Third Edition

ECOLOGICAL CLIMATOLOGY

Concepts and Applications

Gordon Bonan

Ecological Climatology

The third edition of Gordon Bonan's comprehensive textbook introduces an interdisciplinary framework to understand the interaction between terrestrial ecosystems and climate change. Ideal for advanced undergraduate and graduate students studying ecology, environmental science, atmospheric science, and geography, it reviews basic meteorological, hydrological, and ecological concepts to examine the physical, chemical, and biological processes by which terrestrial ecosystems affect and are affected by climate. This new edition has been thoroughly updated with new science and references. The scope has been expanded beyond its initial focus on energy, water, and carbon to include reactive gases and aerosols in the atmosphere. This new edition emphasizes Earth as a system, recognizing interconnections among the planet's physical, chemical, biological, and socioeconomic components, and emphasizing global environmental sustainability. Each chapter contains chapter summaries and review questions, and with over four hundred illustrations, including many in color, this textbook will once again be an essential student guide.

Gordon Bonan is senior scientist and head of the Terrestrial Sciences Section at the National Center for Atmospheric Research, Boulder, Colorado. His research focuses on the interactions of terrestrial ecosystems with climate using models of Earth's biosphere, atmosphere, hydrosphere, and geosphere. He has published more than 120 peer-reviewed articles on land-atmosphere coupling and how changes in vegetation alter climate. He is a member of the American Geophysical Union, American Meteorological Society, and Ecological Society of America. He is a Fellow of the American Geophysical Union and has served on advisory boards for numerous national and international organizations and as an editor for several journals.

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Third edition

Gordon Bonan
*National Center for Atmospheric Research**
Boulder, Colorado

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To Amie, who made this possible
To David, Thomas, and Alice, for family
To Milo, Dancer, and Chloe, for hugs and head pats

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Preface

I began conceiving this book in 1996. At that time, the influence of forests on climate was well-established at the microscale through the study of forest meteorology and biometeorology; that the terrestrial biosphere is essential for climate science and global models of climate was less universally accepted. The first edition was published in 2002. It was an effort to broaden the scope of ecology – to show ecologists the manner in which ecosystems influence climate – and to similarly broaden climate science to recognize the importance of terrestrial ecosystems. The second edition published in 2008 was a marked change from the first edition. It was a complete revision that reflected the expanded scope of the science, improved the organization of the material, and made it more accessible to students.

The third edition is yet another revision and update of the book. The intent has not changed, but the science has so vastly grown. Studies of biosphere–atmosphere interactions at the regional and global scale are now commonplace; all of the major international climate modeling centers include models of the terrestrial biosphere; the carbon cycle and anthropogenic land use and land-cover change are recognized as important facets of climate change; and climate science itself has evolved into a broader perspective of Earth system science. This is seen in the expanded breadth of the book. In the second edition, carbon cycle–climate coupling was still fairly novel. This third edition shows how important that has become to climate science, and additionally includes chapters on the nitrogen cycle, aerosols, and climate change mitigation. A concluding chapter ties together the various topics presented throughout the book. The influences of terrestrial ecosystems on climate must be seen in a larger context of human influences on the global environment and in light of planetary sustainability.

One prominent change over the years has been the extensive growth of the scientific literature. This third edition is not meant to be a survey of all relevant literature; that would be too tedious. Rather, I have highlighted key papers that, with online scholarly databases, provide a springboard to the science. To keep the book manageable, some material had to be deleted from the earlier editions. Many references to scientific studies have been removed or omitted. Nonetheless, this third edition provides a comprehensive survey of the state of the science. The challenge of organizing, synthesizing, and presenting the voluminous material in a comprehensible manner is tempered by the pleasure in seeing the extent to which the science has expanded over the years.

As in the previous editions, this book contains many mathematical equations, but only to illustrate concepts and not with the intent of being a modeling textbook. The book heavily references models, their scientific scope, and their application to understand biosphere–atmosphere interactions. The book also maintains land management, urban planning, and landscape design as a theme. The principles of ecological climatology are applicable to these studies. Unlike global change, land use occurs locally in our communities. It gives substance to environmental issues at spatial and temporal scales to which people can see and respond; we see these changes happen in our communities, often over a period of a few years.

As always, I am indebted to colleagues at the National Center for Atmospheric Research for supporting my efforts to write this third edition, in particular Sam Levis and Keith Oleson, whose long-standing commitment to the development and maintenance of community models, both since 1999, have allowed me to write this book. David Lawrence, too, assumed a leading role in community model development, allowing

me to focus my efforts on writing. And new colleagues – Rosie Fisher, Peter Lawrence, Will Wieder, Danica Lombardozzi, Quinn Thomas, Melannie Hartman, and Liz Burakowski – have similarly grown the science and supported community models. Finally, I am indebted to Matt Lloyd at Cambridge University Press, who has supported this endeavor over the many years.

1

Ecosystems and Climate

1.1 Chapter Summary

When viewed from space, Earth is seen as a blue marble. The dominant features of the planet are the blue of the oceans and the white of the clouds traversing the atmosphere. It is an image of fluids – water and air – in motion. Indeed, the study of Earth's climate is dominated by the geophysical principles of fluid dynamics. With closer inspection, however, one can discern land masses – the continents – and the plants that grow on the land. The blue of the oceans gives way to the emerald green of vegetation. Weather, climate, and atmospheric composition have long been known to determine the floristic composition of these plants, their arrangement into communities, and their functioning as ecosystems. Earth system scientists now recognize that the patterns and processes of plant communities and ecosystems not only respond to weather, climate, and atmospheric composition, but also feedback through a variety of physical, chemical, and biological processes to influence the atmosphere. The geoscientific understanding of planet Earth has given way to a new paradigm of biogeosciences. Ecological climatology is an interdisciplinary framework to study the functioning of terrestrial ecosystems in the Earth system through their cycling of energy, water, chemical elements, and trace gases. Changes in terrestrial ecosystems through natural vegetation dynamics and through human uses of land are a key determinate of Earth's climate.

1.2 Common Science

Ecology is the study of interactions of organisms among themselves and with their environment. It seeks to understand patterns in nature (e.g., the spatial and temporal distribution of organisms) and the processes governing those patterns. Climatology is the study of the physical state of the atmosphere – its instantaneous state, or weather; its seasonal-to-interannual variability; its long-term average condition, or climate; and how climate changes over time. These two fields of scientific study are distinctly different. Ecology is a discipline within the biological sciences and has as its core the principle of natural selection. Climatology is a discipline within the geophysical sciences based on applied physics and fluid dynamics. Both, however, share a common history.

The origin of these sciences is attributed to the Greek scholars Aristotle (ca. 350 BCE) and Theophrastus (ca. 300 BCE) and their books *Meteorologica* and *Enquiry into Plants*, respectively, but their modern beginnings trace back to natural history and plant geography. Naturalists and geographers of the seventeenth, eighteenth, and nineteenth centuries saw changes in vegetation as they explored new regions and laid the foundation for the development of ecology

Table 1.1 Relationship between de Candolle's plant types and Köppen's climate types

de Candolle plant type	Köppen climate type	Dominant vegetation
Megatherms	Humid tropical	Tropical rainforest
		Tropical savanna
Xerophiles	Dry	Desert
		Grassland
Mesotherms	Moist subtropical mid-latitude	Warm temperate deciduous forest
		Warm temperate coniferous forest
		Mediterranean
Microtherms	Moist continental	Cool temperate deciduous forest
		Cool temperate coniferous forest
		Boreal forest
Hekistotherms	Polar	Tundra

Source: Adapted from Colinvaux (1986, p. 326) and Oliver (1996).

and climatology as they sought explanations for these geographic patterns. Alexander von Humboldt, in the early 1800s, observed that widely separated regions have structurally and functionally similar vegetation if their climates are similar. Alphonse de Candolle hypothesized that temperature creates latitudinal zones of tropical, temperate, and arctic vegetation and in 1874 proposed formal vegetation zones with associated temperature limits. This provided an objective basis to map climatic regions, and in 1884 Wladimir Köppen used maps of vegetation geography to produce climate maps. His five primary climate zones shared similar temperature delimitations as de Candolle's vegetation (Table 1.1). The close correspondence between climate and vegetation is readily apparent, and many secondary climate zones such as tropical savanna, tropical rainforest, and tundra are named after vegetation. Although vegetation is no longer used to map the present climate, it is a primary means to reconstruct past climate from relationships of temperature and precipitation with tree-ring width, pollen abundance, and leaf form.

Despite shared origins, twentieth-century advancement of ecology and climatology proceeded not as an integrated and unified science, but rather in the typical disciplinary framework of science into specialized fields of study that favored reductionism. Plant ecology splintered into topical studies of physiology, populations, communities, ecosystems, landscapes, and biogeochemistry. The study of the atmosphere became organized around spatial scales of micrometeorology, mesoscale meteorology, and global climate and topical fields such as boundary layer meteorology, hydrometeorology, radiative transfer, atmospheric dynamics, and atmospheric chemistry.

With lack of communication across disciplines, ecologists and climatologists can draw different insights from the same observations. Pieter Bruegel the Elder's painting "Hunters in the Snow" exemplifies this (Figure 1.1). The painting has been used in climatology textbooks to illustrate climate change (Lamb 1977, pp. 275–276; Lamb 1995, pp. 233–235). Bruegel painted this scene in the winter of 1565 and it depicts, from a climatologist's perspective, Bruegel's impression of the severe winters of that era. It was the beginning of prolonged artistic interest in Dutch winter landscapes that coincided with an extended period of colder than usual European winters. Ecologists have similarly used this painting to illustrate the ecological concept of a landscape (Forman and Godron 1986, pp. 5–6). Instead of a visual record of an unusually cold climate, the ecological perspective perceives an expression of the core tenets of landscape ecology. From an ecological point of view, the painting depicts heterogeneity of

Fig. 1.1 "Hunters in the Snow" (Pieter Bruegel the Elder). Reproduced with permission of the Kunsthistorisches Museum (Vienna).

landscape elements, spatial scale, and movement across the landscape.

Earth, too, has long been viewed differently by ecologists and climatologists. Ecologists have historically seen weather, climate, and atmospheric composition as external forcings that shape plant communities and ecosystem functions. The manner in which ecosystems influence weather, climate, and atmospheric composition was not examined in classic ecology textbooks. Similarly, climatology textbooks emphasized the physics and fluid dynamics of the atmosphere, not the vegetation at the lowest boundary of the atmosphere.

The advent of global models of Earth's climate in the 1970s and 1980s altered the disciplinary study of ecology and climatology. These models require a mathematical representation of the exchanges of energy, water, and momentum between land and atmosphere. These processes are regulated in part by plants, which with their leaves, stomata, and diversity of life do not conform to the mathematics of fluid dynamics. Atmospheric scientists developing climate models had to expand their geophysical framework to a biogeophysical framework (Deardorff 1978; Dickinson et al. 1986; Sellers et al. 1986). The ongoing evolution of climate models to models of the Earth system is marked by recognition of the central role of terrestrial ecosystems in regulating Earth's climate through physical, chemical, and biological processes and the critical influence that human appropriation of ecosystem functions has on climate (Pitman 2003; Bonan 2008; Arneth et al. 2010, 2014; Levis 2010; Seneviratne et al. 2010; Mahowald et al. 2011). Models of Earth's land surface, including its terrestrial ecosystems, for climate simulation have expanded beyond their hydrometeorological heritage (with emphasis on surface energy fluxes and the hydrologic cycle) to include carbon, reactive nitrogen, aerosols, anthropogenic land use, and vegetation dynamics. These models are important research tools to study land–atmosphere interactions and climate feedback from ecological processes.

1.3 Deforestation and Climate – Some Early Views

The notion that vegetation affects climate is not new. Over two thousand years ago, Theophrastus

Fig. 1.2 Sketches of land clearing in western New York (clockwise from top left). In the first panel, the initial forest clearing is small, only to fell trees for the small log house and to raise a few livestock. The second panel shows the settler has cleared a few acres of land. Ten years have passed in the third panel. Thirty to forty acres of land have been cleared, and neighbors have cleared their land. The final panel depicts 45 years following the initial clearing. From Turner (1849, pp. 562–566).

wrote that draining marshes removed the moderating effect of water and created a colder climate, while deforestation exposed the ground to the Sun and warmed climate (Glacken 1967, p. 130; Neumann 1985). The concept that forests increase rainfall can be traced back to the Roman natural philosopher Pliny the Elder and his *Natural History*, written in the first century AD (Andréassian 2004). European naturalists in the seventeenth and eighteen centuries, too, believed that the wide-ranging clearing of forests and cultivation of land in Europe had moderated the climate since antiquity (Fleming 1998). Settlers of the New World carried with them a similar sentiment, and a vigorous debate arose about whether the extensive land clearing (Figure 1.2) was indeed changing the climate of America (Thompson 1980; Fleming 1998). Like debates arose in Australia, where much of the native forest and woodland was cleared following British settlement, and similarly with British colonization of India.

One question concerned whether deforestation and cultivation of land created milder winters. A popular view, espoused by the Scottish philosopher David Hume, was that deforestation opened the land to heating by the Sun during winter. In an essay (ca. 1750) he explained that warmer winters occurred because "the land is at present much better cultivated, and that the woods are cleared, which formerly threw a shade upon the earth, and kept the rays of the sun from penetrating to it" (Hume and Miller 1987, p. 451). Because of this, Hume declared that "our northern colonies in America become more temperate, in proportion as the woods are felled." Similar views are seen in the writings of the New England Puritan minister Cotton Mather, who observed that "our cold is much moderated since the opening and clearing of our woods" (Mather 1721, p. 74). Benjamin Franklin, too, believed that climate was warming because "when a country is clear'd of woods, the sun acts more strongly on the face of the earth" (Franklin and Labaree 1966). Franklin's contemporary Hugh Williamson, a physician, scholar, and politician, predicted that with continued clearing of interior lands "we shall seldom be visited by frosts or snows, but may enjoy such a temperature in the midst of winter, as shall hardly destroy the most tender plants" (Williamson 1771). Thomas Jefferson agreed that "a change in our climate … is taking place very sensibly" (Jefferson 1788, p. 88) and urged climate surveys "to show the effect of clearing and culture towards changes of climate" (Jefferson and Bergh 1905). Samuel Williams, a Congregational minister, professor at Harvard, and founder of the University of Vermont, believed that as trees were cut down and settlements increased "the cold decreases, the earth and air become more warm; and the whole temperature of the climate, becomes more equal, uniform and moderate" (Williams 1794, p. 57). This climate change "is so rapid and constant, that it is the subject of common observation and experience."

Not all agreed with this sentiment. In addition to being a lexicographer, Noah Webster of Connecticut was a political writer. He strongly refuted the notion of such changes in climate in an essay published in 1799. Direct observations of climate change were lacking, he noted, and evidence of a warmer climate relied on anecdotes and personal memories. Yet, he, too, admitted to differences between forests and cleared land that altered local climate (Webster 1843, pp. 145). "While a country is covered with trees," he wrote, "the face of the earth is never swept by violent winds; the temperature of the air is more uniform, than in an open country; the earth is never frozen in winter, nor scorched with heat in summer."

The writings of Europeans attested to the prominence of the forest–climate debate, and also to the difference in opinions. Constantin-François Volney published a book in 1803 based on his travels in eastern North America. He observed that "for some years it has been a general remark in the United States, that very perceptible partial changes in the climate took place, which displayed themselves in proportion as the land was cleared" (Volney 1804, p. 266). Alexander von Humboldt responded in 1807 that "the statements so frequently advanced, although unsupported by measurements, that since the first European settlements in New England, Pennsylvania, and Virginia, the destruction of many forests … has rendered the climate more equable, – making the winters milder and the summers cooler, – are now generally discredited" (Humboldt et al. 1850, p. 103). The *Edinburgh Encyclopaedia* asserted that the theory of land-use climate change "is, we fear, to be regarded rather as the birth of a lively fancy, than the offspring of accurate science" (Brewster 1830, pp. 613–614).

Another belief was that forests contributed to the plentiful rainfall in America and that deforestation decreased rainfall. Christopher Columbus developed such a view from his travels to the New World. His son wrote in a biography that he attributed the rainstorms of Jamaica to "the great forests of that land; he knew from experience that formerly this also occurred in the Canary, Madeira, and Azore Islands, but since the removal of forests that once covered those islands, they do not have so much mist and rain as before" (Colón and Keen 1959, pp. 142–143).

Natural scientists developed similar views. John Evelyn declared in 1664 that forests "render those countries and places more subject to rain and mists" (Evelyn 1801, pp. 29–32). John Clayton, an English naturalist and clergyman, attributed the violent thunderstorms of coastal Virginia to the dense forests (Clayton 1693; Berkeley and Berkeley 1965, pp. 48–49). His compatriot John Woodward described how the "great moisture in the air, was a mighty inconvenience and annoyance to those who first settled in America," but that after clearing the forests "the air mended and cleared up apace: changing into a temper much more dry and serene than before" (Woodward 1699). Samuel Williams accounted for the plentiful rainfall because "the immense forests … supply a larger quantity of water for the formation of clouds, than the more cultivated countries of Europe" (Williams 1794, p. 50). Hugh Williamson described a feedback by which forest evaporation enhances rainfall and cools climate (Williamson 1811, pp. 23–25): "The vapours that arise from forests, are soon converted into rain, and that rain becomes the subject of future evaporation, by which the earth is further cooled." In the Caribbean islands, which had undergone widespread forest clearing to grow sugar cane, forest preserves were created to promote rainfall (Anthes 1984).

Settlement of the Great Plains in the 1870s and 1880s shifted the debate from deforestation to afforestation, with the premise that tree planting would increase rainfall (Emmons 1971; Kutzleb 1971; Thompson 1980; Williams 1989). An official in the United States Department of Interior claimed that "the planting of ten or fifteen acres of forest trees on each quarter section [160 acres] will have a most important effect on the climate, equalizing and increasing the moisture" (United States General Land Office 1867, p. 135). That some official believed that "if one-third the surface of the great plains were covered with forest there is every reason to believe the climate would be greatly improved" (United States General Land Office 1868, p. 197). Congress agreed and enacted the Timber Culture Act of 1873 to promote afforestation. Popular science gazettes, too, advocated tree planting to increase rainfall (Oswald 1877; Anonymous 1879). Samuel Aughey, of the University of Nebraska, promoted the notion that plowing the prairie sod was the cause of an increase in rainfall observed at that time. Cultivation allowed the soil to retain more rainfall, which evaporated and rained back onto the land, he theorized (Aughey 1880, pp. 44–45). Charles Wilber popularized this notion with the phrase "rain follows the plow" (Wilber 1881, p. 68). He described how an "army of frontier farmers … could, acting in concert, turn over the prairie sod, and … present a new surface of green, growing crops instead of the dry, hard-baked earth covered with sparse buffalo grass. No one can question or doubt the inevitable effect of this cool condensing surface upon the moisture in the atmosphere."

A sharply divided debate on forest–climate influences continued in the latter half of the nineteenth century and into the twentieth century. Conservationists, botanists, and foresters argued for such influences. George Perkins Marsh devoted a large portion of his treatise *Man and Nature* to forest–climate influences (Marsh 1864). The botanist Richard Upton Piper agreed that "forests trees should be preserved for their beneficial influence upon the climate" (Piper 1855, p. 51). The fledgling forestry division of the United States Department of Agriculture issued reports supportive of the burgeoning field of forest meteorology and forest–rainfall influences (Hough 1878; Fernow 1902; Zon 1927).

Climatologists of the day, however, dismissed the study of forests and climate. A publication on the climate of the United States asserted that "the great differences of surface character which belong to the deserts, woodlands, and other more striking features, are believed to have their *origin* in climate, and not to be agents of causation themselves" (Blodget 1857, p. 482). The geographer Henry Gannett suggested that faulty reasoning was behind the belief that forests increase rainfall. His analysis of precipitation records found no change in rainfall in regions that had undergone increases and decreases in tree cover, and he complained that "a satisfactory explanation of this supposed phenomenon has never … been offered" (Gannett 1888). He further explained that "it may be that in this case an effect has been mistaken for a cause, or rather, since it is universally recognized that rainfall produces forests, the converse has been incorrectly assumed to be also true" (Anonymous 1888). The eminent meteorologist William Ferrel argued in favor of large-scale control of precipitation by atmospheric circulation, not by surface conditions (Ferrel 1889). His colleague Cleveland Abbe wrote that "rational climatology gives no basis for the much-talked-of influence upon the climate of a country produced by the growth or destruction of forests … and the cultivation of crops over a wide extent of prairie" (Abbe 1889). Abbe believed that "the idea that forests either increase or diminish the quantity of rain that falls from the clouds is not worthy to be entertained by rational, intelligent men" (Moore 1910, p. 7).

Foresters and climatologists in the United States Department of Agriculture were sharply divided over forest influences on rainfall. While the foresters issued reports in support of the science (Hough 1878; Fernow 1902; Zon 1927), their colleagues in the department's Weather Bureau (the predecessor of the present-day National Weather Service) resoundingly dismissed these ideas. Mark Harrington, chief of the Weather Bureau, rejected his forestry colleagues' belief that forests affected climate in any manner (Fernow 1902). Willis Moore, Harrington's successor, also rebuffed studies relating forests and climate with the retort that "while much has been written on this subject, but little of it has emanated from meteorologists" (Moore 1910, p. 3). "Precipitation," he explained, "controls forestation, but forestation has little or no effect upon precipitation" (Moore 1910, p. 37). The caustic rhetoric confused a writer in the journal *Nature*, who reported that "the literature on the subject is somewhat bewildering" (Anonymous 1912).

The views about ongoing climate change in colonial America ultimately proved to be false. Climate was not changing; winters were not becoming milder with land clearing; rainfall was not decreasing because of deforestation or increasing because of tree planting and soil cultivation. However, meteorologists of that era, too, were ultimately proved wrong. As they sought physical explanations for geographic variations in climate, they were too quick to dismiss the precept that forests, grasslands, croplands, and other ecosystems do indeed influence climate. Interest in the climatic effects of deforestation, cultivation, and overgrazing reemerged in the 1970s, with recognition that human activities do indeed change climate and that land use is one such mechanism for climate change (Landsberg 1970; Otterman 1974, 1977; Schneider and Dickinson 1974; Sagan et al. 1979). One hundred years after a writer to *Nature* found the debate to be "belwidering," the journal published another paper that found that tropical forests do indeed increase rainfall (Spracklen et al. 2012).

1.4 Ecological Climatology

Scholars of the eighteenth and nineteenth centuries lacked the scientific tools to properly ascertain forest influences on climate, but scientists in the latter part of the twentieth century had a new tool – global climate models – with which to study how plants and ecosystems affect climate. Scientific interest over the past few decades in the coupling between climate and life has paralleled the trend by atmospheric scientists to recognize the planet as a system of interacting spheres.

