CHAPTER

10

The Global Water Cycle

OUTLINE

INTRODUCTION

The annual circulation of water is the largest movement of a chemical substance at the surface of the Earth. Through evaporation and precipitation, water transfers much of the heat energy received by the Earth from the tropics to the poles, just as a steam-heating system transfers heat from the furnace to the rooms of a house. Movements of water vapor through the atmosphere determine the distribution of rainfall on Earth, and the annual availability of water on land is the single most important factor that determines the growth of plants (Kramer 1982). Where precipitation exceeds evapotranspiration on land, there is runoff. Runoff carries the products of mechanical and chemical weathering to the sea.

In this chapter, we examine a general outline of the global hydrologic cycle and then look briefly at some indications of past changes in the hydrologic cycle and global water balance. Finally, we look somewhat speculatively at changes in the water cycle that may accompany global climate change and other human impacts in the future. These changes will have direct effects on global patterns of plant growth, the height of sea level, and the movement of materials in biogeochemical cycles.

Too often, we forget that adequate freshwater is the most essential resource for human survival. Changes in the water cycle have strong implications for the future of agricultural productivity and for the social and economic well-being of human society (Vörösmarty et al. 2000). Archeologists find the ruins of elaborate water-transport infrastructure throughout human history (e.g., Bono and Boni 1996, Sandor et al. 2007). Widespread drought seems associated with the collapse of the early Mesopotamian civilization in the Middle East around 2200 B.C. (Weiss et al. 1993) and the disappearance of the Mayan civilization in Mexico around 900 A.D. (Hodell et al. 1995, Peterson and Haug 2005, Medina-Elizalde and Rohlijng 2012).

THE GLOBAL WATER CYCLE

The quantities of water in the global hydrologic cycle are so large that it is traditional to describe the pools and transfers in units of km^3 (Figure 10.1). Each km^3 contains 10^{12} liters and weighs 10^{15} g. The flux of water in the water cycle may also be expressed in units of average depth. For example, if all the rainfall on land were spread evenly over the surface, each weather station would record a depth of about 700 mm/yr. Units of depth can also be used to express runoff and evaporation (e.g., Figure 4.17). For instance, annual evaporation from the oceans removes the equivalent of 1000 mm of water each year from the surface area of the sea.

Not surprisingly, the oceans are the dominant pool in the global water cycle (Figure 10.1). Seawater composes more than 97% of all the water at the surface of the Earth. The mean depth of the oceans is 3500 m (Chapter 9). Water held in polar ice caps and in continental glaciers is the next largest contributor to the global pool. About 70% of the world's freshwater is frozen in Antarctica (Parkinson 2006). Discounting the water in the oceans and ice caps, less than 1% of the Earth's water is fresh and accessible in rivers, lakes, and groundwater.

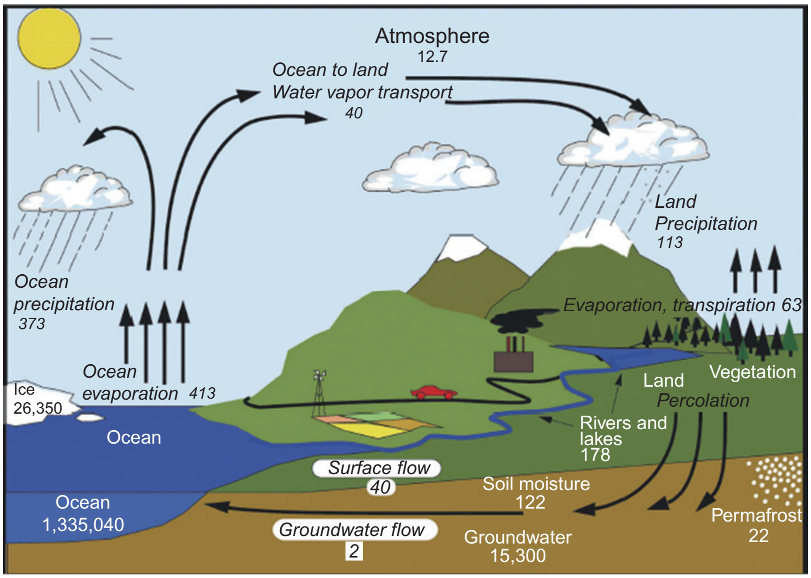

FIGURE 10.1 The global hydrologic cycle, with pools in units of 10^3 km^3 and flux in 10^3 km^3/yr. *Source: Modified from Trenberth (2007). Used with permission of the American Meteorological Society.*

Soils contain 121,800 km^3 of water, of which about 58,100 km^3 is within the rooting zone of plants (Webb et al. 1993). The large pool of freshwater below the unsaturated or vadose zone is known as groundwater (Chapter 7). Global estimates of the volume of groundwater are poorly constrained—4,200,000 to 23,400,000 km^3—but, except as a result of a few deep-rooted plants (e.g., Dawson and Ehleringer 1991) and the ingenuity of humans, groundwater is largely inaccessible to the biosphere. The pool of water in the atmosphere is tiny, equivalent to about 25 mm of rainfall at any given time (Eq. 3.4). Nevertheless, enormous quantities of water move through the atmosphere each year.

Evaporation removes about 413,000 km^3/yr of water from the world's oceans (compare Syed et al. 2010). Thus, the mean residence time of ocean water with respect to losses to the atmosphere is about 3100 years. Only about 373,000 km^3/yr of this water returns to the oceans in rainfall; the rest contributes to precipitation on land, which totals 113,000 km^3/yr. Plant transpiration and surface evaporation return 63,000 $\pm$ 13,100 km^3/yr from soils to the atmosphere (Ryu et al. 2011). The relative importance of plant transpiration (T) varies regionally (Table 10.1), but globally plants appear to be responsible for about 60% of total terrestrial evapotranspiration (ET) to the atmosphere (Choudhury et al. 1998, Alton et al. 2009), with the fraction T/ET increasing from 61% at 25% vegetation cover to 83% at 100% cover in greenhouse experiments (Wang et al. 2010a). With respect to precipitation inputs or evapotranspiration losses, the mean residence time of soil water is about 1 year.

When precipitation is below average, droughts develop relatively quickly, impacting a wide variety of ecosystem processes. During a 2005 drought, the rainforests of the Amazon Basin lost $>1.2 \times 10^{15}$ g C to the atmosphere (Phillips et al. 2009; compare Tian et al. 1998, Potter et al. 2011). Owing to the excess of precipitation over evapotranspiration on land, about 40,000 km^3/yr becomes runoff, derived from surface flow and groundwater. The runoff in rivers is supplemented by the subterranean flow of groundwater to the sea in coastal areas, which is potentially large but difficult to estimate (Moore 1996, 2010). One study estimates this flow as 2200 to 2400 km^3/yr (Zektser et al. 2007).

These global average values obscure large regional differences in the water cycle. Evaporation from the oceans is not uniform, but ranges from 4 mm/day in tropical latitudes to <1 mm/day at the poles (Mitchell 1983). Although much precipitation falls at tropical latitudes, an excess of evaporation over precipitation from the tropical oceans provides a net regional flux of water vapor to the atmosphere. Net evaporative loss accounts for the high salinity in tropical oceans (Figure 9.5), and the movement of water vapor in the atmosphere carries latent heat to polar regions (Trenberth and Caron 2001).

