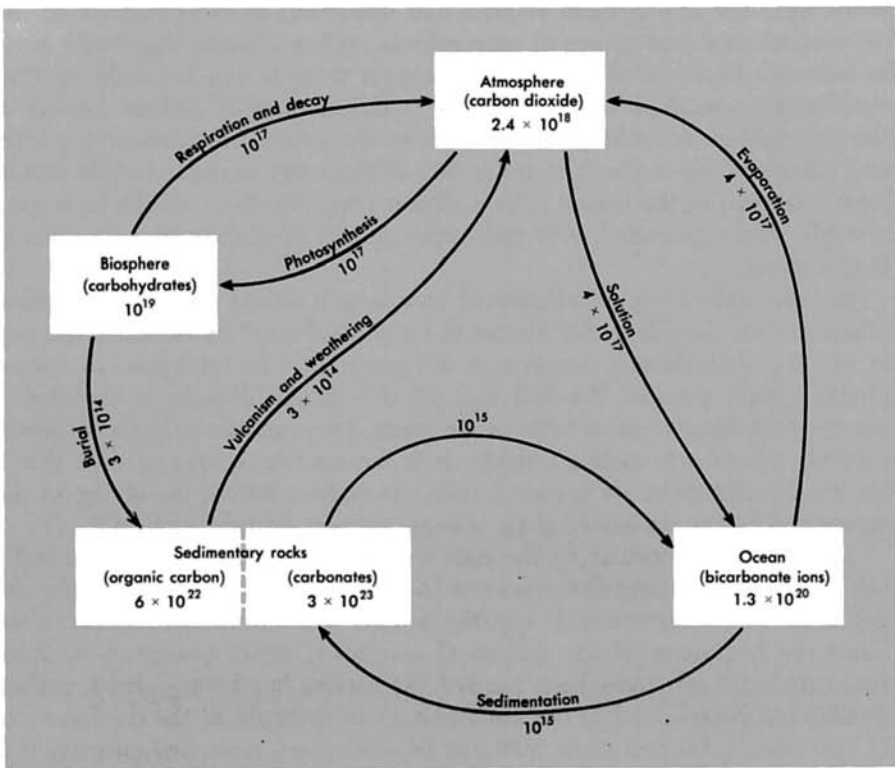


FIGURE 6-1 The cycles of carbon and carbon dioxide. Boxes denote reservoirs of carbon; the contents of the reservoirs are expressed in gm of CO_2 . The arrows denote the transfer of material between reservoirs; the rates of transfer are given in gm of CO_2 per year.

an equilibrium exists between the pressure of carbon dioxide in the atmosphere and the concentration of carbonate ion in the sea. It is principally this equilibrium that governs the amount of carbon dioxide in the atmosphere.

Suppose we were to add 2.4×10^{18} gm of carbon dioxide to the atmosphere by artificial means. The data in Fig. 6-1 show that this amount is equal to the total mass of carbon dioxide in the atmosphere today. Initially, therefore, the amount of carbon dioxide in the atmosphere would be doubled by our addition, but this would not last for long. The rate at which carbon dioxide dissolves in the ocean is proportional to the pressure of carbon dioxide in the atmosphere. This rate would now be doubled, whereas the rate of return of the gas to the atmosphere would not be changed. Since the rate of loss of carbon dioxide to the oceans would exceed the rate at which the oceans release carbon dioxide to the atmosphere, the amount of atmospheric carbon dioxide would decrease. This would go on for a few years, until the atmospheric amount had decreased to the point where the rates of removal and return of atmospheric carbon dioxide were once more in balance. From considerations of reaction rates it can be deduced that equilibrium would be reestablished with the amount of carbon dioxide in the atmosphere increased by only 10% from the value it had before the original addition was made. The other 90% of the extra carbon dioxide would have dissolved in the ocean. This addition (see Fig. 6-1) would be imperceptibly small compared with the large amount of carbon dioxide already in the ocean.

We can make a rough estimate of how long it would take for the atmospheric carbon dioxide to fall almost to its original level by dividing the rate at which carbon dioxide dissolves in the ocean into the total mass of atmospheric carbon dioxide. We find that all the carbon dioxide in the atmosphere could dissolve in as little as six years. Six years, then, is the turnover time of atmospheric carbon dioxide. It is the average length of time that a carbon dioxide molecule spends in the atmosphere before dissolving in the ocean and being replaced by an oceanic carbon dioxide molecule.

