CHAPTER 5: ATMOSPHERIC STRUCTURE AND RADIATION TRANSFER

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The atmosphere. Thunderstorm anvil heads and other clouds extending into the lower atmosphere over Matanitu ko Viti – Republic of the Fiji Islands and the South Pacific Ocean. *(Source: NASA)*

STUDENT LEARNING OUTCOMES

After reading this chapter you should be able to:

- Describe the chemical evolution and current composition of the Earth's atmosphere.
- Define the various layers in the atmosphere based on vertical changes in temperature and chemistry.
- Explain how surface albedo and the atmospheric processes of scattering, absorption, and reflection influence the amount of insolation received at the Earth's surface.
- Discuss the concept of radiation balance and illustrate the patterns of longwave emissions from our planet's surface and atmosphere.
- Portray the patterns of net radiation that exist on our planet's surface.
- Outline how the greenhouse effect aids in the heating the Earth's surface and lower atmosphere.
- Describe how humans are altering the greenhouse effect.

EVOLUTION OF THE EARTH'S ATMOSPHERE

Scientists believe the Earth came into being about 4.6 billion years ago. The Earth formed as gas, particles of interstellar dust, and ice lumped together to form a continually larger mass until about 200 million years had passed. At the end of this period, the growth of the Earth's mass was pretty much complete. Initially, the surface temperature was quite hot having an estimate temperature of about 8000°C (14,400°F). By about 4.4 billion years ago, the surface had cooled enough for the solidification of magma into solid crust. This cooling also resulted in an initial release of gases from the lithosphere, mainly hydrogen and helium. These gases were hot and as a result had high molecular velocities, which enabled them to overcome gravity and travel into space.

Table 5.1 below describes the three major stages of development of the atmosphere. Most of the early atmosphere was created in the first one million years after the solidification of the crust (4.4 billion years ago). Its surface temperature during this period was about 85 to 110°C (185 to 230°F). The main gases of this early atmosphere consisted of carbon dioxide (10-15%), nitrogen (8-10%), and water vapor (60-70%). This mixture had a density that was 10 to 20 times greater than today's atmosphere. This higher density was mostly due to large amounts of evaporated water.

The second stage (secondary atmosphere) in the evolution of the Earth's atmosphere occurred between 4.0 and 3.3 billion years ago. By about 3.8 billion years ago, atmospheric temperatures were cool enough to allow for the condensation of water vapor. Rain events were numerous, heavy, and geographically widespread. At first, the rain evaporated before it hit the ground because of high temperatures in the lower atmosphere. Eventually, the lower atmosphere and ground surface became cool enough to allow for surface accumulations of water forming rivers, lakes, and oceans. The volume of water in the world's oceans leveled off about 3.5 billion years ago. The falling rain also washed out large amounts of carbon dioxide from the air. At the ground surface, this carbon dioxide chemically combined with calcium and magnesium to form sedimentary rocks. Oxygen began accumulating in the atmosphere through the chemical breakdown of water by sunlight (known as **photo-dissociation**), releasing molecular oxygen (O_2), and by way of **photosynthesis**. The emergence of living organisms around 3.6 to 4.0 billion years ago was extremely important in the creation of atmospheric oxygen (O_2) and ozone (O_3).

The last period of atmospheric evolution is known as the living atmosphere. Starting at about 3.3 billion years ago, life began to significantly influence the chemical composition of the atmosphere. Through photosynthesis, organisms produced large amounts of oxygen and this gas began to accumulate at greater concentrations. Most of the build-up of oxygen in the atmosphere occurred between 2.1 and 1.5 billion years ago as a direct result of photosynthesis from ocean-based plants like algae. By about 450 million years ago, there was enough oxygen in the atmosphere to allow for the development of a stratospheric ozone layer that was thick enough to protect terrestrial life from ultraviolet radiation. As a result, evolutionary development and the expansion of terrestrial life began at this time. The composition of the atmosphere remained relatively fixed for most of the last 500 million years. During the last three hundred years, the concentrations of several atmospheric gases have changed significantly because of the activities of humans. We will examine this form of environmental change in greater depth later on in the next section.

COMPOSITION OF THE ATMOSPHERE

GASES

Table 5.2 lists the twelve most common gases found in the Earth's lower atmosphere. Of the gases listed, nitrogen, oxygen, carbon dioxide, methane, nitrous oxide, and ozone are extremely important to the health of the Earth's biosphere. The table indicates that nitrogen and oxygen are the main components of the dry atmosphere by volume. Together, these two gases make up approximately 99% of the dry atmosphere (**Figure 5.1**). Both of these gases have very important associations with life. Nitrogen is removed from the atmosphere and deposited at the Earth's surface mainly by specialized bacteria that can fix nitrogen gas and by way of lightning through precipitation. This addition of nitrogen to the Earth's surface soils and various water bodies supplies much needed nutrition for plant growth. Nitrogen returns to the atmosphere primarily through the natural processes of biomass burning and denitrification. **Denitrification** is a chemical process where solid forms of nitrogen are converted into nitrogen gases (N_2 and N_2O) by bacterial action.

Oxygen is exchanged between the atmosphere and biosphere through the processes of photosynthesis and

TABLE 5.1 Evolution of the Earth's atmosphere.

Name of Stage	Duration of Stage (Billions of Years Ago)	Main Constituents of the Atmosphere	Dominant Processes and Features
Early Atmosphere	4.4 to 4.0	Water vapor (H_2O), hydrogen cyanide (HCN), ammonia (NH_3), methane (CH_4), sulfur, iodine, bromine, chlorine, and argon.	Lighter gases like hydrogen and helium escaped to space. All water was held in the atmosphere as vapor because of high temperatures.
Secondary Atmosphere	4.0 to 3.3	At 4.0 billion H_2O, carbon dioxide (CO_2), and nitrogen (N_2) dominant. By 3.0 billion CO_2, H_2O, N_2 dominant.	Molecular oxygen (O_2) begins to accumulate. Continued release of gases from the lithosphere. Water vapor clouds common in the lower atmosphere. Cooling of the atmosphere causes precipitation and the development of the oceans. Chemosynthetic bacteria appear on the Earth at 3.6 billion years and life begins to modify the atmosphere.
Living Atmosphere	3.3 to Present	N_2 - 78%, O_2 - 21%, Argon - 0.9%, CO_2 - 0.036%	Development, evolution, and growth of life increases the quantity of oxygen in the atmosphere from <1% to 21%. 500 million years ago concentration of atmospheric oxygen levels off at about 21%. Humans begin modifying the concentrations of some gases in the atmosphere beginning around the year 1700.

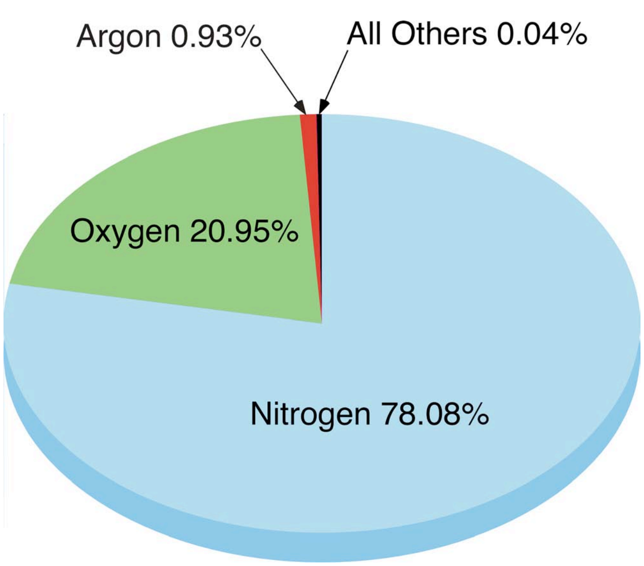

FIGURE 5.1 Proportional concentration of the gases found in the lower atmosphere. By volume, nitrogen and oxygen are the major components. *(Image Copyright: Michael Pidwirny)*

TABLE 5.2 Average composition of the dry atmosphere from the surface to a height of about 11 km (6.8 mi). Variable gases are constituents whose volume in the atmosphere has changed significantly over the last 300 years. This change has been mainly due to human activities.

Gas Name	Chemical Formula	Percent by Volume	Variable Gas
Nitrogen	N_2	78.08%	no
Oxygen	O_2	20.95%	no
Argon	Ar	0.93%	no
Carbon Dioxide	CO_2	0.0376%	yes
Neon	Ne	0.0018%	no
Helium	He	0.00052%	no
Methane	CH_4	0.00017%	yes
Hydrogen	H_2	0.00006%	no
Nitrous Oxide	N_2O	0.00003%	yes
Xenon	Xe	0.000009%	no
Ozone	O_3	0.000004%	yes

respiration. Photosynthesis produces oxygen when carbon dioxide and water are chemically converted into glucose in chloroplasts with the help of sunlight. **Respiration** is the reverse of photosynthesis. In respiration, oxygen is combined with glucose to chemically release energy for metabolism. The products of this reaction are water and carbon dioxide. Respiration and photosynthesis work in tandem to keep oxygen levels in the atmosphere at a static equilibrium.

The fourth most abundant gas in the atmosphere is carbon dioxide. Carbon dioxide is exchanged between the atmosphere and life through the processes of photosynthesis and respiration. The volume of this **greenhouse gas** has increased by about 38% in the last three hundred years (**Figure 5.2**). This increase is primarily due to human induced burning of **fossil fuels**, **deforestation**, and other forms of land-use change. Most scientists believe that this increase is causing an enhancement of the greenhouse effect and global warming. The **greenhouse effect** is a natural process that provides additional heat energy to the Earth's ground surface and lower atmosphere. This additional heat is produced when the various greenhouse gases absorb and re-emit outgoing **longwave radiation** from the Earth.

Prior to 1700, levels of carbon dioxide were about 280 ppm (parts per million). Concentrations of carbon dioxide in the atmosphere are now about 385 ppm (**Figure 5.2**). This increase in carbon dioxide in the atmosphere is primarily due to the activities of humans. Societal changes, brought about by the **Industrial Revolution**, increased the amount of carbon dioxide entering the atmosphere. Emissions from the combustion of fossil fuels account for

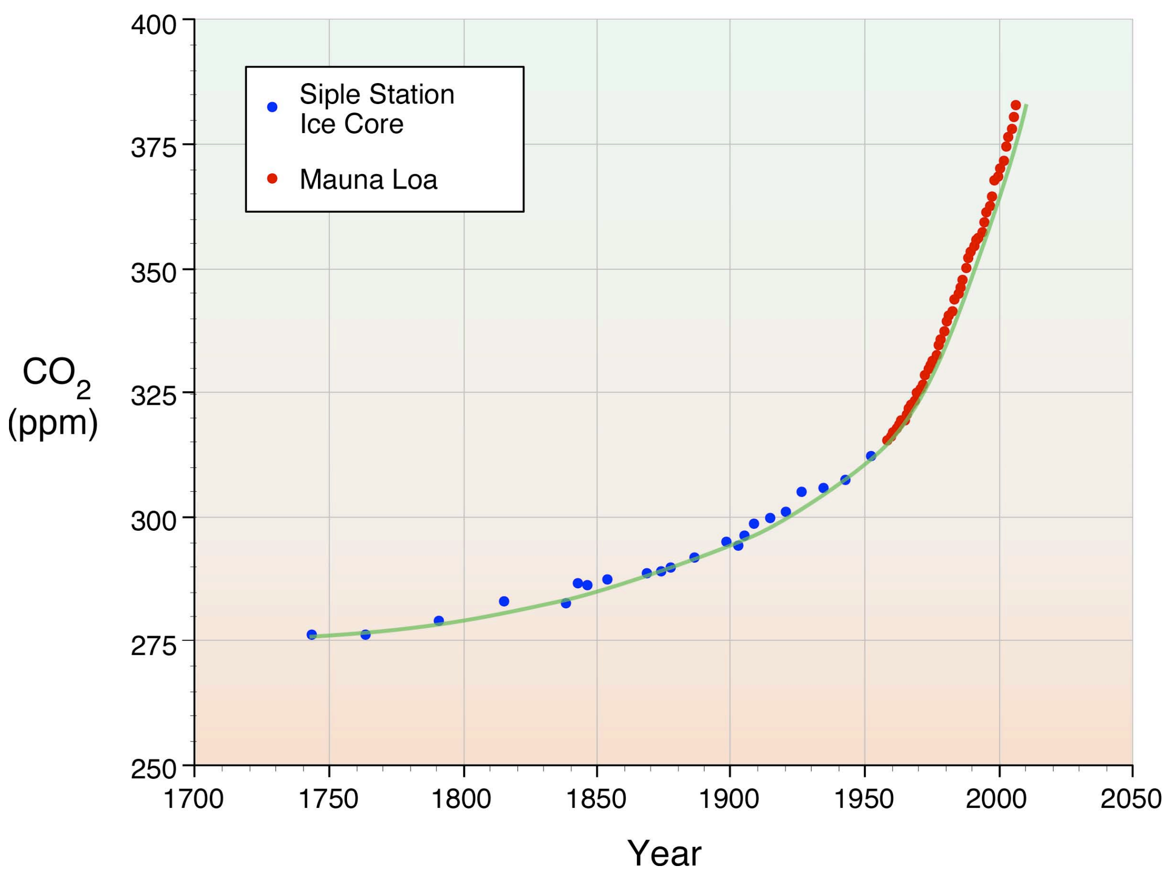

FIGURE 5.2 The following graph illustrates the rise in atmospheric carbon dioxide from 1744 to 2006. Note that the increase in carbon dioxide's concentration in the atmosphere is exponential in nature. An extrapolation into the immediate future would suggest continued annual increases. (*Source: Neftel, A., H. Friedli, E. Moore, H. Lotscher, H. Oeschger, U. Siegenthaler, and B. Stauffer. 1994. Historical carbon dioxide record from the Siple Station ice core. pp. 11-14. In T.A. Boden, D.P. Kaiser, R.J. Sepanski, and F.W. Stoss (eds.) Trends'93: A Compendium of Data on Global Change. ORNL/CDIAC-65. Carbon Dioxide Information Analysis Center, Oak Ridge National Laboratory, Oak Ridge, Tenn. U.S.A. and C.D. Keeling and T.P. Whorf. 2001. Carbon Dioxide Research Group, Scripps Institution of Oceanography, University of California, La Jolla, California 92093-0444, U.S.A.)*

about 65% of the additional carbon dioxide found in the atmosphere. The remaining 35% is derived from deforestation and the conversion of prairie, woodland, and forested ecosystems primarily into less productive agricultural systems. Natural ecosystems can store 20 to 100 times more carbon dioxide per unit area than agricultural systems. Both deforestation and natural land-use change reduce the amount of standing plant mass or biomass found on the Earth's surface. This reduction causes a net export of carbon stored in biomass into the atmosphere through decomposition and burning.

Since 1750, atmospheric concentrations of the greenhouse gas methane have increased more than 150% (**Figure 5.3**). The primary sources for the additional methane added to the atmosphere (in order of importance) are rice cultivation, domestic grazing animals, termites, landfills, oil and gas extraction, and coal mining. **Anaerobic** conditions associated with rice paddy flooding results in the formation of methane gas. An accurate estimate of how much methane is being produced from rice paddies has been difficult to determine. More than 60% of all rice paddies are found in India and China where scientific data concerning emission rates are hard to obtain. Nevertheless, scientists believe that the contribution of rice paddies is large because this type of crop production has more than doubled since 1950. Grazing animals release methane to the environment as a result of herbaceous digestion. Some researchers believe the addition of methane from this source has more than quadrupled over the last century. Termites also release methane through the

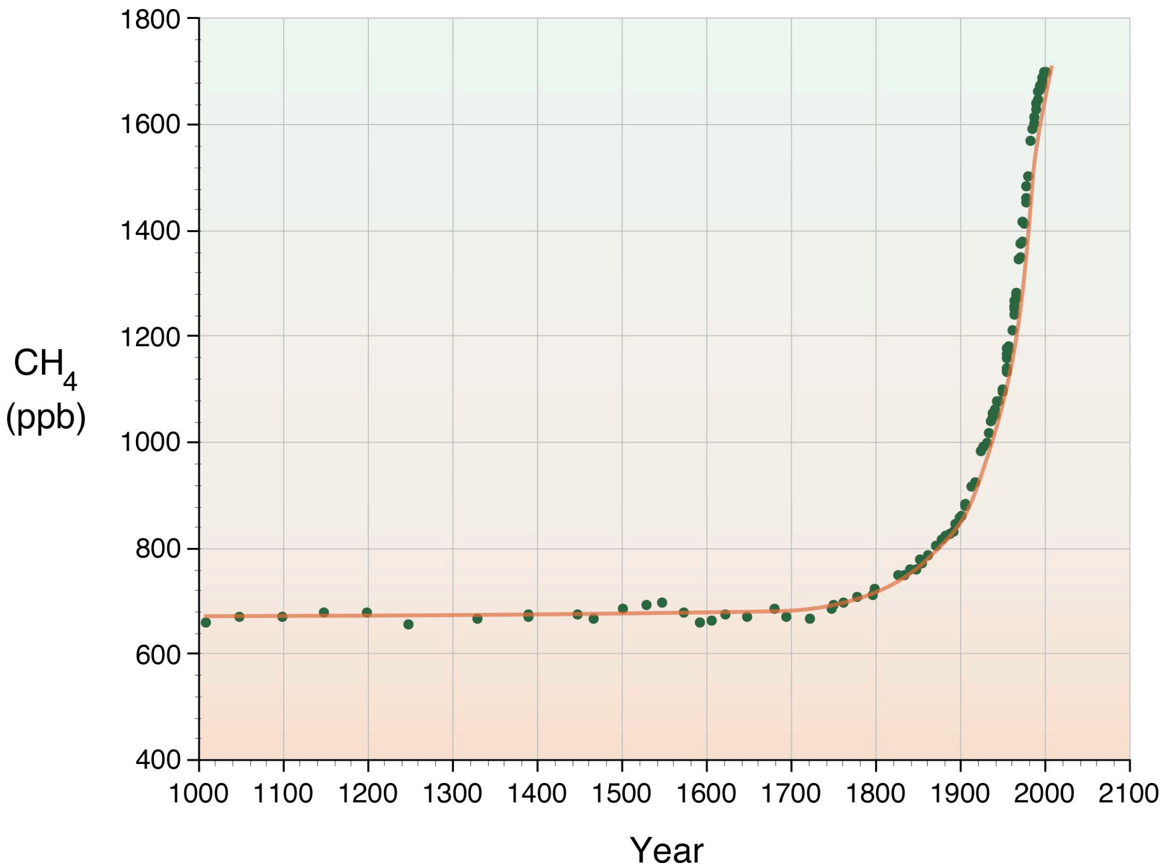

FIGURE 5.3 The following graph illustrates the rise in atmospheric methane from 1008 to 2001. Note that the increase in methane's concentration in the atmosphere is exponential in nature. An extrapolation into the immediate future would suggest continued annual increases. *(Source: D.M. Etheridge, L.P. Steele, R.J. Francey, and R.L. Langenfelds. 2002. Historical CH_4 Records Since About 1000 A.D. From Ice Core Data. In Trends: A Compendium of Data on Global Change. Carbon Dioxide Information Analysis Center, Oak Ridge National Laboratory, U.S. Department of Energy, Oak Ridge, Tenn., U.S.A and Steele, L. P., P. B. Krummel and R. L. Langenfelds. 2002. Atmospheric CH_4 concentrations from sites in the CSIRO Atmospheric Research GASLAB air sampling network (October 2002 version). In Trends: A Compendium of Data on Global Change, Carbon Dioxide Information Analysis Center, Oak Ridge National Laboratory, U.S. Department of Energy, Oak Ridge, TN, U.S.A.)*

digestion of plant material. Land-use change in the tropics, due to deforestation, ranching, and farming, may be causing termite numbers to expand. These types of land conversion produce an environment favorable for termites. Landfills produce methane as organic wastes decompose. Coal, oil, and natural gas deposits release methane to the atmosphere when these deposits are excavated or drilled.