On land the relative balance of precipitation and evaporation differs strongly between regions. In tropical rainforests, precipitation may greatly exceed evapotranspiration. Shuttleworth (1988) calculates that 50% of the rainfall becomes runoff in the Amazon rainforests (Table 10.1). In desert regions, precipitation and evapotranspiration are essentially equal, so there is no runoff and only limited recharge of groundwater (Scanlon et al. 2006). As a global average, rivers carry about one-third of the precipitation from land to the sea (compare Alton et al. 2009). About 11% of precipitation becomes groundwater (12,666 km^3/yr; Doll and Fiedler 2008, Zektser and Loaiciga 1993), and the mean residence time of groundwater is about 1000 years (Slutsky and Yen 1997).

TABLE 10.1 Relative Importance of Pathways Leading to the Loss of Water from Terrestrial Ecosystems (except for T/ET, all data are shown in % of precipitation)

Vegetation	Precipitation (mm/yr)	Transpiration	Evaporation	Runoff and recharge	T/ET × 100	References
Tropical Rainforest						
	1623	45	56		45	Banerjee, cited in Galoux et al. 1981
	2000	49	26	26	65	Salati and Vose 1984
	2000	62	19	19	77	
		40	10	50	80	Shuttleworth 1988
	2209	56	11	32	84	Leopoldo et al. 1995
	2851	31	21		60	Calder et al. 1986
	3725	14	9		61	Frangi and Lugo 1985
Temperate Coniferous Forest						
	366	52	53		52	Pražák et al. 1994[a]
	595	59	36		55	Gash and Stewart 1977
	626	50	49		50	Lutzke and Simon, cited in Galoux et al. 1981
	627	40	48		41	Lutzke and Simon
	710	47	39		55	Rutter, cited in Galoux et al. 1981
	725	37	22	41	63	Tajchman 1972
	1085	49	15	41	77	Waring et al. 1981
	1127	8	12	80	39	Barbour et al. 2005[a]
	1225	49	15	38	77	McNulty et al. 1996
	2175	35	15	46	70	Waring et al. 1981
	2355	16	11	72	50	Waring et al. 1981
	2620	7	23		23	Hudson 1988
Temperate Deciduous Forest						
	349	39	58		40	Mitchell et al. 2009 (and personal communication)
	513	49	36	5	58	Molchanov, cited in Galoux et al. 1981
	549	54	9		86	Ladekarl 1998
	590	28-47				Gebauer et al. 2012[a]
	669	73	27		73	Paço et al. 2009
	763	33	15		69	Granier et al. 2000a[a]

TABLE 10.1 Cont'd

Vegetation	Precipitation (mm/yr)	Transpiration	Evaporation	Runoff and recharge	T/ET × 100	References
	1333	19	14		58	Wilson et al. 2001[a]
	2175	28	12	55	70	Waring et al. 1981
Boreal Forest						
	250	46	25		65	Grelle et al. 1997[a]
	271	51				Cienciala et al. 1997[a]
	872	45	8		85	Telmer and Veizer 2000
	502	39	35		53	Ten studies by Molchanov 1963, cited by Choudhury et al. 1998
	1237	19	26		42	Two studies by Delfs 1967, cited by Choudhury et al. 1998
Mediterranean Shrubland						
	475	60	40		60	Poole et al. 1981
	475	32	51	4	39	
	590	35	55	10	39	
Tropical Grassland						
	180	32	68		32	Hsieh et al. 1998[b]
	570	59	41		59	Chadwick et al. 2003
	1380	61	39		61	
	2500	72	28		72	
Temperate Grassland						
	365	65				Trlica and Biondini 1990[a]
	341	49	51		49	Lauenroth and Bradford 2006
		67	33	0	67	Massman 1992
	350	39	73		39	Hu et al. 2009
	477	37	67		37	
	580	56	83		56	
	580	39	49		44	
Steppe						
	144	45	55	0	45	Floret et al. 1982
	150	34	56	10	38	Paruelo and Sala 1995
	275	55	34		62	Huang et al. 2010[a]

Continued

TABLE 10.1 Relative Importance of Pathways Leading to the Loss of Water from Terrestrial Ecosystems (except for T/ET, all data are shown in % of precipitation)—Cont'd

Vegetation	Precipitation (mm/yr)	Transpiration	Evaporation	Runoff and recharge	T/ET × 100	References
Desert						
	150	35	65		35	Smith et al. 1995
	150	38	62		38	Liu et al. 2012
	165	27	73		27	Lane et al. 1984
	210	72	28		72	Schlesinger et al. 1987
	200	80	20		80	Liu et al. 1995
	260	21	36		37	Cavanaugh et al. 2011[a]
	212	21	27		44	

[a] *Growing season only*
[b] *Values are percentage of soil moisture lost at each site.*

The concept of *potential evapotranspiration* (PET), developed by hydrologists, expresses the maximum evapotranspiration that would be expected to occur under the climatic conditions of a particular site, assuming that water is always present in the soil and plant cover is 100%. Potential evapotranspiration is greater than the evaporation from an open pond, as a result of the plant uptake of water from the deep soil and a leaf area index >1.0 in many plant communities (Chapter 5). In tropical rainforests, PET and actual evapotranspiration (AET) are about equal (Vörösmarty et al. 1989). In deserts, PET greatly exceeds actual AET, owing to long periods when the soils are dry. In southern New Mexico, precipitation averages about 210 mm/year, but the receipt of solar energy could potentially evaporate over 2000 mm/yr from the soil (Phillips et al. 1988).

With higher values in warm, wet conditions, actual evapotranspiration (Figure 10.2) is often useful as a predictor of primary production (Figure 5.8), decomposition (Figure 5.17), and microbial activity (Figure 4.3). Changes in climate that affect rainfall and AET have a dramatic effect on the biosphere. Annual variability in AET is greatest in ecosystems with low AET, causing large year-to-year variations in net primary production related to annual precipitation in deserts (Frank and Inouye 1994, Prince et al. 1998). Actual evapotranspiration is more constant in tundra and boreal forest ecosystems, where wet soils do not constrain the supply of water to plants. The net primary productivity of land plants (60×10^{15} g C/yr) and the plant transpiration of water from land (~60% of 63×10^{18} g/yr) indicate that the global average water-use efficiency of vegetation is about 2.4 mmol of CO_2 fixed per mole of water lost (Eq. 5.3)—somewhat higher than the range measured by physiologists studying individual leaves (Chapter 5).