An experiment similar to the one we have just described has actually been performed during the course of the last hundred years or so. The experiment was unintentional, but the results are informative nonetheless. Since the beginning of the industrial revolution great quantities of fossil fuel (coal and oil) have been burned. As carbon has been burned, carbon dioxide has been added to the atmosphere. An estimate of the total amount of coal and oil burned since 1860 can be made, and from this estimate it is possible to deduce that 5×10^{17} gm of carbon dioxide have been added to the atmosphere in the same period. Actual measurements of the carbon dioxide content of the atmosphere show a smaller increase, however. A part of the added carbon dioxide has dissolved in the ocean.

The burning of fossil fuel is an element of another cycle that affects at-

mospheric carbon dioxide; this is the cycle of photosynthesis and respiration, which is shown on the left in Fig. 6-1. Plants extract carbon dioxide from the atmosphere or ocean and, with the addition of water, manufacture carbohydrates and release oxygen. The reverse reaction occurs as well. Plants and also animals consume atmospheric oxygen and burn their carbohydrates, producing carbon dioxide and water. This process, known as *respiration*, furnishes the organism with energy. Another process that restores to the atmosphere the carbon that has been stored in animals and plants is decay, which occurs after the death of the organism.

The rate at which respiration and decay occur is very nearly equal to the rate of photosynthesis, as the values in Fig. 6-1 show. Nevertheless, there is, on average, a small imbalance that is important. Some of the organic debris is washed out to sea and is buried by sand and mud before it has decayed. As a result, the rate of photosynthesis is slightly faster than the rate of respiration and decay, and there is a continual addition of carbon of organic origin to the sediments at the bottom of the sea. In time these sediments are converted to rocks, and the carbon is incorporated therein, but the process does not stop there. The layer of sedimentary rocks on Earth is not getting thicker all the time. Instead, the rocks are lifted up above the surface of the sea, where they are subjected to weathering and erosion. The carbon in the rocks is oxidized in the process and returned to the atmosphere as carbon dioxide. In this respect, the burning of fossil fuels may be thought of as a greatly accelerated form of weathering. At the present time, the burning of fossil fuel is producing carbon dioxide about thirty times as fast as rock weathering.

Not all the carbon in sedimentary rocks is in the reduced form (carbon without oxygen) that we have been discussing. About 85% of the sedimentary carbon occurs in the form of carbonate rocks such as limestone (CaCO_3). When these rocks are weathered the limestone dissolves, and bicarbonate ions, together with calcium ions, are washed down to the sea. Clearly these ions cannot simply accumulate in the sea. We can see from Fig. 6-1 that carbonate ions are being added as a result of weathering at a rate that is large enough to supply all of the carbonate in the sea in about 100,000 years—a time span much shorter than the age of the Earth. What happens is that the carbonate ions and the calcium ions are deposited on the floor of the ocean as sediments, sediments that are eventually exposed to the atmosphere and weathered. Some of the carbon is returned to the atmosphere through volcanic action. So here again we have a cycle, in this case a cycle of weathering and sedimentation, that links rocks and ocean. By dividing the rate of weathering into the carbonate content of the rocks, we can calculate the average lifetime of the rocks. The values in Fig. 6-1 give a lifetime of 300 million years.

For simplicity we have described the various cycles of carbon as though they proceed always at the same rate. In fact, there is no reason to suppose

that this is so. It must be remembered, however, that the rates of many of the processes involved in the various cycles of carbon can vary without changing the amount of carbon dioxide in the atmosphere. We have described how this amount is largely determined by equilibrium with the sea. This equilibrium depends on factors such as the average temperature of sea water, but it does not depend directly on the rate of volcanic production of carbon dioxide, for example, or on the rate of burial of organic carbon.

Carbon Dioxide on Venus

It is possible that the amount of carbon dioxide in the atmosphere of Venus is also determined by equilibrium, but equilibrium with rocks rather than with an ocean. As an example of the kind of process that can affect the composition of the atmosphere of Venus, let us consider the reaction of carbon dioxide with the mineral wollastonite (calcium silicate). If a quantity of carbon dioxide were added to the atmosphere of Venus, by volcanoes for example, the amount of carbon dioxide would not show a long-term increase. Instead, the additional carbon dioxide would react with wollastonite, which we assume to be present at the surface of the planet, producing two different minerals, calcite (calcium carbonate) and quartz (silicon dioxide). If, on the other hand, a quantity of carbon dioxide were removed from the atmosphere, the reverse reaction would occur; calcite in the rocks of Venus would react with quartz, releasing enough carbon dioxide to restore the atmosphere to its original level.