Today, scientists identify four main components of the Earth system: atmosphere, air; hydrosphere, water; biosphere, living things; and geosphere, solid portion of Earth. The geosphere can be subdivided into other spheres. The lithosphere is the solid outer layer of Earth including the crust and upper mantle. Its outermost layer is called the pedosphere, or soil. Some scientists separately identify the cryosphere, or frozen portion of Earth. The influence of humans is so prevalent, especially after the Industrial Revolution, that a new sphere, the anthroposphere, has been proposed to describe that part of Earth modified by people for human activities or habitats. Earth's climate must be understood in terms of a system of interacting spheres (atmosphere, hydrosphere, biosphere, geosphere, and anthroposphere); the energy, water, and biogeochemical cycles that link these spheres; and the interactions with human systems that alter these cycles. This parallels a progression in atmospheric sciences from (Figure 1.3): atmospheric general circulation models, which considered atmospheric physics and dynamics; to atmosphere–ocean general circulation models, which included the coupling of the atmosphere with models of ocean and sea-ice physics and dynamics; to global climate models, which additionally accounted for hydrometeorological coupling with land; and now to Earth system models, which also include atmospheric chemistry, terrestrial and marine ecology, and biogeochemistry.

At the intersection of these spheres is an interdisciplinary field of study called biogeoscience that bridges the earth and life sciences (Figure 1.4). Biogeoscience is the study of the interactions between life and Earth's atmosphere, hydrosphere, and geosphere. It has long been synonymous with the study of

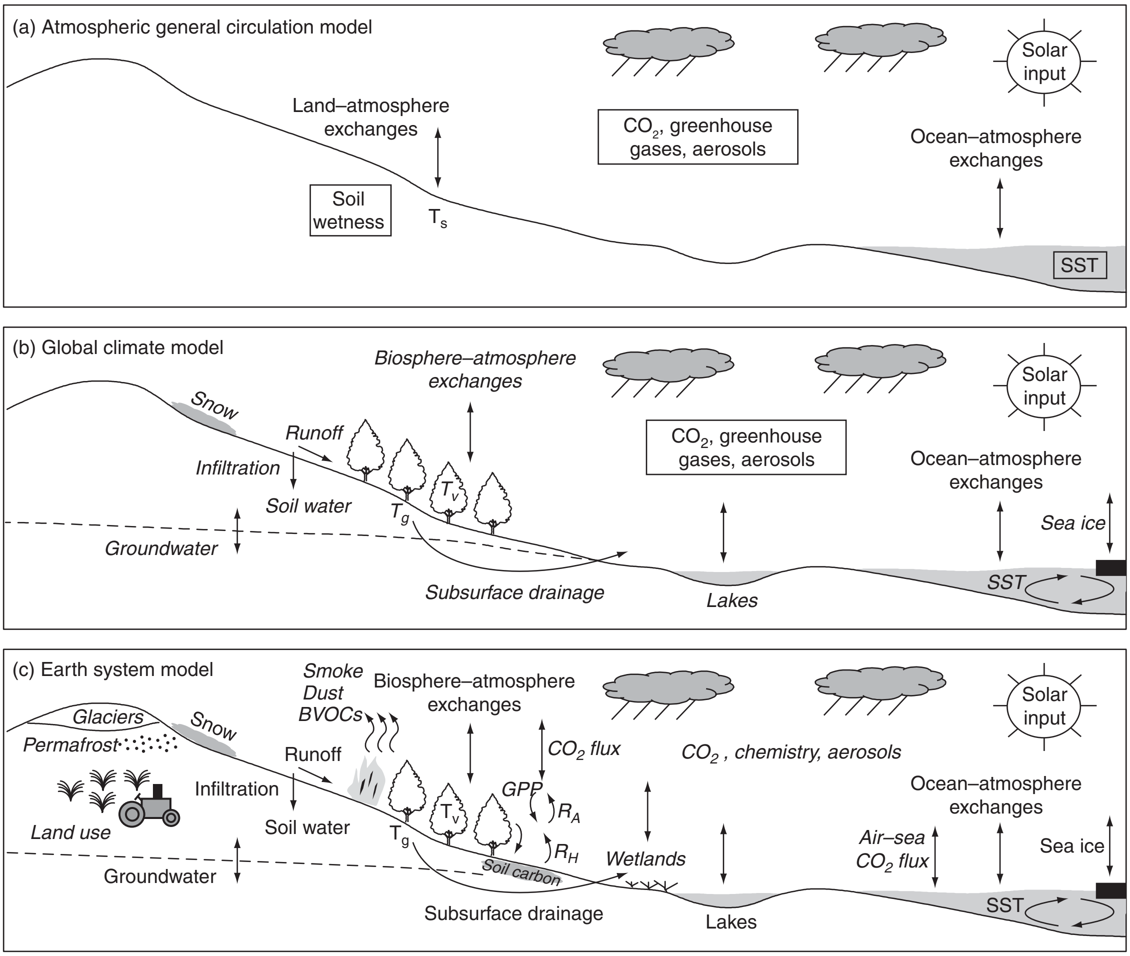

Fig. 1.3 Components of the Earth system, their processes, and interactions as represented in global models. New processes added to each model are highlighted in italics. The distinction among models is not precise, and the transition among models is in fact blurred. (a) Atmospheric general circulation models circa 1970s. These models used prescribed inputs of atmospheric CO_2, other greenhouse gases, and aerosols. They calculated land and ocean physical flux exchanges using prescribed soil wetness and sea surface temperature. (b) Global climate models circa 1990s. These models added the hydrologic cycle on land and plant canopies. They included ocean general circulation models to calculate sea surface temperature, sea ice, and ocean dynamics. (c) Earth system models circa 2010s. These models added the carbon cycle and other biogeochemical processes, anthropogenic land use, wetlands, glaciers and cryospheric processes, atmospheric chemistry, and aerosols.

biogeochemistry, but is broader and includes the interactions between living organisms and the physical environment and the manner in which organisms modify physical systems.

This book examines one element of that science – the commonality between ecological and atmospheric sciences that affect weather, climate, and atmospheric composition. The study of plants and terrestrial ecosystems, and human appropriation of ecosystem functions, is as essential to the study of Earth's climate as is the study of atmospheric physics and dynamics. This book merges the relevant areas of ecology and climatology, broadly defined to include weather, climate, and atmospheric composition, into an overlapping study of ecological climatology. Ecological climatology is an interdisciplinary framework to understand the functioning of plants and terrestrial ecosystems in the Earth

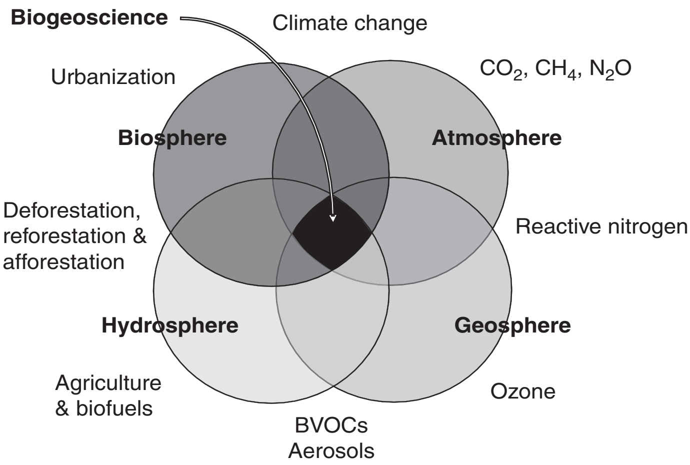

Fig. 1.4 Biogeoscience as the intersection of the atmosphere, hydrosphere, biosphere, and geosphere. Human influences are shown in the text outside the spheres.

system and the physical, chemical, and biological processes by which the biosphere affects atmospheric processes. A central theme of the book is that plants and terrestrial ecosystems, through their cycling of energy, water, chemical elements, and trace gases, are a critical determinate of climate. Changes in terrestrial ecosystems through natural vegetation dynamics, through human land uses, and through climate change itself significantly affect the trajectory of climate change.

Figure 1.5 illustrates five core areas: the *biogeophysical* and *biogeochemical* processes that regulate the exchanges of energy, water, momentum, and chemical materials with the atmosphere over periods of minutes to hours; *watersheds* and *ecosystems* and the hydrological and ecological processes that regulate these exchanges over periods of days to months; and *landscape dynamics* and the ecological and anthropogenic processes controlling the arrangement of plants into communities, the functioning of ecosystems, and temporal changes in response to disturbance over periods of years to centuries.

Biogeophysics is the study of physical interactions of the biosphere and geosphere with the atmosphere. It considers the transfers of heat, moisture, and momentum between land and atmosphere and the meteorological, hydrological, and ecological processes regulating these exchanges. Momentum is transferred when plants and other rough elements of the land surface interfere with the flow of air. Heat and moisture are exchanged when net radiation at the surface (R_n) is returned to the atmosphere as sensible heat (H), latent heat (λE), or stored in the ground (G). Biogeophysical feedbacks are understood through the surface energy balance:

$$R_n = \left(S\downarrow - S\uparrow\right) + \left(L\downarrow - L\uparrow\right) = H + \lambda E + G \qquad (1.1)$$

where $S\downarrow$ and $L\downarrow$ are downwelling solar radiation and longwave radiation onto the surface, respectively, and $S\uparrow$ and $L\uparrow$ are the upward radiative fluxes from the surface. Collectively, these four radiative fluxes comprise net radiation. The typical unit of measurement is the flux of energy per unit area (J s^{-1} m^{-2}, or W m^{-2}).

The surface energy balance highlights several important land–atmosphere interactions. One relates to surface albedo (Figure 1.6a). An increase in surface albedo, which can occur with loss of vegetation cover, increases reflected solar radiation, reduces the absorption of solar radiation at the surface, and cools the surface climate. Less energy returns to the atmosphere as sensible and latent heat, which promotes subsidence of air aloft and may reduce precipitation. Such albedo influence on rainfall is particularly important in semiarid climates. In cold, snowy climates, tall trees protrude above the snowpack and reduce surface albedo. Vegetation masking of the high albedo of snow creates a warmer climate than in the absence of trees.

Another important aspect of land–atmosphere coupling is surface roughness (Figure 1.6b).

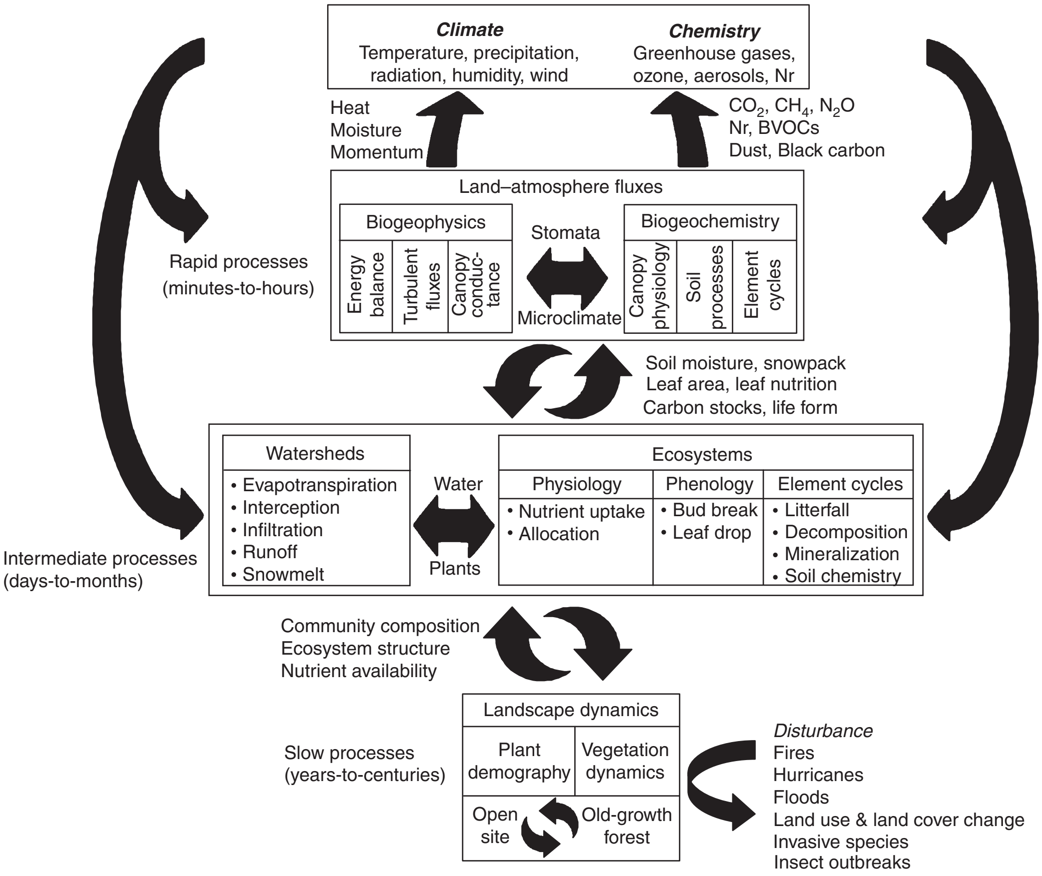

Fig. 1.5 A generalized scope of ecological climatology showing the biogeophysical and biogeochemical processes by which terrestrial ecosystems affect weather, climate, and atmospheric composition, the watershed and ecosystem processes that govern biosphere–atmosphere coupling, and the role of landscape dynamics in initiating change.

Rough surfaces such as forests generate more turbulence and have higher sensible and latent heat fluxes than smoother surfaces such as grasslands, all other factors being equal. A decrease in roughness length, by decreasing turbulence and aerodynamic conductance, can lead to a warmer, drier atmospheric boundary layer.

Biogeochemistry is the study of element cycling among the biosphere, geosphere, and atmosphere. Carbon dioxide (CO_2), methane (CH_4), and nitrous oxide (N_2O) are greenhouse gases regulated in part by terrestrial ecosystems. The net storage of carbon in the biosphere in the absence of fire and other losses, known as net ecosystem production (*NEP*), is the balance of carbon uptake during gross primary production (*GPP*), carbon loss during plant respiration (R_A), and carbon loss during decomposition (R_H):

$$NEP = GPP - R_E = (GPP - R_A) - R_H = NPP - R_H \tag{1.2}$$

The typical unit of measurement is the mass of carbon exchanged per unit area over some period of time (e.g., g C m^{-2} yr^{-1}). The net carbon uptake by plants ($GPP - R_A$) is known as net primary production (*NPP*), and the total ecosystem respiration is $R_E = R_A + R_H$. The signature of terrestrial ecosystems is seen in the annual cycle of atmospheric CO_2, which has low concentration during the growing season when plants absorb CO_2 and high concentration during the dormant season. It is also evident in the uptake

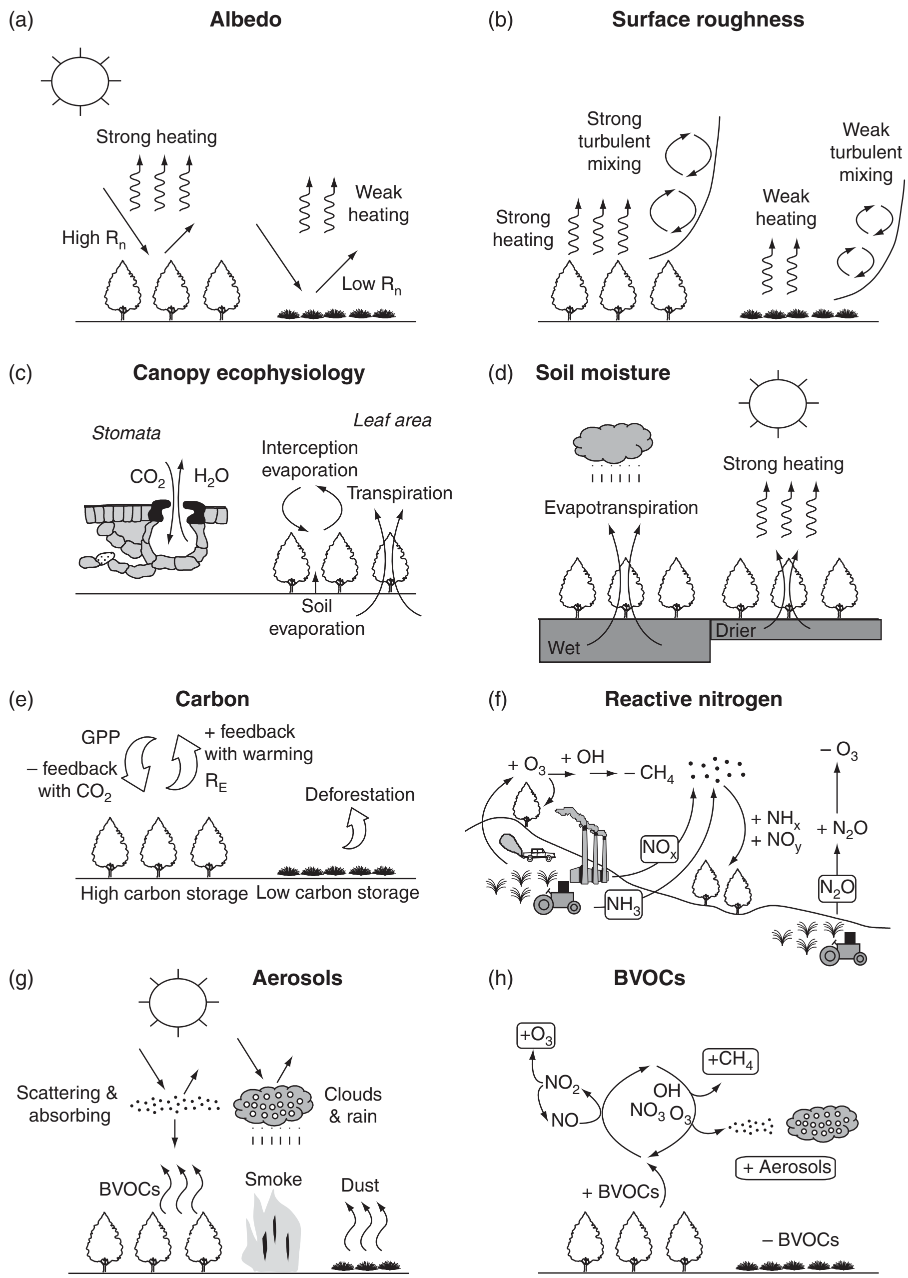

Fig. 1.6 Key land–atmosphere interactions. (a) Surface albedo and net radiation. (b) Surface roughness and turbulent mixing. (c) Canopy ecophysiology, stomata, leaf area, and evapotranspiration. (d) Soil moisture, evapotranspiration, and precipitation. (e) The carbon cycle. (f) Reactive nitrogen, atmospheric chemistry, and aerosols. (g) Aerosols, radiation, and clouds. (h) Biogenic volatile organic compounds, atmospheric chemistry, and secondary organic aerosols.

of anthropogenic CO_2 emissions by terrestrial ecosystems. Only about one-half of current anthropogenic CO_2 emissions remain in the atmosphere. Oceans and terrestrial ecosystems absorb the other half. Two key terrestrial feedbacks that regulate this are the uptake of carbon during photosynthesis with elevated atmospheric CO_2 and the loss of carbon during respiration with a warmer climate. Terrestrial ecosystems differ in these feedbacks, their carbon storage, and their capacity to sequester anthropogenic carbon emissions (Figure 1.6e).

Terrestrial ecosystems similarly regulate the concentrations of CH_4 and N_2O in the atmosphere. Ecosystems are also sources of reactive nitrogen (Nr) that alters atmospheric chemistry and produces aerosols; sources of mineral dust and black carbon (soot) from wildfires, which are important aerosols; and sources of biogenic volatile organic compounds (BVOCs), which produce ozone (O_3), increase CH_4 in the atmosphere, and form aerosols (Figure 1.6f–h).

Biogeophysical and biogeochemical processes do not occur in isolation. For example, stomata open to absorb CO_2 during photosynthesis, but in doing so water diffuses out of the leaf during transpiration (Figure 1.6c). Consequently, water loss during transpiration is tied to carbon uptake during photosynthesis. This is seen in studies that relate leaf photosynthesis, transpiration, and stomatal conductance. The physiology of stomata represents a balance between the conflicting goals of maximizing CO_2 uptake while minimizing water loss.

The exchanges of energy, water, and other materials between biosphere and atmosphere depend on the hydrologic cycle. The fundamental system of study in hydrology is a watershed, or catchment. Over long periods of time, it is commonly assumed that water entering a watershed as precipitation (P) either returns to the atmosphere as evapotranspiration (E) or runs off into streams and rivers (R) so that the annual water balance is:

$$P - E = R \tag{1.3}$$

The typical unit of measurement is the mass of water flowing per unit area over some period of time (e.g., kg H_2O m^{-2} yr^{-1}).

The hydrologic cycle influences climate in many ways. One prominent means is through latent heat flux, or evapotranspiration. A decrease in leaf area reduces the surface area for transpiration and for the interception of rainfall (Figure 1.6c). Evapotranspiration from the plant canopy decreases, but soil evaporation may increase. In general, a reduction in evapotranspiration, which occurs, for example, with deforestation, produces an increase in runoff. A decrease in vegetation cover that reduces latent heat flux warms surface climate and may reduce precipitation. This is particularly prominent in tropical deforestation. Wet soil can sustain a high latent heat flux and creates a cool, moist atmospheric boundary layer – conditions that may feed back to increase precipitation (Figure 1.6d). In contrast, dry soil decreases latent heat flux and amplifies droughts and heat waves.

Numerous topographic, edaphic, and ecological features control the hydrology of a watershed. Hydrologic processes such as evapotranspiration, interception of precipitation by plants, infiltration of water into soil, runoff into streams and rivers, and snowmelt determine soil moisture, snow pack, and saturated areas within the watershed – conditions that vary with a timescale of days to months and that influence surface fluxes.

