CHAPTER 15

Global Warming, Part 1

Recent and Future Climate

Key Questions

- What causes climate change on short time scales?
- How has climate varied over the past 10,000 years?
- Are the observed changes in climate over the past century natural or human-induced?
- How do anthropogenic carbon fluxes compare with natural fluxes?
- How much are atmospheric CO_2 and other greenhouse gases expected to rise over the next few decades to centuries, and how will this rise affect Earth's climate?

Chapter Overview

Earth's climate has remained remarkably stable over the past 10,000 years, although it was slightly warmer 5,000 to 6,000 years ago (the Holocene Climatic Optimum) and slightly cooler from about A.D. 1600 until 1850 (the Little Ice Age). Since 1850, the global average surface temperature has increased by about 0.8°C. The largest part of this change has occurred over the past 40 years and has likely been caused by human activities: release of CO_2 by fossil-fuel burning and deforestation and release of CH_4 and N_2O from agriculture. If no actions are taken to reduce emissions, CO_2 levels are expected to more than double over the next century and could increase by a factor of 6 to 8 over the next few centuries. Temperature increases within the next century are predicted to be only a few degrees Celsius, but long-term climate changes could be substantially larger than this. The oceanic thermohaline circulation might also slow down as a consequence of increased rainfall in the North Atlantic and resulting decreases in salinity of surface water in that region.

INTRODUCTION

We have seen in earlier chapters that Earth's climate has varied on a number of different time scales. Over the past few billion years, the Sun has brightened considerably and Earth's climate has gone through a series of warm and cold periods. Earth has even experienced brief periods where the surface seems to have been completely frozen. But the system has recovered from these catastrophes and has remained within a range conducive to the continued presence of life. Over the past few million years, the climate has oscillated between glacial and interglacial intervals triggered by variations in Earth's orbit and amplified by internal feedbacks within the climate system.

Humans are now altering Earth's climate by adding greenhouse gases to its atmosphere. We saw evidence for this in the plots of atmospheric CO_2 concentration and global mean surface temperature shown in Chapter 1 (Figures 1-3 and 1-4). But how can we know if the observed temperature changes have been caused by humans, or if they simply represent natural fluctuations in the climate system? And how can we tell if the recent increase in atmospheric CO_2, which itself is not disputed, will continue into the future? To answer these questions, we need to do several things: First, we must examine climate change over the past 10,000 years, termed the **Holocene epoch**, so that we can place modern climate change in perspective. Then, we must look more carefully at the modern CO_2 cycle to see how humans are perturbing it. And, finally, we need to look at climate modeling results to understand how high-powered computer models are being used to make projections of future climate change. It is a daunting task, but one for which the reader should be well prepared after making it through the earlier chapters of this book.

HOLOCENE CLIMATE CHANGE

Proxy Climate Data

We have a vast quantity of data with which to describe and analyze the present-day climate system. In particular, an array of measuring stations with accurate thermometers covers much of Earth's land surface, and ships and floating buoys provide measurements for much of the world's oceans. As seen already back in Chapter 1 (Figure 1-4), these data extend back to about A.D. 1850. Where there are gaps in the coverage, we can fill them in today with satellite data. In fact, in terms of quantity, satellites now provide the bulk of the observational climate data set. Consistent and reliable satellite data, however, have been available only since the early 1970s. The data prior to the satellite era have huge gaps over the oceans, and over the land the climate record is highly variable—some regions have several hundred years of observations, others less than 30. With such a short record of observational data, how do we determine climate variability and climate change of the past?

The data problem gets much worse prior to 1850. Accurate, mercury-based thermometers had been invented over 100 years before this time by the German physicist Daniel Gabriel Fahrenheit, but their use was restricted primarily to Europe and North America. To obtain estimates of global temperatures prior to this time, we have to resort to the use of **proxy data**. We have already seen some examples of proxy climate data in earlier chapters. In Chapter 14, for instance, we discussed the geological evidence of the advance and retreat of continental glaciers, such as glacial till deposits and moraines, as well as atmospheric and climate reconstructions derived from ice cores. As we move to time scales covering the past 10,000 years, we make use of other types of evidence in addition to ice cores. Two of the most useful techniques for reconstructing past climates are based on *palynology* and *dendrochronology*.

Palynology is the study of pollen and organic microfossils. Pollen grains are preserved in many different environments (for example, lake sediments and peat bogs). If we drill a core into sediments from a lake or from a peat bog, divide the core into segments going back through time, and then extract the pollen from each layer, we can reconstruct the plant assemblages that lived in the area of that core at each time interval in the past. We then use the present-day distribution of those assemblages to place constraints on what the environment was like in the past. Pollen data from peat bogs have been used to reconstruct climate over the past 30,000 to 35,000 years in the British Isles. Radiocarbon dating (Chapter 5) can be used to date different levels of the core to associate environmental changes with particular time periods.

Dendrochronology is a method of dating trees by counting their annual growth rings. This method uses the growth characteristics of certain tree species. The cross section of a tree trunk consists of a series of rings, each ring representing growth over one year. By counting the annual rings, we know the age of the tree. The width of each ring indicates the amount of growth that occurred during the growing season. In certain circumstances, that amount can be related to the temperature during the growing season or to water availability, both of which can tell us something about the climate while the tree was alive. For example, a climate record that extends back almost 5,500 years has been reconstructed from tree rings of bristlecone pines in the White Mountains of California.

Holocene Warm and Cold Periods

By combining ice-core data with data from these other techniques, we can make estimates of how global surface temperatures have varied over the past few tens of thousands of years. Figure 15-1b, for example, shows how climate has varied over the past 20,000 years. If we zoom in a little closer (Figure 15-1c), we can look at the variations in surface temperature over the past 1,000 years. These changes are compared to the Pleistocene glacial–interglacial changes in Figure 15-1a. As mentioned already, the really big temperature changes occurred prior to about 10,000 years ago. But significant fluctuations, of the order of 1–2°C, have occurred all the way up to the present time. If one looks at the data in more detail, one finds that the pattern of change is not uniform across the globe. It is difficult to generalize, but there appears to be a dominance of temperature changes in the midlatitudes and high latitudes, whereas the tropics and subtropics appear to have experienced greater changes in moisture availability. These changes result partly from orbital effects that enhance seasonality and continentality (which directly affect the temperature regime) and partly from the resulting circulation changes (such as changes in the monsoon circulation) that affect precipitation patterns. Other factors may also have played a role in regional climate changes, as we will see later in this chapter.

Another point to note is that relatively small changes in the mean global temperature are associated with relatively large changes in the physical environment. Mean global temperatures at the height of the last glaciation were probably only 5 to 7°C lower than the 20th-century mean. Eight

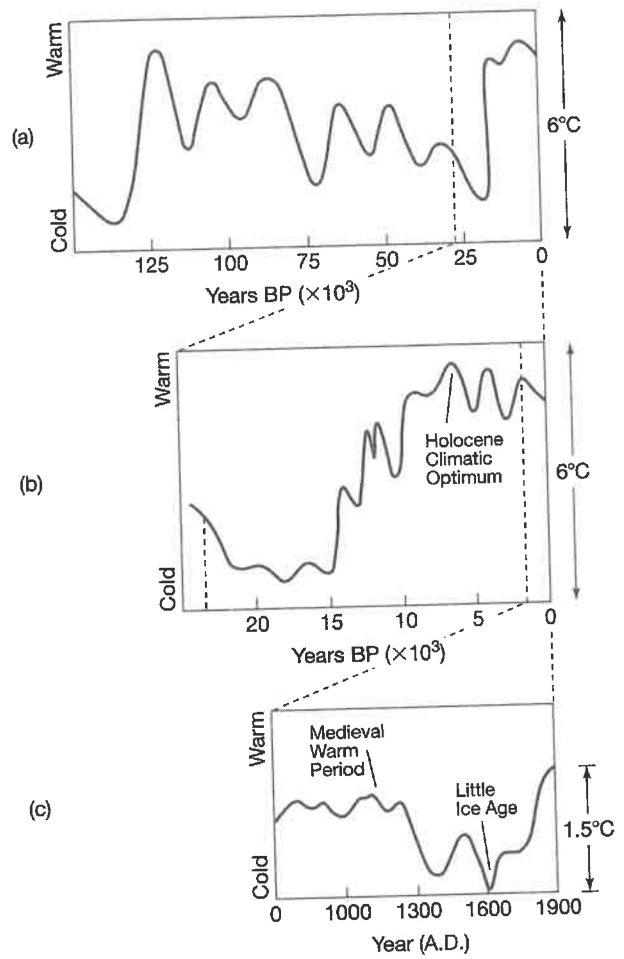


FIGURE 15-1 Mean global temperature change since the last glacial maximum. (a) Generalized oxygen isotope curve from deep-sea sediments. (b) General estimates from pollen data and alpine glaciers (emphasis on midlatitudes from eastern North America and Europe). (c) General estimates from historical documents (emphasis on the North Atlantic region). (*Source:* Adapted from U.S. Committee for GARP, *Understanding Climate Change: A Program for Action*, National Academy of Sciences, Washington, DC, 1975.)

hundred years ago, when the Vikings were able to colonize parts of Greenland and grow sufficient crops to maintain a continuous settlement, mean global temperatures were only 0.5°C or so warmer than they are today. The significance of these numbers is highlighted by the fact that global temperature changes predicted for a doubling of atmospheric CO_2 are on the order of 1.5 to 4°C. In other words, the changes we predict for potential global warming in the 21st century are much greater than anything that has occurred in the past 10,000 years, and those changes are of comparable magnitude to the warming that took place between the last glacial maximum and the present day.

THE HOLOCENE CLIMATIC OPTIMUM The short cold interval of the Younger Dryas event ended with a rapid shift to warmer conditions, followed by a constant climate or relatively slow warming over the next several thousand years—the **Holocene Climatic Optimum**. Evidence suggests that summer temperatures were slightly higher in the Mid-Holocene (5,000 to 6,000 years ago) than are recorded in the recent (20th-century) record. This interval was long thought of as a time of persistent mild conditions with very little climate change. This view originates from early studies in Europe, where the pollen record shows little evidence of major climatic shifts. Elsewhere in the world the picture is less straightforward.

Ancient lake levels in East Africa and the Sahara Desert show evidence of much wetter conditions than exist today. These high lake levels occurred when temperatures were higher—presumably, evaporation would have been greater then. This means that the amount of rainfall in the region must also have been greater than it is today. The evidence suggests a northward shift of the intertropical convergence zone in the Northern Hemisphere summer, bringing higher rainfall to the Sahara and East Africa and an enhanced monsoonal circulation (increased summer rain) in Arabia and northwest India. There is also evidence that the Mediterranean Sea experienced increased summer rainfall during this same interval. Turkey and western Iran, however, appear to have been more arid than at present.

That the Mid-Holocene climate was very different from today's climate is also indicated by archeological evidence. The Tarim Basin, the site of the ancient silk route from China to Europe, is currently a desert, but between 5,000 and 6,000 years ago it was forested and populated with numerous settlements. Nomads grazed cattle in the central Sahara, and the Harappan culture flourished in the Indus River valley. This agricultural society, located in what is now the Rajasthan Desert, may have been the first to cultivate cotton. At first glance, changes in climate would appear to be a likely explanation for the decline of these earlier civilizations. Undoubtedly, the Sahara, for example, is now much drier than it was then. In several cases, however, it is possible that human land-use practices, rather than a change in climate, led to land degradation. In fact, it is likely that the collapse of some of these cultures resulted from the interaction of both factors—where land degradation due to human land use was amplified by changing climate conditions. It would seem reasonable to look back at this, and other warm periods, and suggest that they might be an indicator of what our future climate may be like as global warming progresses. Unfortunately, analogs from the past are of limited use for predicting the future as the forcing factors are different. The climate response, therefore, is also likely to be different.

THE MEDIEVAL WARM PERIOD Temperatures fell after the Holocene Climatic Optimum, reaching a minimum about 3000 years ago, but rose to a new maximum during the European **Medieval Warm Period**. Temperatures in Greenland appear to have increased after A.D. 600 to 650, with the temperatures in the North Atlantic (Greenland and

Iceland) reaching a maximum around A.D. 1100. At this time the Vikings established a self-supporting colony on the southwest coast of Greenland. The colony lasted more than 400 years and, at its largest extent, had 280 farms and a population of 3,000. By the end of the 12th century, however, because of the decreasing temperatures, the sea ice east of Greenland grew more extensive. By the middle of the 14th century, ships had to take a more southerly route to avoid the ice. By A.D. 1410, communication with the Greenland colonies was lost completely.

Farther south, in northern, western, and central Europe, the Medieval Warm Period reached a maximum between A.D. 1150 and A.D. 1300. During this period, wheat was grown in Norway at about 64° N, oats and barley were grown in Iceland, vineyards were cultivated in England, and farm settlements spread to higher elevations in Norway, northern England, and Scotland—all evidence of milder climates than those areas have today. The average temperature of central England during this interval is estimated to have been 0.5 to 0.8°C above the mean for the first half of the 20th century.

From the beginning of the 14th century, the climate became more variable. Wetter (and probably colder) summers in Europe from A.D. 1313 to 1317 led to a succession of failed harvests and widespread famine, and the expansion of farms into the upland regions of northern Europe and Scandinavia came to an abrupt end. The interval from A.D. 1250 to A.D. 1350 was one of numerous large storms and floods. It has been suggested that the storminess was caused by a cooling at high latitudes that caused the sea ice to expand southward, resulting in an increased temperature gradient in the North Atlantic. Flooding along the North Sea coasts of Denmark and Germany was extensive, and 100,000 to 400,000 people were reported to have drowned in the floods that occurred at that time. The impact of these large environmental changes on human societies was then compounded by the arrival in Europe of the bubonic plague in 1346. The plague lasted until 1361, killing an estimated 25 million people (one-quarter of Europe's population).

THE LITTLE ICE AGE The extreme climatic fluctuations that marked the end of the Medieval Warm Period led to an interval of cooling until the early 1500s, after which the climate stayed relatively stable or even began to recover in some areas. The North Atlantic climate, however, entered a renewed episode of rapid cooling in the late 1500s, a period now known as the **Little Ice Age**. Although the cooling was originally thought of as a regional climate fluctuation centered on western Europe and the North Atlantic, a growing body of evidence from the European Alps, Asia and the Himalayas, South America, New Zealand, and Antarctica suggests that changes might have occurred over much of the globe. However, not all parts of the world show evidence of a climate change at this time. Where they do, it is not clear that the changes were synchronous or that they lasted the same length of time everywhere. The Little Ice Age continued through the middle of the 19th century, but the reduced temperatures were not continuous. The Little Ice Age is characterized by considerable variability, with episodic cold spells that varied in timing and duration from place to place.