The average concentration of nitrous oxide is now increasing at a rate of 0.2 to 0.3% per year. Nitrous oxide is another greenhouse gas. Its part in the enhancement of the greenhouse effect is minor relative to the other greenhouse gases already discussed. Nitrous oxide also contributes to the artificial fertilization of ecosystems. In extreme cases, this fertilization can lead to the death of forests, eutrophication of aquatic habitats, and species die-offs. Sources for the increase of nitrous oxide in the atmosphere include land-use conversion, fossil fuel combustion, biomass burning, and soil fertilization. Most of the nitrous oxide added to the atmosphere each year comes from deforestation and the conversion of forest, savanna, and grassland ecosystems into agricultural fields and rangeland. Both of these processes reduce the amount of nitrogen stored in living vegetation and soil through the decomposition of organic matter. Nitrous oxide is also released into the atmosphere when fossil fuels and biomass are burned. The use of nitrate and ammonium fertilizers to enhance plant growth is another important source of nitrous oxide. How much is released from this process has been difficult to quantify. Some estimates suggest that fertilizers may contribute as much as 50% of nitrous oxide added to the atmosphere annually.

Concentrations of ozone gas are found in two different regions of the Earth's atmosphere. The majority of the ozone (about 97%) found in the atmosphere is concentrated in the stratosphere at an altitude of 10 to 50 km (6 to 31 mi) above the Earth's surface. This stratospheric ozone provides an important service to life on the Earth as it absorbs harmful ultraviolet radiation. In recent years, levels of stratospheric ozone have been decreasing due to the buildup of artificially created chlorofluorocarbons (CFCs) in the atmosphere. Since the late 1970s, scientists have noticed a seasonal depletion of ozone in the stratosphere over Antarctica. Satellite measurements have also indicated that the zone from 65°N to 65°S latitude has seen a 3 to 4% decrease in stratospheric ozone since 1978. Ozone also is highly concentrated at the Earth's surface in and around cities. Most of this ozone is created as a by-product of human created photochemical smog. This buildup of ozone is toxic to organisms.

Water vapor was excluded from the gases listed in **Table 5.2** because its concentration varies dramatically in the lower atmosphere both spatially and temporally. Two main reasons for these variations are: warmer air has more internal heat energy increasing the evaporation of liquid water into vapor, and air located over water bodies is supplied with more moisture because of evaporation than air over relatively dry continental masses. Consequently, the highest concentrations of water vapor are found near the equator over the oceans and the humid tropical rain forests. Cold polar areas and subtropical continental deserts are locations where the volume of water vapor can approach zero percent. Water vapor has several very important environmental functions:

- It redistributes heat energy on the Earth through latent heat energy exchange.
- The condensation of water vapor creates precipitation that falls to the Earth's surface providing needed fresh water for plant and animal consumption.
- It helps warm the Earth's atmosphere through the greenhouse effect.

Aerosols

The atmosphere contains a variety of liquid droplets and solid particles that are together known as **aerosols**. Both natural and human processes produce these substances. Water droplets and ice crystals are natural aerosols that are quite familiar to us. We see large accumulations of these particles almost every day in clouds. Other types of natural aerosols include wind-blown soil particles, volcanic dust, pollen, smoke and soot particles from wildfires, and salts from sea spray (**Figure 5.4**). Aerosols of human origin mainly consist of dust hurled into the atmosphere by vehicles and agricultural activities, smoke and soot from the burning of vegetation, and emissions from the combustion of fossil fuels.

The size of aerosols ranges from readily visible particles to specks of matter as small as 0.2 μm (micrometers). Smaller aerosols can remain floating in the atmosphere for long periods of time by the slightest vertical motions of air. Larger particles tend to fall out of the atmosphere in hours or minutes after being lifted. Most aerosols are suspended in the atmosphere for a few days to several weeks. Precipitation is extremely efficient at cleaning the atmosphere of particles in two ways. Aerosols

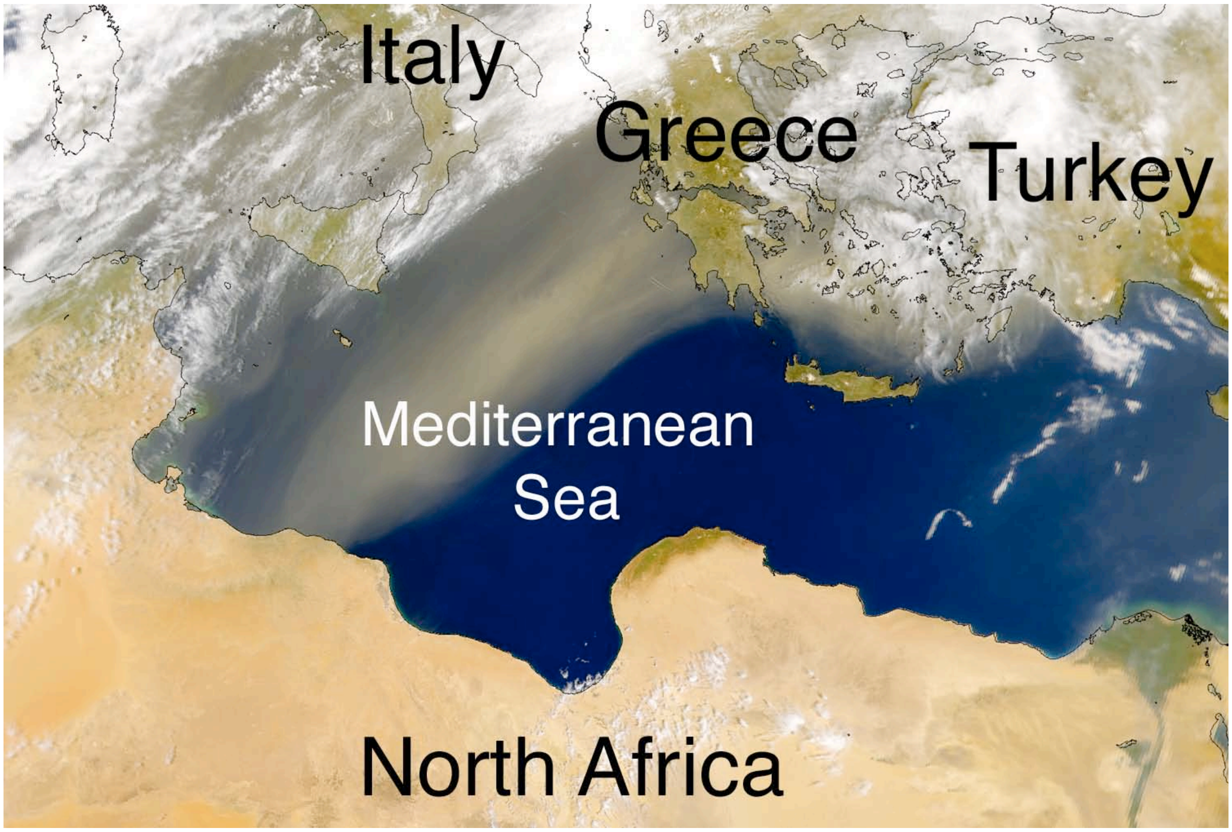

FIGURE 5.4 Movement of large quantities of dust from Northern Africa to southern Europe on March 26, 2001. This image obtained by the SeaWiFS sensor aboard the OrbView-2 satellite shows that Greece is receiving some of the heaviest concentrations of that dust. *(Source: NASA - Earth Observatory)*

act as the nucleus of developing raindrops and snow crystals. Secondly, falling rain droplets and snow crystals capture many particles by collision during their path to the ground surface.

Concentrations of aerosols vary in the atmosphere both spatially and temporally. Some temporal variations are related to daily or seasonal variations in wind speed. On a daily basis, wind normally has a lower velocity during the night. Maximum wind speeds often occur in the afternoon. Spatially, aerosols are most numerous over cities, seacoasts, and over areas of non-vegetated soil. The barren deserts of the world produce large quantities of wind-blown particles (**Figure 5.4**). Concentrations also tend to be greater in the lower atmosphere because the density of air increases as you approach the Earth's ground surface. Higher air densities make aerosols more buoyant.

ATMOSPHERIC LAYERS

Our planet's atmosphere has no clear-cut upper edge. The air just gradually becomes less and less dense with greater height. We can suggest that the atmosphere has a number of layers. These layers can be defined according to vertical changes in air temperature or chemical composition.

THERMAL LAYERS

Figure 5.5 displays the thermal layers found in the atmosphere including four main layers and three transition zones. The first identifiable layer above the Earth's surface is the **troposphere**. The depth of this layer varies considerably with latitude and season. The troposphere is about 16 km (10 mi) thick at the equator. Above the Earth's polar regions, the troposphere reaches an altitude of only 8 km (5 mi). The average thickness of this layer from the surface of the Earth is approximately 11 km (6.8 mi). The troposphere also is thicker during the summer than the winter. During the warm summer season the thermal expansion of the lower atmosphere and the presence of dominant warm updrafts push the upper boundary of this layer to a higher altitude.

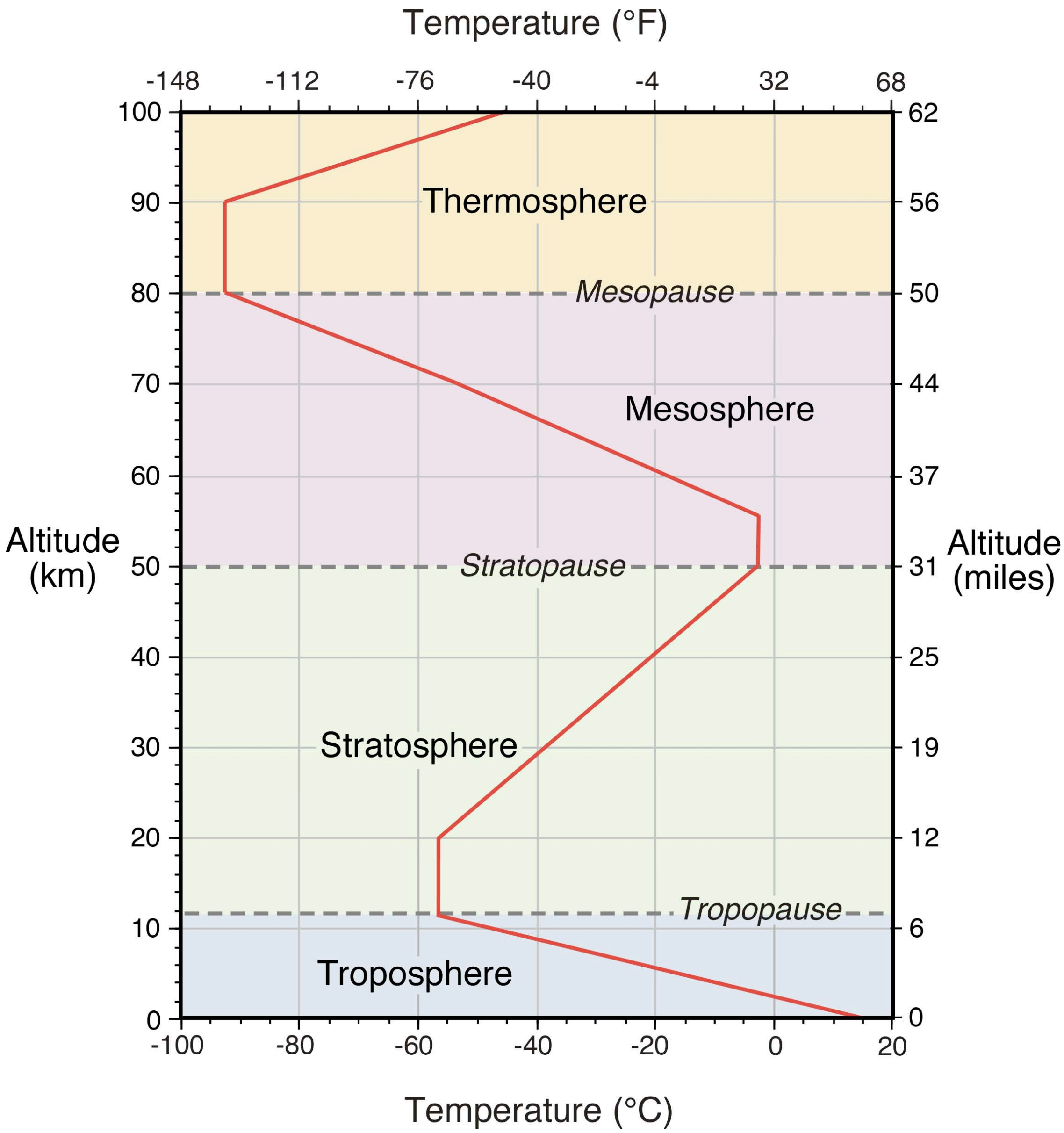

FIGURE 5.5 Structure of the atmosphere as defined by changes in air temperature. Variations in the way temperature changes with altitude indicate the atmosphere is composed of a number of different layers (labeled above). These variations are due to alterations in the chemical and physical nature of the atmosphere with altitude. *(Image Copyright: Michael Pidwirny)*

About 80% of the total mass of the atmosphere is contained in the dense troposphere. The troposphere is also the layer where the majority of our weather takes place. The circulation of air in this zone occurs vertically as well as horizontally. Vertical mixing of the troposphere is most evident during the summer season when intense solar radiation generates convective thermals near the Earth's surface. Some of these buoyant parcels of air can generate **cumulus** or **cumulonimbus** (thunderstorm) clouds under the right conditions (**Figure 5.6**).

Air temperature generally decreases with altitude in the troposphere. The average global temperature near the ground is about 15°C (59°F). With increasing height, air temperature drops uniformly with altitude at an average rate of 6.5°C per 1000 m (3.6°F per 1000 ft). An average temperature of -57°C (-71°F) is reached at the top of the troposphere. Actual rates of tropospheric temperature change vary with altitude, location, and time of year. Each of these rates is referred to as an **environmental lapse rate** (ELR). Meteorologists measure the environmental lapse

FIGURE 5.6 Thunderstorm cloud reaching the top of the troposphere. Summer heating of humid air in the lower troposphere can produce cumulonimbus or thunderstorm clouds. Vertical development of these storm clouds is usually stops at the top of the troposphere. This cumulonimbus cloud was photographed over Africa by an Expedition 16 crew member on the International Space Station. The flat top of the cloud indicates the top of the troposphere. *(Source: NASA)*

rate for various locations around the world twice daily. These measurements are used to forecast future surface air temperatures and to determine the likelihood of thunderstorm formation and other weather conditions. Overnight cooling near the ground, often produces an environmental lapse rate where temperature increases with height for several hundred meters from the surface before beginning to cool again. This occurrence is known as a temperature inversion (**Figure 5.7**). **Temperature inversions** inhibit the upward movement of air currents and can cause a concentration of air pollutants near the Earth's surface.

Near the top of the troposphere, lies a transition zone where air temperature does not change with altitude for a distance of about 9 km (5.6 mi). Such zones in the atmosphere are called **isothermal layers**. This first isothermal layer is called the **tropopause**. The tropopause is also the atmospheric layer where fast moving horizontal streams of meandering air called jet streams occur.