Study of this hypothetical equilibrium between atmospheric gases and rocks on Venus indicates that the reaction with wollastonite is only one of a number of reactions that may control the amount of carbon dioxide in the atmosphere. We cannot be sure just which reactions are the most important until we know more than we do about the properties of the surface of Venus. But uncertainty concerning the detailed reactions does not detract from the conclusion that the pressure of carbon dioxide in the atmosphere of Venus is probably in equilibrium with the rocks.

Examination of chemical reactions that could affect the other gases in the atmosphere of Venus shows that every gas in the atmosphere could be in equilibrium with rocks rather similar in composition to those on Earth. This remarkable result may mean that the chemical composition of the atmosphere of Venus is entirely determined by the properties of the surface. The situation is quite different on Earth where the major gases, oxygen and nitrogen, as well as most of the minor gases are far from being in chemical equilibrium with the ground. Why are Earth and Venus so different in this regard?

There are two factors that must be considered in answering this question. First is the existence of processes tending to drive the atmosphere away from equilibrium, and second is the rate of the chemical reactions tending to restore the atmosphere to equilibrium. Photosynthesis is a good example of a disequilibrating process. In the absence of photosynthesis, atmospheric oxygen would recombine with the organic carbon in rocks, producing carbon dioxide. In equilibrium, there would be no oxygen in the atmosphere. Photosynthesis serves to separate carbon and oxygen, tying the carbon up in living creatures and their remains, and releasing the oxygen to the atmosphere. So, one important difference between Earth and Venus is that Earth has photosynthesizing life, but Venus almost certainly does not. Surface temperatures on Venus are too high to permit life.

The high surface temperature is also responsible for the other factor causing the atmosphere of Venus to be close to chemical equilibrium. At high temperatures, the chemical reactions between atmospheric gases and rocks occur rapidly. Thus, the approach to equilibrium is fast. At terrestrial temperatures, on the other hand, reactions are so slow that most of the gases in the atmosphere never get close to equilibrium.

Why did a high surface temperature develop on Venus? In order to answer this question, we must consider some of the facts and theories concerning the origin and evolution of the solar system. We believe that the planets and the Sun all condensed at about the same time, 4.6 billion years ago, from a great cloud of gas and dust. We can put together a fairly complete picture of the chemical composition of the primordial cloud, partly by spectroscopic analysis of sunlight, and partly by chemical analysis of meteorites and of terrestrial materials.

When we compare the primordial composition with the composition of the Earth, we find that there are striking similarities and also striking differences. In particular, there are a number of elements that are relatively as abundant on Earth as they are in the solar system. These are all elements that take part in chemical reactions and were probably present as solid or liquid compounds in the primordial cloud. On the other hand, the family of inert gases—helium, neon, argon, krypton, and xenon—which hardly react at all and could have existed only as gases, is conspicuously absent. The Earth has 10^6 times less xenon relative to silicon, for instance, than the solar system as a whole.

The implication is either that gases were not incorporated in the Earth when it was formed or else that the entire atmosphere was lost subsequent to formation. In either case there must have been a time when the primitive Earth had no atmosphere. Presumably the same was true of the other inner planets. The very important conclusion is that the atmospheres that exist today on Earth, Mars, and Venus are not remnants of the original solar cloud. Instead, they are the result of the release of gases from the minerals that make up the solid parts of these planets.

Lacking definite information on the chemical composition of the gases released to the atmospheres of Mars and Venus, let us assume that Earth is representative. On Earth the gas that has been released to the atmosphere in the greatest abundance is water vapor, with carbon dioxide the next most abundant. We can see this by comparing the mass of water at the surface of the Earth (1.6×10^{24} gm) with the mass of carbon dioxide that has passed through the atmosphere and is now tied up in sedimentary rocks (4×10^{23} gm). Let us now imagine how atmospheres may have evolved on Venus, Earth, and Mars. What would happen if we were to start with planets that had no atmospheres at all and if we were to release from the solid parts of the planets a mixture of water vapor and carbon dioxide?

The Runaway Greenhouse Effect

We first need to decide what the surface temperatures of the planets would be when they had no atmospheres. These temperatures can be determined by equating the infrared radiation emitted by each planet to the solar radiation absorbed by the planet. The calculation was described in Chapter 3; Equation (3-4) presents an expression for the temperature in terms of the flux of solar radiation and the albedo. We do not know what albedos the primitive planets had, but let us assume that they had the same albedo as Mars has today. The values we obtain for the surface temperatures of the primitive planets are shown on the left-hand side in Fig. 6-2. Venus is the hottest because it is the closest to the Sun; Mars is the coldest.

