Moho
Pangea
plate tectonics
polarity
P wave
radioactive decay
radiometric dating
rock cycle
sea-floor spreading
sediment
sedimentary rock
seismic wave
silicate mineral
slab
subduction
surface waves
S wave
tectonic plate
Wilson cycle

Review Questions
1. Why was the theory of continental drift not immediately embraced by the scientific community in the 1920s?
2. What is the Moho?
3. What are the bases for the two major divisions of Earth's interior—one that distinguishes crust, mantle, and core and the other that distinguishes lithosphere and asthenosphere?
4. Compare and contrast P and S seismic waves.
5. Why are earthquakes focused along plate margins?
6. What are the sources of heat in Earth's interior?
7. What is magnetic polarity? What role did it play in the generation of ideas regarding sea-floor spreading?
8. What are the three types of plate boundaries, and what surface features are characteristic of each?
9. What is erosion?
10. How can radioactivity be used to determine the age of a rock?
11. What are the driving forces for plate movement?
12. What is hypothesized to drive the Wilson cycle of plate fragmentation and reassembly?

Critical-Thinking Problems
1. We have seen that cooling of the oceanic lithosphere causes convection, leading to subsidence of the sea floor away from the axis of spreading. The depth h of the ocean floor, measured in meters, increases with age t, measured in millions of years from the present, according to the following equation (valid for sea floor younger than 80 million years old):

   h = 6t - 300

   where h is the depth of the ocean floor in meters. Graph a cross-section of a mid-ocean ridge that is spreading symmetrically in both directions at a rate of 1 cm/yr (10 km/million yr). The age of the oldest sea floor shown should be 80 million years.

2. Duplicate Figure 7.10 (see color version) and answer the following questions.

Further Reading
General

Advanced
port energy and material to different parts of the Earth system where they are utilized in biological and physical processes. This mixing of Earth's fluid and solid parts also helps accomplish an important task; the recycling of the elements. Essential elements are released to the biosphere (the part of Earth that supports life, including the oceans, atmosphere, land surface, and soils) as rocks weather, volcanoes erupt, and nitrogen is made available from the atmosphere by chemical transformations stimulated by lightning discharge. Compared with the rates of utilization by the biosphere, these releases are very slow; they would support only very low rates of biological activity were there not highly efficient nutrient recycling mechanisms.

**Nutrients** are substances, normally obtained in the diet, that are essential to organisms. Nutrient elements are incorporated into living tissue during growth and rapidly returned to the soil or ocean on death. This cycle is repeated many times before the elements are lost from the biosphere, mostly as constituents of sedimentary rocks. The situation is much like our recycling of aluminum cans; recycling substantially reduces our dependence on the extraction of aluminum from Earth and allows us to produce aluminum products much more rapidly. Similarly, element recycling within the biosphere allows for much higher rates of biological productivity.

A number of important recycling systems operate on Earth. We've already discussed the water cycle in Chapter 4. The cycles of the nutrient elements nitrogen and phosphorus are discussed later in this chapter. The recycling of carbon is especially important. As a major constituent of the greenhouse gases carbon dioxide and methane, it affects not only biological productivity but Earth's climate as well. We focus on the carbon cycle in this chapter because of its overarching importance to the Earth system.

**A Journey through the Terrestrial Organic Carbon Cycle**

As an introduction to just one of the global carbon cycles, imagine that we could follow the carbon atom of a CO₂ molecule as it cycled through the terrestrial (land-based) part of the cycle (Figure 8-1). The carbon atom in CO₂ is **inorganic carbon**—it is not associated with compounds formed by living organisms and it does not contain carbon-carbon or carbon-hydrogen bonds. After spending nearly a decade moving with the winds in the troposphere, the gaseous CO₂ molecule will have visited both the Northern and Southern Hemispheres several times. Then, one spring, during the annual greening of the Northern Hemisphere (Figure 8-2), the CO₂ molecule passes through a small opening in a leaf, the photosynthetic apparatus of a plant. Through a flurry of collisions with other molecules and atoms, the oxygen atoms are ripped from the molecule while hydrogen, nitrogen, and other carbon atoms become attached. Our carbon atom, as part of the leaf, is now **organic carbon**. Some leaves are consumed and digested by animals. The carbon in these leaves is then released back to the atmosphere by the animals' respiration as CO₂.

Summer passes, fall arrives, and the leaf that contains the carbon atom has not been eaten. The nitrifying substances and water that the leaf has received from the tree have ceased to flow. The leaf is released from the branch and settles to the ground. Other leaves fall on top, burying it in a thick mat of decaying matter. The carbon atom is part of the soil, where it will remain for about the next 500 years. By the end of that time, bacteria and fungi will have decomposed the organic matter that contains the atom. The chemical reactions that result transform the carbon atom once again into gaseous CO₂ molecule, which escapes back to the atmosphere.

This life cycle of a carbon atom is repeated nearly 500 times on average before it "leaks" occurs. Once in a while, before the organic matter that contains the carbon atom decomposes, the soil crumbles and is transported by rivers to the oceans. There it settles with other particles to the sea floor and is buried by subsequent sediments or carried with its underlying oceanic plate deep into a subduction zone. Under elevated temperatures and pressures the carbon atom may be converted into gaseous carbon atoms and escape to the surface, or be converted into a component of sedimentary or metamorphic rock.

The carbon atom may spend millions of years in the sedimentary/metamorphic rock reservoir, as mountain belts form, thrusting deeply buried rocks to Earth's surface and beyond to great elevations. Eventually our carbon atom will be transferred from its burial place within Earth's interior to the surface. Here environmental forces, both biological and physical, will cause the sedimentary rock containing the carbon atom to disintegrate during the process of weathering. In this process, the organic carbon reacts with oxygen from the atmosphere and forms (inorganic) CO₂, which escapes as a gas to the atmosphere. The weathering cycle is not complete, however, and is the connecting link in the long path this particular carbon atom has taken—from the atmosphere, to the plant, to the soil, to the sediment, to the sedimentary rock, and back to the atmosphere.

The path the carbon atom has taken encompasses the terrestrial organic carbon cycle, operating on time scales that are short (years to decades) and those that are long (centuries to millennia). Reacting with carbon in both reservoirs responds to perturbations. We will use the response to the release of carbon dioxide from the burning of fossil fuels as an example of the dynamics of the carbon cycle and of material recycling systems in general. Figure 8-4 shows the seasonal fluctuations in the atmospheric CO₂ level for three years (1999-2001) measured from the Mauna Loa, Hawaii. We saw a similar graph in Figure 1-2. In Figure 8-4, however, we focus on the natural seasonal cycle rather than the gradual increase in CO₂ from fossil-fuel burning and deforestation. The CO₂ content falls during the Northern Hemisphere summer, when photosynthesis (and the growth of leaves) surpasses respiration and decomposition. It then rises during...
Atmospheric CO₂

Living biomass
60 GtC

Atmospheric CO₂
780 GtC

Oceanic dissolved CO₂
740 GtC

Oceanic carbonate ion
1300 GtC

Organic carbon in
sediments/soils
1900 GtC

Marine carbonate
sediments
2500 GtC

Fossil fuels
4700 GtC

Organic carbon in
sediiments/soils
10,000,000 GtC

Limestone in sedimentary rocks
40,000,000 GtC

FIGURE 8-3
Reservoirs of carbon at or near Earth’s surface.