The evidence of the Little Ice Age takes numerous forms, among them the readvance of mountain glaciers, the lowering of tree lines, increased erosion and flooding, sea-ice expansion, and the freezing of canals and rivers. The canals of Holland have long been used for transportation, and there are reliable records of freeze-up since 1633. The canals seldom freeze over in today's climate, but in the 15th and 17th centuries it was common for them to be frozen for 3 months at a time. There is also documented evidence of glaciers in the Swiss Alps advancing and covering houses on the outskirts of several villages (Figure 15-2). A climate change is also suggested by indicators of population

(a) (b)

FIGURE 15-2 The Argentiere glacier in the French Alps. (a) An etching made about 1850, showing the extent of the glacier during the waning phase of the Little Ice Age. (b) The same view, photographed in 1966. (*Source:* (a) National Academy Press and (b) National Academy Press.)

decline: the abandonment of settlements and decreasing agricultural productivity. However, the societal evidence is not straightforward; population changes and declining productivity were almost certainly influenced by societal and political factors and by disease.

As with the Younger Dryas event described earlier, the Little Ice Age appears to have had a strong regional focus, yet evidence is mounting that many other parts of the world also experienced a similar cooling. In this case, however, there was no retreating continental ice sheet to help force large changes in the oceanic circulation. Are there any other likely explanations for this particular shift in the climate system?

Volcanoes and Climate

One possibility is that the Little Ice Age was caused by increases in volcanic activity. We saw in earlier chapters that volcanism releases CO_2, which helps warm Earth by contributing to the greenhouse effect. But volcanoes can also cause short-term cooling by injecting sulfur dioxide (SO_2) into the atmosphere. This SO_2 is eventually oxidized to form sulfate aerosols, in much the same way as SO_2 released from coal-burning forms aerosols today (Chapter 1). Some volcanoes, however, are highly explosive and can inject SO_2 high into the stratosphere. Because clouds and rain, which normally remove sulfate aerosols, are not present in the stratosphere, these aerosols can remain aloft for many months and can have a significant cooling effect on Earth's climate. As early as 1784, Benjamin Franklin hypothesized that the abnormally cold winter of 1783–1784 might have been due to the 1783 eruption of Hekla, a volcano on Iceland. Actually, any climate anomaly in that year was more likely due to the 1783 eruption of Laki, another Icelandic volcano. Laki is, in fact, the largest effusive eruption in the historic record, producing about 12 km^3 of basalt lava flows.

The degree to which a volcanic eruption affects climate depends in part on the location of the eruption and on the way the atmospheric circulation distributes the stratospheric aerosols globally. Mt. Pinatubo in the Philippines experienced a major eruption in June 1991, when the equivalent of 3 to 5 km^3 of dense rock was ejected into the atmosphere (Figure 15-3). The ejecta also included approximately 20 Mton of SO_2 gas, which was subsequently converted to sulfuric acid aerosol particles, with the largest concentration occurring in the lower stratosphere at an altitude between 15 and 20 km. The aerosol cloud was distributed very rapidly around the globe by the stratospheric circulation, circling the entire globe in only 22 days. The cooling from this event lasted approximately 1–2 years and can be seen as a brief drop in surface temperature in Figure 1-4 (Chapter 1). The maximum temperature drop in late 1992 was about 0.5°C.

The 1815 Tambora eruption is probably the most renowned eruption in terms of its effect on climate. Large acidity peaks for that time in ice cores from both Greenland and Antarctica show a global distribution of the aerosol cloud, which is estimated to have been about 10^{11} kg (about five times larger than the cloud from the Pinatubo eruption). The following year, 1816, has been referred to as the "year without a summer" in Europe and northeast North America. Hudson Bay remained ice-covered that summer, and estimated average daily temperatures in this region were 5 to 6°C below the long-term average. Although these conditions were maintained only for 2 years, the cold weather from the winter of 1815–1816 through the summer of 1817 produced crop failures in China and bad harvests in India. The failure of the Indian harvest resulted in famine, which was followed by an outbreak of cholera that, over the next two decades, spread through Asia and Europe. However, the cold weather in the eastern Hudson Bay began with a series of anomalously cold years from the winter of 1811–1812, suggesting that not all the cold weather in 1816 was due to the eruption. Nevertheless, several other large eruptions occurred between 1811 and 1814, including the eruption of Vesuvius in 1813. The resulting cold temperatures could have been the cumulative result of the combined volcanic activity.

From the analysis of Greenland ice cores, we know that the interval from A.D. 1250 to 1500 and A.D. 1550 to

FIGURE 15-3 The eruption of Mt. Pinatubo in the Philippines, June 1991. (*Source:* USGS.)

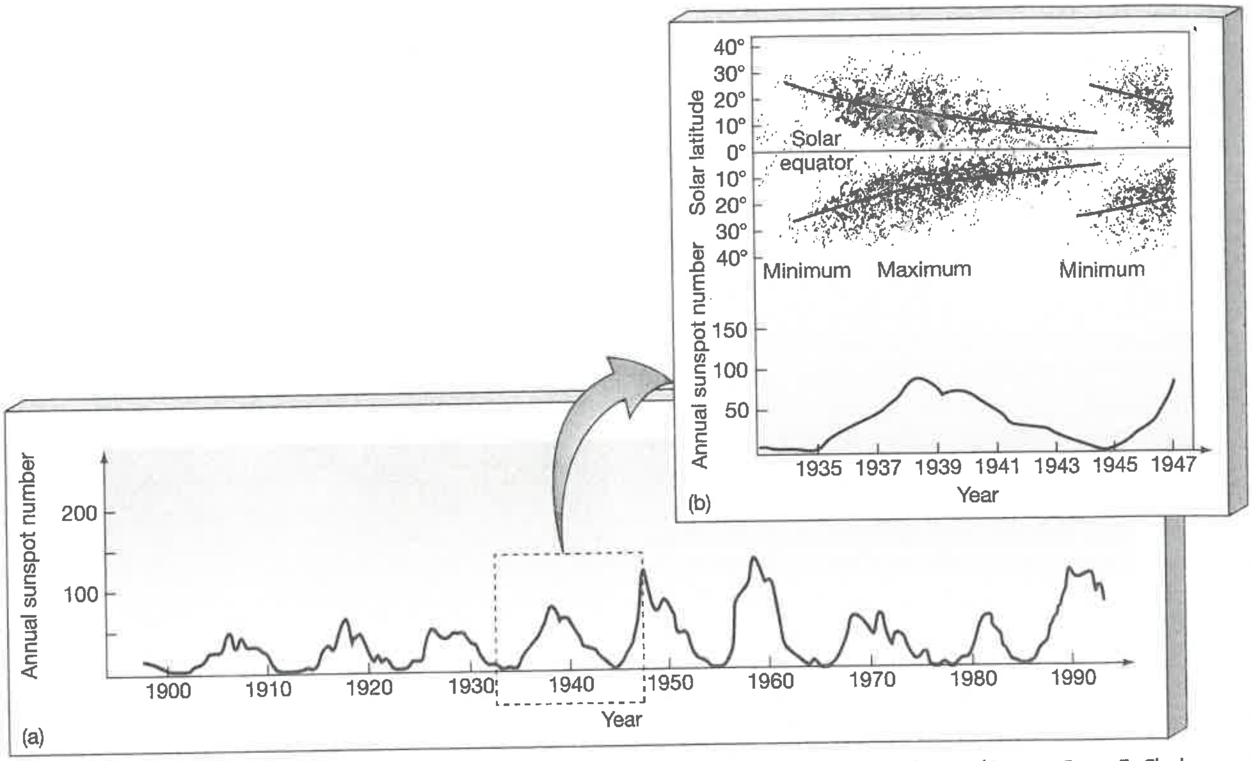


FIGURE 15-4 The butterfly diagram, showing the distribution of sunspots over the Sun's surface. (*Source:* From E. Chaisson and S. McMillan, *Astronomy: A Beginner's Guide to the Universe*, 2/e, 1998. Reprinted by permission of Prentice Hall, Upper Saddle River, N.J.)

1700 were times of high volcanic activity, whereas A.D. 1100 to 1250 had far fewer volcanic events. The first two intervals coincide with the Little Ice Age; A.D. 1100 to 1250 lies in the Medieval Warm Period. Again, the evidence suggests a link between volcanic activity and climate, but that link is still not conclusive.

Solar Variability

Another possible source of Holocene climate fluctuations is changes in solar output. We have already seen that the output of the Sun has changed considerably over the long period of Earth's history. At the short time scales we are considering here, however, we tend to regard the solar output as constant. In fact, the output from the Sun probably varies slightly at *all* time scales—but is the solar variability sufficient to explain the observed climate changes?

The greatest changes in solar output at the decade-to-century time scale accompany variations in the number of sunspots. **Sunspots** are dark areas of lower-than-normal temperatures on the surface of the Sun. Imagine that magnetic field lines wind around the Sun like elastic bands. Where these bands are twisted together, the "knot" at the surface inhibits convection from below, resulting in an area of lower temperatures (sunspots). Telescopes were first used by Galileo to observe sunspots in about 1610, but reliable sunspot data exist only for about the past 150 years. The historic record shows an 11-year cycle in sunspot activity (11 years between the peaks in sunspot abundance) superimposed on a 22-year cycle of magnetic reversals. Just like the magnetic field on Earth, the Sun's magnetic field also experiences periodic reversals, but the Sun's magnetic reversals are more frequent and regular.

The pattern of sunspot activity is illustrated by the *butterfly diagram* in Figure 15-4, which plots the location of sunspots on the Sun's surface through time. In late 1933, there were very few sunspots, but the number increased into 1937. Sunspots appeared first at higher solar latitudes, but, as they increased in abundance, the concentration shifted toward the solar equator. Sunspots then decreased in number until the next minimum in 1944, when, again, the activity switched to the higher latitudes. Plotted in this fashion, the cloud of points in Figure 15-4 resembles a butterfly wing. Note that the period of cycle is not always exactly 11 years. Even when the period is exactly 11 years from minimum to minimum, the curve is not symmetric: Less time passes between a minimum and a maximum than between the maximum and the next minimum.

Somewhat surprisingly, an increase in the abundance of sunspots, which are darker (cooler) areas on the Sun's surface, corresponds to an *increase* in the amount of solar radiation Earth receives. The Stefan–Boltzmann law (Chapter 3) suggests that the opposite should occur: An increase in sunspot abundance should result in decreased luminosity and a drop in solar radiation. The reason for this behavior is that the sunspots are surrounded by bright areas of higher-than-normal temperature called **plages**, and the area of plages is larger than the area of spots. The net result is that the Sun is actually slightly brighter during periods when sunspots increase; hence, the amount of solar radiation emitted also increases. In reality, the dark areas of the sunspots and the bright areas of the plages almost cancel each other out.

How does this variation in sunspot activity relate to Earth's climate? Studies have discovered an 11-year or a 22-year cycle in numerous climate records (records of regional floods, droughts, temperatures, and so on). However, the change in total solar output over the 11-year cycle is very small. Satellite measurements for the 1980s show that the incoming solar radiation decreased by only about 0.1% from the maximum to the minimum in sunspot activity. The direct climatic consequences of such a change are so small that they would be undetectable. At present, there are no commonly accepted mechanisms for explaining how such a small change in insolation could have a measurable impact on climate. That said, it should be pointed out that the amount of ultraviolet radiation, especially at wavelengths of 200 nm and below, changes by as much as a factor of 3 from solar maximum to solar minimum. This radiation is absorbed high up in Earth's atmosphere, and relatively little energy is involved, so it is hard to see how it could affect Earth's surface temperature. But this has not prevented researchers from proposing correlations between sunspot activity and climate.

Systematic records of sunspot activity were initiated in 1848 by Rudolf Wolf, director of the Bern Observatory, after an amateur astronomer, Heinrich Schwabe, published sunspot observations from 1826 to 1843. Wolf also compiled all of the pre-1826 data that were available, extending the record of sunspot abundance back to 1700. He also published the years of sunspot minima and maxima from 1700 back to the invention of the telescope by Galileo in 1610. The pre-1826 data are far from reliable, but they do indicate an interval between 1645 and 1715 when very few sunspots were recorded. The low sunspot abundance during this interval was noted by astronomers at the time and was discussed in detail by two solar astronomers in the late 19th century, Gustav Spörer and E. W. Maunder. This period of low sunspot activity is now referred to as the **Maunder Minimum**.

Large groups of sunspots, which occur during periods of maximum solar activity, can be seen by the naked eye (note that looking at the sun directly can damage your eyes—don't attempt to do this without the correct viewing equipment). Most of the earliest recorded sunspot observations, dating back to 28 B.C., were made in Japan, Korea, and China. More-reliable data come from another form of proxy data: the ^{14}C content of tree rings. The variations in solar output over the sunspot cycle cause changes in the magnetic properties of the solar wind (the output of energetic particles from the Sun), which is thought to affect the production of ^{14}C in Earth's upper atmosphere (see the Box "A Closer Look: Carbon-14—A Radioactive Clock" in Chapter 5). High sunspot activity increases the magnetic field strength of the solar wind, causing a greater deflection of galactic cosmic rays away from Earth. The reduction in number of cosmic rays entering Earth's atmosphere results in low ^{14}C production. Conversely, low sunspot numbers result in high ^{14}C production. The ^{14}C content recorded in annual tree rings is thought to be related to the amount of ^{14}C produced in the atmosphere.

The naked-eye observations and the proxy data indicate two other intervals when few sunspots occurred: The **Spörer Minimum** (1450–1534) and the **Wolf Minimum** (1282–1342). The fact that the Maunder and Spörer minima coincide with one of the major climate changes in the Holocene, the Little Ice Age, has led some researchers to speculate that the two may be related. Furthermore, the 12th and 13th centuries (the Medieval Warm Period) were the interval with the greatest sunspot activity in the record. Detailed studies of ^{14}C tree-ring records have also shown that midlatitude glaciers advanced and temperatures fell when sunspot numbers were low and that temperatures increased and glaciers retreated when sunspot numbers were high. Unfortunately, not all of the changes in the ^{14}C record match suspected global climate changes. So, it may be that the apparent correlations between sunspot activity and climate are simply coincidental. In any case, it is difficult to argue (as some global warming skeptics have done) that changes in solar activity have confused the climate record to such an extent that man's recent influence is obscured. Such opinions are based more on wishful thinking than on sound scientific evidence.

The Last 150 Years

It is against this backdrop of naturally occurring, short-term climate variability that we must view the surface temperature changes that have occurred more recently. Recall from Chapter 1 that the "real" surface temperature record—that is, temperatures measured with thermometers—goes back only as far as about 1850. In Chapter 1 (Figure 1-4) we saw that surface temperatures have increased by about 0.8°C since that time. We also saw that the surface temperature increase has not been uniform in time; rather, temperatures increased gradually from 1850 until 1940, then leveled out or decreased slightly from 1940 to 1970, and finally began to increase more rapidly from 1970 until the present. The slowing, or even reversal, of surface warming between 1940 and 1970 was attributed, tentatively, to cooling caused by sulfate aerosols emitted by the combustion of coal.

A question that has occurred to almost everyone who has studied this temperature record is: How much of the observed surface temperature increase over the past 150 years has been caused by humans, and how much of it might have occurred anyway? After all, if Earth was experiencing a "Little Ice Age" from 1600 until 1850, might it not simply be recovering from that unusually cold period?

