8.10
The Contemporary Carbon Cycle

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8.10.1 INTRODUCTION

The global carbon cycle refers to the exchanges of carbon within and between four major reservoirs: the atmosphere, the oceans, land, and fossil fuels. Carbon may be transferred from one reservoir to another in seconds (e.g., the fixation of atmospheric CO_2 into sugar through photosynthesis) or over millennia (e.g., the accumulation of fossil carbon (coal, oil, gas) through deposition and diagenesis of organic matter). This chapter emphasizes the exchanges that are important over years to decades and includes those occurring over

the scale of months to a few centuries. The focus will be on the years 1980–2000 but our considerations will broadly include the years ~1850–2100. Chapter 8.09, deals with longer-term processes that involve rates of carbon exchange that are small on an annual timescale (weathering, vulcanism, sedimentation, and diagenesis).

The carbon cycle is important for at least three reasons. First, carbon forms the structure of all life on the planet, making up ~50% of the dry weight of living things. Second, the cycling of carbon approximates the flows of energy around the Earth, the metabolism of natural, human, and industrial systems. Plants transform radiant energy into chemical energy in the form of sugars, starches, and other forms of organic matter; this energy, whether in living organisms or dead organic matter, supports food chains in natural ecosystems as well as human ecosystems, not the least of which are industrial societies habituated (addicted?) to fossil forms of energy for heating, transportation, and generation of electricity. The increased use of fossil fuels has led to a third reason for interest in the carbon cycle. Carbon, in the form of carbon dioxide (CO_2) and methane (CH_4), forms two of the most important greenhouse gases. These gases contribute to a natural greenhouse effect that has kept the planet warm enough to evolve and support life (without the greenhouse effect the Earth's average temperature would be −33 °C). Additions of greenhouse gases to the atmosphere from industrial activity, however, are increasing the concentrations of these gases, enhancing the greenhouse effect, and starting to warm the Earth.

The rate and extent of the warming depend, in part, on the global carbon cycle. If the rate at which the oceans remove CO_2 from the atmosphere were faster, e.g., concentrations of CO_2 would have increased less over the last century. If the processes removing carbon from the atmosphere and storing it on land were to diminish, concentrations of CO_2 would increase more rapidly than projected on the basis of recent history. The processes responsible for adding carbon to, and withdrawing it from, the atmosphere are not well enough understood to predict future levels of CO_2 with great accuracy. These processes are a part of the global carbon cycle.

Some of the processes that add carbon to the atmosphere or remove it, such as the combustion of fossil fuels and the establishment of tree plantations, are under direct human control. Others, such as the accumulation of carbon in the oceans or on land as a result of changes in global climate (i.e., feedbacks between the global carbon cycle and climate), are not under direct human control except through controlling rates of greenhouse gas emissions and, hence, climatic change. Because CO_2 has been more important than all of the other greenhouse gases under human control, combined, and is expected to continue so in the future, understanding the global carbon cycle is a vital part of managing global climate.

This chapter addresses, first, the reservoirs and natural flows of carbon on the earth. It then addresses the sources of carbon to the atmosphere from human uses of land and energy and the sinks of carbon on land and in the oceans that have kept the atmospheric accumulation of CO_2 lower than it would otherwise have been. The chapter describes changes in the distribution of carbon among the atmosphere, oceans, and terrestrial ecosystems over the past 150 years as a result of human-induced emissions of carbon. The processes responsible for sinks of carbon on land and in the sea are reviewed from the perspective of feedbacks, and the chapter concludes with some prospects for the future.

Earlier comprehensive summaries of the global carbon cycle include studies by Bolin *et al*. (1979, 1986), Woodwell and Pecan (1973), Bolin (1981), NRC (1983), Sundquist and Broecker (1985), and Trabalka (1985). More recently, the Intergovernmental Panel on Climate Change (IPCC) has summarized information on the carbon cycle in the context of climate change (Watson *et al*., 1990; Schimel *et al*., 1996; Prentice *et al*., 2001). The basic aspects of the global carbon cycle have been understood for decades, but other aspects, such as the partitioning of the carbon sink between land and ocean, are being re-evaluated continuously with new data and analyses. The rate at which new publications revise estimates of these carbon sinks and re-evaluate the mechanisms that control the magnitude of the sinks suggests that portions of this review will be out of date by the time of publication.

8.10.2 MAJOR RESERVOIRS AND NATURAL FLUXES OF CARBON

8.10.2.1 Reservoirs

The contemporary global carbon cycle is shown in simplified form in Figure 1. The four major reservoirs important in the time frame of decades to centuries are the atmosphere, oceans, reserves of fossil fuels, and terrestrial ecosystems, including vegetation and soils. The world's oceans contain ~50 times more carbon than either the atmosphere or the world's terrestrial vegetation, and thus shifts in the abundance of carbon among the major reservoirs will have a much greater significance for the terrestrial biota and for the atmosphere than they will for the oceans.

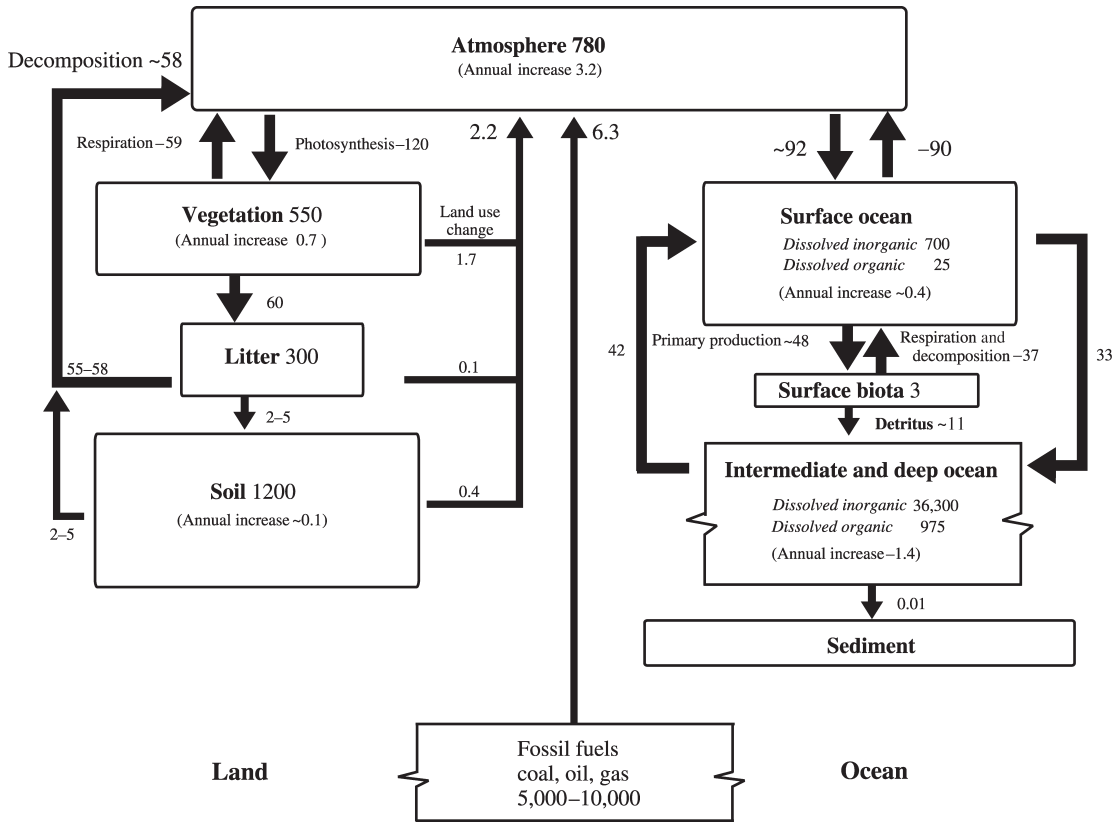

Figure 1 The contemporary global carbon cycle. Units are Pg C or Pg C yr^{-1}.

8.10.2.1.1 The atmosphere

Most of the atmosphere is made up of either nitrogen (78%) or oxygen (21%). In contrast, the concentration of CO_2 in the atmosphere is only ~0.04%. The concentrations of CO_2 in air can be measured to within one tenth of 0.1 ppmv, or 0.00001%. In the year 2000 the globally averaged concentration was ~0.0368%, or 368 ppmv, equivalent to ~780 Pg C (1 Pg = 1 petagram = 10^{15} g = 10^{9} t) (Table 1).

The atmosphere is completely mixed in about a year, so any monitoring station free of local contamination will show approximately the same year-to-year increase in CO_2. There are at least 77 stations worldwide, where weekly flask samples of air are collected, analyzed for CO_2 and other constituents, and where the resulting data are integrated into a consistent global data set (Masarie and Tans, 1995; Cooperative Atmospheric Data Integration Project—Carbon Dioxide, 1997). The stations generally show the same year-to-year increase in concentration but vary with respect to absolute concentration, seasonal variability, and other characteristics useful for investigating the global circulation of carbon.

Most of the carbon in the atmosphere is CO_2, but small amounts of carbon exist in concentrations of CH_4, carbon monoxide (CO), and non-methane hydrocarbons. These trace gases are important because they modify the chemical and/or the radiative properties of the Earth's atmosphere. Methane is present at ~1.7 ppm, two orders of magnitude more dilute than CO_2. Methane is a reduced form of carbon, is much less stable than

Table 1 Stocks and flows of carbon.

Carbon stocks (Pg C)	
Atmosphere	780
Land	2,000
Vegetation	500
Soil	1,500
Ocean	39,000
Surface	700
Deep	38,000
Fossil fuel reserves	10,000
Annual flows (Pg C yr^{-1})	
Atmosphere-oceans	90
Atmosphere-land	120
Net annual exchanges (Pg C yr^{-1})	
Fossil fuels	6
Land-use change	2
Atmospheric increase	3
Oceanic uptake	2
Other terrestrial uptake	3

CO_2, and has an average residence time in the atmosphere of 5–10 years. Carbon monoxide has an atmospheric residence time of only a few months. Its low concentration, ~0.1 ppm, and its short residence time result from its chemical reactivity with OH radicals. Carbon monoxide is not a greenhouse gas, but its chemical reactivity affects the abundances of ozone and methane which are greenhouse gases. Non-methane hydrocarbons, another unstable form of carbon in the atmosphere, are present in even smaller concentrations. The oxidation of these biogenic trace gases is believed to be a major source of atmospheric CO, and, hence, these non-methane hydrocarbons also affect indirectly the Earth's radiative balance.

8.10.2.1.2 Terrestrial ecosystems: vegetation and soils

Carbon accounts for only ~0.27% of the mass of elements in the Earth's crust (Kempe, 1979), yet it is the basis for life on Earth. The amount of carbon contained in the living vegetation of terrestrial ecosystems (550 ± 100 Pg) is somewhat less than that present in the atmosphere (780 Pg). Soils contain 2–3 times that amount (1,500–2,000 Pg C) in the top meter (Table 2) and as much as 2,300 Pg in the top 3 m (Jobbágy and Jackson, 2000). Most terrestrial carbon is stored in the vegetation and soils of the world's forests. Forests cover ~30% of the land surface and hold ~75% of the living organic carbon. When soils are included in the inventory, forests hold almost half of the carbon of the world's terrestrial ecosystems. The soils of woodlands, grasslands, tundra, wetlands, and agricultural lands store most of the rest of the terrestrial organic carbon.

8.10.2.1.3 The oceans

The total amount of dissolved inorganic carbon (DIC) in the world's oceans is $\sim 3.7 \times 10^4$ Pg, and the amount of organic carbon is ~1,000 Pg. Thus, the world's oceans contain ~50 times more carbon than the atmosphere and 70 times more than the world's terrestrial vegetation. Most of this oceanic carbon is in intermediate and deep waters; only 700–1,000 Pg C are in the surface layers of the ocean, that part of the ocean in direct contact with the atmosphere. There are also 6,000 Pg C in reactive ocean sediments (Sundquist, 1986), but the turnover of sediments is slow, and they are not generally considered as part of the active, or short-term, carbon cycle, although they are important in determining the long-term concentration of CO_2 in the atmosphere and oceans.

Carbon dioxide behaves unlike other gases in the ocean. Most gases are not very soluble in water and are predominantly in the atmosphere. For example, only ~1% of the world's oxygen is in the oceans; 99% exists in the atmosphere. Because of the chemistry of seawater, however, the distribution of carbon between air and sea is reversed: 98.5% of the carbon in the ocean–atmosphere systems is in the sea. Although this inorganic carbon is dissolved, less than 1% of it is in the form of dissolved CO_2 (p_{CO_2}); most of the inorganic carbon is in the form of bicarbonate and carbonate ions (Table 3).

About 1,000 Pg C in the oceans (out of the total of 3.8×10^4 Pg) is organic carbon. Carbon in living organisms amounts to ~3 Pg in the sea, in comparison to ~550 Pg on land. The mass of animal life in the oceans is almost the same as on land, however, pointing to the very different trophic structures in the two environments.

Table 2 Area, carbon in living biomass, and net primary productivity of major terrestrial biomes.

Biome	*Area* (10^9 ha)		*Global carbon stocks* (Pg C)						*Carbon stocks* (Mg C ha^{-1})				*NPP* (Pg C yr^{-1})	
	WBGU	*MRS*	*WGBU*			*MRS IGBP*			*WBGU*		*MRS IGPB*		*Ajtay*	*MRS*
			Plants	*Soil*	*Total*	*Plants*	*Soil*	*Total*	*Plants*	*Soil*	*Plants*	*Soil*		
Tropical forests	17.6	17.5	212	216	428	340	214	553	120	123	194	122	13.7	21.9
Temperate forests	1.04	1.04	59	100	159	139	153	292	57	96	134	147	6.5	8.1
Boreal forests	1.37	1.37	88	471	559	57	338	395	64	344	42	247	3.2	2.6
Tropical savannas and grasslands	2.25	2.76	66	264	330	79	247	326	29	117	29	90	17.7	14.9
Temperate grasslands and shrublands	1.25	1.78	9	295	304	23	176	199	7	236	13	99	5.3	7.0
Deserts and semi-deserts	4.55	2.77	8	191	199	10	159	169	2	42	4	57	1.4	3.5
Tundra	0.95	0.56	6	121	127	2	115	117	6	127	4	206	1.0	0.5
Croplands	1.60	1.35	3	128	131	4	165	169	2	80	3	122	6.8	4.1
Wetlands	0.35		15	225	240				43	643			4.3	
Total	15.12	14.93	466	2,011	2,477	654	1,567	2,221					59.9	62.6

Source: Prentice *et al*. (2001).

Table 3 The distribution of 1,000 CO_2 molecules in the atmosphere–ocean.

Atmosphere	15
Ocean	985
CO_2	5
HCO_3^-	875
CO_3^{2-}	105
Total	1,000

Source: Sarmiento (1993).

The ocean's plants are microscopic. They have a high productivity, but the production does not accumulate. Most is either grazed or decomposed in the surface waters. Only a fraction (~25%) sinks into the deeper ocean. In contrast, terrestrial plants accumulate large amounts of carbon in long-lasting structures (trees). The distribution of organic carbon between living and dead forms of carbon is also very different on land and in the sea. The ratio is ~1 : 3 on land and ~1 : 300 in the sea.