The late fall to early spring, when respiration and decomposition of the previous season’s crop of leaves decrease photosynthesis. Because Hawaii is in the Northern Hemisphere (17° N), measurements made at Mauna Loa reflect the annual cycle in that hemisphere (see Chapter 2 for further discussion). Let us cast these observations in terms of systems theory.

Reservoirs. From the systems point of view, the atmosphere is a reservoir of carbon in the form of CO₂ (Figure 8-5). Reservoirs are typically characterized in terms of the amount of material they are holding at any particular time. Their sizes are commonly expressed either in mass units or volume units. (The units are typically molar; refer to the Box “The Concept of the Mole” if you need a refresher.) In Figure 8-5, the amount of carbon is expressed in gigatons (Gt) of carbon, or Gt C. A gigaton is a billion metric tons, and 1 metric ton is 1000 kg. With the notation Gt C, we are keeping track of the mass of only the carbon atoms, not the other atoms to which they are attached.

Reservoirs are temporary repositories for material that flows through them, and their sizes vary in response to imbalances between inflow and outflow. Typically expressed in unit mass, unit volume, or moles per unit time. The inflow to the atmospheric CO₂ reservoir is the combination of the processes of respiration and decomposition.

The outflow from the reservoir is photosynthesis, and the rate of outflow is also expressed as gigatons of carbon per year.

Steady State. If the rates of inflow and outflow were such that the atmospheric CO₂ reservoir remained at a constant value with time, we would say that steady state had been achieved. Steady state is a condition in which the state of a system component is constant with time. Steady state could be achieved if no inflow and no outflow existed, that is, if both processes ceased. A constant level could also be maintained if the rate of inflow of CO₂ into the atmosphere equaled the rate of outflow.

Any imbalance in these rates leads to a change in the level of atmospheric CO₂. When the inflow exceeds the outflow, the atmospheric CO₂ level rises. (This situation is analogous to the Northern Hemisphere winter condition.) When the outflow exceeds the inflow, the level falls (analogous to the Northern Hemisphere summer condition).

In the record of atmospheric CO₂ variations in Figure 8-4, we see that one maximum and one minimum is reached each year. At these times the fluxes are in balance. The system is not really in steady state at these times, however, because the reservoir size is unchanging only for an instant. Averaged over longer times, though, the natural cycle of CO₂ is thought to be close to steady state, despite seasonal imbalances. Because of anthropogenic disturbances, the atmospheric CO₂ level is not currently at steady state, as demonstrated by the steady rise in CO₂ over the past several decades.

Steady state can be maintained over time only if the rates of inflow and outflow are sensitive to changes in the size of the reservoir. In systems terminology, this means that there must be couplings that link the reservoir size to the processes that govern inflow and outflow (see Chapter 2). Consider what would happen if a reservoir in steady state were perturbed by an addition of material. If the coupling governing inflow was negative or if that governing outflow was positive (i.e., inflow decreased or outflow increased), the reservoir would return to its original state. One such negative feedback loop exists between the photosynthetic rate of plants and atmospheric CO₂. As CO₂ levels go up, plants photosynthesize more rapidly; this effect has been called CO₂ fertilization (Figure 8-6). But as they do, CO₂ levels tend to fall, because CO₂ is consumed by plants during photosynthesis.

### USEFUL CONCEPTS

#### The Concept of the Mole

A mole and molecules are typically measured in units called moles. A mole (mol) is the amount of substance containing as many fundamental or chemical units (atoms, molecules, ions, electron pairs, etc.) as there are atoms in 12 g of carbon-12. The mole is the SI base unit of amount of substance.

- **Atomic Weight:** The mass of an atom is its atomic weight (expressed in atomic mass units, AMU or u). The atomic weight of carbon is approximately 12 u.
- **Molar Mass:** The molar mass of a substance is the mass of one mole of that substance. For example, the molar mass of water (H₂O) is approximately 18 g/mol.

#### Chemical Reactions

Consider the chemical reaction for the formation of sodium chloride (NaCl):

\[ 	ext{Na}^+ + 	ext{Cl}^- \rightarrow \text{NaCl} \]

This reaction shows that one atom of sodium (Na) will react with one atom of chlorine (Cl) to form one molecule of NaCl. Therefore, 1 mol of Na will react with 1 mol of Cl to form 1 mol of NaCl. To express this equation in mass units, we would need to use the atomic weights of sodium and chlorine. If we did, we would find that 22.99 g of Na reacts with 35.45 g of Cl to produce 58.44 g of NaCl—a more cumbersome calculation.

### Figure 8-4

Annual variations in atmospheric CO₂ from the Mauna Loa Observatory for 1999-2001. The gradual increase due to fossil-fuel burning and deforestation accounts for the offset from year to year. (Data courtesy Oak Ridge National Laboratory; http://www.esrl.noaa.gov/gmd/ccgg/trends/co2/ois-loa.html)
The Short-Term Organic Carbon Cycle

The short-term organic carbon cycle involves processes ranging from those we can observe and appreciate on a daily to seasonal time scale (see Figure 8-2), such as the processes of photosynthesis and respiration, to processes of decomposition that are somewhat slower (Figure 8-7). The key step in this cycle is the conversion of inorganic carbon (atmospheric CO$_2$) to organic carbon by the process of photosynthesis. We are not so much interested in the process itself but rather in its impact on the global cycle, which is generally expressed as primary productivity. Primary productivity is the amount of organic matter produced by photosynthesis in a unit time over a unit area of Earth's surface. That amount depends on the population size of primary producers—that is, plants (or other types of photosynthesizers or even chemoautotrophs) that provide energy other organisms can use. The relationship is not simple, however, because some primary producer species are much more productive than others. In its simplest representation, primary production involves a chemical reaction between CO$_2$ and water to form organic matter and oxygen:

$$\text{Photosynthesis: } \text{CO}_2 + \text{H}_2\text{O} \rightarrow \text{CH}_2\text{O} + \text{O}_2$$

Here, organic matter is represented by CH$_2$O, the simplest carbohydrate, or complex of carbon, hydrogen, and oxygen. In reality the molecules making up organic matter are much larger than this simple carbohydrate, and they contain small amounts of many other elements, including nitrogen and phosphorus. For our purposes, however, this simpler representation suffices. Photosynthesis does not occur spontaneously at Earth's surface but instead requires an input of energy from the Sun. Plants, algae, and bacteria have evolved pigments that are able to capture the energy of sunlight and convert it to chemical energy, part of which is stored

![Image](image-url)
in living tissues. This chemical energy is then utilized by other organisms that cannot utilize solar energy directly. Such organisms, including animals, are called consumers.