To answer this question, one needs to make use of the tools of modern climatology, some of which we have mentioned in this book. In particular, modern climatologists depend extensively on complex computer models of the climate system called general circulation models (GCMs), which were mentioned earlier in the book and are discussed further in the box titled "A Closer Look: Three-Dimensional General Circulation Models (GCMs)." In recent years, much of this computer modeling work has been coordinated by the Intergovernmental Panel on Climate Change, or IPCC, as mentioned in Chapter 1. The IPCC's answer to this question is shown in Figure 15-5. Figure 15-5a shows the observed surface temperature record (black curve), along with the average of an ensemble of different model calculations (dotted blue curve). The predictions of individual climate models are shown as thin blue curves. These climate models have been executed with a combination of natural and anthropogenic forcings. The natural component includes both solar and volcanic forcings. Major volcanic eruptions (Santa Maria, Agung, El Chichón, Pinatubo) are marked with thin, gray vertical lines. After each one of these, global temperatures decrease briefly by a few tenths of a degree in both the models and the observations, before recovering. As one can see from the figure, the climate models, taken together, agree well with the observations.

What happens, though, if the anthropogenic forcing—most of which consists of increased greenhouse gas concentrations—is removed? The answer is shown in the bottom panel of Figure 15-5b. When the human influence is removed, the models all predict that the climate should have cooled slightly since 1965, whereas the data indicate that it has instead warmed by nearly 0.7°C. Although the models are admittedly not perfect, this is still strong evidence that humans are indeed perturbing the climate. The agreement among the various models and the data also demonstrates that the science of climate modeling has progressed significantly

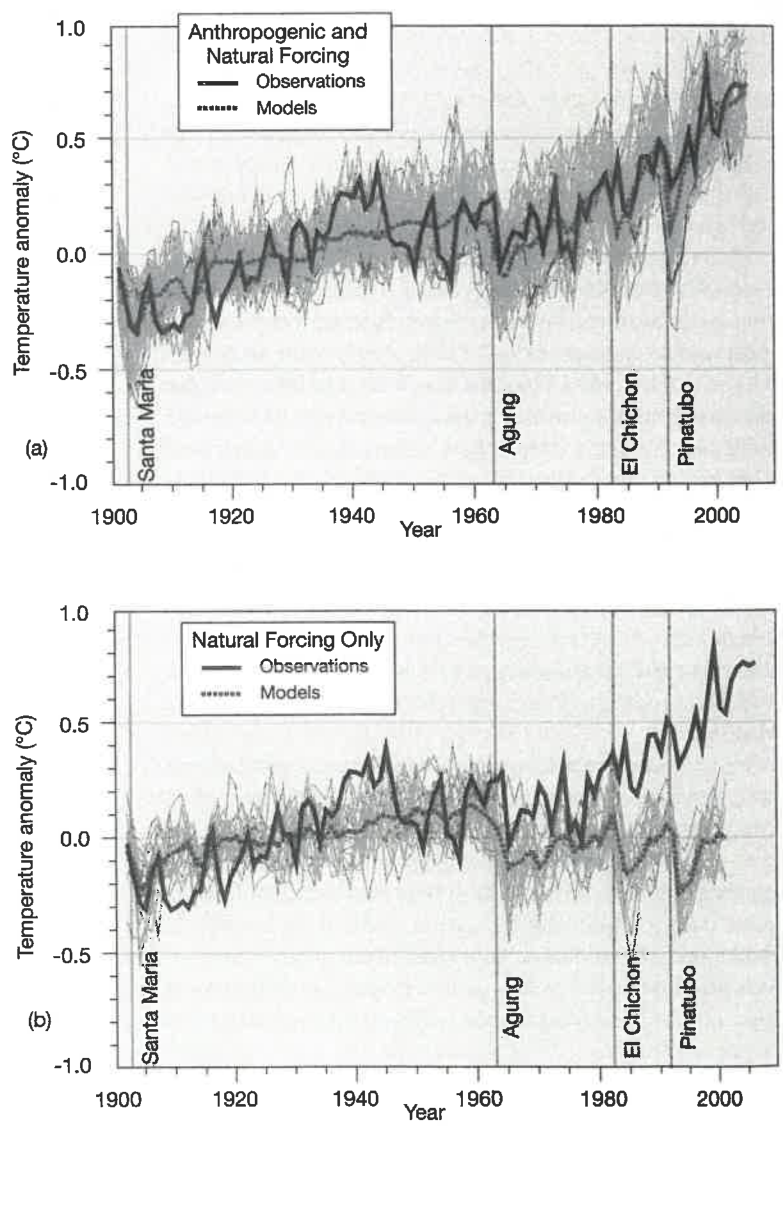


FIGURE 15-5 Global mean surface temperatures derived from observations compared to those predicted by climate models. (a) and (b) The solid black curve shows the observations. The thin curves show the range of model calculations, and the dotted blue curves show the average of those calculations. Vertical gray lines show the timing of major volcanic events. (a) The models are driven by both natural and anthropogenic forcings; that is, greenhouse gases are assumed to be increasing. (b) The models are driven by natural forcings only, and greenhouse gases are held constant. Only in (a) do the models agree with the observations.

within the past 6 years since the previous IPCC report was issued. This gives us confidence that we can predict the temperature changes that may occur in the near future.

CARBON RESERVOIRS AND FLUXES

Let us now examine in more detail how humans are perturbing the climate system. We saw in Chapter 1 (Figure 1-3) that atmospheric CO_2 concentrations have increased by about 25% since the beginning of the 19th century. This change is larger than any natural fluctuation that has occurred since the retreat of the glaciers 11,000 years ago and is almost certainly attributable to human-induced causes, principally the burning of fossil fuels and deforestation. To understand why the human influence is so noticeable, it is useful to compare the sizes of the various carbon reservoirs and the rates at which carbon cycles in both the natural and perturbed systems.

Natural Reservoirs and Fluxes

The carbon cycle is diagrammed in Figure 15-6. This diagram combines the essential features of the organic and inorganic carbon cycles described in Chapter 8. It also includes a box labeled "fossil fuels," which includes several types of compounds—most importantly coal, oil, and natural gas. Fossil fuels are, in many respects, a natural part of the carbon cycle. (It is the rate at which they are being oxidized that is unnatural.) They were created over many millions of years by the accumulation of dead organic matter in soils and sediments. Table 15-1 lists the relative amounts of different types of fossil fuel. Most of it is coal, which is stored in vast quantities in many parts of the world.

The total amount of fossil fuels is uncertain, because new reservoirs are continually being discovered and because whether a given reservoir is counted depends on the price of fuel. Estimates of the amount of fossil fuel that is economically recoverable at 2008 market prices range from 4000 to 6000 Gton(C), 7 to 10 times the amount of carbon that was present in the preindustrial atmosphere as CO_2, 590 Gton(C). The value quoted in Table 15-1, 5050 Gton(C), is in the middle of this range. These numbers, it should be noted, represent the estimated total resources, not just the proven reserves. (The proven reserves are smaller by about a factor of 3.) This gives us our first indication of why the burning of fossil fuel is a potential problem: Simply, a lot of it is available. In Critical-Thinking

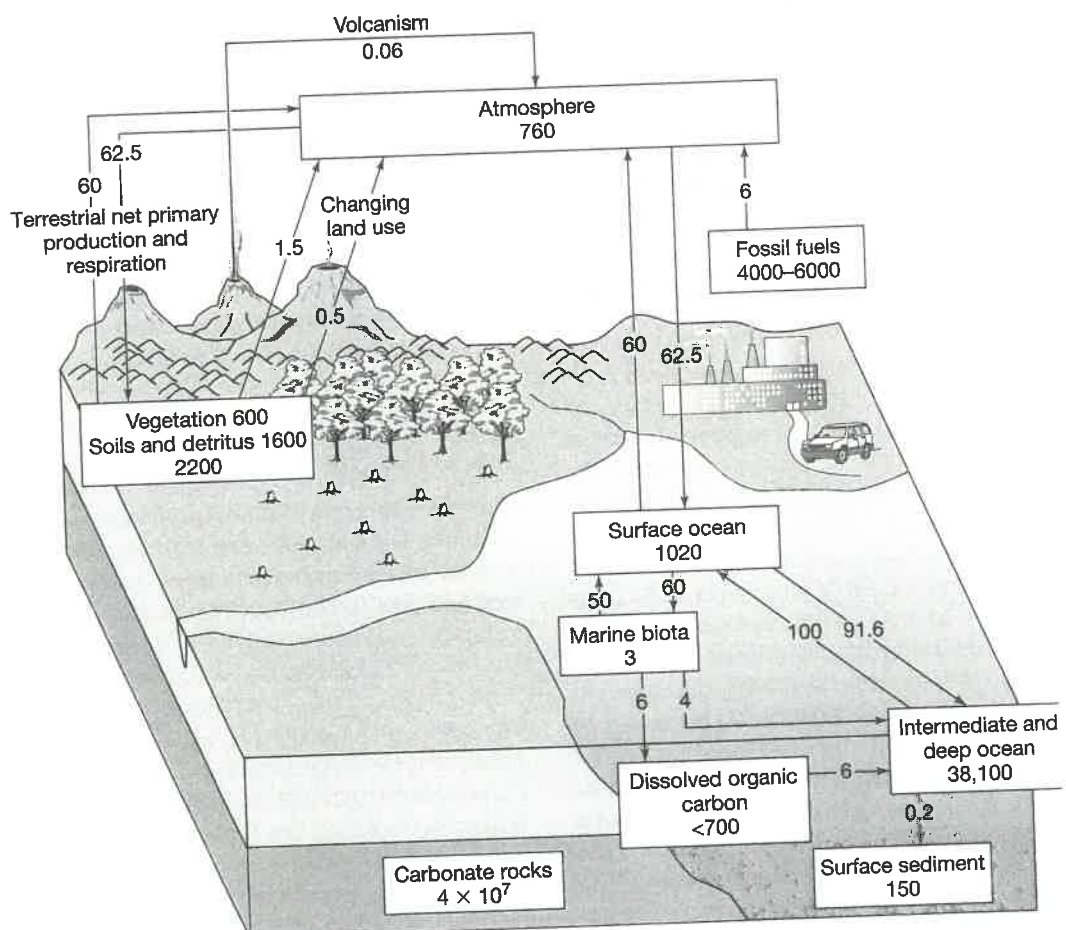


FIGURE 15-6 The major reservoirs and fluxes in the global carbon cycle. Units are Gton(C) for reservoir sizes and Gton(C)/yr for the fluxes. *(Source: Modified from Climate Change 1994: Radiative Forcing of Climate Change, Cambridge: Cambridge University Press, p. 21.)*

TABLE 15-1 Fossil-Fuel Reservoir Sizes and Burning Rates

Reservoir	Size,* Gton(C)	Burning rate, Gton(C)/yr	Projected lifetime** (yrs)
Coal	3800	2.8	1360
Oil	680	3.3	200
Natural gas	570	1.5	380
Cement production		0.3	
Total	5050	7.9	

*Reservoirs have been converted from gtoe (gigatons of oil equivalent) to Gton(C). The following conversion factors were applied: Coal: 1gtoe = 1.12 Gton(C); oil: 1 gtoe = 0.83 Gton(C); gas: 1 gtoe = 0.65 Gton(C).

**Projected lifetime = reservoir size ÷ burning rate.

Sources: Reservoir sizes from Table 9 of H.-H. Rogner, *Annual Review of Energy and the Environment* 22, November 1997, pp. 217–262, doi:10.1146/annurev.energy.22.1.217.; 2004 burning rates from U.S. Carbon Dioxide Information Analysis Center (CDIAC) website, http://cdiac.ornl.gov/trends/emis/tre_glob.htm.

Problem 1, we shall compute what would happen to atmospheric CO_2 concentrations if all the fossil fuel were to be burned instantaneously.

Two additional points should be noted about the fossil-fuel reservoirs listed in Table 15-1. First, the value given for oil reserves, 680 Gton(C), includes both conventional and unconventional reserves. Conventional oil reserves, that is, oil that can be pumped directly out of the ground, make up only 240 Gton(C) of this amount. The rest of the oil is in the form of oil shales, tar sands, and heavy crude oil that is more difficult to access and process. Second, the natural gas values include both conventional (easy to access) and unconventional reserves also, but they do not include methane hydrates. Over the past decade, oceanographers have discovered vast quantities of methane-rich ice on the seafloor just off the continental shelves. Technically, this material is termed **methane clathrate hydrate**. It is a solid in which CH_4 molecules are encased in a lattice of 5 to 6 H_2O molecules. Such methane hydrates form when organic matter decomposes anaerobically at great depths. The overlying water must be at least 300 m deep and must be cold as well in order for methane hydrates to be stable. The total amount of methane hydrate on the seafloor is uncertain but some estimates are as high as 18,000 Gton(C), or more than 3 times higher than the rest of the fossil-fuel reserves! We do not list this in Table 15-1 because it is not recoverable at current market prices, but if someone discovers how to extract it economically the problem of global warming will become even more difficult.

Of equal importance to the carbon reservoirs themselves are the fluxes that connect them. These fluxes are in many cases wholly disproportionate to the size of the reservoirs involved. For example, the huge carbonate rock reservoir is connected to the rest of the system by only a tiny flux, about 0.06 Gton(C)/yr. Shown in Figure 15-6, this volcanic CO_2 flux must, on average, equal the rate at which carbonate sediments are preserved. Despite its large size, this reservoir exerts little influence on the rest of the carbon cycle on any time scale shorter than hundreds of thousands of years. So, processes such as silicate weathering and volcanism, which were extremely important on geologic time scales, should have little effect on atmospheric CO_2 levels over the next few centuries.

The largest fluxes in the natural carbon cycle are the exchange of carbon between the atmosphere and the terrestrial biosphere, about 60 Gton(C)/yr, and between the atmosphere and surface ocean, about 60 Gton(C)/yr. The terrestrial biosphere, largely composed of terrestrial vegetation and soils and detritus, exchanges CO_2 with the atmosphere by way of the short-term organic carbon cycle. In that cycle, the photosynthetic uptake of CO_2 by forests is balanced by the respiration and decay of plant material. This cycle is responsible for the seasonal fluctuations in atmospheric CO_2 (the Keeling curve) discussed in Chapters 1 and 8. The ocean exchanges CO_2 with the atmosphere by diffusion through the surface interface. Once in the surface ocean, CO_2 is taken up by marine photosynthesizers. The net rate of photosynthesis in the oceans, about 60 Gton(C)/yr, is comparable to that on land, but it does not affect the atmosphere in the same way, because the amount of living biomass is very small: The marine biota contain only about 3 Gton of carbon, as compared with approximately 610 Gton of carbon in forests. Free-floating marine organisms and seaweed do not need the massive amounts of structural carbon required by trees. As a result, the marine organic carbon cycle is more closely balanced than the terrestrial cycle and does not contribute appreciably to the seasonal fluctuations of CO_2 in the atmosphere.