8.10.2.1.4 Fossil fuels

The common sources of energy used by industrial societies are another form of organic matter, so-called fossil fuels. Coal, oil, and natural gas are the residuals of organic matter formed millions of years ago by green plants. The material escaped oxidation, became buried in the Earth, and over time was transformed to a (fossil) form. The energy stored in the chemical bonds of fossil fuels is released during combustion just as the energy stored in carbohydrates, proteins, and fats is released during respiration.

The difference between the two forms of organic matter (fossil and nonfossil), from the perspective of the global carbon cycle, is the rate at which they are cycled. The annual rate of formation of fossil carbon is at least 1,000 times slower than rates of photosynthesis and respiration. The formation of fossil fuels is part of a carbon cycle that operates over millions of years, and the processes that govern the behavior of this long-term system (sedimentation, weathering, vulcanism, seafloor spreading) are much slower from those that govern the behavior of the short-term system. Sedimentation of organic and inorganic carbon in the sea, e.g., is ~0.2 Pg C yr^{-1}. In contrast, hundreds of petagrams of carbon are cycled annually among the reservoirs of the short-term, or active, carbon cycle. This short-term system operates over periods of seconds to centuries. When young (nonfossil) organic matter is added to or removed from the atmosphere, the total amount of carbon in the active system is unchanged. It is merely redistributed among reservoirs. When fossil fuels are oxidized, however, the CO_2 released represents a net increase in the amount of carbon in the active system.

The amount of carbon stored in recoverable reserves of coal, oil, and gas is estimated to be 5,000–10,000 Pg C, larger than any other reservoir except the deep sea, and ~10 times the carbon content of the atmosphere. Until ~1850s this reservoir of carbon was not a significant part of the short-term cycle of carbon. The industrial revolution changed that.

8.10.2.2 The Natural Flows of Carbon

Carbon dioxide is chemically stable and has an average residence time in the atmosphere of about four years before it enters either the oceans or terrestrial ecosystems.

8.10.2.2.1 Between land and atmosphere

The inorganic form of carbon in the atmosphere (CO_2) is fixed into organic matter by green plants using energy from the Sun in the process of photosynthesis, as follows:

$$6CO_2 + 6H_2O \leftrightarrows C_6H_{12}O_6 + 6O_2$$

The reduction of CO_2 to glucose ($C_6H_{12}O_6$) stores some of the Sun's energy in the chemical bonds of the organic matter formed. Glucose, cellulose, carbohydrates, protein, and fats are all forms of organic matter, or reduced carbon. They all embody energy and are nearly all derived ultimately from photosynthesis.

The reaction above also goes in the opposite direction during the oxidation of organic matter. Oxidation occurs during the two, seemingly dissimilar but chemically identical, processes of respiration and combustion. During either process the chemical energy stored in organic matter is released. Respiration is the biotic process that yields energy from organic matter, energy required for growth and maintenance. All living organisms oxidize organic matter; only plants and some microbes are capable of reducing CO_2 to produce organic matter.

Approximately 45–50% of the dry weight of organic matter is carbon. The organic carbon of terrestrial ecosystems exists in many forms, including living leaves and roots, animals, microbes, wood, decaying leaves, and soil humus. The turnover of these materials varies from less than one year to more than 1,000 years. In terms of carbon, the world's terrestrial biota is almost entirely vegetation; animals (including humans) account for less than 0.1% of the carbon in living organisms.

Each year the atmosphere exchanges ~120 Pg C with terrestrial ecosystems through photosynthesis and respiration (Figure 1 and Table 1). The uptake of carbon through photosynthesis is gross primary production (GPP). At least half of this production

is respired by the plants, themselves (autotrophic respiration (Rs_a)), leaving a net primary production (NPP) of $\sim$60 Pg C yr^{-1}. Recent estimates of global terrestrial NPP vary between 56.4 Pg C yr^{-1}and 62.6 Pg C yr^{-1} (Ajtay *et al*., 1979; Field *et al*., 1998; Saugier *et al*., 2001). The annual production of organic matter is what fuels the nonplant world, providing food, feed, fiber, and fuel for natural and human systems. Thus, most of the NPP is consumed by animals or respired by decomposer organisms in the soil (heterotrophic respiration (Rs_h)). A smaller amount ($\sim$4 Pg C yr^{-1} globally) is oxidized through fires. The sum of autotrophic and heterotrophic respiration is total respiration or ecosystem respiration (Rs_e). In steady state the net flux of carbon between terrestrial ecosystems and the atmosphere (net ecosystem production (NEP)) is approximately zero, but year-to-year variations in photosynthesis and respiration (including fires) may depart from this long-term balance by as much as 5–6 Pg C yr^{-1}. The annual global exchanges may be summarized as follows:

$$\mathrm{NPP} = \mathrm{GPP} - \mathrm{Rs_a}$$
$$(\sim 60 = 120 - 60\ \mathrm{Pg\,C\,yr^{-1}})$$

$$\mathrm{NEP} = \mathrm{GPP} - \mathrm{Rs_a} - \mathrm{Rs_h}$$
$$(\sim 0 = 120 - 60 - 60\ \mathrm{Pg\,C\,yr^{-1}})$$

$$\mathrm{NEP} = \mathrm{NPP} - \mathrm{Rs_h}$$
$$(\sim 0 = 60 - 60\ \mathrm{Pg\,C\,yr^{-1}})$$

Photosynthesis and respiration are not evenly distributed either in space or over the course of a year. About half of terrestrial photosynthesis occurs in the tropics where the conditions are generally favorable for growth, and where a large proportion of the Earth's land area exists (Table 2). Direct evidence for the importance of terrestrial metabolism (photosynthesis and respiration) can be seen in the effect it has on the atmospheric concentration of CO_2 (Figure 2(a)). The most striking feature of the figure is the regular sawtooth pattern. This pattern repeats itself annually. The cause of the oscillation is the metabolism of terrestrial ecosystems. The highest concentrations occur at the end of each winter, following the season in which respiration has exceeded photosynthesis and thereby caused a net release of CO_2 to the atmosphere. Lowest concentrations occur at the end of each summer, following the season in which photosynthesis has exceeded respiration and drawn CO_2 out of the atmosphere. The latitudinal variability in the amplitude of this oscillation suggests that it is driven largely by northern temperate and boreal ecosystems: the highest amplitudes (up to $\sim$16 ppmv) are in the northern hemisphere with the largest land area. The phase of the amplitude is reversed in the southern hemisphere, corresponding to seasonal terrestrial metabolism there. Despite the high rates of production and respiration in the tropics, equatorial regions are thought to contribute little to this oscillation. Although there is a strong seasonality in precipitation throughout much of the tropics, the seasonal changes in moisture affect photosynthesis and respiration almost equally and thus the two processes remain largely in phase with little or no net flux of CO_2.

8.10.2.2.2 Between oceans and atmosphere

There is $\sim$50 times more carbon in the ocean than in the atmosphere, and it is the amount of DIC in the ocean that determines the atmospheric concentration of CO_2. In the long term (millennia) the most important process determining the exchanges of carbon between the oceans and the atmosphere is the chemical equilibrium of dissolved CO_2, bicarbonate, and carbonate in the ocean. The rate at which the oceans take up or release carbon is slow on a century timescale, however, because of lags in circulation and changes in the availability of calcium ions. The carbon chemistry of seawater is discussed in more detail in the next section.

Two additional processes besides carbon chemistry keep the atmospheric CO_2 lower than it otherwise would be. One process is referred to as the solubility pump and the other as the biological pump. The solubility pump is based on the fact that CO_2 is more soluble in cold waters. In the ocean, CO_2 is $\sim$2 times more soluble in the cold mid-depth and deep waters than it is in the warm surface waters near the equator. Because sinking of cold surface waters in Arctic and Antarctic regions forms these mid-depth and deep waters, the formation of these waters with high CO_2 keeps the CO_2 concentration of the atmosphere lower than the average concentration of surface waters.

The biological pump also transfers surface carbon to the intermediate and deep ocean. Not all of the organic matter produced by phytoplankton is respired in the surface waters where it is produced; some sinks out of the photic zone to deeper water. Eventually, this organic matter is decomposed at depth and reaches the surface again through ocean circulation. The net effect of the sinking of organic matter is to enrich the deeper waters relative to surface waters and thus to reduce the CO_2 concentration of the atmosphere. Marine photosynthesis and the sinking of organic matter out of the surface water are estimated to keep the concentration of CO_2 in air $\sim$30% of what it would be in their absence.

Together the two pumps keep the DIC concentration of the surface waters $\sim$10% lower

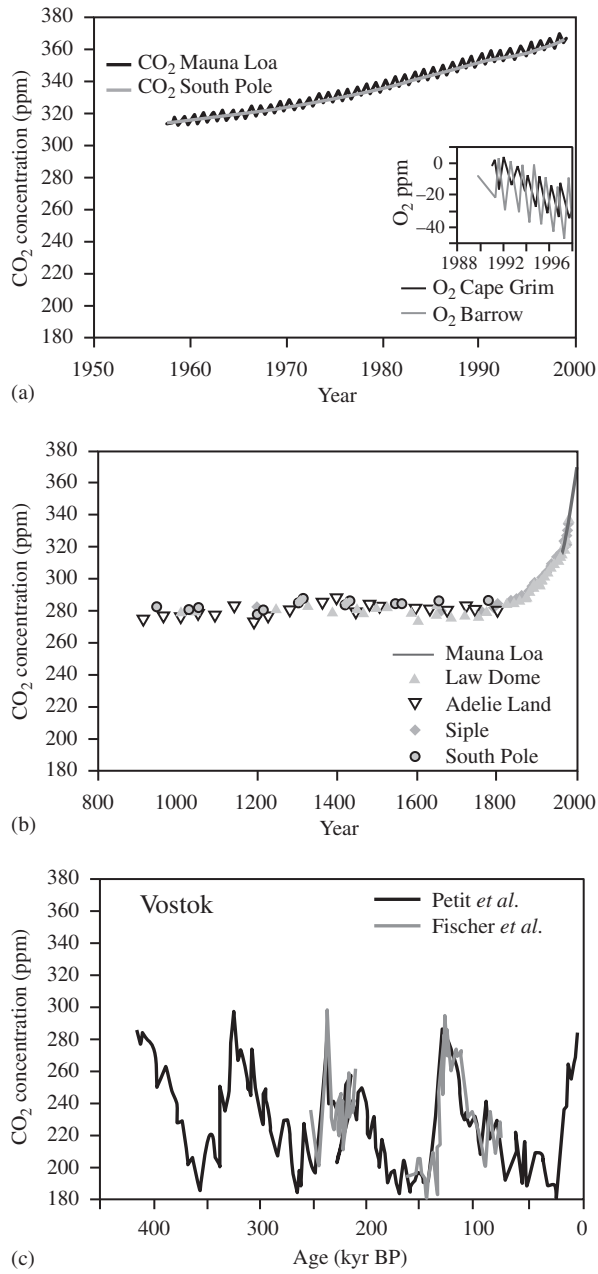

Figure 2 Concentration of CO_2 in the atmosphere: (a) over the last 42 years, (b) over the last 1,000 years, and (c) over the last $\sim 4 \times 10^5$ years (reproduced by permission of Intergovernmental Panel on Climate Change from *Climate Change 2001: The Scientific Basis*, **2001**, pp. 183–237).

than at depth. Ocean models that simulate both carbon chemistry and oceanic circulation show that the concentration of CO_2 in the atmosphere (280 ppmv pre-industrially) would have been 720 ppmv if both pumps were turned off (Sarmiento, 1993).

There is another biological pump, called the carbonate pump, but its effect in reducing the concentration of CO_2 in the atmosphere is small. Some forms of phytoplankton have $CaCO_3$ shells that, in sinking, transfer carbon from the surface to deeper water, just as the biological pump transfers organic carbon to depth. The precipitation of $CaCO_3$ in the surface waters, however, increases the partial pressure of CO_2, and the evasion of this CO_2 to the atmosphere offsets the sinking of carbonate carbon.

Although ocean chemistry determines the CO_2 concentration of the atmosphere in the long term and the solubility and biological pumps act to modify this long-term equilibrium, short-term exchanges of carbon between ocean and

atmosphere result from the diffusion of CO_2 across the air–sea interface. The diffusive exchanges transfer ~90 Pg C yr^{-1} across the air–sea interface in both directions (Figure 1). The transfer has been estimated by two different methods. One method is based on the fact that the transfer rate of naturally produced ^{14}C into the oceans should balance the decay of ^{14}C within the oceans. Both the production rate of ^{14}C in the atmosphere and the inventory of ^{14}C in the oceans are known with enough certainty to yield an average rate of transfer of ~100 Pg C yr^{-1}, into and out of the ocean.

The second method is based on the amount of radon gas in the surface ocean. Radon gas is generated by the decay of ^{226}Ra. The concentration of the parent ^{226}Ra and its half-life allow calculation of the expected radon gas concentration in the surface water. The observed concentration is ~70% of expected, so 30% of the radon must be transferred to the atmosphere during its mean lifetime of six days. Correcting for differences in the diffusivity of radon and CO_2 allows an estimation of the transfer rate for CO_2. The transfer rates given by the ^{14}C method and the radon method agree within ~10%.

The net exchange of CO_2 across the air–sea interface varies latitudinally, largely as a function of the partial pressure of CO_2 in surface waters, which, in turn, is affected by temperature, upwelling or downwelling, and biological production. Cold, high-latitude waters take up carbon, while warm, lower-latitude waters tend to release carbon (outgassing of CO_2 from tropical gyres). Although the latitudinal pattern in net exchange is consistent with temperature, the dominant reason for the exchange is upwelling (in the tropics) and downwelling, or deep-water formation (at high latitudes).

The annual rate of photosynthesis in the world oceans is estimated to be ~48 Pg C (Table 4) (Longhurst *et al*., 1995). About 25% of the primary production sinks from the photic zone to deeper water (Falkowski *et al*., 1998; Laws *et al*., 2000). The gross flows of carbon between the surface ocean and the intermediate and deep ocean are estimated to be ~40 Pg C yr^{-1}, in part from the sinking of organic production (11 Pg C yr^{-1}) and in part from physical mixing (33 Pg C yr^{-1}) (Figure 1).

Table 4 Annual net primary production of the ocean.

Domain or ecosystem	*NPP* (Pg C yr^{-1})
Trade winds domain (tropical and subtropical)	13.0
Westerly winds domain (temperate)	16.3
Polar domain	6.4
Coastal domain	10.7
Salt marshes, estuaries, and macrophytes	1.2
Coral reefs	0.7
Total	48.3

Source: Longhurst *et al*. (1995).

8.10.2.2.3 Between land and oceans

Most of the carbon taken up or lost by terrestrial ecosystems and the ocean is exchanged with the atmosphere, but a small flux of carbon from land to the ocean bypasses the atmosphere. The river input of inorganic carbon to the oceans (0.4 Pg C yr^{-1}) is almost balanced in steady state by a loss of carbon to carbonate sediments (0.2 Pg C yr^{-1}) and a release of CO_2 to the atmosphere (0.1 Pg C yr^{-1}) (Sarmiento and Sundquist, 1992). The riverine flux of organic carbon is 0.3–0.5 Pg C yr^{-1}, and thus, the total flux from land to sea is 0.4–0.7 Pg C yr^{-1}.