Most of the photosynthates that occur each year lead to the formation of tissue that is recycled rapidly, including the leaves of trees. This recycling is the cause of the seasonal variations observed in the CO₂ record of Figure 8-4. However, most of the organic carbon that makes up plant tissues has a residence time of many decades. That is because the bulk of the organic carbon in terrestrial plants is contained not in the leaves but in the roots and trunks of slow-growing trees. In other words, most of the biomass of primary producers on Earth is contained in tree roots and trunks. Biomass is the total mass of organic matter in living organisms in a particular reservoir. In terms of carbon, the total biomass—the combined biomass of all primary producers and consumers—is about equal to the atmospheric carbon reservoir (see Figure 8-3).

Consumer biomass is a small percentage (only about 1%) of the biomass of the producers. Consumers derive their metabolic energy from the chemical energy stored in plants, by eating the tissues and respiring. Respiration is the reverse of photosynthesis: It is the chemical reaction between oxygen and organic tissue that yields CO₂ and water:

\[
\text{Respiration: } \text{CH}_2\text{O} + \text{O}_2 \rightarrow \text{CO}_2 + \text{H}_2\text{O}.
\]

During photosynthesis, plants use solar energy to create tissues. But, like animals, plants produce their metabolic energy through respiration. Hence, respiration is more than just the "breathing" performed by animals. Unlike photosynthesis, respiration would proceed abiotically because it releases rather than requires energy. However, it would do so very slowly. Organisms are able to accelerate this chemical reaction by the use of enzymes, chemical compounds (specifically, proteins) that they synthesize for this purpose.

On land, about half the organic material produced by photosynthesis is respired by animals and by plants themselves. The remaining material is added to the organic-rich upper layers of soil. A host of microscopic bacteria and fungi live in soil. Their metabolic requirements are satisfied through the decomposition of the large store of organic matter that is buried there, fueled by the supply of oxygen from the overlying layers. A biological process that uses oxygen is said to be aerobic, and an organism that carries out aerobic metabolism is an aerobe. The chemical reaction for this aerobic decomposition is identical to the process of respiration.

Because the only source of oxygen is the air, the microorganisms that live well below the surface must be adapted to environments that are devoid of oxygen. A biological process that occurs in the absence of oxygen is said to be anaerobic, and an organism that carries out anaerobic metabolism is an anaerobe. In the O₂-free environment of deep soil live anaerobic bacteria that can decompose organic matter by an overall process known as methanogenesis. Methanogenesis is an anaerobic form of metabolism that involves multiple steps, carried out by different bacteria. One step of this complex process involves simpler forms, including hydrogen (H₂) and acetate that methanogens can then use to form both oxidized carbon (in CO₂) and reduced carbon (methane). The overall process can be represented as:

\[
\text{Methanogenesis: } 2\text{CH}_2\text{O} \rightarrow \text{CO}_2 + \text{CH}_4.
\]

(Other, more complex compounds can also be used and produced during anaerobic metabolism, but for the purposes of studying the carbon cycle, this representation is adequate.) The gases CO₂ and CH₄ can escape to the atmosphere. Once there, the CO₂ continues the short-term cycling path described earlier. Methane, however, is chemically unstable in our O₂-rich atmosphere and is destroyed by a series of oxidation reactions. The carbon contained in CH₄ combines with O₂ to form CO₂. With atmospheric reservoir size of 540 Gt(C) and a supply rate from fermentation of 0.5 Gt(C)/yr (see Figure 8-7), the residence time of CH₄ in the atmosphere is approximately 10 years.

The land surface is continuously stripped of its soil cover by the action of winds and water. On average, about 5 cm of soil is eroded from the land surface every 1,000 years and transported by rivers to the oceans. Although river systems have a substantial capacity for storing sediment in flood plains and deltas, eventually most of this sediment makes its way to the oceans and is deposited at the sea floor. These sediments contain whatever organic matter has survived the trip from land to sea. Thus, there is a transfer of organic carbon from the terrestrial to the marine realm, one that amounts to about 0.3 Gt(C) per year. However, this transfer is small compared with the flux through the oceanic water column of organic matter produced in the ocean.

The Marine Organic Carbon Cycle on Short Time Scales

Producers and Consumers. The dominant primary producers in the ocean are the free-floating, photosynthetic marine microorganisms referred to as phytoplankton. In more general terms, producers are organisms with any type of metabolism that float freely in aquatic environments. These organisms—primarily diatoms (Figure 8-8a) and other algae, such as coccolithophorids (Figure 8-8b)—live in the photic zone. The photic zone is the uppermost part of the oceanic water column where there is sufficient light for photosynthesis: about the upper 100 m of the water column in the open ocean, and in shallower waters near shore, where water clarity is reduced. It roughly corresponds to the "surface ocean," which, as we saw in Chapter 5, is the upper part of the ocean mixed by the wind. Phytoplankton consume CO₂ and produce O₂ through photosynthesis in much the same way as do land-based plants. Although the gases phytoplankton use and produce are not dissolved in seawater, there is continuous gas exchange between the atmosphere and the ocean. Thus, the activities of phytoplankton affect the atmosphere as well as the ocean.

Much of the organic matter produced in the surface ocean by phytoplankton is consumed by zooplankton. Zooplankton are free-floating marine microorganisms, including small invertebrates and microorganisms such as foraminifera (Figure 8-8c) and radiolarians (Figure 8-8d), that cannot photosynthesize. Zooplankton consume food pellets that, together with other large particles of decaying organic matter, settle through the water column to great depths. In contrast to the flux of organic matter from treetops to the ground, though, only about 1% of this material survives the trip to the sea floor. Even then, the material is subject to efficient recycling by aerobic and
Nutrient Limitation. Some elements are classified as nutrients for marine phytoplankton because they are essential for growth and exist at suboptimal concentrations: an increase in their concentration leads to higher rates of primary productivity. Other elements are toxic to marine life—that is, they are poisonous to certain organisms—because their concentrations in seawater are above the optimum for growth. Even nutrient elements can be toxic if their concentrations become too high. There is great variability in the concentration of many elements throughout the world's oceans. Each element, depending on its concentration, can be either a nutrient or a toxic substance. The situation is analogous to the parabolic growth curves for fitness discussed in Chapter 2 (Figure 2-5a), except that element concentration substitutes for temperature. For each element, there is some optimum concentration that favors biological productivity.

Marine phytoplankton incorporate many nutrient elements into their tissues in ratios that appear to be nearly identical in all species. These ratios are called Redfield ratios, in honor of Alfred C. Redfield, the oceanographer who first described this phenomenon. Even more remarkably, the ratios of many of these elements in seawater is nearly identical to that in phytoplankton. For example, the elemental ratio of carbon:nitrogen:phosphorus is very nearly 106:16:1 (Table 8-1). A chicken-or-egg question arises: Does the composition of seawater determine the composition of organisms that live in the sea, or does the composition of marine organisms determine the composition of seawater? To answer this question, we must consider the distribution of primary productivity in the oceans.