Rates of Fossil-Fuel Burning and Deforestation

The currently observed increase in atmospheric CO_2 concentrations is attributable, at least in part, to the combustion of fossil fuels. For natural gas (which is mostly methane), the combustion reaction can be written as

$$CH_4 + 2\,O_2 \rightarrow CO_2 + 2\,H_2O.$$

The combustion reactions of coal and oil are more complicated, as these fuels consist of mixtures of more-complex hydrocarbons, but the main reaction products are the same, CO_2 and H_2O. The rate of fossil-fuel consumption is known fairly accurately (to within 10% or better) because records are kept by companies that produce fossil fuels and by the countries in which these companies operate. The 2004 value was about 7.9 Gton(C)/yr (Table 15-1). (2004 is the last year for which complete information concerning emissions is available from the U.S. Carbon Dioxide

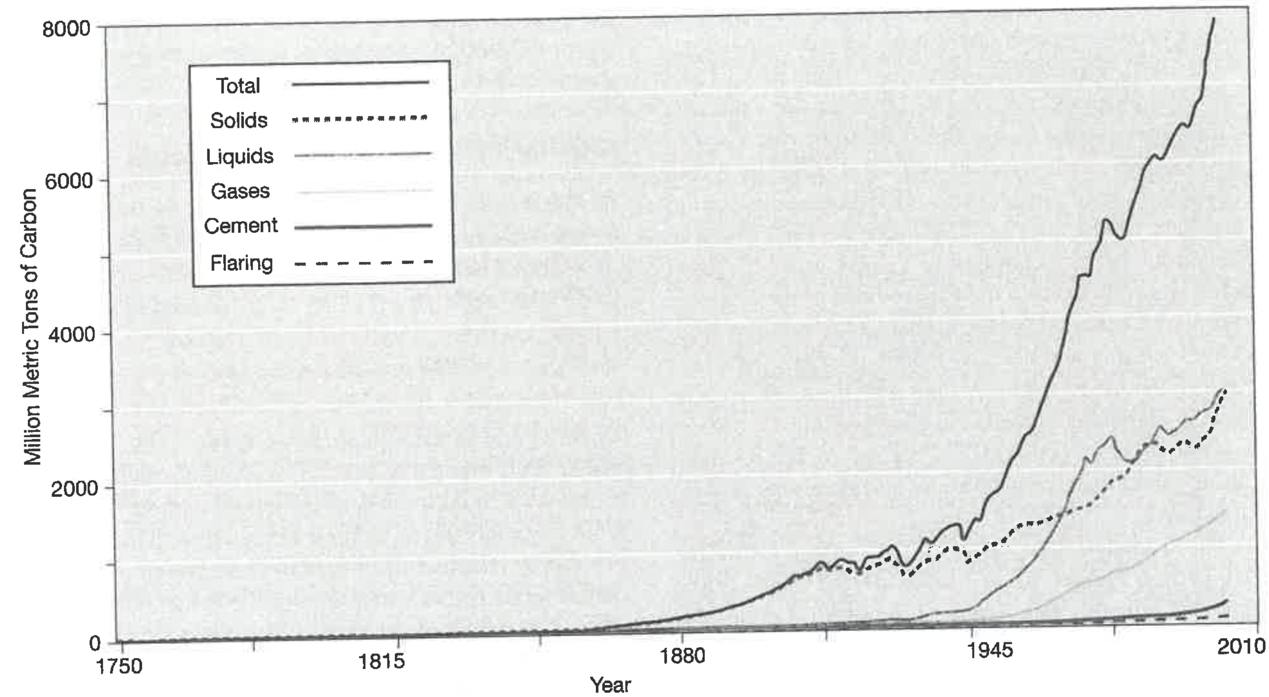


FIGURE 15-7 Coal, oil, and natural gas consumption rates, 1750–present. (*Source:* Carbon Dioxide Information Analysis Center, Oak Ridge National Labs, available online at http://cdiac.ornl.gov/trends/emis/glo.htm.)

Information Analysis Center. We will cite some more recent figures below.) World population is currently about 6 billion, so this consumption rate amounts to 1.3 metric tons of carbon per person per year. The rate of fossil-fuel burning is about 10 times less than the rate of CO_2 exchange with the terrestrial biosphere but 100 times larger than the rate of CO_2 release by volcanos. Its effect on the atmosphere is disproportionately large, however, because this part of the global carbon cycle is not in balance.

All three of the major fossil fuels are being consumed at appreciable rates, but oil leads the way at 3.3 Gton(C)/yr, followed by coal at 2.8 Gton(C)/yr and natural gas at 1.5 Gton(C)/yr (Table 15-1 and Figure 15-7). This implies that the different forms of fossil fuel have vastly different projected lifetimes. Recall from Chapter 8 that the residence time of a reservoir is just the reservoir size divided by the outgoing flux. As fossil fuels are not forming at appreciable rates, their residence times indicate how long they will last at current burning rates. From Table 15-1, we see that oil has the shortest projected lifetime, ~200 years, while coal has the longest, ~1,360 years. These values can be misleading, though. Conventional oil will be exhausted in less than 75 years at current burning rates and could be gone much sooner than that if world oil consumption increases. Global warming is not the only potential problem on the horizon. Depletion of oil reserves is also something that must be considered.

Fossil-fuel consumption is distributed unequally among the various nations of the world. As shown in Table 15-2, most of the fossil fuel is burned in the Northern Hemisphere, with North America and the Far East essentially tied for the lead at nearly 1.8 Gton(C)/yr apiece. The United States, with only about 5% of the world's population, accounts for some 1.65 Gton(C)/yr, or 21% of global CO_2 emissions. Thus, our *per capita* emissions (emissions per person) are about 4 times the world average. The large amount of CO_2 emitted reflects our high standard of living and energy-intensive lifestyle. Other countries are catching up, however. China, in particular, has a rapidly growing economy and an increasing appetite for fossil fuels to go with it. With its large population, extensive coal reserves, and newfound thirst for oil as well, China became the world's largest CO_2 emitter in 2006–2007.

TABLE 15-2 Fossil-Fuel Consumption by Geographic Region

Region	Consumption rate, Gton(C)/yr
North America	1.8
Central and South America	0.4
Western Europe	0.9
Eastern Europe (incl. Russia)	0.8
Middle East	0.4
Africa	0.3
Oceania (Australia and Japan)	0.4
China and Vietnam	1.4

Source: 2004 burning rates from U.S. Carbon Dioxide Information Analysis Center (CDIAC) website, http://cdiac.ornl.gov/trends/emis/tre_glob.htm.

The increase in global CO_2 emissions over the past 15 years has been so rapid that it deserves additional comment. In 1990, global CO_2 emissions from fossil fuels were about 6.2 Gton(C)/yr. That was the year for which the emission limits for the Kyoto Protocol, to be discussed in the next chapter, were standardized. (Actually, the Kyoto agreement asks developed countries to reduce emissions to 5% less than 1990 values by 2020—but we shall save that discussion for later.) According to a 2007 paper by Josep Canadell and colleagues in the Proceedings of the National Academy of Sciences (PNAS), global CO_2 emissions from fossil-fuel burning and cement production for 2006 were 8.4 Gton(C)/yr, or 35% higher than 1990 values! Such an increase in emissions exceeds the most pessimistic projections of the previous (2001) IPCC report. We are proceeding along the climate change highway faster than anyone had imagined.

The Canadell study provides an estimate for the global carbon budget for the time period 2000–2006. During this interval, an average of about 7.6 Gton(C)/yr was produced by the combustion of fossil fuels and by cement production. An additional 1.5 Gton(C)/yr was produced by deforestation, primarily in the tropics. The clearing of forests and the utilization of land for agricultural purposes generally results in a substantial release of carbon into the atmosphere, both from the trees themselves and from the soil beneath them. Thus, the total anthropogenic CO_2 source during this interval was about 9.1 Gton(C)/yr. Of this, about 4.1 Gton(C)/yr accumulated in the atmosphere. This number can be calculated directly from the observed 1.9 ppm/yr rate of atmospheric CO_2 increase. (We shall do so in Critical-Thinking Problem 2.) Another 2.2 Gton(C)/yr is estimated to have been absorbed by the oceans, as the imbalance in fluxes between the atmosphere and the surface ocean (Figure 15-6) indicates. The remainder, about 2.8 Gton(C)/yr, is thought to be accumulating in forests and soils. So, surprisingly, despite the deforestation that is occurring in the tropics, the biosphere as a whole is currently acting as a net sink for carbon! The reasons are discussed below.

CO_2 REMOVAL PROCESSES AND TIME SCALES

How fast the CO_2 released from fossil-fuel burning will disappear and where it will ultimately go are perhaps the most confusing aspects of the whole global warming issue. The confusion arises because the carbon cycle is complex and involves processes that operate on a variety of time scales. Scientists have developed computer models of the carbon cycle that are capable of simulating many of these processes and that can be used to make projections of future atmospheric CO_2 concentrations. Some of these model predictions will be described in the following section. To understand these predictions, however, and to gain some idea of how reliable they might be, we must consider the physical processes involved. Next, we describe the major CO_2-uptake processes in decreasing order of the speed at which they are expected to occur.

Northern Hemisphere Reforestation

As the previous carbon budget numbers demonstrate, the fastest uptake process for anthropogenic CO_2 is photosynthesis by the terrestrial biosphere. As discussed in Chapter 8, we can represent photosynthesis by the simplified reaction

$$CO_2 + H_2O \rightarrow CH_2O + O_2,$$

where CH_2O represents more-complex forms of organic matter. This organic matter can accumulate either in living biomass or in soils. When it does so, carbon is removed from the atmosphere, at least temporarily. This carbon is eventually returned to the atmosphere when trees are cut down and burned or when soil organic matter decays.

Whether photosynthesis will act as a sink for anthropogenic CO_2 over the next few centuries depends largely on whether forests expand or shrink in size. Deforestation causes them to shrink, so the terrestrial biosphere acts as a CO_2 source. Reforestation causes them to expand, so the biosphere acts as a CO_2 sink. Studies suggest that a significant amount of reforestation, about 0.5 Gton(C)/yr, is occurring in temperate parts of the Northern Hemisphere. Recall from Chapter 1 that deforestation of North America during the 19th century, the *pioneer effect,* was responsible for most of the rise in atmospheric CO_2 between 1800 and 1850. Much of the deforested land was converted to farms and remains farmland today, so that land cannot be where the carbon is going. But the mountain ridges in Pennsylvania were stripped of trees to provide fuel for making steel and for powering trains. The demand for wood is now lower, and many areas are protected from logging by the state and federal governments. Consequently, the forests are regrowing and are probably contributing to the Northern Hemisphere CO_2 sink. Most of these forests will not reach maturity for another century or more, so this sink should remain active for some time into the future.

Northern Hemisphere forest regrowth, however, cannot be the only place where CO_2 is disappearing. According to the carbon budget numbers presented, the total biospheric uptake of CO_2 is approximately 2.8 Gton(C)/yr. Northern Hemisphere reforestation is only about 20% of that figure. Where is the rest of the CO_2 going?

CO_2 and Nitrogen Fertilization

A substantial part of the missing carbon may be going into existing forests as a consequence of higher atmospheric CO_2 concentrations. Most plants raised under greenhouse conditions, where plenty of water and other nutrients are available, grow faster when exposed to higher CO_2 levels. (But about 5% of plants, termed C_4 plants, do *not*, as discussed in the next chapter.) This process is termed **CO_2 fertilization**. The increased growth rate is caused, in part, by the fact that CO_2 is a limiting nutrient under these conditions. But there is an indirect stimulation effect as well. Plants tend to use water more efficiently under elevated CO_2 conditions. Plants have small openings, called *stomata,* on the undersides of leaves that allow air to pass in and out of them; these stomata need not open as wide as normal when CO_2 levels are high. As CO_2 from the atmosphere enters the leaf, water vapor from inside can escape. Not opening their stomata as wide allows plants to survive under drier conditions, allowing the plants to grow faster under high CO_2 levels. The number of stomata per leaf also tends to decrease in plants grown under elevated CO_2 concentrations. Whether forests and other natural ecosystems should respond similarly to CO_2 fertilization is a question that has aroused considerable debate. Ecologists have noted that many natural ecosystems are limited by other factors, such as nutrient availability and competition for sunlight. In such cases, higher atmospheric CO_2 concentrations should have little effect on plant growth. Even if CO_2 fertilization is occurring, other factors might come into play in the future as the climate warms. One concern is that soil carbon might decay more rapidly under such conditions. Tropical soils, for example, are deficient in carbon as a result of rapid rates of decomposition. Because most of the carbon in temperate forests resides in the soil rather than in the trees themselves, the total amount of carbon stored in forest ecosystems could actually decrease in the future even if the trees themselves grow faster. In any case, most computer model simulations now suggest that the terrestrial biosphere will become less efficient at absorbing CO_2 as the climate warms. This could lead to a faster rate of CO_2 accumulation in the atmosphere.

A related phenomenon that might be encouraging CO_2 uptake by the terrestrial biosphere is **nitrogen fertilization**. Nitrogen is an essential nutrient for all organisms, and it is often in short supply because it is difficult to convert atmospheric N_2 into fixed nitrogen that organisms can use. Humans have been helping out in this regard by adding nitrogen fertilizer to agricultural fields and by emitting large amounts of nitrogen oxides from combustion. Agricultural activity does not normally lead to net uptake of CO_2, because the crops that are grown are harvested and eaten and the remaining organic matter is burned or decays. But the nitrogen oxides released into the atmosphere are oxidized to nitric acid, HNO_3, and are removed by precipitation. If the concentration of nitric acid in rainwater becomes too high, the resulting acid rain is harmful to plants and to other organisms, especially fish (although nitric acid generally contributes less to acid rain than does the sulfuric acid generated from SO_2 released by coal burning). At more modest concentrations, the nitric acid becomes a source of fixed nitrogen, which can stimulate plant growth. So, some of the increased forest growth currently taking place may be a response to anthropogenic nitrogen emissions.

Dissolution in the Oceans

The next-fastest mechanism for removing anthropogenic CO_2 from the atmosphere is dissolution in the oceans. This process, too, is complex, because CO_2 reacts chemically when it dissolves, unlike N_2 or O_2. The chemical process by which the oceans dissolve CO_2 is described in the Box "A Closer Look: The Chemistry of CO_2 Uptake." According to Figure 15-6, the flux of CO_2 between the atmosphere and oceans is on the order of 60 Gton(C)/yr, so the residence time for atmospheric CO_2 with respect to this process, t_{OA}, is on the order of 12 years.