8.10.3 CHANGES IN THE STOCKS AND FLUXES OF CARBON AS A RESULT OF HUMAN ACTIVITIES

8.10.3.1 Changes Over the Period 1850–2000

8.10.3.1.1 Emissions of carbon from combustion of fossil fuels

The CO_2 released annually from the combustion of fossil fuels (coal, oil, and gas) is calculated from records of fuel production compiled internationally (Marland *et al*., 1998). Emissions of CO_2 from the production of cement and gas flaring add small amounts to the total industrial emissions, which have generally increased exponentially since ~1750. Temporary interruptions in the trend occurred during the two World Wars, following the increase in oil prices in 1973 and 1979, and following the collapse of the former Soviet Union in 1992 (Figure 3). Between 1751 and 2000, the total emissions of carbon are estimated to have been ~275 Pg C, essentially all of it since 1860. Annual emissions averaged 5.4 Pg C yr^{-1} during the 1980s and 6.3 Pg C yr^{-1} during the 1990s. Estimates are thought to be known globally to within 20% before 1950 and to within 6% since 1950 (Keeling, 1973; Andres *et al*., 1999).

The proportions of coal, oil, and gas production have changed through time. Coal was the major contributor to atmospheric CO_2 until the late 1960s, when the production of oil first exceeded that of coal. Rates of oil and gas consumption grew rapidly until 1973. After that their relative rates of growth declined dramatically, such that emissions of carbon from coal were, again, as large as those from oil during the

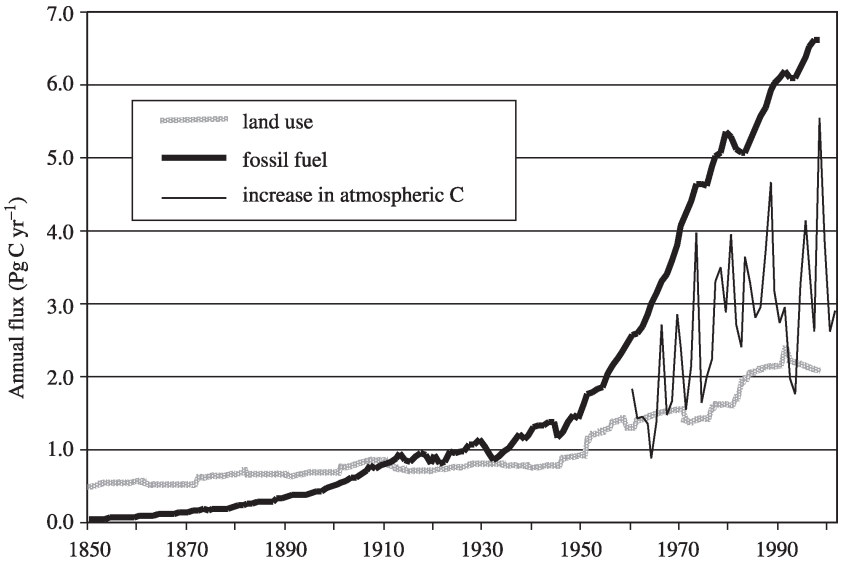

Figure 3 Annual emissions of carbon from combustion of fossil fuels and from changes in land use, and the annual increase in atmospheric CO_2 (in Pg C) since ~1750 (interannual variation in the growth rate of atmospheric CO_2 is greater than variation in emissions).

second half of the 1980s and in the last years of the twentieth century.

The relative contributions of different world regions to the annual emissions of fossil fuel carbon have also changed. In 1925, the US, Western Europe, Japan, and Australia were responsible for ~88% of the world's fossil fuel CO_2 emissions. By 1950 the fraction contributed by these countries had decreased to 71%, and by 1980 to 48%. The annual rate of growth in the use of fossil fuels in developed countries varied between 0.5% and 1.4% in the 1970s. In contrast, the annual rate of growth in developing nations was 6.3% during this period. The share of the world's total fossil fuel used by the developing countries has grown from 6% in 1925, to 10% in 1950, to ~20% in 1980. By 2020, the developing world may be using more than half of the world's fossil fuels annually (Goldemberg *et al*., 1985). They may then be the major source of both fossil fuel and terrestrial CO_2 to the atmosphere (Section 8.10.3.1.5).

Annual emissions of CO_2 from fossil fuel combustion are small relative to the natural flows of carbon through terrestrial photosynthesis and respiration (~120 Pg C yr^{-1}) and relative to the gross exchanges between oceans and atmosphere (~90 Pg C yr^{-1}) (Figure 1). Nevertheless, these anthropogenic emissions are the major contributor to increasing concentrations of CO_2 in the atmosphere. They represent a transfer of carbon from the slow carbon cycle (see Chapter 8.09) to the active carbon cycle.

8.10.3.1.2 The increase in atmospheric CO_2

Numerous measurements of atmospheric CO_2 concentrations were made in the nineteenth century (Fraser *et al*., 1986), and Callendar (1938) estimated from these early measurements that the amount of CO_2 had increased by 6% between 1900 and 1935. Because of geographical and seasonal variations in the concentrations of CO_2, however, no reliable measure of the rate of increase was possible until after 1957 when the first continuous monitoring of CO_2 concentrations was begun at Mauna Loa, Hawaii, and at the South Pole (Keeling *et al*., 2001). In 1958 the average concentration of CO_2 in air at Mauna Loa was ~315 ppm. In the year 2000 the concentration had reached ~368 ppm, yielding an average rate of increase of ~1 ppm yr^{-1} since 1958. However, in recent decades the rate of increase in the atmosphere has been ~1.5 ppm yr^{-1} (~3 Pg C yr^{-1}).

During the early 1980s, scientists developed instruments that could measure the concentration of atmospheric CO_2 in bubbles of air trapped in glacial ice. Ice cores from Greenland and Antarctica show that the pre-industrial concentration of CO_2 was between 275 ppm and 285 ppm (Neftel *et al*., 1985; Raynaud and Barnola, 1985; Etheridge *et al*., 1996) (Figure 2(b)). The increase between 1700 and 2000, therefore, has been ~85 ppm, equivalent to ~175 Pg C, or 30% of the pre-industrial level.

Over the last 1,000 years the concentration of CO_2 in the atmosphere has varied by less than 10 ppmv (Figure 2(b)). However, over the last 4.2×10^5 years (four glacial cycles), the concentration of CO_2 has consistently varied from ~180 ppm during glacial periods to ~280 ppm during interglacial periods (Figure 2(c)). The correlation between CO_2 concentration and the surface temperature of the Earth is evidence for the greenhouse effect of CO_2, first advanced almost a century ago by the Swedish climatologist Arrhenius (1896). As a greenhouse gas, CO_2 is more transparent to the Sun's energy entering the Earth's atmosphere than it is to the re-radiated heat energy leaving the Earth. Higher concentrations of CO_2 in the atmosphere cause a warmer Earth and lower concentrations a cooler one. There have been abrupt changes in global temperature that were not associated with a change in CO_2 concentrations (Smith *et al*., 1999), but never in the last 4.2×10^5 years have concentrations of CO_2 changed without a discernible change in temperature (Falkowski *et al*., 2000). The glacial–interglacial difference of 100 ppm corresponds to a temperature difference of ~10 °C. The change reflects temperature changes in the upper troposphere and in the region of the ice core (Vostoc, Antarctica) and may not represent a global average. Today's CO_2 concentration of 368 ppm represents a large departure from the last 4.2×10^5 years, although the expected increase in temperature has not yet occurred.

It is impossible to say that the increase in atmospheric CO_2 is entirely the result of human activities, but the evidence is compelling. First, the known sources of carbon are more than adequate to explain the observed increase in the atmosphere. Balancing the global carbon budget requires additional carbon sinks, not an unexplained source of carbon (see Section 8.10.3.1.4). Since 1850, ~275 Pg C have been released from the combustion of fossil fuels and another 155 Pg C were released as a result of net changes in land use, i.e., from the net effects of deforestation and reforestation (Section 8.10.3.1.5). The observed increase in atmospheric carbon was only 175 Pg C (40% of total emissions) over this 150-year period (Table 5).

Table 5 The global carbon budget for the period 1850 to 2000 (units are Pg C).

Fossil fuel emissions	275
Atmospheric increase	175
Oceanic uptake	140
Net terrestrial source	40
Land-use net source	155
Residual terrestrial sink	115

Second, for several thousand years preceding 1850 (approximately the start of the industrial revolution), the concentration of CO_2 varied by less than 10 ppmv (Etheridge *et al*., 1996) (Figure 2(b)). Since 1850, concentrations have increased by 85 ppmv (~30%). The timing of the increase is coincident with the annual emissions of carbon from combustion of fossil fuels and the net emissions from land-use change (Figure 3).

Third, the latitudinal gradient in CO_2 concentrations is highest at northern mid-latitudes, consistent with the fact that most of the emissions of fossil fuel are located in northern mid-latitudes. Although atmospheric transport is rapid, the signal of fossil fuel combustion is discernible.

Fourth, the rate of increase of carbon in the atmosphere and the distribution of carbon isotopes and other biogeochemical tracers are consistent with scientific understanding of the sources and sinks of carbon from fossil fuels, land, and the oceans. For example, while the concentration of CO_2 has increased over the period 1850–2000, the ^{14}C content of the CO_2 has decreased. The decrease is what would be expected if the CO_2 added to the system were fossil carbon depleted in ^{14}C through radioactive decay.

Concentrations of other carbon containing gases have also increased in the last two centuries. The increase in the concentration of CH_4 has been more than 100% in the last 100 years, from background levels of less than 0.8 ppm to a value of ~1.75 ppm in 2000 (Prather and Ehhalt, 2001). The temporal pattern of the increase is similar to that of CO_2. There was no apparent trend for the 1,000 years before 1700. Between 1700 and 1900 the annual rate of increase was ~1.5 ppbv, accelerating to 15 ppb yr^{-1} in the 1980s. Since 1985, however, the annual growth rate of CH_4 (unlike CO_2) has declined. The concentration is still increasing, but not as rapidly. It is unclear whether sources have declined or whether atmospheric sinks have increased.

Methane is released from anaerobic environments, such as the sediments of wetlands, peatlands, and rice paddies and the guts of ruminants. The major sources of increased CH_4 concentrations are uncertain but are thought to include the expansion of paddy rice, the increase in the world's population of ruminants, and leaks from drilling and transport of CH_4 (Prather and Ehhalt, 2001). Atmospheric CH_4 budgets are more difficult to construct than CO_2 budgets, because increased concentrations of CH_4 occur not only from increased sources from the Earth's surface but from decreased destruction (by OH radicals) in the atmosphere as well. The increase in atmospheric CH_4 has been more significant for the greenhouse effect than it has for the carbon budget. The doubling of CH_4 concentrations since 1700 has amounted to only ~1 ppm, in comparison to

the CO_2 increase of almost 90 ppm. Alternatively, CH_4 is, molecule for molecule, ~15 times more effective than CO_2 as a greenhouse gas. Its atmospheric lifetime is only 8–10 years, however.

Carbon monoxide is not a greenhouse gas, but its chemical effects on the OH radical affect the destruction of CH_4 and the formation of ozone. Because the concentration of CO is low and its lifetime is short, its atmospheric budget is less well understood than budgets for CO_2 and CH_4. Nevertheless, CO seems to have been increasing in the atmosphere until the late 1980s (Prather and Ehhalt, 2001). Its contribution to the carbon cycle is very small.

8.10.3.1.3 Net uptake of carbon by the oceans

As discussed above, the chemistry of carbon in seawater is such that less than 1% of the carbon exists as dissolved CO_2. More than 99% of the DIC exists as bicarbonate and carbonate anions (Table 3). The chemical equilibrium among these three forms of DIC is responsible for the high solubility of CO_2 in the oceans. It also sets up a buffer for changes in oceanic carbon. The buffer factor (or Revelle factor), ξ, is defined as follows:

$$\xi = \frac{\Delta p_{CO_2}/p_{CO_2}}{\Delta \Sigma CO_2/\Sigma CO_2}$$

where p_{CO_2} is the partial pressure of CO_2 (the atmospheric concentration of CO_2 at equilibrium with that of seawater), ΣCO_2 is total inorganic carbon (DIC), and Δ refers to the change in the variable. The buffer factor varies with temperature, but globally averages ~10. It indicates that p_{CO_2} is sensitive to small changes in DIC: a change in the partial pressure of CO_2 (p_{CO_2}) is ~10 times the change in total CO_2. The significance of this is that the storage capacity of the ocean for excess atmospheric CO_2 is a factor of ~10 lower than might be expected by comparing reservoir sizes (Table 1). The oceans will not take up 98% of the carbon released through human activity, but only ~85% of it. The increase in atmospheric CO_2 concentration by ~30% since 1850s has been associated with a change of only ~3% in DIC of the surface waters. The other important aspect of the buffer factor is that it increases as DIC increases. The ocean will become increasingly resistant to taking up carbon (see Section 8.10.4.2.1).

Although the oceans determine the concentration of CO_2 in the atmosphere in the long term, in the short term, lags introduced by other processes besides chemistry allow a temporary disequilibrium. Two processes that delay the transfer of anthropogenic carbon into the ocean are: (i) the transfer of CO_2 across the air–sea interface and (ii) the mixing of water masses within the sea. The rate of transfer of CO_2 across the air–sea interface was discussed above (Section 8.10.2.2.2). This transfer is believed to have reduced the oceanic absorption of CO_2 by ~10% (Broecker *et al*., 1979).

The more important process in slowing the oceanic uptake of CO_2 is the rate of vertical mixing within the oceans. The mixing of ocean waters is determined from measured profiles of natural ^{14}C, bomb-produced ^{14}C, bomb-produced tritium, and other tracers. Profiles of these tracers were obtained during extensive oceanographic surveys: one called Geochemical Ocean Sections (GEOSECS) carried out between 1972 and 1978), a second called Transient Tracers in the Ocean (TTO) carried out in 1981, and a third called the Joint Global Ocean Flux Study (JGOSFS) carried out in the 1990s. The surveys measured profiles of carbon, oxygen, radioisotopes, and other tracers along transects in the Atlantic and Pacific Oceans. The differences between the profiles over time have been used to calculate directly the penetration of anthropogenic CO_2 into the oceans (e.g., Gruber *et al*., 1996, described below). As of 1980, the oceans are thought to have absorbed only ~40% of the emissions (20–47%, depending on the model used; Bolin, 1986).

Direct measurement of changes in the amount of carbon in the world's oceans is difficult for two reasons: first, the oceans are not mixed as rapidly as the atmosphere, so that spatial and temporal heterogeneity is large; and, second, the background concentration of dissolved carbon in seawater is large relative to the change, so measurement of the change requires very accurate methods. Nevertheless, direct measurement of the uptake of anthropogenic carbon is possible, in theory if not practically, by two approaches. The first approach is based on measurement of changes in the oceanic inventory of carbon and the second is based on measurement of the transfer of CO_2 across the air–sea interface.

Measurement of an increase in oceanic carbon is complicated by the background concentration and the natural variability of carbon concentrations in seawater. The total uptake of anthropogenic carbon in the surface waters of the ocean is calculated by models to have been ~40 $\mu mol\,kg^{-1}$ of water. Annual changes would, of course, be much smaller than 40 $\mu mol\,kg^{-1}$, as would the increase in DIC concentrations in deeper waters, where less anthropogenic carbon has penetrated. By comparison, the background concentration of DIC in surface waters is 2,000 $\mu mol\,kg^{-1}$. Furthermore, the seasonal variability at one site off Bermuda was 30 $\mu mol\,kg^{-1}$. Against this background and variability, direct measurement of change is a challenge. Analytical techniques add to uncertainties, although current techniques are capable

of a precision of 1.5 $\mu mol\,kg^{-1}$ within a laboratory and 4 $\mu mol\,kg^{-1}$ between laboratories (Sarmiento, 1993).