The question we must now consider is how the surface temperatures increase, as a result of the greenhouse effect, when water vapor and carbon dioxide are released to the atmosphere. The greenhouse effect was also discussed in Chapter 3; Equation (3-10) gives the value of the ground temperature in terms of the effective temperature, which we have just evaluated, and the optical thickness of the atmosphere. We shall assume that in this early era water was sufficiently abundant to dominate the absorption and emission of infrared radiation by the atmospheres of Venus, Earth, and Mars. It is then possible to relate the optical thickness of the atmosphere approximately to the vapor pressure of water at the surface, in this way converting Eq. (3-10) to an expression for surface temperature as a function of surface vapor pressure. With this expression we can calculate the way the surface temperatures of Venus, Earth, and Mars evolved as water and carbon dioxide gradually accumulated in their atmospheres.

The evolution is shown in Fig. 6-2, where surface pressure is on the horizontal axis and surface temperature is on the vertical axis. As time

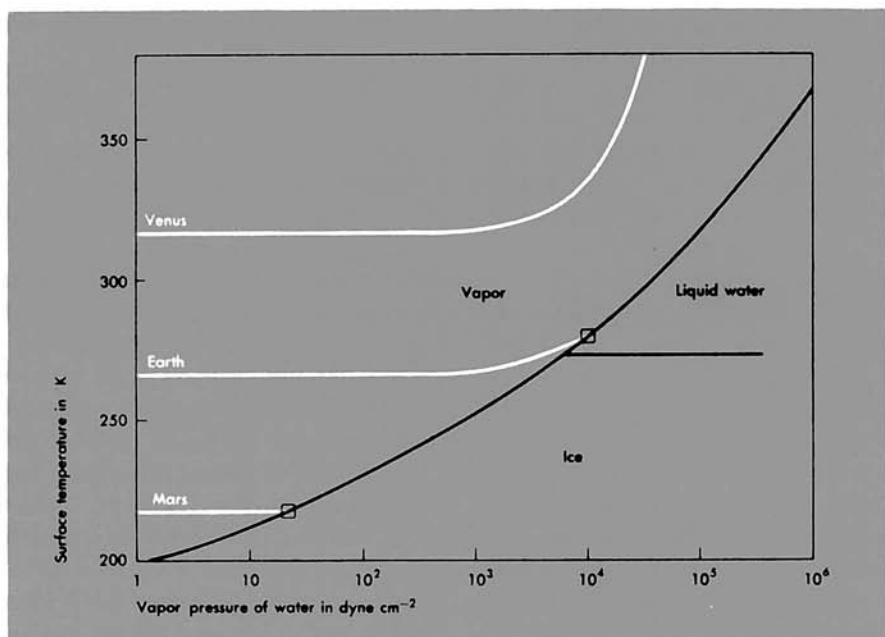


FIGURE 6-2 *The runaway greenhouse effect. The light curves show how surface temperatures increase, due to the greenhouse effect, as water vapor accumulates in the atmospheres of the inner planets. On Mars and on Earth the increase is halted when the water vapor pressure is equal to the saturated vapor pressure (shown as the dark curve), and freezing or condensation occurs. Temperatures are higher on Venus because Venus is closer to the Sun and saturation is never achieved. The temperature runs away. Note that the temperatures on the left-hand axis are not the same for Earth and Venus as the effective temperatures in Table 3-1. They differ because a different albedo has been used. (After S. I. Rasool and C. DeBergh, 1970.)*

passes, and gases accumulate in the atmosphere, we move across the figure from the left-hand side toward the right. The figure also shows the saturated vapor pressure curve of water, which was presented in Fig. 5-1. We must consider how the water in our evolving planetary atmospheres is affected by condensation.

On Mars we find that the temperature is so low that gases do not accumulate in the atmosphere for long before the pressure reaches the saturated vapor pressure of ice. Because the pressure of water vapor in the atmosphere cannot exceed the saturated vapor pressure, any additional water released to the atmosphere condenses on the surface in the form of frost, and the surface temperature ceases to rise. On Mars, therefore, we have a small

greenhouse effect associated with a low pressure of water vapor in the atmosphere.

Earth is warmer than Mars because Earth is closer to the Sun. For this reason, ice does not form at the average temperature of the ground. Nevertheless, a point is reached in the evolution of the atmosphere where the pressure is equal to the saturated vapor pressure, and water condenses to form oceans. From this point on, the situations on Earth and Mars are similar. Release of additional water to the atmosphere does not lead to an increase in surface pressure. The additional water condenses. Moreover, in the absence of increases in water vapor pressure, the greenhouse effect cannot produce further substantial increases in ground temperature.