This time scale, however, is somewhat misleading. Recall that the ocean can be thought of as consisting of two layers: a well-mixed surface layer approximately 75 m thick and a poorly mixed deep-ocean layer nearly 4 km thick. These layers can be represented schematically by separate boxes, as shown in Box Figure 15-1. Only the shallow surface layer is capable of rapid CO_2 exchange with the atmosphere. Because of its small volume, its CO_2 uptake capacity is relatively low. The uptake capacity is determined largely by the abundance of carbonate ion (see the Box "A Closer Look: The Chemistry of CO_2 Uptake"). The deep ocean has a much larger volume and, hence, a much larger CO_2 uptake capacity, but its turnover time, t_{SD}, is on the order of 1,000 years. So, CO_2 exchange between the atmosphere and the oceans occurs on a variety of time scales, ranging from 8 years to more than 1,000 years. Because the uptake of CO_2 by the oceans is a chemical process, the lifetime of anthropogenic CO_2 depends on how much of it we produce. The first puffs of CO_2 released at the dawn of the industrial age were taken up almost immediately by the surface ocean. But the more CO_2 we release, the deeper it must penetrate into the ocean in order to be buffered by reaction with carbonate ion. The lifetime of CO_2 released today is estimated to be only about 60 years, but the lifetime of CO_2 released in the future is predicted to be much longer. Computer models that take this chemistry into account are required to calculate how fast the ocean will actually take up anthropogenic CO_2.

The limited CO_2 uptake capacity of the surface ocean can be viewed in another way—one that makes clear its importance for marine ecosystems. In reality, one can continue to pump CO_2 into seawater, even after its natural carbonate ion content is depleted. But doing so causes the pH of the water to decrease substantially, that is, it makes the water more acidic. And this, in turn, can be destructive to organisms such as foraminifera or corals which form shells or other habitats out of calcium carbonate, because it causes them to dissolve. Model simulations suggest that the pH of the surface ocean has already dropped by about 0.1 unit since the beginning of the industrial revolution in 1850. Because pH is evaluated on a log scale ($pH = -\log[H^+]$), this corresponds to a 30% increase in the H^+ concentration. One simulation discussed in the 2007 IPCC report (Vol. 1, p. 529) suggests that

A CLOSER LOOK

The Chemistry of CO_2 Uptake

The rate at which CO_2 can be taken up by the oceans depends on ocean chemistry as well as ocean mixing. The reason is that CO_2 does not simply dissolve in seawater as would a gas such as N_2 or O_2. As we discussed in Chapter 8, when CO_2 dissolves in water, it forms carbonic acid, H_2CO_3, which then dissociates into bicarbonate ions, HCO_3^-, and carbonate ions, CO_3^{2-}. Long before humans began perturbing the Earth system, the ocean contained substantial quantities of carbonate and bicarbonate ions as part of the natural inorganic carbon cycle. The presence of these ions in solution makes it possible for seawater to absorb more anthropogenic CO_2 than would otherwise be possible. The reason is that these ions, carbonate ion in particular, moderate the change in the ocean's acidity as CO_2 is added. A chemist would say that they serve as a pH **buffer**, a dissolved substance that helps maintain a stable pH. At pH values that are typical of the surface ocean (pH of about 8), the chemical reaction that occurs can be written as

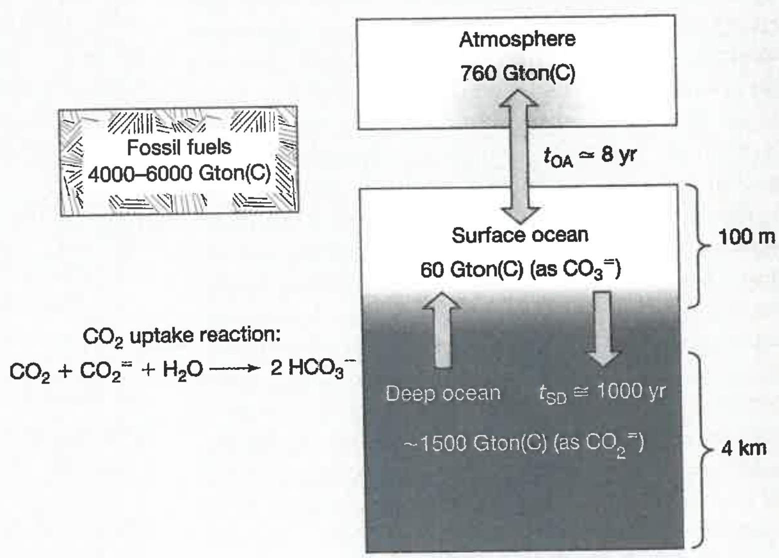
$$CO_2 + CO_3^{2-} + H_2O \rightarrow 2\,HCO_3^-$$

Each anthropogenic CO_2 molecule that enters the ocean combines with one carbonate ion and one water molecule, yielding two bicarbonate ions. A similar reaction converts borate ion ($H_2BO_3^-$) into boric acid (H_3BO_3). The presence of borate increases the ocean's buffering capacity by an additional 25%. The fact that such chemical reactions occur implies that the ocean's capacity to absorb CO_2 is limited. If the amount of anthropogenic CO_2 added to the ocean exceeds the amounts of carbonate ion and borate ion that were initially present, the ocean's buffering capacity will be exhausted and its ability to absorb CO_2 will be greatly diminished. We can estimate the CO_2 uptake capacity of the ocean by measuring the dissolved carbonate ion concentration in the surface and deep ocean and multiplying by the volumes of the respective reservoirs. (We shall do so in "Critical-Thinking" Problem 3.) The results are shown in Box Figure 15-1. The surface ocean has only a small buffering capacity compared with the amount of CO_2 that could be produced from the burning of fossil fuels. The deep ocean contains enough carbonate and borate ion to react with approximately 30% of the fossil-fuel reservoir.

The oceans can also absorb anthropogenic CO_2 by dissolving carbonate sediment on the seafloor. The chemical reaction involved is

$$CO_2 + CaCO_3 + H_2O \rightarrow Ca^{++} + 2\,HCO_3^-$$

This reaction is similar to that by which seawater itself absorbs CO_2, except that the required carbonate ion is initially attached to a calcium ion. Neither of these two processes is a permanent sink for CO_2 because, even after the reactions have occurred, the CO_2 is still present in the oceans as bicarbonate. This bicarbonate will eventually be removed when enough calcium ions have been provided by silicate weathering to reprecipitate it as carbonate sediments. Only then will the anthropogenic CO_2 truly be gone.



BOX FIGURE 15-1 Two-box ocean model illustrating the capacity of the ocean for CO_2 uptake. The numbers in the ocean boxes represent the amount of carbonate ion that can react with CO_2.

surface ocean pH could drop by more than 0.7 unit—corresponding to a fivefold increase in $[H^+]$—over the next few centuries if CO_2 emissions continue unabated. This would likely wreak havoc with marine ecosystems by making it extremely difficult for carbonate-secreting organisms to survive.

Dissolution of Seafloor Carbonates

Carbon dioxide can also be taken up by the dissolution of carbonate sediments on the seafloor, spurred by the higher H^+ content of CO_2-enriched seawater. This may at first seem surprising, as we learned previously that the precipitation of carbonate sediments, in conjunction with silicate weathering, removes CO_2 from the atmosphere–ocean system. Carbonate sediments are the long-term sink for CO_2. On shorter time scales, however, just the opposite occurs: Atmospheric CO_2 is taken up when carbonate sediments dissolve, because both compounds are converted to bicarbonate.

Carbonate sediments dissolve when CO_2-enriched seawater comes in contact with the sediments. For this to occur, the CO_2-rich water needs to be carried down into the deep ocean. As discussed earlier, this is a slow process, requiring many hundreds of years. Furthermore, the sediments on the seafloor need to be stirred to expose fresh surface area for the reaction. This stirring, or **bioturbation**, is accomplished by burrowing organisms, such as worms, that make their homes in the sediments. Bioturbation occurs primarily in the uppermost 10 cm of sediments, but repeated episodes of carbonate dissolution followed by renewed burrowing can eventually cause the uppermost 40 to 50 cm of marine sediments to dissolve. As Wallace Broecker of Lamont-Doherty Earth Observatory has observed, "even the worms will do their part in [taking up fossil-fuel CO_2]!" Thus, seafloor carbonate dissolution may eventually play an important role in CO_2 removal, but it is not likely to prevent atmospheric CO_2 from rising over the next few decades to centuries.

Weathering of Continental Rocks

The slowest, but most permanent, sink for anthropogenic CO_2 involves the weathering of silicate rocks on the continents, followed by the precipitation of carbonate sediments on the seafloor. As we discussed in Chapter 8, the combination of these two processes can be represented by the chemical reaction

$$CaSiO_3 + CO_2 \rightarrow CaCO_3 + SiO_2,$$

where $CaSiO_3$ represents a variety of more-complicated silicate minerals. This process is much slower than the carbonate dissolution process discussed because it requires that Ca^{2+} ions produced by weathering accumulate in the ocean. The time required to precipitate anthropogenic CO_2 as carbonate can be estimated by dividing the total amount of carbon in the combined atmosphere–ocean system, about 38,000 Gton, by the rate at which CO_2 is consumed by silicate weathering, 0.06 Gton(C)/yr. This time scale, which is in excess of half a million years, is the characteristic response time of the carbonate–silicate cycle. Our human-induced perturbation to the natural carbon cycle is likely to last at least that long.

Carbonate rocks on the continents can also be weathered and dissolved. This process is analogous to seafloor carbonate dissolution and provides another "temporary" sink for CO_2. The CO_2 is not really gone, because it is stored in the oceans as bicarbonate. Carbonate rocks dissolve more rapidly than do silicate rocks, so this loss process could become important on time scales of only a few thousands of years.

PROJECTIONS OF FUTURE ATMOSPHERIC CO_2 CONCENTRATIONS AND CLIMATE

Once we have understood the present-day sources and sinks for CO_2, the really difficult task begins: We need to project this information into the future to try to estimate future atmospheric CO_2 levels. Doing so involves making various assumptions about how much fossil fuel people will consume and how much deforestation/reforestation will take place. If we input this information into a computer model of the global carbon cycle that includes the various CO_2 removal processes just described, we can attempt to predict how atmospheric CO_2 concentrations will change. Then, this information can be used in global climate models to predict how future climate may be affected. However, we must consider other greenhouse gases as well because CO_2 is not the only greenhouse gas that is increasing.

Atmospheric CO_2 Levels for Different Emission Scenarios

Rather than try to make our own estimates for how much fossil fuel will be burned over the next few decades, we will rely on projections made, or adopted, by the IPCC. Although many different scenarios have been analyzed, the 2007 report focused on three in particular, labeled A2, A1B, and B1. These correspond, respectively, to high, medium, and low rates of future CO_2 emissions. The projected emission rates and corresponding predicted atmospheric CO_2 concentrations are shown in Figure 15-8.

Let us begin with the most optimistic scenario. In the "low" emissions case, CO_2 emissions rise from their current level of about 8 Gton(C)/yr to a peak of about 12 Gton(C)/yr in 2040, then gradually decrease to under 5 Gton(C)/yr by the end of the century (Figure 15-8b). Atmospheric CO_2 for this case rises gradually to about 550 ppm by the end of the century. This is very close to double the preindustrial CO_2 concentration of 280 ppm. Atmospheric CO_2 looks as if it has nearly stabilized by this time; however, as we will see later, that would happen only if emissions were cut to approximately 2 Gton(C)/yr over the next two centuries.

In the "medium" scenario, CO_2 emissions rise to considerably higher values, about 17 Gton(C)/yr (or twice the current value) by 2050, then decline to about 14 Gton(C)/yr by 2100. Atmospheric CO_2 climbs more rapidly in this case, reaching just over 700 ppm by the end of the century. CO_2 concentrations are clearly still growing at this time, so further increases and associated climate warming would be expected. CO_2 doubling from preindustrial levels occurs by the year 2050.

In the "high" scenario, CO_2 emissions track the medium case up through 2050, but then continue to climb

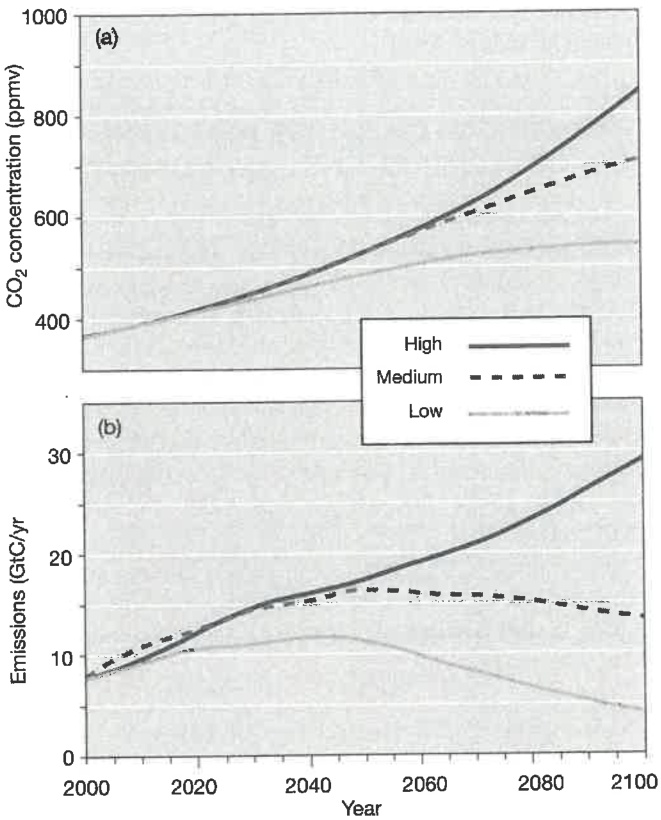


FIGURE 15-8 Estimated CO_2 emissions (b) and atmospheric CO_2 concentrations (a) for the next century for different assumptions about population and economic growth. (*Source:* http://www.globalwarmingart.com/wiki/Image:Carbon_Dioxide_Emissions_Scenarios_png.)

throughout the latter half of the century, reaching nearly 30 Gton(C)/yr by 2100. The cumulative CO_2 emissions over the next century in this case are roughly 2000 Gton(C), or almost half the recoverable fossil-fuel reserves. (By comparison, about 500 Gton[C] had already been emitted prior to 2006.) Atmospheric CO_2 concentrations increase steeply as well, climbing to almost 850 ppm by this time. Although this scenario may seem extreme, it was not the most pessimistic case considered by the IPCC. The A1F1 scenario (not shown) pushed atmospheric CO_2 concentrations to nearly 1000 ppm by the end of the century. Considering that we have already exceeded the IPCC's highest emissions estimate over the past 5 or 6 years, such a scenario is not out of the question.

By how much would we expect climate to change in these various cases? We can derive a preliminary answer from the one-dimensional, radiative–convective climate models discussed in Chapter 3, and then return to the question with more sophisticated models later once we've assessed the projected trends in other greenhouse gases. Doubling the atmospheric CO_2 concentration in such a model produces about a 2.5°C increase in global mean surface temperature when the water vapor feedback is included. This temperature increase could occur by the year 2050 in the IPCC's high- and medium-emissions cases or by 2100 in the low-emissions case. Doubling CO_2 a second time would produce approximately the same amount of warming, yielding 5°C total. The high scenario does not quite reach quadrupled CO_2 by the year 2100, so we should reduce this number slightly. Hence, we estimate that global temperatures could increase by as much as 4°C (7°F) over the next century in the worst-case scenario. In the most favorable case, the warming would be about 2°C. Both of these numbers should be compared with the warming of 0.8°C that has taken place over the past 100 years. Evidently, climate change during the next century is likely to be considerably more rapid than it was in the last century. A prediction of this nature, however, is not considered good enough for making policy decisions because it overlooks several factors that should affect Earth's climate over the next few decades. These factors include radiative forcing by trace gases other than CO_2, the oversimplification of one-dimensional models, and the thermal properties of the ocean.