A second method for directly measuring carbon uptake by the oceans, measurement of the air–sea exchange, is also made difficult by spatial and temporal variability. The approach measures the concentration of CO_2 in the air and in the surface mixed layer. The difference defines the gradient, which, together with a model that relates the exchange coefficient to wind speed, enables the rate of exchange to be calculated. An average air–sea difference (gradient) of 8 ppm, globally, is equivalent to an oceanic uptake of 2 Pg C yr^{-1} (Sarmiento, 1993), but the natural variability is greater than 10 ppm. Furthermore, the gas transfer coefficient is also uncertain within a factor of 2 (Broecker, 2001).

Because of the difficulty in measuring either changes in the ocean's inventory of carbon or the exchange of carbon across the air–sea interface, the uptake of anthropogenic carbon by the oceans is calculated with models that simulate the chemistry of carbon in seawater, the air–sea exchanges of CO_2, and oceanic circulation.

Ocean carbon models. Models of the ocean carbon cycle include three processes that affect the uptake and redistribution of carbon within the ocean: the air–sea transfer of CO_2, the chemistry of CO_2 in seawater, and the circulation or mixing of the ocean's water masses.

Three tracers have been used to constrain models. One tracer is CO_2 itself. The difference between current distribution of CO_2 in the ocean and the distribution expected without anthropogenic emissions yields an estimate of oceanic uptake (Gruber *et al*., 1996). The approach is based on changes that occur in the chemistry of seawater as it ages. With age, the organic matter present in surface waters decays, increasing the concentration of CO_2 and various nutrients, and decreasing the concentration of O_2. The hard parts ($CaCO_3$) of marine organisms also decay with time, increasing the alkalinity of the water. From data on the concentrations of CO_2, O_2, and alkalinity throughout the oceans, it is theoretically possible to calculate the increased abundance of carbon in the ocean as a result of the increased concentration in the atmosphere. The approach is based on the assumption that the surface waters were in equilibrium with the atmosphere when they sank, or, at least, that the extent of disequilibrium is known. The approach is sensitive to seasonal variation in the CO_2 concentration in these surface waters.

A second tracer is bomb ^{14}C. The distribution of bomb ^{14}C in the oceans (Broecker *et al*., 1995), together with an estimate of the transfer of $^{14}CO_2$ across the air–sea interface (Wanninkhof, 1992) (taking into account the fact that $^{14}CO_2$ equilibrates ~10 times more slowly than CO_2 across this interface), yields a constraint on uptake. A third constraint is based on the penetration of CFCs into the oceans (Orr and Dutay, 1999; McNeil *et al*., 2003).

Ocean carbon models calculate changes in the oceanic carbon inventory. When these changes, together with changes in the atmospheric carbon inventory (from atmospheric and ice core CO_2 data), are subtracted from the emissions of carbon from fossil fuels, the result is an estimate of the net annual terrestrial flux of carbon.

Most current models of the ocean reproduce the major features of oceanic carbon: the vertical gradient in DIC, the seasonal and latitudinal patterns of p_{CO_2} in surface waters, and the interannual variability in p_{CO_2} observed during El Niños (Prentice *et al*., 2001). However, ocean models do not capture the spatial distribution of ^{14}C at depth (Orr *et al*., 2001), and they do not show an interhemispheric transport of carbon that is suggested from atmospheric CO_2 measurements (Stephens *et al*., 1998). The models also have a tight biological coupling between carbon and nutrients, which seems not to have existed in the past and may not exist in the future. The issue is addressed below in Section 8.10.4.2.2.

8.10.3.1.4 Land: net exchange of carbon between terrestrial ecosystems and the atmosphere

Direct measurement of change in the amount of carbon held in the world's vegetation and soils may be more difficult than measurement of change in the oceans, because the land surface is not mixed. Not only are the background levels high (~550 Pg C in vegetation and ~1,500 in soils), but the spatial heterogeneity is greater on land than in the ocean. Thus, measurement of annual changes even as large as 3 Pg C yr^{-1}, in background levels 100 times greater, would require a very large sampling approach. Change may be measured over short intervals of a year or so in individual ecosystems by measuring fluxes of carbon, as, e.g., with the eddy flux technique (Goulden *et al*., 1996), but, again, the results must be scaled up from 1 km^2 to the ecosystem, landscape, region, and globe.

Global changes in terrestrial carbon were initially estimated by difference, i.e., by estimates of change in the other three reservoirs. Because the global mass of carbon is conserved, when three terms of the global carbon budget are known, the fourth can be determined by difference. For the period 1850–2000, three of the terms (275 Pg C released from fossil fuels, 175 Pg C accumulated in the atmosphere, and 140 Pg C taken up by the oceans) define a net

terrestrial uptake of 40 Pg C (Table 5). Temporal variations in these terrestrial sources and sinks can also be determined through inverse calculations with ocean carbon models (see Section 8.10.3.1.3). In inverse mode, models calculate the annual sources and sinks of carbon (output) necessary to produce observed concentrations of CO_2 in the atmosphere (input). Then, subtracting known fossil fuel sources from the calculated sources and sinks yields a residual flux of carbon, presumably terrestrial, because the other terms have been accounted for (the atmosphere and fossil fuels directly, the oceans indirectly). One such inverse calculation or deconvolution (Joos *et al*., 1999b) suggests that terrestrial ecosystems were a net source of carbon until ~1940 and then became a small net sink. Only in the early 1990s was the net terrestrial sink greater than 0.5 Pg C yr^{-1} (Figure 4).

8.10.3.1.5 Land: changes in land use

At least a portion of terrestrial sources and sinks can be determined more directly from the large changes in vegetation and soil carbon that result from changes in land use, such as the conversion of forests to cleared lands. Changes in the use of land affect the amount of carbon stored in vegetation and soils and, hence, affect the flux of carbon between land and the atmosphere. The amount of carbon released to the atmosphere or accumulated on land depends not only on the magnitude and types of changes in land use, but also on the amounts of carbon held in different ecosystems. For example, the conversion of grassland to pasture may release no carbon to the atmosphere because the stocks of carbon are unchanged. The net release or accumulation of carbon also depends on time lags introduced by the rates of decay of organic matter, the rates of oxidation of wood products, and the rates of regrowth of forests following harvest or following abandonment of agriculture land. Calculation of the net terrestrial flux of carbon requires knowledge of these rates in different ecosystems under different types of land use. Because there are several important forms of land use and many types of ecosystems in different parts of the world, and because short-term variations in the magnitude of the flux are important, computation of the annual flux requires a computer model.

Changes in terrestrial carbon calculated from changes in land use. Bookkeeping models (Houghton *et al*., 1983; Hall and Uhlig, 1991; Houghton and Hackler, 1995) have been used to calculate net sources and sinks of carbon resulting from land-use change in all the world's regions.

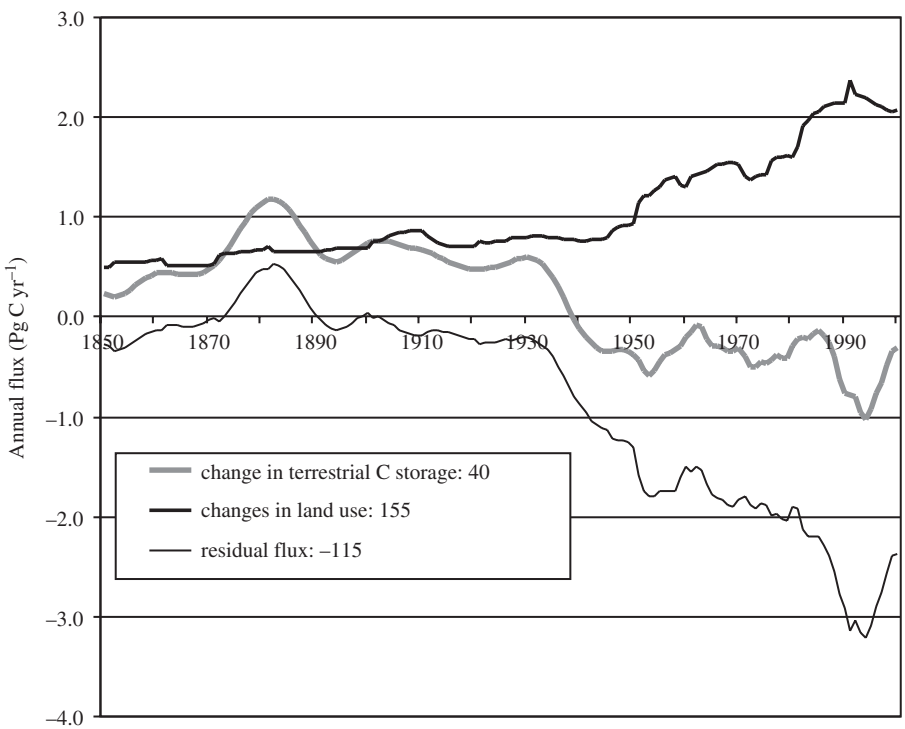

Figure 4 The net annual flux of carbon to or from terrestrial ecosystems (from inverse calculations with an ocean model (Joos *et al*., 1999b), the flux of carbon from changes in land use (from Houghton, 2003), and the difference between the net flux and the flux from land-use change (i.e., the residual terrestrial sink). Positive values indicate a source of carbon from land and negative values indicate a terrestrial sink.

Calculations are based on two types of data: rates of land-use change and per hectare changes in carbon stocks that follow a change in land use. Changes in land use are defined broadly to include the clearing of lands for cultivation and pastures, the abandonment of these agricultural lands, the harvest of wood, reforestation, afforestation, and shifting cultivation. Some analyses have included wildfire because active policies of fire exclusion and fire suppression have affected carbon storage (Houghton *et al*., 1999).

Bookkeeping models used to calculate fluxes of carbon from changes in land use track the carbon in living vegetation, dead plant material, wood products, and soils for each hectare of land cultivated, harvested, or reforested. Rates of land-use change are generally obtained from agricultural and forestry statistics, historical accounts, and national handbooks. Carbon stocks and changes in them following disturbance and growth are obtained from field studies. The data and assumptions used in the calculations are more fully documented in Houghton (1999) and Houghton and Hackler (2001).

The calculated flux is not the net flux of carbon between terrestrial ecosystems and the atmosphere, because the analysis does not consider ecosystems undisturbed by direct human activity. Rates of decay and rates of regrowth are defined in the model for different types of ecosystems and different types of land-use change, but they do not vary through time in response to changes in climate or concentrations of CO_2. The processes explicitly included in the model are the ecological processes of disturbance and recovery, not the physiological processes of photosynthesis and respiration.

The worldwide trend in land use over the last 300 years has been to reduce the area of forests, increase the area of agricultural lands, and, therefore, reduce the amount of carbon on land. Although some changes in land use increase the carbon stored on land, the net change for the 150-year period 1850–2000 is estimated to have released 156 Pg C (Houghton, 2003). An independent comparison of 1990 land cover with maps of natural vegetation suggests that another 58–75 Pg C (or ~30% of the total loss) were lost before 1850 (DeFries *et al*., 1999).

The net annual fluxes of carbon to the atmosphere from terrestrial ecosystems (and fossil fuels) are shown in Figure 3. The estimates of the net flux from land before 1800 are relatively less reliable, because early estimates of land-use change are often incomplete. However, the absolute errors for the early years are small because the fluxes themselves were small. There were no worldwide economic or cultural developments in the eighteenth century that would have caused changes in land use of the magnitude that began in the nineteenth century and accelerated to the present day. The net annual biotic flux of carbon to the atmosphere before 1800 was probably less than 0.5 Pg and probably less than 1 Pg C until ~1950.

It was not until the middle of the last century that the annual emissions of carbon from combustion of fossil fuels exceeded the net terrestrial source from land-use change. Since then the fossil fuel contribution has predominated, although both fluxes have accelerated in recent decades with the intensification of industrial activity and the expansion of agricultural area.

The major releases of terrestrial carbon result from the oxidation of vegetation and soils associated with the expansion of cultivated land. The harvest of forests for fuelwood and timber is less important because the release of carbon to the atmosphere from the oxidation of wood products is likely to be balanced by the storage of carbon in regrowing forests. The balance will occur only as long as the forests harvested are allowed to regrow, however. If wood harvest leads to permanent deforestation, the process will release carbon to the atmosphere.

In recent decades the net release of carbon from changes in land use has been almost entirely from the tropics, while the emissions of CO_2 from fossil fuels were almost entirely from outside the tropics. The highest biotic releases were not always from tropical countries. The release of terrestrial carbon from the tropics is a relatively recent phenomenon, post-1945. In the nineteenth century the major sources were from the industrialized regions—North America, Europe, and the Soviet Union—and from those regions with the greatest numbers of people—South Asia and China.

8.10.3.1.6 Land: a residual flux of carbon

The amount of carbon calculated to have been released from changes in land use since the early 1850s (156 Pg C) (Houghton, 2003) is much larger than the amount calculated to have been released using inverse calculations with global carbon models (40 Pg C) (Joos *et al*., 1999b) (Section 8.10.3.1.4). Moreover, the net source of CO_2 from changes in land use has generally increased over the past century, while the inversion approach suggests, on the contrary, that the largest releases of carbon from land were before 1930, and that since 1940 terrestrial ecosystems have been a small net sink (Figure 4).

The difference between these two estimates is greater than the errors in either one or both of the analyses, and might indicate a flux of carbon from processes not related to land-use change. The approach based on land-use change includes only the sources and sinks of carbon directly attributable to human activity; ecosystems not directly

modified by human activity are left out of the analysis (assumed neither to accumulate nor release carbon). The approach based on inverse analyses with atmospheric data, in contrast, includes all ecosystems and all processes affecting carbon storage. It yields a net terrestrial flux of carbon. The difference between the two approaches thus suggests a generally increasing terrestrial sink for carbon attributable to factors other than land-use change. Ecosystems not directly cut or cleared could be accumulating or releasing carbon in response to small variations in climate, to increased concentrations of CO_2 in air, to increased availability of nitrogen or other nutrients, or to increased levels of toxins in air and soil resulting from industrialization. It is also possible that management practices not considered in analyses of land-use change may have increased the storage of carbon on lands that have been affected by land-use change. These possibilities will be discussed in more detail below (Section 8.10.4.1). Interestingly, the two estimates (land-use change and inverse modeling) are generally in agreement before 1935 (Figure 4), suggesting that before that date the net flux of carbon from terrestrial ecosystems was largely the result of changes in land use. Only after 1935 have changes in land use underestimated the net terrestrial carbon sink. By the mid-1990s this annual residual sink had grown to $\sim$3 Pg C yr^{-1}.

8.10.3.2 Changes Over the Period 1980–2000

The period 1980–2000 deserves special attention not because the carbon cycle is qualitatively different over this period, but because scientists have been able to understand it better. Since 1980 new types of measurements and sophisticated methods of analysis have enabled better estimates of the uptake of carbon by the world's oceans and terrestrial ecosystems. The following section addresses the results of these analyses, first at the global level, and then at a regional level. Attention focuses on the two outstanding questions that have concerned scientists investigating the global carbon cycle since the first carbon budgets were constructed in the late 1960s (SCEP, 1970): (i) How much of the carbon released to the atmosphere from combustion of fossil fuels and changes in land use is taken up by the oceans and by terrestrial ecosystems? (ii) What are the mechanisms responsible for the uptake of carbon? The mechanisms for a carbon sink in terrestrial ecosystems have received considerable attention, in part because different mechanisms have different implications for future rates of CO_2 growth (and hence future global warming).