Terrestrial ecosystems are an expression of an ecological system. All ecosystems have structure – the arrangement of materials in pools and reservoirs – and function – the flows and exchanges among these pools. For example, the carbon cycle is commonly described by pools such as foliage, stem, and root biomass and decomposing soil organic matter. Functions include carbon uptake during photosynthesis and carbon loss during respiration. A variety of ecological processes operating at timescales of days to months influence ecosystem function. The amount of leaf area is an important determinant of photosynthesis, absorption of solar radiation, heat and momentum fluxes, evapotranspiration, and interception. In many plant communities, the presence of leaves varies seasonally in relation to temperature or moisture stress. Other processes such as litterfall, decomposition, mineralization of organically bound nutrients, nutrient uptake, and the allocation of resources to growth influence carbon storage. Short-term functioning of terrestrial ecosystems is seen in the fluxes of photosynthesis and respiration in relation to the diurnal cycle of solar radiation, temperature, and humidity and day-to-day variability arising from the passage

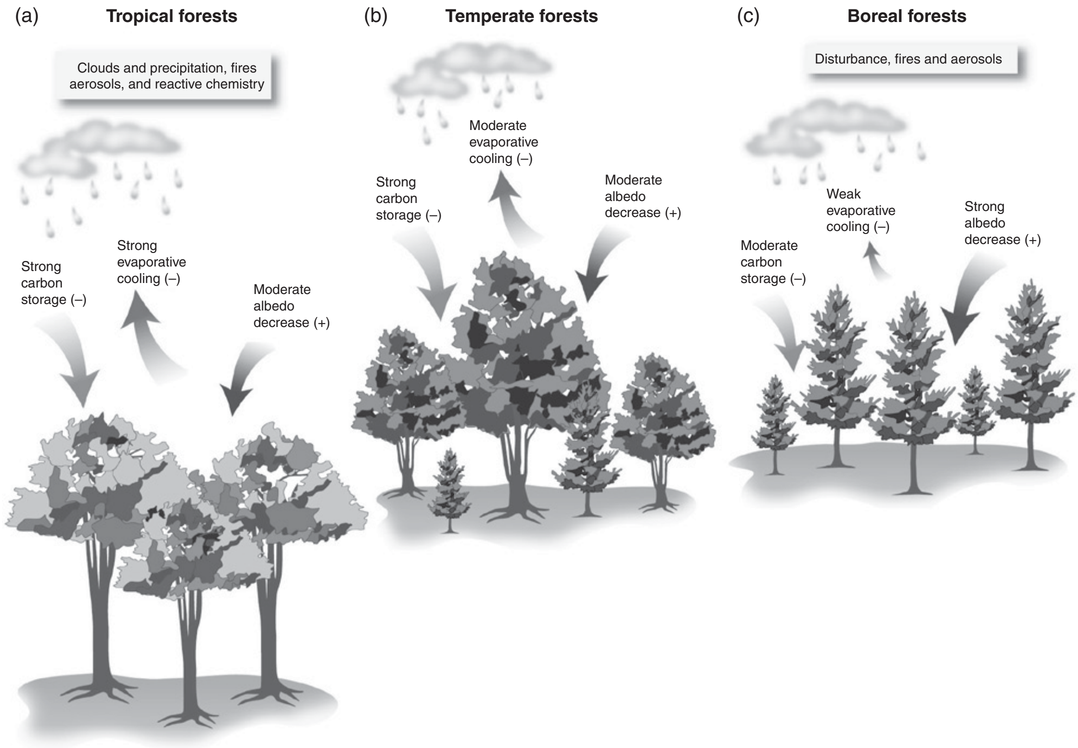

Fig. 1.7 Climate services of (a) tropical, (b) temperate, and (c) boreal forests in terms of albedo, evapotranspiration, and carbon storage. Symbols indicate a positive climate forcing (+, warming) or negative climate forcing (–, cooling). Text boxes indicate other key processes with uncertain climate influences. Adapted from Bonan (2008).

of weather systems. Interannual variability in these functions is seen in the interannual variability of atmospheric CO_2. Long-term functioning manifests in the relationships of climate with net primary production and soil carbon turnover.

Ecosystems are not just static elements of the landscape; they are dynamic. The abundance and biomass of plant species change over periods of years to centuries. Disturbances such as floods, fires, and hurricanes initiate temporal change in ecosystems known as succession. The life history patterns of plants have evolved in part as a result of recurring disturbances. Many plant species are ephemeral members of the landscape, adapted to recently disturbed sites. Others dominate old-growth ecosystems in the late stages of succession. Changes in climate and atmospheric composition affect ecosystems. Long-term changes in temperature, precipitation, atmospheric CO_2, the chemistry of rainfall, and the deposition of chemical elements onto leaves and soil alter the conditions for vegetation growth. Resulting changes in species composition, ecosystem structure, and nutrient availability each feeds back to affect climate. In particular, forest growth absorbs carbon from the atmosphere while deforestation releases carbon to the atmosphere. Human activities also alter ecosystems through clearing of land for agriculture, farm abandonment, and introduction of invasive species.

The outcome of these physical, chemical, and biological processes can be seen in the influence of the world's forests on climate, shown in Figure 1.7 (Bonan 2008). Tropical rainforests provide a negative climate forcing that cools climate. High rates of carbon storage reduce the accumulation of anthropogenic CO_2 emissions in the atmosphere (reducing greenhouse gas warming). Evaporation of the plentiful rainfall augments this with strong evaporative cooling.

In boreal forests, strong absorption of solar radiation (low surface albedo) may outweigh carbon sequestration so that the boreal forest warms global climate (positive climate forcing) compared with removal of the forest. The net climate forcing of temperate forests is more uncertain. Reforestation and afforestation sequester carbon, but biogeophysical processes augment or diminish the negative biogeochemical forcing of climate. Low surface albedo, especially during winter in snowy climates, contributes to warming, while rates of high evapotranspiration during summer contribute to cooling.

These forest–atmosphere interactions can dampen or amplify anthropogenic climate change. However, an integrated assessment of forest influences entails an evaluation beyond albedo, evapotranspiration, and carbon to include other greenhouse gases, biogenic aerosols, and reactive gases. Forests, in addition to being carbon sinks, also act as sources for aerosol particles. The combined carbon cycle and biogeophysical effect of tropical forests cools climate, but fires, biogenic aerosols, and reactive gases in these forests also affect clouds and precipitation. Biogenic aerosols are also important in boreal forests, where the net forcing from fire must also be considered.

An emerging research frontier is to link the biogeophysical and carbon cycle influences of terrestrial ecosystems with a full depiction of biogeochemical feedbacks mediated through atmospheric chemistry. Terrestrial ecosystems are sources of CH_4 and N_2O. Both are powerful, long-lived greenhouse gases. Additional reactive nitrogen increases N_2O emissions, can produce tropospheric ozone, increase the oxidation capacity of the troposphere (OH radical), decrease CH_4, form aerosols, and increase the deposition of nitrogen onto land (Figure 1.6f). Biomass burning during wildfires injects black carbon (soot) and primary organic aerosols into the atmosphere and also many short-lived gases that affect atmospheric chemistry and air quality (Figure 1.6g). Mineral aerosols (dust) are another important type of aerosol. Plants emit numerous biogenic volatile organic compounds (mostly as isoprene and monoterpenes) that increase the lifetime of CH_4 in the atmosphere and that produce ozone and secondary organic aerosols (Figure 1.6h). Tropospheric ozone is a greenhouse gas that warms climate. The effects of aerosols on climate are complex. Many aerosols scatter solar radiation back to space and cool climate; black carbon absorbs solar radiation and warms climate. Aerosols can also increase cloud brightness (reflecting more solar radiation to space) and suppress rainfall.

1.5 Timescales of Climate–Ecosystem Interactions

The coupling of ecosystems and climate occurs over a continuum of timescales from minutes to seasons to millennia (Table 1.2). At short timescales, the seasonal emergence and senescence of leaves alters the absorption of radiation, the dissipation of energy into latent and sensible heat, and CO_2 uptake. The effect of these changes can be seen in air temperature, humidity, and the seasonal drawdown of CO_2 in the atmosphere. At seasonal to interannual timescales, photosynthesis, respiration, evapotranspiration, and reactive gas fluxes influence the physical and chemical state of the atmosphere. Interannual variability in temperature and precipitation alter ecosystem metabolism, which is again evident in the concentration of CO_2 in the atmosphere.

Over several decades, people shape the landscape through clearing of land for agriculture, reforestation of abandoned farmland, and through urbanization. These land uses alter surface energy fluxes, biogeochemical cycles, and the hydrological cycle, and they produce a discernible signal in temperature, precipitation, the concentrations of CO_2, CH_4, and N_2O in the atmosphere, and the deposition of atmospheric pollutants such as reactive nitrogen, ozone, and black carbon aerosols onto land. At longer timescales of decades to centuries, slower successional changes in response to disturbances control community composition and ecosystem structure and so alter surface energy fluxes, carbon storage, and trace gas emissions. Coupled climate–ecosystem dynamics are particularly evident over periods of centuries to millennia.

Table 1.2 Timescales of vegetation change and associated atmospheric impact

Vegetation change	Timescale	Ecological signal	Controlling processes	Atmospheric signal
Leaf emergence	Seasonal to interannual	Leaf area index Canopy conductance	Air temperature Soil moisture Life form	Cooler temperature Higher humidity CO_2 drawdown Lower albedo Greater latent heat Less sensible heat
Ecosystem metabolism	Seasonal to interannual	Leaf area index Carbon storage Water balance Element cycles	Air temperature Soil moisture Humidity Solar radiation Atmospheric CO_2 Nr deposition Ozone	CO_2 drawdown Albedo Latent heat Sensible heat BVOCs Ozone
Land use → Succession → Biogeography →	Decadal Decadal to century Century to millennial	Leaf area index Carbon storage Species composition Nutrient availability Ecosystem structure	People Life history Disturbance Climate Atmospheric CO_2 Nr deposition	Temperature Precipitation Energy balance CO_2, CH_4, Nr BVOCs Aerosols Ozone
Evolution	Millennial	Stomatal density Leaf form	Atmospheric CO_2	Leaf temperature Leaf energy fluxes

Temperature, precipitation, and atmospheric CO_2 are the chief determinants of the geographic distribution of vegetation across the planet. In turn, this biogeography affects climate and atmospheric CO_2 concentration. The outcome of climate–vegetation interactions can also be seen in the evolutionary record. There is a close relationship between leaf shape and climate. Vascular plants introduced numerous biotic feedbacks on climate, primarily related to plant responses to CO_2 that affected stomatal conductance and leaf form.

1.6 Scientific Tools

The influence of plants and terrestrial ecosystems on the atmosphere can be discerned through environmental monitoring, experimental manipulation, or with the use of numerical models of weather and climate. Intensive field campaigns with ground-based measurements of biosphere–atmosphere flux exchanges and aircraft measurements of atmospheric composition provide datasets to analyze biosphere–atmosphere coupling over short time periods (up to several weeks). Environmental monitoring techniques include eddy covariance flux towers that provide continuous measurements of biosphere–atmosphere exchanges of energy, moisture, and trace gases at fast timescales (subhourly). The longest such sites have been continually operating for over two decades. Ecosystem and watershed studies provide monitoring of carbon and elemental stocks and fluxes and the hydrologic cycle, typically at longer timescales (e.g., annual). Such studies extend over several decades. Satellite observations of leaf area, surface albedo, surface temperature, and other properties provide global coverage at a high spatial resolution for a period extending now for almost three decades. Atmospheric CO_2 observations at numerous locations throughout the world provide information about the seasonal dynamics of land–atmosphere carbon exchange and continental-scale fluxes on timescales of years to decades. The longest such record, at Mauna Loa, Hawaii, dates back to 1958. Ice core measurements reveal the history of CO_2, CH_4, N_2O, and dust in the atmosphere over the past several hundred thousand years and variations with glacial–interglacial cycles.

Whole-ecosystem experimental manipulations provide insight to ecosystem responses to environmental change. Such experiments warm ecosystems or exclude rainfall to study responses to climate change, enrich the air with CO_2 to study responses to elevated atmospheric CO_2 concentrations, and fertilize the soil with nutrients (e.g., nitrogen and phosphorus) to examine responses to perturbed biogeochemical cycles. Watershed manipulation studies that remove vegetation show the biotic control of the hydrologic cycle.

The influence of plants and ecosystems on large-scale climate is difficult to establish directly through observations. Careful examination of climatic data can sometimes reveal an ecological influence, such as the effect of leaf emergence on springtime evapotranspiration and air temperature. Eddy covariance flux towers and field experiments provide local-scale insight to ecosystem–atmosphere interactions, and advances in remote sensing science aid extrapolation to larger spatial scales. More often, however, our understanding of how plants and ecosystems affect climate comes from atmospheric models and their numerical parameterizations of Earth's biosphere. Paired climate simulations, one serving as a control to compare against another simulation with altered vegetation, demonstrate an ecological influence on climate.

1.7 Overview of the Book

The book describes Earth's climate, the processing shaping climate, how climate changes over time, and the manner in which terrestrial ecosystems influence climate. It is divided into six sections on the Earth system, global physical climatology, hydrometeorology, biometeorology, terrestrial plant ecology, and terrestrial forcings and feedbacks. The first section describes component spheres of the Earth system (Chapter 2) and the energy, water, and biogeochemical cycles that link these spheres (Chapter 3).

The second section reviews climate, climate variability, and climate change. The radiative balance of the atmosphere, its geographic variation, and its annual cycle are an important determinant of climate (Chapter 4). Geographic and seasonal variation in the radiative balance drives the general circulation of the atmosphere (Chapter 5). This gives rise to Earth's macroclimates, and within which mountains, lakes, and vegetation create local climates (Chapter 6). The realized temperature and precipitation in any year can deviate markedly from the long-term climatology because of seasonal-to-interannual atmospheric variability such as the El Niño/Southern Oscillation and North Atlantic Oscillation (Chapter 7). Climate also changes over longer timescales of centuries and millennia in response to changes in insolation, greenhouse gases, and numerous feedbacks within the Earth system (Chapter 8).

The third section on hydrometeorology reviews the hydrologic cycle, surface energy fluxes, and the interactions between the hydrosphere and atmosphere. Soils store vast amounts of energy and water, and this modulates the diurnal and annual cycle of temperature, provides water for evapotranspiration, and regulates the hydrologic cycle on land (Chapter 9). The hydrologic cycle on land is reviewed in terms of point processes (Chapter 10) and watershed processes (Chapter 11). The hydrologic cycle regulates surface energy fluxes. The energy balance at Earth's land surface requires that energy gained from net radiation be balanced by the fluxes of sensible and latent heat to the atmosphere and the storage of heat in soil (Chapter 12). The fluxes of sensible and latent heat occur because turbulent mixing of air transports heat and moisture, typically away from the surface (Chapter 13). Soil moisture exerts a strong control on the partitioning of net radiation into sensible and latent heat fluxes, and through this affects the atmospheric boundary layer (Chapter 14).

The fourth section reviews biometeorology. The exchanges of sensible heat, latent heat, and CO_2 between land and atmosphere are regulated by the physiology and micrometeorology of plant canopies. Individual leaves absorb radiation and exchange sensible heat and latent heat with the surrounding air (Chapter 15). The uptake of CO_2 during photosynthesis is tightly coupled to the loss of water during transpiration (Chapter 16). Both occur through stomatal openings on the leaf surface. The aggregate flux from vegetation is the integral of the individual leaf fluxes over the depth of the canopy (Chapter 17).

The fifth section reviews terrestrial plant ecology. It extends the discussion of plant physiology from the previous section to an overview of whole-plant allocation and plant strategies (Chapter 18), the arrangement of plant species into populations and communities (Chapter 19), and the functioning of ecosystems (Chapter 20). The weathering of rocks, the decomposition of soil organic material, and soil formation are part of the biogeochemical cycling of carbon, nitrogen, and other elements among the atmosphere, biosphere, and geosphere (Chapter 21). Vegetation changes over time in response to recurring disturbance (Chapter 22). Landscapes represent another level of ecological organization, merging the concepts of populations, communities, ecosystems, and plant dynamics. Spatial gradients in the environment combine with natural and anthropogenic disturbances to create a mosaic of plant communities and ecosystems across the landscape (Chapter 23). The structure and composition of vegetation and the functioning of terrestrial ecosystems, which at a local scale are shaped by environmental factors such as temperature and moisture, are also influenced by global climate. This is seen in the biogeography of vegetation and in the global carbon cycle, especially net primary production (Chapter 24).

The final section examines terrestrial forcings and feedbacks in the Earth system, especially how natural and human changes in land cover and land use affect climate. Numerous global climate model experiments have demonstrated the role of land surface hydrology and terrestrial vegetation in determining regional and global climate. Chapter 25 reviews the representation of land surface processes in global models. Soil moisture, snow, and vegetation contribute to climate variability (Chapter 26). Vegetation dynamics in response to climate change alters climate by changing surface albedo, net radiation, and evapotranspiration. The boreal forest–tundra ecotone and the Sahel of North Africa are prominent examples of coupled climate–vegetation dynamics (Chapter 27). Deforestation, reforestation, degradation of drylands, and cultivation of croplands are case studies of how human uses of land alter climate through biogeophysical processes (Chapter 28). In addition, terrestrial ecosystems are coupled to climate through various biogeochemical cycles. The carbon cycle is a prominent feedback with climate change (Chapter 29), as are reactive nitrogen (Chapter 30) and aerosols (Chapter 31). Urbanization also alters climate and the hydrologic cycle (Chapter 32). Because terrestrial ecosystems have such a significant impact on climate through albedo, evapotranspiration, carbon storage, and other processes, they can be managed to mitigate the undesirable effects of anthropogenic climate change (Chapter 33). Greater understanding of Earth and its climate requires that all components of the Earth system – physical, chemical, biological, socioeconomic – be considered. Terrestrial ecosystems – nature's technology – are a critical aspect of planetary habitability in an ever increasing technological world (Chapter 34).

1.8 Review Questions

1. Scientists have found fossilized remains of tree foliage in Antarctica that date to 100 million years ago. What does this indicate about the climate of that era?

2. A 15 km^2 region has areas of forest and open land, with an average elevation of 50 m above sea level. Within this region 11 meteorological stations were established in the open land and in small clearings within the forests, all at a height of 1.07 m above the ground. The measurements, obtained in 1886, showed annual precipitation in the open land averaged 383 mm ($n = 4$); the forests averaged 448 mm of rainfall ($n = 7$). Does this support the notion that forests increase rainfall?

3. In the semiarid Sahel region of North Africa, annual rainfall is low and the vegetation is sparse with low annual productivity. Years with high rainfall anomalies have greater vegetation cover and productivity. What does this relationship indicate about rainfall and vegetation?

4. A paired watershed study compares one catchment that is forested and an adjacent catchment that is deforested. Both receive similar annual rainfall, but runoff to streams is greater in the deforested catchment. What does this show about forest–climate influences?

5. Describe changes in ecosystem structure and function with reforestation that affect climate.

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Part I

The Earth System

Components of the Earth System

2.1 Chapter Summary

Earth's climate is understood in terms of a system of several interacting spheres and the energy, water, and biogeochemical cycles that link these spheres. The main components of the Earth system are: atmosphere, air; hydrosphere, water; cryosphere, frozen portion of Earth; biosphere, living organisms; pedosphere, soil; and anthroposphere, humans. People are important agents of environmental change through land use and land-cover change and co-option of the hydrologic cycle and biogeochemical cycles. Numerous physical, chemical, and biological processes within the Earth system feed back to accentuate or mitigate climate change. Many of these feedbacks relate to terrestrial ecosystems and human activities. Greater understanding of Earth and its climate requires that all components of the Earth system – physical, chemical, biological, socioeconomic – be considered.

2.2 Atmosphere

The atmosphere is the air that surrounds Earth. It is comprised primarily of nitrogen (N_2) and oxygen (O_2), which together account for 99 percent of the volume of the atmosphere (Table 2.1). Many other gases occur in trace amounts that when combined comprise less than 1 percent of the volume of the atmosphere. Although they occur in minor quantities, some of these gases play an important role in Earth's radiation balance through the greenhouse effect.

Air pressure is a measure of the mass of air above a given point. The total pressure exerted by a parcel of air is the sum of the pressures of all the individual gases in the parcel. Nitrogen, which comprises 78 percent of the air, exerts the most partial pressure, followed by oxygen (21%). Water vapor typically comprises 1–4 percent of air. For example, the atmospheric pressure near sea level is about 1000 hectopascals (hPa, 1 hPa = 100 Pa = 1 millibar). The partial pressure of nitrogen is 780 hPa and oxygen is 210 hPa. If water vapor comprises 1 percent of the parcel, its partial pressure is 10 hPa or 1000 Pa. Because water vapor is only a small constituent of air, vapor pressure is only a small component of total air pressure. Carbon dioxide has a partial pressure of about 40 Pa.

Greenhouses gases are poor absorbers of solar radiation, but are strong absorbers of longwave radiation. As a result, the Sun's radiation passes through the atmosphere and heats the surface, but greenhouse gases in the atmosphere absorb the longwave radiation emitted by the surface. The majority of this longwave radiation is emitted back to the surface, warming the surface. This reemission of terrestrial longwave radiation back to the surface is the greenhouse effect that warms the surface.

The principal greenhouse gases are water vapor (H_2O), carbon dioxide (CO_2), methane (CH_4), and nitrous oxide (N_2O). The amount of water vapor in the atmosphere varies

Table 2.1 Chemical composition of the atmosphere

Gas	Chemical symbol	Percent (by volume)	
		Current (2011)	Preindustrial (1750)
Nitrogen	N_2	78.08%	
Oxygen	O_2	20.95%	
Argon	Ar	0.93%	
Trace gases		<1%	
Water vapor	H_2O	0–5%	
Carbon dioxide	CO_2	390 ppm	278 ppm
Methane	CH_4	1803 ppb	722 ppb
Nitrous oxide	N_2O	324 ppb	270 ppb

Note: Of the many gases that occur in trace amounts, only the four major greenhouse gases are shown. One part per million (ppm, μmol mol^{-1}) = 0.0001%. One part per billion (ppb, nmol mol^{-1}) = 0.0000001%. CO_2, CH_4, and N_2O are from Hartmann et al. (2013).

geographically and seasonally and can be as high as 5 percent of the atmosphere, with more water vapor in the warm tropics than in colder polar regions and more water vapor in warm seasons than in cold seasons. The concentrations of CO_2, CH_4, and N_2O vary over time. The concentration of CO_2 in the atmosphere averaged for the year 2011 was 390 parts per million (ppm, or μmol mol^{-1}; more precisely, this is the mole fraction, defined as the number of CO_2 molecules in a given number of molecules of dry air) – a 40 percent increase from preindustrial levels, primarily as a result of human activities such as fossil fuel burning and deforestation. The concentrations of CH_4 and N_2O have similarly increased over the past few centuries. Although these gases occur in lower concentration than CO_2, their global warming potential is much greater. Global warming potential measures the effectiveness of gases in absorbing outgoing terrestrial radiation combined with their lifetime in the atmosphere. It is a measure of the total energy added to the climate system relative to that added by CO_2. The 100-year global warming potential of CH_4 is 28 times that of CO_2, and N_2O has a 100-year global warming potential 265 times that of CO_2 (Myhre et al. 2013). Ozone and halocarbons (chlorofluorocarbons (CFCs), hydrofluorocarbons (HFCs), and other carbon compounds containing fluorine, chlorine, bromine, or iodine) are other greenhouse gases.

In addition to these gases, the atmosphere contains microscopic particles ranging in size from a few nanometers to tens of micrometers, known as aerosols (Boucher et al. 2013). Primary aerosols enter the atmosphere directly as dust from land, salts from ocean spray, black carbon (soot) from fires, and volcanic ash. Wind erosion from arid and semiarid environments carries the largest amount of mineral aerosols into the atmosphere. Sea salt produced by breaking waves is a similarly large source of aerosols. Secondary aerosols form when chemical reactions in the atmosphere convert emitted gases to particles. Biological processes on land and in oceans emit sulfate and organic particles. Over land, organic condensates form during chemical reactions involving the emission of nonmethane hydrocarbons from terrestrial vegetation. Human activities produce a wide variety of primary and secondary aerosols, though the total emission is small compared with natural emissions. Sulfate aerosols produced through the combustion of sulfur-containing fossil fuels are particularly important. Airborne particles such as dust and sulfate aerosols directly affect climate by absorbing or scattering solar radiation. In addition, sulfate aerosols alter climate indirectly by influencing the number and size of cloud droplets. In this way, clouds brighten, reducing the sunlight reaching the surface. Black carbon deposited on snow and ice darkens the surface and enhances solar heating of the surface.

