

Third Edition

Chapter 3

THE EARTH SYSTEM

Lee R. Kump

*The Pennsylvania State University
Department of Geosciences
and Earth System Science Center*

James F. Kasting

*The Pennsylvania State University
Department of Geosciences and Meteorology
and Earth System Science Center*

Robert G. Crane

*The Pennsylvania State University
Department of Geography
and Earth System Science Center*

Prentice Hall

San Francisco Boston New York
Cape Town Hong Kong London Madrid Mexico City
Montreal Munich Paris Singapore Sydney Tokyo Toronto

CONTENTS

About the Authors viii

Preface ix

Chapter 1 Global Change 1

Introduction 1

Global Change on Short Time Scales 3

■ A Closer Look: Are Hurricanes Getting Stronger with Time? 9

■ A Closer Look: The Discovery of the Antarctic Ozone Hole 12

Global Change on Long Time Scales 13

Chapter 2 Daisyworld: An Introduction to Systems 21

The Systems Approach 21

■ Thinking Quantitatively: Stability of Positive Feedback Loops 25

The Daisyworld Climate System 26

■ Useful Concepts: Graphs and Graph Making 28

External Forcing: The Response of Daisyworld to Increasing Solar Luminosity 30

Chapter 3 Global Energy Balance: The Greenhouse Effect 36

Introduction 36

Electromagnetic Radiation 37

Temperature Scales 40

Blackbody Radiation 41

Planetary Energy Balance 43

■ A Closer Look: Planetary Energy Balance 44

Atmospheric Composition and Structure 44

■ Thinking Quantitatively: How the Greenhouse Effect Works: The One-Layer Atmosphere 45

Physical Causes of the Greenhouse Effect 48

Effect of Clouds on the Atmospheric Radiation Budget 50

Introduction to Climate Modeling 52

Climate Feedbacks 53

Chapter 4 The Atmospheric Circulation System 57

The Global Circulatory Subsystems 57

The Atmospheric Circulation 58

■ A Closer Look: The Relationships between Temperature, Pressure, and Volumes—
The Ideal Gas Law 59

■ A Closer Look: How Hurricanes (Tropical Cyclones) Work 67

Global Distributions of Temperature and Rainfall 70

Chapter 5 The Circulation of the Oceans 84

Winds and Surface Currents 84

■ A Closer Look: Vorticity 90

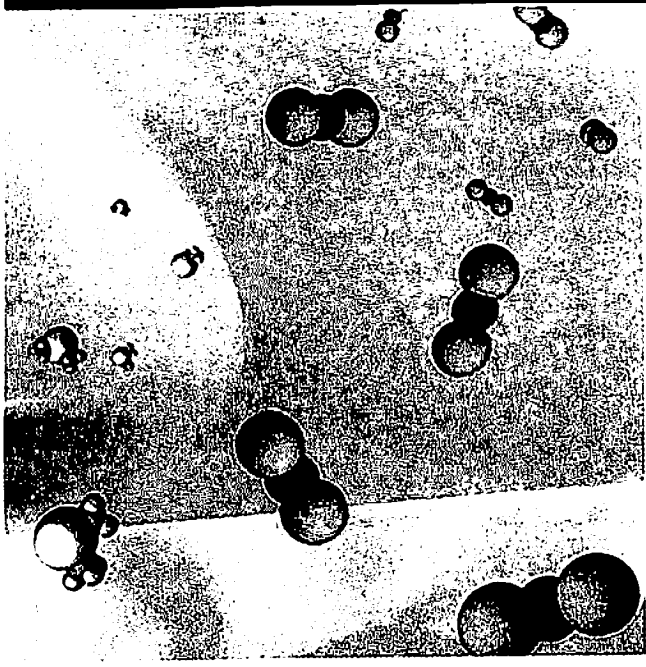
■ A Closer Look: The 1982–1983 and 1997–1998 ENSO Events 95

The Circulation of the Deep Ocean 96

■ A Closer Look: The Salt Content of the Oceans and the Age of Earth 97

■ Useful Concepts: Isotopes and Their Uses 102

■ A Closer Look: Carbon-14—A Radioactive Clock 103



CHAPTER 3

Global Energy Balance

The Greenhouse Effect

Key Questions

- What are the basic characteristics of electromagnetic radiation?
- What causes the greenhouse effect, and how is the magnitude of this effect determined?
- How do clouds affect the atmospheric radiation budget?
- What are the most fundamental feedbacks in the climate system?

Chapter Overview

Earth is heated by visible radiation from the Sun and cools by radiating infrared energy back to space. Earth's surface temperature depends on the amount of incident sunlight, the planet's reflectivity, and the greenhouse effect of its atmosphere. Certain gases in the atmosphere absorb outgoing infrared radiation and reradiate part of that energy back down to the surface. If this process did not occur, Earth's average surface temperature would be well below the freezing point of water, and life could not survive. Both the greenhouse effect and the amount of absorbed sunlight are strongly influenced by the presence of clouds. Clouds can either warm or cool the surface, depending on their altitude and thickness. Built-in feedback loops involving atmospheric water vapor and the extent of snow and ice cover are also fundamental aspects of the climate system. All these factors need to be considered in order to predict the response of Earth's climate to future increases in greenhouse gas concentrations.

INTRODUCTION

That Earth is suitable for life is largely a consequence of its temperate climate. A fundamental requirement for life as we know it is liquid water, and Earth is the only

planet in our solar system that has liquid water at its surface. Venus, our nearest neighbor toward the Sun, has an average surface temperature of 460°C (860°F), hot enough to melt lead. Mars, the closest planet away from the Sun, has an average surface temperature of -55°C (-67°F), the coldest temperatures experienced at the South Pole. Earth's average surface temperature is 15°C (59°F), the same as the mean annual temperature of San Francisco. Earth is not only habitable, it is a relatively pleasant place to live.

Why is Venus too hot, Mars too cold, and Earth just right? This question is sometimes called the "Goldilocks problem" of comparative planetology. Intuition suggests that the answer is that Earth happens to lie at the right distance from the Sun (and hence would receive exactly the right amount of sunlight), whereas Venus and Mars do not (Figure 3-1). A closer look, however, reveals that it is not just the amount of sunlight that a planet receives that determines its surface temperature. A planet's surface is also warmed by the greenhouse effect of its atmosphere. As we saw in Chapter 1, a planet's atmosphere allows sunlight to come in but slows down the rate at which heat is lost. Without this greenhouse effect, Earth's average surface temperature would be about 33°C (59°F) colder than the observed value. Earth would be an icy, desolate world.



FIGURE 3-1 Venus, Earth, and Mars, shown roughly to scale. (Source: NASA (left and center) and Photodisc/Getty Images (right).)

In this chapter, we discuss how the greenhouse effect works. We begin by considering the nature of electromagnetic radiation and why the Sun emits primarily one form of radiation (visible light), whereas Earth emits another (infrared radiation). We show that the incoming solar energy and outgoing infrared energy must be approximately in balance, and we demonstrate how this balance allows us to calculate the magnitude of the atmospheric greenhouse effect. Next, we discuss how both forms of energy are affected by atmospheric gases and by clouds, and we explain why some gases are greenhouse gases but others are not. Finally, we use the systems notation developed in Chapter 2 to introduce real climate feedback mechanisms, and we show why it is necessary to understand these feedbacks in order to estimate the climate changes that are occurring now as well as those that may occur in the future.

ELECTROMAGNETIC RADIATION

What exactly does it mean to say that Earth is heated by radiation from the Sun? From our sense of sight, we know that the Sun emits about 50% of its energy in the form of *visible light*. Let us start by considering what makes up light and other forms of *electromagnetic radiation*.

PROPERTIES OF ELECTROMAGNETIC RADIATION A physicist would describe an electromagnetic wave as a propagating disturbance consisting of oscillating (regularly fluctuating) electric and magnetic fields that are perpendicular to each other. For our purposes, we can think of **electromagnetic radiation** as a self-propagating electric and magnetic wave that is similar to a wave that moves on the surface of a pond. A wave of any form of electromagnetic radiation—such as light, ultraviolet, or infrared radiation—moves at a fixed speed c (the “speed of light”). The numerical value of c for a light wave in a vacuum is 3.00×10^8 m/s. The wave consists of a series of crests and troughs (Figure 3-2). The distance between two adjacent crests is called the **wavelength**. It is typically denoted

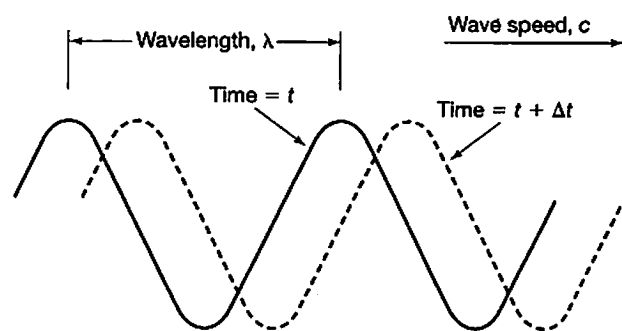


FIGURE 3-2 Simplified representation of an electromagnetic wave, illustrating the concept of wavelength. The solid curve shows the position of the wave at some time t . The dashed curve shows the wave at time $t + \Delta t$.

by the Greek letter λ (lambda). An observer standing at a fixed point in the path of the wave would be passed by a given number of crests in one second. This number is called the **frequency** of the wave. It is represented by the Greek letter ν (nu).

If we neglect complexities like polarization, an electromagnetic wave can be described by these three characteristics: speed, wavelength, and frequency. Not all of these characteristics are independent. The speed of the wave must equal the product of the number of wave crests that pass a given point each second (the frequency) and the distance between the crests (the wavelength). We can express this relationship mathematically as

$$\lambda \nu = c,$$

or equivalently as

$$\nu = \frac{c}{\lambda}.$$

The longer the wavelength of an electromagnetic wave, the lower must be its frequency, and vice versa. Conversely, the shorter the wavelength, the higher the frequency.

Photons and Photon Energy

Although we can think of electromagnetic radiation as a wave, at times it behaves more like a stream of particles. A single “particle,” or pulse, of electromagnetic radiation is referred to as a **photon**. A photon is the smallest discrete (independent) amount of energy that can be transported by an electromagnetic wave of a given frequency. The energy E of a photon is proportional to its frequency:

$$E = h\nu = \frac{hc}{\lambda},$$

where h is a constant called *Planck’s constant*, after the famous German physicist Max Planck. Its numerical value is 6.63×10^{-34} J-s (joule-seconds). Thus, high-frequency (short-wavelength) photons have high energy, and low-frequency (long-wavelength) photons have low energy. This difference in photon energy becomes important when electromagnetic radiation interacts with matter, because high-energy and low-energy photons have very different effects. High-energy photons can break molecules apart and hence cause chemical reactions to occur, whereas low-energy photons merely cause molecules to rotate faster or vibrate more strongly.

The fact that electromagnetic radiation behaves both as a particle and as a wave was one of the great discoveries of physics of the early part of the 20th century. This *wave-particle duality* is not restricted to electromagnetic waves. Rather, it is a general characteristic of matter and energy.

The Electromagnetic Spectrum

The full range of forms of electromagnetic radiation, which differ by their wavelengths (or by their frequencies), makes up the **electromagnetic spectrum** (Figure 3-3). Wavelengths in the visible range are typically measured in

nanometers (nm). One *nanometer* is one-billionth of 1 meter. **Visible radiation**, or visible light, consists of a relatively narrow range of wavelengths, from about 400 to 700 nm. Within this range, the color of the light depends on its wavelength. Anyone who has observed a rainbow has witnessed this phenomenon. The longest visible wavelengths appear to our eyes as the color red, whereas the shortest wavelengths register as blue to violet. The colors of the rainbow—in other words, the range of component wavelengths of visible light—are referred to as the **visible spectrum**. The term “spectrum” indicates that the light has been separated into its component wavelengths.