The sources of water contributing to precipitation also differ greatly among different regions of the Earth. Nearly all the rainfall over the oceans is derived from the oceans. On land, much of the rainfall in maritime and monsoonal climates is also derived from evaporation from the sea. Estimates of the percentage of rainfall derived from evapotranspiration from

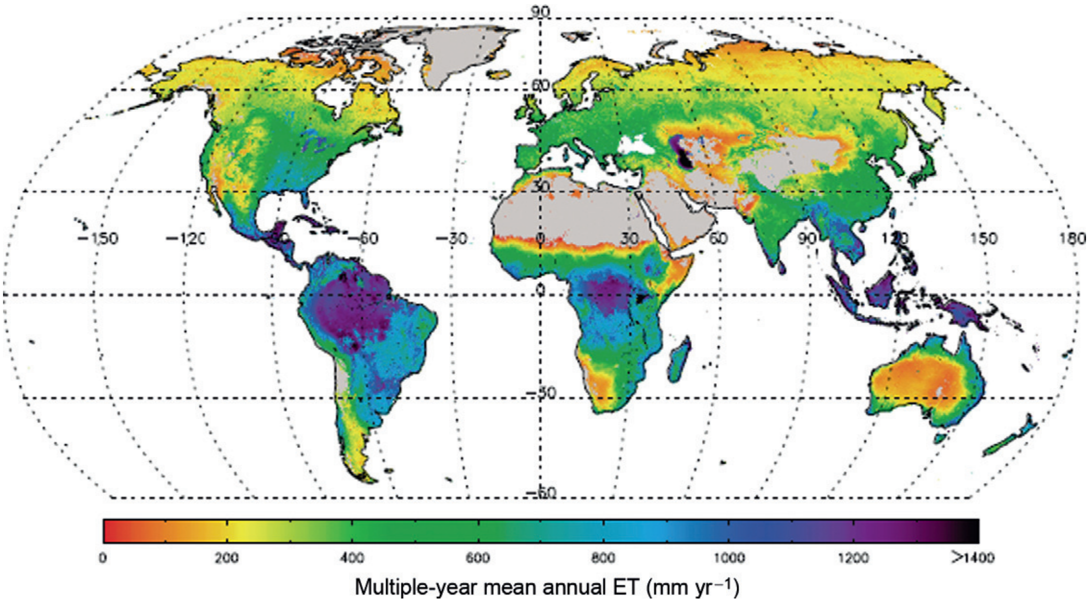

FIGURE 10.2 Global loss of water from the land surface by evapotranspiration. *Source: From Zhang et al. (2010). Used with permission of the American Geophysical Union.*

land are poorly constrained and regionally variable, with global estimates varying by 10 to 60% (Trenberth 1998, van der Ent 2010).[1]

Between 25 to 50% of the water falling in the Amazon Basin may be derived from evapotranspiration within the basin, with the rest derived from long-distance atmospheric transport from other regions (Salati and Vose 1984, Eltahir and Bras 1994). Evapotranspiration in Amazon forests is maximized by deep-rooted plants (Nepstad et al. 1994), and the regional importance of evapotranspiration in the Amazon basin speaks strongly for the long-term implications of forest destruction in that region. Using a general circulation model of the Earth's climate, Lean and Warrilow (1989) show that a replacement of the Amazon rainforest by a savanna would decrease regional evaporation and precipitation and increase surface temperatures (compare Shukla et al. 1990).

Regional cooling from plant transpiration is well known from comparisons of urban and rural areas (Juang et al. 2007). In semiarid regions, precipitation may decline as a result of the removal of vegetation, leading to soil warming (Balling 1989, Kurc and Small 2004, He et al. 2010) and increasing desertification (Schlesinger 1990, Chahine 1995, Koster et al. 2004). Thus, the transpiration of land plants is an important factor determining the movement of water in the hydrologic cycle and Earth's climate.

Estimates of global riverflow range from 33,500 to 47,000 km^3/yr (Lvovitch 1973, Speidel and Agnew 1982). A recent satellite-derived global estimate indicates 36,055 km^3 of annual

[1] Note the scale-dependence of such a statistic. At a local scale, nearly all the precipitation will be derived outside of the local area, whereas at larger regional scales an increased percentage will be derived within the area of interest (Trenberth 1998).

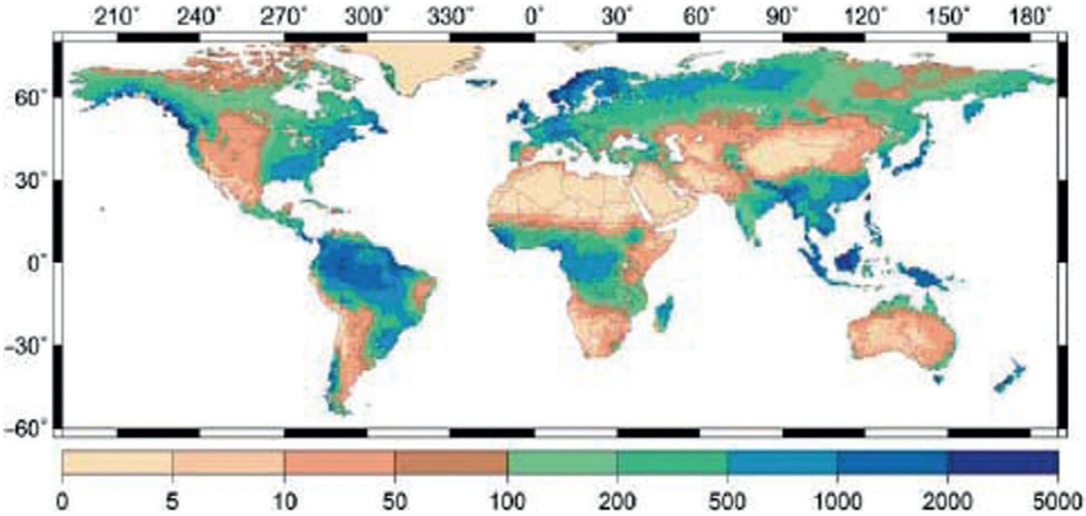

FIGURE 10.3 Annual runoff to the oceans, in mm/yr. *Source: From Oki and Kanae (2006). Used with permission of the American Association for the Advancement of Science.*

riverflow to the sea between 1994 and 2006 (Syed et al. 2010). For global models, many workers assume a value of about 40,000 km^3/yr (Figure 10.1). The distribution of flow among rivers is highly skewed. The 50 largest rivers carry about 43% of total riverflow; the Amazon alone carries ~20%. Reasonable estimates of the global transport of organic carbon, inorganic nutrients, and suspended sediments from land to sea can be based on data from a few large rivers. As a result of the positions of the continents and their surface topography, there are large regional differences in the delivery of runoff to the sea (Figure 10.3). The average runoff from North America is about 320 mm/yr, whereas the average runoff from Australia, which has a large area of internal drainage and deserts, is only 40 mm/yr (Tamrazyan 1989). Thus, the delivery of dissolved and suspended sediment to the oceans varies greatly between rivers draining different continents (Table 4.8).

In the Northern Hemisphere, 77% of the runoff is carried in rivers in which the flow is now regulated by dams and other human structures (Dynesius and Nilsson 1994, Nilsson et al. 2005), which strongly affect the sediment transport to the sea. The mean transit time of water to the sea has increased by an average of 60 days in rivers with significant impoundments and reservoirs (Vörösmarty and Sahagian 2000). Postel et al. (1996) calculate that humans now use 17% of the volume of rivers globally (~7000 km^2), converting a large portion of it to water vapor as a result of irrigated agriculture (Rost et al. 2008).