The next thermal layer in the atmosphere is called the **stratosphere**. This layer contains about 20% of the total mass of the atmosphere and extends to an altitude of 20 to

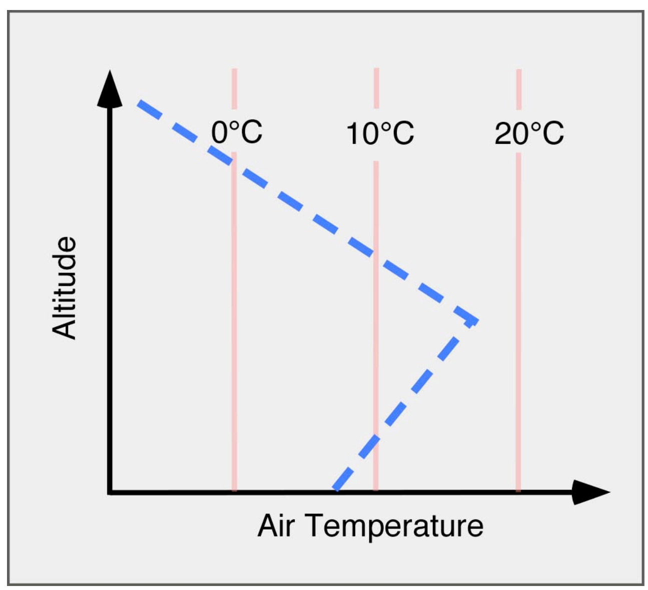

FIGURE 5.7 Temperature inversion vertical profile. A temperature inversion is a layer in the troposphere where temperature increases with height. *(Image Copyright: Michael Pidwirny)*

50 km (12 to 30 mi) above the Earth's surface. Minimal weather and vertical mixing occur in this layer of the atmosphere. Occasionally, the top of a thunderstorm may breach the bottom of this layer. In the stratosphere, temperature increases with altitude because a localized concentration of **ozone** gas molecules (tri-atomic oxygen - 03) absorbs ultraviolet radiation from the Sun creating heat energy. Stratospheric temperatures range from –57°C (-70°F) to 0°C (32°F) at the layer's upper boundary. Separating the next thermal layer in the atmosphere from the stratosphere is another isothermal layer called the **stratopause**.

On top of the stratosphere is the **mesosphere**. The mesosphere extends from 50 to 90 km (30 to 56 mi) above the Earth's surface. The air here is very thin and can get extremely cold. The top of this layer has an average temperature of about -90°C (-130°F). Above the mesosphere is the last isothermal layer called the **mesopause**.

The last atmospheric layer has an altitude greater than 90 km (56 mi) and is called the **thermosphere**. Temperatures in this layer can be as high as 1200°C (2200°F). These high temperatures are generated from the absorption of intense solar radiation by oxygen molecules (O_2). While these temperatures seem extreme, the amount of heat energy in the low-density air is very small. In Chapter 3, we learned that the amount of heat stored in a substance is controlled in part by its mass. The air in the thermosphere is extremely thin with individual gas molecules being separated from each other by large distances. Consequently, measuring the temperature of the thermosphere with a thermometer is a very difficult process. Thermometers measure the temperature of bodies via the movement of heat energy. Normally, this process takes a few minutes for the conductive transfer of kinetic energy from countless molecules in the body of a substance to the expanding liquid inside the thermometer. In the thermosphere, our thermometer would lose more heat energy from radiative emission then what it would gain from making occasional contact with extremely hot gas molecules.

Chemical Layers

We can also organize the atmosphere into a variety of layers based on chemical characteristics. From the Earth's surface to an altitude of 50 km (31 mi), the atmosphere has a fairly uniform mixture of the principal gases: nitrogen, oxygen, argon, and carbon dioxide. This zone of homogeneous composition is known as the **homosphere** (**Figure 5.8**). Above the homosphere is a layer called the heterosphere. In the **heterosphere**, the gases nitrogen, oxygen, helium, and hydrogen are concentrated at distinct altitudes within this layer. The order sequence of these gases with height is controlled by their atomic weight. The heaviest of these gases is nitrogen and it is found concentrated at the bottom of this layer. The next two sub-layers are occupied by oxygen and helium, respectively. Positioned at the top of the heterosphere is a sub-layer of hydrogen. Hydrogen is the lightest gas of this group.

We can also identify two other layers in the atmosphere based on chemistry. Between the altitudes of 10 to 50 km (6 to 31 mi) exists a layer in the atmosphere where ozone gas is concentrated. This zone is called the **ozone layer** (**Figure 5.9**). Within this layer, ozone reaches its highest concentrations at about 25 km (15 mi) above the surface. The ozone layer is important to organisms at the Earth's surface as it protects them from the harmful effects of the Sun's ultraviolet radiation. Without the ozone layer life, as we know it, could not exist on the Earth's surface.

The **ionosphere** occurs at a height between 60 and 400 km (40 to 250 mi) (**Figure 5.9**). In this relatively thick layer there is a concentration of ions. In the ionosphere, ions are positively charged because of the energizing effects of solar radiation on gas atoms and molecules. This same process also creates an abundance of free electrons. We use the electrically charged ionosphere to help transport radio waves. Certain layers of the ionosphere have the ability to reflect radio waves. By bouncing radio waves off these atmospheric regions we are able to extend transmissions over hundreds of kilometers (**Figure 5.10**). This process works best at night because of a unique property of the ionosphere. The ionosphere is divided into three sub-layers: the D-, E-, and F-layers. The lower D- and E-layers differ from the higher F-layer in two ways. First, these two sub-layers only exist during daylight hours. Second, they also have the ability to absorb some of the radio transmission. This absorption weakens the radio transmission requiring that radio stations increase signal strength after sunrise.

The ionosphere also has a role in the creation of the **Aurora Borealis** (Northern Lights) and **Aurora Australis** (Southern Lights). The aurora process begins with the release of clouds of sub-atomic particles (electrons and protons) from solar flares on the Sun into space. Some of the Sun's subatomic particles are intercepted and captured by the Earth's magnetic field in outer space. The magnetic field then redirects these solar particles, sending them off to the magnetic poles. At the Earth's magnetic poles, the electrons enter the ionosphere where the energy they

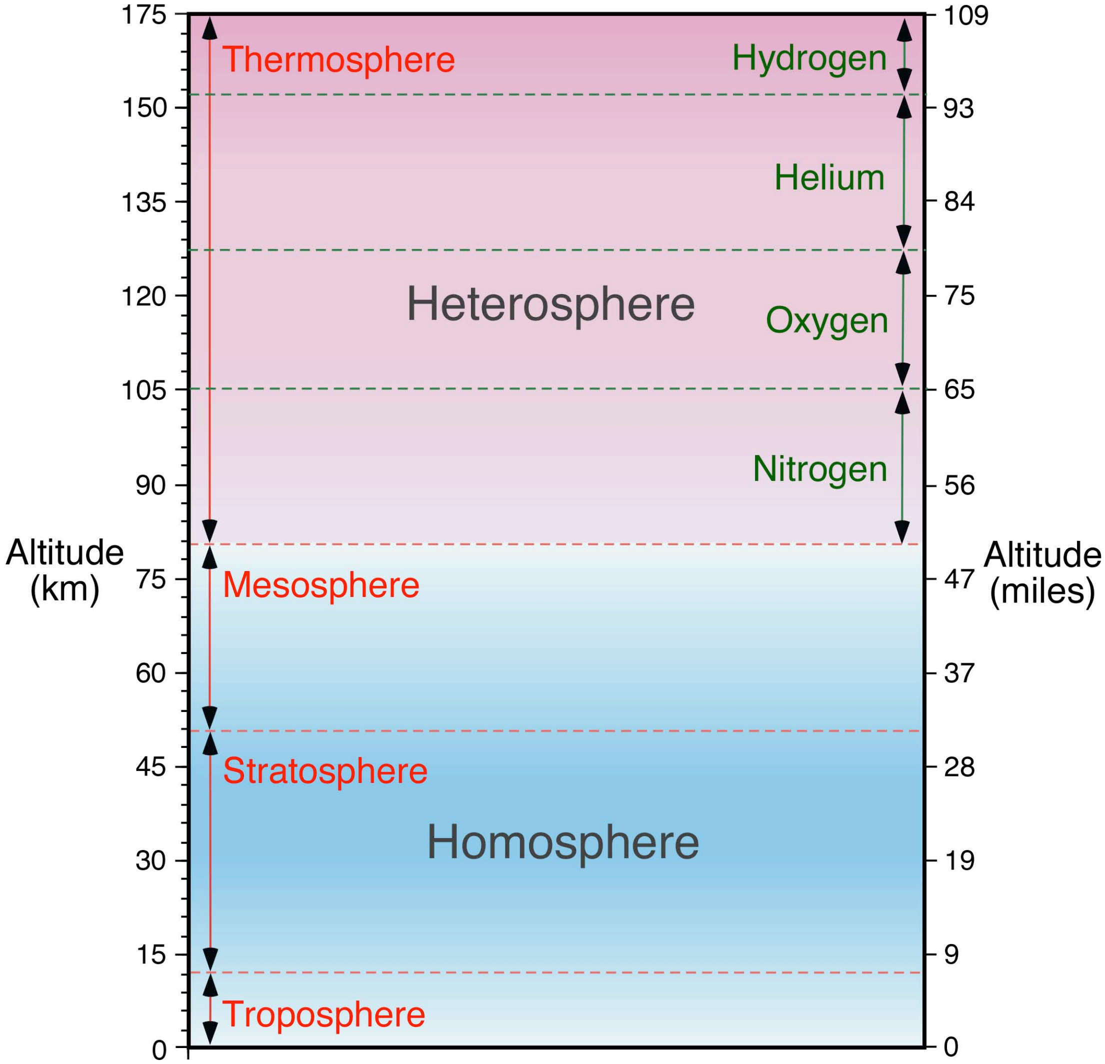

FIGURE 5.8 Chemical layers in the atmosphere. The homosphere is a layer in the atmosphere that is defined by uniform mixture of gases. It extends from the Earth's surface to an altitude of about 80 kilometers. Above this homosphere is the heterosphere. In the heterosphere the gases are layered vertically according to molecular mass. *(Image Copyright: Michael Pidwirny)*

contain is transferred to nitrogen and oxygen gas molecules in the upper atmosphere. When enough energy is absorbed, the gas molecules begin radiating energy and we see this as visible color bands of light (the aurora).

ABSORPTION, REFLECTION, AND SCATTERING

Solar radiation passing through the space reaches the edge of the Earth's outer atmosphere altered only in terms of intensity. This alteration is caused by the fact that light travels away from the Sun's surface like an expanding sphere. The **Inverse Square Law** tells us that for every unit distance traveled the intensity of solar radiation decreases to one-quarter of its original quantity.

The Earth's atmosphere is not transparent to incoming solar radiation. The atmosphere contains a variety of gases, liquid droplets of varying chemistry, and different types of solid particles. Because of these gases and particles, three atmospheric processes modify the sunlight passing through our atmosphere: scattering, absorption, and reflection.

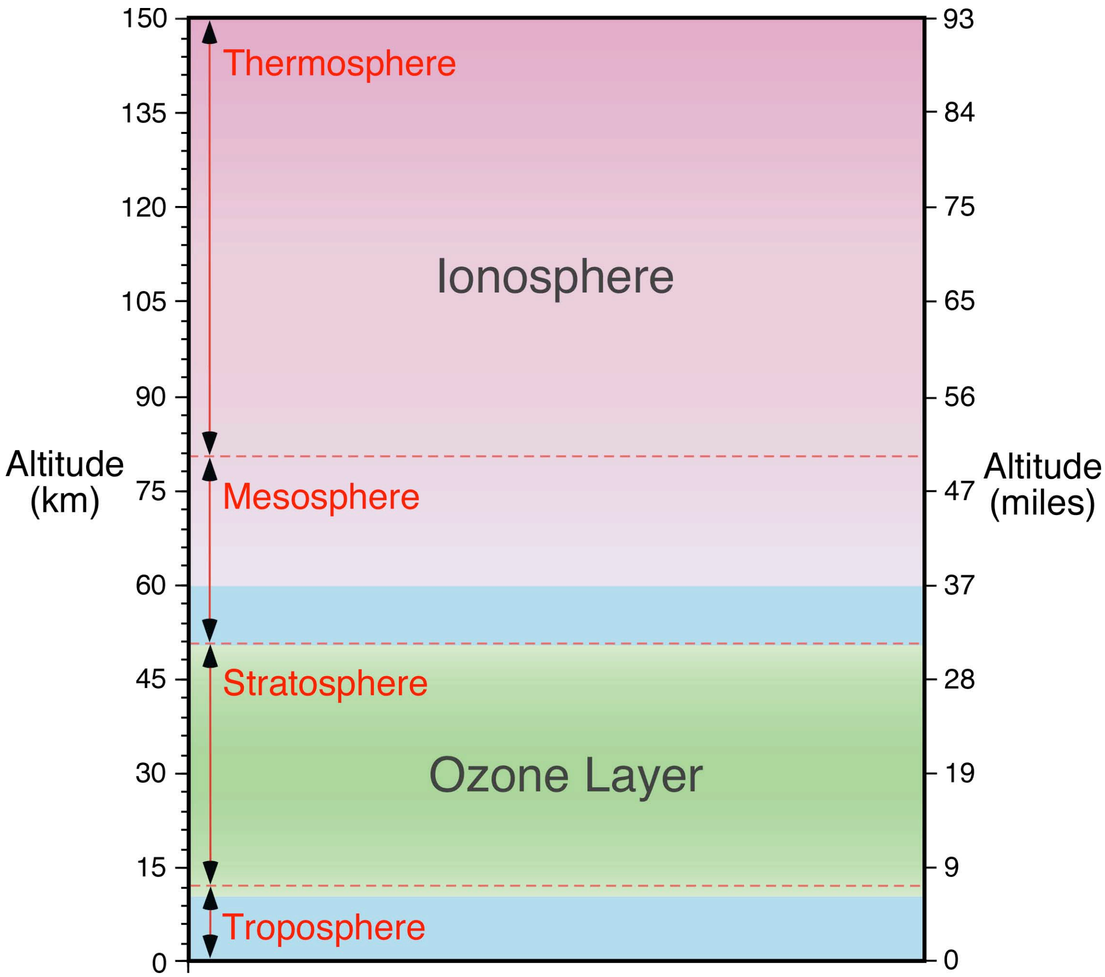

FIGURE 5.9 The ozone layer and ionosphere. The ozone layer is a zone in the atmosphere where ozone is concentrated. The ionosphere is a zone that contains large numbers of electrically charged particles (atoms and molecules) known as ions. *(Image Copyright: Michael Pidwirny)*

Atmospheric Scattering

The process of **scattering** occurs when small particles and gas molecules diffuse part of the incoming solar radiation in random directions. This process does not alter the wavelength of the solar rays being redirected (**Figure 5.11**). Yet, scattering does reduce the amount of incoming solar radiation reaching the Earth's surface. About 20% of scattered shortwave solar radiation is redirected back to space. This process is called **backscattering**.

Atmospheric scattering actually involves three separate processes. **Rayleigh scattering** occurs when solar insolation interacts with gas molecules at a height of about 10 km (6.2 mi) in the atmosphere. Rayleigh scattering does not affect the various wavelengths of solar radiation uniformly. This process tends to be most effective with ultraviolet and shorter wavelengths of visible spectrum. As a result, Rayleigh scattering mainly redirects wavelengths that correspond to the color blue when the Sun is well above the horizon. Without Rayleigh scattering in our atmosphere the daylight sky would appear black (**Figure 5.12**). This particular type of scattering is also responsible for the orange or red skies seen just before sunset and immediately after sunrise (**Figure 5.13**). When the Sun is positioned near the horizon, its rays must travel over a much longer atmospheric path. This increased distance causes the blue wavelengths to be completely scattered out by the time it reaches an observer on the Earth. Consequently, the light that finally reaches the observer consists mainly of orange and red wavelengths.