Increases in Other Trace Gases

Carbon dioxide is not the only atmospheric greenhouse gas that is currently increasing in concentration. Methane and nitrous oxide have also been increasing in concentration over the past 200 years, as evidenced by measurements of their concentrations in air bubbles trapped in polar ice cores (See Chapter 1, Figure 1-3). Methane has strong anthropogenic sources, mostly cattle raising and rice cultivation, that account for 60 to 70% of its total emissions. Consequently, methane has more than doubled from its preindustrial concentration of 750 ppb to a modern value of nearly 1800 ppb (1.8 ppm). About 30% of nitrous oxide emissions come from bacterial denitrification in fertilized soils; the remaining 70% is natural. Hence, the increase in N_2O concentrations has been more modest—only about 50 ppb, or roughly 20% since preindustrial times. Chlorofluorocarbon compounds (CFCs) have also been increasing over the past several decades, but here the future projections are quite different. Most conventional CFCs, such as freon-11 and freon-12, have now been banned in order to protect the ozone layer (see Chapter 17). Thus, they are not expected to contribute to future global warming. Freon replacement gases are beginning to accumulate, and these gases could eventually contribute to global warming, but so far this is not a big problem.

Interestingly, the increase in atmospheric methane concentrations appears to have slowed, or even stopped, over the past few years. Figure 15-9a shows the measured, global average CH_4 concentration from 1983 until 2006. Figure 15-9b shows the rate of change in CH_4 (technically, the derivative, for those who are familiar with calculus). Evidently, the rate of change in CH_4 concentrations has oscillated around zero since about the year 2000. This may reflect the fact that most arable land, especially that suitable for growing rice, is already being cultivated. This

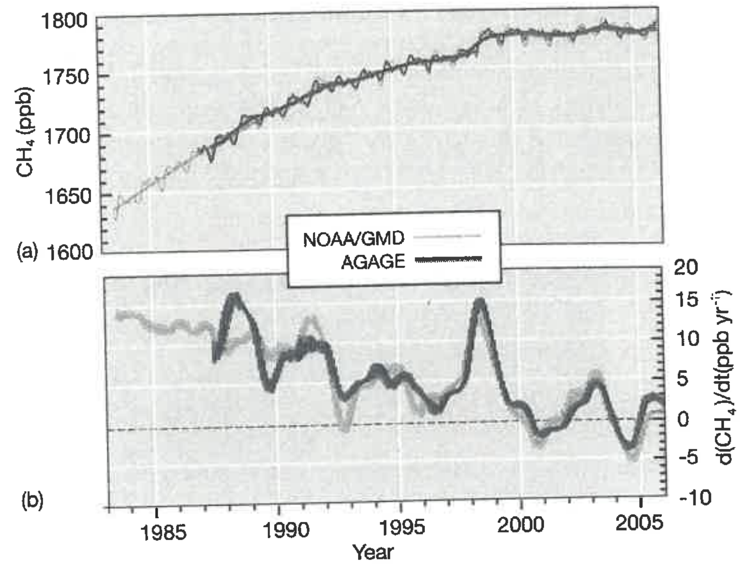


FIGURE 15-9 (a) Globally averaged atmospheric CH_4 measurements since 1983. (b) The rate of change of the measurements shown in (a). Figure 15-9 shows that the increase in CH_4 has virtually stopped since the year 2000.

suggests that while CH_4 may have contributed to global warming over the past century, its additional effect over the next century may be minimal.

By contrast, nitrous oxide concentrations have continued to rise over this same time period (Figure 15-10). Because N_2O production is enhanced by use of nitrogen-based fertilizers, its production rate can continue to increase even if the total area of cultivated land remains constant. Fortunately, because of its relatively low abundance, N_2O is less of a greenhouse threat than either CO_2 or CH_4, as we shall see below.

The Concept of Radiative Forcing

The contributions of each of these different gases to the atmospheric greenhouse effect can be measured in terms of

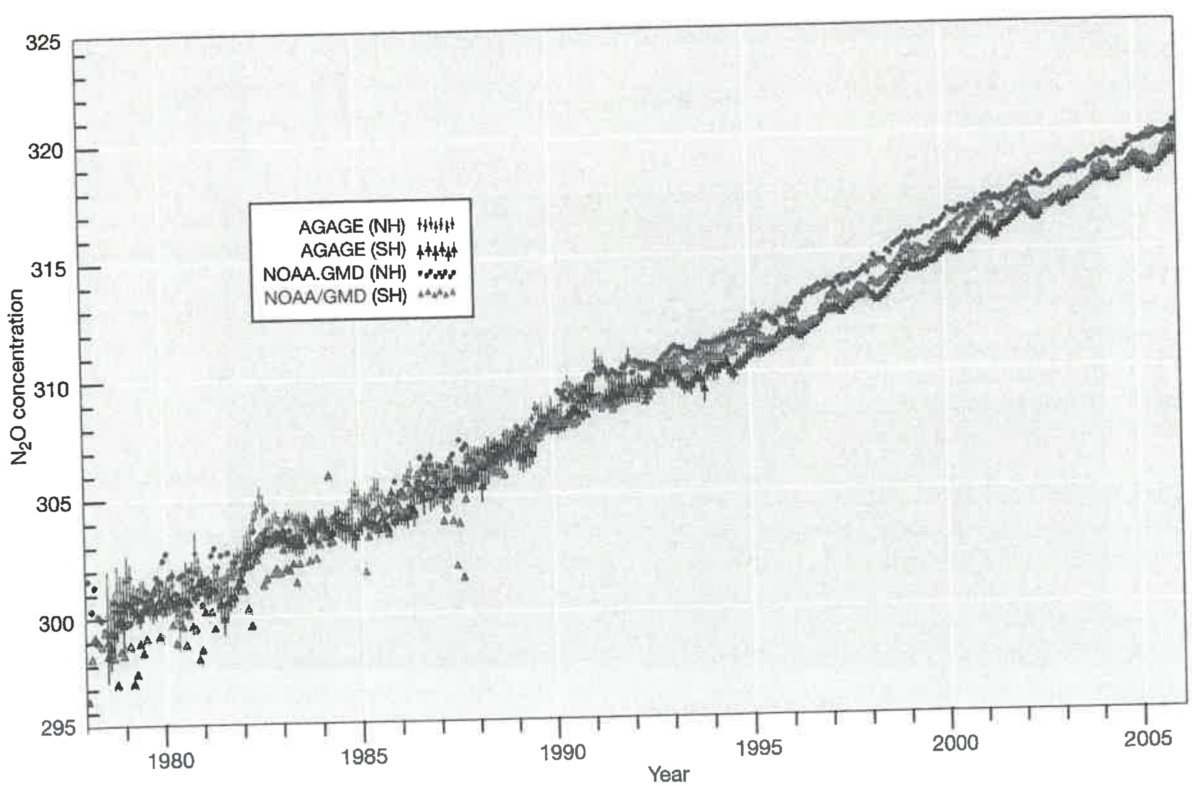


FIGURE 15-10 Measurements of atmospheric N_2O made since 1978. Slightly different values are reported for the Northern Hemisphere (NH) and Southern Hemisphere (SH).

a quantity termed **radiative forcing**. Radiative forcing refers to the change in the outgoing infrared flux caused by a change in the concentration of a particular greenhouse gas. A doubling of atmospheric CO_2 levels produces a radiative forcing of about 4.4 W/m^2. We have already seen that such a forcing produces a surface temperature increase of about 2.5°C in radiative-convective climate models. In three-dimension climate models (discussed below), the predicted change is 2–4.5°C.

The top panel in Figure 15-11a shows the radiative forcing produced by the increases in different greenhouse gases between 1750 and 2005. As one can see,

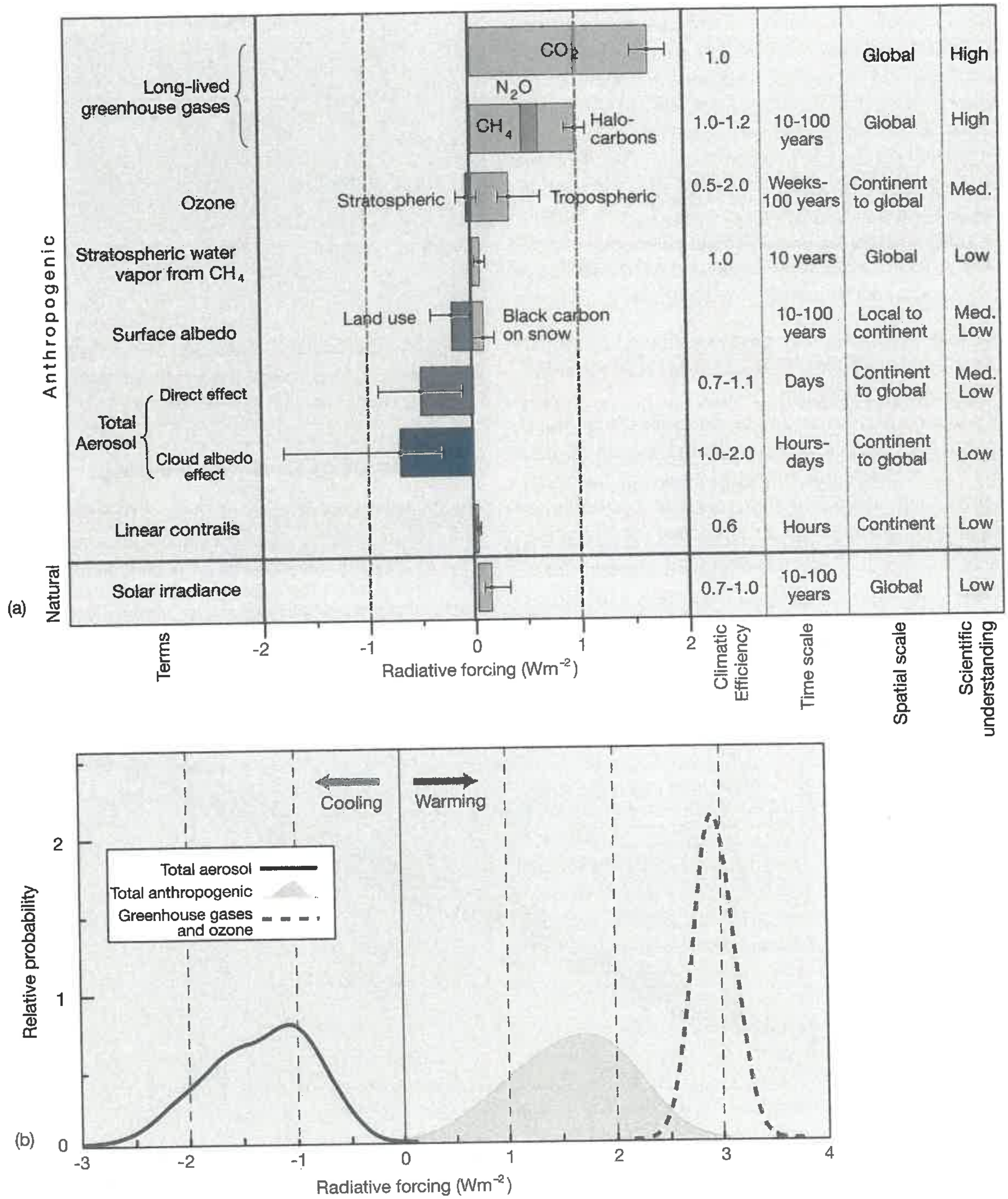


FIGURE 15-11 (a) Net radiative forcing by different atmospheric trace gases and other factors. The confidence in these estimates decreases from top to bottom. (b) Net radiative warming from greenhouse gases (right-hand dashed curve), net cooling from aerosols (left-hand dashed curve), and combined radiative effect (filled curve). (*Source:* IPCC 2007, Chapter 2, Figure 2.20.)

the increases in radiative forcing are about 1.7 W/m^2 for CO_2, 0.5 W/m^2 for CH_4, 0.15 W/m^2 for N_2O, and about 0.4 W/m^2 for halocarbons (CFCs). Other factors contribute to radiative forcing as well (Figure 15-11a). The forcings from greenhouse gases such as CO_2, CH_4, and N_2O are positive. Other forcings, however, are negative. Increases in sulfur dioxide (SO_2) gas emissions, largely from coal-fired power plants, for example, lead to increased concentrations of sulfate aerosols. Such particles cool the surface by increasing Earth's albedo. Sulfate aerosols may also cool the surface by acting as **cloud condensation nuclei (CCN)**, thereby increasing the reflectivity of clouds (see bar labeled "cloud albedo effect" in Figure 15-11a). The different radiative forcing factors are listed in order of how well we understand them. Those near the top of the diagram (which includes the greenhouse gases) are relatively well understood whereas those near the bottom are poorly understood.

The combined effects of these different radiative forcings can be estimated by adding them together, as has been done in Figure 15-11b. The dashed curve on the right shows the radiative forcing caused by increases in greenhouse gases, including ozone. The width of the curve indicates the uncertainty in the estimate. The greenhouse gas curve is narrow because the forcing is well understood. The solid curve on the left part of the diagram represents the radiative forcing from sulfate aerosols (both direct and via clouds). This curve is broad because the uncertainties are much greater. The filled curve in between the two dashed curves shows the combined radiative forcing of greenhouse gases plus aerosols. Although the uncertainties are large, the most likely value is about +1.8 W/m^2, indicating that the net forcing is positive and should therefore cause warming.

One can better appreciate the climatic effects of aerosols by comparing the net radiative forcing, +1.8 W/m^2, with that from greenhouse gases alone, +2.9 W/m^2. Evidently, about one-third of the forcing caused by greenhouse gases has been offset by increases in aerosols. The most visible manifestation of this effect was the slowing of surface temperature increases between 1940 and 1970 (Figure 15-5). Unfortunately, this offsetting influence on climate is likely to be less important in the future, for two reasons. First, the aerosols have short lifetimes, typically a week or two, whereas some greenhouse gases—CO_2 in particular—have very long lifetimes. And, second, other nations are likely to want to reduce their SO_2 and hydrocarbon emissions in the future, just as the United States and Europe have done over the past 40 years. The motivation for this change was to reduce air pollution and acid rain. We should be encouraging developing countries to follow this same path. But, by doing so, we may unavoidably exacerbate the problem of global warming.