Where are rates of primary productivity greatest in the oceans? We now have a sophisticated way of answering this question, by satellite. With a color scanner, researchers can quantify the color of seawater from space (Figure 8-10). The color of the ocean surface is strongly influenced by the density of phytoplankton, which contain photosynthetic pigments. In the early 1960s and 1970s, oceanographers began to study the relationship between the concentration of these pigments and the abundance and productivity of near-surface-dwelling phytoplankton. The researchers found that regions of the ocean that have low concentrations of chlorophyll (a}

### Table 8-1

<table>
<thead>
<tr>
<th>Redfield Ratios</th>
<th>Relative number of atoms in living phytoplankton</th>
</tr>
</thead>
<tbody>
<tr>
<td>Carbon</td>
<td>106</td>
</tr>
<tr>
<td>Nitrogen</td>
<td>16</td>
</tr>
<tr>
<td>Phosphorus</td>
<td>1</td>
</tr>
<tr>
<td>Iron</td>
<td>0.01</td>
</tr>
</tbody>
</table>
The Long-Term Organic Carbon Cycle

The processes we have discussed thus far affect the atmospheric CO₂ balance on time scales shorter than a century. On longer time scales, these processes must be viewed in close balance: Because the fluxes involved are large, persistent imbalances would lead to intolerable fluctuations in atmospheric CO₂. Geologic processes come in as the important controls on atmospheric CO₂ on longer time scales (Figure 8-12). The fluxes of carbon involved in these processes are small, and the reservoirs involved are large. Together these two adjustments in sources and sinks mean that the importance of the geological processes affecting the sedimentary reservoir and the atmosphere is minimal on long time scales (on the order of 1,000 years to 1 million years).

Carbon Burial in Sedimentary Rocks

The flux of land-derived and marine sediments to the seafloor fills sedimentary basins, many of which flank the margins of the continents. The continuous supply of carbon fixation in these basins leads to the burial of previously deposited material. Eventually, as the process continues, sediments become buried to a depth of a few kilometers below the sea floor and become lithified. The organic carbon associated with these sediments is then cycled in the sedimentary rock until weathering liberates the material to the biosphere.

Carbon Leaks and Oxygen Replenishment

This organic carbon burial represents a leak of material from the short-term organic carbon cycle (see Figure 8-12). It is this leak, rather than photosynthesis alone (which is nearly balanced by respiration and decay), that maintains the O₂ content of the atmosphere. Oxygen is continually replenished from the atmosphere by chemical reactions with reduced materials (especially organic matter) that are preserved in rocks exposed at Earth's surface and with reduced volcanic gases such as hydrogen, sulfur dioxide, and carbon monoxide. The loss of O₂ is very slow, but the O₂ concentration would reach zero in a few million years if that gas were not supplied from other sources. Oxygen is replenished by the leak of organic matter into the sedimentary rock reservoir. For every carbon atom that enters this reservoir as organic carbon, one oxygen molecule is left behind. This is because the O₂ liberated during the photosynthesis of that carbon was not utilized during respiration or decomposition, and thus the gas remains in the atmosphere.

Formation of Fossil Fuels

During burial, the organic material in the sediment also undergoes significant changes in its structure and chemistry, and fossil fuels may...
The Sedimentary Organic Carbon Reservoir. Sedimentary rocks contain by far the greatest quantity of organic carbon on Earth: approximately 10^12 Gton C (see Figure 8-3). Most of the organic carbon is found in shales, which are fine-grained sedimentary rocks formed by the lithification of muds. The residence time for the sedimentary organic carbon reservoir is about 200 million years.

Weathering of Organic Carbon in Sedimentary Rocks

Weathering of the organic carbon in sedimentary rocks is an oxidation process requiring atmospheric O_2 either by direct exposure to the atmosphere or by exposure to groundwaters containing dissolved O_2. The oxidation of this material can be represented by the same stoichiometric chemical reaction that we used previously for respiration and aerobic decomposition. The organic matter reacts with oxygen, releasing carbon dioxide to the atmosphere or groundwater.

In this sense, the mining, pumping, and combustion of fossil fuels represent merely an acceleration of the weathering process. The rocks from which humans have removed these fuels would likely have become exposed on the surface and undergone oxidation to form CO_2 and oxygen in the distant future. Human intervention, however, has speeded up this process by a factor of a million or more for these fossil-fuel deposits. The release of organic matter from sedimentary rocks is occurring much faster than it can be replaced. Hence fossil fuels represent only a short-term energy source.

Summary of the Organic Carbon Cycle

We have now explored the entire global organic carbon cycle. Pathways exist to recycle carbon from all reservoirs (see Figure 8-12). Every reservoir in this cycle is directly connected to the atmosphere. Thus, the CO_2 concentration of the atmosphere changes continuously in response to changes in the flux of carbon to and from these reservoirs. The responses to these changes are rapid for the large fluxes associated with the biota, soils, and marine sediments but slow for the small fluxes from the sedimentary rock reservoir. This observation will prove to be important in later chapters when we consider the fate of CO_2 added to the atmosphere from the burning of fossil fuels.

The Inorganic Carbon Cycle

The photosynthesis of CO_2 to reduced carbon in organic matter, and its subsequent respiration to CO_2 through respiration, decomposition, and weathering, is central to the organic carbon cycle. But there are other sources and sinks for atmospheric CO_2. Carbon dioxide readily dissolves in rainwater and seawater and then undergoes rapid chemical reactions to other ionic forms of inorganic carbon. The oxidized carbon in these waters is chemically reactive and becomes involved in a number of chemical processes. Because they do not involve organic carbon directly, these processes are together referred to as the inorganic carbon cycle.

The important reservoirs of inorganic carbon are the atmosphere, which we have discussed at length; the oceans; sediments, and sedimentary rocks (see Figure 8-3). The sediment and sedimentary-rock carbon reservoirs consist primarily of limestone. Limestone is a rock composed largely of calcium carbonate (CaCO_3), generally in the form of the mineral calcite. The magnesium-rich carbonate mineral dolomite, CaMg(CO_3)_2, is abundant in older sedimentary rocks.

Carbon Exchange between Ocean and Atmosphere

Carbon dioxide is continuously exchanged between the atmosphere and the ocean. The distribution of sources and sinks of CO_2 is tied to the circulation and productivity patterns of the ocean (Figure 8-13). In regions of the ocean where high rates of primary productivity have created surface waters with low CO_2, CO_2 diffuses from the atmosphere to the ocean. In other words, the net flow of CO_2 is down the concentration gradient—from regions of higher CO_2 concentration (in this case, the atmosphere) to regions of lower concentration (the ocean). Conversely, upwelling regions, such as the equatorial Pacific surface waters, have high CO_2 concentrations because deep wa-
Chapter 8 Recycling of the Elements: The Carbon Cycle

The release of the second ion converts bi-carbonate ion in the previous reaction to carbonate ion (CO$_3^{2-}$): \[
\text{HCO}_3^- \rightarrow \text{H}^+ + \text{CO}_3^{2-} \tag{3}
\]
(Note that we have previously used the term "carbonate" to refer to calcium and magnesium carbonate minerals.) For a given hydrogen ion concentration ([H$^+$]), the relative amounts of bicarbonate and carbonate ions are adjusted until equilibrium is achieved. A pH above neutral favors the carbonate ion, whereas a pH below neutral favors the bicarbonate ion.