While the growth of surface pressure and temperature on Mars and on Earth is arrested either by freezing or by condensation of water vapor, no similar end to the growth occurs on Venus. Because Venus is closer to the Sun than Earth or Mars, it is warmer, as Fig. 6-2 shows. The saturated vapor pressure of water increases as the temperature increases, which means that a greater pressure of water is required on Venus in order to cause condensation than is required on Earth or Mars. But it turns out that the greenhouse effect caused by the greater pressure of water is so large that condensation never occurs. As more water is released to the atmosphere, the surface pressure continues to increase, and because of the greenhouse effect, the surface temperature also continues to increase.

This behavior of Venus' atmosphere has been called the *runaway greenhouse effect*. There are uncertainties in the theory particularly as regards the albedo of the primitive Venus and the nature of the gases released to the atmosphere from the solid portion of the planet. The account of the evolution of Venus' atmosphere that we are giving is speculative, as are all theories in this field. Nevertheless, the runaway greenhouse theory provides a satisfying explanation of how atmospheric evolution on Earth and Venus happened to diverge in such a striking way.

The idea is that the growth of the Earth's atmosphere was halted at moderate pressure and temperature by the formation of the oceans. Water condensed, and carbon dioxide dissolved in the water, furnishing a suitable environment for the onset of life. On Venus, however, the runaway greenhouse effect prevented oceans from ever forming. Surface temperatures rose so high that atmospheric gases could react rapidly with surface rocks, leading to the chemical equilibrium between atmosphere and surface that we have already described. In the course of these chemical reactions enough condensible material accumulated in the atmosphere to form the extensive clouds that cover Venus today.

According to this theory, there were at one time large quantities of water in the atmosphere of Venus. Today the atmosphere is very dry, as we related in Chapter 1. Is there some process that could have removed the water from Venus' atmosphere?

Escape of Atmospheric Gases to Space

There is such a process. Both Venus and Earth are losing water all the time, although not very fast. Water vapor (H_2O) in the upper atmospheres of the planets is photodissociated by solar ultraviolet radiation into hydrogen atoms (H) and hydroxyl radicals (OH). These species are involved in a number of chemical reactions that ultimately lead to reformation of the water vapor. Not all of the hydrogen atoms recombine, however. Some of them diffuse upward into the thermosphere (see Fig. 3-1) and, even higher, into the region of the atmosphere known as the *exosphere*.

The exosphere is the region at the top of the atmosphere where collisions between molecules are very infrequent. The density of the atmosphere decreases steadily as the altitude increases, according to the barometric law described in Chapter 1. The distance that a given atmospheric molecule can travel before colliding with another molecule gets longer as the density decreases and there are fewer molecules with which to collide. Eventually this distance, the *mean free path*, becomes so long that a molecule traveling upward can go all the way out to space without colliding at all. The altitude at which the mean free path first becomes long enough for escape to space to be possible is known as the *critical level*. The atmosphere above the critical level is the exosphere. On Earth the critical level is at a height of about 500 km, but the height increases and decreases in response to changes in upper atmosphere density. Densities high in the atmosphere of Venus are less than on Earth, and the critical level occurs at a height of about 200 km.

Although atoms in the exosphere of a planet can travel a long way without colliding, they will not escape from the planet unless they are travelling sufficiently fast. The force of gravity pulls most atoms back down to the critical level just as it pulls a baseball or rocket back down to the ground (see Fig. 6-3). In order to escape from the gravitational attraction of a planet, any body must have more than a certain minimum velocity, called the *escape velocity*. We can derive the escape velocity from the law of conservation of energy.

Consider a body of mass m gm travelling at a velocity v cm sec⁻¹. Its kinetic energy is $\frac{1}{2}mv^2$ erg. In order to escape from the Earth's gravitational field, work must be done against the force of gravity. If energy is to be conserved, we must draw upon the kinetic energy to provide this work. Thus, if the particle is to escape, the kinetic energy must exceed the work required to take the body far from the planet.

The work required for a body of mass m gm originally at the surface of a planet of radius R cm and gravitational acceleration g cm sec⁻² is mgR