In some regions, such as the southwestern United States, human appropriation of surface runoff is as high as 76% (Sabo et al. 2010). Human extraction of groundwater, 145 to 283 km^3/yr (Wada et al. 2010, Konikow 2011), is less than 2% of groundwater recharge, but it is centered in arid and semiarid regions, which show large groundwater depletion (Rodell et al. 2009). Much of this groundwater is used for irrigation (Siebert et al. 2010, Wada et al. 2012). In the central valley of California, groundwater withdrawal of about 3 km^3/yr lowered the water table by ~20.3 mm/yr between 2003 and 2010 (Figure 10.4).

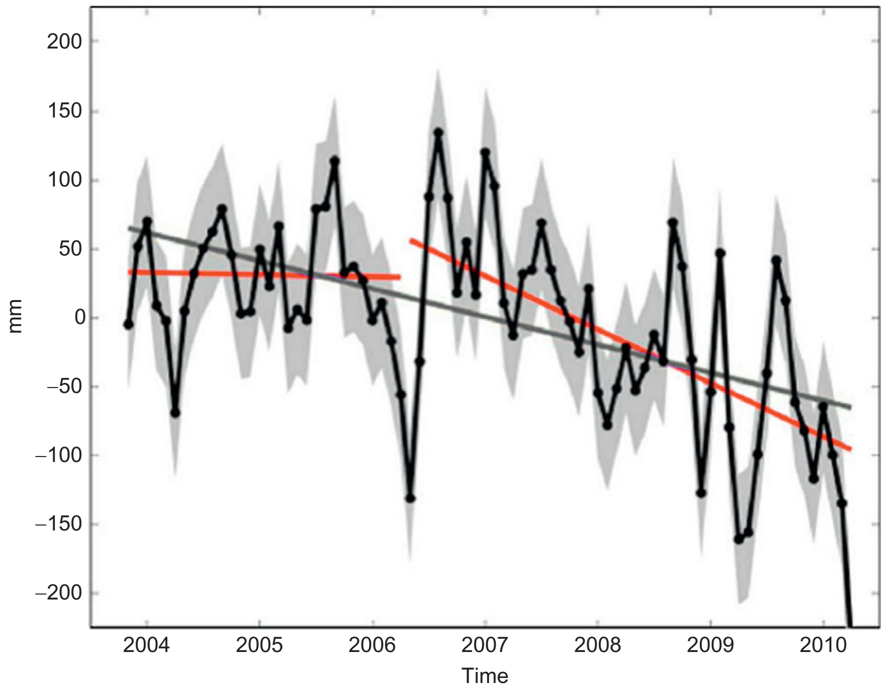

FIGURE 10.4 Groundwater levels in the Central Valley of California, showing changes in mm from a 2004 baseline value. *Source: From Famiglietti et al. (2011). Used with permission of the American Geophysical Union.*

The mean residence time of the oceans with respect to riverflow is about 34,000 years. The mean residence times differ among ocean basins. The mean residence time for water in the Pacific Ocean is 57,500 years—significantly longer than that for the Atlantic (18,000 years; Speidel and Agnew 1982, p. 30). This is consistent with the greater accumulation of nutrients in deep Pacific waters and a shallower carbonate compensation depth in the Pacific Ocean (Chapter 9). Despite the enormous riverflow of the Amazon, continental runoff to the Atlantic Ocean is less than the loss of water through evaporation. Thus, the Atlantic Ocean has a net water deficit, which accounts for its greater salinity (refer to Figure 9.5). Conversely, the Pacific Ocean receives a greater proportion of the total freshwater returning to the sea each year. Ocean currents carry water from the Pacific and Indian oceans to the Atlantic Ocean to restore the balance (refer to Figure 9.3)

MODELS OF THE HYDROLOGIC CYCLE

A variety of models have been developed to predict the movement of water in hydrologic cycles. Watershed models follow the fate of water received in precipitation and calculate runoff after subtraction of losses due to evaporation and plant uptake (Waring et al. 1981, Moorhead et al. 1989, Ostendorf and Reynolds 1993). In these models, the soil is considered as a collection of small boxes, in which the annual input and output of water must be equal.

Water entering the soil in excess of its water-holding capacity is routed to the next lower soil layer or to the next downslope soil unit via subsurface flow (Chapter 8). Models of water movement in the soil can be coupled to models of soil chemistry to predict the loss of elements in runoff (e.g., Nielsen et al. 1986, Knight et al. 1985, Furrer et al. 1990).

A major challenge in building these models is the calculation of plant uptake and transpiration loss. This flux is usually computed using a formulation of the basic diffusion law, in which the loss of water is determined by the gradient, or vapor pressure deficit, between plant leaves and the atmosphere. The loss is mediated by a resistance term, which includes stomatal conductance and wind speed (Chapter 5). In a model of forest hydrology, Running et al. (1989) assume that canopy conductance decreases to zero when air temperatures fall below 0°C or soil water potential declines below -1.6 MPa. Their model appears to give an accurate regional prediction of evapotranspiration and primary productivity for a variety of forest types in western Montana.

Larger-scale models have been developed to assess the contribution of continental land areas to the global hydrologic cycle. For example, Vörösmarty et al. (1989) divided South America into 5700 boxes, each $1/2° \times 1/2°$ in size. Large-scale maps of each nation were used to characterize the vegetation and soils in each box, and data from local weather stations were used to characterize the climate. A model (Figure 10.5) was used to calculate the water balance

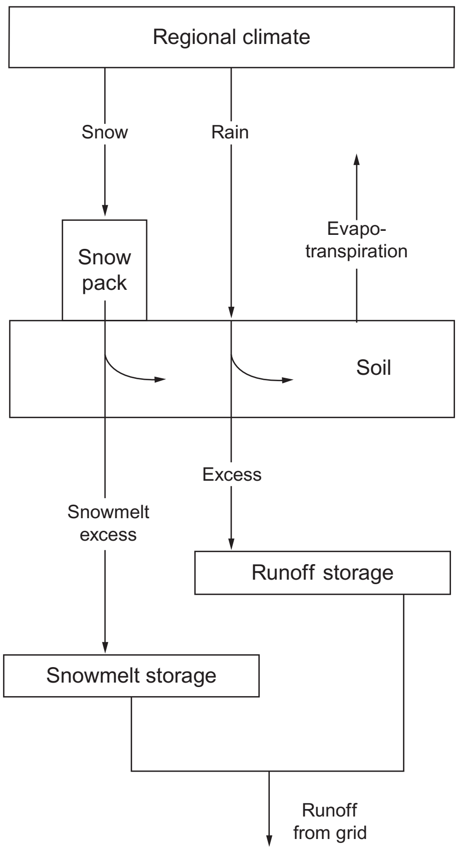

FIGURE 10.5 Components of a model for the hydrologic cycle of South America. *Source: From Vörösmarty et al. (1989). Used with permission of the American Geophysical Union.*

in each unit. During periods of rainfall, soil moisture storage is allowed to increase up to a maximum water-holding capacity determined by soil texture. During dry periods, water is lost to evapotranspiration, with the rate becoming a declining fraction of PET as the soil dries.