The previous section addressed the major reservoirs of the global carbon cycle, one at a time. This section addresses the methods used to determine changes in the amount of carbon held on land and in the sea, the two reservoirs for which changes in carbon are less well known. In contrast, the atmospheric increase in CO_2 and the emissions from fossil fuels are well documented. The order in which methods are presented is arbitrary. To set the stage, top-down (i.e., atmospherically based) approaches are described first, followed by bottom–up (ground-based) approaches (Table 6). Although the results of different methods often differ, the methods are not entirely comparable. Rather, they are complementary, and discrepancies sometimes suggest mechanisms responsible for transfers of carbon (Houghton, 2003; House *et al*., 2003). The results from each method are presented first, and then they are added to an accumulating picture of the global carbon cycle. Again, the emphasis is on, first, the fluxes of carbon to and from terrestrial ecosystems and the ocean and, second, the mechanisms responsible for the terrestrial carbon sink.

Table 6 Characteristics of methods use to estimate terrestrial sinks.

	Geographic limitations	*Temporal resolution*	*Attribution of mechanism(s)*	*Precision*
Inverse modeling: oceanic data	No geographic resolution	Annual	No	Moderate
Land-use models	Data limitations in some regions	Annual	Yes	Moderate
Inverse modeling: atmospheric data	Poor in tropics	Monthly to annual	No	High: North–South Low: East–West
Forest inventories	Nearly nonexistent in the tropics	5–10 years	Yes (age classes)	High for biomass; variable for soil carbon
CO_2 flux	Site specific (a few km^2); difficult to scale up	Hourly to annual	No	Some problems with windless conditions
Physiologically based models	None	Hourly to annual	Yes	Variable; difficult to validate

8.10.3.2.1 The global carbon budget

(i) Inferring changes in terrestrial and oceanic carbon from atmospheric concentrations of CO_2 and O_2. According to the most recent assessment of climate change by the IPCC, the world's terrestrial ecosystems were a net sink averaging close to zero (0.2 Pg C yr^{-1}) during the 1980s and a significantly larger sink (1.4 Pg C yr^{-1}) during the 1990s (Prentice *et al*., 2001). The large increase during the 1990s is difficult to explain. Surprisingly, the oceanic uptake of carbon was greater in the 1980s than the 1990s. The reverse would have been expected because atmospheric concentrations of CO_2 were higher in the 1990s. The estimates of terrestrial and oceanic uptake were based on changes in atmospheric CO_2 and O_2 and contained a small adjustment for the outgassing of O_2 from the oceans.

One approach for distinguishing terrestrial from oceanic sinks of carbon is based on atmospheric concentrations of CO_2 and O_2. CO_2 is released and O_2 taken up when fossil fuels are burned and when forests are burned. On land, CO_2 and O_2 are tightly coupled. In the oceans they are not, because O_2 is not very soluble in seawater. Thus, CO_2 is taken up by the oceans without any change in the atmospheric concentration of O_2. Because of this differential response of oceans and land, changes in atmospheric O_2 relative to CO_2 can be used to distinguish between oceanic and terrestrial sinks of carbon (Keeling and Shertz, 1992; Keeling *et al*., 1996b; Battle *et al*., 2000). Over intervals as short as a few years, slight variations in the seasonality of oceanic production and decay may appear as a change in oceanic O_2, but these variations cancel out over many years, making the method robust over multiyear intervals (Battle *et al*., 2000).

Figure 5 shows how the method works. The individual points show average annual global CO_2/O_2 concentrations over the years 1990–2000.

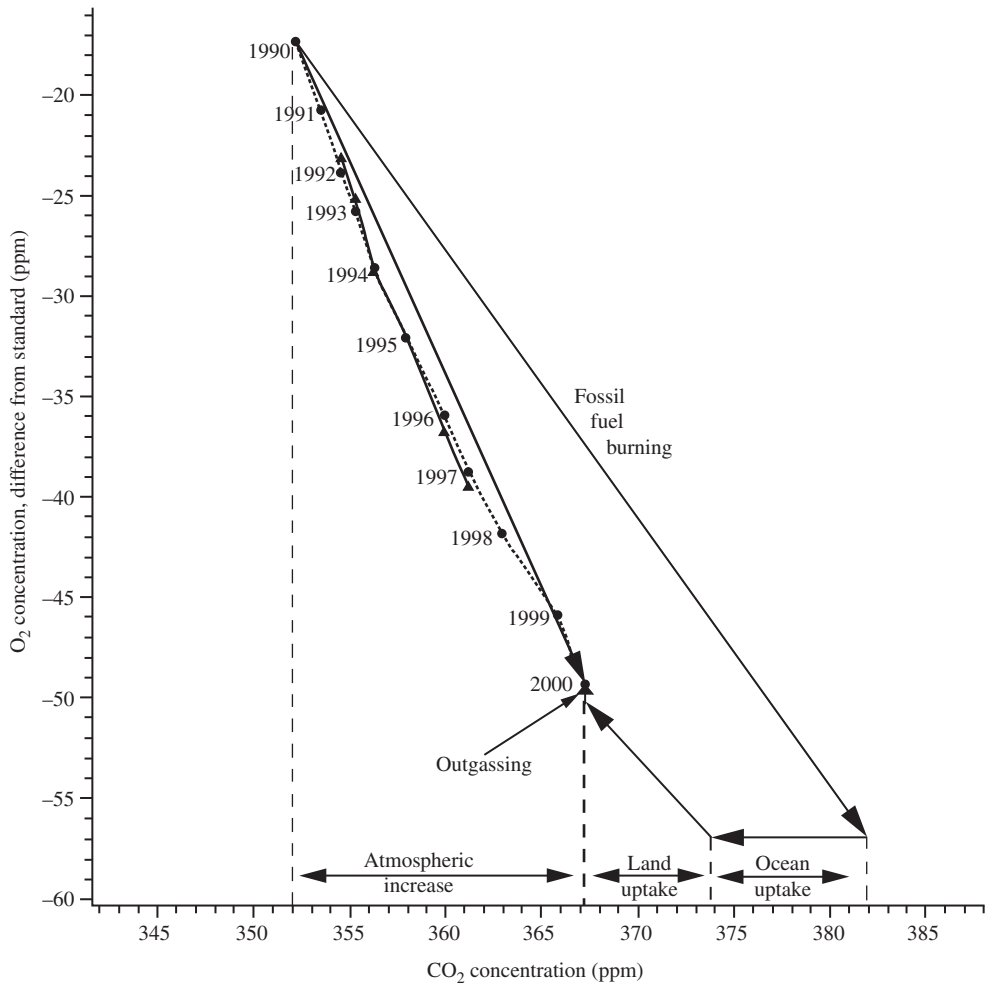

Figure 5 Terrestrial and oceanic sinks of carbon deduced from changes in atmospheric concentrations of CO_2 and O_2 (reproduced by permission of the Intergovernmental Panel on Climate Change from *Climate Change 2001: The Scientific Basis*, **2001**, pp. 183–237).

Changes in the concentrations expected from fossil fuel combustion (approximately 1 : 1) during this interval are drawn, starting in 1990. The departure of these two sets of data confirms that carbon has accumulated somewhere besides the atmosphere. The oceans are assumed not to be changing with respect to O_2, so the line for the oceanic sink is horizontal. The line for the terrestrial sink is approximately parallel to the line for fossil fuel, and drawn through 2000. The intersection of the terrestrial and the oceanic lines thus defines the terrestrial and oceanic sinks. According to the IPCC (Prentice *et al*., 2001), these sinks averaged 1.4 Pg C yr^{-1} and 1.7 Pg C yr^{-1}, respectively, for the 1990s. The estimate also included a small correction for outgassing of O_2 from the ocean (in effect, recognizing that the ocean is not neutral with respect to O_2).

Recent analyses suggest that such outgassing is significantly larger than initially estimated (Bopp *et al*., 2002; Keeling and Garcia, 2002; Plattner *et al*., 2002). The observed decadal variability in ocean temperatures (Levitus *et al*., 2000) suggests a warming-caused reduction in the transport rate of O_2 to deeper waters and, hence, an increased outgassing of O_2. The direct effect of the warming on O_2 solubility is estimated to have accounted for only 10% of the loss of O_2 (Plattner *et al*., 2002). The revised estimates of O_2 outgassing change the partitioning of the carbon sink between land and ocean. The revision increases the oceanic carbon sink of the 1990s relative to that of the 1980s (average sinks of 1.7 Pg C yr^{-1} and 2.4 Pg C yr^{-1}, respectively, for the 1980s and 1990s). The revised estimates are more consistent with estimates from ocean models (Orr, 2000) and from analyses based on $^{13}C/^{12}C$ ratios of atmospheric CO_2 (Joos *et al*., 1999b; Keeling *et al*., 2001). The revised estimate for land (a net sink of 0.7 Pg C yr^{-1} during the 1990s) (Table 7) is half of that given by the IPCC (Prentice *et al*., 2001). The decadal change in the terrestrial sink is also much smaller (from 0.4 Pg C yr^{-1} to 0.7 Pg C yr^{-1} instead of from 0.2 Pg C yr^{-1} to 1.4 Pg C yr^{-1}).

(ii) Sources and sinks inferred from inverse modeling with atmospheric transport models and atmospheric concentrations of CO_2, $^{13}CO_2$, and O_2. A second top-down method for determining oceanic and terrestrial sinks is based on spatial and temporal variations in concentrations of atmospheric CO_2 obtained through a network of flask air samples (Masarie and Tans, 1995; Cooperative Atmospheric Data Integration Project—Carbon Dioxide, 1997). Together with models of atmospheric transport, these variations are used to infer the geographic distribution of sources and sinks of carbon through a technique called inverse modeling.

Variations in the carbon isotope of CO_2 may also be used to distinguish terrestrial sources and sinks from oceanic ones. The ^{13}C isotope is slightly heavier than the ^{12}C isotope and is discriminated against during photosynthesis. Thus, trees have a lighter isotopic ratio (−22 ppt to −27 ppt) than does air (−7 ppt) (ratios are expressed relative to a standard). The burning of forests (and fossil fuels) releases a disproportionate share of the lighter isotope, reducing the isotopic ratio of $^{13}C/^{12}C$ in air. In contrast, diffusion of CO_2 across the air–sea interface does not result in appreciable discrimination, so variations in the isotopic composition of CO_2 suggest terrestrial and fossil fuels fluxes of carbon, rather than oceanic.

Spatial and temporal variations in the concentrations of CO_2, $^{13}CO_2$, and O_2 are used with models of atmospheric transport to infer (through inverse calculations) sources and sinks of carbon at the Earth's surface. The results are dependent upon the model of atmospheric transport (Figure 6; Ciais *et al*., 2000).

The interpretation of variations in ^{13}C is complicated. One complication results from isotopic disequilibria in carbon pools (Battle *et al*., 2000). Disequilibria occur because the $\delta^{13}C$ taken up by plants, e.g., is representative of the $\delta^{13}C$ currently in the atmosphere (allowing for discrimination), but the $\delta^{13}C$ of CO_2 released through decay represents not the $\delta^{13}C$ of the current atmosphere but of an atmosphere several decades ago. As long as the $\delta^{13}C$ of the atmosphere is changing, the $\delta^{13}C$ in pools will reflect a mixture of earlier and current conditions. Uncertainties in the turnover of various carbon pools add uncertainty to interpretation of the $\delta^{13}C$ signal.

Table 7 The global carbon budget (Pg C yr^{-1}).

	1980s	*1990s*
Fossil fuel emissions[a]	5.4 ± 0.3	6.3 ± 0.4
Atmospheric increase[a]	3.3 ± 0.1	3.2 ± 0.2
Oceanic uptake[b]	−1.7 + 0.6 (−1.9 ± 0.6)	−2.4 + 0.7 (−1.7 ± 0.5)
Net terrestrial flux[b]	−0.4 + 0.7 (−0.2 ± 0.7)	−0.7 + 0.8 (−1.4 ± 0.7)
Land-use change[c]	2.0 ± 0.8	2.2 ± 0.8
Residual terrestrial flux	−2.4 + 1.1 (−2.2 ± 1.1)	−2.9 + 1.1 (−3.6 ± 1.1)

[a] Source: Prentice *et al*. (2001). [b] Source: Plattner *et al*. (2002) (values in parentheses are from Prentice *et al*., 2001). [c] Houghton (2003).

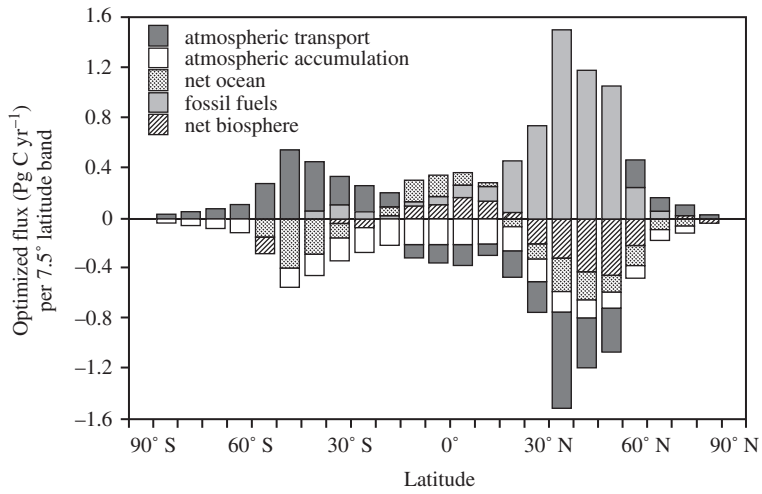

Figure 6 Terrestrial and oceanic sources and sinks of carbon inferred from inverse calculations with an atmospheric transport model and spatial and temporal variations in CO_2 concentrations. The net fluxes inferred over each region have been averaged into 7.5°-wide latitude strips (reproduced by permission of the Ecological Society of America from *Ecol. Appl.*, **2000**, *10*, 1574–1589).

Another complication results from unknown year-to-year variations in the photosynthesis of C_3 and C_4 plants (because these two types of plants discriminate differently against the heavier isotope). C_4 plants discriminate less than C_3 plants and leave a signal that looks oceanic, thus confounding the separation of land and ocean exchanges. These uncertainties of the $\delta^{13}C$ approach are most troublesome over long periods (Battle *et al*., 2000); the approach is more reliable for reconstructing interannual variations in sources and sinks of carbon.

An important distinction exists between global approaches (e.g., O_2, above) and regional inverse approaches, such as implemented with ^{13}C. In the global top-down approach, changes in terrestrial or oceanic carbon *storage* are calculated. In contrast, the regional inverse method yields *fluxes* of carbon between the land or ocean surface and the atmosphere. These fluxes of carbon include both natural and anthropogenic components. Horizontal exchange between regions must be taken into account to estimate changes in storage. For example, the fluxes will not accurately reflect changes in the amount of carbon on land or in the sea if some of the carbon fixed by terrestrial plants is transported by rivers to the ocean and respired there (Sarmiento and Sundquist, 1992; Tans *et al*., 1995; Aumont *et al*., 2001).