Mie scattering occurs with atmospheric particulates that are one to ten times larger than the wavelength of the solar radiation. This type of scattering generally influences

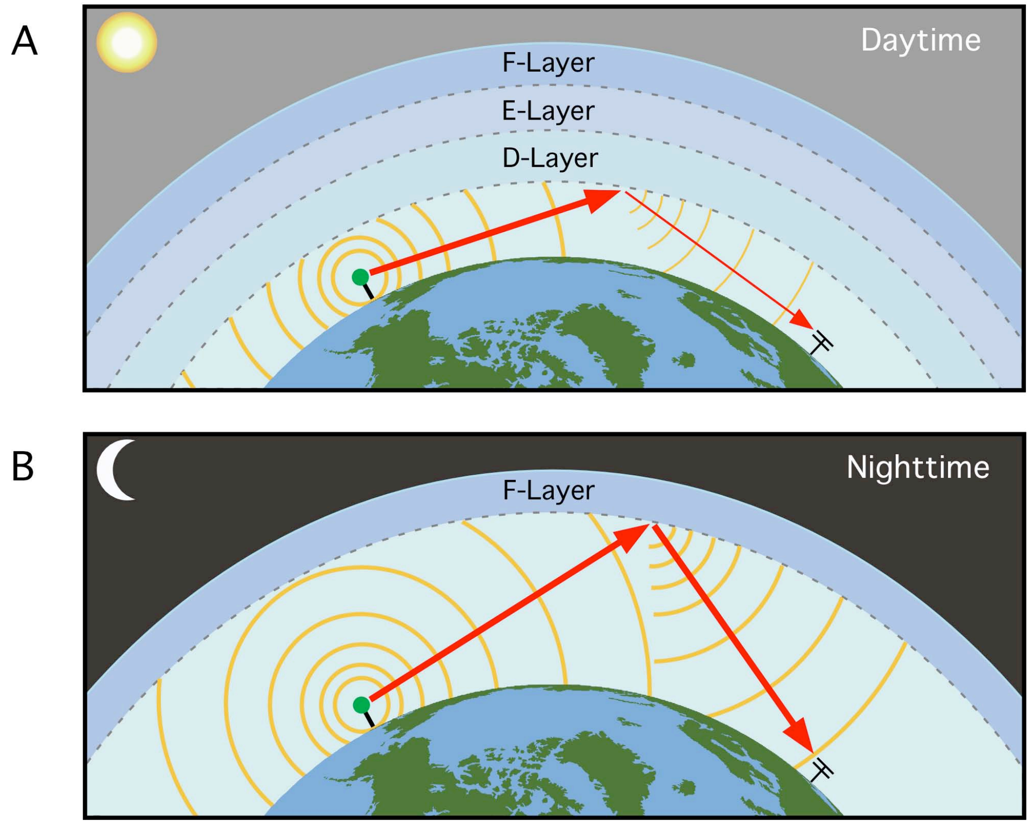

FIGURE 5.10 We use the ionosphere to help broadcast radio transmissions over long distances. By reflecting radio waves off the ionosphere, surface locations normally obscured by the Earth's curvature can receive transmissions. During the day (A), radio signals must be strengthened because the D- and E-layers have the ability to partially absorb radio waves. At night (B), the D- and E-layers dissipate and only the F-layer is used to reflect radio signals. Because very little absorption occurs in the F-layer, radio transmissions received on the Earth at night are less distorted and stronger. *(Image Copyright: Michael Pidwirny)*

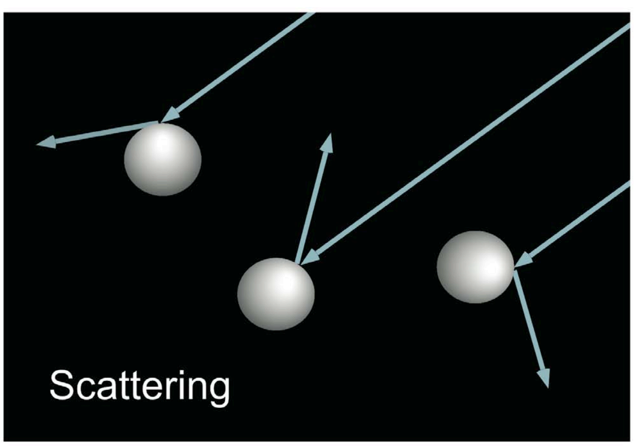

FIGURE 5.11 The process of atmospheric scattering causes rays of sunlight to be redirected to a new direction after hitting a particle in the atmosphere. In this illustration, we see how three particles send striking light rays off into three different directions. Scattering does not change the striking light ray's wavelength or intensity. *(Image Copyright: Michael Pidwirny)*

FIGURE 5.12 On the moon, Rayleigh scattering does not occur because of an extremely thin atmosphere. As a result, the moon's sky appears black even when the sun is directly overhead. *(Source: NASA)*

wavelengths of sunlight that are longer. Dust, pollen, smoke, and water droplets are the main types of particles that cause this form of scattering. Much like Rayleigh scattering, most of the incident radiation is redirected in a forward direction. Mie scattering normally occurs at altitudes from 0 to 5 km (0 to 3.1 mi) where large particulate matter tends to concentrate. Finally, Mie scattering also is important in the creation of red sky color at sunrise and sunset.

Sunsets are sometimes made more sensational when the atmosphere is loaded with additional particles from, dust storms, forest fires, and exploding volcanoes. After the eruption of Krakatau in 1883 observers all over the world described brilliant red sunsets for about three years. The increased scattering and reflection of solar insolation caused by volcanic eruptions can be very effective in reducing the amount of sunlight received at the Earth's surface. Satellite measurements for a significant period of time after the 1991 eruption of Mount Pinatubo in the Philippines recorded an increase in the amount of sunlight returned to space by the atmosphere.

Non-selective scattering takes place when large atmospheric particulates interact with incoming solar radiation. Water droplets with diameters between 5 to 100 micrometers in diameter are very effective at this type of scattering. Non-selective scattering equally influences all wavelengths in the visible and near infrared spectrum. It produces a scatter that has a color that ranges from blue to white. When this type of scatter is white in color reduced visibility occurs (**Figure 5.14**). You may have observed non-selective scattering when shining a light into fog. The water droplets in the fog redirect the light in all directions producing a white haze and poor visibility.

FIGURE 5.13 Red skies at sunset and sunrise occur because of greater atmospheric scattering. When the Sun is at a low angle, the incoming sunlight has to travel through a much thicker layer of atmosphere. As a result, the rays of sunlight undergo much more scattering and when the light finally reaches an observer on the Earth's surface it appears colored orange to red.

Atmospheric Absorption

Some gases and particles in the atmosphere also have the ability to absorb incoming insolation (**Figure 5.15**). Atmospheric **absorption** is defined as a process in which solar radiation is retained by a substance found in the atmosphere and converted into heat energy. The creation of heat energy also causes the substance to emit its own radiation. In general, the absorption of solar radiation by substances in the Earth's atmosphere results in temperatures that can get as high as 1200°C (2200°F) in the thermosphere. According to **Wien's Law**, bodies with temperatures at this level or lower would emit their radiation in the longwave band. Further, this emission of radiation is in all directions, so a sizable proportion of this energy is lost to space.

Absorption can occur due to the presence of atmospheric gases, aerosols, clouds, and precipitation

FIGURE 5.14 Non-selective scattering. The poor visibility or haze seen in this photograph is the result non-selective scattering of sunlight by minute water droplets in the atmosphere. The air above the snow is well above freezing and has high quantities of evaporated water. The cold surface of the snow is causing the warm moist air to cool forming a fine mist of water droplets.

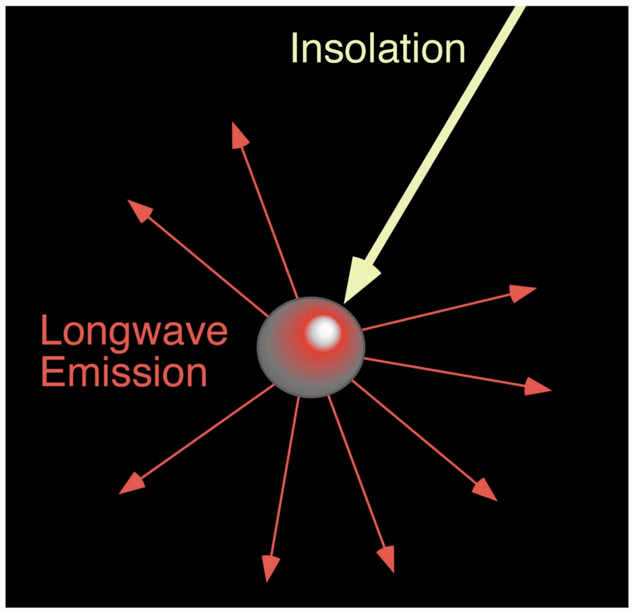

FIGURE 5.15 Atmospheric absorption. In this process, sunlight is absorbed by an atmospheric particle, transferred into heat energy, and then converted into longwave radiation emissions that come from the particle. *(Image Copyright: Michael Pidwirny)*

particles. Solar radiation in the near-infrared band is especially susceptible to atmospheric absorption. Near-infrared radiation represents nearly 50% of the energy emitted from the Sun. Water vapor and carbon dioxide absorb almost all of this type of solar radiation within the atmosphere.

Atmospheric Reflection

The final process in the atmosphere that modifies incoming solar radiation is reflection (**Figure 5.16**). Atmospheric **reflection** is a process where sunlight is redirected by 180 degrees after it strikes an atmospheric particle. This redirection often causes a 100% loss of the insolation to space. Most of the reflection in our atmosphere occurs in clouds when light is intercepted by particles of liquid and frozen water (**Figure 5.17**). The reflectivity of clouds ranges between 40 to 90%, with an average value of about 60%. Exactly how much light is redirected from a cloud depends on two factors: cloud thickness and the relative abundance of water droplets and ice crystals. Thicker clouds contain more particles increasing the chance of a ray of sunlight being reflected off a water droplet or ice crystal. When water changes from liquid to solid there is a significant increase in its reflectivity.

Atmospheric Transmission

The passage of radiation through the atmosphere without it being absorbed, reflected, or backscattered to space is called atmospheric **transmission**. It is important for us to realize that transmission in the atmosphere varies both spatially and temporally. Both of these types of variability are due to changes in the clarity of the atmosphere. A number of natural and human mediated processes influence the transparency of the atmosphere. Some important natural processes include cloud development, vegetation fires, sulfide emissions from marine plankton, wind transported dust, and volcanic eruptions. The magnitude of the effect of these natural processes on atmospheric transmission can be very significant. For example, measurements from the June 15th, 1991 volcanic eruption of Mount Pinatubo (Luzon Island, Philippines) indicate that this single event released 20 million metric tons of sulfur dioxide gas into the stratosphere in just a few days (**Figure 5.18**). This sulfur dioxide then reacted with water in the stratosphere to form a cloud of sulfuric acid particles (**Figure 5.19**). By mid-August of 1991, this cloud of particles was partially obscuring 42% of the Earth's surface from receiving sunlight. A few more months later the cloud completely covered the planet. With less sunlight available for heating the atmosphere near the ground surface, mean global air temperature dropped by 0.5°C (0.9°F) in the following year. The cloud completely dissipated after three years.

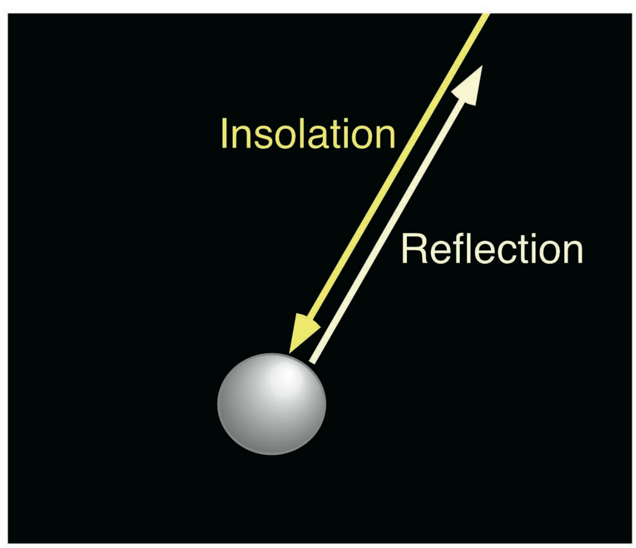

FIGURE 5.16 Atmospheric reflection. In this process, the solar radiation striking an atmospheric particle is redirected back to space unchanged. *(Image Copyright: Michael Pidwirny)*

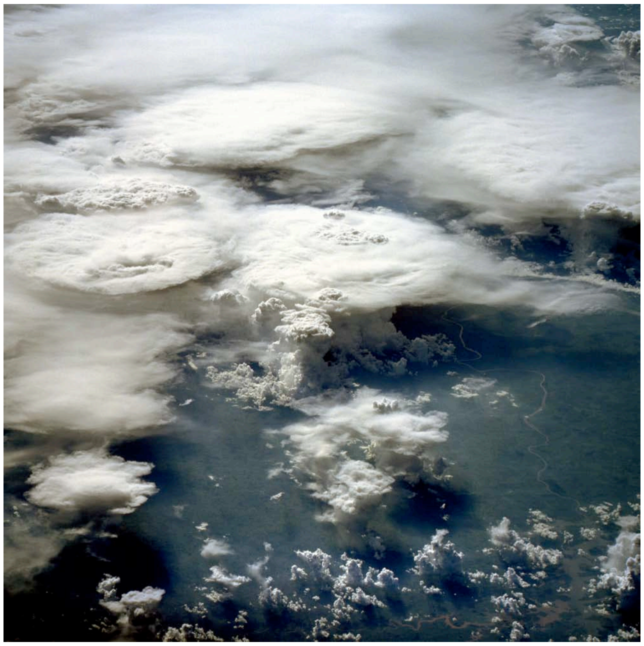

FIGURE 5.17 Most of the reflection that takes place in the Earth's atmosphere is caused by clouds. This overhead image from space illustrates how reflective clouds can be. The bright white color of the these cumulus clouds indicates that all wavelengths of light are being reflected back to space from their surface. *(Source: NASA)*

Humans also influence the transmission of solar radiation through the atmosphere. A number of gases and aerosols are released into the atmosphere from human activities. Some of the activities that release large quantities of particles in the atmosphere include biomass burning, fossil fuel burning, and vehicle transportation. Quantities of these particles are generally highest around cities where human populations are concentrated. The emission of these particles may be significantly reducing the amount of solar radiation received in populated regions of the world. This reduction may be counteracting some of the increased warming predicted from the human mediated enhancement of the greenhouse effect. Scientists refer to this process as **global dimming**.

SURFACE ALBEDO AND INSOLATION

Insolation (sunlight) reaching the Earth's surface unmodified by any of the atmospheric processes just discussed is termed **direct solar radiation**. Solar radiation that reaches the Earth's surface after it was altered by the process of scattering is called **diffused solar radiation**. Not all of the direct and diffused radiation available at the Earth's surface is absorbed. As in the atmosphere, some of the sunlight received at the Earth's surface is redirected back to the atmosphere or space by reflection. The reflectivity of the Earth's surface varies considerably. Quantitative measurements of the reflectivity of a surface are called **albedo**. **Table 5.3** lists the albedos of some of the common surface types found on our planet.

Visually we can estimate the albedo of an object's surface from its tone or color. This method suggests that albedo becomes higher as an object gets lighter in shade. The data in **Table 5.3** verifies this idea. Light toned surfaces like snow do have high albedos. Low albedos are associated with surfaces that appear dark colored to our eyes. Some dark colored surfaces include black top roads, coniferous forest, and dark soil. **Table 5.3** also indicates that the albedo of water varies with Sun angle. When Sun angles are high, water tends to absorb more than 95% of

FIGURE 5.18 Eruption of Mount Pinatubo. The first of series of major explosive eruptions of Mount Pinatubo began at 8:51 am on June 12, 1991. The largest eruption occurred on June 15. The eruption of Mount Pinatubo released millions of metric tons of ash and gas into the Earth's atmosphere. *(Source: United States Geological Survey)*

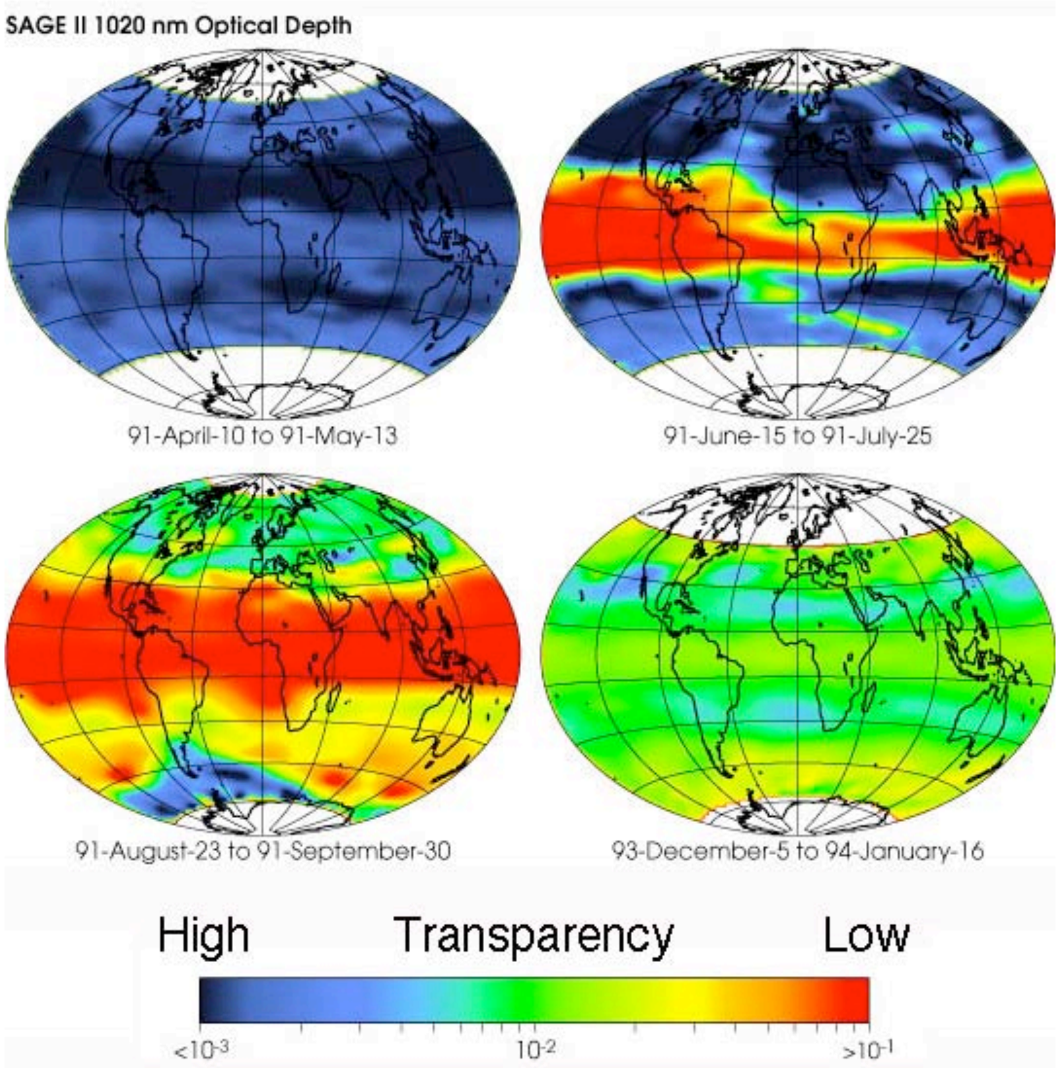

FIGURE 5.19 Mount Pinatubo aerosol cloud viewed from space. This series of images compares the optical transparency of the stratosphere as observed by NASA's SAGE II satellite before and after the June 1991 Pinatubo eruption. Initially, the ejected volcanic materials were concentrated in the stratosphere above the tropics. Within a few months, winds in the upper atmosphere spread the aerosol cloud across the entire globe. The aerosol cloud persisted in the stratosphere for almost three years. *(Source: NASA)*

the insolation falling on it. At low Sun angles, the surface of water becomes much more reflective.