This type of model can be coupled to other models, including general circulation models of the Earth's climate (Chapter 3), to predict global biogeochemical phenomena. For example, a monthly prediction of soil moisture content for the South American continent can be used with known relationships between soil moisture and denitrification (Figure 6.17) to predict the loss of N_2O and the total loss of gaseous nitrogen from soils to the atmosphere (Potter et al. 1996). The excess water in the water-balance model is routed to stream channels, where it can be used to predict the flow of the major rivers draining the continent (Russell and Miller 1990, Milly et al. 2005). Changes in land use and the destruction of vegetation are easily added to these models, allowing a prediction of future changes in continental-scale hydrology and biogeochemistry.

THE HISTORY OF THE WATER CYCLE

As we learned in Chapter 2, water was delivered to the primitive Earth during its accretion from planetesimals, meteors, and comets. The accretion of the planets was largely complete by 4.5 billion years ago (bya). Then water was released from the Earth's mantle in volcanic eruptions (i.e., degassing), which continue to deliver water vapor to the atmosphere today—equivalent to $\sim 2.5\ km^3/yr$ (Wallman 2001). As long as the Earth's temperature was greater than the boiling point of water, water vapor remained in the atmosphere. When the Earth cooled, nearly all the water condensed to form the oceans. Even then, a small amount of water vapor and CO_2 remained in the Earth's atmosphere—enough to maintain the temperature of the Earth above freezing via the greenhouse effect (Chapter 3). Today, water vapor and clouds account for 75% of the greenhouse effect on Earth (Lacis et al. 2010, Schmidt et al. 2010a). Without this greenhouse effect the surface of the Earth might be coated with a thick layer of ice, and biogeochemistry would be much less interesting.

There is good evidence of liquid oceans on Earth as early as 3.8 bya, and it is likely that the volume of water in the hydrologic cycle has not changed appreciably since that time. The total inventory of volatiles at the surface of the Earth (refer to Table 2.3) indicates that about 155×10^{22} g of water has been degassed from its crust. The difference between this value and the total of the pools of water shown in Figure 10.1 is largely contained in sedimentary rocks. Each year about 0.3 to 0.4 km^3 of water is carried to the Earth's mantle by subducted sediments (Wallace 2005, Wallman 2001, Parai and Mukhopadhyay 2012, Alt et al. 2012). This is less than the return of water vapor to the atmosphere in volcanic emissions since much of the water in subducted marine sediments is degassed before it reaches the lower mantle (Dixon et al. 2002, Green et al. 2010, van Keken et al. 2011, Alt et al. 2012). Nevertheless, a large volume of water, perhaps several times the current volume of the oceans, is retained in Earth's mantle (Marty 2012).

Owing to the low content of water vapor in the stratosphere, the Earth appears to have lost only a small amount of H_2O by photolysis, perhaps less than 35% of its total degassed water through Earth's history (Yung et al. 1989; Yung and DeMore 1999, p. 346; Pope et al. 2012). Much larger quantities appear to have been lost from Venus, where all water remains as vapor and exposed to ultraviolet light (Chapter 2). The loss of water from Venus is consistent with the high D/H ratio of its current inventory, reflecting the slower loss of its heavier isotope (Chapter 2). The accumulation of O_2 in the atmosphere and in oxidized minerals of the Earth's crust suggests that about 2% of Earth's water has been consumed by net photosynthesis through geologic time (refer to Table 2.3).

Throughout Earth's history, changes in relative sea level have accompanied periods of tectonic activity that increase (or decrease) the volume of submarine mountains. Changes in sea level also accompany changes in global temperature that lead to glaciations. The geologic record shows large changes in ocean volume during the 16 continental glaciations that occurred during the Pleistocene Epoch, extending to 2 million years ago (Bintanja et al. 2005, Dutton and Lambeck 2012). During the most recent glaciation, which reached a peak 22,000 years ago, an increment of $52{,}500 \times 10^3$ km^3 of seawater was sequestered in the polar and other glacial ice (Lambeck et al. 2002, Yokoyama et al. 2000). This represents nearly 4% of the ocean volume, and it lowered the sea level about 120 to 130 m from that of present day (Fairbanks 1989, Siddall et al. 2003). As we saw in Chapter 9, the Pleistocene glaciations are recorded in calcareous marine sediments. During periods of glaciation, the ocean was relatively rich in $H_2{}^{18}O$, which evaporates more slowly than $H_2{}^{16}O$. Calcium carbonate precipitated in these oceans shows higher values of $\delta^{18}O$, which can be used as an index of paleotemperature (refer to Figure 9.31).

Although many causes have been suggested, most climate scientists now believe that ice ages are related to small variations in the Earth's orbit around the Sun (Harrington 1987). These variations lead to differences in the receipt of solar energy, particularly in polar regions, and large changes in the Earth's hydrologic cycle. Once polar ice begins to accumulate, the cooling accelerates because snow has a high reflectivity or albedo to incoming solar radiation. Low concentrations of atmospheric CO_2 (refer to Figure 1.2) and high concentrations of sulfate aerosols (Legrand et al. 1991) and atmospheric dust (Lambert et al. 2008) during the last ice age were probably an effect, rather than a cause, of global cooling; however, these changes in the atmosphere may have reinforced the rate and onset of the cooling trend (Harvey 1988, Shakun et al. 2012).

Earth appears to enter each ice age relatively slowly ($\sim$50,000 years), whereas the warming to interglacial conditions occurs over a relatively short period ($<$1000 years; Figure 1.2). Some climate change is remarkably rapid; records of paleoclimate show periods when the mean annual temperature in Greenland rose as much as 9°C over a couple of decades (Taylor 1999, Severinghaus and Brook 1999). At the present time, the Earth is unusually warm; we are about halfway through an interglacial period, which should end about 12,000 A.D. There is substantial evidence that the Earth may have undergone several periods of frozen conditions in the Precambrian, and perhaps even more recently, yielding a "snowball" Earth, on which ice covered most of the planet (Hoffman et al. 1998, Kirschvink et al. 2000).

Continental glaciations represent a major disruption—a loss of steady-state conditions—in Earth's water cycle. Global cooling yields lower rates of evaporation, reducing the circulation of moisture through the atmosphere and reducing precipitation. One model of global climate suggests that during the last glacial epoch, total precipitation was 14% lower than that of today (Gates 1976). Throughout most of the world, the area of deserts expanded, and total net primary productivity and plant biomass on land were much lower (Chapter 5). Greater wind erosion of desert soils contributed to the accumulation of dust in ocean sediments, polar ice caps, and loess deposits (Yung et al. 1996, Lambert et al. 2008). The southwestern United States appears to have been an exception. Over most of this desert area, the climate of 18,000 years ago was wetter than today (Van Devender and Spaulding 1979, Wells 1983, Marion et al. 1985).

Changes in the rate of global riverflow produce changes in the delivery of dissolved and suspended matter to the sea. Broecker (1982) suggests that erosion of exposed continental shelf sediments during the glacial sea-level minimum may have led to a greater nutrient content of seawater and higher marine net primary productivity in glacial times. Worsley and Davies (1979) show that deep-sea sedimentation rates throughout geologic time have been greatest during periods of relatively low sea level, when a greater area of continents is displayed. These observations are consistent with the current imbalance in the ocean nitrogen cycle—the oceans may still be responding to large nitrogen inputs during the last glacial period (McElroy 1983; Figure 9.21).