An example of inverse calculations is the analysis by Tans *et al*. (1990). The concentration of CO_2 near the Earth's surface is ~3 ppm higher over the northern mid-latitudes than over the southern hemisphere. The "bulge" in concentration over northern mid-latitudes is consistent with the emissions of carbon from fossil fuel combustion at these latitudes. The extent of the bulge is also affected by the rate of atmospheric mixing. High rates of mixing would dilute the bulge; low rates would enhance it. By using the latitudinal gradient in CO_2 and the latitudinal distribution of fossil fuel emissions, together with a model of atmospheric transport, Tans *et al*. (1990) determined that the bulge was smaller than expected on the basis of atmospheric transport alone. Thus, carbon is being removed from the atmosphere by the land and oceans at northern mid-latitudes. Tans *et al*. estimated removal rates averaging between 2.4 Pg C yr^{-1} and 3.5 Pg C yr^{-1} for the years 1981–1987. From p_{CO_2} measurements in surface waters, Tans *et al*. calculated that the northern mid-latitude oceans were taking up only 0.2–0.4 Pg C yr^{-1}, and thus, by difference, northern mid-latitude lands were responsible for the rest, a sink of 2.0–3.4 Pg C yr^{-1}. The range resulted from uncertainties in atmospheric transport and the limited distribution of CO_2 sampling stations. Almost no stations exist over tropical continents. Thus, Tans *et al*. (1990) could not constrain the magnitude of a tropical land source or sink, but they could determine the magnitude of the northern sink relative to a tropical source. A large tropical source, as might be expected from deforestation, implied a large northern sink; a small tropical source implied a smaller northern sink.

The analysis by Tans *et al*. (1990) caused quite a stir because their estimate for oceanic uptake was only 0.3–0.8 Pg C yr^{-1}, while analyses based on ocean models yielded estimates of 2.0 ± 0.8 Pg C yr^{-1}. The discrepancy was subsequently reconciled (Sarmiento and Sundquist, 1992) by accounting for the effect of skin temperature on the calculated air–sea exchange, the effect of atmospheric transport and oxidation

of CO on the carbon budget, and the effect of riverine transport of carbon on changes in carbon storage (see below). All of the adjustments increased the estimated oceanic uptake of carbon to values obtained by ocean models and lowered the estimate of the mid-latitude terrestrial sink.

Similar inverse approaches, using not only CO_2 concentrations but also spatial variations in O_2 and $^{13}CO_2$ to distinguish oceanic from terrestrial fluxes, have been carried out by several groups since 1990. An intercomparison of 16 atmospheric transport models (the TransCom 3 project) by Gurney *et al*. (2002) suggests average oceanic and terrestrial sinks of 1.3 Pg C yr^{-1} and 1.4 Pg C yr^{-1}, respectively, for the period 1992–1996.

The mean global terrestrial sink of 1.4 Pg C yr^{-1} for the years 1992–1996 is higher than that obtained from changes in O_2 and CO_2 (0.7 Pg C yr^{-1}) (Plattner *et al*., 2002). However, the estimate from inverse modeling has to be adjusted to account for terrestrial sources and sinks of carbon that are not "seen" by the atmosphere. For example, the fluxes inferred from atmospheric data will not accurately reflect changes in the amount of carbon on land or in the sea if some of the carbon fixed by terrestrial plants or used in weathering minerals is transported by rivers to the ocean and respired and released to the atmosphere there. Under such circumstances, the atmosphere sees a terrestrial sink and an oceanic source, while the storage of carbon on land and in the sea may not have changed. Several studies have tried to adjust atmospherically based carbon budgets by accounting for the river transport of carbon. Sarmiento and Sundquist (1992) estimated a pre-industrial net export by rivers of 0.4–0.7 Pg C yr^{-1}, balanced by a net terrestrial uptake of carbon through photosynthesis and weathering. Aumont *et al*. (2001) obtained a global estimate of 0.6 Pg C yr^{-1}. Adjusting the net terrestrial sink obtained through inverse calculations (1.4 Pg C yr^{-1}) by 0.6 Pg C yr^{-1} yields a result (0.8 Pg C yr^{-1}) similar to the estimate obtained through changes in the concentrations of O_2 and CO_2 (Table 8). The two top-down methods based on atmospheric measurements yield similar global estimates of a net terrestrial sink (~0.7 (±0.8) Pg C yr^{-1} for the 1990s).

(iii) Land-use change. Another method, independent of those based on atmospheric data and models, that has been used to estimate terrestrial sources and sinks of carbon, globally, is a method based on changes in land use (see Section 8.10.3.1.5). This is a ground-based or bottom-up approach. Changes in land use suggest that deforestation, reforestation, cultivation, and logging were responsible for a carbon source, globally, that averaged 2.0 Pg C yr^{-1} during the 1980s and 2.2 Pg C yr^{-1} during the 1990s (Houghton, 2003). The approach includes emissions of carbon from the decay of dead plant material, soil, and wood products and sinks of carbon in regrowing ecosystems, including both vegetation and soil. Analyses account for delayed sources and sinks of carbon that result from decay and regrowth following a change in land use.

Other recent analyses of land-use change give results that bound the results of this summary, although differences in the processes and regions included make comparisons somewhat misleading. An estimate by Fearnside (2000) of a 2.4 Pg C yr^{-1} source includes only the tropics. A source of 0.8 Pg C yr^{-1} estimated by McGuire *et al*. (2001) includes changes in global cropland area but does not include either the harvest of wood or the clearing of forests for pastures, both of which contributed to the net global source. The average annual release of carbon attributed by Houghton (2003) to changes in the area of croplands (1.2 Pg C yr^{-1} for the 1980s) is higher than the estimate found by McGuire *et al*. (0.8 Pg C yr^{-1}).

The calculated *source* of 2.2 (±0.8) Pg C yr^{-1} for the 1990s (Houghton, 2003) is very different from the global net terrestrial *sink* determined from top-down analyses (0.7 Pg C yr^{-1}) (Table 8). Are the methods biased? Biases in the inverse calculations may be in either direction. Because of the "rectifier effect" (the seasonal covariance

Table 8 Estimates of the annual terrestrial flux of carbon (Pg C yr^{-1}) in the 1990s according to different methods. Negative values indicate a terrestrial sink.

	O_2 and CO_2	*Inverse calculations CO_2, $^{13}CO_2$, O_2*	*Forest inventories*	*Land-use change*
Globe	−0.7 (±0.8)[a]	−0.8 (±0.8)[b]		2.2 (±0.6)[c]
Northern mid-latitudes		−2.1 (±0.8)[d]	−0.6 to −1.3[e]	−0.03 (±0.5)[c]
Tropics		1.5 (±1.2)[f]	−0.6 (±0.3)[g]	0.5 to 3.0[h]

[a] Plattner *et al*. (2002). [b] −1.4 (±0.8) from Gurney *et al*. (2002) reduced by 0.6 to account for river transport (Aumont *et al*., 2001). [c] Houghton, 2003. [d] −2.4 from Gurney *et al*. (2002) reduced by 0.3 to account for river transport (Aumont *et al*., 2001). [e] −0.65 in forests (Goodale *et al*., 2002) and another 0.0–0.65 assumed for nonforests (see text). [f] 1.2 from Gurney *et al*. (2002) increased by 0.3 to account for river transport (Aumont *et al*., 2001). [g] Undisturbed forests: −0.6 from Phillips *et al*. (1998) (challenged by Clark, 2002). [h] 0.9 (range 0.5–1.4) from DeFries *et al*. (2002) 1.3 from Achard *et al*. (2002) adjusted for soils and degradation (see text) 2.2 (±0.8) from Houghton (2003). 2.4 from Fearnside (2000).

between the terrestrial carbon flux and atmospheric transport), inverse calculations are thought to underestimate the magnitude of a northern mid-latitude sink (Denning *et al*., 1995). However, if the near-surface concentrations of atmospheric CO_2 in northern mid-latitude regions are naturally lower than those in the southern hemisphere, the apparent sink in the north may not be anthropogenic, as usually assumed. Rather, the anthropogenic sink would be less than 0.5 Pg C yr^{-1} (Taylor and Orr, 2000).

In contrast to the unknown bias of atmospheric methods, analyses based on land-use change are deliberately biased. These analyses consider only those changes in terrestrial carbon resulting directly from human activity (conversion and modification of terrestrial ecosystems). There may be other sources and sinks of carbon not related to land-use change (such as caused by CO_2 fertilization, changes in climate, or management) that are captured by other methods but ignored in analyses of land-use change. In other words, the flux of carbon from changes in land use is not necessarily the same as the net terrestrial flux from all terrestrial processes.

If the net terrestrial flux of carbon during the 1990s was 0.7 Pg C yr^{-1}, and 2.2 Pg C yr^{-1} were emitted as a result of changes in land use, then 2.9 Pg C yr^{-1} must have accumulated on land for reasons not related to land-use change. This residual terrestrial sink was discussed above (Table 7 and Figure 4). That the residual terrestrial sink exists at all suggests that processes other than land-use change are affecting the storage of carbon on land. Recall, however, that the residual sink is calculated by difference; if the emissions from land-use change are overestimated, the residual sink will also be high.

8.10.3.2.2 Regional distribution of sources and sinks of carbon: the northern mid-latitudes

Insights into the magnitude of carbon sources and sinks and the mechanisms responsible for the residual terrestrial carbon sink may be obtained from a consideration of tropical and extratropical regions separately. Inverse calculations show the tropics to be a moderate source, largely oceanic as a result of CO_2 outgassing in upwelling regions. Some of the tropical source is also terrestrial. Estimates vary greatly depending on the models of atmospheric transport and the years included in the analyses. The net global oceanic sink of 1.3 Pg C yr^{-1} for the period 1992–1996 is distributed in northern (1.2 Pg C yr^{-1}) and southern oceans (0.8 Pg C yr^{-1}), with a net source from tropical gyres (0.5 Pg C yr^{-1}) (Gurney *et al*., 2002).

The net terrestrial sink of ~0.7 Pg C yr^{-1} is not evenly distributed either. The comparison by Gurney *et al*. (2002) showed net terrestrial sinks of 2.4 ± 0.8 Pg C yr^{-1} and 0.2 Pg C yr^{-1} for northern and southern mid-latitude lands, respectively, offset to some degree by a net tropical land source of 1.2 ± 1.2 Pg C yr^{-1}. Errors are larger for the tropics than the nontropics because of the lack of sampling stations and the more complex atmospheric circulation there.

River transport and subsequent oceanic release of terrestrial carbon are thought to overestimate the magnitude of the atmospherically derived northern terrestrial sink by 0.3 Pg C yr^{-1} and underestimate the tropical source (or overestimate its sink) by the same magnitude (Aumont *et al*., 2001). Thus, the northern terrestrial sink becomes 2.1 Pg C yr^{-1}, while the tropical terrestrial source becomes 1.5 Pg C yr^{-1} (Table 8).

Inverse calculations have also been used to infer east–west differences in the distribution of sources and sinks of carbon. Such calculations are more difficult because east–west gradients in CO_2 concentration are an order of magnitude smaller than north–south gradients. Some estimates placed most of the northern sink in North America (Fan *et al*., 1998); others placed most of it in Eurasia (Bousquet *et al*., 1999a,b). More recent analyses suggest a sink in both North America and Eurasia, roughly in proportion to land area (Schimel *et al*., 2001; Gurney *et al*., 2002). The analyses also suggest that higher-latitude boreal forests are small sources rather than sinks of carbon during some years.

The types of land use determining fluxes of carbon are substantially different inside and outside the tropics (Table 9). As of early 2000s, the fluxes of carbon to and from northern lands are dominated by rotational processes, e.g., logging and subsequent regrowth. Changes in the area of forests are small. The losses of carbon from decay of wood products and slash (woody debris generated as a result of harvest) are largely offset by the accumulation of carbon in regrowing forests (reforestation and regrowth following harvest). Thus, the net flux of carbon from changes in land use is small: a source of 0.06 Pg C yr^{-1} during the 1980s changing to a sink of 0.02 Pg C yr^{-1} during the 1990s. Both the US and Europe are estimated to have been carbon sinks as a result of land-use change.

Inferring changes in terrestrial carbon storage from analysis of forest inventories. An independent estimate of carbon sources and sinks in northern mid-latitudinal lands may be obtained from forest inventories. Most countries in the northern mid-latitudes conduct periodic inventories of the growing stocks in forests. Sampling is designed to yield estimates of total growing stocks (volumes of merchantable wood) that are

Table 9 Estimates of the annual sources (+) and sinks (−) of carbon resulting from different types of land-use change and management during the 1990s (Pg C yr^{-1}).

Activity	*Tropical regions*	*Temperate and boreal zones*	*Globe*
Deforestation	2.110[a]	0.130	2.240
Afforestation	−0.100	−0.080[b]	−0.190
Reforestation (agricultural abandonment)	0[a]	−0.060	−0.060
Harvest/management	0.190	0.120	0.310
Products	0.200	0.390	0.590
Slash	0.420	0.420	0.840
Regrowth	−0.430	−0.690	−1.120
Fire suppression[c]	0	−0.030	−0.030
Nonforests			
Agricultural soils[d]	0	0.020	0.020
Woody encroachment[c]	0	−0.060	−0.060
Total	2.200	0.040	2.240

[a] Only the net effect of shifting cultivation is included here. The gross fluxes from repeated clearing and abandonment are not included.
[b] Areas of plantation forests are not generally reported in developed countries. This estimates includes only China's plantations. [c] Probably an underestimate. The estimate is for the US only, and similar values may apply in South America, Australia, and elsewhere. [d] These values include loss of soil carbon resulting from cultivation of new lands; they do not include accumulations of carbon that may have resulted from recent agricultural practices.

estimated with 95% confidence to within 1–5% (Powell *et al*., 1993; Köhl and Päivinen, 1997; Shvidenko and Nilsson, 1997). Because annual changes due to growth and mortality are small relative to the total stocks, estimates of wood volumes are relatively less precise. A study in the southeastern US determined that regional growing stocks (m^3) were known with 95% confidence to within 1.1%, while changes in the stocks ($m^3\ yr^{-1}$) were known to within 39.7% (Phillips *et al*., 2000). Allometric regressions are used to convert growing stocks (the wood contained in the boles of trees) to carbon, including all parts of the tree (roots, stumps, branches, and foliage as well as bole), nonmerchantable and small trees and nontree vegetation. Other measurements provide estimates of the carbon in the forest floor (litter) and soil. The precision of the estimates for these other pools of carbon is less than that for the growing stocks. An uncertainty analysis for 140×10^6 ha of US forests suggested an uncertainty of 0.028 Pg C yr^{-1} (Heath and Smith, 2000). The strength of forest inventories is that they provide direct estimates of wood volumes on more than one million plots throughout northern mid-latitude forests, often inventoried on 5–10-year repeat cycles. Some inventories also provide estimates of growth rates and estimates of mortality from various causes, i.e., fires, insects, and harvests. One recent synthesis of these forest inventories, after converting wood volumes to total biomass and accounting for the fate of harvested products and changes in pools of woody debris, forest floor, and soils, found a net northern mid-latitude terrestrial sink of between 0.6 Pg C yr^{-1} and 0.7 Pg C yr^{-1} for the years around 1990 (Goodale *et al*., 2002). The estimate is ~30% of the sink inferred from atmospheric data corrected for river transport (Table 8). Some of the difference may be explained if nonforest ecosystems throughout the region are also accumulating carbon. Inventories of nonforest lands are generally lacking, but in the US, at least, nonforests are estimated to account for 40–70% of the net terrestrial carbon sink (Houghton *et al*., 1999; Pacala *et al*., 2001).