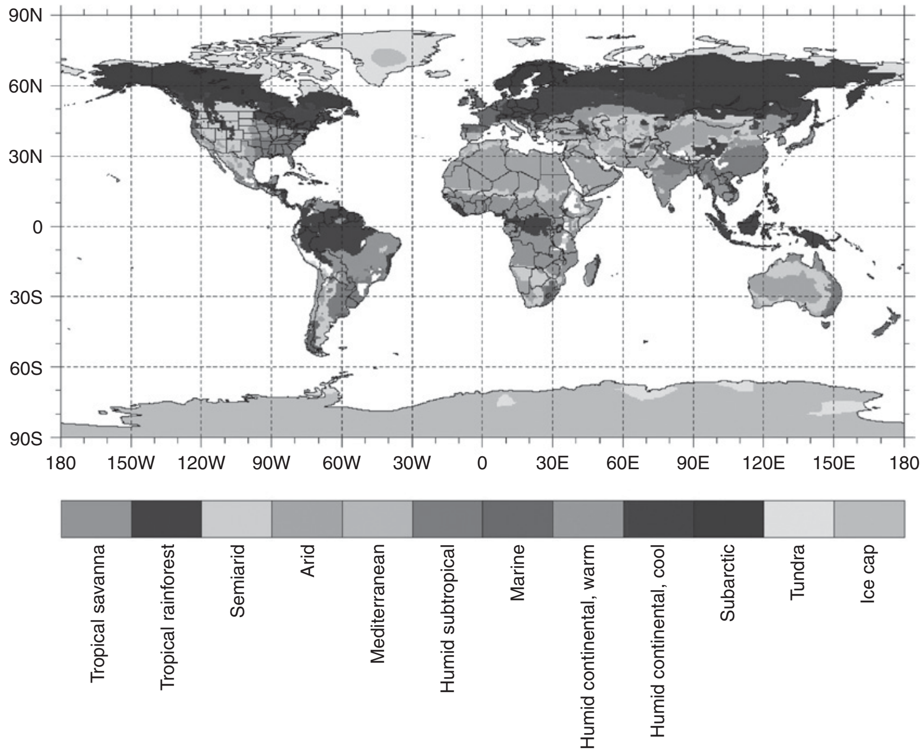

Fig. 2.1 Earth's major climate zones following the Köppen classification as modified by Trewartha (Chapter 6). See color plate section.

The atmosphere is constantly in motion. These motions arise from the heating of the air, land, and oceans by the Sun and the resultant geographic redistribution of heat by atmospheric and oceanic circulations. These motions can be seen over periods of a few minutes in the fluttering of flags in a strong breeze, over the course of a day in the formation of cumulus clouds and afternoon thunderstorms, or in the passage of warm or cold fronts. Atmospheric phenomena at these short timescales are referred to as weather. Whereas weather can be thought of as describing the instantaneous state of the atmosphere, climate describes the average weather, or long-term state of the atmosphere, over periods of many years.

Earth's surface has distinct climate zones defined by temperature and precipitation (Figure 2.1). Tropical climates occur near the equator where monthly mean temperatures are warmer than 18°C. Polar climates form near the poles, where the warmest month of the year is colder than 10°C. Polar climates are categorized as tundra or ice cap based on temperature. In between are the middle latitude climates with both warm and cold seasons. Several such mid-latitude climates form depending on temperature. For example, the subarctic climate zone has less than four months with monthly mean temperature greater than 10°C. Similarly, there are major geographic precipitation patterns. Tropical regions along the equator receive abundant rainfall year-round (tropical rainforest climate). Other tropical regions receive less annual rainfall and have pronounced wet and dry seasons (tropical savanna). Middle latitudes are generally moist, though arid climates, categorized as semiarid or arid based on decreasing moisture, develop along the subtropical high pressures at latitude 30° in both hemispheres and in regions far removed from sources of atmospheric moisture. The Mediterranean climate is a distinct mid-latitude climate with a dry season in summer. Polar regions are

Table 2.2 Water storage on Earth

Pool	Volume (km^3)	Percent of total water	Percent of freshwater
Ocean	1,335,040,000	97.0	–
Freshwater	41,984,700	3.0	–
Icecaps, glaciers, and permafrost	26,372,000	1.9	62.8
Groundwater	15,300,000	1.1	36.4
Lakes and rivers	178,000	0.01	0.42
Soil water	122,000	0.01	0.29
Atmosphere	12,700	0.001	0.03

Source: From Trenberth et al. (2007). See also Oki and Kanae (2006).

generally dry because the cold air holds little moisture.

Earth's climate has changed in the past, and is changing still. Global mean planetary temperature increased by 0.85°C over the period 1880–2012 (Hartmann et al. 2013). Human activities that increase greenhouse gases and aerosols in the atmosphere have contributed to this climate change (Bindoff et al. 2013).

2.3 Hydrosphere

The hydrosphere describes the water on Earth held in rivers, lakes, oceans, the ground, and air. Earth holds about 1377 million km^3 of water. This is an amount that if spread over Earth's 510 million km^2 surface area would be about 2700 m deep. Oceans hold 97 percent of this water (Table 2.2). This is an average depth of about 3700 m spread over the 360 million km^2 surface area of oceans. Another 1.9 percent is frozen in polar icecaps, glaciers, and permafrost. Only 1.1 percent of Earth's water is liquid on land. Rivers, lakes, and wetlands contain 178,000 km^3 of water on the surface. Over 15 million km^3 of water is below ground. Deep aquifers that comprise groundwater hold most of this water; the soil near the surface holds very little water (122,000 km^3). The atmosphere has the least amount of water, about 13,000 km^3 or approximately 25 mm of water spread over Earth's surface area.

Water is an important part of Earth's climate. The oceans store and transport heat, redistributing the uneven geographic heating of Earth by the Sun. Carbon storage in the oceans regulates atmospheric CO_2 concentration. Water vapor is the most important greenhouse gas both in concentration and radiative warming. Water vapor condenses to form clouds, which can precipitate rainfall and which affect the planetary radiation budget by reflecting solar radiation and absorbing and emitting longwave radiation. The latent heat released during condensation provides considerable energy to fuel storms. The hydrologic cycle among the atmosphere, ocean, and land regulates the amount of water vapor in the air. Rates of precipitation and evaporation depend on temperature and other climatic factors so that as climate changes the amount of water vapor in the atmosphere also changes.

2.4 Cryosphere

The cryosphere is the frozen portion of Earth including glaciers, sea ice, freshwater ice, snow, and permafrost. Glaciers are large, thick masses of ice accumulated from snowfall. They can be geographically small such as mountain glaciers or extensive ice sheets such as those that cover Greenland and Antarctica. Glaciers are the largest store of freshwater (Table 2.2). The melting of glaciers with a warmer climate contributes to sea level rise. Conversely, sea level was lower during the last glacial maximum some 21,000 years

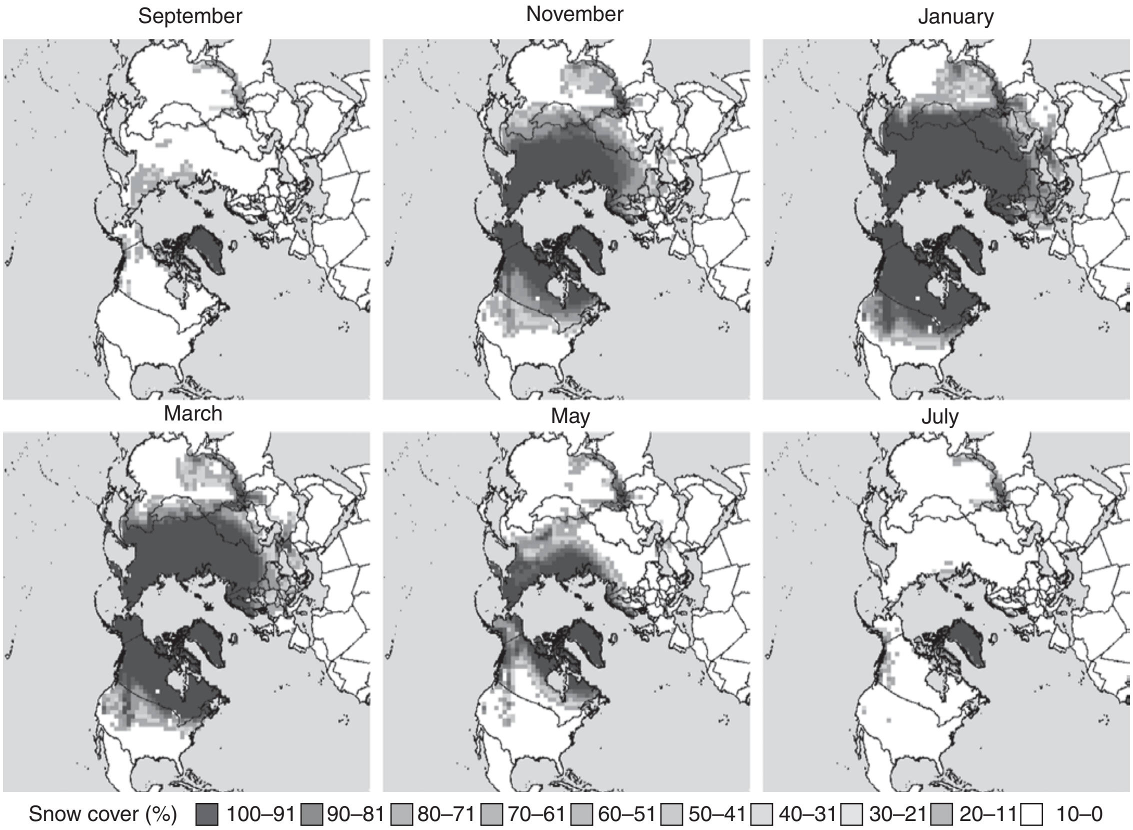

Fig. 2.2 Monthly snow cover climatology in the Northern Hemisphere for the period 1966–1999. Maps provided courtesy of David Robinson (Global Snow Lab, Rutgers University). See color plate section.

before present when glaciers covered a large portion of North America and Eurasia. Glaciers also affect climate through the planetary energy balance. Glaciers have a much larger albedo than do oceans, soil, or vegetation, which decreases the amount of solar radiation absorbed by the surface and contributes to planetary cooling.

Snow cover and sea ice extent vary seasonally. At its winter maximum, snow covers about 46 million km^2 of land in the Northern Hemisphere (Figure 2.2). Sea ice covers 14–16 million km^2 in the Arctic Ocean and 17–20 million km^2 in the oceans around Antarctica at its seasonal maximum. This sea ice melts during summer to 6–8 million km^2 in the Arctic and 3–4 million km^2 around Antarctica. Similar to glaciers, snow and sea ice affect planetary energetics through high albedo. Fresh snow, for example, can reflect 80–90 percent of incoming solar radiation compared with 10–20 percent for soil or vegetation.

Permafrost is soil or rock where temperatures remain below freezing for two or more years. It is common throughout the Arctic over vast regions of tundra and boreal forest vegetation. Many factors influence the geographic distribution of permafrost, such as air temperature, snow cover, vegetation cover, the presence of an insulating organic layer, and soil water. Soil temperature and permafrost control ecological and biogeochemical processes at high latitudes. Thawing of permafrost with warmer temperatures alters the biogeography, productivity, and carbon balance of arctic ecosystems. Permafrost restricts infiltration and drainage, leading to wet soils and standing surface water, and the degradation of permafrost impacts runoff of freshwater to the Arctic Ocean.

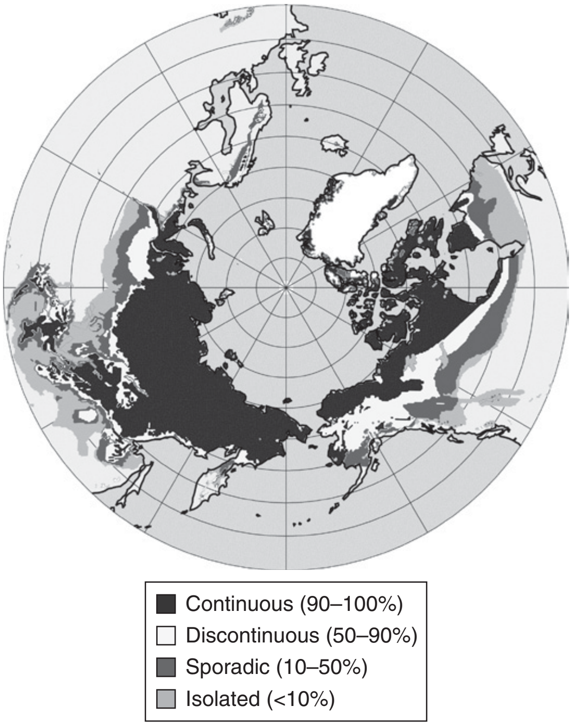

Fig. 2.3 Geographic distribution of permafrost showing the extent of continuous, discontinuous, sporadic, and isolated patches of permafrost in terms of percentage area. Reproduced from Lawrence and Slater (2005).

Permafrost occupies approximately 23 million km^2, or 24 percent of land in the Northern Hemisphere (Figure 2.3). Continuous permafrost, where permafrost underlies more than 90 percent of the land, extends over some 11 million km^2 including northern regions of Canada, Alaska, and Russia and is widespread throughout Siberia extending to northern China and Mongolia. In more southern locations, permafrost is discontinuous (50–90% of land), sporadic (10–50%), or occurs in isolated patches (<10%) depending on local climate and site factors.

2.5 Biosphere

The biosphere describes the plants, animals, fungi, bacteria, algae, and other living organisms that inhabit Earth. Of particular interest in the atmospheric sciences are terrestrial ecosystems. Much of the carbon in biologically active pools is stored in vegetation and soil (Ciais et al. 2013). It is estimated that plant biomass contains 450–650 Pg (1 Pg = 10^{15} g) of carbon. Soils hold more than three times as much carbon (1500–2400 Pg), with an additional ~1700 Pg C or more locked in permafrost. For comparison, the atmosphere contains about 800 Pg of carbon. The uptake of carbon during photosynthesis and the release of carbon during respiration by plants and soil microorganisms are important parts of the global carbon cycle. In addition, terrestrial ecosystems regulate other biogeochemical cycles that are important to climate (e.g., CH_4, N_2O), the emission of mineral aerosols into the atmosphere, and the emission of chemically reactive gases that affect air quality (e.g., biogenic volatile organic compounds). Plants, by regulating evapotranspiration, infiltration, and runoff, are a key component of the hydrologic cycle.

The natural vegetation of Earth has a distinct geographic pattern that corresponds to climate zones (Figure 2.4). Forests grow in tropical rainforest, humid subtropical, marine, humid continental, and subarctic climates. In these regions, precipitation is abundant year-round. Trees cannot survive in the bitter cold of tundra climates. Instead, small shrubs, herbaceous plants, and mosses grow in the short summers. Extensive grasslands occur in the semiarid and savanna climates of central North America, northern and central South America, central and southern Africa, central Asia, and Australia. Here climate is hot and dry. Short, dense woody bushes form chaparral (also known as Mediterranean) vegetation in the Mediterranean climate where summers are hot and dry and winters are mild and moist. Deserts, with sparse or widely spaced scrubby plants, establish in arid climates. The close correspondence between climate zones and vegetation zones is readily apparent. Climate zones such as tropical savanna, tropical rainforest, Mediterranean, and tundra are named after vegetation.

2.6 Pedosphere

The lithosphere is the solid outer layer of Earth including the crust and upper mantle. Its

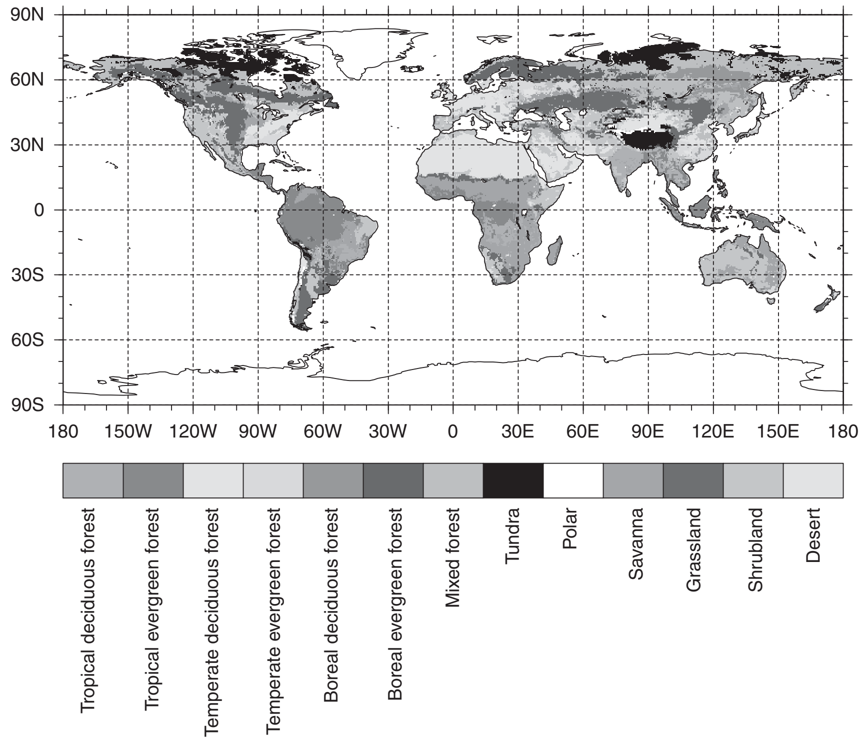

Fig. 2.4 Biogeography of natural vegetation prior to human land use. Data from Ramankutty and Foley (1999). See color plate section.

outermost layer is called the pedosphere, or soil. Soil is the interface in the cycling of energy and materials between the atmosphere and land. It is the matrix through which energy, water, biomass, and nutrients flow. It is the location of large transformations of energy as radiation absorbed by the surface becomes sensible heat, latent heat, or is stored in the ground. Soil is the source of water and nutrients for plant growth. The recycling of nitrogen and other nutrients from vegetation back to the soil is crucial to both soil development and plant growth. Soils store vast amounts of carbon that is slowly released to the atmosphere during decomposition. The transformation of this material to decomposed humus releases nutrients that support plant growth. Numerous microflora and microfauna facilitate the cycling of carbon and nutrients among soil, living biomass, and air.

Soils develop over time in relation to climate, vegetation, and the underlying geologic parent material. There are twelve broad classes of soil, known as soil orders, that differ in color, texture, structure, and chemical and mineralogical properties. Many of these twelve soil orders closely relate to climate and vegetation (Figure 2.5). Entisols and aridisols are soils with little organic matter or soil development and are common in arid climates in association with deserts. Gelisols are cold soils underlain with permafrost and develop throughout the Arctic in association with tundra. Histosols are organic soils with thick peat layers that develop in wet conditions associated with marshes, swamps, and bogs. They are most prevalent in cold climates. Spodosols develop in boreal and cool temperate needleleaf evergreen forests, where the acidic litter enhances leaching. Mollisols are the thick soils with high organic matter content found in association with prairie vegetation. Ultisols are highly weathered clay soils that develop in warm to tropical climates. Oxisols are the most highly weathered soils, with high clay content, and develop where the climate is

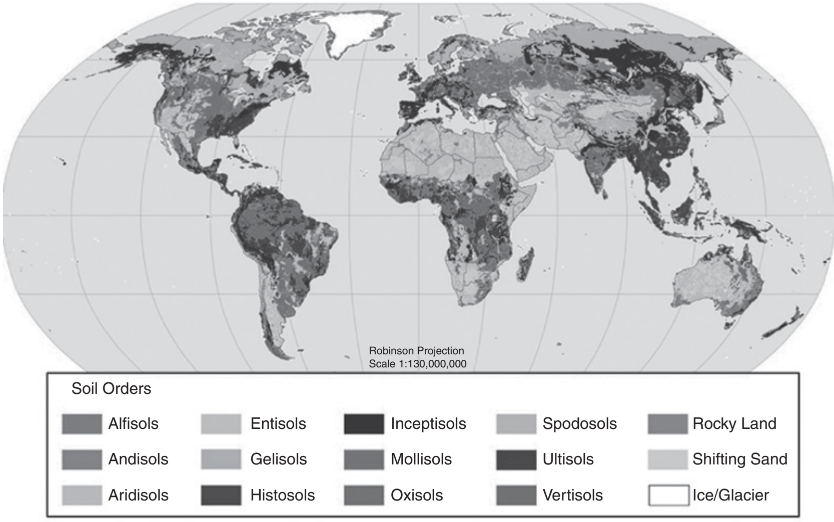

Fig. 2.5 Geographic distribution of the twelve soil orders. Image provided courtesy of the U.S. Department of Agriculture Natural Resources Conservation Service, Soil Survey Division, World Soil Resources, Washington D.C. See color plate section.

hot and wet throughout the year in association with tropical rainforests.

2.7 Anthroposphere

The world's population in 2005 was estimated to be 6.5 billion people. This is more than a sevenfold increase from that prior to the industrial era, and population is expected to increase further to 8–10 billion by 2050 (Figure 2.6). The ever increasing spread of humanity across the planet has lead to growing recognition of people as agents of environmental change (Vitousek et al. 1997; Foley et al. 2005). Very little of the planet is untouched in some form by human activities. The anthroposphere represents humankind, our socioeconomic systems, and our activities. It is our cities, towns, and villages; the agriculture, energy, and water to sustain the populace; our transportation systems; and our collective influence on the environment. One of the more prominent anthropogenic effects is the emission of CO_2 to the atmosphere from fossil fuel combustion (Figure 2.7). This CO_2 emission has increased to more than 9 Pg C yr^{-1}, and atmospheric CO_2 concentration has increased as a result. The outcome of human activities on planetary functioning is also seen in the elevated levels of atmospheric CH_4, N_2O, and other greenhouse gases over the preindustrial era; by increasing emissions and subsequent deposition of reactive nitrogen; and through land use practices such as agriculture, deforestation, afforestation, and reforestation. By altering biogeochemical cycles, the hydrologic cycle, and energy fluxes, human activities have a significant impact on climate. Inclusion of past and future changes in human activities is an important component of climate change simulations.

Over the past 300 years, human uses of land removed 7–11 million km^2 of forest worldwide (Foley et al. 2005). Managed forests replaced an additional 1.9 million km^2 of natural forest land. Presently, croplands and pastures cover more than one-third of the land surface (Figure 2.8). Croplands cover significant portions of central North America, Europe, and Asia. Pastures and rangelands are extensive throughout the western United States, central Asia, and the tropics. Regionally, large tracts of land have been urbanized, as seen in the nighttime lights of the world (Figure 2.9). It is estimated that about 84,000 km^2 of land in the conterminous United States (an area approximately equal to the state of South Carolina) is covered by buildings, roofs,

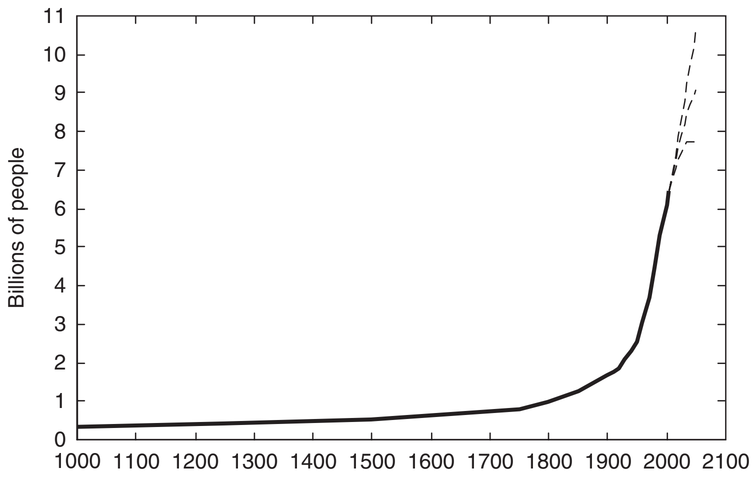

Fig. 2.6 World population from 1000 CE to 2005. Projected population through 2050 is shown for low, medium, and high growth (dashed lines). Data provided courtesy of the Population Division, Department of Economic and Social Affairs, United Nations Secretariat.