About 40% of the Sun’s energy is emitted at wavelengths longer than the visible limit of 700 nm in a region referred to as the *infrared* region of the electromagnetic spectrum. Wavelengths of **infrared (IR) radiation** are significantly longer than those of visible light. Hence, it is convenient to keep track of them in units called *micrometers* (μm) rather than in nanometers. One micrometer (formerly called *micron*) equals one-millionth (10^{-6}) of 1 meter. So, 1 μm equals 1000 nm. Infrared wavelengths range from 0.7 to 1000 μm . At even longer wavelengths within the electromagnetic spectrum, radiation is transmitted in the form of *microwaves* and *radio waves*. Radio waves can have wavelengths of many meters.

About 10% of the Sun’s energy is emitted at wavelengths shorter than those of visible light as **ultraviolet (UV) radiation**. Wavelengths of the ultraviolet region extend from 400 nm down to about 10 nm. At shorter wavelengths are *X-rays* and *gamma rays*. These high-energy forms of electromagnetic radiation have little effect on our story here, but they do affect the chemistry of the uppermost atmosphere. X-rays, of course, are also used in medicine because they can penetrate skin and muscle tissue and allow us to see the underlying bones.

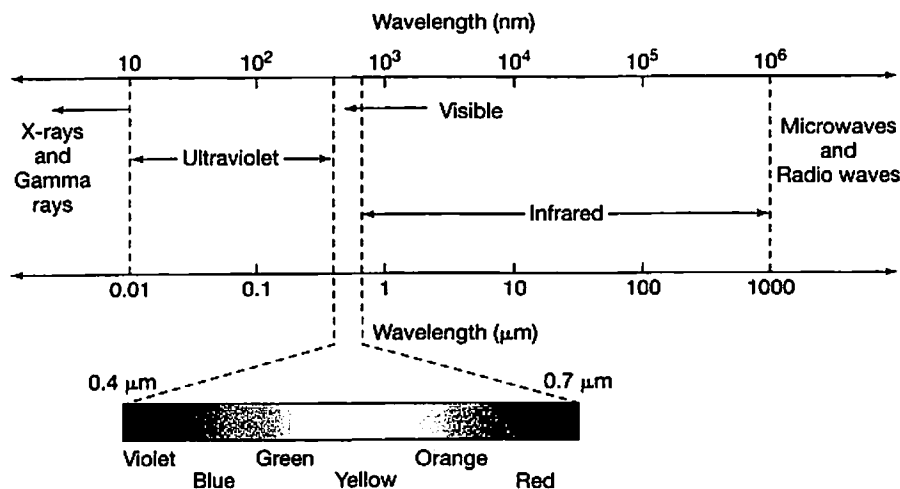


FIGURE 3-3 [See color section]
The electromagnetic spectrum.

The regions of the electromagnetic spectrum that are most important to climate are the visible and the infrared. The Sun emits energy in both of these spectral regions. Earth, as we shall see, emits primarily in the infrared. Solar ultraviolet radiation also affects the Earth system significantly by driving atmospheric chemistry. In addition, UV radiation would be lethal to most forms of life were it not almost totally blocked out by oxygen and ozone in Earth's atmosphere.

Flux

We will need one other basic concept from electromagnetic theory in order to proceed: the notion of flux. In general terms, *flux* (F) is the amount of energy or material that passes through a given area (perpendicular to that area) per unit time. In terms of fluid flow, for example, the flux is the volume of fluid that flows perpendicularly into or out of a unit area per unit time. Applied to electromagnetic radiation, *flux* is the amount of energy (or number of photons) in an electromagnetic wave that passes perpendicularly through a unit surface area per unit time.

To demonstrate the concept of flux, let us consider the light given off by an electric lightbulb. A typical, small

lightbulb is labeled "60 watts." A *watt* (W) is a unit of *power*—formally, the rate at which work is done; informally, the intensity of the bulb in the SI system. One watt equals one joule (J) per second. Suppose that a person is standing some distance from such a lightbulb and holds up a sheet of paper directly facing the light (Figure 3-4a). The paper is illuminated by radiant energy from the bulb. The radiation crosses the paper perpendicularly from the lightbulb at a certain flux, or intensity per unit area. That flux is measured in watts per square meter (W/m^2). The magnitude of the flux depends on how far from the lightbulb the person is standing, but it does *not* depend on how big the paper is because flux is defined as the intensity per *unit* area.

The fact that flux is measured perpendicular to the direction the wave is traveling is important. Suppose that the person is holding the paper at an angle, rather than perpendicularly, to the light (Figure 3-4b). Although the total area of the paper remains the same, the flux of radiant energy reaching its surface is less because less radiation strikes a given unit area. This simple concept has direct, familiar consequences for Earth's climate. The polar regions are cooler than the tropics because the Sun's rays strike the ground at a higher angle at the poles. Summer temperatures are warmer than winter temperatures

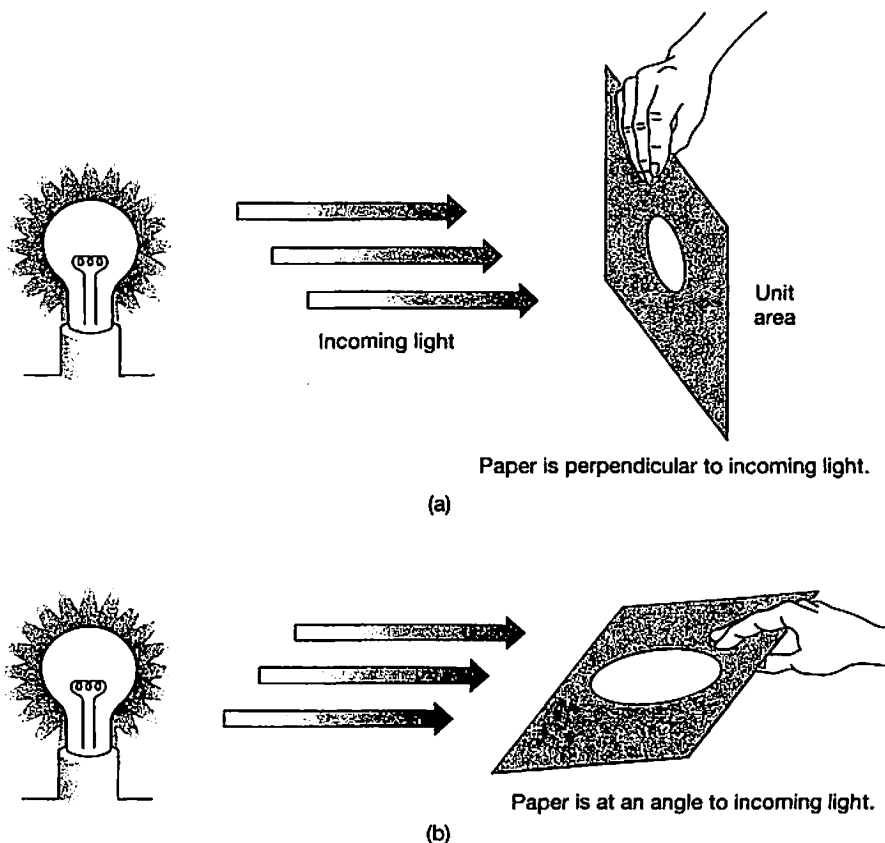


FIGURE 3-4 Schematic diagram of the concept of flux. The flux of radiation into the paper is reduced when the paper is tilted at an angle to the incoming light.

because the Sun is higher in the sky during summer. We discuss these fundamental features of climate at greater length in Chapter 4. For now, we simply need to understand the concept of flux.

The Inverse-Square Law

Figure 3-4 demonstrates that the flux of radiant energy from a lightbulb depends on how far away the observer (the person holding the paper) is standing. Likewise, the flux of solar energy decreases as distance from the Sun increases; that is why Venus is illuminated more strongly than Earth. The rate at which this *solar flux* decreases with increasing distance is described by a simple relationship. This relationship, called an **inverse-square law**, is expressed mathematically as

$$S = S_0 \left(\frac{r_0}{r} \right)^2,$$

where S represents the solar flux at some distance r from the source, and S_0 represents the flux at some reference distance r_0 (Figure 3-5).

The inverse-square law has a straightforward physical interpretation: If we double the distance from the source to the observer, the intensity of the radiation decreases by a factor of $(1/2)^2$, or $1/4$. Similarly, if we reduce the distance from the source to the observer by a factor of 3, the radiation intensity increases by a factor of 3^2 , or 9.

As an example, consider a hypothetical planet, Planet X, located twice as far from the Sun as is Earth. What would be the solar flux hitting Planet X? Refer to Figure 3-5, and let the Sun be at the center of the two circles. Also, let the inner circle represent Earth's orbit and the outer circle represent the orbit of Planet X. Then r_0 is the average distance from Earth to the Sun, which is 149,600,000 km, defined as one *astronomical unit* (AU), and S_0 is the solar flux at Earth's orbit, 1366 W/m^2 . (The value of S_0 is determined by satellite measurements.) So

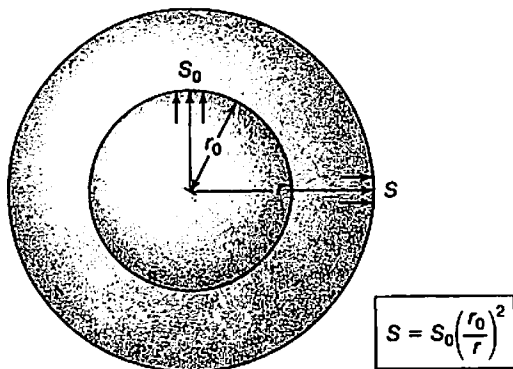


FIGURE 3-5 Diagram illustrating the inverse-square law.

according to the inverse-square law, for this example $r = 2 \text{ AU}$, the solar flux incident at Planet X is

$$\begin{aligned} S &= 1366 \text{ W/m}^2 \left(\frac{1 \text{ AU}}{2 \text{ AU}} \right)^2 \\ &= 341.5 \text{ W/m}^2. \end{aligned}$$

We would have gotten precisely the same answer if we had expressed the distances in kilometers, but the arithmetic would have been harder.

The inverse-square law is of fundamental importance to the study of planetary climates. It allows us to determine quantitatively why Earth's climate differs from that of Venus and Mars. It also plays a crucial role in our understanding of the causes of the glacial-interglacial cycles of the last 3 million years of Earth's history. As we will see later, small variations in the shape of Earth's orbit, combined with the inverse-square relationship between the distance from Earth to the Sun and solar flux, cause large changes in the climate of the polar regions and in the size and extent of the polar ice sheets.

TEMPERATURE SCALES

To understand climate, which is the prevailing weather patterns of a planet or region over time, we must first understand the concept of temperature. *Temperature* is a measure of the internal heat energy of a substance. Heat energy, in turn, is determined by the average rate of motion of individual molecules in that substance. For a solid, these motions consist of regular vibrations, whereas for a gas or liquid they are just random movements of molecules. The faster the molecules in a substance move, the higher its temperature.

Most areas of the world measure temperature (T) by the *Celsius* (formerly, *centigrade*) scale, which is measured in degrees Celsius ($^{\circ}\text{C}$) and is part of the SI system of units. In the United States, temperature is typically measured in degrees *Fahrenheit* ($^{\circ}\text{F}$). Scientists, particularly those studying climate, often use the **Kelvin (absolute) temperature scale**, measured in units called *kelvins* (K). (Note that temperatures in the Kelvin scale are given simply as kelvins, not as degrees Kelvin.)

The Celsius temperature scale is defined in terms of the freezing and boiling points of water at sea level (Table 3-1). At sea-level pressure, the freezing point is 0°C , and the boiling point is 100°C . Atmospheric pressure decreases with altitude, as we will see later in the chapter, so it makes a difference where the boiling point is determined. Water boils when its vapor pressure exceeds the overlying atmospheric pressure. Thus, the boiling point decreases with altitude. (This is why it takes longer to hard-boil an egg when you are camping in the mountains.)