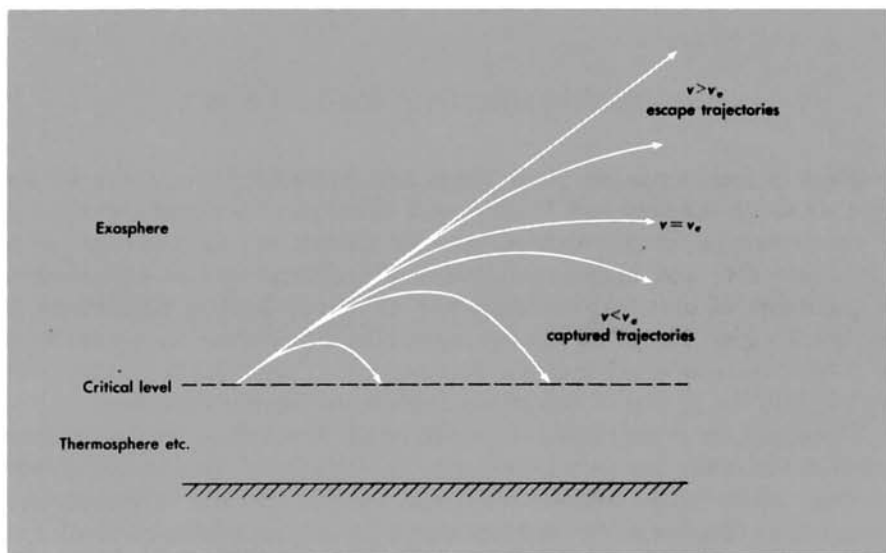


FIGURE 6-3 Only the atoms whose velocity exceeds the escape velocity v_e can escape from the planet. This may be a minute fraction of the total number of atoms at the critical level.

erg. We can understand this result by means of the argument that follows. At the surface of the planet the force on the body is mg dyne. Work is equal to force times distance. Thus, to move the body a distance of one centimeter from the surface requires an amount of energy equal to $mg \times 1$ erg. As we go farther from the center of the planet the force of gravity decreases; it follows the inverse square law as does the intensity of solar radiation (see Chapter 1). Thus, the second centimeter requires less work than the first, and so on. When the body is at a distance R from the surface, it is at a distance $2R$ from the center, and the work per centimeter of displacement has fallen by $2^2 = 4$. We may obtain a reasonable estimate of the work required to take the body far from the planet by ignoring gravity for distances from the surface greater than R , while assuming that gravity is constant for smaller distances. By this means we arrive at $mg \times R$ erg as an estimate of the energy that a body must have in order to escape from the planet. The escape velocity v_e is the velocity of a body that has this minimum amount of kinetic energy,

$$\frac{1}{2} mv_e^2 = mgR \quad (6-1)$$

or

$$v_e = \sqrt{2gR} \quad (6-2)$$

Note that the value of the escape velocity that we have derived does not depend on the mass of the body; it is the same for an atom or for a space ship. Calculated values of escape velocity are shown in Table 6-1. The differences from planet to planet may not seem large, but the number of atoms able to escape from an atmosphere depends in a critical way on the relative magnitudes of the escape velocity v_e and the most probable atomic velocity v_0 .

If we could measure the velocities of a sufficient number of gas atoms, we would find some atoms with any chosen velocity, no matter how small or how large. The number of atoms with extreme velocities, however, would be small. For the most part atoms tend to have velocities close to the most probable velocity

$$v_0 = \sqrt{\frac{2kT}{m}} \quad (6-3)$$

where k is Boltzmann's constant (1.38×10^{-16} erg deg⁻¹), T is the temperature, and m is the atomic mass. Values of v_0 for a range of temperatures are shown in Table 6-2.

Let us now consider escape of gases from the moon. The escape velocity is 2.3 km sec⁻¹. From Table 6-2 we see that the most probable velocity for hydrogen exceeds this figure for all temperatures above 300°K. The day side of the moon is as hot as this, so most hydrogen atoms on the day side of the moon have enough kinetic energy to escape. The moon's gravitational field is clearly too weak to retain an atmosphere of hydrogen.

Since every species has a few atoms with velocities in excess of the escape velocity, all gases on all planets leak more or less rapidly to space. Thus, if a species is to remain on a planet for a long period of time, it must

Table 6-1

Escape Velocities

Planet	Gravitational Acceleration (cm sec ⁻²)	Radius (km)	Escape Velocity (V_e) (km sec ⁻¹)
Mercury	376	2,439	4.3
Venus	888	6,049	10.3
Earth	981	6,371	11.2
Moon	162	1,738	2.3
Mars	373	3,390	5.0
Jupiter	2,620	69,500	60
Saturn	1,120	58,100	36
Uranus	975	24,500	22
Neptune	1,134	24,600	24

Table 6-2

Most Probable Velocities $v_0 = \sqrt{\frac{2kT}{m}}$ in km sec⁻¹

Atom	Atomic Weight (amu)	Temperature (°K)		
		300	600	900
H	1	2.24	3.16	3.87
He	4	1.12	1.58	1.94
O	16	0.56	0.79	0.97

be because the proportion of atoms with velocities exceeding v_0 is very small.