It is also possible that the accumulation of carbon below ground, not directly measured in forest inventories, was underestimated and thus might account for the difference in estimates. However, the few studies that have measured the accumulation of carbon in forest soils have consistently found soils to account for only a small fraction (5–15%) of measured ecosystem sinks (Gaudinski *et al*., 2000; Barford *et al*., 2001; Schlesinger and Lichter, 2001). Thus, despite the fact that the world's soils hold 2–3 times more carbon than biomass, there is no evidence, as of early 2000s, that they account for much of a terrestrial sink.

The discrepancy between estimates obtained from forest inventories and inverse calculations might also be explained by differences in the dates of measurements. The northern sink of 2.1 Pg C yr^{-1} from Gurney *et al*. (−2.4 + 0.3 for riverine transport) is for 1992–1996 and would probably have been lower (and closer to the forest inventory-based estimate) if averaged over the entire decade (see other estimates in Prentice *et al*., 2001). Top-down measurements based on atmospheric data are sensitive to large year-to-year variations in the growth rate of CO_2 concentrations.

Both forest inventories and inverse calculations with atmospheric data show terrestrial ecosystems to be a significant carbon sink, while changes in

land use show a sink near zero. Either the analyses of land-use change are incomplete, or other mechanisms besides land-use change must be responsible for the observed sink, or some combination of both. With respect to the difference between forest inventories and land-use change, a regional comparison suggests that the recovery of forests from land-use change (abandoned farmlands, logging, fire suppression) may either overestimate or underestimate the sinks measured in forest inventories (Table 10). In Canada and Russia, the carbon sink calculated for forests recovering from harvests (land-use change) is greater than the measured sink. The difference could be error, but it is consistent with the fact that fires and insect damage increased in these regions during the 1980s and thus converted some of the boreal forests from sinks to sources (Kurz and Apps, 1999). These sources would not be counted in the analysis of land-use change, because natural disturbances were ignored. In time, recovery from these natural disturbances will increase the sink above that calculated on the basis of harvests alone, but as of early 2000s the sources from fire and insect damage exceed the net flux associated with harvest and regrowth.

In the three other regions (Table 10), changes in land use yield a sink that is smaller than measured in forest inventories. If the results are not simply a reflection of error, the failure of past changes in land use to explain the measured sink suggests that factors not considered in the analysis have enhanced the storage of carbon in the forests of the US, Europe, and China. Such factors include past natural disturbances, more subtle forms of management than recovery from harvest and agricultural abandonment (and fire suppression in the US), and environmental changes that may have enhanced forest growth. It is unclear whether the differences between estimates (changes in land use and forest inventories) are real or the result of errors and omissions. The differences are small, generally less than 0.1 Pg C yr^{-1} in any region. The likely errors and omissions in analyses of land-use change include uncertain rates of forest growth, natural disturbances, and many types of forest management (Spiecker *et al*., 1996).

8.10.3.2.3 Regional distribution of sources and sinks of carbon: the tropics

How do different methods compare in the tropics? Inverse calculations show that tropical lands were a net source of carbon, 1.2 ± 1.2 Pg C yr^{-1} for the period 1992–1996 (Gurney *et al*., 2002). Accounting for the effects of rivers (Aumont *et al*., 2001) suggests a source of 1.5 ($\pm$1.2) Pg C yr^{-1} (Table 8).

Forest inventories for large areas of the tropics are rare, although repeated measurements of permanent plots throughout the tropics suggest that undisturbed tropical forests are accumulating carbon, at least in the neotropics (Phillips *et al*., 1998). The number of such plots was too small in tropical African or Asian forests to demonstrate a change in carbon accumulation, but assuming the plots in the neotropics are representative of undisturbed forests in that region suggests a sink of 0.62 ($\pm$0.30) Pg C yr^{-1} for mature humid neotropical forests (Phillips *et al*., 1998). The finding of a net sink has been challenged, however, on the basis of systematic errors in measurement. Clark (2002) notes that many of the measurements of diameter included buttresses and other protuberances, while the allometric regressions used to estimate biomass were based on above-buttress relationships. Furthermore, these stem protuberances display disproportionate rates of radial growth. Finally, some of the plots were on floodplains where primary forests accumulate carbon. When plots with buttresses were excluded (and when recent floodplain (secondary) forests were excluded as well), the net increment was not statistically different from zero (Clark, 2002). Phillips *et al*. (2002) counter that the errors are minor, but the results remain contentious.

Thus, the two methods most powerful in constraining the northern net sink (inverse analyses and forest inventories) are weak or lacking in the tropics (Table 14), and the carbon balance of the tropics is less certain.

Direct measurement of CO_2 flux. The flux of CO_2 between an ecosystem and the atmosphere can be calculated directly by measuring the

Table 10 Annual net changes in the living vegetation of forests (Pg C yr^{-1}) in northern mid-latitude regions around the year 1990. Negative values indicate an increase in carbon stocks (i.e., a terrestrial sink).

Region	*Land-use change*[a]	*Forest inventory*[b]	*Sink from land-use change relative to inventoried sink*
Canada	−0.025	0.040	0.065 (larger)
Russia	−0.055	0.040	0.095 (larger)
USA	−0.035	−0.110	0.075 (smaller)
China	0.075	−0.040	0.115 (smaller)
Europe	−0.020	−0.090	0.070 (smaller)
Total	−0.060	−0.160	0.100 (smaller)

[a] Houghton (2003). [b] From Goodale *et al*. (2002).

covariance between concentrations of CO_2 and vertical wind speed (Goulden *et al*., 1996). The approach is being applied at ~150 sites in North America, South America, Asia, and Europe. The advantage of the approach is that it includes an integrated measure for the whole ecosystem, not only the wood or the soil. The method is ideal for determining the short-term response of ecosystems to diurnal, seasonal, and interannual variations of such variables as temperature, soil moisture, and cloudiness. If measurements are made over an entire year or over a significant number of days in each season, an annual carbon balance can be determined. The results of such measured fluxes have been demonstrated in at least one ecosystem to be in agreement with independent measurements of change in the major components of the ecosystem (Barford, 2001).

As NEP is often small relative to the gross fluxes of photosynthesis and ecosystem respiration, the net flux is sometimes less than the error of measurement. More important than error is bias, and the approach is vulnerable to bias because both the fluxes of CO_2 and the micrometeorological conditions are systematically different day and night. Wind speeds below 17 cm s^{-1} in a temperate zone forest, e.g., resulted in an underestimate of nighttime respiration (Barford *et al*., 2001). A similar relationship between nighttime wind speed and respiration in forests in the Brazilian Amazon suggests that the assumption that lateral transport is unimportant may have been invalid (Miller *et al*., in press).

Although the approach works well where micrometeorological conditions are met, the footprint for the measured flux is generally less than 1 km^2, and it is difficult to extrapolate the measured flux to large regions. Accurate extrapolations require a distribution of tower sites representative of different flux patches, but such patches are difficult to determine *a priori*. The simple extrapolation of an annual sink of 1 Mg C ha yr^{-1} (based on 55 days of measurement) for a tropical forest in Brazil to all moist forests in the Brazilian Amazon gave an estimated sink of ~1 Pg C yr^{-1} (Grace *et al*., 1995). In contrast, a more sophisticated extrapolation based on a spatial model of CO_2 flux showed a basin-wide estimate averaging only 0.3 Pg C yr^{-1} (Tian *et al*., 1998). The modeled flux agreed with the measured flux in the location of the site; spatial differences resulted from variations in modeled soil moisture throughout the basin.

Initially, support for an accumulation of carbon in undisturbed tropical forests came from measurements of CO_2 flux by eddy correlation (Grace *et al*., 1995; Malhi *et al*., 1998). Results showed large sinks of carbon in undisturbed forests, that, if scaled up to the entire tropics, yielded sinks in the range of 3.9–10.3 Pg C yr^{-1} (Malhi *et al*., 2001), much larger than the sources of carbon from deforestation. Tropical lands seemed to be a large net carbon sink. Recent analyses raise doubts about these initial results.

When flux measurements are corrected for calm conditions, the net carbon balance may be nearly neutral. One of the studies in an old-growth forest in the Tapajós National Forest, Pará, Brazil, showed a small net CO_2 source (Saleska *et al*., in press). The results in that forest were supported by measurements of biomass (forest inventory) (Rice *et al*., in press). Living trees were accumulating carbon, but the decay of downed wood released more, for a small net source. Both fluxes suggest that the stand was recovering from a disturbance several years earlier.

The observation that the rivers and streams of the Amazon are a strong source for CO_2 (Richey *et al*., 2002) may help balance the large sinks measured in some upland sites. However, the riverine source is included in inverse calculations based on atmospheric data and does not change those estimates of a net terrestrial source (Gurney *et al*., 2002).

Changes in land use in the tropics are clearly a source of carbon to the atmosphere, although the magnitude is uncertain (Detwiler and Hall, 1988; Fearnside, 2000; Houghton, 1999, 2003). The tropics are characterized by high rates of deforestation, and this conversion of forests to nonforests involves a large loss of carbon. Although rotational processes of land use, such as logging, are just as common in the tropics as in temperate zones (even more so because shifting cultivation is common in the tropics), the sinks of carbon in regrowing forests are dwarfed in the tropics by the large releases of carbon resulting from permanent deforestation.

Comparisons of results from different methods (Table 8) suggest at least two, mutually exclusive, interpretations for the net terrestrial source of carbon from the tropics. One interpretation is that a large release of carbon from land-use change (Fearnside, 2000; Houghton, 2003) is partially offset by a large sink in undisturbed forests (Malhi *et al*., 1998; Phillips *et al*., 1998, 2002). The other interpretation is that the source from deforestation is smaller (see below), and that the net flux from undisturbed forests is nearly zero (Rice *et al*., in press; Saleska *et al*., in press). Under the first interpretation, some sort of growth enhancement (or past natural disturbance) is required to explain the large current sink in undisturbed forests. Under the second, the entire net flux of carbon may be explained by changes in land use, but the source from land-use change is smaller than estimated by Fearnside (2000) or Houghton (2003).

A third possibility, that the net tropical source from land is larger than indicated by inverse

calculations (uncertain in the tropics), is constrained by the magnitude of the net sink in northern mid-latitudes. The latitudinal gradient in CO_2 concentrations constrains the difference between the northern sink and tropical source more than it constrains the absolute fluxes. The tropical source can only be larger than indicated by inverse calculations if the northern mid-latitude sink is also larger. As discussed above, the northern mid-latitude sink is thought to be in the range of 1–2.6 Pg C yr^{-1}, but the estimates are based on the assumption that the pre-industrial north–south gradient in CO_2 concentrations was zero (similar concentrations at all latitudes). No data exist for the pre-industrial north–south gradient in CO_2 concentrations, but following Keeling *et al*. (1989), Tayor and Orr extrapolated the current CO_2 gradient to a zero fossil fuel release and found a negative gradient (lower concentrations in the north). They interpreted this negative gradient as the pre-industrial gradient, and their interpretation would suggest a northern sink larger than generally believed. In contrast, Conway and Tans (1999) interpret the extrapolated zero fossil fuel gradient as representing the current sources and sinks of carbon in response to fossil fuel emissions and other human activities, such as present and past land-use change. Most investigators of the carbon cycle favor this interpretation.

The second interpretation of existing estimates (a modest source of carbon from deforestation and little or no sink in undisturbed forests) is supported by satellite-based estimates of tropical deforestation. The high estimates of Fearnside (2000) and Houghton (2003) were based on rates of deforestation reported by the FAO (2001). If these rates of deforestation are high, the estimates of the carbon source are also high. Two new studies of tropical deforestation (Achard *et al*., 2002; DeFries *et al*., 2002) report lower rates than the FAO and lower emissions of carbon than Fearnside or Houghton. The study by Achard *et al*. (2002) found rates 23% lower than the FAO for the 1990s (Table 11). Their analysis used high resolution satellite data over a 6.5% sample of tropical humid forests, stratified by "deforestation hot-spot areas" defined by experts. In addition to observing 5.8×10^6 ha of outright deforestation in the tropical humid forests, Achard *et al*. also observed 2.3×10^6 ha of degradation. Their estimated carbon flux, including changes in the area of dry forests as well as humid ones, was 0.96 Pg C yr^{-1}. The estimate is probably low because it did not include the losses of soil carbon that often occur with cultivation or the losses of carbon from degradation (reduction of biomass within forests). Soils and degradation accounted for 12% and 26%, respectively, of Houghton's (2003) estimated flux of carbon for tropical Asia and America and would yield a total flux of 1.3 Pg C yr^{-1} if the same percentages were applied to the estimate by Archard *et al*.

A second estimate of tropical deforestation (DeFries *et al*., 2002) was based on coarse resolution satellite data (8 km), calibrated with high-resolution satellite data to identify percent tree cover and to account for small clearings that would be missed with the coarse resolution data. The results yielded estimates of deforestation that were, on average, 54% lower than those reported by the FAO (Table 11). According to DeFries *et al*., the estimated net flux of carbon for the 1990s was 0.9 (range 0.5–1.4) Pg C yr^{-1}.

If the tropical deforestation rates obtained by Archard *et al*. and DeFries *et al*. were similar, there would be little doubt that the FAO estimates are high. However, the estimates are as different from each other as they are from those of the FAO (Table 11). Absolute differences between the two studies are difficult to evaluate because Achard *et al*. considered only humid tropical forests, whereas DeFries *et al*. considered all tropical forests. The greatest differences are in tropical Africa, where the percent tree cover mapped by DeFries *et al*. is most unreliable because of the large areas of savanna. Both studies suggest that the FAO estimates of tropical deforestation are high, but the rates are still in question (Fearnside and Laurance, 2003; Eva *et al*., 2003). The tropical emissions of carbon estimated by the two studies (after adjustments for degradation and soils) are

Table 11 Annual rate of change in tropical forest area[a] for the 1990s.

	Tropical humid forests			*All tropical forests*		
	FAO (2001) (10^6 ha yr^{-1})	*Achard et al. (2002)*		*FAO (2001)* (10^6 ha yr^{-1})	*DeFries et al. (2002)*	
		10^6 ha yr^{-1}	*% lower than FAO*		10^6 ha yr^{-1}	*% lower than FAO*
America	2.7	2.2	18	4.4	3.179	28
Asia	2.5	2.0	20	2.4	2.008	16
Africa	1.2	0.7	42	5.2	0.376	93
All tropics	6.4	4.9	23	12.0	5.563	54

[a] The net change in forest area is not the rate of deforestation but, rather, the rate of deforestation minus the rate of afforestation.

about half of Houghton's estimate: 1.3 Pg C yr^{-1} and 0.9 Pg C yr^{-1}, as opposed to 2.2 Pg C yr^{-1} (Table 8).

8.10.3.2.4 Summary: synthesis of the results of different methods

Top-down methods show consistently that terrestrial ecosystems, globally, were a small net sink in the 1980s and 1990s. The sink was in northern mid-latitudes, partially offset by a tropical source. The northern sink was distributed over both North America and Eurasia roughly in proportion to land area. The magnitudes of terrestrial sinks obtained through inverse calculations are larger (or the sources smaller) than those obtained from bottom-up analyses (land-use change and forest inventories). Is there a bias in the atmospheric analyses? Or are there sinks not included in the bottom-up analyses?