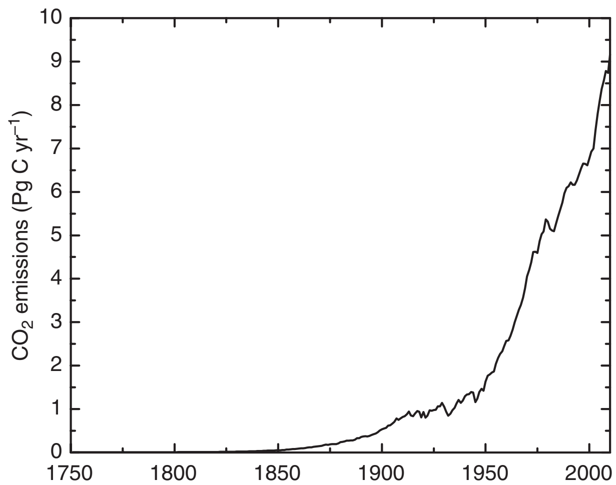

Fig. 2.7 World CO_2 emissions from fossil fuel combustion and cement production. Andres et al. (2012) describe the data. Data provided by the Carbon Dioxide Information Analysis Center (Oak Ridge National Laboratory, Oak Ridge, Tennessee).

roads, parking lots, and other impervious surfaces (Elvidge et al. 2007); globally, impervious surfaces cover an area of about 580,000 km^2 (larger than France). Other data analysis suggests that the area of cities worldwide may be 660,000 km^2 (Schneider et al. 2009).

Anthropogenic land-cover change and human uses of land alter ecosystem functions. About 20–30 percent of the net primary production of the world's terrestrial ecosystems is managed by humans (Vitousek et al. 1986; Rojstaczer et al. 2001; Imhoff et al. 2004), and the amount has increased over time (Krausmann et al. 2013). The clearing of forests for agricultural land releases much of the carbon stored in trees and soils to the atmosphere. It is estimated that from 1750 to 2011 an amount of carbon equal to one-half that emitted during the combustion of fossil fuels over the same period was released to the atmosphere as a result of changes in land use (Ciais et al. 2013). Vast sums of available renewable freshwater are withdrawn annually for agricultural, industrial, and municipal uses (Postel et al. 1996; Oki and Kanae 2006).

The production of nitrogen fertilizers, the cultivation of legumes and other nitrogen-fixing crops, and to a lesser extent the combustion of fossil fuels have increased the amount of reactive nitrogen in the Earth system (Galloway et al. 2003, 2004, 2008; Gruber and Galloway 2008; Erisman et al. 2011). The Haber–Bosch process to produce ammonia (NH_3) from N_2 for

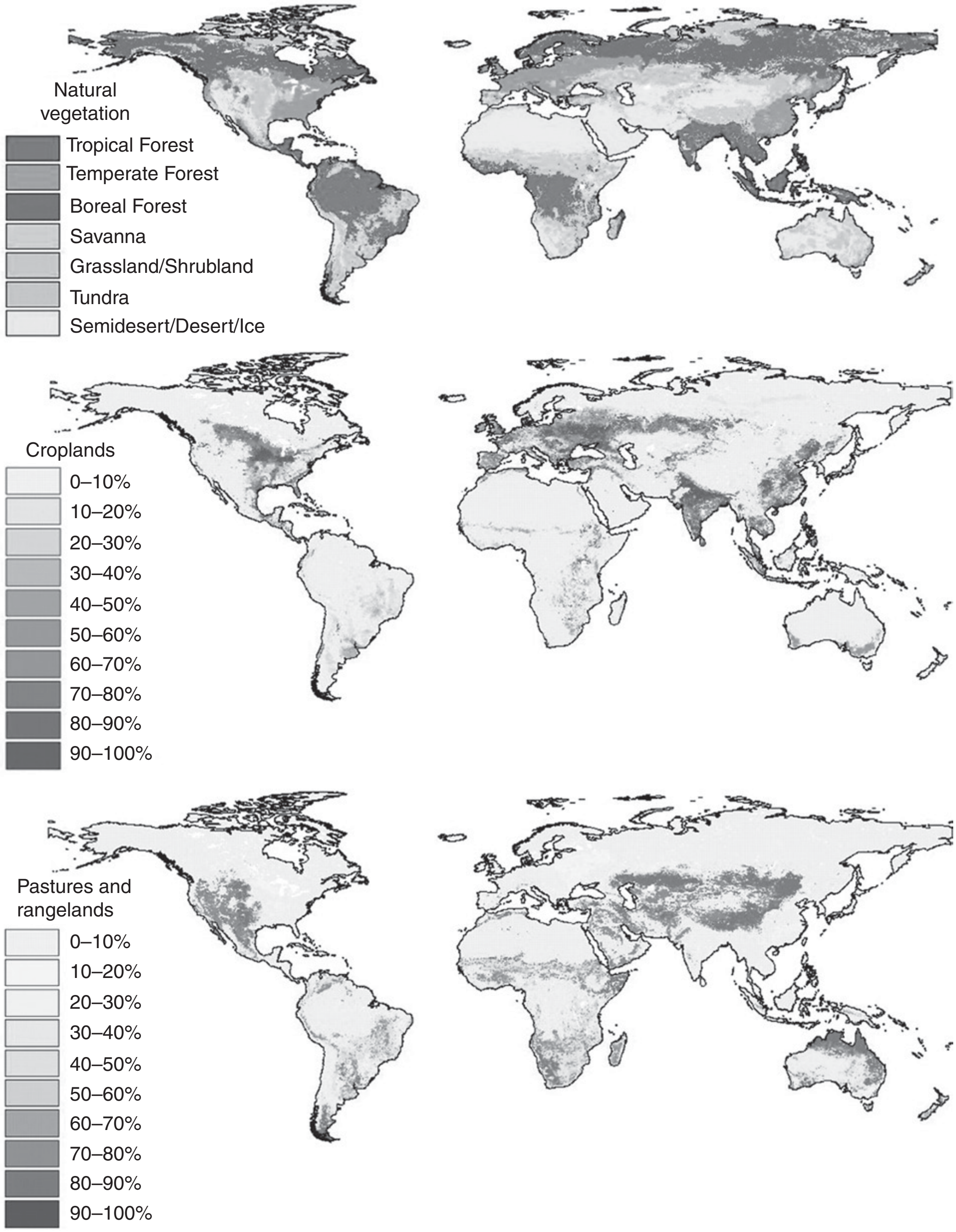

Fig. 2.8 Natural vegetation prior to human land use (top) and the extent of agricultural land during the 1990s. The panels for croplands (middle) and pastures (bottom) show the percentage of the land surface occupied by these land cover types. Reproduced from Foley et al. (2005). See color plate section.

fertilizers and other uses is the single largest source of reactive nitrogen (Figure 2.10). In the late-twentieth century, human production of reactive nitrogen became larger than natural inputs. The Haber–Bosch process produces over 100 Tg N yr^{-1} (1 Tg = 10^{12} g). Cultivation of crops adds additional nitrogen, and fossil fuel combustion adds still more nitrogen. Food production

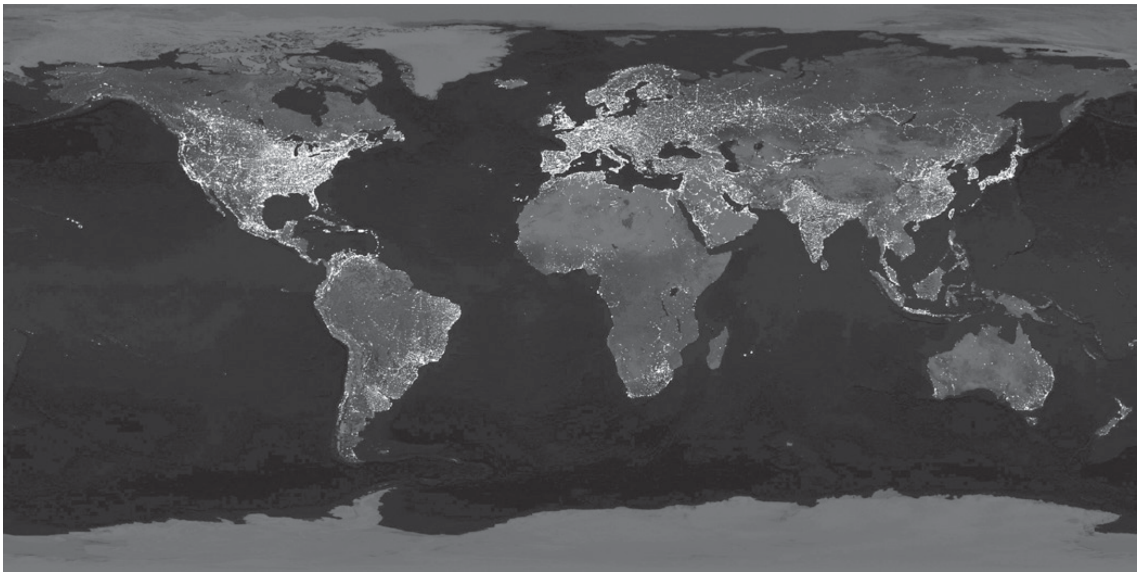

Fig. 2.9 Urban areas as seen by satellite in nighttime lights of the world. Image from the Defense Meteorological Satellite Program and provided by the National Geophysical Data Center (National Oceanic and Atmospheric Administration, Boulder, Colorado) and Goddard Space Flight Center (National Aeronautics and Space Administration, Greenbelt, Maryland) courtesy of Marc Imhoff (GSFC) and Christopher Elvidge (NGDC). The nighttime lights dataset provides a means to monitor urban land cover (Elvidge et al. 1997; Imhoff et al. 1997; Gallo et al. 2004; Small et al. 2005).

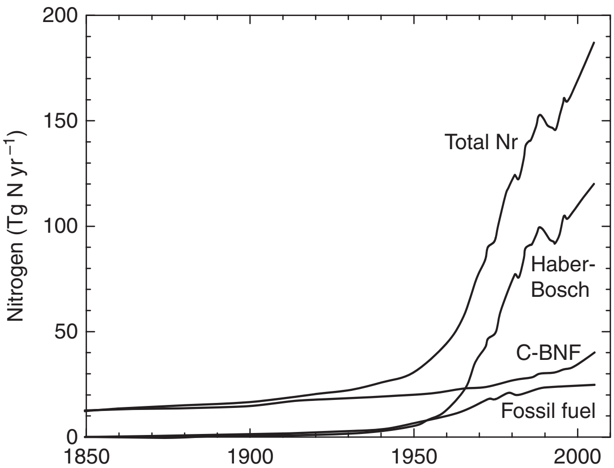

Fig. 2.10 Reactive nitrogen (Nr) creation from fossil fuel combustion, cultivation of biological nitrogen-fixing crops, and the Haber–Bosch process. Adapted from Galloway et al. (2003).

accounts for three-quarter of the reactive nitrogen created by humans. The increased availability of nitrogen contributes to poor air quality and anthropogenic climate change.

Human activities will continue to evolve in the future. Representative concentration pathways (RCPs) are a comprehensive set of four concentration and emission scenarios developed for use as input to models to assess anthropogenic climate change over the twenty-first century (Moss et al. 2010; van Vuuren et al. 2011a). They depict plausible scenarios of how the future may look with respect to population growth, socioeconomic change, technological change, energy consumption, and land use; resulting emissions of greenhouse gases, reactive gases,

Table 2.3 Main characteristics of each representative concentration pathway (RCP)

Scenario component	RCP2.6	RCP4.5	RCP6.0	RCP8.5
Radiative forcing	2.6 W m^{-2}	4.5 W m^{-2}	6.0 W m^{-2}	8.5 W m^{-2}
Greenhouse gas emissions	Very low	Medium-low	Medium-high	High
Agricultural area	Medium for cropland and pasture	Very low for cropland and pasture	Medium for cropland and very low for pasture	Medium for cropland and pasture
Air pollution	Medium-low	Medium	Medium	Medium-high

Source: From van Vuuren et al. (2011a).

and aerosol precursors; and the concentration of atmosphere constituents (Table 2.3). The four RCPs span a range of radiative forcing values at year 2100 from high to low. (Radiative forcing is the change in the balance between incoming and outgoing radiation at the tropopause due to a change in atmospheric constituents, such as CO_2.) The various radiative forcing pathways are achieved through a range of socioeconomic and technological development scenarios. The RCPs represent a high emission scenario (8.5 W m^{-2}, RCP8.5), two medium stabilization scenarios (6.0 W m^{-2}, RCP6.0; 4.5 W m^{-2}, RCP4.5), and one mitigation scenario leading to low radiative forcing (2.6 W m^{-2}, RCP2.6).

RCP8.5 is a baseline scenario in the absence of climate change policy (Riahi et al. 2011). Increasing greenhouse gas emissions and high greenhouse gas concentrations characterize RCP8.5 (Figure 2.11). Radiative forcing increases to 8.5 W m^{-2} by 2100, with atmospheric CO_2 increasing to 936 ppm, CH_4 to 3751 ppb, and N_2O to 435 ppb. It describes high fossil fuel energy consumption as a result of a large increase in the global population, slow income growth, and modest technological change and energy efficiency improvements in the absence of climate change policies. Global population increases to over 10 billion people in 2050 and to 12 billion people by 2100. Global cropland area increases, primarily in Africa and South America, to meet the growing food demand. Increasing use of fertilizers and intensification of agricultural production give rise to high N_2O emissions. More livestock and rice production result in high CH_4 emissions. Global forestland decreases, but bioenergy use increases, primarily as a result of wood harvest, giving rise to secondary managed forests. Although RCP8.5 depicts an absence of climate mitigation policies, air quality policies greatly affect the depiction of pollutant emissions. Emissions of sulfur dioxide (SO_2), nitrogen oxides (NO_x), black carbon aerosols, and organic carbon aerosols decline due to clean air legislation and technology improvements.

RCP6.0 is a climate policy intervention scenario (Masui et al. 2011). Total radiative forcing stabilizes at 6.0 W m^{-2} after 2100, with atmospheric CO_2 increasing to 670 ppm, CH_4 decreasing somewhat, and N_2O increasing to 406 ppb (Figure 2.11). Population increases to about 10 billion at 2100, and stabilization is achieved through a variety of technologies and strategies designed to reduce greenhouse gas emissions. However, the degree of greenhouse gas emission mitigation is small compared with RCP4.5 and RCP2.6. Global cropland area increases to feed the growing population while forest area remains relatively constant.

RCP4.5 is another intervention scenario in which total radiative forcing stabilizes at 4.5 W m^{-2} after 2100 (Thomson et al. 2011). Atmospheric CO_2 increases to 538 ppm, CH_4 decreases, and N_2O increases to 372 ppb (Figure 2.11). Global population peaks at 9 billion before declining. The necessary reduction in emissions is achieved through imposition of

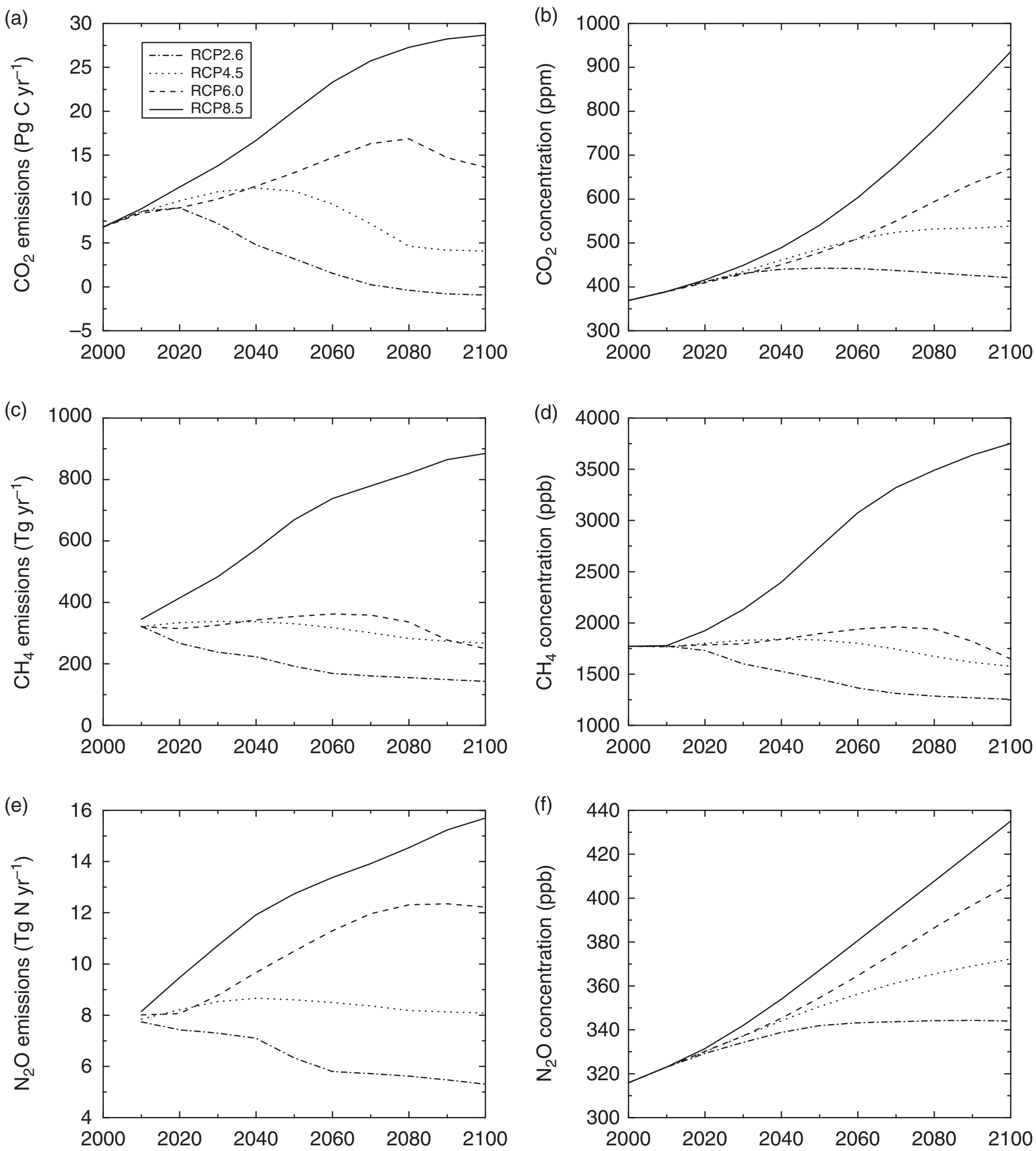

Fig. 2.11 Anthropogenic emissions of (a) fossil fuel CO_2, (c) CH_4, and (e) N_2O for the four RCPs. Also shown are the atmospheric concentrations of (b) CO_2, (d) CH_4, and (f) N_2O. Data from Prather et al. (2013).

climate policies that drive declines in overall energy use and fossil fuel CO_2 emissions with substantial increases in renewable energy and carbon capture systems. Global forest cover increases for carbon storage as part of the overall emissions mitigation strategy. RCP4.5 is the only scenario in which cropland area declines, because of reforestation policies and assumed increases in crop yield.

RCP2.6 is a scenario that leads to very low greenhouse gas concentrations (van Vuuren et al. 2011b). Radiative forcing peaks at ~3 W m^{-2} before 2100 and declines to 2.6 W m^{-2} by 2100 (CO_2, 421 ppm; CH_4, ~30% lower than

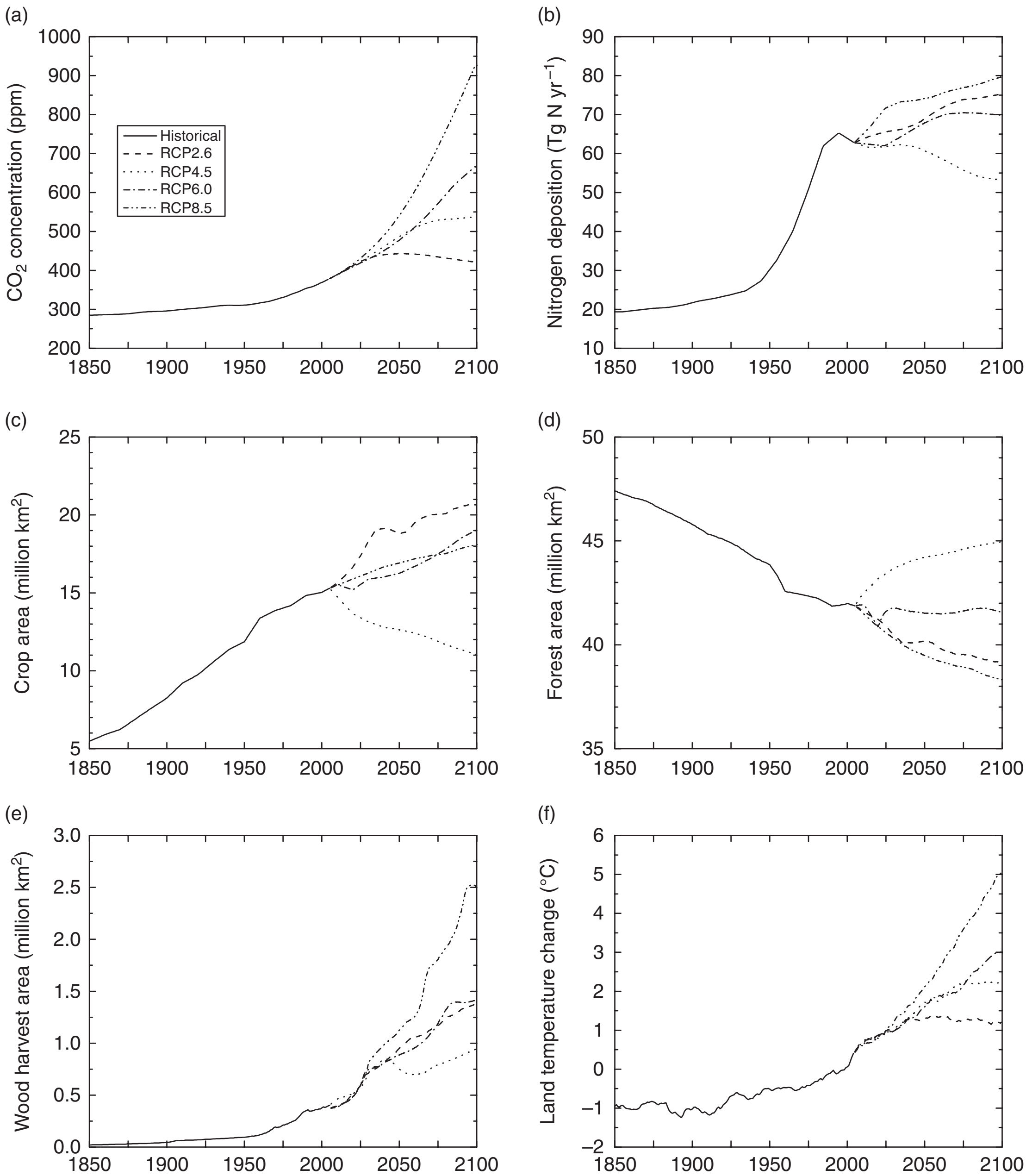

Fig. 2.12 Climate forcings and simulated temperature change in an Earth system model. (a) Atmospheric CO_2 concentration. (b) Annual nitrogen deposition on land. (c) Cropland area. (d) Forest area. (e) Annual wood harvest area. (f) Change in land temperature (relative to 1986–2005). Data from Lawrence et al. (2012).