Figure 6-4 gives the proportion of atoms whose velocities exceed a given velocity v in terms of v/v_0 . We see that for $v = v_0$ the proportion is close to $\frac{1}{2}$ as we might have anticipated. But for $v/v_0 = 4$, we find that only one atom in every 10^6 travels faster than v . Now take the case of oxygen on Earth at 600°K . The escape velocity v_e is 11.2 km sec^{-1} while the most probable velocity v_0 is 0.79 . Figure 6-4 shows that, under this circumstance, only 1 atom in every 10^{84} has a velocity larger than the escape velocity. The fraction is so small that the leakage of atomic oxygen from the Earth is

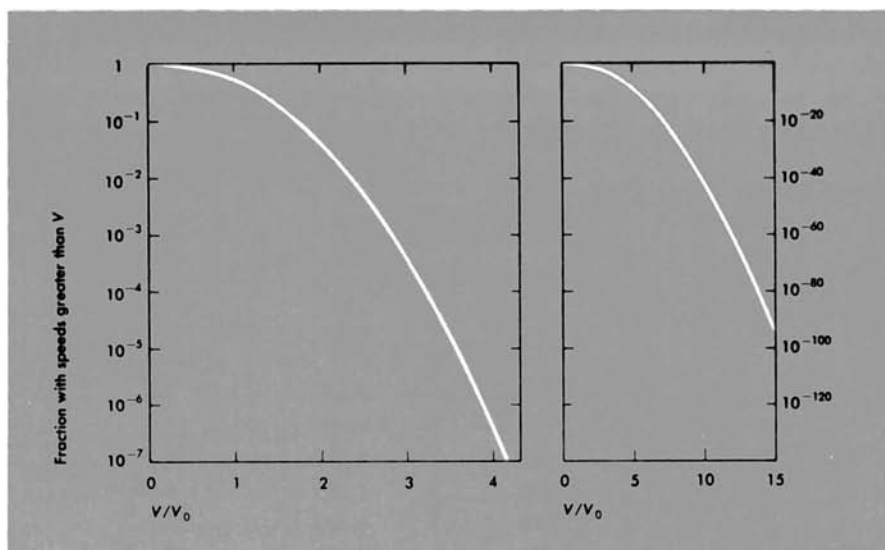


FIGURE 6-4 The fraction of atmospheric molecules with speeds greater than v in terms of the ratio of v to the most probable velocity v_0 . The two parts of the figure are the same, apart from the scale of v/v_0 .

negligible, and even if all of the Earth's oxygen could reach the critical escape level, it would not escape the pull of gravity in the age of the planetary system. For practical purposes, therefore, we can neglect the leakage and consider oxygen to be a permanent constituent of the Earth's atmosphere.

The situation is different for hydrogen, however. Hydrogen is present in the Earth's atmosphere almost exclusively in the form of water. Any large amount of free hydrogen that may once have existed has escaped during the history of the planet. For hydrogen, v_0 is four times as great as for oxygen. This difference is enough to increase the proportion of atoms whose velocities exceed v_e by a factor of 10^{79} , thereby making hydrogen an ephemeral species, at least on a cosmic time scale.

Now we see how Venus may have lost its water. The runaway greenhouse effect could have driven the water into the atmosphere where it would have been subject to photodissociation by solar ultraviolet radiation. The hydrogen produced could then have escaped to space, leaving behind oxygen, which could have been consumed in reactions with surface minerals. This explanation is possible but we cannot, of course, be sure that it is correct.

Whereas hydrogen is an ephemeral species on the inner planets, conditions are very different on the outer planets. For these planets low temperatures (see Table 3-1) give low values for the most probable velocity v_0 , and large surface gravities (see Table 1-5) give large values for the escape velocity v_e . The result is a negligibly small fraction of hydrogen atoms able to escape. The outer planets evidently formed mainly from hydrogen, which was the most abundant constituent of the primordial solar system, and this hydrogen has remained dominant in their atmospheres to the present day (see Table 1-2).

On the other hand, because of their small gravities, the satellites of the inner planets should retain no atmospheres at all. In the case of Mercury, the gravitational acceleration is small and the temperature is high; we do not anticipate any atmosphere, but discriminating measurements have not yet been made.

Terrestrial Life and Atmospheric Oxygen

Living organisms on Earth are another factor having a great effect on atmospheric composition. As an example of the interaction between the atmosphere and the biosphere we shall discuss the cycles of atmospheric oxygen (see Fig. 6-5).

When water is photodissociated and the hydrogen escapes to space, oxygen is left behind in the atmosphere. This source is shown at the top of

Fig. 6-5. It is unimportant compared with the source provided by photosynthesis (the process discussed earlier in this chapter in connection with the cycles of carbon dioxide).