USEFUL CONCEPTS

The hydrogen ion [$H^+$] is the smallest of all cations. The small size and ionic charge make hydrogen ions extremely reactive. They tend to form infinite solids, breaking bonds and causing the reactions that make up the solids to dissolve. Solutions (liquids with dissolved materials) with high concentrations of hydrogen ions are called acids. Vinegar and hydrochloric acid are common acids. Solutions with low concentrations of hydrogen ions are called bases (or alkalis). Baking soda and lime dissolved in water form common basic solutions.

Strong acids are solutions that completely dissociate. When dissolved in water, the anions separate completely from the hydrogen cations. This dissociation leads to high hydrogen ion concentrations. For example, the strong acid hydrochloric acid (HCl) dissociates to form hydrogen ions and chloride ions:

\[
\text{HCl} \rightarrow \text{H}^+ + \text{Cl}^- \]

Other strong acids include nitric acid (HNO$_3$) and sulfuric acid (H$_2$SO$_4$). Weak acids, such as carbonic acid (H$_2$CO$_3$) and hydrocyanic acid (HCN), only partially dissociate when dissolved in water.

The concentration of dissociated hydrogen ions (expressed in moles per liter of solution) determines the acidity of the solution. Acidity is commonly measured by the pH scale, which is a logarithmic scale. The pH of a solution is a close approximation of the negative of the log of the hydrogen ion concentration: \[
\text{pH} = -\log[H^+] \]

(We say "approximation" because chemists define pH in terms of the activity of hydrogen, which is the concentration of hydrogen ion available for chemical reaction. A small fraction of the hydrogen ions are involved in ionic interactions with other ions.) For example, a solution with a hydrogen ion concentration of $10^{-7}$ mol per liter would have a pH of 7, because the logarithm of $10^{-7}$ is -4.

The pH of the surface and deep oceans is slightly basic about 8 and 7.5, respectively. Rainwater that is in equilibrium with atmospheric CO$_2$ has a slightly acidic pH between 5 and 6. Most lakes, rivers, and streams range in pH from about 6 to 9, that is, from slightly acidic to mildly basic.
Perturbation of this equilibrium (e.g., by the diffusion of anthropogenic CO₂ from the atmosphere to the ocean) changes the pH of seawater. These changes in pH affect the relative concentrations of carbonic acid, bicarbonate ion, and carbonate ion in the following way: CO₂ dissolves, forming carbonic acid (reaction 1); the carbonic acid dissociates, forming bicarbonate and hydrogen ions (reaction 2), which causes the pH to drop. Hydrogen ions then react with carbonate ion, forming another bicarbonate ion (the reverse of reaction 3). The overall chemical reaction describing the uptake of atmospheric CO₂ is the sum of these three reactions (reactions 1, 2, and the reverse of 3):

\[
\text{CO}_2 + \text{H}_2\text{O} + \text{CO}_2 \rightleftharpoons 2\text{HCO}_3^-. 
\]

That, the ocean’s capacity to take up CO₂ derived from fossil fuels is enhanced relative to what it would be if it simply equilibrated with the atmosphere, because it is converted to other forms of inorganic carbon. There is a limitation on the amount that can be taken up, however: the amount of carbonate ion in the ocean, which is smaller than the total available fossil fuel. Thus, if we continue to utilize fossil fuels, the ocean’s capacity to take it up will become depleted, and a larger fraction will remain in the atmosphere.

Chemical Weathering

A similar chemistry applies when atmospheric CO₂ dissolves in raindrops, making them naturally acidic. As we noted earlier, the unbuffered pH of rainwater is generally between 5 and 6. Rocks exposed at Earth’s surface undergo chemical attack from this rain of dilute acid, a process known as chemical weathering.

Crystalline rocks are composed mainly of two types of minerals: carbonates and silicates. Carbonate minerals, such as calcite, contain carbon in combination with oxygen and other elements. Calcite and dolomite are the most abundant carbonates at Earth’s surface, occurring as the dominant minerals in limestones and dolostones and as minor minerals in a host of other rock types. These minerals have fairly simple chemical formulas: CaCO₃ and CaMg(CO₃)₂, respectively. Silicate minerals contain compounds of silicon and oxygen. They tend to have rather complicated compositions. They are most abundant in igneous rocks but are also common minerals in sedimentary and metamorphic rocks. (See Chapter 6 for a further discussion of rocks and minerals.)

In discussing the carbon cycle, we are most interested in the weathering of the calcium-bearing minerals, because calcium ions released by weathering are used by organisms in the construction of calcium carbonate shells and skeletons. Important calcium-bearing silicates include anorthite, CaAl₂Si₂O₈, and hornblende, NaCa₂(Al₃Fe₂, Al₂(Al₂Si₃O₈)(OH)₄. However, we will use wollastonite, CaSiO₃, which has a much simpler formula, to represent calcium silicates. It is the relative amount of calcium and silica that matter to our discussion of chemical weathering, not the detailed compositions of the weathered minerals.

When exposed to rain, both carbonates and silicates weather chemically, although the carbonate minerals do so much more rapidly (Figure 8-14). Chemical weathering neutralizes the acidity of carbonic acid in much the same way that an antacid neutralizes the acidity of your stomach.

Carbonate weathering:

\[
\text{CaCO}_3 + \text{H}_2\text{O} \rightleftharpoons \text{Ca}^{2+} + 2\text{HCO}_3^- 
\]

\[
\text{Ca}^{2+} + 2\text{HCO}_3^- \rightleftharpoons \text{CaCO}_3 + \text{H}_2\text{O} + \text{CO}_2 
\]

\[
\text{Ca}^{2+} + \text{H}_2\text{O} + \text{CO}_2 \rightleftharpoons 2\text{HCO}_3^- 
\]

\[
\text{CaCO}_3 + \text{H}_2\text{O} + \text{CO}_2 \rightleftharpoons 2\text{HCO}_3^- 
\]

In both weathering reactions, the products include Ca²⁺ and HCO₃⁻. Silicate weathering also yields dissolved silica (SiO₂), which, together with Ca²⁺ and HCO₃⁻, is transported to rivers and ultimately to the oceans. Silicate weathering consumes twice as much dissolved CO₂ (in the form of carbonic acid) as does carbonate weathering. This fact will prove important later in this chapter when we attempt to balance the carbon cycle.