2000; N_2O, 344 ppb). Greenhouse gas emissions decrease substantially over time in order to reach this low radiative forcing (Figure 2.11), necessitating substantial changes in energy use. The mitigation strategies utilize bioenergy, with consequences for global land use. RCP2.6 has the largest increase in global cropland area.

Figure 2.12 illustrates some of these historical and future forcings used to simulate climate for the twentieth and twenty-first centuries.

RCP8.5 has the largest warming, followed by RCP6.0 and RCP4.5; RCP2.6 has the smallest warming. A particularly important difference among these is the depiction of future land use (Hurtt et al. 2011).

2.8 Terrestrial Feedbacks

Numerous physical, chemical, and biological processes within the atmosphere, ocean, and land regulate Earth's climate, and many of these processes feed back to accentuate or mitigate climate change. A positive feedback accentuates a perturbation within the system while a negative feedback mitigates the perturbation. Water vapor is an example of a positive feedback. As temperature increases, the amount of water vapor that can be held in air increases. Because water vapor is a powerful greenhouse gas, any increase in water vapor in the atmosphere as a result of warming will feed back to accentuate the warming. Clouds are another important feedback because they increase planetary albedo (cooling climate) and reduce longwave radiation lost to space (warming climate). The effect depends on the type of cloud.

The amount of snow and ice covering the surface is a key feedback. Sea ice reflects more solar radiation than open water. Glaciers and snow reflect more solar radiation than soil or vegetation. By reflecting more solar radiation, sea ice, glaciers, and snow cool climate because less radiation is available to heat the surface. Consequently, growth of ice and snow with a colder glacial climate reinforces the cold climate. Conversely, climate warming melts sea ice, glaciers, and snow. This reinforces the warming, because the darker surface absorbs more solar radiation.

Many other climate feedbacks relate to processes that occur on land. Particularly prominent biogeophysical feedbacks relate to albedo, surface roughness, and leaf area (Figure 1.6). Others relate to the hydrologic cycle and biogeochemical cycles. A greater discharge of freshwater into the North Atlantic shuts down the thermohaline circulation. The melting of glaciers provides much of this freshwater but also influences climate through the ice albedo feedback. The geological and biological carbon cycles regulate atmospheric CO_2, and terrestrial ecosystems regulate CH_4 and N_2O concentrations as well. High dust emissions during glacial times may deposit more iron in oceans, which may stimulate phytoplankton growth, reduce atmospheric CO_2, and cool climate. Changes in the geographic distribution of vegetation alter biogeochemical cycles, the hydrologic cycle, and surface energy fluxes. Such feedbacks can be observed in the geological record. In addition, it is widely accepted that glacial–interglacial cycles are a response to orbital forcing, but terrestrial feedbacks within the climate system amplify the response to this forcing. The Arctic is thought to be particularly sensitive to climate change due to numerous feedbacks among climate, the hydrologic cycle, the biosphere, and the cryosphere.

2.9 Review Questions

1. A change in atmospheric CO_2 concentration of 1 part per million (ppm) is equivalent to about 2.1 Pg C. How much carbon has accumulated in the atmosphere since the preindustrial era?

2. The Amazon basin of South America has a tropical rainforest climate. What type of climate would be expected in the future if annual precipitation decreased and if the seasonality of rainfall increased to distinct wet and dry seasons?

3. By how much would the average ocean depth increase if all of the freshwater in glaciers was put into oceans? Use an ocean area of 360,000,000 km^2. The Greenland ice sheet contains about 2,900,000 km^3 of frozen water. By how much would sea level rise if all this water melted into oceans?

4. What percentage of the water on Earth is available for human uses?

5. Arctic sea ice is widely expected to decrease in extent with a warmer climate. Why should we care?

6. What type of vegetation would be favored under the climate change scenario of question 2? What type of vegetation would be reduced in extent?

7. By how much would atmospheric CO_2 concentration increase if the biosphere stored no carbon in plants and soil? Assume none of the carbon is taken up by oceans.

8. The Sahel region of Africa lies between the Sahara desert to the north and tropical rainforest to the south. How has human land use altered the vegetation of this region? Use Figure 2.8 for reference.

9. RCP8.5 is considered a business-as-usual scenario in the absence of changes in global energy use. The high energy demand and fossil fuel use in RCP8.5 implies that achieving climate stabilization will require a substantial reduction of emissions and change in energy use. Discuss ways in which this transformation can be achieved.

10. RCP2.0 is designed to limit global temperature warming to about 2°C. It requires significant reduction in CO_2 emissions. Are these reductions feasible?

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3

Global Cycles

3.1 Chapter Summary

The functioning of Earth as a system is seen in the global cycling of energy, water, and carbon, and in other biogeochemical cycles. This chapter introduces the fundamental scientific concepts of energy, water, and biogeochemical cycles that regulate climate and link the atmosphere, hydrosphere, cryosphere, biosphere, pedosphere, and anthroposphere. Heat flows between materials due to temperature differences. Heat is transferred in the atmosphere by radiation, conduction, and convection. These flows of heat determine the balance of energy gained, lost, or stored. For the planet as a whole and averaged over the year, the solar radiation absorbed by Earth is equal to the longwave radiation emitted to space. That is, the net radiation absorbed by Earth is zero in the annually averaged planetary mean. The hydrologic cycle describes the cycling of water among land, ocean, and air, principally in terms of precipitation, evaporation, and the runoff of freshwater from land into oceans. The hydrologic cycle regulates the amount of water vapor in the air, which is a key greenhouse gas. The increased capacity of air to hold moisture as temperature increases is an important thermodynamic principle that affects climate. In addition, the change of water among its solid, liquid, and vapor states requires considerable energy. These phase changes provide energy to fuel storms. Atmospheric gases interact with radiant energy flowing through the atmosphere to determine the planetary energy budget. Principal among these are carbon dioxide (CO_2), methane (CH_4), and nitrous oxide (N_2O). These gases cycle among the atmosphere, ocean, and land, regulated in part by biological and geochemical processes. Human activities modify their natural cycles, which can be seen in the rising concentration of these greenhouse gases in the atmosphere.

3.2 Scientific Units

All units of measurement are derived from four basic units (Table 3.1): length, measured in meters (m); mass, measured in kilograms (kg); time, measured in seconds (s); and temperature, measured in kelvin (K). An additional quantity, mole (mol), is used in chemistry to measure the amount of a substance. Mass and moles are related by the molecular mass of the material (kg mol^{-1}). Force is a quantity that accelerates an object (force = mass × acceleration). The scientific unit for force is the newton (N). One newton is defined as the force needed to accelerate a mass of one kilogram to a speed of one meter per second in one second (1 N = 1 kg m s^{-2}). Work is done when a force acts on an object to move it over a certain distance (work = force × distance). The scientific unit for work is the joule (J). One joule is the work needed to move an object with the force of one newton over a distance of one meter in the direction of the

Table 3.1 Basic and derived scientific units

Quantity name	Unit name	Dimension symbol	Unit symbol	Base units
Length	meter	L	m	–
Mass	kilogram	M	kg	–
Time	second	T	s	–
Temperature	kelvin	K	K	–
Amount	mole	–	mol	–
Area	square meter	L^2	m^2	m^2
Volume	cubic meter	L^3	m^3	m^3
Density	kilogram per cubic meter	$M\ L^{-3}$	$kg\ m^{-3}$	$kg\ m^{-3}$
Velocity	meter per second	$L\ T^{-1}$	$m\ s^{-1}$	$m\ s^{-1}$
Acceleration	meter per second per second	$L\ T^{-2}$	$m\ s^{-2}$	$m\ s^{-2}$
Force	newton	$M\ L\ T^{-2}$	N	$kg\ m\ s^{-2}$
Energy	joule	$M\ L^2\ T^{-2}$	J	$kg\ m^2\ s^{-2} = N\ m$
Power	watt	$M\ L^2\ T^{-3}$	W	$kg\ m^2\ s^{-3} = J\ s^{-1}$
Pressure	pascal	$M\ L^{-1}\ T^{-2}$	Pa	$kg\ m^{-1}\ s^{-2} = N\ m^{-2}$

Note: °C = K – 273.15 so that a change in temperature of 1°C = 1 K.

Table 3.2 Metric prefixes

Multiple	Prefix	Symbol	Multiple	Prefix	Symbol
10^{-1}	deci-	d	10^{1}	deca-	da
10^{-2}	centi-	c	10^{2}	hecto-	h
10^{-3}	milli-	m	10^{3}	kilo-	k
10^{-6}	micro-	μ	10^{6}	mega-	M
10^{-9}	nano-	n	10^{9}	giga-	G
10^{-12}	pico-	p	10^{12}	tera-	T
10^{-15}	femto-	f	10^{15}	peta-	P

force (1 J = 1 N m = 1 kg $m^2\ s^{-2}$). Energy is the capacity to do work and has the same units as work. One joule of energy supports 1 N m of work. Power is defined as the rate at which work is done. The scientific unit for power is the watt (W), which is equal to the rate of working one joule per second (1 W = 1 J s^{-1}). Pressure is the force per unit area. The scientific unit for pressure is the pascal (Pa), equal to a force of one newton over an area of one square meter (1 Pa = 1 N m^{-2} = 1 kg $m^{-1}\ s^{-2}$).

These units of measure can have a prefix that indicates multiples or fractions of the units. For example, the prefix kilo- denotes multiplication by one thousand; one kilometer equals one thousand meters (1 km = 1000 m). The prefix milli- denotes division by one thousand; one millimeter equals one-thousandth of a meter (or conversely, 1000 mm = 1 m). Other common prefixes are: micro- (μ, 10^{-6}); hecto- (h, 10^2); mega- (M, 10^6); tera- (T, 10^{12}), and peta- (P, 10^{15}). Table 3.2 lists common scientific prefixes.

3.3 Energy Fluxes

Energy is the ability to do work. It exists in a variety of forms, but there are two basic categories. Potential energy is stored energy that results from an object's position. It is the work that must be done to move an object from some reference point to another position. For example, a ball at rest on the top of a hill contains gravitational potential energy. A stretched spring has elastic potential energy arising from its deformation. Kinetic energy is the energy of motion. It is the energy an object possesses because of its motion. Kinetic energy is a measure of the amount of work an object in motion can do as a result of its motion. A moving automobile will do work as it hits another vehicle. A ball thrown at a window will do work as it strikes the window.

Temperature is a measure of the energy of motion, or kinetic energy, in the movement of molecules in a substance. All materials are composed of molecules. These molecules are in motion. This motion is most evident for gases or liquids, but the molecules in solids are also in motion through vibrations. As the motion of molecules increases, kinetic energy increases and temperature rises. Temperature is perceived in terms of the relative warmth or coolness of an object.

Heat is a form of kinetic energy that flows from one object to another due to temperature differences between them. In this way, energy transfers from hot to cold materials. In the atmosphere, heat is transferred by radiation, conduction, and convection.

3.3.1 Radiation

All materials with temperature greater than absolute zero (0 K, or –273.15°C) emit energy in the form of electromagnetic radiation. We perceive this energy principally as visible light and in the warmth of the Sun's rays. Terrestrial objects also emit radiation but at wavelengths that are longer than the Sun's and that are not visible to the eye. At lower temperatures, we do not sense this energy because of its low amount. At higher temperatures, such as an electric heater, the radiant energy increases and we feel its warmth. At very high temperatures (1000 K), such as a wood fire, the radiant energy is strong, the emitted radiation shifts to shorter wavelengths in the visible spectrum, and objects glow. Electromagnetic radiation is called radiant energy and has the unit joules (J). Radiant energy emitted or received per unit time is called the radiant flux and has the unit of joules per second, or watts (1 W = 1 J s^{-1}). Radiant flux density is the radiant flux per unit surface area (W m^{-2}). Irradiance is the radiant flux density incident on an object; emittance is that emitted by an object.

Radiant energy travels in waves with peaks and troughs. The distance between successive peaks (or troughs) is the wavelength. Electromagnetic radiation transfers energy in discrete units called quanta or photons. The energy of a photon (e_p, J) is related to wavelength (Λ, m) by:

$$e_p = hc / \Lambda \tag{3.1}$$

where $h = 6.626 \times 10^{-34}$ J s is Planck's constant and $c = 3 \times 10^8$ m s^{-1} is the speed of light. The longer the wavelength, the lesser the energy of the photon. The energy of a photon of blue light with wavelength 0.450 μm is 4.42×10^{-19} J. A photon of red light (0.680 μm) has 2.92×10^{-19} J.

Planck's law defines the radiant flux density per unit wavelength ($E(\Lambda)$, W m^{-2} m^{-1}) emitted by a blackbody in relation to wavelength and temperature:

$$E(\Lambda) = \frac{2\pi hc^2}{\Lambda^5\left[\exp(hc / k\Lambda T) - 1\right]} \tag{3.2}$$

where $k = 1.38 \times 10^{-23}$ J K^{-1} is the Boltzmann constant and T is temperature (K). Figure 3.1 illustrates this for two bodies with the approximate temperatures of the Sun and Earth. It is readily evident that objects with a higher temperature emit radiation at a greater rate than objects with a lower temperature. It is also evident from Figure 3.1 that the spectral distribution of radiant energy depends on the temperature of the object. The higher the object's temperature, the shorter the wavelength of emitted radiation. The Sun, with a temperature of 6000 K, emits radiation in short wavelengths between 0.2 and

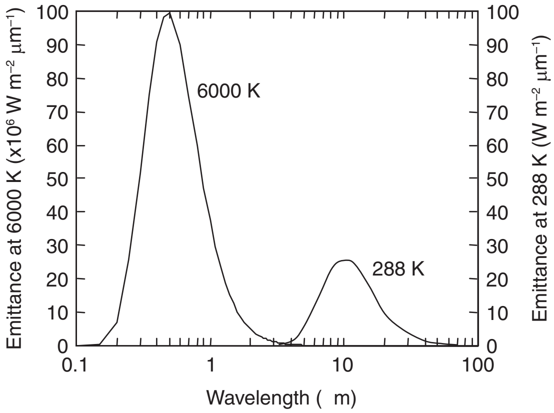

Fig. 3.1 Spectral distribution of radiation emitted by blackbodies at temperatures of 6000 K (Sun, left-hand axis) and 288 K (Earth, right-hand axis). Note that the left-hand axis is larger than the right-hand axis by the factor 10^6.

4 µm. Radiation in these wavelengths is known as solar, or shortwave, radiation. Solar radiation is divided into ultraviolet radiation with a wavelength of less than about 0.4 µm (containing 10% of the Sun's energy), visible radiation between about 0.4 and 0.7 µm (40% of the Sun's energy), and near-infrared radiation at wavelengths of greater than about 0.7 µm (50% of the Sun's energy). Visible radiation is further divided into violet, blue, green, yellow, orange, and red (in order of increasing wavelength). Earth, with an effective temperature of about 288 K, emits less radiation than the Sun and in longer wavelengths from about 3 to 100 µm. Radiation at these high wavelengths is called infrared, or longwave, radiation.

The wavelength at which maximum emission occurs similarly decreases with higher temperature. Maximum emission for the Sun occurs at a wavelength of about 0.5 µm. For Earth, peak emission occurs at a wavelength of 10 µm. Wien's displacement law relates the wavelength of maximum emission to temperature:

$$\Lambda_{max} = 2897\ \mu\text{m K} / T \tag{3.3}$$

This is evident when watching wood burn in a fireplace. As the fire burns out and grows cold, the embers turn from violet and blue when very hot, and more commonly yellow, to orange and then red while continuing to give off heat. The radiation spectrum shifts from shorter wavelengths when the embers are hot to longer wavelengths as they cool.

The Stefan–Boltzmann law relates the radiant flux density emitted by an object to its temperature, obtained by integrating $E(\Lambda)$ over all wavelengths:

$$E = \varepsilon\sigma T^4 \tag{3.4}$$

where E is emittance (W m^{-2}), $\sigma = 5.67 \times 10^{-8}$ W m^{-2} K^{-4} is the Stefan–Boltzmann constant, and ε is the broadband emissivity. This equation is formulated for a blackbody, which is an object that is a perfect absorber of radiation at all wavelengths and that emits the maximum possible energy at all wavelengths for a given temperature. For a blackbody, $\varepsilon = 1$. Most objects are not blackbodies and emit less radiation. Instead, the Stefan–Boltzmann law describes emission of radiation, but blackbody emittance is reduced by the object's emissivity (ε). Emissivity is defined as the ratio of the actual emittance to the blackbody emittance. Most objects have a broadband emissivity of 0.95–0.98 when integrated over all wavelengths.

An object absorbs radiant energy from the Sun. The radiation incident on an object is absorbed, reflected, or transmitted through the material so that:

$$a_\Lambda + r_\Lambda + t_\Lambda = 1 \tag{3.5}$$

Here, a_Λ is the fraction of incident radiation at a specified wavelength that is absorbed (absorptivity, or absorptance); r_Λ is the fraction of incident radiation at a specified wavelength that is reflected (reflectivity, or reflectance); and t_Λ is the fraction of incident radiation at a specified wavelength that is transmitted (transmissivity, or transmittance). These optical properties can vary strongly with wavelength. Green leaves, for example, typically absorb 85 percent of the solar radiation in the visible waveband (a_{vis} = 0.85) but absorb less than 50 percent of the radiation in the near-infrared waveband (a_{nir} < 0.50). Objects that are sufficiently thick are opaque (t_Λ = 0), and absorptivity is $a_\Lambda = 1 - r_\Lambda$. The solar radiation absorbed by an object is commonly calculated from this relationship.

Terrestrial bodies also absorb longwave radiation. Equation (3.5) describes absorptance, reflectance, and transmittance of this radiation. Additionally, Kirchhoff's law states that the absorptivity of a material is equal to its emissivity:

$$a_\Lambda = \varepsilon_\Lambda \tag{3.6}$$

A blackbody is a perfect absorber and emitter of radiation so that $a_\Lambda = \varepsilon_\Lambda = 1$ and $r_\Lambda = t_\Lambda = 0$. An opaque gray body ($t_\Lambda = 0$) has reflectance $r_\Lambda = 1 - a_\Lambda = 1 - \varepsilon_\Lambda$.

A body with a broadband reflectance r (integrated over all wavelengths) absorbs a fraction of the incoming solar radiation given by $1 - r$. The longwave radiative heat transfer between the body and its surrounding environment is the balance between the energy radiated by the body and that absorbed from the environment. The body absorbs a fraction ε (integrated over all wavelengths) of the energy radiated by its surrounding environment and reflects a fraction $1 - \varepsilon$. The net radiation (R_n, W m^{-2}) absorbed is:

$$R_n = (1 - r)S\downarrow + \varepsilon L\downarrow - \varepsilon\sigma T^4 \tag{3.7}$$

where $S\downarrow$ is solar radiation (W m^{-2}) and $L\downarrow$ is the radiant energy from the environment (W m^{-2}).

For photosynthesis, the number of photons, not energy, is important. A photon of light with blue wavelength has more energy than a photon of light with red wavelength, but both have the same effect on photosynthesis. Only radiation with wavelengths between 0.4 and 0.7 μm, known also as photosynthetically active radiation, is used during photosynthesis. The energy of a mole of photons is obtained by multiplying the energy per photon by Avogadro's number (6.022×10^{23} mol^{-1}). A mole of photons with wavelength 0.55 μm has the energy:

$$\frac{(6.626 \times 10^{-34}\ \text{J s})\,(3 \times 10^{8}\ \text{m s}^{-1})}{0.55 \times 10^{-6}\ \text{m}}(6.022 \times 10^{23}\ \text{mol}^{-1})$$
$$= 0.218 \times 10^{6}\ \text{J mol}^{-1}$$

Therefore, photosynthetically active radiation (with average wavelength 0.55 μm) is converted from W m^{-2} (J s^{-1} m^{-2}) to photosynthetic photon flux density with units μmol photon m^{-2} s^{-1} using the factor 4.6 μmol J^{-1}.

3.3.2 Conduction

Conduction is the transfer of heat within a material or between materials arising from molecular vibration without any motion of the material itself. Consider, for example, a metal spoon placed in a pot of hot water. The end of the spoon not in the water becomes hot. As the molecules in the part of the spoon placed in the water absorb heat from the water, they vibrate faster. These molecules cause adjacent molecules to also vibrate faster. This process repeats until all the molecules in the spoon vibrate rapidly. If a person touches the spoon, heat flows from the hot spoon to the skin, causing the molecules in the person's skin to vibrate faster. The person perceives that the spoon is hot. In this way, heat is conducted from the water to the spoon to the skin, flowing from high temperature to low temperature.

The rate of heat transfer in one dimension by conduction (Q, W m^{-2}) is:

$$Q = -\kappa(\Delta T / \Delta z) \tag{3.8}$$

where κ is the thermal conductivity of the material (W m^{-1} K^{-1}) and $\Delta T / \Delta z$ is the temperature gradient (°C m^{-1}, or K m^{-1}). The rate of heat flow between two points separated by some distance (Δz) is proportional to the temperature difference between the points (ΔT). Thermal conductivity determines the rate of heat transfer for a unit temperature gradient. The negative sign denotes that the flux is positive for a negative temperature gradient.

The type of material affects the rate of heat transfer. Metals are good conductors of heat and have a high thermal conductivity. Wood is a poor conductor and has low thermal conductivity. A metal spoon placed in hot water feels hotter to the touch than does a wooden spoon in the same water because it conducts heat to the hand much more rapidly than the wooden spoon. Materials with low thermal conductivity reduce heat loss by conduction and are effective insulators. Styrofoam is a poor conductor of heat. Air is also a very poor conductor of heat. Double-paned glass windows with an inner layer of air are a very effective insulator.

The thermal conductivity of air, for example, is 0.02 W m^{-1} K^{-1}. A 5°C temperature difference between a body and overlying air over a distance of 10 mm transfers 10 W m^{-2} by conduction. The thermal conductivity of water is 0.57 W m^{-1} K^{-1} and produces a heat flux that is almost thirty times larger (285 W m^{-2}). This is why a body immersed in cold water loses heat rapidly.