THE WATER CYCLE AND CLIMATE CHANGE

Our ability to monitor ongoing changes in the global water cycle has improved dramatically with the deployment of the GRACE (Gravity Recovery and Climate Experiment) satellite system, which measures small changes in the distribution of Earth's gravity associated with changes in its mass, including that of water in soils, groundwater, and ice packs (Syed et al. 2008). For example, GRACE measurements show increasing water storage in the Amazon Basin during the seasonal period of heavy rains (Figure 10.6), and GRACE allows the calculation of evapotranspiration by subtracting changes in soil moisture from precipitation. The GRACE satellite has also allowed measurements of groundwater depletion in California (refer to Figure 10.4). Additionally, changes in sea level and ocean currents are accurately monitored by the TOPEX/Poseidon satellite (Chapter 9).

Rise in Sea Level

It is widely believed that global warming will cause a melting of the polar ice caps, leading to a rise in sea level and a flooding of coastal areas during the next century. Arctic rivers now carry a significantly enhanced flow of water compared to the early 1960s (Peterson et al. 2006,

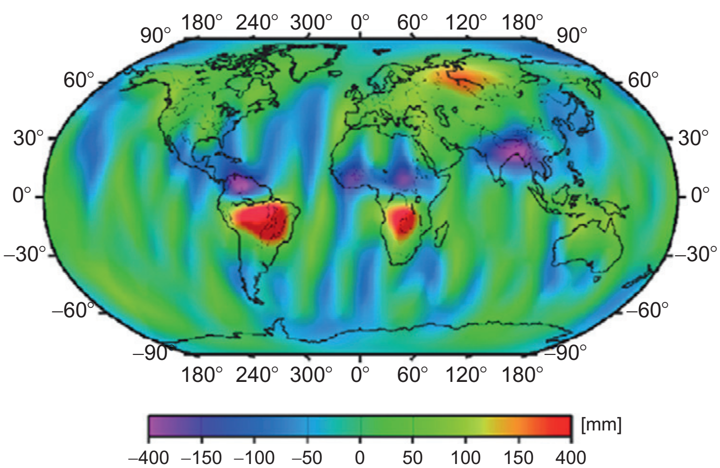

FIGURE 10.6 Changes in water content of the Earth's land surface, as measured by the GRACE satellite, April to August 2003. Note the large increase in water storage during the wet season in the Amazon. *Source: From Schmidt et al. (2006).*

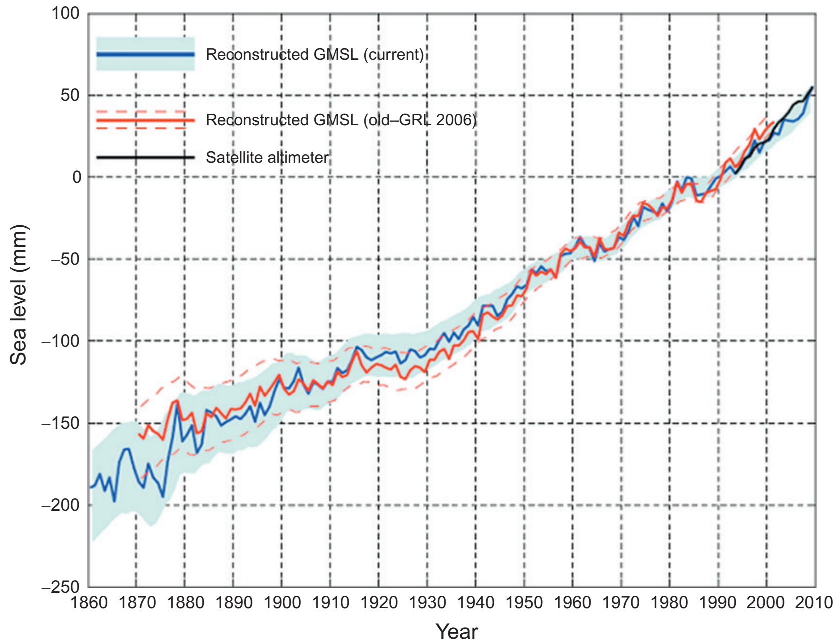

FIGURE 10.7 Global average sea level from 1860 to 2009. *Source: From Church and White (2011). Used with permission of Springer.*

McClelland et al. 2006). Deduced from a variety of methods, the long-term average rise in sea level has been between 1 to 2 mm/yr for the last 100 years (Church and White 2011, Merrifield et al. 2009, Kemp et al. 2011; Figure 10.7).

Observations of sea-level rise are complicated by the continuing isostatic adjustments of continental elevations in response to the melting of ice from the last continental glaciation (Sella et al. 2007). After removing this factor, Peltier and Tushingham (1989) found a rise of 2.4 mm/yr from 1920 to 1970, which they related to the effects of global warming on continental glaciers. Recent measurements by the TOPEX/Poseidon satellite suggest that relative sea level rose at a rate of 3.3 mm/yr for 1993 to 2009 (Cazenave and Llovel 2010), which is consistent with tide-gauge measurements during the same interval (Merrifield et al. 2009, Church and White 2011).

Sea-surface temperatures have also risen over the last 100 years (Levitus et al. 2001, 2005; Barnett et al. 2005), so some of the rise in sea level must be attributed to the thermal expansion of water at warmer temperatures (Miller and Douglas 2004, Antonov et al. 2005). Some rise in sea level may also stem from human activities, including the extraction of groundwater, which is then delivered to the sea by rivers (Sahagian et al. 1994, Konikow 2011, Pokhrel et al. 2012). The remaining rise in sea level is likely dominated by the melting of mountain glaciers throughout the world—an indication of a global warming trend (Oerlemans 2005, Meier et al. 2007, Jacob et al. 2012).

Loss of polar ice caps is associated with dramatic rise in sea level during past glacial cycles (Raymo and Mitrovica 2012, Deschamps et al. 2012). GRACE satellite measurements show that the ice pack on Greenland is in decline (van den Broeke et al. 2009), and the melting on Greenland is consistent with observations of lower salinity in North Atlantic waters (Curry and Mauritzer 2005, Dickson et al. 2002). The ice pack on Greenland shows the greatest decline at its margins and may actually be accumulating some mass in northern and interior regions (Krabill et al. 2000, Pritchard et al. 2009). However, the recent net loss of ice has accelerated to more than 250 km^3/yr (Velicogna and Wahr 2006a, Chen et al. 2006a, Rignot et al. 2011).

Overall, the massive ice cap on Antarctica shows smaller changes than in Greenland, although it also appears to be decreasing in volume (Velicogna and Wahr 2006a, Chen et al. 2009). Rapid melting on the Antarctic Peninsula may be balanced by slight accumulations of snowfall in East Antarctica (Rignot and Thomas 2002, Ramillien et al. 2006, Rignot et al. 2008). Although the record is not long, measurements from GRACE show an accelerating loss of ice mass from West Antarctica, now approaching 246 km^3/yr (Velicogna 2009, Rignot et al. 2011). The Ross Sea shows declining salinity that is consistent with these trends (Jacobs et al. 2002).