TABLE 3-1 Freezing and Boiling Points of Water on Various Temperature Scales

Temperature Scale	Freezing Point	Boiling Point (at sea level)
Fahrenheit	32°	212°
Celsius	0°	100°
Kelvin (absolute)	273.15	373.15

The boiling water is not as hot, so it takes longer to cook the egg.)

The Fahrenheit temperature scale was originally defined on the basis of the temperature of a mixture of snow and table salt (0°F) and the temperature of the human body (about 100°F). Like the Celsius scale, it is defined today in terms of the physical properties of water: The freezing point is 32°F, and the boiling point is 212°F. The following relations allow us to convert temperatures between the Celsius and Fahrenheit scales:

$$T(^{\circ}\text{C}) = \frac{T(^{\circ}\text{F}) - 32}{1.8}$$

$$T(^{\circ}\text{F}) = [T(^{\circ}\text{C}) \times 1.8] + 32.$$

Note that converting a temperature *change* from one system of units to the other is easier, because the effect of the different zero points is removed. Thus, a temperature change of 1°C is equal to a change of 1.8°F. Conversely, a change of 1°F is equal to a change of 0.5556 (=1/1.8)°C.

Absolute temperature—that is, temperature on the Kelvin scale—is defined in terms of the heat energy of a substance relative to the energy it would have at a temperature of absolute zero. At *absolute zero*, the

molecules of a substance are at rest (or, more precisely, are in their lowest possible energy state). A temperature change of 1 K is equal to a temperature change of 1°C. The zero point of the Kelvin scale is, however, lower than that of the Celsius scale by 273.15°. To convert temperature in degrees Celsius to kelvins, we use the following equation:

$$T(\text{K}) = T(^{\circ}\text{C}) + 273.15.$$

Thus, a temperature of absolute zero corresponds to a Celsius reading of -273.15°C.

BLACKBODY RADIATION

To fully understand the greenhouse effect, we need one final concept from the world of physics: the concept of blackbody radiation. A **blackbody** is something that emits (or absorbs) electromagnetic radiation with 100% efficiency at all wavelengths. Consider a cast-iron ball (Figure 3-6). At room temperature, the ball looks black because it absorbs most of the light incident on it and gives off little visible radiation of its own. If we heat the ball, however, it begins to glow a dull red. If we heat the ball further, it eventually glows white hot because it radiates at all visible wavelengths. Recall that white light is a mixture of all the colors of the spectrum.

The radiation emitted by a blackbody is called **blackbody radiation**. It has a characteristic wavelength distribution that depends on the body's absolute temperature. This distribution can be described mathematically by a relation called the Planck function. The *Planck function* relates the intensity of radiation from a blackbody to its wavelength, or frequency. When shown graphically, this relation is also known as the *blackbody radiation curve*

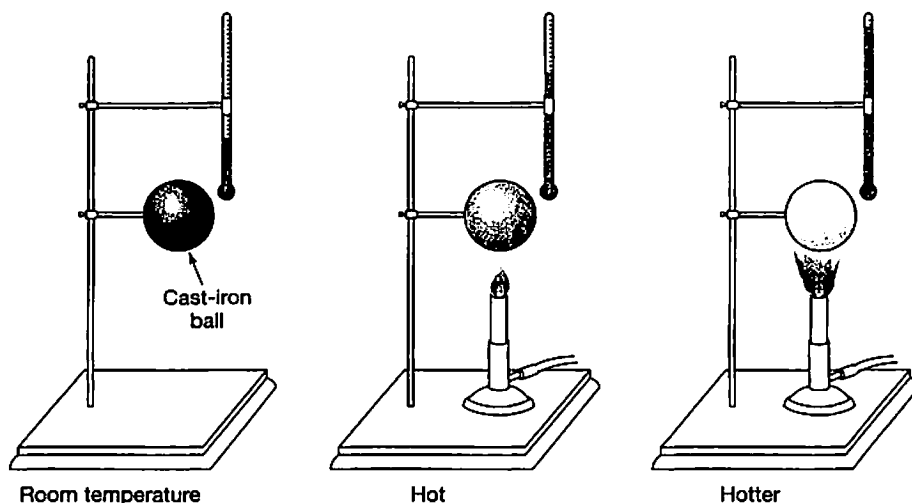


FIGURE 3-6 Change in emitted radiation by a blackbody as it is warmed.

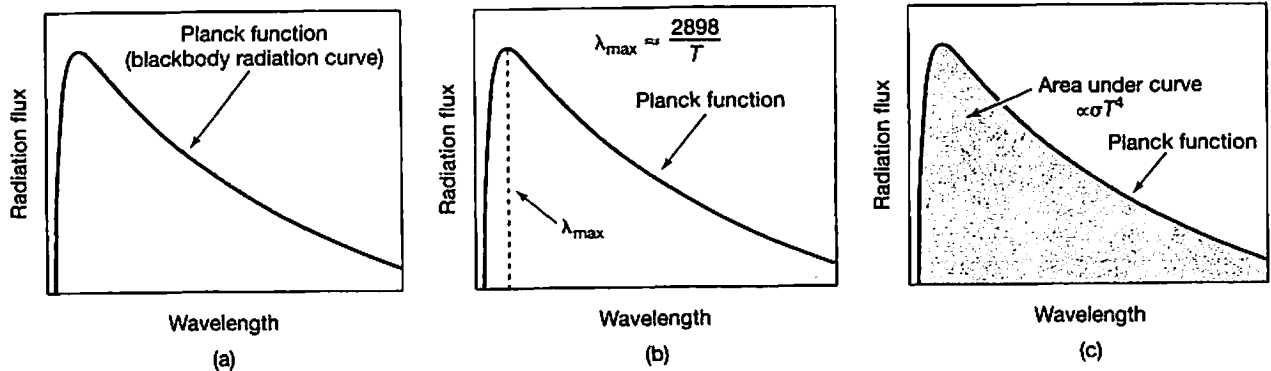


FIGURE 3-7 (a) The Planck function, or blackbody radiation curve; (b) Wien's law; (c) the Stefan-Boltzmann law.

(Figure 3-7a). The Planck function itself is mathematically complicated and is beyond the scope of our discussion here. We can, however, use this relation to derive two simpler rules that are fundamental to an understanding of climate.

Wien's Law

The first rule derived from the Planck function that will assist us in studying climate is called Wien's law. **Wien's law** states that the flux of radiation emitted by a blackbody reaches its peak value at a wavelength λ_{\max} , which depends inversely on the body's absolute temperature. According to this rule, hotter bodies emit radiation at shorter wavelengths than do colder bodies. Wien's law may be written as

$$\lambda_{\max} \approx \frac{2898}{T},$$

where T is the temperature in kelvins and λ_{\max} is the wavelength of maximum radiation flux in micrometers (Figure 3-7b).

Wien's law allows us to understand why the Sun's radiation peaks in the visible part of the electromagnetic spectrum and why Earth radiates at infrared wavelengths. The Sun emits most of its energy, including visible radiation, from its surface layer, called the **photosphere**. The temperature of the photosphere is about 5780 K. Thus, according to Wien's law, the Sun's radiation flux should maximize at $2898 \mu\text{m}/5780 \approx 0.5 \mu\text{m}$, or 500 nm (Figure 3-8). This is right in the middle of the visible spectrum. (The fact that the solar radiation flux peaks in the visible is no coincidence; our sense of vision presumably evolved as it did to take advantage of the solar spectrum.) Earth, meanwhile, has a surface temperature of about 288 K, so its radiation peaks at $2898 \mu\text{m}/288 \approx 10 \mu\text{m}$ —well into the infrared range. In reality, neither Earth nor the Sun is a perfect blackbody, so their emitted radiation flux is not exactly described by the Planck function. Nevertheless, Wien's

law is still useful in predicting the wavelength at which most of their radiant energy is emitted.

The Stefan-Boltzmann Law

A second rule derived from the Planck function that will prove useful in climate studies is called the Stefan-Boltzmann law. The **Stefan-Boltzmann law** states that the energy flux emitted by a blackbody is related to the fourth power of the body's absolute temperature:

$$F = \sigma T^4,$$

where T is the temperature in kelvins and σ (the lowercase Greek letter sigma) is a constant with a numerical value of $5.67 \times 10^{-8} \text{ W/m}^2/\text{K}^4$. The total energy flux per unit area is proportional to the area under the blackbody radiation curve (Figure 3-7c).

As an example of how the Stefan-Boltzmann law can be applied, consider a hypothetical star that has a surface temperature twice that of the Sun. (We shall use stars

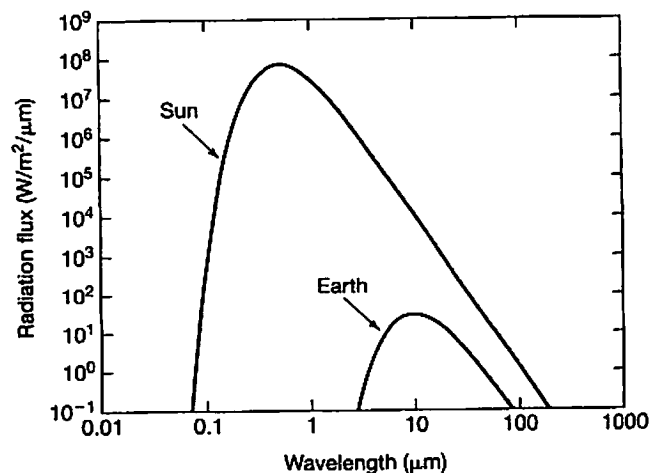


FIGURE 3-8 Blackbody emission curves for the Sun and Earth. The Sun emits more energy per unit area at all wavelengths.

rather than planets in this example because the radiation emitted from stars is more nearly approximated as black-body radiation.) Our Sun has a surface temperature of about 5780 K, so the energy flux per unit area is

$$F_{\text{Sun}} = \sigma(5780 \text{ K})^4 \approx 6.3 \times 10^7 \text{ W/m}^2.$$

The other star releases energy at a rate of

$$\begin{aligned} F_{\text{star}} &= \sigma(2 \times 5780 \text{ K})^4 \\ &= 2^4 \times \sigma(5780 \text{ K})^4 \\ &= 16 F_{\text{Sun}}. \end{aligned}$$

Thus, the amount of energy released per unit area per unit time by the hot star is 2^4 , or 16, times greater than that released by the Sun. Evidently, the amount of radiation emitted by a blackbody is a very sensitive function of its temperature.

PLANETARY ENERGY BALANCE

We now have all the tools necessary to analyze Earth's average climate in a quantitative manner. What we need to do next is put them together. The principle that we will apply is that of *energy balance*. To a first approximation, the amount of energy emitted by Earth must equal the amount of energy absorbed. In reality, this cannot be exactly true; if it were, Earth's average surface temperature would never change. We showed in Chapter 1 that the average surface temperature is changing—specifically, it is getting warmer. But it is getting warmer precisely because Earth's energy budget is slightly out of balance: The flux of incoming solar energy exceeds the outgoing IR flux by an almost imperceptible amount (a few hundredths of a percent). The imbalance may be caused by the increase in CO_2 and other greenhouse gases in the atmosphere, or it may be caused by natural fluctuations within the climate system. The latest evidence indicates strongly that the increase in greenhouse gases is dominating this change. When the climate system eventually reaches *steady state*, that is, when the surface temperature stops changing, the amount of energy going out will exactly equal the amount of energy coming in. But this will not happen until greenhouse gas levels are stabilized and the climate system is allowed sufficient time (at least 20 additional years) to come to equilibrium.