Surface albedo of the Earth varies with geographic region and time of year. Most of the Earth's surface is either covered by water, vegetation, bare soil, or rock. Generally, these surfaces have albedos between 5 to 45%. Over time, the albedo of vegetation can vary significantly. Plants can shed their dark colored leaves in the cold and dry seasons. This process often exposes a surface that has a different albedo as it is covered by bare ground, low growing plants, and decaying leaves. The color of plants also changes with moisture conditions. Drought often causes plant leaves to become dried out and lighter in color. Subsequent rainfall after drought can cause the albedo of vegetated surfaces to become less reflective because of new leaf growth. In the middle and high latitudes, the albedo of ground surfaces can be altered significantly because of temperature change and snowfall. When temperatures drop below freezing, precipitation falls as highly reflective snow (**Figure 5.20**).

Satellite Measurements of Albedo

Global measurements of the Earth's surface albedo can be best measured with the aid of sensors aboard orbiting space satellites. NASA's Earth Radiation Budget

TABLE 5.3 Albedo of some common surface types found on the Earth.

Surface Type	Percent Reflectivity
Snow, fresh-fallen	75-95
Snow, several days old	40-70
Water, high Sun angle	3-5
Water, low Sun angle	10-50
Sea Ice	30-40
Sand dune, dry	35-45
Sand dune, wet	20-30
Soil, dark	5-15
Soil, dry light sand	25-45
Road, black top	5-10
Forest, deciduous	10-20
Forest, coniferous	5-15
Grassland	15-25
Tundra	15-20
Cropland	15-25

Experiment (ERBE) was one of the first attempts of making such measurements. This experiment used a variety of satellite sensors aboard Nimbus-7, NOAA-9, and the Earth Radiation Budget Satellite (ERBS) to monitor the Earth's albedo for a period of about four years. Figure 5.21 shows the monthly average surface albedo of the Earth for January and July, 1987. In this figure, most of the reflective properties of the atmosphere have been removed. The patterns seen here are probably representative for most other years. For both January and July, the lowest surface albedos occur over oceans in a zone that covers more than 100 degrees of latitude. Albedo values of this zone are between 8 and 13%, and the center of this zone shifts seasonally. In July, the low albedo zone is located approximately at the Tropic of Cancer (23.5°N), while in January it migrates to the Tropic of Capricorn (23.5°S). At the higher latitudes, the albedo of the ocean surface increases significantly because of low Sun angles or the presence of sea ice. In the July image, the region occupied by the Arctic Ocean has an albedo between 45 to 60%. On the Earth's terrestrial surface, vegetated areas have an albedo from 15 to 25%. Non-vegetated regions like the Sahara Desert reflect about 30 to 40% of the Sun's incoming light. Other land surfaces with high albedos are

FIGURE 5.20 Highly reflective snow. A winter snowfall has quickly changed the albedo of these trees. *(Image Copyright: Michael Pidwirny)*

glaciers and seasonal snowfields. The large glaciers covering Greenland and Antarctica reflect as much as 75% of the insolation falling on their surfaces. Comparing the January and July images, we can see that the albedos of areas with a latitude greater than 45°N vary annually because of seasonal snowfall. In these areas, summer albedos typically are around 20%, while winter values jump to as high as 70%

Figure 5.21 shows the surface reflectivity of the Earth for January and July 1987. Comparing this figure to **Figure 5.22**, which measures combined surface and atmosphere reflectivity (or **planetary albedo**), illustrates the contribution that clouds have on reflecting incoming sunlight back to space. Significant bands of reflective cloud exist over at the equator and in the mid-latitudes. Skies are generally clear of cloud over the major deserts, sub-tropical oceans, and the large continental glaciers of Greenland and Antarctica.

GLOBAL RADIATION TRANSFERS

We can model the fate of incoming **shortwave radiation** from the Sun (insolation) for the entire planet with a simple cascade diagram (**Figure 5.23**). This diagram quantifies the effects of surface albedo and atmospheric scattering, absorption, and reflection on the reception of insolation at the Earth's ground surface over a period of one year. In this model, we begin with an input of insolation available just outside the Earth's atmosphere. For the purpose of simplification, this input is given a value of 100%. Of all the sunlight that passes through the atmosphere annually, only 50% of the available radiation is

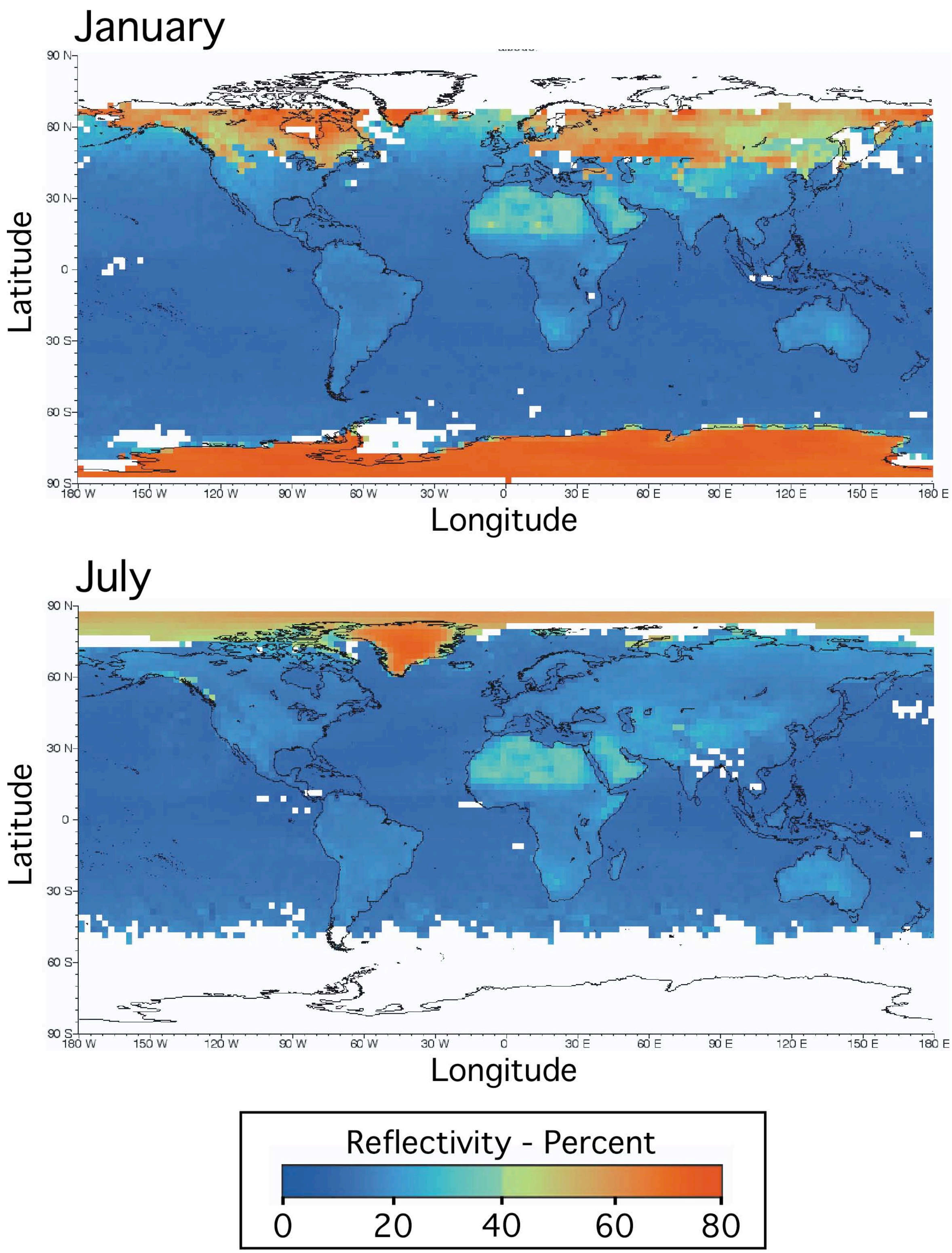

FIGURE 5.21 Surface reflectivity of the Earth for January and July 1987. Cells with missing data are colored white. Measured by sensors aboard a variety of satellites for NASA's Earth Radiation Budget Experiment (ERBE). *(Source: NASA - Earth Radiation Budget Experiment)*

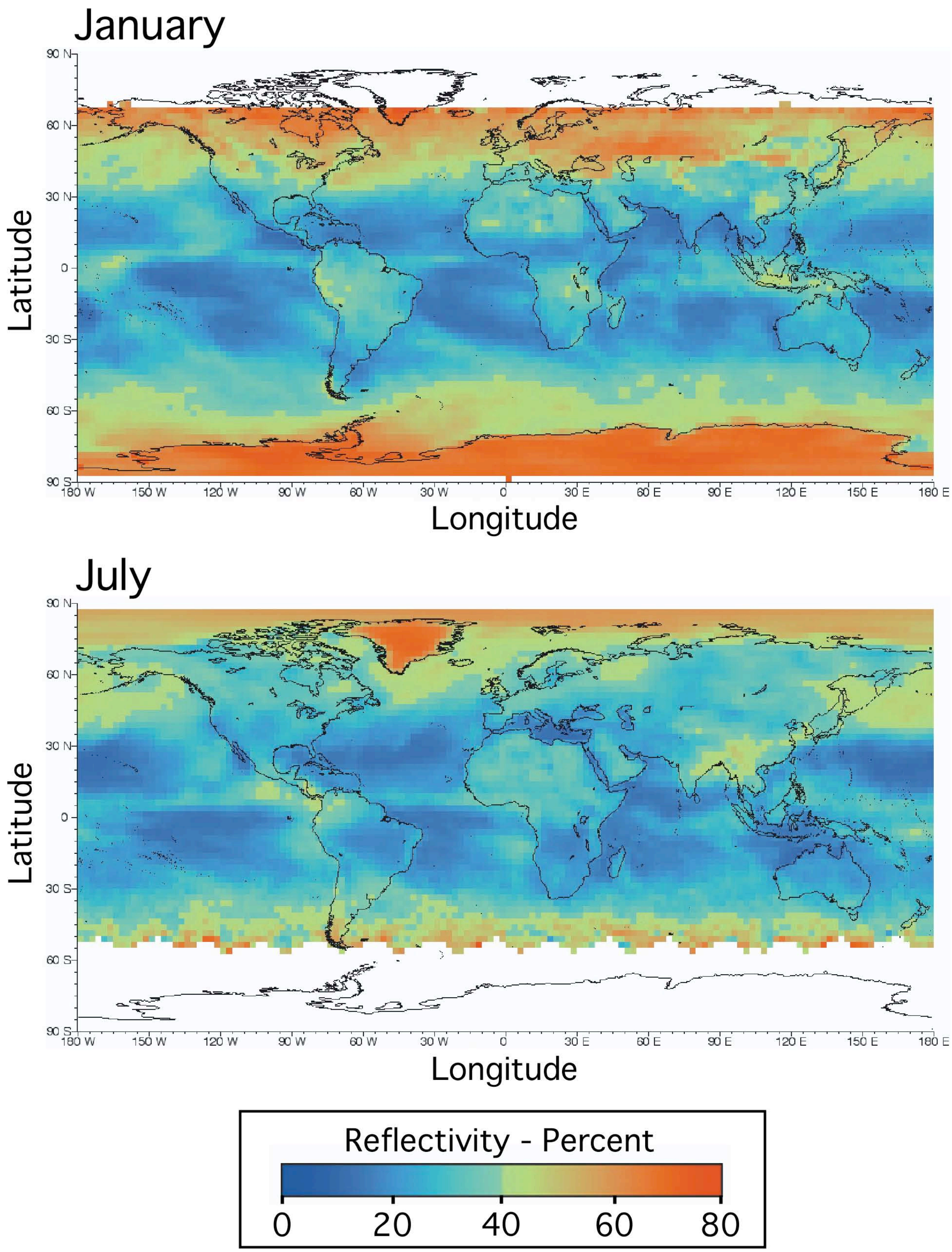

FIGURE 5.22 Combined surface and atmosphere reflectivity (or planetary albedo) of the Earth for January and July 1987. Cells with missing data are colored white. Measured by sensors aboard a variety of satellites for NASA's Earth Radiation Budget Experiment (ERBE). *(Source: NASA - Earth Radiation Budget Experiment)*

absorbed at the Earth's surface to do work. This energy is used to heat the Earth's surface and lower atmosphere, melt and evaporate water, and fuel photosynthesis in plants. Of the other 50%, 5% is reflected back to space by the Earth's surface, 5% is backscattered, and 20% is reflected to space by clouds and atmospheric particles. The combination of these three losses of energy back to space totals 30%. This quantity represents the Earth's planetary albedo. We are now left with 20% unaccounted for. Our planet's atmospheric gases, particles, and clouds absorb this remaining amount. The total quantity of energy absorbed by the Earth's atmosphere and surface is 70% of the incoming sunlight.

SPATIAL PATTERNS OF SURFACE INSOLATION INPUT

The insolation cascade model for the Earth helped us to visualize how the Sun's incoming energy is partitioned. We also need to appreciate the fact that the spatial patterns of insolation received at the Earth's surface are not uniform. The combined effect of Earth-Sun relationships (**angle of incidence** and day length variations) and the modification of the Sun's beam as it travels through the atmosphere produces specific global patterns. **Figure 5.24** describes the insolation received on the Earth for the year 1987. For this typical year, the highest values of insolation received (about 360 Wm^{-2}) occur just outside the equator over ocean surfaces. At the equator, the more frequent presence of greater cloud cover, especially over land, reduces the amount of insolation available by about 60 to 80 Wm^{-2}. Outside these maximum zones, insolation absorbed decreases with increasing latitude. In the Earth's polar regions, annual insolation received can be as low as 50 Wm^{-2}.

Figure 5.25 shows the patterns of insolation absorbed at the ground surface for January and July 1987. In this image, we can see the seasonal migration of the zone of maximum received insolation because of changes in Earth's tilt relative to the Sun. For both months, solar radiation receipts generally decrease outside these maximum zones with increasing latitude. This pattern is

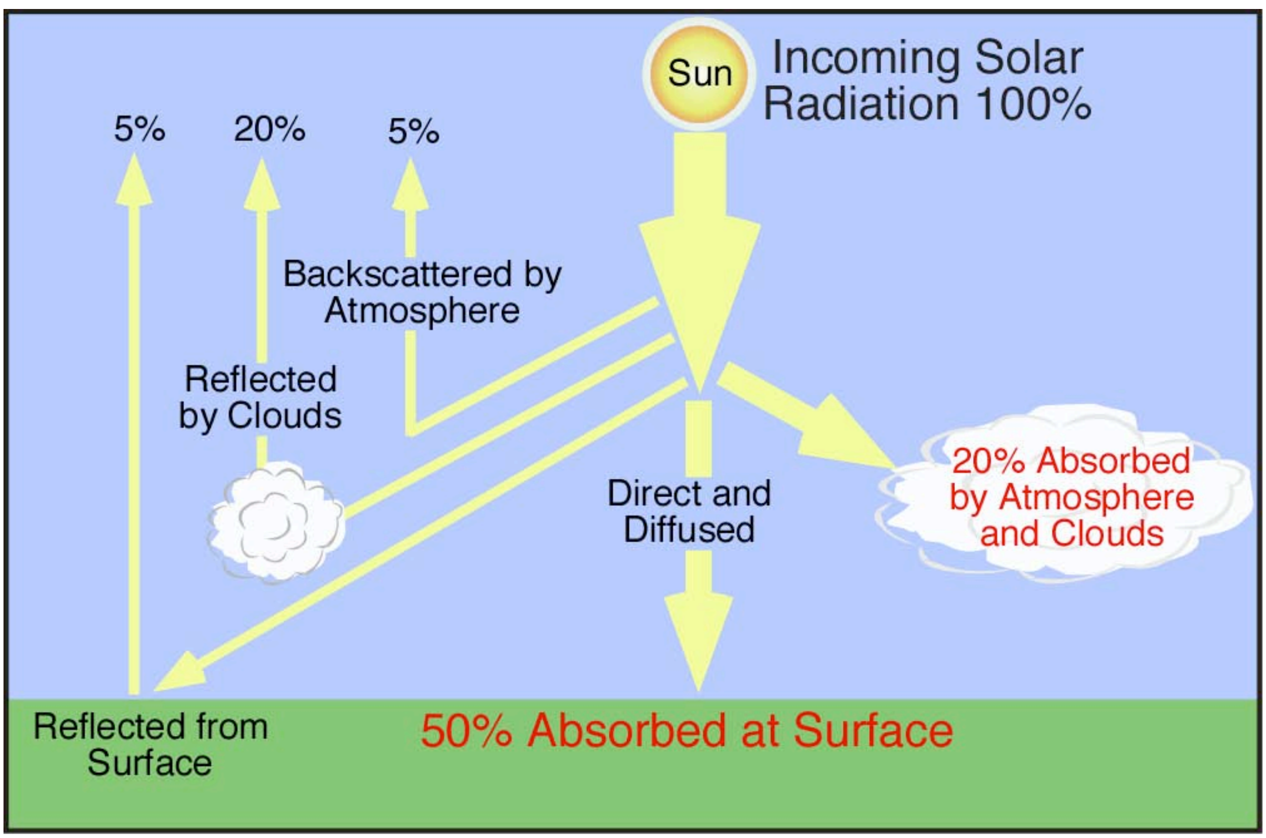

FIGURE 5.23 Average annual global partitioning of incoming solar radiation. pproximately, 50% of the radiation from sun is absorbed at the Earth's surface as direct and diffuse light. Of the remaining 50%, about 20% of the sunlight is absorbed by clouds and other atmosphere gases and particles. Roughly 30% is returned to space through surface reflection (5%), atmospheric backscattering (5%), and reflection off clouds (20%). *(Image Copyright: Michael Pidwirny)*