The measured losses of ice volume from Greenland, Antarctica, and continental glaciers yield a global rise in sea level of about 1.5 to 1.8 mm/yr (Meier et al. 2007, Jacob et al. 2012), about half of the observed rate. Presumably the thermal expansion of warmer seawater explains the difference. By 2100, losses from these ice packs may yield a total sea level rise >1 m, yielding widespread flooding of coastal environments and most of the world's major cities. Overall the Greenland ice pack contains the equivalent of 7 m of sea level rise, whereas Antarctica contains about 65 m. Thus, the expected melting during this century has released only a small fraction of the water in frozen ice packs.

Other human impacts on sea level include the extraction of groundwater, which adds water to the hydrologic cycle at the Earth's surface, and impoundments of water in reservoirs, which delays the delivery of water to the oceans. Impoundments now contain 8070 km^3 of water, and new dams are planned or under construction in many areas of Asia and South America (Lehner et al. 2011). Reservoirs and irrigation are thought to have reduced global riverflow by about 2% (930 km^3/yr; Biemans et al. 2011, Haddeland et al. 2006) and reduced the rise in sea level by 0.55 mm/yr (Chao et al. 2008). Groundwater extraction, on the other hand, may have increased the rise in sea level by 0.3 mm/yr (Sahagian et al. 1994, Konikow 2011).

Sea Ice

Just as the volume of a glass of water is not affected by ice cubes that may melt within it, sea level is not affected by changes in the area or volume of ice, known as *sea ice,* which floats on the ocean surface. Nevertheless, trends in sea ice are a useful index of Earth's climate. Dramatic losses of sea ice are seen in the Arctic, which may have ice-free summers within a few decades (Serreze et al. 2007, Comiso et al. 2008, Parkinson and Cavalieri 2008; Figure 10.8). The remaining ice has also lost thickness (Kwok and Rothrock 2009). In contrast, in the Southern Ocean, satellite measurements have shown relatively little change in the area of sea ice surrounding Antarctica during the past couple of decades (Zwally et al. 2002, Cavalieri and Parkinson 2008). Meanwhile, historical records of whaling near Antarctica show an

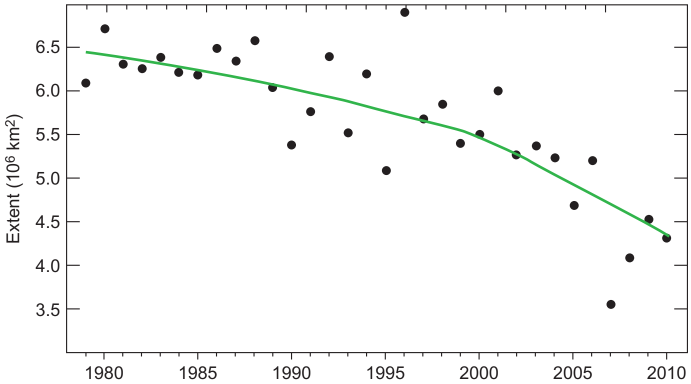

FIGURE 10.8 September sea ice extent, 1979 to 2010 in the Arctic. *Source: From Stroeve et al. (2012). Used with permission of Springer.*

increasing frequency of kills at extreme southern latitudes, consistent with a long-term decline in the extent of sea ice in the Southern Ocean (de la Mare 1997, Cotte and Guinet 2007). Loss of the natural reflectivity or albedo of snow cover and sea ice in the Arctic is estimated to have contributed 0.1 to 0.45 W m^{-2} to global warming in the interval from 1979 to 2008—reinforcing the ongoing climate change due to the accumulation of greenhouse gases in the atmosphere (Hudson 2011, Flanner et al. 2011).[2]

Terrestrial Water Balance

In response to future global warming, most climate models predict a more humid world, in which the movements of water in the hydrologic cycle through evaporation, precipitation, and runoff are enhanced (Loaiciga et al. 1996, Huntington 2006). Increased cloudiness may moderate the degree of warming, but a new steady state in Earth's temperature will be found at a higher value than that of today (Chapter 3). Not all areas of the land will be affected equally. Most of the anticipated temperature change is confined to high latitudes, and Manabe and Wetherald (1986) show that large areas of the central United States and Asia may experience a reduction in soil moisture, leading to more arid conditions. Due to the thermal buffering capacity of water, the oceans may warm more slowly than the land surface. Because most precipitation is generated from the oceans, land areas may experience severe drought during the transient period of global warming (Rind et al. 1990, Dirmeyer and Shulka 1996). Such changes in precipitation and temperature will lead to large-scale adjustments in the distribution of vegetation and global net primary production (Emanuel et al. 1985b, Smith et al. 1992b, Neilson and Marks 1994).

Analyzing the rainfall records of 1487 weather stations, Bradley et al. (1987) found an increase in precipitation over most of the midlatitudes in the Northern Hemisphere during

[2] Compare to the 2.3 W m^{-2} estimated to derive from the current accumulation of greenhouse gases from human activities (Chapter 3).

a 30- to 40-year period—consistent with changes expected for a warmer planet. Measured changes in global precipitation are small, since some areas with greater precipitation partially compensate for those reporting lesser amounts (Smith et al. 2006). Wentz et al. (2007) report a 2.8% increase in global precipitation between 1987 and 2006 (compare Dai et al. 1997).

In many areas, precipitation also seems to be becoming more variable; that is, floods and droughts are more frequent—consistent with the predictions of several general circulation models of future climate (Min et al. 2011). Over much of the world, the historical record of precipitation is scanty, and we must hope that global estimates of precipitation will improve dramatically with further application of satellite remote sensing (Petty 1995), especially the anticipated launch of the Global Precipitation Measurement (GPM) satellite in 2013.[3] Because water vapor absorbs microwave energy, the relative transmission of microwave radiation through the atmosphere is related to water vapor content and rainfall, and satellite remote sensing of the microwave emission from Earth can measure the rainfall and soil moisture over large areas (e.g., Weng et al. 1994).

One might expect that a warmer land surface would increase rates of evaporation and a warmer atmosphere might contain more water vapor. Water-balance studies of the terrestrial watersheds show an increase in evapotranspiration during the past 50 years (Walter et al. 2004a), correlated with an increased occurrence of drought (Dai et al. 2004), especially in the southwestern United States (Andreadis and Lettenmaier 2006, Seager et al. 2007). Analyses of long-term records show increasing evaporation in recent years (Szilagyi et al. 2001, Golubev et al. 2001, Brutsaert 2006). Global evapotranspiration has increased during most of the past 25 years, except during periods of drought associated with El Niño events (Jung et al. 2010). There are also indications of increases in humidity (Willett et al. 2007), particularly in regions where irrigated agriculture is widespread (Sorooshian et al. 2011).

Greater precipitation should lead to greater runoff from land (Miller and Russell 1992). Probst and Tardy (1987) found a 3% increase in stream flow in major world rivers over the last 65 years (compare Labat et al. 2004, McCabe and Wolock 2002, Syed et al. 2010). This increased stream flow may be an indication of global climate change, but it may also relate to the human destruction of vegetation leading to greater runoff (DeWalle et al. 2000, Brown et al. 2005, Rost et al. 2008, Peel et al. 2010). We might also speculate that greater stream flow is expected due to greater water-use efficiency by vegetation growing in a high-CO_2 atmosphere (Gedney et al. 2006, Betts et al. 2007; Chapter 5). In cold temperate regions, wintertime snowpack will be smaller and the spring runoff earlier in a warmer climate (Barnett et al. 2005, Burns et al. 2007).