Physically, Earth's surface temperature depends on three factors: (1) the solar flux available at the distance of Earth's orbit, (2) Earth's reflectivity, and (3) the amount of warming provided by the atmosphere (i.e., the greenhouse effect). The solar flux, S , as mentioned earlier, is the amount of solar energy reaching the top of Earth's atmosphere. Not all this energy is absorbed, however. About 30% of the incident energy is reflected back to space, mostly by clouds. As we saw in Chapter 2, the reflectivity

of a planet is called its *albedo*. It is usually expressed as the fraction of the total incident sunlight that is reflected from the planet as a whole. We shall designate albedo by the letter A .

To calculate the magnitude of the third factor, the greenhouse effect, it is convenient to treat Earth as a blackbody even though this is not exactly true. (As we discuss later, the atmosphere radiates and absorbs energy better at some wavelengths than at others because of the presence of gases such as CO_2 and H_2O .) We do this by defining a quantity T_e that represents the **effective radiating temperature** of the planet. This temperature is the temperature that a true blackbody would need to radiate the same amount of energy that Earth radiates. With this definition in place, we can use the Stefan-Boltzmann law to calculate the energy emitted by Earth. By balancing the energy emitted with the energy absorbed, we obtain the following formula (see the Box "A Closer Look: Planetary Energy Balance"):

$$\sigma T_e^4 = \frac{S}{4} (1 - A).$$

This formula expresses the planetary energy balance between outgoing infrared energy and incoming solar energy.

Magnitude of the Greenhouse Effect

What is the significance of the effective radiating temperature? We can think of this quantity as the temperature at the height in the atmosphere from which most of the outgoing infrared radiation derives (see "Critical-Thinking," Problem 4). We can also think of it as the average temperature that Earth's surface would reach if the planet had no atmosphere (assuming that the albedo remained constant). To get a better understanding, let us calculate its value for the present Earth. We can solve the planetary energy balance equation for T_e by dividing both sides of the equation by σ and then taking the fourth root of each side:

$$T_e = \sqrt[4]{\frac{S}{4\sigma} (1 - A)}.$$

If we insert the known values of S (1366 W/m^2), A (30%, or 0.3), and σ ($5.67 \times 10^{-8} \text{ W/m}^2/\text{K}^4$), we get $T_e \approx 255 \text{ K}$. Thus, Earth's effective radiating temperature is a relatively chilly -18°C , or 0°F .

We saw earlier, however, that the actual mean surface temperature of Earth, T_s , is 288 K , or about 15°C . The difference between the actual surface temperature and the effective radiating temperature is caused by the greenhouse effect of Earth's atmosphere. We can represent this mathematically by letting

$$\Delta T_g = T_s - T_e.$$

A CLOSER LOOK

Planetary Energy Balance

The derivation of the planetary energy balance equation is not difficult, but it does require that we consider the geometry of the Earth–Sun system. The starting point for the derivation is the relation

energy emitted by Earth = energy absorbed by Earth.

Let us first calculate the energy emitted by Earth. If we treat Earth as a blackbody with an effective radiating temperature T_e , the Stefan–Boltzmann law tells us that the energy emitted per unit area must be equal to σT_e^4 . Earth radiates over its entire surface area, $4\pi R_{\text{Earth}}^2$, where R_{Earth} represents Earth's radius (Box Figure 3-1). Thus, the total energy emitted by Earth is

$$\text{energy emitted} = 4\pi R_{\text{Earth}}^2 \times \sigma T_e^4.$$

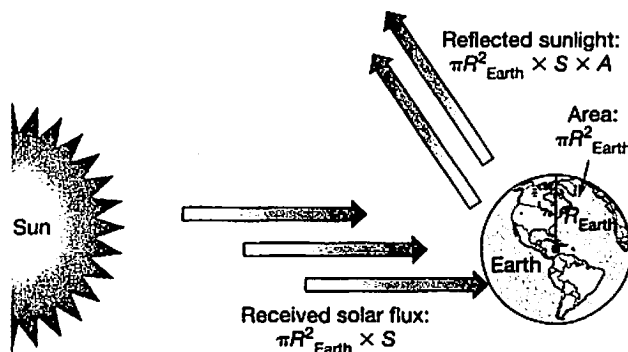
Now, let us calculate the energy absorbed by Earth. From the Sun, Earth would look like a circle with radius R_{Earth} and area πR_{Earth}^2 . Note that it is the area of Earth projected against the Sun's rays that enters here, not half of the surface area of Earth. (Half of Earth's surface area would be $2\pi R_{\text{Earth}}^2$, but the Sun's rays do not strike all of this area perpendicularly.) The total energy intercepted must be equal to the product of Earth's projected area and the solar flux (S), or $\pi R_{\text{Earth}}^2 S$. The reflected energy is equal to this incident energy times the albedo (A). The difference between these two quantities is the energy absorbed by Earth:

$$\begin{aligned} \text{energy absorbed} &= \text{energy intercepted} - \text{energy reflected} \\ &= \pi R_{\text{Earth}}^2 S - \pi R_{\text{Earth}}^2 SA \\ &= \pi R_{\text{Earth}}^2 S(1 - A). \end{aligned}$$

where ΔT_g is the magnitude of the greenhouse effect. Thus, $\Delta T_g = 15^\circ\text{C} - (-18^\circ\text{C}) = 33^\circ\text{C}$. A heuristic (*i.e.*, somewhat simplified) explanation for what causes this greenhouse warming is given in the box titled “Thinking Quantitatively: How the Greenhouse Effects Works: The One-Layer Atmosphere.”

To place this value in context, we can carry out similar calculations for Venus and Mars from known data of the albedos, surface temperatures, and orbital distances of these planets. (See “Critical-Thinking,” Problem 2.) The results show that the solution to the Goldilocks problem posed at the beginning of this chapter is more complicated than we might have guessed. Evidently, a planet's greenhouse effect is at least as important in determining that planet's surface temperature as is its distance from the Sun.

We can also apply the planetary energy balance equation to the faint young Sun paradox mentioned in Chapter 1. Recall that solar luminosity, and thus S , is



BOX FIGURE 3-1 The amount of sunlight received by and reflected by Earth.

All that remains is for us to equate the outgoing and incoming energy. Using the expressions just calculated, we get

$$4\pi R_{\text{Earth}}^2 \times \sigma T_e^4 = \pi R_{\text{Earth}}^2 S(1 - A).$$

Canceling out πR_{Earth}^2 on both sides of this equation and dividing both sides by 4, we obtain the desired equation,

$$\sigma T_e^4 = \frac{S}{4} (1 - A).$$

estimated to have been 30% lower early in the solar system's history. It is easy to demonstrate that Earth's average surface temperature would have been below the freezing point of water under such circumstances, if the planetary albedo and the atmospheric greenhouse effect had remained unchanged (see “Critical-Thinking,” Problem 5). We have already seen, though, that the early Earth had both liquid water and life on its surface. In later chapters, we discuss ways to resolve this apparent paradox.

ATMOSPHERIC COMPOSITION AND STRUCTURE

Atmospheric Composition

To understand the greenhouse effect in more detail, along with other aspects of climate and Earth's radiation budget, we must learn a few fundamental facts about the composition and structure of Earth's atmosphere. Table 3-2

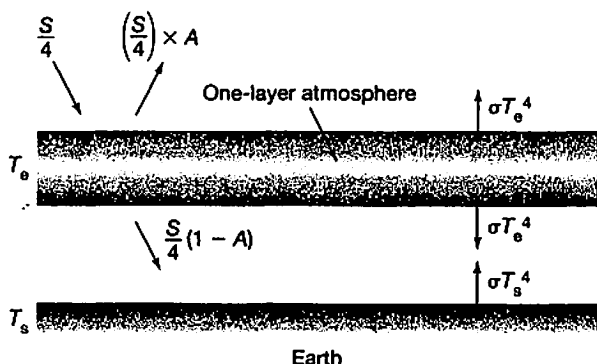
THINKING QUANTITATIVELY

How the Greenhouse Effect Works: The One-Layer Atmosphere

Although we calculated that Earth's greenhouse effect provides 33°C of surface warming, our method of obtaining this result (by subtracting the calculated effective radiating temperature from the observed mean surface temperature) provides little insight into the physical mechanism that causes the warming. We can remedy this by doing a simple calculation that demonstrates how the greenhouse effect actually works.

Suppose we treat the atmosphere as a single layer of gas and that this gas absorbs (and reemits) all of the infrared radiation incident on it (Box Figure 3-2). Let us assume that it absorbs and emits infrared radiation equally well at all wavelengths, so that we can treat it as a blackbody, and that it has an albedo A in the visible spectrum, just like that of the real Earth. What are the temperatures of the gas layer and of the surface beneath it? We will call the layer temperature T_e and the surface temperature T_s , because these quantities are exactly analogous to those discussed in the text.

We can determine the values of T_e and T_s by balancing the energy absorbed and emitted by both the surface and the one-layer atmosphere. Let the amount of sunlight striking the planet be equal to $S/4$ (the globally averaged solar flux). The surface absorbs an amount of sunlight equal to $S/4 \times (1 - A)$, along with a flux of downward infrared radiation from the atmosphere equal to σT_e^4 . The atmosphere absorbs an amount of upward infrared radiation from the ground equal to σT_s^4 , and it emits infrared radiation in both the upward and downward directions at a rate of σT_e^4 . (The real atmosphere also absorbs some of the incoming solar radiation, but we ignore that complication here.) Thus, we can write the overall energy balance in the form of two equations:



BOX FIGURE 3-2 The greenhouse effect of a one-layer atmosphere.

For the surface,

$$\sigma T_s^4 = \frac{S}{4}(1 - A) + \sigma T_e^4;$$

for the atmosphere,

$$\sigma T_s^4 = 2\sigma T_e^4.$$

(The factor 2 in the second equation arises because the atmosphere radiates in both the upward and downward directions.) If we now substitute the second equation into the left-hand side of the first equation and subtract σT_e^4 from both sides, we obtain

$$\sigma T_e^4 = \frac{S}{4}(1 - A),$$

which is just the familiar energy-balance formula. But dividing the atmospheric energy-balance equation by σ and then taking the fourth root of both sides yields an additional result:

$$T_s = 2^{1/4} T_e.$$

Thus, the surface temperature is higher than the one-layer-atmosphere temperature by a factor of the fourth root of 2, or about 1.19. For $T_e = 255$ K, as on Earth at present, we get $T_s = 303$ K, and we calculate a greenhouse effect of

$$\Delta T_g = T_s - T_e = 48 \text{ K}.$$

This is higher than the actual greenhouse effect on Earth by about 15 K.

This example is not meant to be realistic. The real atmosphere is not perfectly absorbing at all infrared wavelengths, so some of the outgoing IR radiation from the surface leaks through to space. This effect tends to make ΔT_g smaller. Conversely, a more accurate calculation would subdivide the atmosphere into a number of different layers. Including more layers tends to make ΔT_g bigger and is the reason why a thick atmosphere, like that of Venus, can produce a really huge amount of surface warming. The calculation does, however, illustrate the basic nature of the greenhouse effect: By absorbing part of the infrared radiation radiated upward from the surface and reemitting it in both the upward and downward directions, the atmosphere allows the surface to be warmer than it would be if the atmosphere were not present.

Table 3-2 Major Constituents of Earth's Atmosphere Today

Name and Chemical Symbol	Concentration (% by volume)
Nitrogen, N ₂	78
Oxygen, O ₂	21
Argon, Ar	0.9
Water vapor, H ₂ O	0.00001 (South Pole)–4 (tropics)
Carbon dioxide, CO ₂	0.039*

*In 2008.

lists the main constituents of Earth's present atmosphere and their relative abundances.

As Table 3-2 indicates, the three most abundant constituents of our atmosphere are nitrogen, oxygen, and argon. Nitrogen is a relatively *inert* (chemically unreactive) gas, but when split into its constituent atoms, it plays an important role in biological cycles. Oxygen, which is highly reactive, is the essential gas that all animals must breathe; it is required by many other life-forms as well. Argon is almost completely inert; it is the product of the radioactive decay of potassium, K, in Earth's interior. These three constituents—nitrogen, oxygen, and argon—are not greenhouse gases. In other words, they do not contribute to Earth's greenhouse effect.