Carbonate Mineral Deposition

The oceans eventually receive the products—solid particles and dissolved materials—of the chemical and physical weathering and erosion of the land surface. Most of the particulate material is deposited near the mouths of rivers in deltas, beaches, and other deposits near shore. In contrast, the dissolved materials mix with seawater and, if their residence times are longer than the ocean's mixing time, are distributed throughout the world's oceans. The constant flux of dissolved materials from land to sea would gradually increase the salinity of seawater by dilution were it not for processes that continuously remove material from the sea (see Chapter 5).

Some organisms, such as diatoms (Figure 8-8a), radiolarians (Figure 8-8d), and sponges, remove dissolved silica from seawater. They convert it into solid (opaline) silica as the structural part of their skeletons. Other organisms, such as foraminifera (Figure 8-8c), coccolithophorids (Figure 8-8b), corals, and shellfish, produce solid CaCO₃ in forming their shells and skeletons. Although these minerals can form abiotically (without the aid of organisms), most of the CaCO₃ precipitated from the ocean today is formed by such organisms.

These carbonate-producing marine organisms remove Ca²⁺ and HCO₃⁻ from seawater and precipitate CaCO₃ as a shell or skeleton. The overall chemical reaction is essentially the reverse of the carbonate weathering reaction:

Carbonate precipitation:

\[
\text{Ca}^{2+} + 2\text{HCO}_3^- \rightleftharpoons \text{CaCO}_3 + \text{H}_2\text{O} + \text{CO}_2 
\]

\[
\text{CaCO}_3 + \text{H}_2\text{O} + \text{CO}_2 \rightleftharpoons 2\text{HCO}_3^- 
\]

\[
\text{CaCO}_3 + \text{H}_2\text{O} + \text{CO}_2 \rightleftharpoons 2\text{HCO}_3^- 
\]

Calcium carbonate producers cause a shift in the ocean's carbon chemistry. By precipitating CaCO₃ shells or skeletons, they enhance HCO₃⁻ and, because of the equilibria shown, dissolved CO₂ concentrations and reduce HCO₃⁻ concentrations, as well as pH. By increasing the concentration of dissolved CO₂ in surface waters, carbonate-producing organisms produce a CO₂ gradient between the oceans and the atmosphere. This gradient promotes the diffusion of CO₂ from the oceans to the atmosphere.

Some carbonate producers (e.g., coccolithophorids) are also phytoplankton. They photosynthesize organic matter, which tends to drive the dissolved carbon system in the opposite direction, toward lower CO₂ concentrations; hence these microorganisms have opposing effects on atmospheric CO₂ concentrations. However, the majority of phytoplankton do not produce carbonate skeletons. Thus, the overall effect of biological production in the surface ocean is in favor of reduced CO₂ concentrations. On average, the ratio of organic matter to carbonate mineral production for plankton is about 4:1.

When pelagic die, their shells or skeletons sink through the water column but are less subject to destruction during the trip to the sea floor than is their organic debris. In regions where the water depth is less than 4 km or so, carbonate particles accumulate more or less intact on the sea floor. These waters are said to be saturated with respect to CaCO₃. This shallower-water deposition of planktonic carbonate is part of the material that eventually becomes limestone. Other limestones form in very shallow tropical waters, where reefs and other carbonate-producing organisms live on the sea floor. Deeper waters, however, have higher concentrations of dissolved CO₂ (carbonic acid), due to the decomposition of organic matter (Box Figure 8-1). As a result, these waters are corrosive to CaCO₃; they are said to be undersaturated with respect to CaCO₃. The level at which the rate of dissolution of carbonate sediment balances the flux of carbonate settling through the water column is called the calcification boundary. Below this depth, carbonate materials, in the form of shells or skeletons, dissolve in transit as they settle through the water column and do not accumulate on the sea floor. Thus, the deep-ocean basins are devoid of the cover of plankton-deposited carbonate sediments that occupy the shallower parts of the ocean, for example, along the mid-ocean ridges (Figure 8-15). Interestingly, carbonate sediments deposited on mid-ocean ridges are carried slowly to depths below the CCD as the sea floor spreads away from the ridge, cools, and subducts (see Chapter 7). If it weren't for a protective layer of ooze, the carbonate sediments would dissolve as they were carried below the CCD by sea-floor spreading.

Summary of the Inorganic Carbon Cycle

In the past few sections, we have discussed several processes that affect the transfer of inorganic carbon. These processes can be summarized in a mass flow diagram for the inorganic carbon cycle (Figure 8-16). The free exchange of CO₂ between the atmosphere and the ocean
The Carbon-Silicate Geochemical Cycle

After inorganic carbon is involved in chemical weathering and carbonate mineral precipitation and is removed by sedimentary burial, plate tectonics provides the needed return flux of CO$_2$ in the form of metamorphic and volcanic CO$_2$ inputs to the atmosphere. Mantle-derived CO$_2$ is released to the ocean-atmosphere system at mid-ocean ridges and along convergent margins (Figure 8-17). This carbon is derived from the mantle and so is, in a sense, "new." At convergent plate margins (deep-sea trenches), some of the sediments resting on the downwelling slab are subducted along with the plate (see Chapter 7). The plate and its sediment cover to depths as great as hundreds of kilometers within the mantle, where high temperatures and pressures promote chemical reactions.
that transform the sediments into metamorphic rock. Among these reactions is the reaction between sedimentary carbonate minerals and silica-rich sediments that forms silicate minerals and releases CO₂.

Carbonate metamorphism:

\[
\text{CaCO}_3 + \text{SiO}_2 \rightarrow \text{CaSiO}_4 + \text{CO}_2
\]

This process is termed carbonate metamorphism. As before, we use the mineral wollastonite to represent the more complex silicate minerals that are typically generated by this process.

If sufficiently high temperatures are reached at depth during carbonate metamorphism, magnas are generated. These magnas may erupt in volcanoes at the surface, releasing CO₂ to the atmosphere. The CO₂ in these volcanoes probably includes some mantle-derived CO₂ and some CO₂ from the subducted crust and sediments, but scientists do not yet know the relative proportions of these two sources. Under metamorphic conditions the CO₂ produced can migrate as a fluid toward the surface. Although a substantial fraction of it reacts with minerals along the way, some CO₂ is released through springs and seeps to the atmosphere.

Together, silicate weathering, carbonate precipitation, and ocean-atmosphere exchange are the reverse of carbonate metamorphism. (Compare the equation for the net result of these processes, given in the previous section, with the carbonate metamorphism equation.) Without a fairly close balance between the inflows and outflows of carbon, the supply of this important greenhouse gas to the atmosphere and ocean would, on geological time scales, quickly be depleted. Earth would soon become a frozen ball of ice.

That may have been the fate of Mars, which appears to once have had flowing water on its surface and, perhaps, a CO₂-rich atmosphere with a stronger greenhouse effect. In contrast, on Earth the release of CO₂ after carbonate metamorphism and volcanism has essentially balanced the consumption of CO₂ during silicate weathering over the history of the planet. What has caused this balance? We cannot call on simple chemical equilibrium, because the reactions involved are representative of a whole host of processes rather than a single chemical reaction. Rather, we must look for feedback loops that, according to the amount of CO₂ in the atmosphere, adjust the rates of CO₂ input by volcanism or CO₂ removal by silicate weathering and thereby keep the reservoir in steady state.