The historical pattern of runoff for each continent and for the world as a whole shows a cyclic pattern (Probst and Tardy 1987). The cycles for the continents are not synchronous, so the trends in the global record are "damped," relative to those on each continent. Predictions of future runoff show differences between major regions of the world, with some

[3] http://pmm.nasa.gov/GPM/.

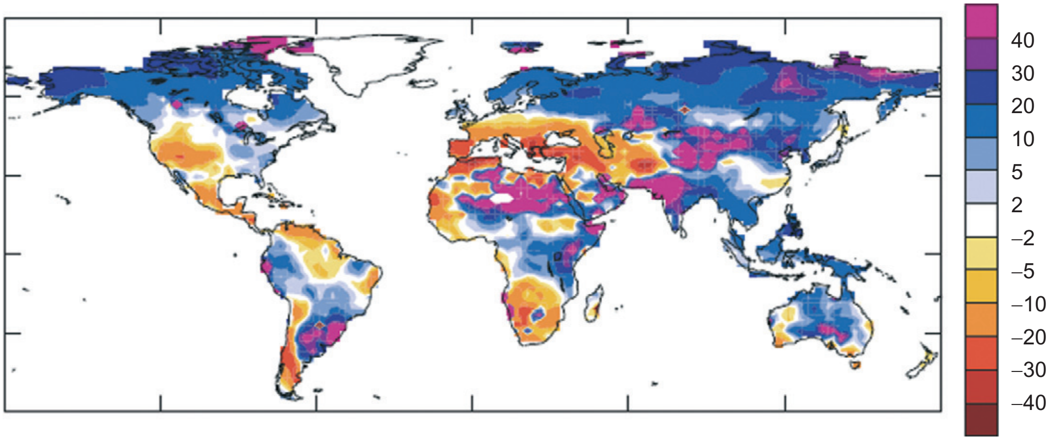

FIGURE 10.9 Projected percent changes in runoff from the Earth's land surface for 2041–2060, compared to the mean for 1900–1970. *Source: From Milly et al. (2005). Used with permission of the American Association for the Advancement of Science.*

showing increases and some showing extensive decreases (Figure 10.9). Impoundments of water in reservoirs may reduce regional impacts on the availability of water to humans.

In sum, the recent increases in precipitation, evaporation, and stream flow are consistent with predicted changes in the water cycle with global warming, but such observations must be evaluated in the context of long-term cycles in climate that have occurred through geologic time.

SUMMARY

Through evaporation and precipitation, the hydrologic cycle transfers water and heat throughout the global system. Receipt of water in precipitation is one of the primary factors controlling net primary production on land. Changes in the hydrologic cycle through geologic time are associated with changes in global temperature. All evidence suggests that movements in the hydrologic cycle were slower in glacial time and that they would be likely to increase with global warming. Movements of water on the surface of the Earth affect the rate of rock weathering and other biogeochemical phenomena.

Management of water resources will be critical to sustain the world's growing population. Data suggest that there is substantial room for improved management options. People use water very differently in different nations, with direct and indirect per capita consumption ranging from 700 m^3/yr in China to 2480 m^3/yr in the United States (Hoekstra and Chapagain 2007). Decreases in the per capita availability of water are likely to accompany changes in climate and human population during the next several decades, leaving substantial areas of the world with an increasing scarcity of water (Vörösmarty et al. 2000; Figure 10.10).

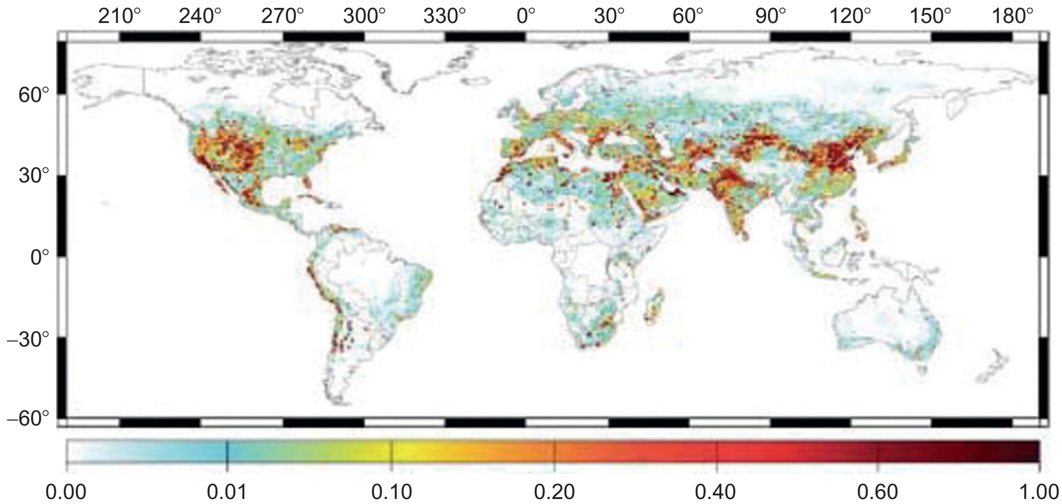

FIGURE 10.10 A water scarcity index for the Earth's land surface. Water scarcity is defined as the withdrawal from surface water divided by recycling, adjusted for regional supplements from desalinization. *Source: From Oki and Kanae (2006).*

Recommended Readings

Baumgartner, A., and E. Reichel. 1975. *The World Water Balance*. R. Olenburg.
Browning, K.A., and R.J. Gurney (Eds.). 1999. *Global Energy and Water Cycles*. Cambridge University Press.
National Academy of Sciences. 2012. *Challenges and Opportunities in the Hydrologic Sciences*. National Academy Press.
Oliver, H.R., and S.A. Oliver. 1995. *The Role of Water and the Hydrologic Cycle in Global Change*. Springer-Verlag.
Sumner, G. 1988. *Precipitation: Process and Analysis*. Wiley.
Van der Leeden, F., F.L. Troise, and D.K. Todd. 1990. *The Water Encyclopedia*. Lewis Publishers.
Ward, R.C. 1967. *The Principles of Hydrology*. McGraw-Hill.

PROBLEMS

1. If the 3% rise in global riverflow during the last 65 years (Probst and Tardy 1987) is all derived from a greater water-use efficiency of land vegetation in response to rising atmospheric CO_2, what has been the change in the water-use efficiency calculated on a global basis?
2. If the surface ocean (0–100 m) should increase in temperature by 3°C during the next century, how much will the thermal expansion of seawater contribute to a rise in sea level? (For this question, you will need to look up the equation for the thermal expansion of water as a function of temperature; try the *Handbook of Chemistry and Physics*).
3. How much larger would be the total inventory of water on Earth if some of it had not been consumed (i.e., split by photosynthesis) to produce molecular oxygen?
4. Calculate the amount of heat released to the atmosphere (calories) in the northward flow of water in the Gulf Stream ($\sim$31 Sv) as it leaves the tropical Atlantic at a temperature of 30°C and sinks to form North Atlantic Deep Water (NADW) at approximately 3°C. How does this compare to the heat that is released to Earth's atmosphere from the current annual combustion of fossil fuels?