Although they appear at the bottom of Table 3-2, water vapor and carbon dioxide are two of the most important atmospheric constituents. Besides being directly used by organisms, they are also strong greenhouse gases. We will soon see what makes a particular gas a greenhouse gas.

In addition to the major constituents listed in Table 3-2, Earth's atmosphere also contains a number of minor (or "trace") constituents that affect climate. The most important of these are methane, nitrous oxide, ozone, and freons. Their concentrations are generally much lower than those of the major constituents. Despite their low concentrations, these trace gases are important greenhouse gases. Table 3-3 lists the major greenhouse gases. (Note that water vapor and carbon dioxide are

Table 3-3 Important Atmospheric Greenhouse Gases

Name and Chemical Symbol	Concentration (ppm by volume)
Water vapor, H ₂ O	0.1 (South Pole)–40,000 (tropics)
Carbon dioxide, CO ₂	390
Methane, CH ₄	1.7
Nitrous oxide, N ₂ O	0.3
Ozone, O ₃	0.01 (at the surface)
Freon-11, CCl ₃ F	0.00026
Freon-12, CCl ₂ F ₂	0.00048

repeated here.) It is convenient to keep track of these gases in units of parts per million (ppm), which we defined in Chapter 1. Take a moment to convince yourself that the 0.00001–4% value of water vapor and the 0.039% value of CO₂ given in Table 3-2 are equivalent to the 0.1–40,000 ppm and 390 ppm values of water vapor and of CO₂ given in Table 3-3.

Table 3-3 is by no means a complete list of greenhouse gases. Several other gases affect climate to some extent or are otherwise important in atmospheric chemistry. The gases listed in Table 3-3, however, are the ones that are most important to the modern problem of global warming, and hence they are the ones on which we focus.

Atmospheric Structure

HOW ATMOSPHERIC PRESSURE VARIES WITH ALTITUDE Other characteristics of Earth's atmosphere that influence climate and the radiation budget are its pressure and temperature structure. *Pressure* may be defined as the force per unit area exerted by a gas or liquid on some surface with which it is in contact. The pressure exerted by the atmosphere at sea level is defined as one *atmosphere* (atm). A pressure of 1 atm is equivalent to about 15 lb/in² in the English system and to 1.013 *bar*, or 1013 *millibars* (mbar), in the metric system. (The pressure unit in the SI system is the *pascal* /Pa, but this unit is cumbersome in atmospheric work: 1 Pa = 1 × 10⁻⁵ bar ≈ 9.9 × 10⁻⁶ atm.) An instrument used to measure atmospheric pressure is called a *barometer*, the name of which derives from the metric unit of measure "bar."

At higher levels in the atmosphere, the pressure decreases markedly (Figure 3-9a). This change in pressure is what makes your ears pop in an airplane. (The cabin is pressurized, or the popping would be much worse.) The decrease by altitude follows the *barometric law*, which states that atmospheric pressure decreases by about a factor of 10 for every 16-km increase in altitude. Thus, the pressure is about 0.1 bar at 16 km above the surface, 0.01 bar at 32 km, and so on. In more precise terms, the barometric law says that pressure decreases *exponentially* with altitude. Note from Figure 3-9a that the exponential decrease in pressure appears almost like a straight line when pressure is plotted on a logarithmic scale. The slight deviation from linearity is caused by the variation in temperature with altitude (discussed next). Pressure decreases faster with height in regions where the air is colder.

HOW ATMOSPHERIC TEMPERATURE VARIES WITH ALTITUDE The vertical temperature structure of the atmosphere is more complicated than the vertical pressure structure (Figure 3-9b). This temperature profile is the basis for distinguishing four regions within Earth's

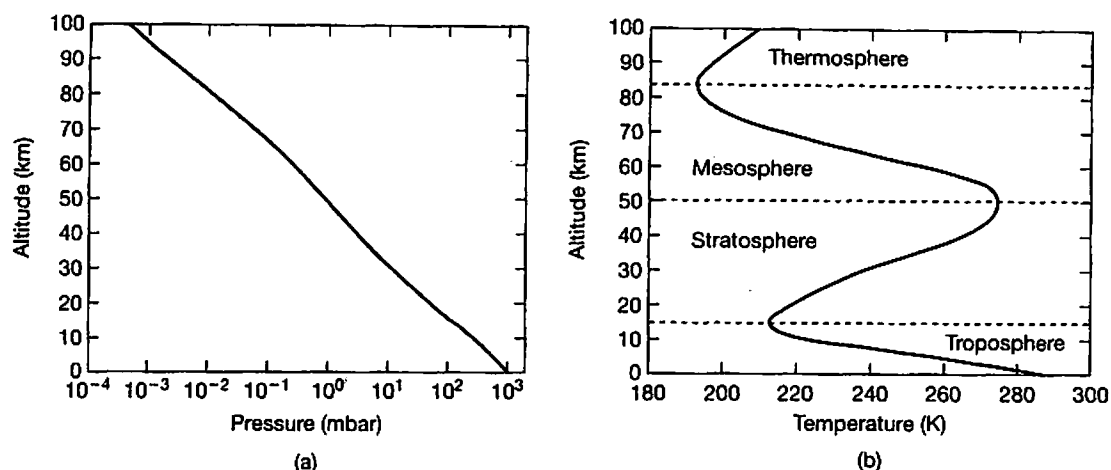


FIGURE 3-9 (a) How pressure varies with altitude in Earth's atmosphere. (b) How temperature varies with altitude in Earth's atmosphere. The different regions of the atmosphere, determined by temperature regimes, are labeled.

atmosphere: the troposphere, the stratosphere, the mesosphere, and the thermosphere. Temperature decreases rapidly with altitude in the lowermost layer of the atmosphere, the **troposphere**, which extends from the surface up to 10–15 km (higher in the tropics, lower near the poles). Immediately above the troposphere is the **stratosphere**, which is located from about 10–15 to 50 km above the surface and in which temperature increases with altitude. Above the stratosphere, temperature decreases with altitude in the **mesosphere** (from about 50–90 km) and then increases once again in the uppermost layer, the **thermosphere** (above about 90 km). These temperature-based “spheres” overlap with atmospheric layers based on other characteristics. For example, the **ionosphere** (a layer that reflects radio waves) includes parts of both the thermosphere and the mesosphere. The very outermost fringe of the atmosphere, where the gas is so tenuous that collisions between molecules become infrequent, is often termed the **exosphere**.

The Troposphere The atmospheric layers that are most important to climate studies are the two lowermost ones: the troposphere and the stratosphere. The troposphere is where most of the phenomena that we call weather—such as clouds, rain, snow, and storm activity—occur. It differs from the other atmospheric layers in that it is well mixed by convection. Convection is a process in which heat energy is transported by the motions of a *fluid* (a liquid or a gas). Such motions are generated when a fluid in a gravitational field, like that of Earth, is heated from below. A familiar example is the convective motion that occurs when a pot of water is heated on the burner of a stove (Figure 3-10). The warm water at the bottom of the pot is less dense than the cooler water at the top. As a result of this imbalance, the fluid overturns (it circulates, or convects) and will continue to do so as long as the pot is being

heated. If the water were heated uniformly or from above, convection would not occur.

We note for completeness that a third mode of heat transfer (in addition to radiation and convection) is **conduction**. Conduction is the transfer of heat energy by direct contact between molecules. The coils of the electric burner shown in Figure 3-10 heat the bottom of the pot by conduction. Conduction plays little role in atmospheric (or oceanic) heat transfer, however, so we will make no further mention of it.

The troposphere is convective because roughly half the incoming sunlight is absorbed by the ground and by the ocean surface. The energy from this light is eventually reradiated to space as IR radiation, but it cannot make its way directly from the surface in this form because IR

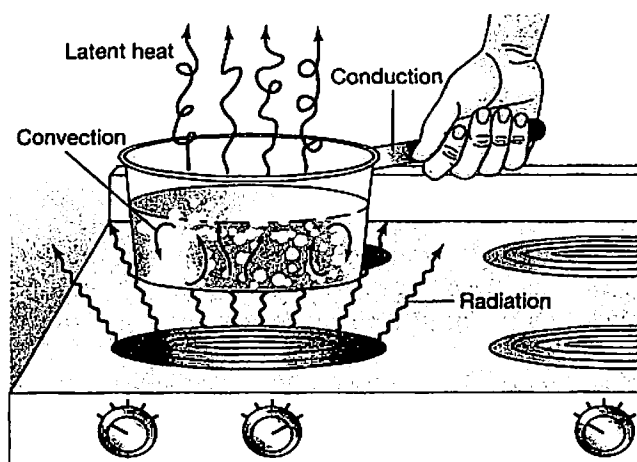


FIGURE 3-10 A pot of water on a stove, illustrating convection. The fluid circulates because it is heated from below. (Source: From R. W. Christopherson, *Geosystems: An Introduction to Physical Geography*, 3/e, 1997. Reprinted by permission of Prentice Hall, Upper Saddle River, N.J.)

radiation is absorbed by atmospheric greenhouse gases and by clouds. So, the energy is instead transported by fluid motions until it reaches an altitude where the atmosphere is more transparent to IR radiation. Only then can the heat energy radiate away from Earth.

As we shall see in Chapter 4, the upward convection of warm, moist air plays a major role in the global energy balance. Convection of heat in a moist atmosphere is more complicated than that in a dry atmosphere, because water can condense or evaporate. When water is evaporated from the ocean surface or from rivers and lakes, energy is taken up by the resulting vapor. This energy is referred to as the latent heat of vaporization. When the water vapor condenses to form clouds, the same amount of latent heat is released to the atmosphere. In more general terms, **latent heat** is the heat energy released or absorbed during the transition from one phase—gaseous, liquid, or solid—to another.

The Stratosphere The stratosphere differs from the troposphere in several respects. The pressure is substantially lower in the stratosphere, in accordance with the barometric law. The two layers differ in composition as well. The stratosphere contains most of Earth's ozone. Stratospheric air is also very dry, containing less than 5 ppm of water vapor on average. Thus, condensation of water vapor does not occur, and so clouds and precipitation are absent. (An exception occurs in the polar regions during winter, where tenuous *polar stratospheric clouds* can form. These clouds play a key role in the development of the Antarctic ozone hole, as we will see in Chapter 17.) Stratospheric air is not convective and is therefore less well mixed than tropospheric air. Indeed, the name "stratosphere" derives from the word "stratified," which means layered.

The Vertical Temperature Profile Why does the vertical temperature profile in Figure 3-9b exhibit all those curves? The reason has to do primarily with where the atmosphere is heated—that is, where solar energy is absorbed. The high temperatures near the ground are caused by the absorption of sunlight at Earth's surface, which then heats the atmosphere above it. The high temperatures near 50 km are caused by the absorption of solar UV radiation by ozone. The ozone concentration actually peaks some 20 km lower, in the middle stratosphere (Figure 3-11), but the heating rate is highest in the upper stratosphere because more UV radiation is available at those altitudes. The vertical heating distribution also explains why the stratosphere is not convective: The maximum heating occurs at the top of the layer, so there is no tendency for the air to rise. Above 50 km, both the ozone concentration and the heating rate decline, so the temperature decreases with altitude in the mesosphere. Finally, the temperature rise above 90 km in the thermosphere is

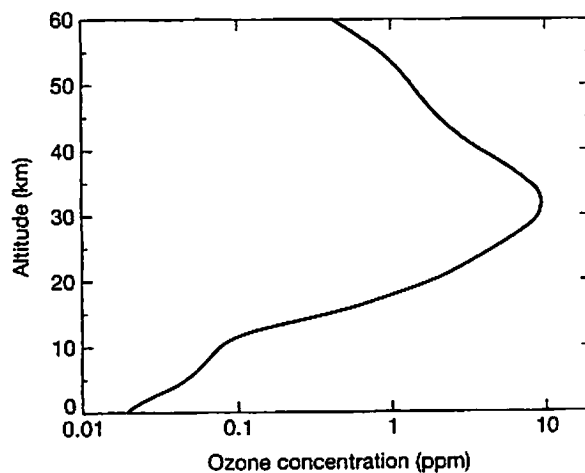


FIGURE 3-11 An approximate profile of the vertical variation of ozone concentration in Earth's atmosphere.

caused by the absorption of short-wavelength UV radiation by molecular oxygen, O_2 .