Long-Term Feedbacks in the Carbonate–Silicate Cycle

Because it is driven largely by heat flow from the Earth's interior, the rate of volcanism is probably not very sensitive either to the amount of CO₂ in the atmosphere or to the climate of Earth's surface. In contrast, many climatic factors affect the rate of chemical weathering. The regulation of atmospheric CO₂ on long time scales (millions of years) is the consequence of the feedback between climate and rates of silicate weathering. The climatic factors that help regulate the chemical weathering rates of silicate rocks include the following:

- Temperature: rates of reactions, including chemical weathering, tend to increase as temperature increases.
- Net rainfall: weathering requires water as a medium both for the dissolution of minerals and for the transport of the dissolved material to the oceans, and thus weathering rates rise as precipitation increases.
- Temperature: rates of reactions, including chemical weathering, tend to increase as temperature increases.
- Net rainfall: weathering requires water as a medium both for the dissolution of minerals and for the transport of the dissolved material to the oceans, and thus weathering rates rise as precipitation increases.

These environmental factors are responsive to atmospheric CO₂ levels. Recall from Chapter 3 that, as a result of the greenhouse effect, global temperatures rise as the atmospheric content of CO₂ increases and that rates of evaporation increase with increasing temperature. We know from Chapter 4 that water that evaporates from the ocean must fall as precipitation. For these reasons, we would expect that a warmer world would be a wetter world. Net precipitation should increase as temperature increases. Thus, the silicate weathering rate should increase as the atmospheric CO₂ level rises. Figure 8-18 shows a feedback loop for these processes. On the other side of the feedback loop, increased silicate weathering rates tend to reduce atmospheric CO₂ levels because silicate weathering uses up carbonic acid.

A CLOSER LOOK

Biological Enhancement of Chemical Weathering

A particularly active and controversial area of research in geochemistry is the role of biological processes in accelerating chemical weathering. Scientists who study the lithosphere both at Earth's surface and in the laboratory have concluded that biologically mediated weathering is an important process on Earth, and that it could be an important contributor to the overall weathering rate on the planet.

Microbial decomposition and root metabolism are significant factors in increasing the rate of weathering. For example, the release of organic acids from plants can increase the rate of weathering by dissolving minerals more quickly. Similarly, the release of carbonic acid from plants can increase the rate of weathering by dissolving minerals more quickly. Similarly, the release of carbonic acid from plants can increase the rate of weathering by dissolving minerals more quickly.

Field experiments have shown that the presence of living organisms can significantly enhance the rate of weathering. For example, the presence of lichens on rock surfaces can significantly increase the rate of weathering. Similarly, the presence of fungi on rock surfaces can significantly increase the rate of weathering. Similarly, the presence of bacteria on rock surfaces can significantly increase the rate of weathering.

BOX FIGURE 8-3

[Box Figure 8-3: The lithosphere as a dynamic system. The interactions of the lithosphere, atmosphere, and biosphere are illustrated in this figure.

The Carbon–Silicate Geochemical Cycle

[Figure 8-17: Diagram of the carbon–silicate geochemical cycle.]

[Figure 8-18: Diagram of the feedback loop showing the balance between atmospheric CO₂ concentration and climate on long time scales.]

[Figure 8-19: Diagram of the biological enhancement of chemical weathering.]

[Figure 8-20: Diagram of the role of biological processes in accelerating chemical weathering.]

[Figure 8-21: Diagram of the interactions between the lithosphere, atmosphere, and biosphere.]

[Figure 8-22: Diagram of the feedback loop showing the balance between atmospheric CO₂ concentration and climate on long time scales.]
The overall feedback loop, as shown in Figure 8-18, is negative. The feedback tends to stabilize Earth’s climate against perturbations, as we will see later.

Links between the Organic and Inorganic Carbon Cycle

Although we have differentiated between the terrestrial and marine organic carbon cycles and between the organic and inorganic carbon cycles, all these cycles are intricately linked as parts of the global carbon cycle. Changes that occur on land rapidly affect the oceans through the transport of carbon and nutrients by rivers and through variations in atmospheric CO₂. Changes in the recycling of organic matter affect atmospheric and oceanic CO₂ and in turn the whole of the carbon cycle.

Thus, the cycle divisions we have made are artificial, but they help us represent a complicated system in simple terms.

Chapter Summary

1. The global carbon cycle involves processes that occur on land and in the oceans and involve both biological and non-biological chemical reactions.
2. The terrestrial and marine organic carbon cycles operate on a variety of time scales. On the short time scale (hours to hundreds of years):
   a. Carbon dioxide is removed from the atmosphere during photosynthesis on land and returned during respiration and decomposition; methane is released to the atmosphere from soils, where anaerobic metabolisms is taking place.
   b. A small amount of terrestrial organic carbon survives respiration and decomposition and is buried in sedimentary basins on land or is transported to the sea.
   c. In the oceans, phytoplankton produce organic carbon that is consumed by zooplankton and decomposed by aerobic and anaerobic bacteria.
   d. A small fraction of the organic matter settling through the water column is not decomposed and is instead buried in marine sediments.
3. On longer time scales (millions of years), the organic matter buried in sediments undergoes lithification with the sediment. Most of these sediments are muds, and the rocks formed are shales.
4. When concentrations of organic matter are very high in the sediment, fossil fuels may form during burial and lithification.
5. The sedimentary rocks may eventually undergo uplift through tectonic processes, exposure, and weathering. During weathering, organic matter undergoes oxidation, producing CO₂, which escapes to the atmosphere.
6. An inorganic carbon cycle, involving oxidized forms of carbon, is important on both short and long time scales (millions of years):
   a. Inorganic CO₂ exchanges with CO₂ dissolved in the surface ocean on a time scale of decades. The uptake of CO₂ is enhanced by reactions among the forms of dissolved inorganic carbon in seawater.
   b. Atmospheric CO₂ dissolves in rainwater, creating an acidic solution. When the rain falls on the land surface, reactions with carbonate and silicate minerals convert carbonic acid to bicarbonate ion. The bicarbonate ion is carried by rivers to the ocean.
   c. In the oceans, carbonate-accumulating organisms use the bicarbonate ion in the construction of their shells or skeletons. This material may dissolve in transit or it may become part of the sediment that covers the sea floor.
   d. The regulation of atmospheric CO₂ on long time scales (millions of years) is the consequence of the feedback between climate factors and rates of chemical weathering of silicate rocks, as part of the long-term inorganic carbon cycle, as well as to the carbonate-silicate geochemical cycle.
   e. Any disturbance in the amount of atmospheric CO₂ affects climate through the greenhouse effect. Changes in climate affect silicate weathering rates and the input of CO₂ into the atmosphere.
   f. The overall feedback loop is negative, implying that on long time scales the climate system is stable against a wide range of perturbations.
7. The terrestrial and marine organic carbon cycles, the organic and inorganic carbon cycles, and the carbonate-silicate geochemical cycle are intricately linked as parts of the global carbon cycle.
8. Changes on land are “communicated” rapidly to the oceans by rivers, transport of carbon and nutrients, and through variations in atmospheric CO₂.
9. Changes in the recycling of organic matter affect atmospheric and oceanic CO₂ and in turn the whole of the carbon cycle.