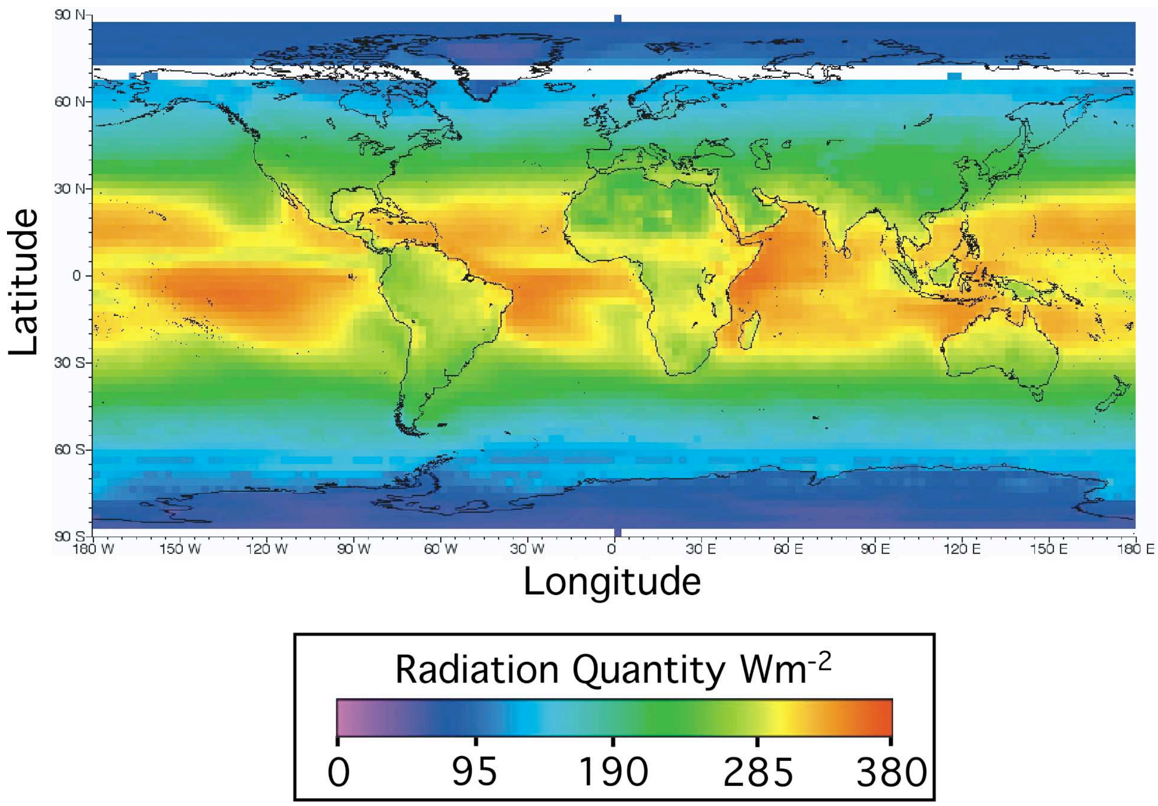

FIGURE 5.24 Annual absorption of insolation at the Earth's surface and in the atmosphere for 1987. Cells with missing data are colored white. Measured by sensors aboard a variety of satellites for NASA's Earth Radiation Budget Experiment (ERBE). *(Source: NASA - Earth Radiation Budget Experiment)*

mainly the result of Earth-Sun geometric relationships and their effect on the duration and intensity of solar radiation received. Locations poleward of the Arctic Circle (66.5°N) and the Antarctic Circle (66.5°S) have insolation values of zero in January and July, respectively. At these times, these areas of the world are in complete darkness.

Concept of Radiation Balance

In Chapter 4, we learned that a one-to-one relationship exists between an object's ability to absorb and emit radiation. As a result of this relationship, the absorption of solar radiation by the Earth should be balanced by the emission of a similar quantity of radiation back to space; however, the quality or wavelength of this emission may differ. The wavelength of emission is controlled by temperature and the Earth's temperature is much lower than the Sun's. Our planet's emission spectrum is in the infrared band, and the wavelength of maximum emission is at about 10 µm.

Earth Space Bound Emission

Figure 5.26 describes the annual space bound longwave emission that occurs from the Earth's ground surface and atmosphere. Note that these patterns of output do not mirror the patterns of input seen in **Figure 5.24**. Longwave output from our planet tends to be more evenly spread out than shortwave input. This finding suggests that after the Sun's radiation is absorbed by the Earth's surface and atmosphere some of this energy is redistributed about the planet by other mechanisms. These mechanisms include the greenhouse effect and the transfer of heat energy by the horizontal circulation of the atmosphere and ocean waters. We will explore these phenomena in greater detail in upcoming chapters.

Outgoing emissions of longwave radiation in **Figure 5.26** are greatest where skies are generally clear and surface temperatures are high. Low values of emission are related to cooler surface temperatures and/or the presence of cloud cover. Clouds block the transmission of some of the longwave radiation to space because of the presence of

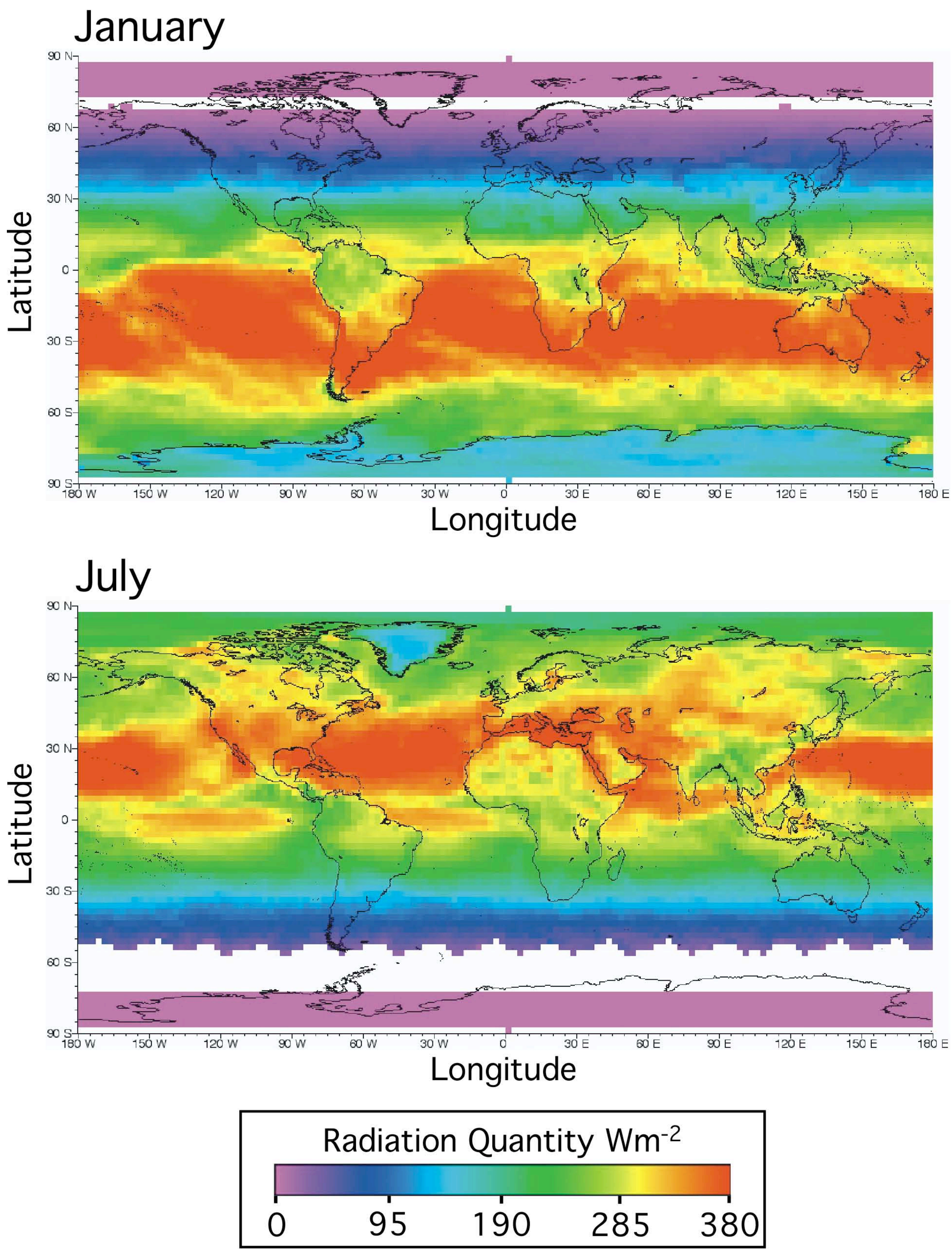

FIGURE 5.25 Absorption of insolation at the surface and in the atmosphere of the Earth for January and July 1987. Cells with missing data are colored white. Measured by sensors aboard a variety of satellites for NASA's Earth Radiation Budget Experiment (ERBE). *(Source: NASA - Earth Radiation Budget Experiment)*

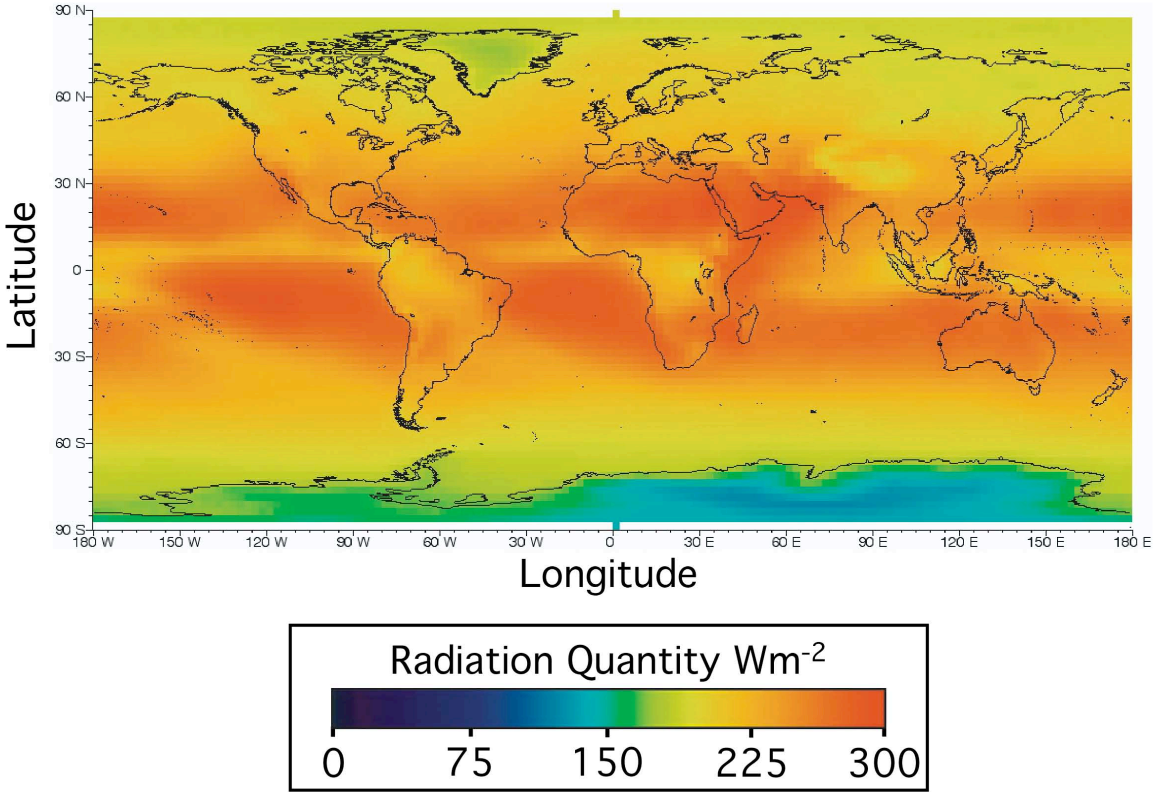

FIGURE 5.26 Annual space bound emission of longwave radiation from the surface and atmosphere of the Earth for 1987. Measured by sensors aboard a variety of satellites for NASA's Earth Radiation Budget Experiment (ERBE). *(Source: NASA - Earth Radiation Budget Experiment)*

water. Both water vapor and liquid water droplets have the ability to absorb infrared emissions. This absorbed energy is then re-radiated in the atmosphere and most of this emission is sent back down to the Earth's surface.

The patterns of combined surface and atmosphere outgoing longwave emission for January and July are shown in **Figure 5.27**. Seasonal variations in longwave output from the Earth tend to be subtle. In January, longwave emissions from the middle and high latitudes of the Northern Hemisphere are lower than the same locations in the Southern Hemisphere. Southern Hemisphere locations emit more longwave because they are experiencing summer and higher temperatures at that time. Summer in the Northern Hemisphere occurs in July. During this month, middle and high latitude locations north of the equator emit more outgoing longwave then the same regions in the Southern Hemisphere.

Annual Global Radiation Cascades

To gain a better understanding of the balance between incoming and outgoing energy we need to model the processes involved in global shortwave and longwave radiation cascade diagrams. **Figure 5.28** illustrates a slightly more complex version of the annual global shortwave radiation cascade. This model suggests that as the energy from the Sun passes through our atmosphere a number of things occur. About 25 units of the energy available is reflected or scattered back to space by clouds and particles. Atmospheric absorption by clouds, gases (like ozone), and particles consume another 20 units. A large amount of this absorption (17 out 20 units) occurs in the troposphere where most of the atmosphere's mass resides. Of the remaining 55 units, 5 units are reflected from the ground surface back to space. On average, about 50 units of the Sun's radiation reach the surface. This energy is then used in a number of processes including the heating of the ground surface and lower atmosphere, the melting of ice and snow, the evaporation of water, and plant photosynthesis.

Figure 5.29 describes the global longwave radiation cascade. This cascade indicates that energy leaves the Earth's surface through three different processes. Seven

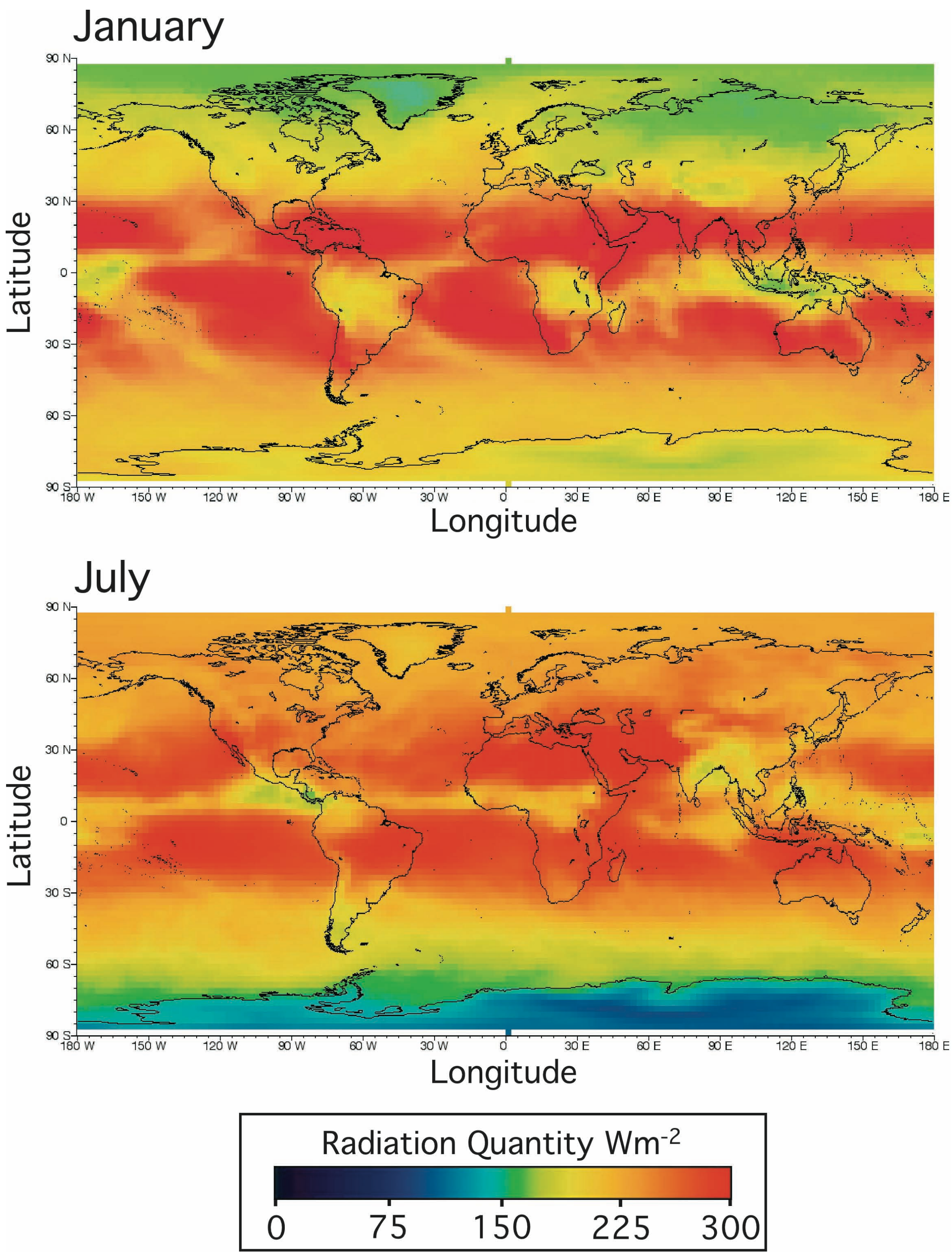

FIGURE 5.27 Space bound emission of longwave radiation from the surface and atmosphere of the Earth for January and July 1987. Measured by sensors aboard a variety of satellites for NASA's Earth Radiation Budget Experiment (ERBE). *(Source: NASA - Earth Radiation Budget Experiment)*