PHYSICAL CAUSES OF THE GREENHOUSE EFFECT

We determined earlier that Earth's greenhouse effect warms the surface by some 33°C compared with the temperature we would expect if there were no atmosphere. This warming has been attributed to the presence of greenhouse gases, especially H_2O and CO_2 . Why do some gases contribute to the greenhouse effect whereas others, such as O_2 and N_2 , do not?

Molecular Motions and the Greenhouse Gases H_2O and CO_2

The defining property of a greenhouse gas is its ability to absorb or emit infrared radiation. Gas molecules can absorb or emit radiation in the IR range in two different ways. One way is by changing the rate at which the molecules rotate. The theory of *quantum mechanics* describes the behavior of matter on a microscopic scale—that is, the size of molecules and smaller. According to this theory, molecules can rotate only at certain discrete frequencies, just as most house fans can operate only at certain speeds. The rotation frequency is the number of revolutions that a molecule completes per second. Consider one photon of an electromagnetic wave that is incident on an individual molecule (Figure 3-12). If the incident wave has just the right frequency (corresponding to the difference between two allowed rotation frequencies), the molecule can absorb the photon. In the process, the molecule's rotation rate increases. Conversely, the rotation rate slows down when the molecule emits a photon.

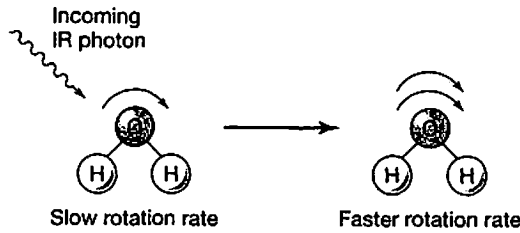


FIGURE 3-12 The rotation rate of an individual H_2O molecule increases when the molecule absorbs a photon of infrared radiation.

The frequency (or wavelength) of the radiation that can be absorbed or emitted depends on the molecule's structure. The H_2O molecule is constructed in such a manner that it absorbs IR radiation of wavelengths of about $12\text{ }\mu\text{m}$ and longer. This interaction gives rise to a very strong absorption feature in Earth's atmosphere called the H_2O rotation band. It can clearly be seen in Figure 3-13, which shows the percentage of radiation at different wavelengths that is absorbed during vertical passage through the atmosphere. Virtually 100% of infrared radiation longer than $12\text{ }\mu\text{m}$ is absorbed, although some of this absorption is caused by CO_2 (see below). The H_2O rotation band extends all the way into the microwave region of the electromagnetic spectrum (above a wavelength of $1000\text{ }\mu\text{m}$), which is why a microwave oven is able to heat up anything that contains water.

A second way in which molecules can absorb or emit IR radiation is by changing the amplitude with which they vibrate. Molecules not only rotate, they also vibrate—their constituent atoms move toward and away from each other. Again consider an electromagnetic wave that is incident on a molecule. If the frequency at which the molecule vibrates matches the frequency of the wave, the molecule can absorb a photon and begin to vibrate more vigorously. (Similarly, a vibrating tuning fork will induce vibrations in a second tuning fork if the pitches of the two instruments

are the same. The pitch is proportional to the frequency of the sound wave.)

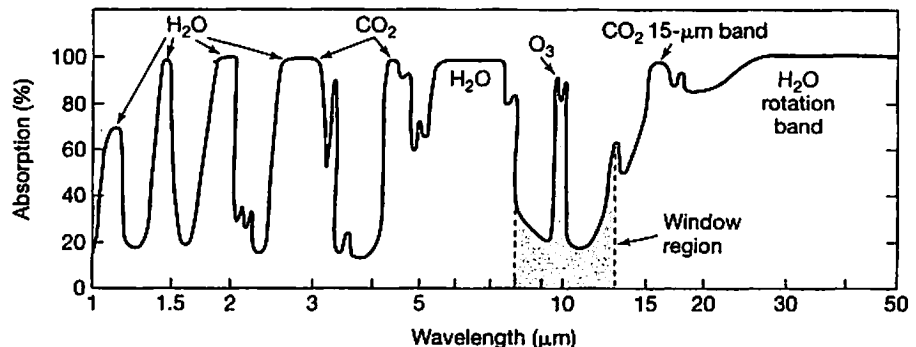
The *triatomic* (three-atom) CO_2 molecule can vibrate in three ways. We need to concern ourselves only with the *bending mode* of vibration (Figure 3-14). This vibration has a frequency that allows the molecule to absorb IR radiation at a wavelength of about $15\text{ }\mu\text{m}$. It gives rise to a strong absorption feature in Earth's atmosphere called the $15\text{-}\mu\text{m}$ CO_2 band. The $15\text{-}\mu\text{m}$ CO_2 band overlaps the H_2O rotation band and, hence, is hard to distinguish in Figure 3-13. It is, however, easily seen by satellites that look down at Earth's atmosphere from above. Because it occurs fairly near the peak of Earth's outgoing radiation, this absorption band is particularly important to climate. Earth's surface emits strongly in this wavelength region, but very little of this radiation is able to escape directly to space because it is absorbed by CO_2 molecules in the atmosphere. This is why CO_2 is such an important contributor to the greenhouse effect.

Other Greenhouse Gases

Water vapor and CO_2 are the most important greenhouse gases in Earth's atmosphere, but several other trace gases—notably CH_4 , N_2O , O_3 , and freons—also contribute to greenhouse warming (Table 3-3). These gases have more of an effect on outgoing radiation than their small concentrations would suggest because they absorb at different wavelengths than do H_2O and CO_2 . Freons, for example, have absorption bands within the 8- to $12\text{-}\mu\text{m}$ *window region*, where both H_2O and CO_2 are poor absorbers (see Figure 3-13). Thus, one molecule of Freon-11 contributes much more to the greenhouse effect than does one CO_2 molecule. Ozone also has an absorption band in this region centered at $9.6\text{ }\mu\text{m}$. Thus, O_3 is a good greenhouse gas as well.

Now, recall that we asked the question, Why are O_2 and N_2 poor absorbers of IR radiation and, thus, do not

FIGURE 3-13 Percentage of radiation absorbed during vertical passage through the atmosphere. Absorption of 100% means that no radiation penetrates the atmosphere. The nearly complete absorption of radiation longer than $13\text{ }\mu\text{m}$ is caused by absorption by CO_2 and H_2O . Both of these gases also absorb solar radiation in the near infrared (wavelengths between about 0.7 and $5\text{ }\mu\text{m}$). The absorption feature at $9.6\text{ }\mu\text{m}$ is caused by ozone. (Source: Data originally from R. M. Goody and Y. L. Yung, *Atmospheric Radiation*, 2nd ed., New York: Oxford University Press, 1989, Figure 1.1.)



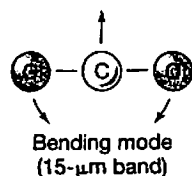


FIGURE 3-14 The bending mode of vibration of the CO_2 molecule.

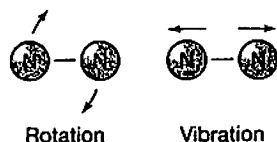


FIGURE 3-15 Rotation and vibration for a diatomic molecule, such as N_2 or O_2 .

contribute significantly to the greenhouse effect? We are now ready to answer that question. *Diatomic* (two-atom) molecules can rotate and vibrate just like the more complicated molecules, H_2O and CO_2 , discussed earlier (Figure 3-15). The O_2 and N_2 molecules, however, are perfectly symmetric: Both of their constituent atoms are identical. Hence, there is no separation of positive and negative electric charges within the molecule. As noted earlier, an electromagnetic wave actually consists of oscillating electric and magnetic fields. To a first approximation, these fields cannot interact with a totally symmetric molecule; the electromagnetic wave passes by such a molecule without being absorbed. (Note that CO_2 is a symmetric molecule because the three atoms are, on average, arranged in a line. However, the symmetry is broken when the molecule bends, allowing $15\text{-}\mu\text{m}$ radiation to be absorbed or emitted.)

EFFECT OF CLOUDS ON THE ATMOSPHERIC RADIATION BUDGET

Gases are not the only constituents of the atmosphere that affect its radiation balance; that balance is also influenced by the presence of clouds and aerosols. We will postpone the discussion of aerosols until Chapter 15, because their climatic effects are usually rather small and short term. (Sulfate aerosols, for example, cooled Earth by about 0.5°C [1°F] for a year or two after the Mt. Pinatubo eruption, as described in Chapter 2.) The effects of clouds, though, are large and cannot be ignored (Figure 3-16). Unfortunately, these effects cannot always be calculated reliably either, and this leads to significant problems in climate prediction.

Types of Clouds

The effect of clouds on Earth's radiation budget is difficult to calculate quantitatively, partly because there are many



FIGURE 3-16 Photo of Earth from space, showing clouds. (Source: NASA Headquarters.)

different types of clouds (Figure 3-17). *Cumulus clouds* are the familiar puffy, white clouds that look like balls of cotton. They are composed of droplets of liquid water and are formed in convective updrafts. *Cumulonimbus clouds* are big, tall cumulus clouds that give rise to thunderstorms. *Stratus clouds* are gray, low-level water clouds that are more or less continuous. They cover much of the eastern United States during winter (especially central Pennsylvania, where the three authors of this book reside!). *Cirrus clouds* are high, wispy clouds composed of ice crystals rather than liquid water, because the temperature of the upper troposphere is well below the freezing point.

Opposing Climatic Effects of Clouds

Have you ever noticed that cloudy days are relatively cool, yet cloudy nights are relatively warm? That is because clouds affect planetary energy balance in two opposing ways. Clouds cool Earth during the daytime by reflecting incident sunlight back to space. We noted earlier that at present the planetary albedo is about 0.3. A large fraction of this is caused by clouds. In fact, without clouds, Earth's albedo would probably be closer to 0.1. According to the planetary energy balance equation, reducing the albedo from 0.3 to 0.1 would raise the effective radiating temperature of Earth by about 17°C (30°F). The increase in surface temperature on a cloud-free Earth would be smaller than this, however, because clouds also absorb and reemit outgoing infrared radiation and, thus, contribute significantly to the greenhouse effect. This effect dominates at night and helps keep cloudy nights warm.

To complicate matters further, the effect of any particular cloud depends on its height and thickness. Low, thick clouds, such as stratus clouds, generally cool the surface because their primary influence is to reflect incoming solar radiation. High, thin clouds, such as cirrus clouds, tend to warm the surface because they contribute



(a)



(b)

FIGURE 3-17 Photos of (a) stratus and (b) cirrus clouds. (Source: (a) Ralph F. Kresge/NOAA and (b) Dean Bergmann/iStockphoto.)

more to the greenhouse effect than to the planetary albedo. The reason for the difference is twofold: First, the elongated ice crystals of which cirrus clouds are composed allow much of the incident solar radiation to pass through but absorb most of the outgoing IR radiation. In contrast, stratus clouds reflect much of the incoming visible radiation, in addition to absorbing radiation at IR wavelengths. Second, cirrus clouds occur higher in the troposphere than do stratus clouds and are therefore colder (Figure 3-18). According to the Stefan-Boltzmann law, cirrus clouds therefore radiate less IR energy to space. Because they absorb the upward-directed IR radiation from the warm surface and reradiate it at a lower temperature, cirrus clouds make a large contribution to the atmospheric greenhouse effect. Lower-lying stratus clouds

do this as well, but their radiating temperature is higher and so their contribution to the greenhouse effect is not as large.