Key Terms

acid
base
biomass
bioreactor
carbonate
characteristic response time
carbonate metamorphism
consumers
inorganic carbon

Critical-Thinking Problems

1. Which of the following carbon reservoirs has the longest residence time: plants, the oceans, or sedimentary limestone?
2. One or more of the following processes involves organic carbon; identify them: the precipitation of a calcite skeleton, the exchange of carbon between the oceans and the atmosphere, dissolution of the sea floor, or oxidation during weathering?
3. Describe the biological pump.
4. Why is plate tectonics critical to the maintenance of an atmosphere-ocean reservoir rich in carbon?
5. Limestone (carbonate) weathering does not lead to the net removal of carbon dioxide from the atmosphere. Why not?

Review Questions

1. Which of the following carbon reservoirs has the longest residence time: plants, the oceans, or sedimentary limestone?
2. One or more of the following processes involves organic carbon; identify them: the precipitation of a calcite skeleton, the exchange of carbon between the oceans and the atmosphere, dissolution of the sea floor, or oxidation during weathering?
3. Describe the biological pump.
4. Why is plate tectonics critical to the maintenance of an atmosphere-ocean reservoir rich in carbon?
5. Limestone (carbonate) weathering does not lead to the net removal of carbon dioxide from the atmosphere. Why not?

Critical-Thinking Problems

1. The key to stability feedback between the reservoir and the fitness in and out of the reservoir. Assume that the rate of outflow from a reservoir depends on the size of the reservoir according to the following relationship: outflow rate = k x (size of reservoir), where k is a constant.
   a. Reservoir of water has a volume of 5000 liters, and the rate of outflow at steady state is 25 liters per minute. What is k? (Give both the numerical value and its units.)
   b. What is the residence time? What is the relationship between k and the residence time?
   c. The inflow rate is 25 liters per minute. Describe graphically and in words how the reservoir size would change with time, beginning with a reservoir size of zero and continuing until the reservoir reaches steady state.
   d. Use Figure 8-4 to answer the following questions:
      a. During which months is the rate of photosynthesis greatest, relative to the combined rate of respiration and decomposition, and during which months is it smallest?
      b. On the basis of your answer to part (a), estimate, for each year, the maximum net rate of photosynthesis and the maximum net rate of respiration/decomposition for each of the three years shown.
5. Are there significant differences in these rates from year to year? If so, for which years?
5. A giant meteor crashes into Earth, causing devastating environmental changes that kill off all life in the oceans.
   a. Describe how the vertical distribution of dissolved oxygen, carbon, and nutrients would respond.
   b. Would the temperature and salinity of the ocean be affected by the loss of the biological pump? Why or why not?
   c. If global warming from CO₂ released to the atmosphere from the meteor impact site caused, instead of the complete loss of marine life, the sudden cessation of thermohaline circulation in the oceans, what would be the effect on the vertical and spatial distribution of dissolved nutrients, carbon, and oxygen in the world’s oceans?
   d. The atmosphere consists of 78% N₂, 21% O₂, 1% Ar, and about 0.003% CO₂. What is the mean molecular weight of air? Round your answer to three significant figures, and use the following table of atomic weights:

<table>
<thead>
<tr>
<th>Element</th>
<th>Atomic Weight</th>
</tr>
</thead>
<tbody>
<tr>
<td>C (Carbon)</td>
<td>12.011</td>
</tr>
<tr>
<td>N (Nitrogen)</td>
<td>14.0067</td>
</tr>
<tr>
<td>O (Oxygen)</td>
<td>15.994</td>
</tr>
<tr>
<td>Ar (Argon)</td>
<td>39.948</td>
</tr>
</tbody>
</table>

The total mass of the atmosphere is about 5 x 10¹⁵ kg. How many moles each of CO₂ and O₂ are present in the atmosphere? (Note: Calculate the total two answers from the first one rather than by computing the masses of O₂ and CO₂. The values listed in part (a) for the various gases are abundances by volume, not by mass. This fact, and the fact that a mole of any gas takes up the same volume as a given gram-mole, is usually true, and is the basis on which this calculation is done.)

| Forests contain about 400 Gt (giga-tonnes) in the form of wood and leaves. Suppose that all the wood in the world were burned down instantaneously. By how much would atmospheric CO₂ increase? By how much would O₂ decrease? Express your answers in percentages. Assume that the equation for burning is the same as that for respiration (given earlier in this chapter).
5. Explain why lakes and rivers have slightly basic pH values, whereas rainwater (the ultimate source of water for lakes and rivers) is slightly acidic.
Further Reading

General


Advanced


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Focus on the Biota: Metabolism, Ecosystems, and Biodiversity

Key Questions

- What are the characteristics of life on Earth that allow it to interact with physical processes at the global scale in such a significant way that it creates a habitable planet?
- How is the biosphere structured?
- How is energy transferred within the biosphere?
- What is an ecosystem?
- What is biodiversity and how is it measured?
- How is the diversity of interactions between the biota and the physical world related to the stability of the Earth system?

Chapter Overview

In this chapter we highlight the role that life plays in the operation of the Earth system. We begin with a general discussion of life and its unique characteristics, and then explore the varied metabolic pathways different forms of life take to grow and reproduce. Organisms interact at a variety of scales, so we find that populations of organisms group into communities, which at a larger scale interact among themselves and with their environment in ecosystems. The level of diversity of ecosystems can be expressed in a variety of ways. We can simply count the number of species or we can take into account the more complex diversity of interactions that take place between species and between organisms and their environment. This diversity of interactions, a defining characteristic of life on Earth, is important in our understanding of the feedbacks between the biota and the physical world that create a habitable planet, and helps us further understand the complexity of the Earth system.

Life on Earth

Characteristics of Life

Earth is unique among the planets in our solar system in that it apparently is the only one to support life. Earth more than supports life, it fosters it. Life is involved in almost every process occurring at the surface of the planet. Some fundamental characteristics of life allow it to have such an influence.

- Life spreads exponentially. The rate of population growth depends on the number of individuals reproducing at a particular time. This characteristic leads to the phenomenon of exponential growth. If left unchecked, 2 individuals become 4 in one generation, 4 become 8 in two generations, 8 become 16 in three generations, and 16 become 32 in four generations. In nature, however, exponential growth ceases as resources become limiting.
- Life needs energy. Photosynthesizers use solar energy, chemosynthesizers use chemical energy, and most other