3.3.3 Convection

Diffusion is the transport and mixing of heat and mass through the movement of a gas or fluid. Such mixing occurs along a gradient from high to low concentration. The vertical transport and mixing of heat and mass through the movement of air is a form of diffusion termed convection. In the atmospheric sciences, convection refers to vertical motion in the atmosphere. The horizontal transport of heat is termed advection.

Fick's law describes the rate of mass transfer per unit area of a gas (e.g., H_2O, CO_2) in one dimension along a gradient from high to low concentration. The diffusive flux (F_j, kg m^{-2} s^{-1}) between two points relates to the concentration difference ($\Delta\rho_j$, kg m^{-3}) multiplied by a conductance (g'_j, m s^{-1}) or divided by a resistance (r'_j, s m^{-1}):

$$F_j = \Delta\rho_j g'_j = \Delta\rho_j / r'_j \tag{3.9}$$

The conductance accounts for molecular or turbulent motions that mix the fluid.

For diffusive flux calculations, the concentration of a gas (ρ_j) is defined as its mass (m_j, kg) per unit volume of mixture (V, m^3). The ideal gas law describes the volume occupied by a gas. It relates the volume occupied by n moles of a gas at a given pressure (P, Pa = N m^{-2} = J m^{-3}) and temperature (T, K) as:

$$PV = n\Re T \tag{3.10}$$

where $\Re$ is the universal gas constant (8.314 J K^{-1} mol^{-1}). For example, one mole of air occupies a volume $V = 0.0236$ m^3 for a standard atmosphere at sea level. (The standard atmosphere defines sea level as T = 288.15 K (15°C) and P = 1013.25 hPa, which is denoted STP for standard temperature and pressure.) The inverse (the number of moles per unit volume) is termed molar density (ρ_m, mol m^{-3}), and:

$$\rho_m = \frac{n}{V} = \frac{P}{\Re T} \tag{3.11}$$

Molar density at a given pressure and temperature is constant for all gases and equals 42.3 mol m^{-3} at STP. The density of a gas at a given pressure and temperature is equal to its mass divided by the volume occupied by the gas. An equivalent form of the ideal gas law is:

$$\text{density} = \frac{m_j}{V} = \frac{nM_j}{V} = \frac{P}{\Re T}M_j = \rho_m M_j \tag{3.12}$$

where M_j (kg mol^{-1}) is the molecular mass. For example, dry air (molecular mass, 28.97 g mol^{-1}) has a density of 1.225 kg m^{-3} at STP. If pressure remains constant, any increase in temperature results in a decrease in density. A decrease in temperature results in an increase in density at the same pressure.

Air is comprised of N_2, O_2, Ar, H_2O, CO_2, and other gases in trace amounts (Table 2.1). Each individual gas follows the ideal gas law, and:

$$\rho_j = \frac{m_j}{V} = \frac{n_j M_j}{V} = \frac{P_j}{\Re T}M_j \tag{3.13}$$

with m_j the mass, n_j the number of moles, and M_j the molecular mass of the gas. Here, ρ_j is the mass concentration of the gas and P_j is the partial pressure of the gas. Partial pressure is the pressure that a gas would exert if it alone occupied the same volume as the mixture and at the same temperature.

Volume changes with temperature and pressure so that changes in mass concentration (ρ_j) can occur independent of changes in mass. An alternative measure of concentration, independent of volume, is the mole fraction (c_j, mol mol^{-1}). This is the number of moles of a gas (n_j) in a given volume expressed as a fraction of the total number of moles (n) in the same volume:

$$c_j = \frac{n_j}{n} = \frac{P_j}{P} = \frac{\rho_j}{M_j}\frac{\Re T}{P} = \frac{\rho_j}{M_j\rho_m} \tag{3.14}$$

For example, CO_2 (molecular mass, 44.01 g mol^{-1}) with mole fraction c_j = 390 µmol mol^{-1} in the atmosphere at pressure 1013.25 hPa has partial pressure P_j = 39.5 Pa and mass concentration ρ_j = 0.73 g m^{-3} at 15°C. Inserting Eq. (3.14) into Eq. (3.9), an equivalent form of the diffusive flux equation is:

$$F_j / M_j = \rho_m \Delta c_j g'_j \tag{3.15}$$

Dividing the mass flux (F_j) with units kg m^{-2} s^{-1} by molecular mass (M_j) gives the molar flux with units mol m^{-2} s^{-1}.

The principles of diffusion apply to moisture and heat transfer in the atmosphere. With e the partial pressure of water vapor (Pa), the diffusive flux equation for evaporation (E, kg m^{-2} s^{-1}) is:

$$E / M_w = \rho_m \frac{\Delta e}{P} g'_w \tag{3.16}$$

with M_w the molecular mass of water (18.02 g mol^{-1}) and g'_w the conductance for water vapor (m s^{-1}). Heat transfer by convection (H, W m^{-2}) is analogous to diffusion. The term $\rho c'_p \Delta T$ replaces $\Delta\rho_j$ in Eq. (3.9), and:

$$H = \rho c'_p \Delta T\, g'_h \tag{3.17}$$

with ρ the density of moist air (kg m^{-3}), c'_p the specific heat of moist air at constant pressure (J kg^{-1} K^{-1}), ΔT (K) the temperature difference, and g'_h the conductance for heat (m s^{-1}). The density of air varies with temperature, pressure, and vapor pressure. A representative value is ρ = 1.22 kg m^{-3} at sea level (1013.25 hPa) and 15°C. The specific heat of air also varies with humidity. A representative value is c'_p = 1010 J kg^{-1} K^{-1}.

The meteorological community expresses diffusive fluxes in terms of a mass flux (F_j) with units kg m^{-2} s^{-1} and conductance (g'_j) with units m s^{-1}, or resistance r'_j (s m^{-1}). Molar units are common in the plant physiological literature, where conductance is preferred because it is directly proportional to the flux and because conductance and flux have the same units. A conductance g'_j with units m s^{-1} is converted to g_j with units mol m^{-2} s^{-1} by multiplying by the molar density (ρ_m) with units mol m^{-3}, given by Eq. (3.11). At STP, 1 mol = 0.0236 m^3 and 1 m s^{-1} = 42.3 mol m^{-2} s^{-1}. The conversion for conductance is:

$$g'_j \left[\frac{\text{m}}{\text{s}}\right] \times \frac{P}{\Re T}\left[\frac{\text{mol}}{\text{m}^3}\right] = g_j \left[\frac{\text{mol}}{\text{m}^2\text{s}}\right] \tag{3.18}$$

Equivalent forms of the diffusive fluxes (Eq. (3.15), Eq. (3.16), and Eq. (3.17)) are:

$$F_j / M_j = \Delta c_j g_j \tag{3.19}$$

$$E / M_w = \frac{\Delta e}{P} g_w \tag{3.20}$$

$$H = c_p \Delta T\, g_h \tag{3.21}$$

In Eq. (3.19) and Eq. (3.20), the fluxes (F_j / M_j and E / M_w) and conductances (g_j and g_w) have units mol m^{-2} s^{-1}. Equation (3.21) is the corresponding form of the convective heat flux, with c_p the molar specific heat of moist air at constant pressure (J mol^{-1} K^{-1}) and g_h the molar conductance for heat (mol m^{-2} s^{-1}). A representative value is c_p = 29.2 J mol^{-1} K^{-1}. Molar units and conductances are used in this book, as in Cowan (1977) and Campbell and Norman (1998).

Two types of convection are distinguished in meteorological studies. Free convection occurs due to temperature differences that affect the density, and therefore buoyancy, of air. One means by which convection occurs in the atmosphere is that warm air is less dense than cold air. Warm air, therefore, tends to rise in the atmosphere while cold air sinks. In the process, heat transfers from a warm surface to the colder air above. This transport of heat by vertical motion is called sensible heat because it is heat we can feel. A common example is the warmth felt as warm air rises from a radiator. Convective activity is readily seen in many regions of the world on a hot summer day. Solar radiation heats air near the ground. The warm, moist air rises, where it cools with greater height. Cold air holds less moisture than warm air. As the parcel of air rises and cools, it becomes saturated with moisture. Water vapor condenses, and cumulus clouds form. Forced convection is transport caused by wind. Wind moving across a warm body carries away heat to the cooler air. This is why a breeze is refreshing on a hot summer day. Thick clothes increase the resistance to convective heat loss, diminishing the loss, which is why short-sleeved shirts and shorts are comfortable on hot days. The effect of free and forced convection on mass and heat transport in the atmosphere is represented by the conductances g_j, g_w, and g_h.

3.3.4 Heat Storage

Heat capacity is the amount of energy needed to raise the temperature of a unit volume of material by one degree. A volume with unit area stores the energy:

$$\Delta Q = c_v(\Delta T / \Delta t)\Delta z \qquad (3.22)$$

where ΔQ is the heat absorbed by the volume (W m^{-2}), ΔT is the change in temperature (°C, or K) over the time period Δt (s), Δz is thickness (m), and c_v is heat capacity (J m^{-3} K^{-1}). The heat capacity of water (4.18 MJ m^{-3} K^{-1}) is about twice that of soil. Water has to absorb considerably more energy than soil to warm one degree. The presence of large bodies of waters such as lakes or oceans, therefore, acts to modulate the surrounding climate.

The first law of thermodynamics describes the conservation of energy. It states that in a closed system energy can change from one form to another but it cannot be created or destroyed; the total amount of energy in the system is conserved. Consider, for example, a system in which energy input is balanced by energy output and change in stored energy:

$$\text{energy input} = \text{energy output} + \Delta Q \qquad (3.23)$$

If the system loses the same amount of energy that it gains, there can be no change in storage ($\Delta Q = 0$) and the temperature of the system remains constant ($\Delta T = 0$). If the system gains more energy than it loses, the excess energy is stored in the system as thermal energy, raising the temperature of the system. Conversely, the temperature of the system decreases if it loses more energy than it gains.

3.3.5 Planetary Energy Balance

The principle of energy conservation can be seen in the planetary energy balance (Figure 3.2). Annually, Earth receives approximately 341 W m^{-2} of solar radiation at the top of the atmosphere. Clouds, gases, and aerosols absorb 78 W m^{-2} (23%), and Earth's surface absorbs an additional 161 W m^{-2} (47%). The remainder, 102 W m^{-2} (30%), is reflected back to space by the atmosphere and surface. The absorbed solar radiation warms Earth, which emits longwave radiation. At the surface, Earth emits 396 W m^{-2} of longwave radiation. Clouds, water vapor, CO_2, and other gases in the atmosphere absorb most of this radiation (356 W m^{-2}); only 40 W m^{-2} escapes to space. The gases, particles, and other material suspended in the atmosphere emit longwave radiation. This radiation travels in all directions with some lost to space and some reaching the surface. A total of 199 W

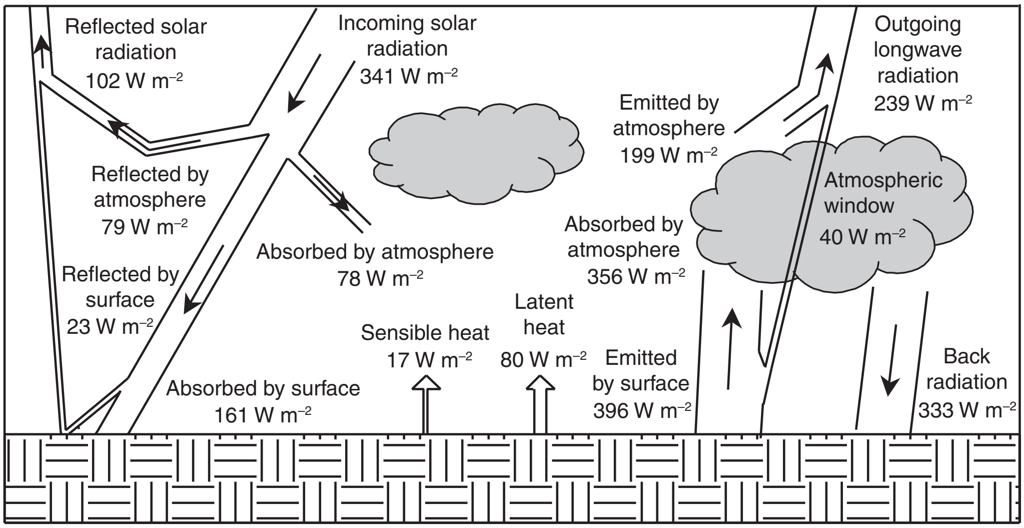

Fig. 3.2 Earth's annual mean global energy budget showing solar radiation (left), sensible and latent heat fluxes (middle), and longwave radiation (right). Data from Trenberth et al. (2009).

m^{-2} escapes to space, which together with the 40 W m^{-2} from the surface balances the 239 W m^{-2} solar radiation absorbed by the atmosphere and surface. That is, the net radiative balance of Earth is zero; the absorbed solar radiation equals the longwave radiation emitted to space.

The fluxes of sensible heat (17 W m^{-2}) and latent heat (80 W m^{-2}), while small compared with radiative fluxes, are important terms in the planetary energy balance (Figure 3.2). Earth's energy budget shows the atmosphere has a deficit of energy while the surface has a surplus. The atmosphere absorbs 78 W m^{-2} of solar radiation and 356 W m^{-2} of longwave radiation from the surface; it emits 199 W m^{-2} of longwave radiation to space and 333 W m^{-2} to the surface. The excess loss of radiation compared with absorption is –98 W m^{-2}. Earth's surface, in contrast, gains 161 W m^{-2} of solar radiation and 333 W m^{-2} of longwave radiation from the atmosphere while emitting 396 W m^{-2} of longwave radiation. This gives the surface a surplus of 98 W m^{-2}. This surplus energy is returned to the atmosphere as sensible heat and latent heat. These fluxes arise as winds carry heat (sensible heat) and moisture (latent heat) away from the surface.

A simple planetary energy balance model provides a descriptor of Earth's temperature and the role of greenhouse gases to warm the surface. Solar radiation heats the planet, and longwave radiation emitted to space cools the planet. This radiative balance determines the mean planetary temperature. The energy balance at the top of the atmosphere (F, W m^{-2}) is:

$$F = \frac{S_c}{4}(1-r) - \sigma T_s^4 = 0 \tag{3.24}$$

The first term of this equation is the solar radiation absorbed by the atmosphere and surface. In this equation, S_c = 1364 W m^{-2} is the amount of radiation emitted by the Sun. The division by four arises because an area of πy^2 intercepts solar radiation (y is the radius of Earth), and $S_c(1-r)\pi y^2$ is the radiant flux (W) received by Earth; but an area of $4\pi y^2$ (i.e., the surface area of a sphere) emits longwave radiation, and $\sigma T_s^4 4\pi y^2$ is the energy flux (W) emitted by Earth. Hence, $S_c / 4$ = 341 W m^{-2} is the incoming solar radiation at the top of the atmosphere averaged over Earth's surface area. The term r is the planetary albedo, which is the fraction of incoming solar radiation reflected to space (r = 0.30). The second term is the outgoing longwave radiation. The emission of longwave radiation to space is given in terms of a global mean surface temperature (T_s, K). The calculated temperature is T_s = 255 K (–18°C). Earth's temperature is in fact about 288 K (15°C). The difference in temperature (33°C) is due to the presence of the atmosphere and the absorption within the atmosphere by water vapor, CO_2, and other greenhouse gases of longwave radiation emitted by Earth's surface.

The greenhouse effect can be understood by extending the planetary energy balance model to include an atmosphere (Figure 3.3). In this model, the atmosphere is perfectly transparent to solar radiation (the surface absorbs all the radiation) and also absorbs all the longwave radiation from the surface. The atmosphere radiates up to space and down onto the surface so that an equal amount of energy (239 W m^{-2}) is lost in both directions. To maintain thermal equilibrium, the surface must emit longwave radiation to balance the 239 W m^{-2} from the Sun and an additional 239 W m^{-2} from the atmosphere. Its temperature is 303 K (30°C). The warmer surface temperature with an atmosphere arises because of atmospheric longwave radiation onto the surface; the surface has to be warm enough to emit twice as much radiation as without an atmosphere. The calculated temperature is warmer than the actual temperature of Earth (288 K) because the model neglects the complexity of the atmosphere. However, the model illustrates the basic principle of the greenhouse effect. The surface emits longwave radiation to maintain thermal equilibrium. The atmosphere absorbs this energy and emits some to space and some back onto the surface. The reemission of longwave radiation back to the surface is the greenhouse effect that warms the surface.

The model presented in Figure 3.3b also illustrates how a gain in energy in the system warms the surface. If the system is perturbed such that the surface gains an additional amount of

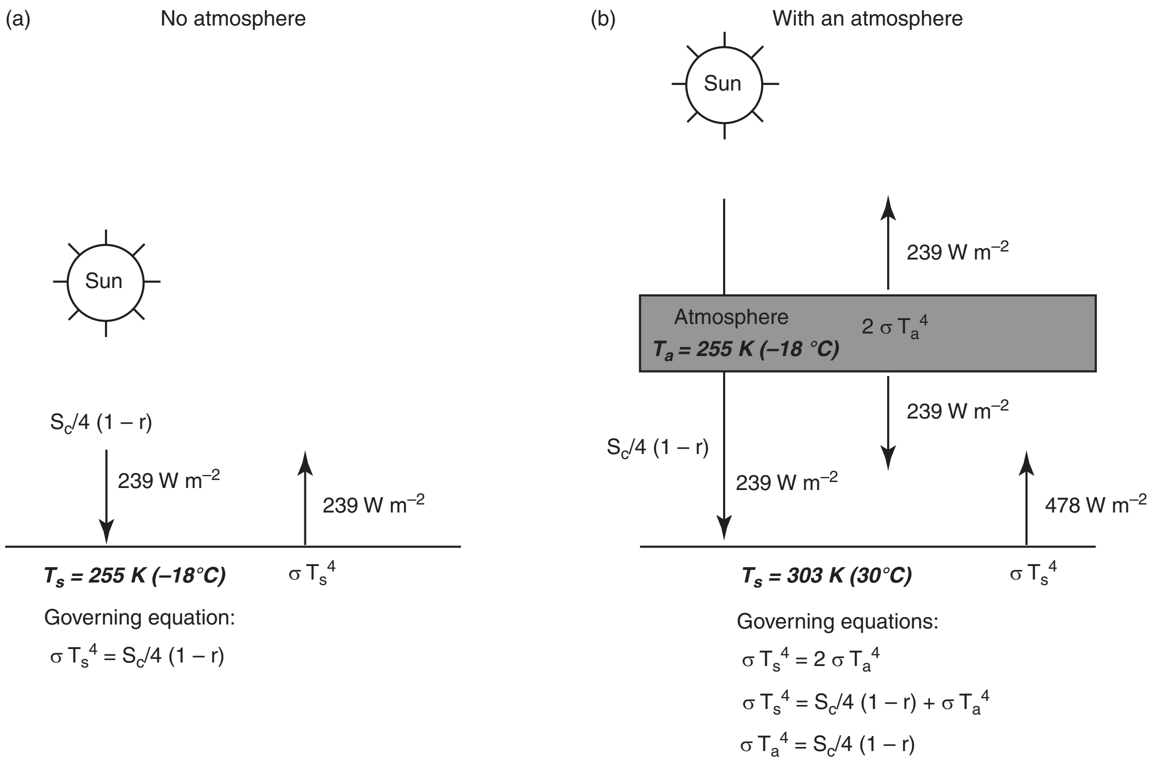

Fig. 3.3 Planetary energy balance and temperature calculation shown (a) without an atmosphere and (b) with an atmosphere.

energy (ΔF), the surface will warm to emit the additional ΔF. One-half of this radiation will be lost to space, and one-half will be emitted back onto the surface. The surface will warm so that it radiates the additional $\Delta F/2$, but the atmosphere returns one-half of this to the surface. The surface must warm still more to emit another $\Delta F/4$, and the process continues until thermal equilibrium is achieved. Of the total energy perturbation to the system (ΔF), an amount equal to $\Delta F + \Delta F/2 + \Delta F/4 + \Delta F/8 + \Delta F/16 + \cdots$ is gained by the surface (equal to $2\Delta F$) and an amount equal to $\Delta F/2 + \Delta F/4 + \Delta F/8 + \Delta F/16 + \cdots$ (i.e., ΔF) is emitted at the top of the atmosphere.

Equation (3.24) helps to explain the change in planetary temperature to a perturbation of energy in the system. Climate sensitivity is defined as the change in temperature for some change in forcing applied to the system. For Eq. (3.24), and neglecting dependences on other climate processes, the climate sensitivity factor is:

$$\frac{\Delta T_s}{\Delta F} = -\left(\frac{\partial F}{\partial T_s}\right)^{-1} = \left(\frac{S_c}{4}\frac{\partial r}{\partial T_s} + 4\sigma T_s^3\right)^{-1} \qquad (3.25)$$

Ignoring the dependence of albedo on temperature, climate sensitivity is 0.27 K (W m^{-2})$^{-1}$ when Earth is treated as a blackbody. A 1 W m^{-2} gain in energy produces about one-quarter of a degree increase in temperature. This measure of climate sensitivity considers only the blackbody emission of longwave radiation. In fact, the climate sensitivity of Earth is two to three times as large. Feedback mechanisms associated with water vapor, clouds, albedo, and other processes amplify climate sensitivity.