Earth's Global Energy Budget

The various factors that we have just discussed can be combined to calculate a global energy budget for Earth, as in Figure 3-19. The incident solar flux in this diagram is normalized to 100 arbitrary "units" of radiation. These 100 units of incoming energy are balanced by 30 units of reflected solar energy (25 reflected by the atmosphere and 5 reflected by the surface) and 70 units of outgoing infrared radiation. About half the incident solar radiation makes it down to the surface; the other half is either

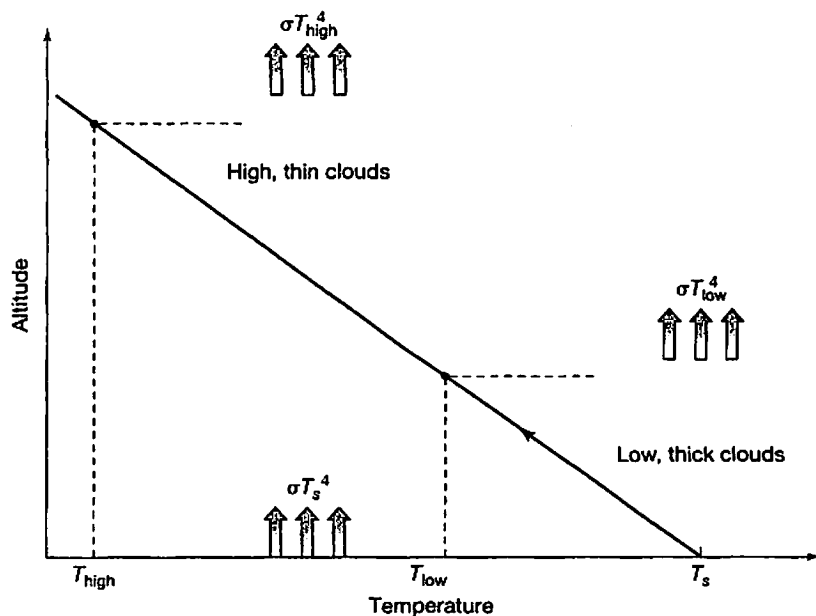


FIGURE 3-18 The different effects of high and low clouds on the atmospheric radiation budget. High, thin clouds are more transparent to incoming sunlight and radiate at a lower temperature than do low, thick clouds. The expressions σT_{high}^4 , σT_{low}^4 , and σT_s^4 represent the radiation flux at the temperature of high, thin clouds, at the temperature of low, thick clouds, and at the surface temperature, respectively.

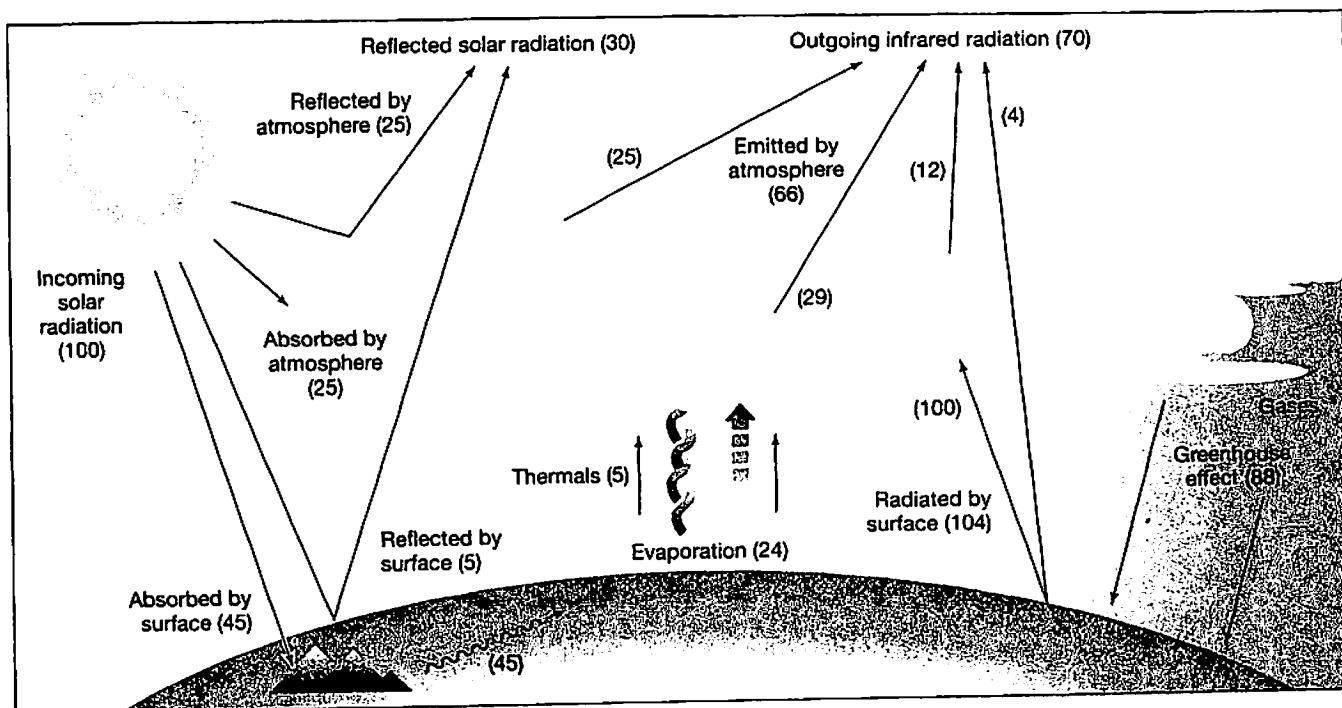


FIGURE 3-19 Earth's globally averaged atmospheric energy budget. All fluxes are normalized relative to 100 arbitrary units of incident radiation. (Source: S. Schneider, "Climate Modeling," *Scientific American* 256, no. 5, 1987, pp. 72–80.)

reflected or absorbed by the atmosphere. Within the atmosphere, energy is transported by a combination of radiation, convection, and the latent heat associated with the evaporation and condensation of water vapor. This latter process is a very important one: Roughly half of the solar energy absorbed by the surface (24 out of 45 units) goes directly into evaporating water.

The greenhouse effect is shown here as an additional 88 units of downward-directed infrared radiation. Thus, the total energy flux absorbed by the surface is 133 units (= 45 units of solar radiation + 88 units of IR radiation). This value is almost twice the net amount of energy absorbed by Earth (70 units). The reason is that infrared radiation is absorbed and reemitted multiple times within the atmosphere, so the internal fluxes can actually be higher than the net input of energy. *At the top of the atmosphere, however, the net downward solar radiation flux (incoming minus reflected) must equal the outgoing infrared flux.* This statement is the principle of *planetary energy balance*.

INTRODUCTION TO CLIMATE MODELING

How can we utilize our knowledge of Earth's current energy budget to predict what Earth's surface temperature might have been in the past or how it might vary in the future? Because the climate system is complex, we need some sort of computer model to keep track of all the intricacies.

Instead of estimating the magnitude of the greenhouse effect by subtraction, as we did earlier in this chapter, we need to be able to calculate it directly from the measured or predicted concentrations of greenhouse gases.

Such a calculation must take into account the rotational and vibrational absorption bands of all the different greenhouse gases. Doing so quantitatively requires that we combine the predictions of quantum mechanics with laboratory measurements of the strengths of different absorption bands. Fortunately for climate modelers, a great deal of effort has gone into obtaining the required parameters. As a result, a voluminous database of information on the absorption characteristics of various molecules of atmospheric interest is now available.

Armed with these data, the climate modeler must next decide how to incorporate them into a computer model of the atmosphere. The most complete type of model is called an **atmospheric general circulation model (GCM)**, also sometimes referred to as a *global climate model*. These elaborate computer models include a three-dimensional representation of the atmosphere (or oceans) that simulates winds (or currents), moisture transport, and energy balance. Atmospheric GCMs are replete with clouds, winds, snow, rain, and most of the other phenomena that we call weather. Thus, they are capable of predicting how climate varies on a regional basis. But GCMs have a number of drawbacks, the most serious of which is that they require large amounts of runtime on the world's fastest computers to simulate

even a few years of global climate. We describe GCMs in some detail in Chapter 15 because they play a central role in climate policymaking today. We begin here, however, with models that are somewhat less complicated and, hence, easier to understand.

One-Dimensional Climate Models—RCMs

For many purposes it is sufficient to construct simpler climate models that require less effort to program and less computer time to run. We have already seen one such model—the one-layer atmosphere model described in the box earlier in this chapter. But that model did not produce a good, quantitative estimate of the greenhouse effect, nor did it account for the contributions of different greenhouse gases. The simplest model that is capable of doing both of these things reliably is called a **radiative-convective model (RCM)**. In an RCM, the climate system is approximated by averaging the incoming solar and outgoing IR radiation over Earth's entire surface. The vertical structure of the atmosphere (Figure 3-9) is taken into account (unlike in the one-layer model), but horizontal variations are ignored. Thus, such models are sometimes called one-dimensional climate models, in contrast with the three-dimensional GCMs. The vertical dimension (altitude) is then divided into a number of layers. The RCM calculates the temperature of each layer by taking into account the amount of energy received or emitted in the form of radiation, along with the effects of convection and latent heat release in the lowermost layers.

RADIATIVE EFFECT OF DOUBLING ATMOSPHERIC CO₂

Although RCMs are quite simple compared with the real climate system, they allow us to estimate the magnitude of the greenhouse effect as a function of the concentrations of various greenhouse gases in Earth's atmosphere. These models correctly predict that the greenhouse-induced temperature difference ΔT_g for the present atmosphere is 33°C, in agreement with the estimate derived earlier by the subtraction $\Delta T_g = T_s + T_e$. (This is *not* a trivial result. We would have to work through a lot of the relevant physics to come up with this answer.) More importantly, RCMs allow us to predict the average surface temperature increase that should result from an increase in the concentration of greenhouse gases. A commonly cited benchmark is the temperature change that would result from a doubling of the atmospheric CO₂ concentration from 300 ppm (its value near the turn of the 20th century) to 600 ppm. RCM calculations show that, all other factors being equal, such a change in CO₂ would produce an increase of about 1.2°C (2.2°F) in the global average surface temperature. In the terminology developed in Chapter 2, this value is the temperature change ΔT_0 that would result in the absence of any feedbacks in the climate system.

In reality, we would expect other factors in the climate system to change as atmospheric CO₂ increases, and so a temperature change of +1.2°C is not the best estimate we could make of the effect of CO₂ doubling. To obtain a better estimate, we must consider what those climate feedbacks might be and how strongly they affect our answer.

CLIMATE FEEDBACKS

Climate feedbacks are extremely important because they can either amplify or moderate the radiative effect of changes in greenhouse gas concentrations. That is why we devoted most of Chapter 2 to explaining how they work. There, we dealt with an imaginary feedback system involving the percentage of daisy cover on the hypothetical planet Daisyworld. Here, we discuss several feedback processes that affect climate on Earth.

The Water Vapor Feedback

One of the most important feedbacks in the climate system involves the concentration of atmospheric water vapor. As noted earlier, water vapor is an excellent absorber of IR radiation and, hence, a good greenhouse gas. Unlike CO₂, however, water vapor is typically close to its *condensation point*—the temperature at which a vapor condenses to form a liquid. If Earth's surface temperature were to decrease for some reason, water vapor would condense out in the form of rain or snow, leaving less water vapor behind in the atmosphere. This reduction in atmospheric water vapor would cause a corresponding decrease in the greenhouse effect, which, in turn, would lower the surface temperature still further. Conversely, an increase in surface temperature would cause an increase in the rate at which water vapor evaporates from the oceans. This would increase the concentration of water vapor in the atmosphere, thereby increasing the greenhouse effect and further warming Earth's surface.