Because of the dominant role of photosynthesis as a source of oxygen, there is a close connection between the oxygen cycles and the carbon cycles. Photosynthesis releases oxygen to the atmosphere and removes carbon dioxide. Respiration and decay remove oxygen from the atmosphere and release carbon dioxide. If life were to cease on Earth and all living things were to decay, the carbon they contain would consume about 10^{19} gm of atmospheric oxygen.

Most organisms do eventually decay, consuming just as much atmospheric oxygen as was released to the atmosphere by photosynthesis when they were growing. Only the small fraction of organic debris that is buried in sediments without decaying corresponds to an overall addition of oxygen to the atmosphere. The oxygen source corresponding to the burial of organic carbon is indicated in the bottom right corner of Fig. 6-5. Oxygen, of course, does not go on accumulating in the atmosphere indefinitely. Ultimately the organic carbon is brought back to the surface as mountains

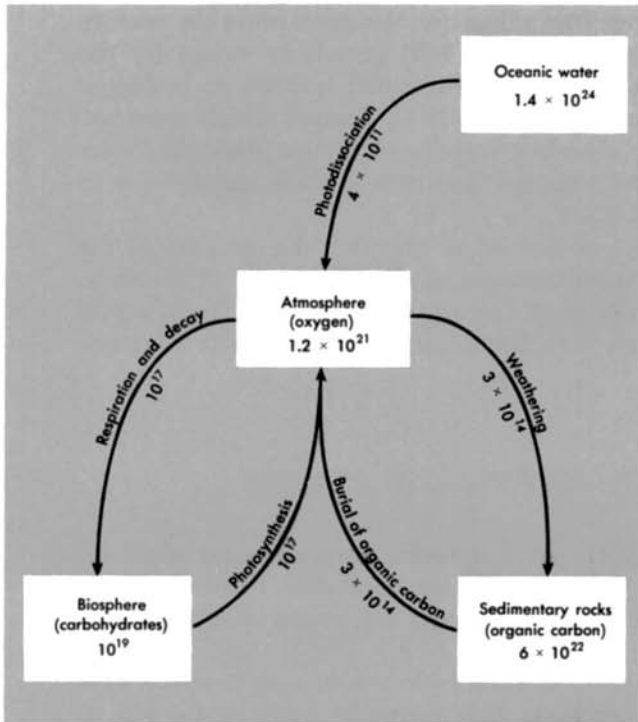


FIGURE 6-5 The cycles of oxygen. Boxes denote reservoirs of oxygen; the contents of the reservoirs are expressed in gm of O_2 . The arrows denote the transfer of material between reservoirs; the rates of transfer are given in gm of O_2 per year.

are raised up and eroded. Once the sedimentary rocks are exposed to the atmosphere, the organic carbon they contain reacts with oxygen to produce carbon dioxide, thus closing the cycle of burial and weathering.

The interesting difference between oxygen and carbon dioxide lies in the stability of the amount in the atmosphere. We have already described how the amount of carbon dioxide in the atmosphere is controlled by the ocean, and we have pointed out the stabilizing effect of the enormous oceanic reservoir. A large amount of carbon dioxide can be released to the atmosphere without causing long-term changes because the extra gas dissolves in the ocean, but there is no equivalent stabilizing reservoir for oxygen. The oxygen content of the ocean is relatively small because oxygen is much less soluble than carbon dioxide. It is possible, therefore, that there have been marked fluctuations in the level of atmospheric oxygen in the course of geologic history.

In particular, in the time before life evolved on Earth there can have been very little oxygen in the atmosphere at all. The small amount produced as a result of photodissociation of water vapor and escape of hydrogen must have been consumed almost at once by the weathering of rocks. Without oxygen there can have been no protective screen of ozone in the atmosphere, and solar ultraviolet radiation in the 2000 Å to 3000 Å wavelength range may have penetrated to the ground. Since this radiation is lethal to simple organisms, conditions were most inhospitable.

In spite of this, primitive plants did develop and start to photosynthesize. Their remains have been found in rocks that are at least two billion years old. In time the plants were able to produce more oxygen by photosynthesis than was consumed by rock weathering, and oxygen began to accumulate in the atmosphere. We do not know when this happened, but it must have preceded the flowering of life that marks the opening of the Paleozoic era, about 600 million years ago.

We still have much to learn about the history of the Earth's atmosphere. The ideas we have described are speculative, but they are the first steps in a process that will enhance our understanding of the factors that have made life on Earth possible.