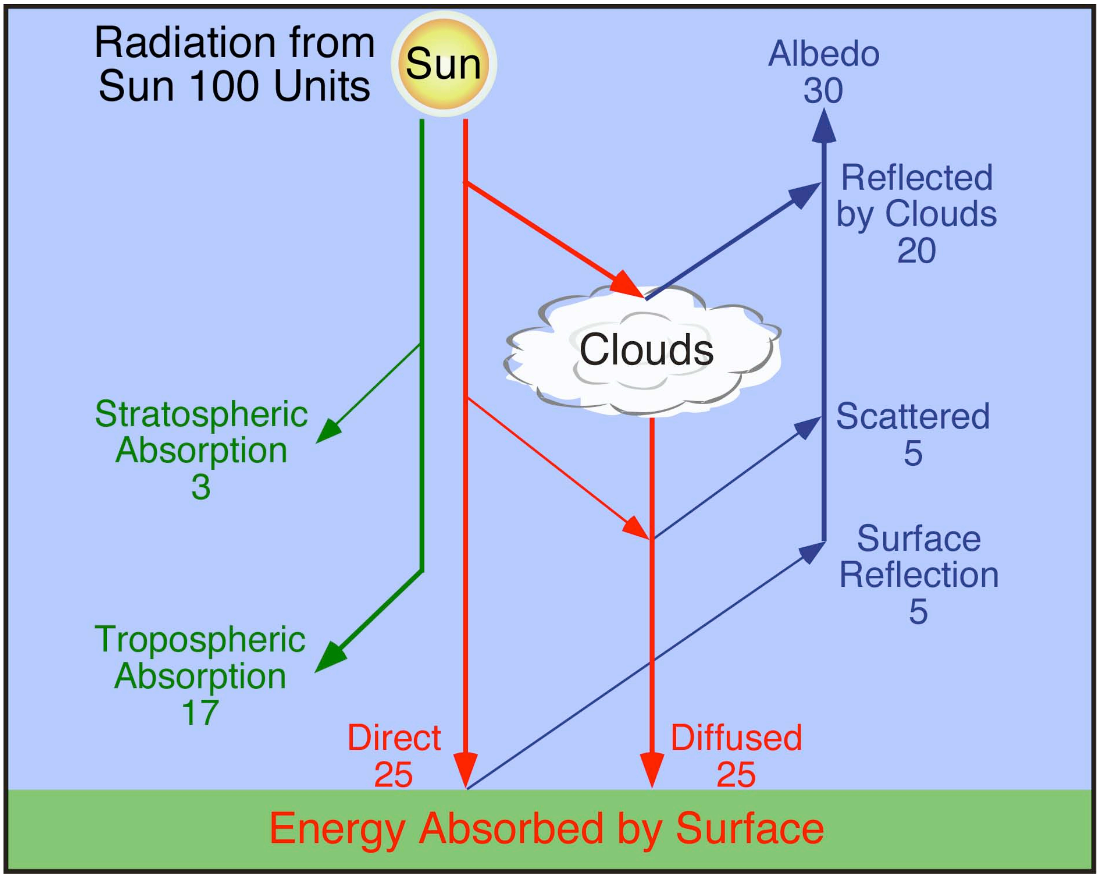

FIGURE 5.28 Global shortwave radiation cascade. For each process in the cascade quantities of shortwave radiation used is given in percentage units. The shortwave radiation cascade begins with 100 units of insolation available just outside the Earth's atmosphere. *(Image Copyright: Michael Pidwirny)*

units leave the surface as sensible heat. This heat is transferred into the atmosphere by conduction, convection, and advection (sensible heat transfer). The evaporation of water at the Earth's surface incorporates 23 units energy into the atmosphere as latent heat. This energy is released into the atmosphere when the water condenses or becomes solid (latent heat transfer). Eventually, both sensible heat and latent heat become incorporated in the emission of longwave radiation by the atmosphere and its clouds.

The surface of the Earth emits 114 units of longwave radiation. Of this emission only 12 units are directly lost to space. The other 102 units are absorbed by greenhouse gases and converted into heat energy. This heat energy is then changed into longwave radiation that is emitted from the atmosphere. The atmosphere emits a total of 152 units of longwave energy. The additional 50 units of longwave energy (152 – 102 = 50) entered the atmosphere via three processes:

- Absorption of shortwave radiation from the Sun by gases and aerosols. In a previous section, we learned that 20 units of solar energy (see **Figure 5.28**) that enters the Earth atmosphere are absorbed in the atmosphere.

- Transfer of sensible heat energy from the Earth's surface into the atmosphere through conduction and the convective movement of parcels of air upward.

- Transfer of latent heat energy stored in water vapor (23 units). This water was originally in solid or liquid form at the Earth's surface. With the absorption of radiation it was converted into vapor. This water vapor was then transported vertically and horizontally into the air by atmospheric circulation. In the atmosphere, this latent energy is later released

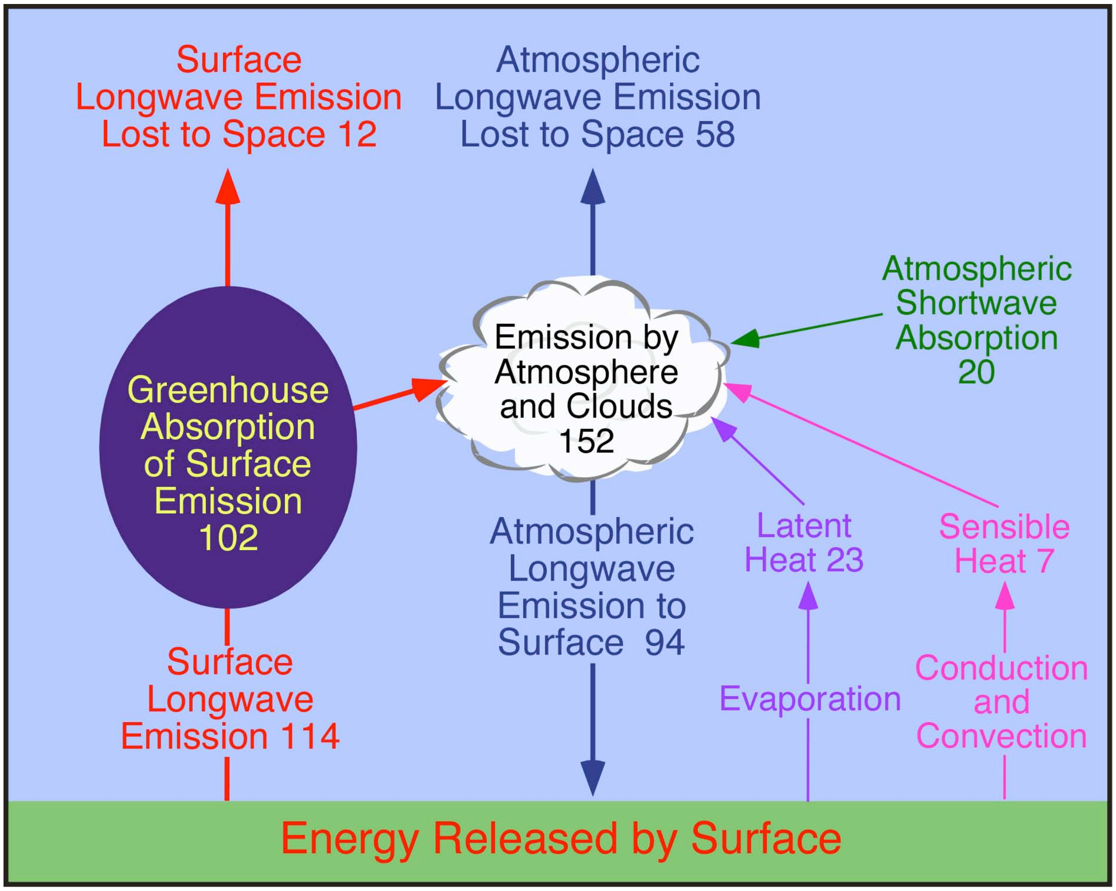

FIGURE 5.29 Global longwave radiation cascade. For each process in the cascade quantities of heat energy or longwave radiation used is given in percentage units. *(Image Copyright: Michael Pidwirny)*

into the surrounding atmospheric environment when condensation and freezing take place.

Atmospheric infrared emissions travel in two directions. About 58 units of the atmospheric emission are lost directly to space. The remaining 94 units travel to the Earth's surface where they are absorbed and transferred into heat energy. This movement of longwave radiation back to the Earth's surface is called **counterradiation**. Note that if this energy were not added back to the surface, longwave output from the ground would only be 20 units! Thus the greenhouse effect, the re-emission of longwave radiation by greenhouse gases back to the Earth's surface, re-circulates 74 units of energy.

To balance the surface energy exchanges in the longwave cascade we have to account for 50 units of missing energy [atmosphere and cloud longwave emission (94 units) minus surface longwave emission (114 units) plus latent heat transfer (23 units) plus sensible heat transfer (7 units) 94-114+23+7 = -50 units of energy]. This missing component to the longwave radiation cascade is the 50 units of energy absorbed at the Earth's surface as direct and diffused insolation in the shortwave cascade (see **Figure 5.28**). The total amount of energy lost to space in the global longwave radiation cascade is 70 units (surface emission 12 units + atmospheric emission 58 units). This is the same amount of energy that was added to the Earth's atmosphere and surface by the global shortwave radiation cascade.

Net Radiation

Net radiation can be defined as the total quantity of radiation of all wavelengths available to do work. We can suggest that the net radiation available at the Earth's surface and atmosphere is composed of the following parts:

The net radiation of the entire Earth's surface and atmosphere should equal zero because the planet is in radiative balance (input = output). Examining the spatial patterns of annual net radiation on our planet indicates a geographical imbalance (**Figure 5.30**). Net radiation tends to be positive in the tropics and subtropics and negative in the middle and high latitudes. The zone of positive net radiation stretches from about 35°N and S with maximum values located near the equator. These positive values of net radiation occur mainly because the amount of incoming shortwave radiation is greater than outgoing longwave radiation. Within this zone of positive values, there are some areas that are irregular. In the Sahara Desert, the Arabian Peninsula, and central Australia net radiation is negative. These locations experience large quantities of longwave radiation loss to space because of clear skies. Areas of localized deficit also occur in the subtropical oceans just off the western coasts of South and North America. Once again, these deficits are attributed to the general absence of cloud cover.

Incoming Shortwave - Outgoing Reflected Shortwave + Incoming Longwave - Outgoing Longwave

Zones of negative net radiation extend from 35°N to 90°N and 35°S to 90°S. Inside these zones, the amount of outgoing longwave radiation exceeds incoming shortwave radiation. These outgoing longwave emissions are energized by the net radiation surplus that occurs between 35°N to 35°S. As mentioned previously, the latitudinal transfer of sensible and latent heat by the atmospheric circulation and ocean currents moves the surplus energy that occurs at the tropics to the poles. If this redistribution of energy did not occur, temperatures would increase in the low latitudes and drop in the high latitudes.

Figure 5.31 describes global patterns of net radiation for January and July 1987. This figure captures the effect the seasonal migration of the Sun has on the Earth's net radiation balance. In January, most of the Southern Hemisphere is experiencing positive net radiation values because of the corresponding higher Sun angles and longer days of summer. The zone of positive net radiation shifts to the Northern Hemisphere in July.

THE GREENHOUSE EFFECT AND GLOBAL WARMING

The greenhouse effect is a naturally occurring process that aids in heating the Earth's surface and atmosphere. It results from the fact that certain atmospheric gases, such as water vapor (H_2O), carbon dioxide (CO_2), methane (CH_4), and nitrous oxide (N_2O) are able to change the energy balance of the planet by absorbing longwave radiation emitted from the Earth's surface. Without the greenhouse effect much of the life found on our planet probably would perish as the average temperature of the Earth would be a chilly -18°C (0°F), rather than the present 15°C (59°F).

The greenhouse effect process begins with the heating of the Earth's ground surface by incoming sunlight. This heating causes the ground surface to become an energy radiator in the longwave band. This emission of energy is generally directed to space (**Figure 5.32**); however, only a small portion of this initial emission of energy actually makes it back to space. About 90% of the outgoing infrared radiation is intercepted and absorbed by the greenhouse gases. Absorption of longwave radiation by these gases causes additional heat energy to be added to the Earth's atmospheric system. The now warmer atmospheric greenhouse gas molecules begin radiating longwave energy in all directions. Most of this emission of longwave energy is directed back to the Earth's surface where it once again is absorbed by the surface. The heating of the ground by the longwave radiation causes the ground surface to once again radiate, repeating the cycle described above, again and again, until no more longwave is available for absorption. The net result of this complex process is an average global temperature that is 33°C (59°F) warmer than it would be without the greenhouse effect.

The amount of heat energy added to the atmosphere by the greenhouse effect is controlled by the concentration of greenhouse gases in the Earth's atmosphere. All of the major greenhouse gases have increased in concentration since the beginning of the Industrial Revolution (**Table 5.4**). Most scientists believe that these higher gas concentrations are causing an enhancement of the greenhouse effect. One piece of supporting evidence for this theory is the fact that the Earth's average global temperature has warmed by about 0.3 to 0.6°C (0.5 to 1.1°F) since 1900. This greenhouse effect related warming is commonly referred to as **global warming**.

Global warming due to the enhancement of the greenhouse effect is predicted to continue for at least the next century. This forecast is being made because human activity will most likely continue to increase the concentrations of greenhouse gases in our planet's atmosphere. Of the gases responsible for the enhancement of the Earth's greenhouse effect, the single most important gas is carbon dioxide (CO_2). It alone accounts for about 63% of the recent change in the intensity of the greenhouse effect process. The enhancement contributions of the other

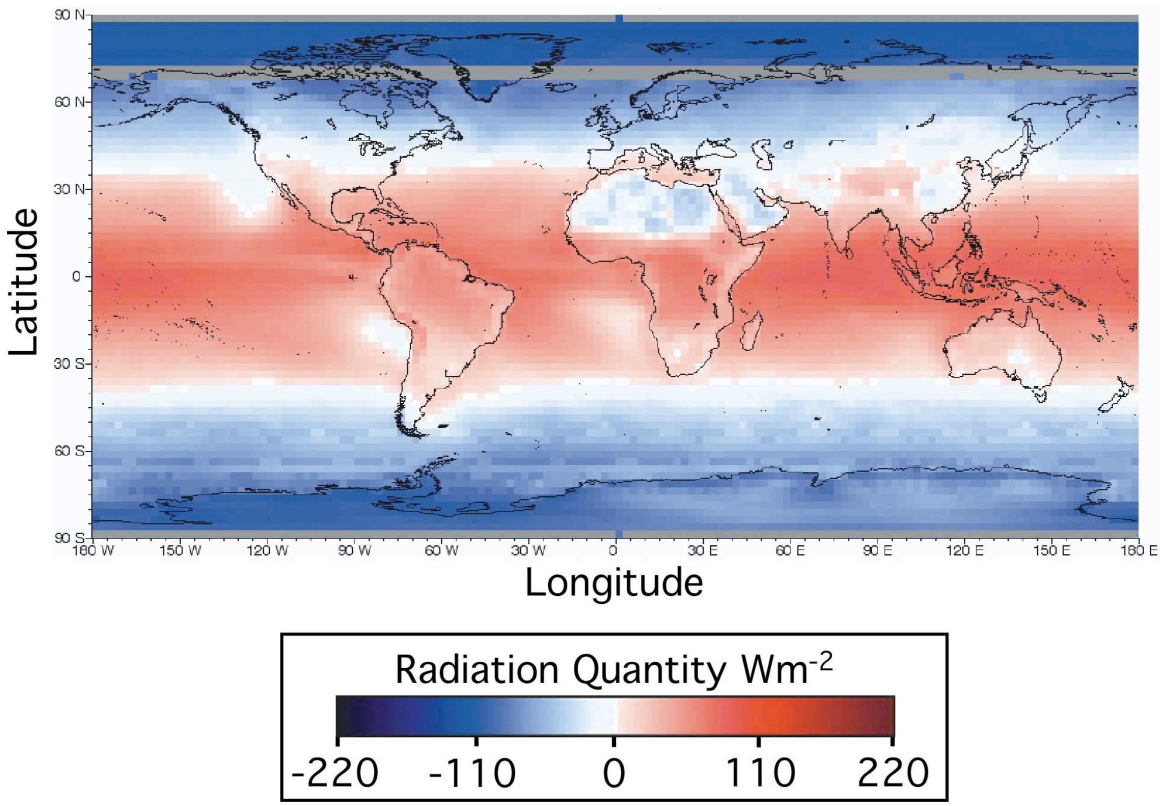

FIGURE 5.30 Annual net radiation of the Earth's combined surface and atmosphere for 1987. Cells with missing data are colored grey. Measured by sensors aboard a variety of satellites for NASA's Earth Radiation Budget Experiment (ERBE). *(Source: NASA - Earth Radiation Budget Experiment)*

gases are 18% for methane (CH_4), 10% for chlorofluorocarbons (CFCs), and 6% for nitrous oxide (N_2O) (Forster et al., 2007). Researchers expect that the concentration of carbon dioxide may reach between 450 to 600 ppm by the year 2100. Methane and nitrous oxide are also predicted to become much more plentiful in the future. Concentrations of CFCs are now declining because of imposed environmental regulations related to ozone depletion.