The net result of this interaction between water vapor abundance and Earth's surface temperature is a positive feedback loop that tends to amplify small temperature perturbations (Figure 3-20). This feedback loop can be incorporated in RCMs by assuming a fixed *relative humidity*

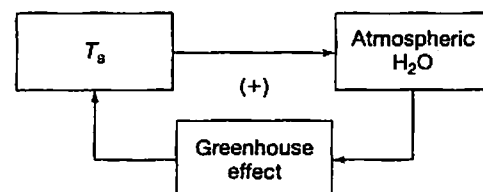


FIGURE 3-20 Systems diagram showing the positive feedback loop that includes atmospheric water vapor.

profile in the troposphere. **Relative humidity** is the concentration of water vapor in an air parcel divided by the concentration that would be present if the air parcel were *saturated* with water vapor (i.e., on the verge of condensation). When such a calculation is performed, the RCM predicts that the equilibrium change in surface temperature for CO₂ doubling, ΔT_{eq} , is about twice that which would have occurred otherwise. Recall from Chapter 2 that we can write

$$\Delta T_{eq} = \Delta T_0 + \Delta T_f,$$

where ΔT_0 is the temperature change with no feedbacks and ΔT_f is the change caused by the feedback. For the problem of CO₂ doubling, $\Delta T_0 = 1.2^\circ\text{C}$ (2.2°F), $\Delta T_{eq} \approx 2.4^\circ\text{C}$ (4.4°F), so the temperature change caused by the water vapor feedback is approximately 1.2°C . Furthermore, the feedback factor f is given by

$$f = \frac{\Delta T_{eq}}{\Delta T_0} = \frac{2.4^\circ\text{C}}{1.2^\circ\text{C}} = 2.$$

A feedback factor of 2 indicates that this is a strong, positive feedback on the climate system.

Snow and Ice Albedo Feedback

A second feedback loop that is expected to have some impact on modern global warming, but is especially important for glacial–interglacial variations, involves albedo changes caused by snow and ice. As Earth's climate cools, the extent of wintertime snow and ice cover increases in temperate regions. On longer time scales, the permanent ice cap in the northern polar regions expands toward the equator, resulting in the periods of glaciation known as the Ice Ages. Snow and ice have a much higher albedo than does land or water (refer to Table 2-1). Therefore, increases in snow and ice cover should cause further decreases in surface temperature. The result is a positive feedback loop that tends to amplify induced changes in Earth's surface temperature (Figure 3-21). As snow and ice cover are restricted to middle and high latitudes, modeling this feedback loop quantitatively requires the use of two-dimensional or three-dimensional computer models.

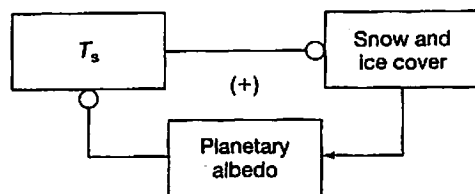


FIGURE 3-21 Systems diagram showing the positive feedback loop that includes snow and ice cover.

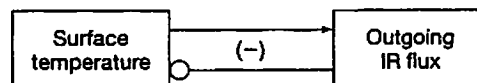


FIGURE 3-22 Systems diagram illustrating the negative feedback loop between surface temperature and the outgoing flux of infrared radiation. This feedback is the fundamental reason that Earth's climate is stable.

The IR Flux/Temperature Feedback

Both of the feedbacks discussed so far are positive. But systems that contain only positive feedback loops are unstable. Does this mean that Earth's climate is unstable? No! You would not be sitting there reading this book if that were true. Earth's climate system contains a very strong negative feedback that is so basic that it is often overlooked. The feedback loop that stabilizes Earth's climate on short time scales is the relationship between surface temperature and the flux of outgoing IR radiation (Figure 3-22). (We have already hinted in Chapter 1 that there is another feedback loop that stabilizes Earth's climate on long time scales, but that is not what we are talking about here.) If Earth's surface temperature were to increase for some reason, the outgoing IR flux from the top of the atmosphere would also increase. But if the outgoing IR flux were to increase, the surface temperature would tend to decrease, because more energy would be lost from the Earth system. This feedback loop might appear to be trivial, but it is not; there are situations in which it can fail. In particular, the positive correlation between surface temperature and the outgoing IR flux can break down if the atmosphere contains a very large amount of water vapor. This, we think, is what happened to our sister planet, Venus, and it led to what is sometimes called a *runaway greenhouse*. But we will save that story for Chapter 19.

The Uncertain Feedback Caused by Clouds

Another important feedback process in the climate system is that provided by changes in clouds. Unfortunately, this feedback process is not as easy to quantify as the ones just discussed. You already know that clouds can either warm the surface, or cool it, depending on their height. This alone should provide a hint that estimating their feedback effect might be difficult. In addition to this problem, clouds are inherently three-dimensional: they form at some locations and not at others because of the way the winds blow. As we will see in Chapter 15, most current 3-D climate models suggest that the net feedback effect of clouds is positive for the doubled CO₂ problem because the increase in high-altitude cirrus clouds (which cause warming) outweighs any increase in low-altitude stratus clouds (which cause cooling). But climate modelers agree that this ranks as one of their least secure predictions, and hence adds significant uncertainty to projections of future global warming.

In summary, we have now examined Earth's climate system in enough detail to understand how the atmospheric greenhouse effect warms the planet and how the planet's average surface temperature may respond to a human-induced increase in greenhouse gases. But Earth's climate cannot be described by just its average surface temperature. The term "climate" includes many other related factors, such

as latitudinal and seasonal temperature gradients, winds, and precipitation. To study these phenomena, we need to broaden our spatial perspective and consider the Earth system from a three-dimensional perspective. The next two chapters describe how the transport of heat from one location to another by the atmosphere and oceans determines these other important features of Earth's global climate.

Chapter Summary

1. Earth is warmed by the absorption of visible radiation from the Sun and is cooled by the emission of infrared radiation to space.
 - a. Much of the infrared radiation emitted by Earth's surface is absorbed and reemitted by atmospheric gases.
 - b. The result is a greenhouse effect that warms the surface by about 33°C. Without this natural greenhouse effect, Earth would be too cold to support life.
2. Only certain atmospheric gases, most importantly H₂O and CO₂, contribute to the greenhouse effect. These gases absorb infrared radiation by changing the rate at which individual molecules rotate or vibrate. Other trace gases, such as freons, can contribute substantially to the greenhouse effect by absorbing radiation at different wavelengths than do H₂O and CO₂.
3. Clouds affect the atmospheric radiation budget both by reflecting incident sunlight and by contributing to the greenhouse effect. Low, thick clouds tend to cool the surface; high, thin clouds tend to warm it.
4. Earth's climate system contains several well-understood feedbacks that play important roles in regulating climate change.
 - a. The climate system is stabilized by a strong negative feedback loop between surface temperature and the outgoing infrared flux.
 - b. The system is destabilized by a positive feedback loop involving atmospheric water vapor. Because it acts on short time scales, this feedback is likely to play an important role in contemporary global warming. Climate models predict a surface temperature response to CO₂ doubling that is twice that of models in which this feedback is neglected.
 - c. The system is also destabilized by a positive feedback loop involving the extent of snow and ice cover due to the effect of albedo.
 - d. Clouds may also contribute to climate feedback, but their effect is not well understood.

Key Terms

barometric law	general circulation model (GCM)	relative humidity
blackbody	H ₂ O rotation band	Stefan-Boltzmann law
blackbody radiation	infrared (IR) radiation	stratosphere
conduction	inverse-square law	thermosphere
convection	Kelvin (absolute) temperature scale	troposphere
effective radiating temperature	latent heat	ultraviolet (UV) radiation
electromagnetic radiation	mesosphere	visible radiation
electromagnetic spectrum	photon	visible spectrum
15- μ m CO ₂ band	photosphere	wavelength
flux	radiative-convective model (RCM)	Wien's law
frequency		

Review Questions

1. How are the wavelength and frequency of an electromagnetic wave related?
2. What is a photon?
3. What physical law describes the manner in which the intensity of sunlight changes as the observer moves away from the Sun?
4. Name two physical laws that apply to blackbody radiation. What do these laws tell us about the nature of the emitted radiation?
5. What is the major contributor to Earth's albedo?
6. What are the three most abundant gases in Earth's atmosphere?

7. List the four layers of Earth's atmosphere. How are they defined?
8. Name three mechanisms by which heat energy can be transferred. Which two are important in Earth's global energy budget?
9. Identify two physical processes by which gases can absorb infrared radiation. Give examples of each process.
10. Why are O_2 and N_2 not greenhouse gases?
11. Describe the different ways in which climate is affected by high and low clouds.
12. Identify two positive feedback loops in Earth's climate system. Why is Earth's climate stable despite these destabilizing, positive feedbacks?

Critical-Thinking Problems

1. a. Given that a 300-K blackbody radiates its peak energy at a wavelength of about $10\ \mu\text{m}$, at what wavelength would a 600-K blackbody radiate its peak energy?
b. If the two bodies in part (a) were the same size, what would be the ratio of the heat emitted by the hotter object to the heat emitted by the colder one?
2. a. Venus and Mars orbit the Sun at average distances of 0.72 and 1.52 AU, respectively. What is the solar flux at each planet?
b. Venus has a planetary albedo of 0.8, and Mars has an albedo of 0.22. Using the answer to part (a), determine the effective radiating temperatures of these planets.
c. How do the effective radiating temperatures determined in part (b) compare with the value for Earth, and why is this result surprising?
d. The mean surface temperatures of Venus and Mars are 730 and 218 K, respectively. Using the answer to part (b), determine the magnitude of the greenhouse effect on each planet.
e. How do the results of (d) compare with the magnitude of the greenhouse effect on Earth?
3. a. The Sun radiates at an effective temperature of 5780 K and has a radius of about 696,000 km. Remembering that 1 AU = 149,600,000 km, derive the approximate value of the solar flux at Earth's orbit.
b. Compare your answer with the value given in the text.
4. The tropospheric lapse rate (the rate at which temperature decreases with altitude) is approximately 6°C (11°F) per kilometer. Given that the mean surface temperature of Earth is 288 K and the effective radiating temperature is 255 K, from what altitude does most of the emitted radiation derive?
5. Solar luminosity is estimated to have been 30% lower than today at the time when the solar system formed, 4.6 billion years ago.
- a. If Earth's albedo was the same as it is now ($A = 0.3$), what would have been its effective radiating temperature at that time?
b. If the magnitude of the greenhouse effect had also remained unchanged ($\Delta T_g = 33\ \text{K}$), what would Earth's average surface temperature have been? How does this compare with today's value?
6. For atmospheric CO_2 concentrations not too different from the present value, the radiative forcing of CO_2 can be expressed by the formula

$$\Delta F = -6.3 \ln\left(\frac{C}{C_0}\right),$$
 where $C_0 = 300\ \text{ppm}$ is the CO_2 concentration near the turn of the 20th century, C is the CO_2 concentration at some other time, and ΔF is the change (in watts per square meter) in the outgoing infrared flux caused by the change in CO_2 concentration. The function $\ln(x)$ denotes the natural logarithm of a given number x . Any scientific calculator has this function key.
 - a. By how much would the outgoing infrared flux decrease if the atmospheric CO_2 concentration were increased from 300 to 600 ppm (i.e., if $C = 600\ \text{ppm}$)?
 - b. By how much would surface temperature have to increase in order to bring the radiation budget back into balance in part (a), assuming that the planetary albedo and the amount of water vapor in the atmosphere do not change? (*Hint:* Use the planetary energy balance equation to calculate how much T_e would have to change to balance the radiation budget. Remember that the left-hand side of this equation represents the outgoing infrared flux. The quantity T_s will change by the same amount as T_e if the amount of water vapor is held constant.)