3.4 Hydrologic Cycle

The hydrologic cycle describes the cycling of water among land, ocean, and air. Evaporation is the physical process by which liquid water in

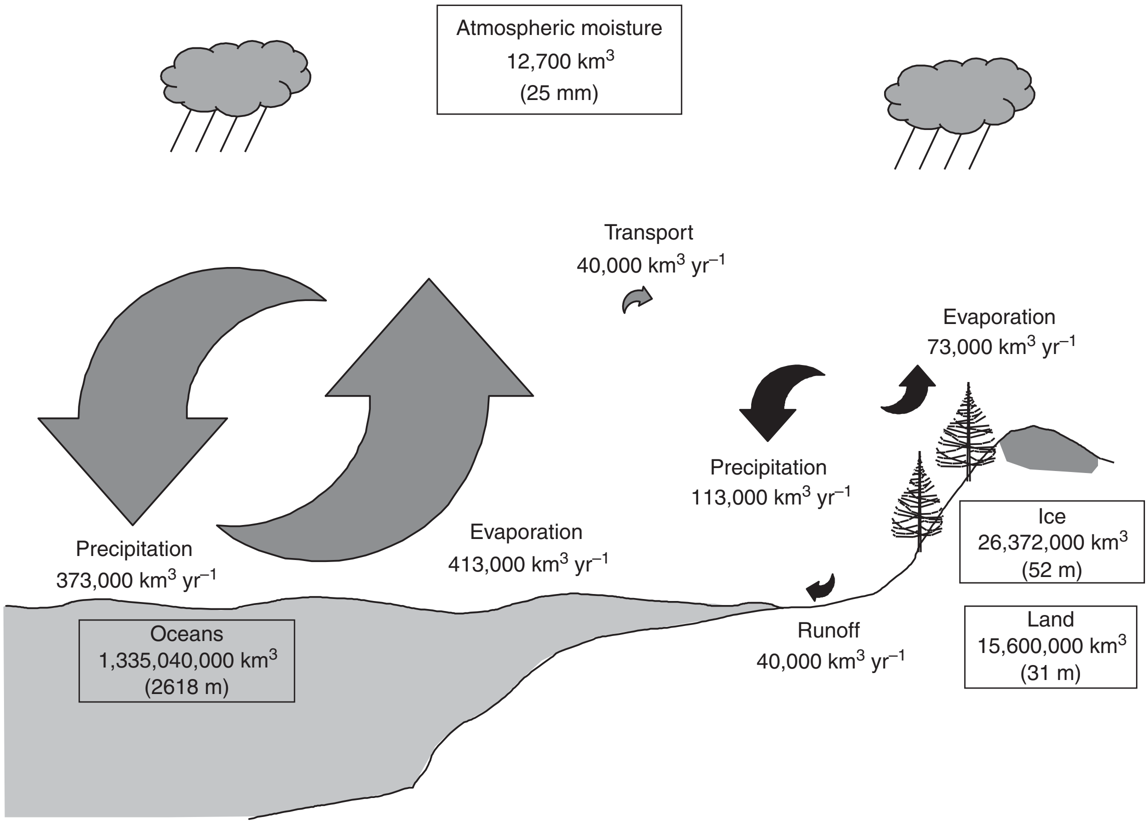

Fig. 3.4 The global hydrologic cycle. Units are km^3 of water or, in parentheses, the depth of water spread over Earth's 510 million km^2 surface area. Data from Trenberth et al. (2007). See also Oki and Kanae (2006).

the oceans or on land changes to vapor in the air. It occurs when unsaturated air comes into contact with a moist surface. Evaporation provides the atmospheric moisture that returns to the surface as rain or snow. Evaporation also consumes an enormous amount of heat, which helps to cool the evaporating surface. Once in the atmosphere, water condenses, forming clouds, and if conditions are right the water falls back to the surface as precipitation. Heat is released as water vapor condenses and changes from vapor to liquid. This heat is a source of energy that drives atmospheric circulation and fuels storms. Oceans are the largest source of water for evaporation. Soils contain less than 1 percent of the unfrozen freshwater on Earth. However, soil water is an important determinant of surface energy fluxes and the climate near the ground. Additionally, the discharge of freshwater from rivers into oceans prevents oceans from becoming saltier, which in turn influences ocean heat transport.

3.4.1 Global Water Balance

Water flows among the oceans, land, and atmosphere (Figure 3.4). The amount of water and its transfers can be measured by volume (m^3) and equivalently by depth (m), mass (kg), or moles (mol) per unit area. These are related by the density of water (1 m^3 H_2O = 1000 kg) and the molecular mass of water (1 mol H_2O = 18.02 g). One kilogram of water spread over an area of one square meter (1 kg m^{-2}) is equivalent to a depth of 1 mm, a volume of 0.001 m^3, and 55.5 mol m^{-2}.

Annually, about 486,000 km^3 of water falls from the atmosphere as precipitation. The same amount of water returns to the atmosphere annually as evaporation. Although Earth as a whole balances water, oceans and land differ in precipitation and evaporation. Approximately 373,000 km^3 of water falls over the oceans each year as precipitation. However, more water evaporates from the oceans (413,000 km^3), resulting in an annual surface deficit of 40,000 km^3 of water. Runoff from land replenishes this imbalance. Over land, precipitation exceeds evaporation. About 113,000 km^3 of water falls on land as precipitation, and 73,000 km^3 of water evaporates from land. The surplus water at the land surface (40,000 km^3) runs off to streams and rivers where it flows to the oceans to replenish the net loss of water. About 65 percent of the water reaching the land surface as precipitation returns to the atmosphere as evaporation, and 35 percent runs off to the ocean.

The average length of time a parcel of water spends in a reservoir can be calculated from the ratio of water storage to inflow or outflow (inflow and outflow are equal assuming steady state). This ratio is known as residence time or turnover time and measures the time required to replace all the water in the reservoir. For example, the atmosphere holds 12,700 km^3 of water but precipitates 486,000 km^3 of water per year. The turnover time is 9.5 days. For oceans, the residence time is 3000 years or so. The residence time is days for water in the upper soil to thousands of years for deep aquifers.

3.4.2 Atmospheric Humidity

It is convenient to distinguish water vapor and dry air. The latter is a general term for all gases other than H_2O, and the sum of these two components is the moist air. The total pressure of air (P, Pa) is the sum of dry air and water vapor:

$$P = P_d + e \tag{3.26}$$

where P_d is the partial pressure of dry air (Pa) and e is the partial pressure of water vapor (Pa), commonly called vapor pressure. Dry air with partial pressure $P_d = P - e$ follows the ideal gas law with density (ρ_d, kg m^{-3}):

$$\rho_d = \frac{P - e}{\Re T} M_a \tag{3.27}$$

where M_a is the molecular mass of dry air (28.97 g mol^{-1}). The density of water vapor (ρ_v, kg m^{-3}) is:

$$\rho_v = \frac{e}{\Re T} M_w = \frac{0.622e}{\Re T} M_a \tag{3.28}$$

with M_w the molecular mass of water (18.02 g mol^{-1}) and $M_w / M_a = 0.622$.

The density of moist air is the sum of that for dry air and water vapor:

$$\rho = \rho_d + \rho_v = \frac{P}{\Re T} M_a \left(1 - 0.378 \frac{e}{P}\right) \tag{3.29}$$

The density of moist air is less than the density of dry air at the same temperature and pressure. For a parcel of air at STP, $\rho = 1.225$ kg m^{-3} with $e = 0$ and $\rho = 1.217$ kg m^{-3} with $e = 1704$ Pa (i.e., saturated). Because moist air is less dense than dry air, water vapor is a source of buoyancy in the atmosphere.

Mass mixing ratio and specific humidity are common measures of the moisture in air. Mass mixing ratio (χ_v, kg kg^{-1}) is defined as the ratio of the mass of water vapor (m_v) in a parcel of air to the mass of dry air (m_d) in the parcel (i.e., excluding the water vapor). It is related to vapor pressure by:

$$\chi_v = \frac{m_v}{m_d} = \frac{\rho_v}{\rho_d} = \frac{0.622e}{P - e} \tag{3.30}$$

Specific humidity (q, kg kg^{-1}) is defined as the ratio of the mass of water vapor in a parcel of air to the total mass of the air. Specific humidity is related to vapor pressure by:

$$q = \frac{m_v}{m_d + m_v} = \frac{\rho_v}{\rho_d + \rho_v} = \frac{0.622e}{P - 0.378e} \tag{3.31}$$

Saturation vapor pressure is the maximum amount of water vapor that a parcel of air can hold. Saturation vapor pressure increases exponentially with warmer temperature, and warm air can hold considerably more water vapor when saturated than can cold air (Figure 3.5). For example, a parcel of air with a temperature of 30°C has a saturation vapor pressure of 4243 Pa; a parcel of air with a temperature of 18.5°C has a saturation vapor pressure of 2129 Pa. One kilogram of air can hold about

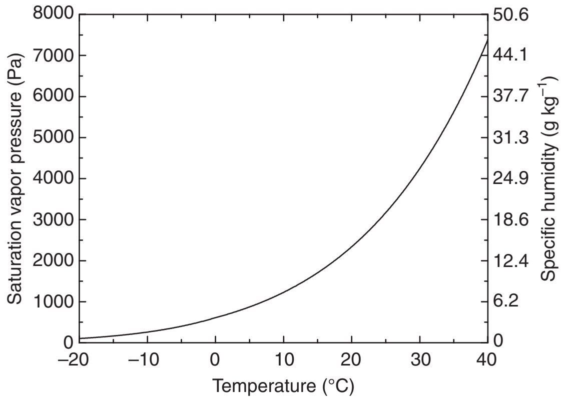

Fig. 3.5 Saturation vapor pressure as a function of temperature. The left-hand axis shows vapor pressure in pascals. The right-hand axis shows specific humidity.

13 g of water when saturated at 18.5°C and 26 g at 30°C.

Relative humidity is a measure of how saturated the air is with water. It is the ratio of e, the actual vapor pressure, to $e_*(T)$, the saturated vapor pressure at temperature T, expressed as a percentage:

$$RH = 100\, e/e_*(T) \tag{3.32}$$

For a constant vapor pressure, relative humidity decreases as temperature, and therefore saturated vapor pressure, increases. A parcel of air with a vapor pressure of 2129 Pa is saturated at a temperature of 18.5°C (i.e., RH = 100%), but has a relative humidity of only 50% at 30°C.

The vapor pressure deficit, $e_*(T) - e$, is the difference between the saturation vapor pressure (i.e., the maximum amount of water vapor that can be held in the air) and the actual vapor pressure. It is a measure of the drying potential of air and is an indication of evaporative potential.

3.4.3 Phase Change

The hydrologic cycle is a transfer of water, measured by volume or mass (1 m^3 H_2O = 1000 kg H_2O). It is also an exchange of energy between the surface and atmosphere. Water occurs in three forms: solid (ice), liquid, or gas (vapor). Energy is required to melt ice (solid to liquid) or to evaporate water (liquid to gas). This energy does not change the temperature of the water. Rather, it only changes the molecular state of the water. This energy is stored in the water molecules and released in the reverse process as water vapor condenses to liquid or liquid water freezes to ice. This stored energy is called latent heat because the temperature of water does not change with the gain or loss of heat; only the state of the water molecules changes.

Considerable amounts of energy are required to change water among its solid, liquid, and vapor states (Table 3.3). The energy absorbed in the evaporation of water is called the latent heat of vaporization. At 15°C, 2466 J are required to change one gram of water from liquid to vapor. For the Earth as a whole, 80 W m^{-2} are used annually in evaporation – more than three-quarters of the 98 W m^{-2} net radiation at the surface (Figure 3.2). For land, more than one-half (39 W m^{-2}) of the annual net radiation at the surface (66 W m^{-2}) is used to evaporate water (Trenberth et al. 2009). This stored energy is released as latent heat of condensation when water vapor condenses back to liquid. Absorption of energy cools the evaporating surface while condensation releases energy to the surrounding environment. The latent heat of vaporization decreases with warmer temperatures because water molecules contain more internal energy at warmer temperatures and less energy is required for them to evaporate.

The energy required to melt frozen water is called the latent heat of fusion. At 0°C it takes 334 J to melt one gram of water. During melting, this energy is absorbed by the water molecules, changing their phase from solid to liquid rather than warming the environment. The same heat is released when liquid water freezes. This is the reason commercial fruit growers spray orchards with water when a cold freeze is imminent. The change from liquid to ice releases heat, and while the water is freezing its temperature remains at 0°C. The latent heat of fusion decreases with colder temperatures because the water molecules contain less internal energy and less energy is released when they freeze. A third type of phase change, sublimation, occurs when ice changes directly to vapor without passing through the liquid phase.

Table 3.3 Latent heat and saturation vapor pressure in relation to temperature

Temperature (°C)	Latent heat ($J\ g^{-1}$)			$e_*(T)$ (Pa)	$de_*(T)/dT$ ($Pa\ K^{-1}$)
	Vaporization	Fusion	Sublimation		
–20	2550	289	2839	103	10
–10	2525	312	2837	260	23
0	2501	334	2835	611	44
5	2490			872	61
10	2478			1227	82
15	2466			1704	110
20	2454			2337	146
25	2442			3167	189
30	2430			4243	243
35	2419			5624	311
40	2407			7378	393

Note: Multiply by the molecular mass of water (18.02 $g\ mol^{-1}$) to convert latent heat from $J\ g^{-1}$ to $J\ mol^{-1}$.

3.5 Biogeochemical Cycles

The chemical composition of the atmosphere is a crucial determinant of climate. Atmospheric gases interact with radiant energy flowing through the atmosphere to affect climate through the greenhouse effect. Chief among these gases are carbon dioxide (CO_2), methane (CH_4), and nitrous oxide (N_2O). The concentration of these gases in the atmosphere is the result of natural and anthropogenic processes.

3.5.1 Carbon Dioxide

Carbon dioxide is the most widely recognized greenhouse gas. Its average concentration in the atmosphere for the year 2011 was 390 ppm (Table 2.1). This represents the balance of geological processes, biological processes, and human activities. Of the some 10^{23} g of carbon on Earth, all but a small portion is buried in sedimentary rocks. Only about 0.04 percent of the carbon (40,000 Pg) is in biologically active pools near Earth's surface, and oceans hold the vast majority of the biologically active carbon (38,000 Pg C). Soils store most of the biologically active carbon on land. It is estimated that plants contain 450–650 Pg C worldwide while soils hold 1500–2400 Pg C with an additional ~1700 Pg C or more in permafrost. The atmosphere has the least carbon, slightly more than 800 Pg C during the period 2000–2009.

The global carbon cycle represents the interactions of two superimposed cycles: the geological carbon cycle, in which carbon cycles among atmosphere, oceans, and continents in response to the chemical weathering of rocks over a period of millions of years; and the biological carbon cycle, in which carbon cycles among the atmosphere and marine and terrestrial organisms through biological and physical processes over shorter timescales of days, seasons, years, and decades. The annual carbon fluxes in the geological carbon cycle are small compared with the biological fluxes.

In the preindustrial era, the carbon cycle is thought to have been in balance with no net carbon gain by the land or ocean (Figure 3.6a). The biosphere was a small annual sink of carbon (1.7 Pg C yr^{-1}) because photosynthetic uptake exceeded losses from respiration and wildfire. This and additional carbon from rock weathering (0.4 Pg C yr^{-1}) washed into rivers and lakes, where it returned to the atmosphere in freshwater outgassing (1.0 Pg C yr^{-1}), was buried in sediments (0.2 Pg C yr^{-1}), or was

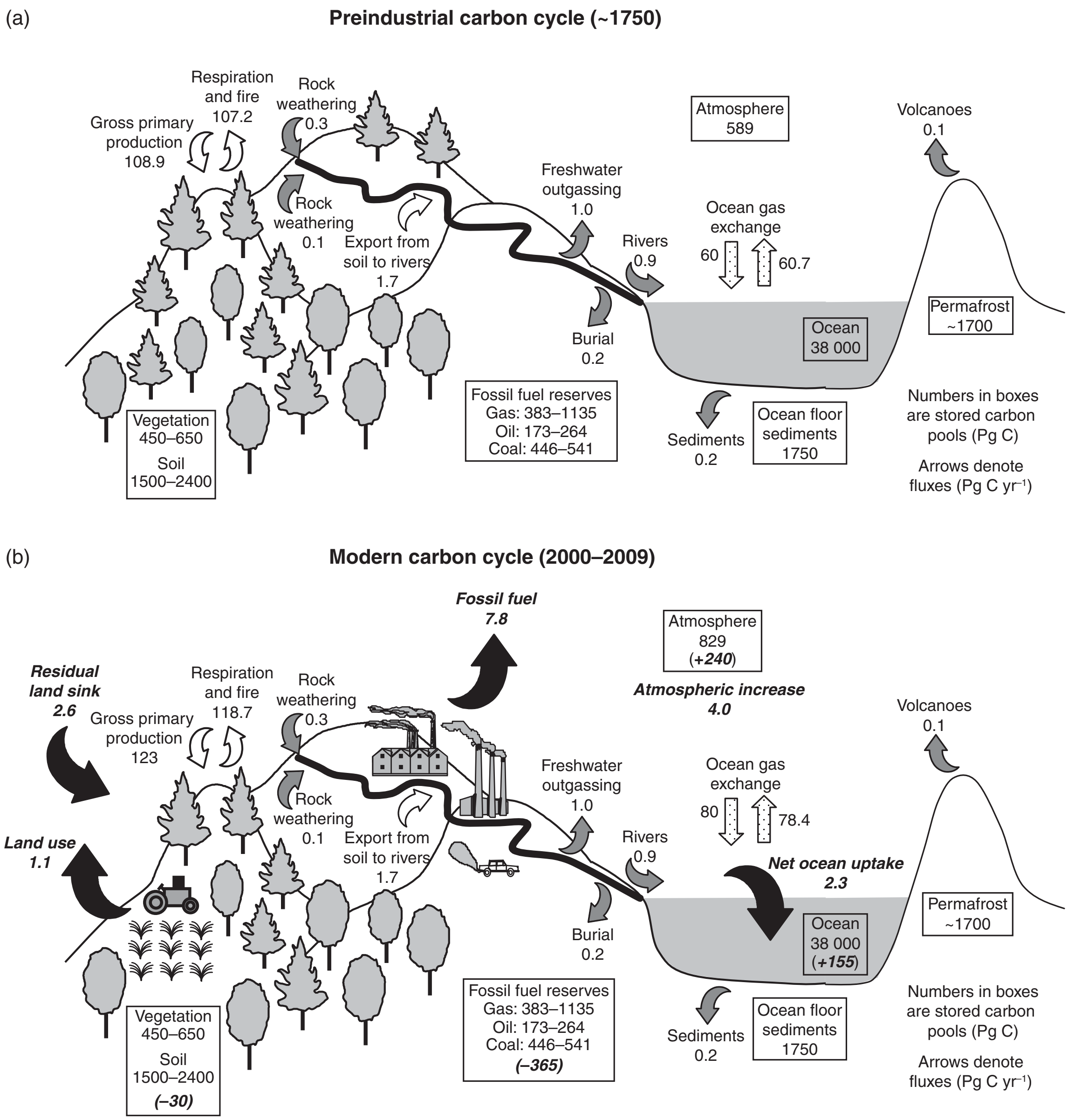

Fig. 3.6 Global carbon cycle for (a) the preindustrial era (~1750) and (b) the modern carbon cycle (2000–2009). Gray arrows denote fluxes in the geologic cycle. Black arrows and bold italic letters show the major changes in fluxes since 1750. Boxes show carbon pools and the cumulative change in carbon since 1750. Adapted from Ciais et al. (2013).

carried into oceans (0.9 Pg C yr^{-1}). The carbon input balanced a small net loss from air–sea exchange (0.7 Pg C yr^{-1}) and burial on the ocean floor (0.2 Pg C yr^{-1}).

Human activities have significantly impacted the global carbon cycle (Figure 3.6b). The burning of oil and coal to generate heat and electricity, the combustion of gasoline for transportation, and other industrial processes release CO_2 to the atmosphere. During the period 2000–2009, these activities emitted 7.8 Pg C yr^{-1}. In addition, human uses of land, particularly forest clearing, emitted another 1.1 Pg C yr^{-1}. Slightly less than half of the total anthropogenic emission

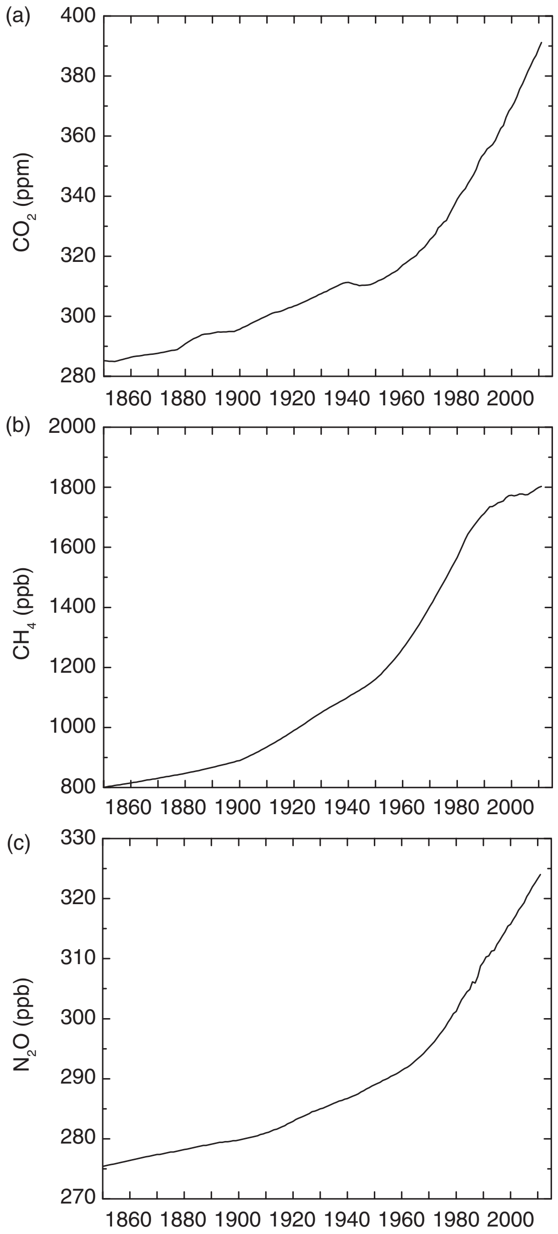

Fig. 3.7 Atmospheric CO_2, CH_4, and N_2O concentration from 1850 to 2011. Data updated from Hansen et al. (1998) and Hansen and Sato (2004) and provided by the NASA Goddard Institute for Space Studies (New York City, New York).

(4.0 Pg C yr^{-1}) remained in the atmosphere. The rest was taken up by the oceans (2.3 Pg C yr^{-1}) and terrestrial ecosystems (2.6 Pg C yr^{-1}). As a result of these processes, the concentration of CO_2 in the atmosphere has increased since the preindustrial era (Figure 3.7a).

The land uptake is the balance among several large, but highly uncertain processes. It is estimated that terrestrial plants gain 123 Pg C yr^{-1} during photosynthesis (Beer et al. 2010). It is thought that half of this carbon returns to the atmosphere during autotrophic respiration, by which plants maintain and grow new biomass. The remainder is the annual net primary productivity of terrestrial plants. Leaves, twigs, and other plant debris fall to the ground and decompose. Decomposition of plant debris, wildfire, and other non-respiratory losses return much of this carbon to the atmosphere. From our current understanding of the global carbon cycle, the land must have gained 2.6 Pg C yr^{-1} over the period 2000–2009 after accounting for these carbon losses. Respiration and fire must emit 118.7 Pg C yr^{-1} to balance the photosynthetic carbon gain and export of carbon to rivers. In fact, however, more carbon may be flowing through the aquatic system because of larger carbon export from soils (Raymond et al. 2013; Regnier et al. 2013). The magnitude, geographic location, and causes of the terrestrial carbon sink are the subject of considerable scientific research.

Oceans are thought to have a net uptake of 2.3 Pg C yr^{-1}. Carbon dioxide is exchanged between air and sea through a variety of physical and chemical processes that affect the solubility of CO_2 in water. In addition, growth of phytoplankton absorbs CO_2. The carbon is buried in sediments as the organisms die and settle on the ocean floor. The oceanic biological pump is an important regulator of atmospheric CO_2. Without it, atmospheric CO_2 concentrations would be considerably greater.

3.5.2 Methane

Methane is another important greenhouse gas that cycles among atmosphere, ocean, and terrestrial pools. Its atmospheric concentration, like CO_2, has increased since the beginning of the industrial era (Figure 3.7b). The largest natural source of CH_4 is from wetlands, where anaerobic decomposition in waterlogged soils produces CH_4 (Table 3.4). Termites and other