AOGCM Predictions of Global Warming

The IPCC has made estimates for how much atmospheric CH_4 and N_2O will change over the next century and for how future production of SO_2 will affect sulfate aerosol concentrations. These estimates (which are not shown here) are generally in accord with the assumptions made regarding emissions of CO_2. In the most pessimistic case (the high-emissions case in Figure 15-8), CH_4 increases by a factor of about 2 over the next century (from 1.7 to ~3.4 ppm), while N_2O increases by about 50% (from 310 to 450 ppb). As we have already seen (Figure 15-9), that assumption may be overly pessimistic. Similarly, although nitrous oxide concentrations are still going up (Figure 15-10), they can only increase by a modest amount because a significant fraction of the present N_2O sources are nonanthropogenic. These factors are taken into account in the IPCC estimates. Various assumptions are made regarding SO_2 emissions as well. In the low-emissions model, SO_2 emissions remain roughly constant for the next 40 years, then decline gradually until the end of the century, whereas in the other two models, SO_2 emissions increase at first and then slowly decline. In the two higher-emission cases, the growth in SO_2 emissions significantly reduces the climate warming expected by the middle of the next century.

To convert these estimates for future trace gas emissions into predicted temperature changes, it is necessary to use numerical climate models. However, as mentioned earlier, the most reliable climate model predictions do not come from simple, one-dimensional climate models, but rather from three-dimensional general circulation models, or GCMs. (See the Box "A Closer Look: Three-Dimensional General Circulation Models [GCMs].") GCMs can predict the geographical distribution of future climate change, along with changes in other important variables such as precipitation. Furthermore, GCMs that include the ocean as well as the atmosphere, so-called AOGCMs (atmosphere–ocean general circulation models) can simulate another important effect as well. As the global climate warms, the ocean is expected to heat up more slowly than the atmosphere. It then acts as a brake on how fast the atmosphere itself can warm. Technically, it exerts a **thermal drag** on the system. Thus, the *transient* (time-dependent) response of the atmosphere to greenhouse gas increases is smaller than the *equilibrium* response.

The various climate-modeling groups within the IPCC have taken the greenhouse gas scenarios described previously and used them as input for time-dependent climate models. Some of these models are true AOGCMs. AOGCMs are very time-consuming to run, however, so not all of the scenarios have been modeled by multiple AOGCMs. Some climate predictions have been generated with simpler climate models that are "tuned" to reproduce the behavior of a true AOGCM. All such models include the thermal drag of the ocean, which reduces the rate at

A CLOSER LOOK

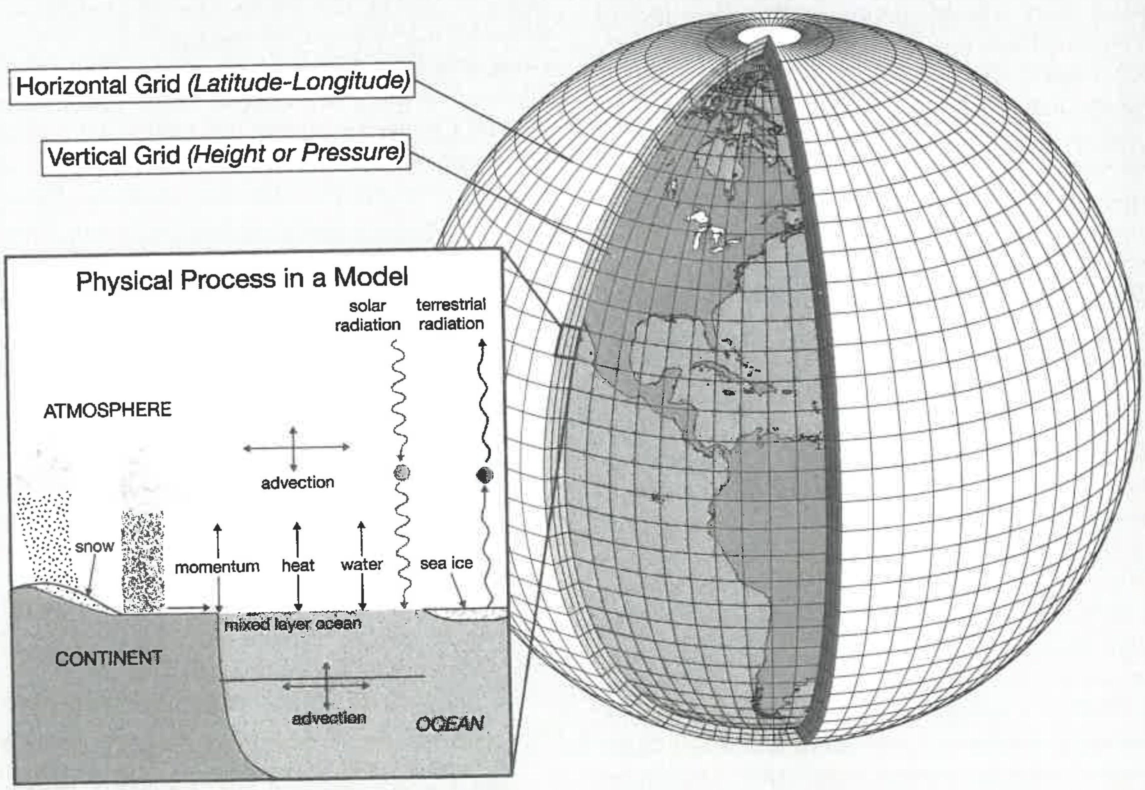
Three-Dimensional General Circulation Models (GCMs)

The most accurate and detailed predictions of future climate are made with three-dimensional general circulation models, or GCMs. These models were originally developed for weather forecasting. They can, however, be used for longer-term projections of climate. Fortuitously, the acronym GCM also stands for "global climate model," so we shall use these terms interchangeably.

In a GCM, the surface of the globe is divided into a two-dimensional array of longitude–latitude cells, as shown in Box Figure 15-2. The atmosphere above each cell is also divided into discrete layers, so that the model itself is comprised of a three-dimensional grid of boxes. Using these models we can now do the same sorts of calculations as described in Chapter 3, but each box can also exchange energy with the boxes above and below, and on all four sides (east–west and north–south). Not only can we include the effects of different latitudes, but we can also take geography into account. Whether, for example, we are over land or water, mountains or plains, deserts or tropical forests has an effect on the local climate. Putting all of these together allows for a much more realistic simulation of the global climate and how it is distributed.

To understand how such a model works, imagine that solar radiation enters a grid box at the top of the atmosphere. The model calculates how much radiation is reflected back to space by air molecules, how much is absorbed within the cell, and how much is transmitted down to the next level. The same calculations are made as the radiation is transmitted downward through each level until it reaches the surface. There, radiation is either reflected back upward or absorbed. The absorbed radiation heats the surface, which then emits longer wavelength infrared radiation upward as a function of its temperature. This is an application of the Stefan–Boltzmann law from Chapter 3. Some of this emitted radiation is absorbed by clouds, water vapor, and other greenhouse gases, thereby raising the temperature of the atmospheric grid boxes. Each box, in turn, emits infrared radiation both upward and back down toward the surface. The downward radiation, of course, is what we term the "greenhouse effect."

The radiation absorbed at the surface also results in evaporation and the transfer of sensible and latent heat upward into the atmosphere. As the model develops differences in temperature, vertically or horizontally, this causes changes in air density that result in the movement of air between adjacent boxes on each side (horizontal winds) or vertically (uplift or convection, and subsidence). As the air moves between these boxes it carries with it



BOX FIGURE 15-2 Schematic diagram of a three-dimensional general circulation model (GCM). These models are also called global climate models. (*Source:* http://en.wikipedia.org/wiki/Global_climate_model.)

energy, mass (including water vapor and aerosols), and momentum. Depending on the temperature and humidity of the box, the water may stay as a gas or it may condense to form clouds. The clouds produce rain that falls through the atmosphere. It may evaporate as it falls, or it may reach the surface as precipitation, which can occur as rain or snow, depending on the temperature. The changing characteristics of the box (in terms of water vapor, aerosols, temperature, clouds) all change the radiative properties of the box, which further changes the transmission, absorption, and reflection of the solar and terrestrial radiation fluxes and the transfer of sensible and latent heat.

All of this sounds complicated enough, but present-day global climate models include many other details, as well. The variables seem endless. The surface cells can be land or water. If they are land cells they have an elevation, so the model takes into consideration the surface topography, as well as soil and vegetation cover. When rain falls on the surface, some evaporates, some infiltrates the soil, and some runs off over the surface. How much depends on the soil characteristics, the surface temperature, how much water is already present, and the surface slope. The soil is divided into layers, allowing for water and heat to be transferred up and down in the soil and for water to flow laterally between adjacent boxes below the surface. Each model grid cell has a characteristic surface cover that includes a predominant vegetation type.

The vegetation type determines how deeply roots penetrate the soil (which influences water infiltration rates and evapotranspiration), and each vegetation type has a characteristic stem and leaf structure (which also affects evapotranspiration). In addition, the type of vegetation helps determine the surface albedo and also the surface roughness. How "rough" the surface is affects the friction between the low-level wind and the surface, which influences air movement (turbulence) and affects the rate at which heat is exchanged between the surface and the atmosphere. If precipitation falls as snow, this not only changes the albedo, but may also smooth out the surface as the vegetation becomes covered by snow (thus changing heat fluxes). Each of these processes and interactions is described by an equation or series of equations that can be solved by the model.

Some 3-D models, termed atmosphere–ocean general circulation models, or AOGCMs, include ocean circulation models that are as detailed as the atmospheric component. The ocean is divided into layers. Energy, mass, and momentum are exchanged at the ocean surface. The wind drives ocean currents and mixes the surface layers, and the model tracks temperature and salinity changes that determine water density and drive the deep–ocean (thermohaline) circulation. Convergence and divergence result in downwelling and upwelling water that connects the deep ocean and the surface layers. Sea ice forms where the surface ocean temperature is at or below the freezing point. Again, within each cell, the ice is divided into layers, heat is transferred through the ice, and the ice thickens and thins through the seasons as the water and air temperatures change.

This still does not give a full description of all the processes and calculations that are included in a global climate model. Hopefully, however, this description does illustrate the point that these are very sophisticated models that take into account all of the processes described in Chapters 3, 4, and 5, plus others that go beyond the level of detail we have discussed in the text.

There is one somewhat more technical wrinkle that deserves to be discussed here. In the past, these models were constructed essentially the way we described—by dividing the globe into boxes and calculating all of the transfers across the adjacent cell boundaries. All of the model equations were expressed as finite difference equations; that is, the transfers of energy, mass, and momentum between boxes were calculated as some function of the difference in various quantities (e.g., temperature) between the boxes. These models are, therefore, referred to as finite difference models. Because the atmosphere is unbounded at the sides (i.e., it is continuous around the globe), however, it is possible to express all of these equations very differently, as wave functions. In a model that uses wave functions, the number of waves resolved determines the spatial resolution of the model—the equivalent to the grid size in the finite difference models. These models are referred to as spectral models. Such models have certain numerical, or computational, advantages, compared to finite difference models; hence, most current GCMs are spectral models. Their output, however, is still usually presented as a gridded product.

which the global climate can warm. The results of these simulations for the three cases described earlier are shown in Figure 15-12. Somewhat surprisingly, the results are remarkably similar to the estimates given earlier, which were based on simple one-dimensional climate model calculations. For the optimistic, low-emissions case, the climate warms by about 2°C relative to the year 2000, or about 2.7°C overall. (Remember that the surface temperature in 2000 was already 0.7°C warmer than its 1850 value.) For the pessimistic high-emissions case, the predicted warming from 2000 to 2100 is about 3.8°C. Also shown is a case in which both greenhouse gases and aerosols were held constant at 2000 levels. The climate still warms by about 0.6°C over the next century because of the thermal inertia of the oceans, which are still catching up with the atmosphere. The various curves represent the average surface temperatures calculated by a number of different climate models. Not all of the models agree with each other. The range of uncertainty for the amount of warming produced by the year 2100 is shown by the error bars in Figure 15-12. The total range of uncertainty for all climate models and all greenhouse gas scenarios is shown by the shaded region of the figure. When all the uncertainties are taken into account, the predicted range of climate warming by the year 2100 (relative to 2000) is 1.4–4.0°C.

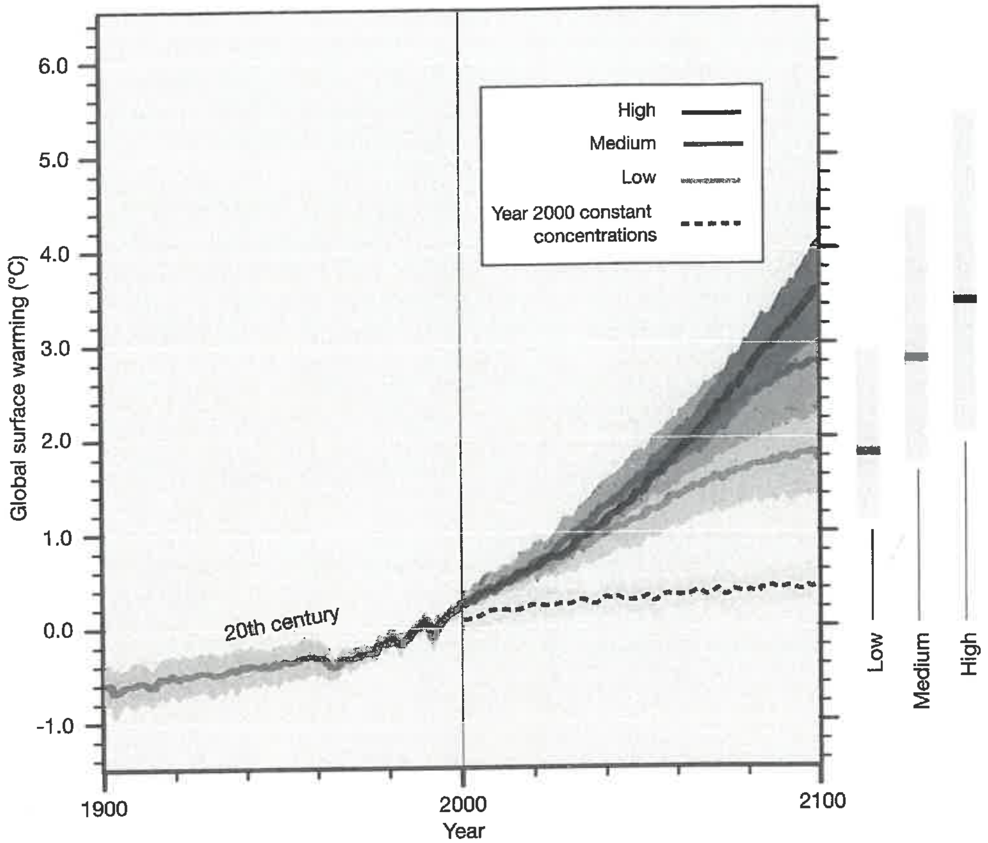


FIGURE 15-12 Predicted trends in global surface temperature over the next century for the three cases shown in Figure 15-8. The shaded area shows the total range of temperatures predicted by different climate models for the various scenarios. The bottommost curve shows the temperature change predicted when greenhouse gases and other forcings are held constant at 2000 values. (*Source:* IPCC 2007, Summary for Policy Makers, Fig. SPM.5.)