Predicting the amount of future global warming is accomplished by sophisticated computer climate modeling. Computer models suggest that a doubling of the concentration of just the main greenhouse gas, carbon dioxide, may raise the average global temperature between 1 and 3°C (1.8 to 5.4°F) warmer than it is today. A few scientists have voiced concern about the accuracy of these forecasts. These objections are based on the fact that the numeric equations of computer models may not precisely simulate the effects of possible negative and positive feedback mechanisms that exist in the various components of the Earth's climate system. One component that is not simulated very well in global climate models is the effect of clouds. Yet, clouds are very important to our global climate.

Increasing the Earth's temperature through global warming should cause the oceans to evaporate greater amounts of water into the atmosphere. More water vapor in the atmosphere would then cause climate to become even warmer because this substance is a greenhouse gas. But, higher concentrations of water vapor would also cause more clouds to develop. If these extra clouds would reflect a greater proportion of the Sun's energy back to space reducing the amount of solar radiation absorbed by the atmosphere and the Earth's surface. With less solar energy being absorbed at the surface, the Earth's temperature would cool. Current, global climate models do not simulate the interrelated dynamics of these two opposing processes very well. Thus, researchers are not entirely sure if the presence of more atmospheric moisture will cause a net warming or cooling of the Earth.

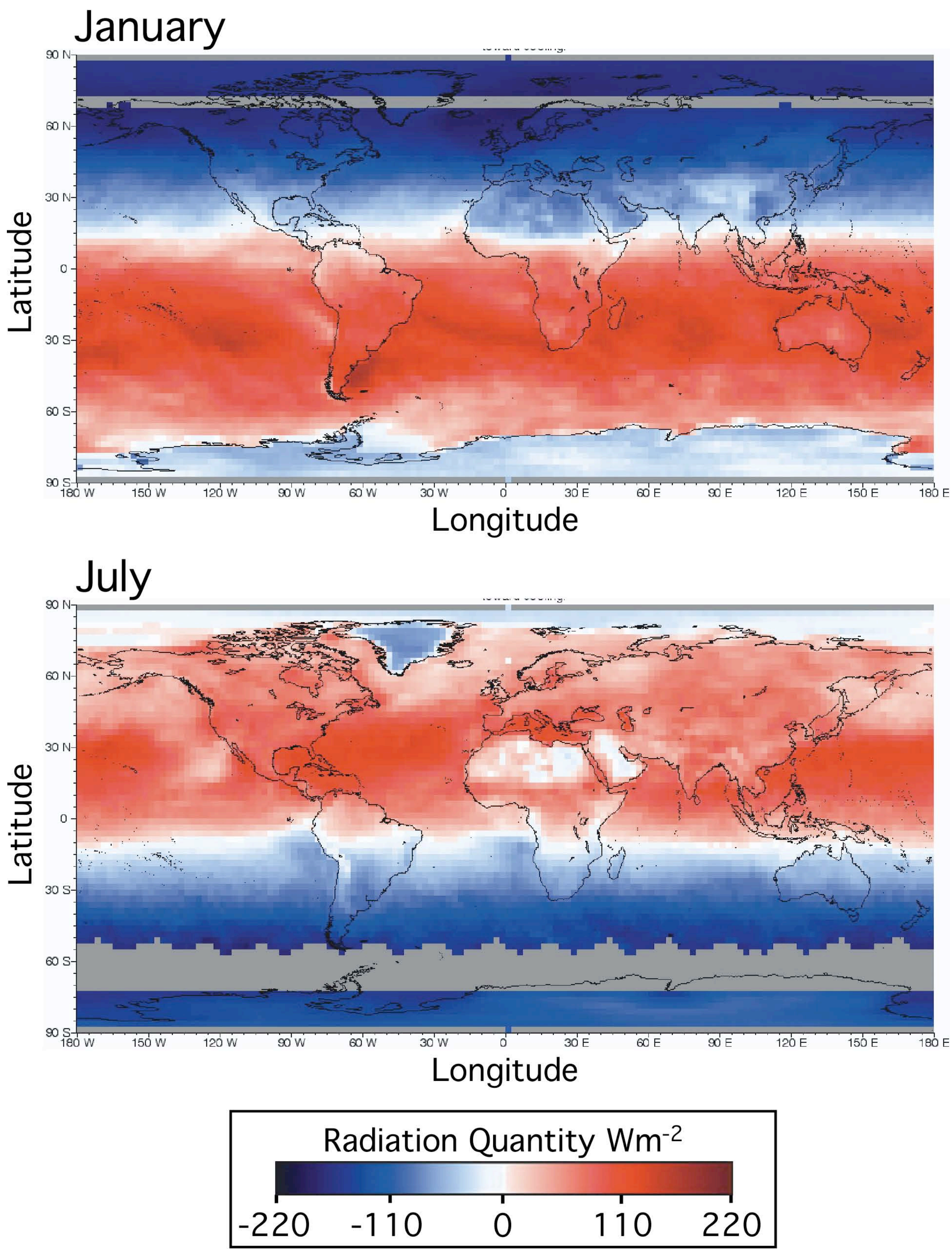

FIGURE 5.31 Net radiation of the Earth's combined surface and atmosphere for January and July 1987. Cells with missing data are colored grey. Measured by sensors aboard a variety of satellites for NASA's Earth Radiation Budget Experiment (ERBE). *(Source: NASA - Earth Radiation Budget Experiment)*

TABLE 5.4 Gases involved in the greenhouse effect: Past and present concentration and sources. Note that chlorofluorocarbons (CFCs) are not natural but produced by a variety of industrial processes. CFCs first began entering the Earth's atmosphere in the 1920s. [ppm = parts per million; ppb = parts per billion; ppt = parts per trillion].

Greenhouse Gas	Concentration Pre-Industrial	Concentration 2005	Percent Change	Anthropogenic Sources
Carbon Dioxide	280 ppm	385 ppm	38%	Biomass fires, burning fossil fuels, deforestation, and land-use change.
Methane	700 ppb	*1796 ppb	157%	Natural gas and oil extraction, biomass burning, rice cultivation, livestock, and refuse landfills.
Nitrous Oxide	270 ppb	*321 ppb	19%	Soil cultivation, fertilizers, biomass burning, and burning of fossil fuels.
Chlorofluorocarbons (CFCs)	0	*861 ppt	Not applicable.	Refrigerators, aerosol spray propellants, and cleaning solvents.
Ozone	25 ppb	34 ppb	36%	Created naturally by the action of sunlight on molecular oxygen and artificially through photochemical smog.

* These values represent an average of measurements taken at Mace Head, Ireland, a mid-latitude Northern Hemisphere site, and Cape Grim, Tasmania, a mid-latitude Southern Hemisphere site.

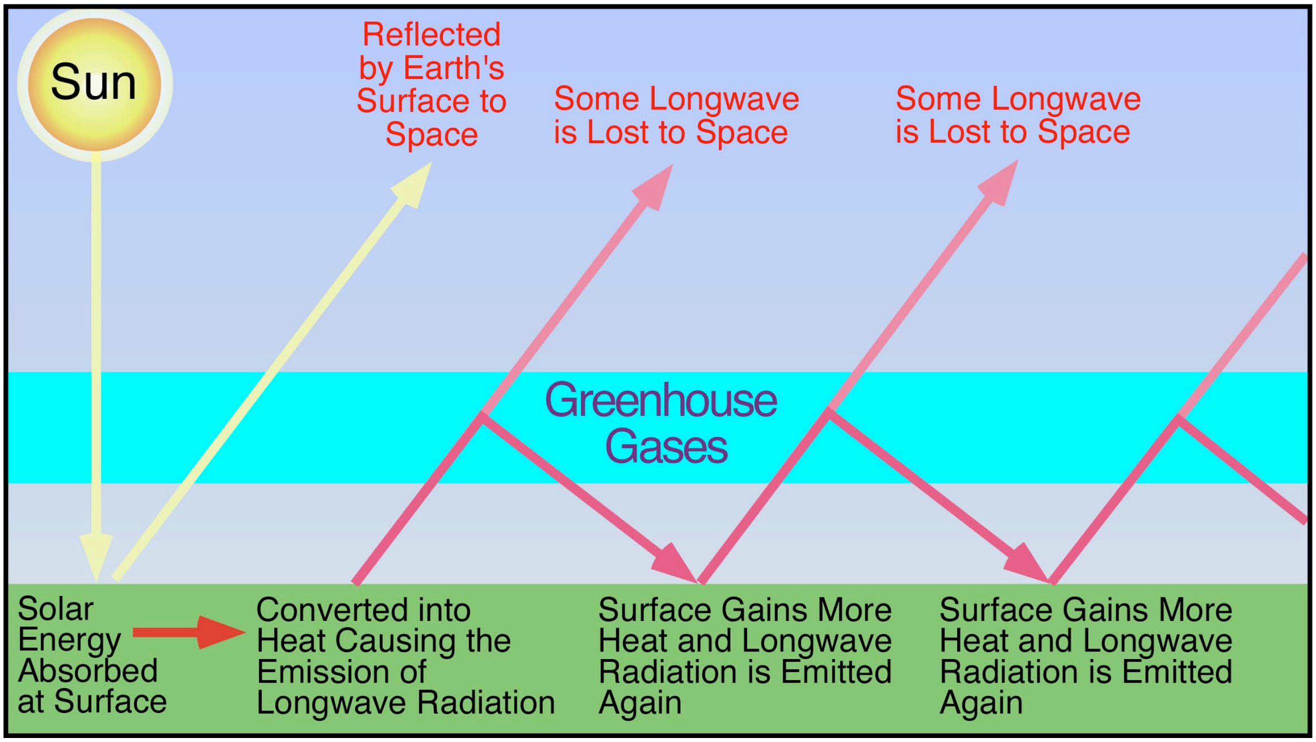

FIGURE 5.32 Earth's greenhouse effect. The diagram above illustrates the greenhouse effect. This process begins with the absorption of shortwave radiation from the sun. Absorption causes the solar energy to be converted into sensible heat at the Earth's surface. Some of this heat is transferred to the lower atmosphere by conduction and convection. After the heating of the ground and the lower atmosphere, these surfaces become radiators of infrared or longwave radiation and they begin to cool. This emission of energy is directed to space.; however, only a portion of this energy actually makes it through the atmosphere. About 90% of the longwave radiation emitted from the Earth's surface is absorbed by the atmosphere's greenhouse gases. Absorption of this energy causes heat energy to be added to the Earth's atmospheric system through the warming of greenhouse gas molecules. The greenhouse gas molecules then begin radiating longwave energy primarily back to the Earth's surface where it once again creates heat energy. The heating of the ground by the longwave radiation causes the ground surface to once again radiate, repeating the cycle described above, again and again, until no more infrared radiation is available for surface absorption. In conclusion, the net result of the greenhouse effect is an increase in the creation and storage of heat energy to the Earth's atmosphere and ground surface. *(Image Copyright: Michael Pidwirny)*

CHAPTER SUMMARY

- This chapter begins with an examination of the composition of the Earth’s atmosphere. We discovered that our planet’s atmosphere has undergone three stages of chemical evolution.
- In the first stage, the atmosphere was dominated by water vapor, carbon dioxide, and nitrogen.
- With the cooling of our planet much of the water vapor in the atmosphere condensed out. This process marks the beginning of the second stage of evolution that occurred between 4.0 and 3.3 billion years ago.
- In the last stage, life begins to significantly change the chemistry of the atmosphere through the addition of oxygen. Today, the concentration of oxygen in the atmosphere is about 21%.
- Nitrogen and oxygen together make up about 99% of the volume of the modern atmosphere. The remaining 1% consists of a mixture of many different gases.
- Several of the atmosphere’s gases have the ability to absorb infrared radiation from the Earth's surface and atmosphere giving rise to the greenhouse effect.
- Our planet’s atmosphere also contains various liquid and solid particles. These aerosols enter the air through a variety of natural and human processes.
- The atmosphere consists of a number of horizontal layers based on temperature change or chemical composition.
- There are four layers that have different temperature characteristics: troposphere, stratosphere, mesosphere, and thermosphere. Of these four, the troposphere is the most important because it is the layer where most of our planet’s weather occurs.
- We also can identify distinct atmospheric layers according to chemical composition. Some of the chemically unique zones in the Earth’s atmosphere include the homosphere, heterosphere, ionosphere, and ozone layer.
- As solar radiation passes through the atmosphere, the processes of scattering, absorption, and atmospheric reflection act to reduce the intensity of the shortwave beam received at the Earth’s surface.
- Scattering is a process where some of the rays of incoming sunlight are redirected in terms of direction of travel by atmospheric particles. This process does not alter the quality of the incoming sunlight, but it does cause about 15% of the scattered light to be redirected back to space.
- Atmospheric absorption is a process where sunlight is absorbed by an airborne substance creating heat energy. This heat energy is then converted into longwave radiation and some of this emission is lost to space.
- Atmospheric reflection acts like a mirror redirecting the light back to space. Finally, not all of the insolation received at the Earth’s surface is absorbed. Like the atmosphere, the various substances found on our planet’s surface reflect a portion of the light towards space.
- The Earth maintains an overall balance between the radiation absorbed by the surface and atmosphere and the quantity of longwave radiation emission that is sent back to space; however, this balance is not uniform geographically.
- Net radiation is a measure that describes the quantitative difference between incoming absorbed radiation with outgoing longwave emission.
- At latitudes lower than 35 degrees net radiation is positive. These positive values indicate that incoming shortwave radiation exceeds outgoing terrestrial radiation creating a surplus of energy at these latitudes.
- Negative values of net radiation exist from 35°N and S to the North and South Poles, respectively. This spatial imbalance is created because atmospheric circulation and ocean currents transfer large amounts of latent and sensible heat from the tropics to higher latitudes.
- The greenhouse effect causes the atmosphere to trap more heat energy at the Earth's surface and within the atmosphere by absorbing and re-emitting longwave energy. Of the longwave energy emitted back to space, 90% is intercepted and absorbed by greenhouse gases.
- Without the greenhouse effect the Earth's average global temperature would be -18°C (0°F), rather than the present 15°C (59°F).
- In the last few centuries, the activities of humans have directly or indirectly caused the concentration of the major greenhouse gases to increase. Scientists predict that this increase may enhance the greenhouse effect making the planet warmer.
- Some experts estimate that the Earth's average global temperature has already increased by 0.3 to 0.6°C (0.5 to 1.1°F), since the beginning of this century, because of this enhancement. Predictions of future climates indicate that by the year 2050 the Earth's global temperature may be 1 to 3°C (1.8 to 5.4°F) higher than today.

IMPORTANT TERMS

Aerosols
Albedo
Anaerobic
Angle of Incidence
Atmospheric absorption
Atmospheric reflection
Atmospheric transmission
Aurora Borealis
Aurora Australis
Backscattering
Counterradiation
Cumulus
Cumulonimbus
Deforestation
Denitrification
Diffused solar radiation
Direct solar radiation
Environmental lapse rate (ELR)
Fossil fuel
Global dimming
Global warming
Greenhouse effect
Greenhouse gas
Heterosphere
Homosphere
Ionosphere
Industrial Revolution
Insolation
Inverse Square Law
Isothermal layer
Land-use change
Longwave radiation
Mesopause
Mesosphere
Mie scattering
Net radiation
Non-selective scattering
Ozone
Ozone layer
Photo-dissociation
Photosynthesis
Planetary albedo
Rayleigh scattering
Respiration
Scattering
Shortwave radiation
Stratopause
Stratosphere
Temperature inversion
Thermosphere
Tropopause
Troposphere
Wien's Law

CHAPTER REVIEW QUESTIONS

1. Describe the three stages in the evolution of the Earth's atmosphere. What gases were dominant in each stage?
2. How are the various gases that make up today's atmosphere important to life on our planet?
3. Which gases in the atmosphere have varied in their concentration over the last 300 years? Why?
4. What are aerosols? Why do the concentrations of aerosols vary over time and space?
5. Name and describe the characteristics of the various layers found in the atmosphere according to vertical change in air temperature.
6. Describe and name the various chemical layers that exist in the Earth's atmosphere.
7. How is the incoming shortwave solar radiation from the Sun modified by the atmospheric processes of absorption, scattering, and reflection?
8. What is surface albedo? How does it vary at the Earth's surface?
9. Describe how incoming insolation is partitioned in the global shortwave radiation cascade?
10. How are longwave radiation and heat energy partitioned in the global longwave radiation cascade?
11. Describe the annual patterns of insolation received on the Earth's surface.
12. Explain the global patterns of annual outgoing surface and atmosphere longwave emission for the Earth.
13. Discuss the relationship between global patterns of annual net radiation and the transfer of latent and sensible heat by atmospheric circulation and ocean currents.
14. Describe how the greenhouse effect works? How has human activity over the last few centuries enhanced this natural process?

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