Long-Term Climate Warming

The climate calculations discussed so far extend only to the year 2100. But, as described in the Box "A Closer Look: Long-Term CO_2 Projections," atmospheric CO_2 could continue to rise for several centuries beyond that time and its concentration might eventually reach 2100 ppm—almost eight times its preindustrial value. What effect would this extended rise have on Earth's climate?

The greenhouse effect of CO_2 is roughly logarithmic, which means that each factor-of-2 increase in CO_2 produces roughly the same amount of warming. An eightfold increase in CO_2 should therefore cause about three times as much warming as would a twofold increase (because $8 = 2^3$.) Thus, a GCM that predicted 3°C of warming for doubled-CO_2 conditions should produce about 9°C of warming for an eightfold CO_2 increase.

The actual range of GCM responses to doubled atmospheric CO_2 is between 2.0 and 4.5°C of warming. The corresponding range for an eightfold CO_2 increase is 6.0 to 13.5°C of warming. It is instructive to compare these numbers to estimated surface temperature changes during the past 100 million years of Earth history. The warmest part of the Mesozoic is thought to have been about 6 to 10°C warmer than today on a global average. The coldest part of the Pleistocene was probably about 10°C cooler than today. Thus, regardless of whether we believe the low climate model estimates or the high ones, the warming from an eightfold CO_2 increase could make Earth warmer than it has been for tens of millions of years.

Possible Changes in the Thermohaline Circulation

Finally, we should note that future changes in climate could also cause changes in ocean circulation. One particular aspect of ocean circulation that has been studied extensively with regard to global warming is the Atlantic Conveyor Belt. This is the thermohaline circulation pattern described in Chapter 5 in which deep water forms in the North Atlantic, spreads out globally at depth, then is upwelled and returns to the North Atlantic as a warm surface current. Recall that this circulation pattern is responsible for keeping western Europe warm in the wintertime. Without it, winters there would be much colder. In the previous chapter, we saw that this circulation pattern probably did cease for almost a thousand years during the Younger Dryas event at the end of the last Ice Age. The cause of that shutdown is thought to have been a pulse of freshwater from the melting of the Laurentide ice sheet that flowed down the

A CLOSER LOOK

Long-Term CO_2 Projections

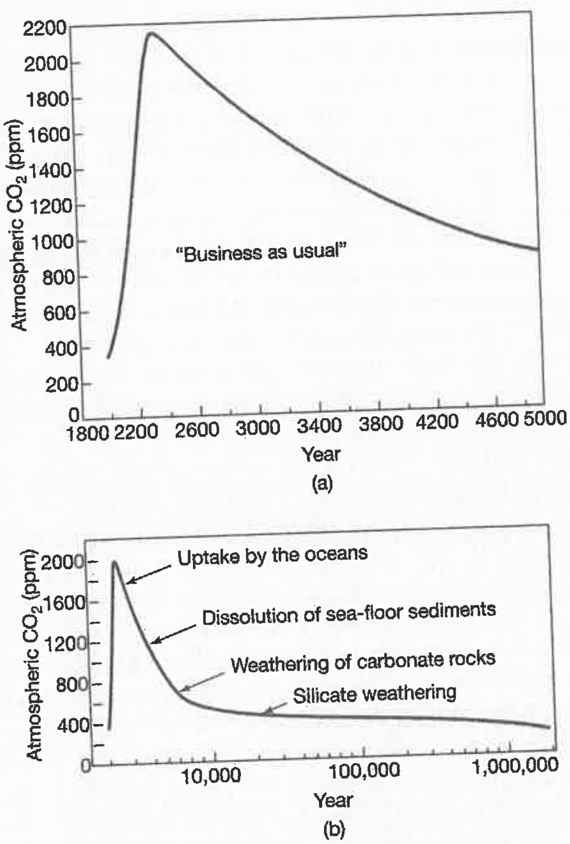
What will happen to atmospheric CO_2 levels in the distant future if we continue to burn fossil fuels? This question can be studied by using specially designed computer models that are able to take large time steps. For illustrative purposes, let us assume that most of the fossil-fuel reserve listed in Table 15-1, 4200 Gton(C), is consumed during the next 400 years. Let us further assume that current deforestation trends continue until only 30% of the world's forests remain.

This particular scenario has been investigated with a computer model that includes a six-box ocean, along with separate boxes to represent the atmosphere, forests, and carbonate sediments on the seafloor. The results of the simulation are shown in Box Figure 15-3. Box Figure 15-3a shows calculated atmospheric CO_2 concentrations over the next 3,000 years. The computer model predicts that atmospheric CO_2 levels will increase to a peak of about 2100 ppm by the year A.D. 2300. At that time, the fossil-fuel reserves will be exhausted and atmospheric CO_2 will begin to decline.

If we extend this calculation further into the future, the atmospheric CO_2 concentration must eventually return to its preindustrial value of 280 ppm, because the model assumes that the carbon cycle was balanced at that time. The amount of time required to return to that steady-state level is shown in Box Figure 15-3b. There, the date is displayed on a logarithmic scale extending millions of years into the future. The processes that are responsible for CO_2 uptake on different time scales are indicated. Most of the CO_2 is removed over the next few thousand years by dissolution in the deep ocean and by the dissolution of carbonate sediments and rocks. The last vestiges of the CO_2 pulse are removed by silicate weathering over a period of more than 1 million years.

Although the calculations shown here are speculative, they suggest an intriguing possibility. If the climate record of the past few million years were extended into the future, we would expect that Earth should experience at least 10 major glacial–interglacial cycles over the next 1 million years. But atmospheric CO_2 levels were relatively low, 200 to 280 ppm, during the previous glacial cycle, whereas they are predicted to exceed 350 ppm during most of the next 1 million years. Could the addition of this much anthropogenic CO_2 break the glacial–interglacial cycle? If it did, the results would be at least partly beneficial; after all, no one looks forward to the beginning of the next Ice Age. Yet, the accompanying increase in sea level over this time could cause shorelines to move substantially inland and force massive relocations of people. The possible long-term effects of the burning of fossil fuels on the Earth system are evidently quite large and will eventually need to be considered.

In 2005, David Archer of the University of Chicago published similar long-term simulations of the effects of fossil-fuel burning. The results of his simulations for total burns of 1000 Gton(C) and 5000 Gton(C) can be found in the 2007 IPCC report (Chapter 7, Figure 7-12). The peak CO_2 concentration reached in the 5000 Gton(C) simulation is 1750 ppm, about 10% lower than shown here, probably because of a more realistic treatment of seafloor carbonate dissolution. The time scale for recovery, however, is quite similar to that in the Walker and Kasting model shown here, and so the basic conclusions are the same: We probably *do not* want to do this!



BOX FIGURE 15-3 Long-term projections for atmospheric CO_2 for (a) the next 3,000 years and (b) the next several million years. The total amount of fossil fuel consumed is equivalent to 4200 Gton(C). (*Source:* Walker and Kasting, *Global and Planetary Change* 97, 1992, p. 151.)

St. Lawrence River and into the North Atlantic. This created a *freshwater cap* that was too buoyant to sink easily. Recall that Europe grew much colder again during this time, leading to the reappearance of alpine flowers like the dryas.

In at least some AOGCMs, a similar phenomenon is predicted to occur in the future, although to a lesser extent and for a somewhat different reason. Melting of the Greenland ice cap could, of course, supply freshwater that might duplicate the effect of the meltwater from the Laurentide ice sheet. That effect, if it occurred, would likely be several centuries in the future. In some AOGCMs, however, another similar phenomenon takes place much earlier, within the next 100–200 years. Global warming causes increased evaporation in the low-latitude to midlatitude Atlantic Ocean, putting more moisture into the air. This moisture is transported northward by winds and falls out as rain in the North Atlantic. The increased rainfall freshens the surface layer, reducing its density, and making it less likely to sink. In at least some models this causes the thermohaline circulation to slow appreciably within the next century. This is not a robust prediction at this time, as this type of couple ocean–atmosphere calculation is still considered to be difficult. If this were to occur, however, something really paradoxical might happen: Europe might cool temporarily as the rest of the world warmed! It is unlikely that Europe would experience a new Ice Age, as the global climate by this time would be significantly warmer than it was 11,000 years ago, but the predicted temperature drops could still be severe. This is yet one more example of why we must consider the entire Earth as a system. It takes a tightly coupled systems model, an AOGCM in this case, to predict this type of counterintuitive behavior.

Chapter Summary

1. Climate has remained remarkably stable over the last 10,000 years (the Holocene), although small changes have occurred. Temperatures were slightly higher during the Holocene Climatic Optimum, 5,000–6,000 years B.P., and again during the Medieval Warm Period, A.D. 1150–1300. They then decreased until the early 1500s. The Little Ice Age, A.D. 1600–1850, was unusually cold. Since then, surface temperatures have been increasing.
   a. One of the contributing causes of the colder temperatures during the Little Ice Age may be the effect of a higher frequency of volcanic eruptions.
   b. Decreased solar activity (as evidenced by decreases in sunspots) has also been linked to the Little Ice Age.
2. Volcanoes cause short-term climatic cooling.
   a. The sulfur dioxide injected into the stratosphere by an eruption hydrolyzes, forming sulfuric acid droplets, which both reflect solar radiation and absorb some of the long-wave radiation emitted from the troposphere.
   b. The result is lower-than-normal surface air temperatures for 1 to 2 years after a major eruption.
3. The observed rise in atmospheric CO_2 concentrations is caused chiefly by the burning of fossil fuels and, to a lesser extent, by tropical deforestation. The anthropogenic CO_2 source is much smaller than the rate at which CO_2 is released by respiration and decay but much larger than the rate at which CO_2 is emitted by volcanos. It therefore represents a substantial perturbation to the global carbon cycle.
4. The CO_2 generated by human activities can be removed from the atmosphere by several mechanisms.
   a. The fastest of these mechanisms is photosynthesis, but this sink will be effective only if forests are replanted or if CO_2 fertilization of plant growth continues to cause additional carbon to be stored in forests and soils.
   b. Much of the fossil-fuel CO_2 will dissolve in the oceans. The rate of CO_2 uptake is limited by the mixing rate of the deep ocean and by the chemical buffering capacity of seawater. Only about 30 to 40% of the available carbon in fossil fuels can be absorbed in this manner.
   c. Additional CO_2 can be removed by the dissolution of carbonate sediments on the seafloor and of carbonate rocks on land, but these processes occur over hundreds to thousands of years.
   d. The fossil-fuel CO_2 pulse would be completely removed by silicate weathering on a time scale of about 1 million years.
5. Computer models of the global carbon cycle predict that atmospheric CO_2 levels will double within the next 50 to 100 years and that CO_2 concentrations could exceed 2000 ppm within a few centuries if nothing is done to limit emissions. Radiative forcing of climate could be accelerated by increases in other trace greenhouse gases, although the increase in CH_4 appears to have slowed or stopped over the past few years. Earth's climate could warm by several degrees Celsius over the next century and by as much as 10 to 15°C in the long term. The warming will be unevenly distributed, with the polar regions warming the most and the tropics warming the least. Potentially damaging consequences of such warming include the drying out of continental interiors, the spread of insect pests and tropical diseases, and substantial increases in sea level.
6. Oceanic circulation could also change as the climate warms. Freshening of the North Atlantic from increased rainfall and melting of Greenland ice may tend to slow down the thermohaline circulation, causing Europe to cool transiently even as the rest of the world warms. Detailed understanding of the coupling between the atmosphere and oceans is needed to know whether such predictions are robust.

Key Terms

bioturbation
buffer
cloud condensation nuclei (CCN)
CO_2 fertilization
dendrochronology
Holocene epoch
Holocene Climatic Optimum
Little Ice Age
Maunder Minimum
Medieval Warm Period
methane clathrate hydrate
nitrogen fertilization
palynology
plages
proxy data
radiative forcing
Spörer Minimum
sunspots
thermal drag
Wolf Minimum

Review Questions

1. What is the Holocene epoch?
2. What are proxy data? Describe several examples of proxy climate data.
3. Briefly describe the Younger Dryas, the Holocene Climatic Optimum, the European Medieval Warm Period, and the Little Ice Age.
4. How do volcanoes affect climate?
5. What are sunspots? Why are they thought to have a possible effect on climate?
6. How does the amount of CO_2 produced by fossil-fuel consumption compare to the natural flux of CO_2 in the carbon cycle?
7. What are the major processes that can remove CO_2 from the atmosphere? What are the approximate time scales for these processes to be effective?
8. What is the size of the fossil-fuel reservoir compared with the atmospheric CO_2 reservoir?
9. Why does the ocean have a limited capacity for CO_2 uptake?
10. By how much is global temperature predicted to rise over the next century?
11. Why do some climate models predict that the thermohaline circulation might shut down?

Critical-Thinking Problems

1. a. The present atmosphere contains approximately 700 Gton(C) in the form of CO_2. Earth's total recoverable fossil-fuel reserves contain at least 4200 Gton(C), mostly in the form of coal. (We shall use the value 4200 Gton[C] to be specific.) At present, about half the CO_2 produced by the burning of fossil fuels stays in the atmosphere. The other half dissolves in the oceans or is taken up by the terrestrial biosphere. If this ratio remained constant and we burned up all of our fossil fuels instantaneously, by how much would atmospheric CO_2 concentrations rise? (Express your answer in terms of the new CO_2 level divided by the old one.)
   b. Climate models predict that each doubling of the atmospheric CO_2 concentration will cause the mean global temperature to increase by 1.5 to 4.5°C. (The range is due largely to uncertainties about how clouds will respond.) By how much would the mean temperature increase for the scenario described in part a? Express your answer as a temperature range in degrees Celsius and in degrees Fahrenheit.
   c. The actual problem of global warming could be more severe than we have just calculated. Forests and soils together contain an additional 2100 Gton of carbon that might go into the atmosphere if deforestation is not prevented. The ocean becomes more acidic as it absorbs CO_2, so it might not be able to continue taking up as much CO_2 as it has been until now. If we burned up all our fossil fuels and deforested one-third of the globe without losing any CO_2 to the ocean (or to CO_2 fertilization), by how much would atmospheric CO_2 and temperature increase?
2. The atmospheric CO_2 concentration is currently increasing by about 1.9 ppm/yr. How many gigatons of carbon are being added to the atmosphere each year? (*Hint:* The total mass of the atmosphere is 5×10^{18} kg, and its mean molecular weight is about 29. You will need to do the calculation in moles and then convert back to mass units.)
3. The surface ocean contains about 2.6×10^{16} liters of water with a carbonate ion content of about 2×10^{-4} mol/L. The deep ocean contains about 1.4×10^{21} L of water with a carbonate ion content of roughly 9×10^{-5} mol/L. If each mole of carbonate reacts with 1 mol of CO_2 according to the reaction

$$CO_2 + CO_3^{=} + H_2O \rightarrow 2\,HCO_3^{-},$$

what percentage of the fossil-fuel reservoir, 4200 Gton(C), can be neutralized by the surface ocean? By the deep ocean?

Further Reading

General

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