For the northern mid-latitudes, when estimates of change in nonforests (poorly known) are added to the results of forest inventories, the net sink barely overlaps with estimates determined from inverse calculations. Changes in land use yield smaller estimates of a sink. It is not clear how much of the discrepancy is the result of omissions of management practices and natural disturbances from analyses of land-use change, and how much is the result of environmentally enhanced rates of tree growth. In other words, how much of the carbon sink in forests can be explained by age structure (i.e., previous disturbances and management), and how much by enhanced rates of carbon storage? The question is important for predicting future concentrations of atmospheric CO_2 (see below).

In the tropics, the uncertainties are similar but also greater because inverse calculations are more poorly constrained and because forest inventories are lacking. Existing evidence suggests two possibilities. Either large emissions of carbon from land-use change are somewhat offset by large carbon sinks in undisturbed forests, or lower releases of carbon from land-use change explain the entire net terrestrial flux, with essentially no requirement for an additional sink. The first alternative (large sources and large sinks) is most consistent with the argument that factors other than land-use change are responsible for observed carbon sinks (i.e., management or environmentally enhanced rates of growth). The second alternative is most consistent with the findings of Caspersen *et al*. (2000) that there is little enhanced growth. Overall, in both northern and tropical regions changes in land use exert a dominant influence on the flux of carbon, and it is unclear whether other factors have been important in either region. These conclusions question the assumption used in predictions of climatic change, the assumption that the current terrestrial carbon sink will increase in the future (see below).

8.10.4 MECHANISMS THOUGHT TO BE RESPONSIBLE FOR CURRENT SINKS OF CARBON

8.10.4.1 Terrestrial Mechanisms

Distinguishing between regrowth and enhanced growth in the current terrestrial sink is important. If regrowth is dominant, the current sink may be expected to diminish as forests age (Hurtt *et al*., 2002). If enhanced growth is important, the magnitude of the carbon sink may be expected to increase in the future. Carbon cycle models used to calculate future concentrations of atmospheric CO_2 from emissions scenarios assume the latter (that the current terrestrial sink will increase) (Prentice *et al*., 2001). These calculated concentrations are then used in general circulation models to project future rates of climatic change. If the current terrestrial sink is largely the result of regrowth, rather than enhanced growth, future projections of climate may underestimate the extent and rate of climatic change.

The issue of enhanced growth versus regrowth can be illustrated with studies from the US. Houghton *et al*. (1999) estimated a terrestrial carbon sink of 0.15–0.35 Pg C yr^{-1} for the US, attributable to changes in land use. Pacala *et al*. (2001) revised the estimate upwards by including additional processes, but in so doing they included sinks not necessarily resulting from land-use change. Their estimate for the uptake of carbon by forests, e.g., was the uptake measured by forest inventories. The measured uptake might result from previous land use (regrowth), but it might also result from environmentally enhanced growth, e.g., CO_2 fertilization (Figure 7). If all of the accumulation of carbon in US forests were the result of recovery from past land-use practices (i.e., no enhanced growth), then the measured uptake should equal the flux calculated on the basis of land-use change. The residual flux would be zero. The study by Caspersen *et al*. (2000) suggests that such an attribution is warranted because they found that 98% of forest growth in five US states could be attributed to regrowth rather than enhanced growth. However, the analysis by Houghton *et al*. (1999) found that past changes in land use accounted for only 20–30% of the observed accumulation of carbon in trees. The uptake calculated for forests recovering from agricultural abandonment, fire suppression, and earlier harvests was only 20–30% of the uptake measured by forest inventories (~40% if the uptake attributed to woodland "thickening" (0.26 Pg C yr^{-1}; Houghton, 2003) is included (Table 12)). The results are inconsistent with

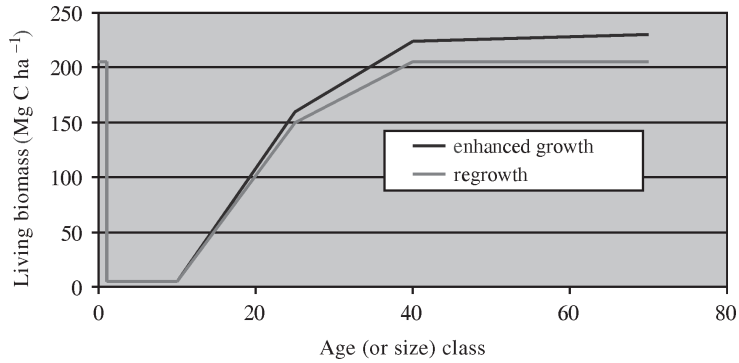

Figure 7 Idealized curves showing the difference between enhanced growth and regrowth in the accumulation of carbon in forest biomass.

Table 12 Estimated rates of carbon accumulation in the US (Pg C yr^{-1} in 1990).

	Pacala et al.[a] *(2001)*		*Houghton*[b] *et al. (1999)*	*Houghton*[b] *(2003)*	*Goodale et al. (2002)*
	Low	*High*			
Forest trees	−0.11	−0.15	−0.072[c]	−0.046[d]	−0.11
Other forest organic matter	−0.03	−0.15	0.010	0.010	−0.11
Cropland soils	0.00	−0.04	−0.138	0.00	NE
Woody encroachment	−0.12	−0.13	−0.122	−0.061	NE
Wood products	−0.03	−0.07	−0.027	−0.027	−0.06
Sediments	−0.01	−0.04	NE	NE	NE
Total sink	−0.30	−0.58	−0.35	−0.11	−0.28
% of total sink neither in forests nor wood products	43%	36%	74%	55%	NE

NE is "not estimated". Negative values indicate an accumulation of carbon on land.
[a] Pacala *et al.* (2001) also included the import/export imbalance of food and wood products and river exports. As these would create corresponding sources outside the US, they are ignored here. [b] Includes only the direct effects of human activity (i.e., land-use change and some management). [c] 0.020 Pg C yr^{-1} in forests and 0.052 Pg C yr^{-1} in the thickening of western pine woodlands as a result of early fire suppression. [d] 0.020 Pg C yr^{-1} in forests and 0.026 Pg C yr^{-1} in the thickening of western pine woodlands as a result of early fire suppression.

those of Caspersen *et al.* (2000). Houghton's analysis requires a significant growth enhancement to account for the observed accumulation of carbon in trees; the analysis by Caspersen *et al.* suggests little enhancement.

Both analyses merit closer scrutiny. Joos *et al.* (2002) have pointed out, e.g., that the relationship between forest age and wood volume (or biomass) is too variable to constrain the enhancement of growth to between 0.001% and 0.01% per year, as Caspersen *et al.* claimed. An enhancement of 0.1% per year fits the data as well. Furthermore, even a small enhancement of 0.1% per year in NPP yields a significant sink (~2 Pg C yr^{-1}) if it applies globally (Joos *et al.*, 2002). Thus, Caspersen *et al.* may have underestimated the sink attributable to enhanced growth.

However, Houghton's analysis of land-use change (Houghton *et al.*, 1999; Houghton, 2003) most likely underestimates the sink attributable to regrowth. Houghton did not consider forest management practices other than harvest and subsequent regrowth. Nor did he include natural disturbances, which in boreal forests are more important than logging in determining the current age structure and, hence, rate of carbon accumulation (Kurz and Apps, 1999). Forests might now be recovering from an earlier disturbance. A third reason why the sink may have been underestimated is that Houghton used net changes in agricultural area to obtain rates of agricultural abandonment. In contrast, rates of clearing and abandonment are often simultaneous and thus create larger areas of regrowing forests than would be predicted from net changes in agricultural area. It is unclear how much of the carbon sink in the US can be attributed to changes in land use and management, and how much can be attributed to enhanced rates of growth.

The mechanisms responsible for the current terrestrial sink fall into two broad categories (Table 13 and Figure 7): (i) enhanced growth from physiological or metabolic factors that affect rates of photosynthesis, respiration, growth, and decay and (ii) regrowth from past disturbances, changes in land use, or management, affecting the mortality of forest stands, the age structure of forests, and hence their rates of carbon accumulation. What evidence do we have that

Table 13 Proposed mechanisms for terrestrial carbon sinks.[a]

Metabolic or physiological mechanisms
CO_2 fertilization
N fertilization
Tropospheric ozone, acid deposition
Changes in climate (temperature, moisture)
Ecosystem mechanisms
Large-scale regrowth of forests following human disturbance (includes recovery from logging and agricultural abandonment)[b]
Large-scale regrowth of forests following natural disturbance[b]
Fire suppression and woody encroachment[b]
Decreased deforestation[b]
Improved agricultural practices[b]
Erosion and re-deposition of sediment
Wood products and landfills[b]

[a] Some of these mechanisms enhance growth; some reduce decomposition. In some cases these same mechanisms may also yield sources of carbon to the atmosphere. [b] Mechanisms included in analyses of land-use change (although not necessarily in all regions).

Table 14 Increases observed for a 100% increase in CO_2 concentrations.

Increased rates
60% increase in *photosynthesis* of young trees
33% average increase in net primary productivity (*NPP*) of crops
25% increase in *NPP* of a young pine forest
Increased stocks
14% average increase in *biomass* of grasslands and crops
~0% increase in the carbon content of mature forests

these mechanisms are important? Consider, first, enhanced rates of growth.

8.10.4.1.1 Physiological or metabolic factors that enhance rates of growth and carbon accumulation

CO_2 fertilization. Numerous reviews of the direct and indirect effects of CO_2 on photosynthesis and plant growth have appeared in the literature (Curtis, 1996; Koch and Mooney, 1996; Mooney *et al*., 1999; Körner, 2000), and only a very brief review is given here. Horticulturalists have long known that annual plants respond to higher levels of CO_2 with increased rates of growth, and the concentration of CO_2 in greenhouses is often deliberately increased to make use of this effect. Similarly, experiments have shown that most C_3 plants (all trees, most crops, and vegetation from cold regions) respond to elevated concentrations of CO_2 with increased rates of photosynthesis and increased rates of growth.

Despite the observed stimulative effects of CO_2 on photosynthesis and plant growth, it is not clear that the effects will result in an increased storage of carbon in the world's ecosystems. One reason is that the measured effects of CO_2 have generally been short term, while over longer intervals the effects are often reduced or absent. For example, plants often acclimate to higher concentrations of CO_2 so that their rates of photosynthesis and growth return to the rates observed before the concentration was raised (Tissue and Oechel, 1987; Oren *et al*., 2001).

Another reason why the experimental results may not apply to ecosystems is that most experiments with elevated CO_2 have been conducted with crops, annual plants, or tree seedlings. The few studies conducted at higher levels of integration or complexity, such as with mature trees and whole ecosystems, including soils as well as vegetation, suggest much reduced responses. Table 14 summarizes the results of experiments at different levels of integration. Arranged in this way (from biochemical processes to ecosystem processes), observations suggest that as the level of complexity, or the number of interacting processes, increases, the effects of CO_2 fertilization are reduced. This dampening of effects across ever-increasing levels of complexity has been noted since scientists first began to consider the effects of CO_2 on carbon storage (Lemon, 1977).

In other words, a CO_2-enhanced increase in photosynthesis is many steps removed from an increase in carbon storage. An increase in NPP is expected to lead to increased carbon storage until the carbon lost from the detritus pool comes into a new equilibrium with the higher input of NPP. But, if the increased NPP is largely labile (easily decomposed), then it may be decomposed rapidly with little net carbon storage (Davidson and Hirsch, 2001). Results from a loblolly pine forest in North Carolina suggest a very small increase in carbon storage. Elevated CO_2 increased litter production (with a turnover time of about three years) but did not increase carbon accumulation deeper in the soil layer (Schlesinger and Lichter, 2001). Alternatively, the observation that microbes seemed to switch from old organic matter to new organic matter after CO_2 fertilization of a grassland suggests that the loss of carbon may be delayed in older, more refractory pools of soil organic matter (Cardon *et al*., 2001).

The central question is whether natural ecosystems will accumulate carbon as a result of elevated CO_2, and whether the accumulation will persist. Few CO_2 fertilization experiments have been carried out for more than a few years in whole ecosystems, but where they have, the results generally show an initial CO_2-induced increment in biomass that diminishes after a few years. The diminution of the initial CO_2-induced

effect occurred after two years in an arctic tundra (Oechel *et al*., 1994) and after three years in a rapidly growing loblolly pine forest (Oren *et al*., 2001). Other forests may behave differently, but the North Carolina forest was chosen in part because CO_2 fertilization, if it occurs anywhere, is likely to occur in a rapidly growing forest. The longest CO_2 fertilization experiment, in a brackish wetland on the Chesapeake Bay, has shown an enhanced net uptake of carbon even after 12 years, but the expected accumulation of carbon at the site has not been observed (Drake *et al*., 1996).

Nitrogen fertilization. Human activity has increased the abundance of biologically active forms of nitrogen (NO_x and NH_4), largely through the production of fertilizers, the cultivation of legumes that fix atmospheric nitrogen, and the use of internal combustion engines. Because the availability of nitrogen is thought to limit NPP in temperate-zone ecosystems, the addition of nitrogen through human activities is expected to increase NPP and, hence, terrestrial carbon storage (Peterson and Melillo, 1985; Schimel *et al*., 1996; Holland *et al*., 1997). Based on stoichiometric relations between carbon and nitrogen, many physiologically based models predict that added nitrogen should lead to an accumulation of carbon in biomass. But the extent to which this accumulation occurs in nature is unclear. Adding nitrogen to forests does increase NPP (Bergh *et al*., 1999). It may also modify soil organic matter and increase its residence time (Fog, 1988; Bryant *et al*., 1998). But nitrogen deposited in an ecosystem may also be immobilized in soils (Nadelhoffer *et al*., 1999) or lost from the ecosystem, becoming largely unavailable in either case (Davidson, 1995).

There is also evidence that additions of nitrogen above some level may saturate the ecosystem, causing: (i) increased nitrification and nitrate leaching, with associated acidification of soils and surface waters; (ii) cation depletion and nutrient imbalances; and (iii) reduced productivity (Aber *et al*., 1998; Fenn *et al*., 1998). Experimental nitrogen additions have had varied effects on wood production and growing stocks. Woody biomass production increased in response to nitrogen additions to two New England hardwood sites, although increased mortality at one site led to a net decrease in the stock of woody biomass (Magill *et al*., 1997, 2000). Several studies have shown that chronic exposure to elevated nitrogen inputs can inhibit forest growth, especially in evergreen species (Tamm *et al*., 1995; Makipaa, 1995). Fertilization decreased rates of wood production in high-elevation spruce-fir in Vermont (McNulty *et al*., 1996) and in a heavily fertilized red pine plantation (Magill *et al*., 2000). The long-term effects of nitrogen deposition on forest production and carbon balance remain uncertain. Furthermore, because much of the nitrogen deposited on land is in the form of acid precipitation, it is difficult to distinguish the fertilization effects of nitrogen from the adverse effects of acidity (see below).

Atmospheric chemistry. Other factors besides nitrogen saturation may have negative effects on NPP, thus reducing the uptake of carbon in ecosystems and perhaps changing them from sinks to sources of carbon. Two factors that have received attention are tropospheric ozone and sulfur (acid rain). Experimental studies show leaf injury and reduced growth in crops and trees exposed to ozone. At the level of the ecosystem, elevated levels of ozone have been associated with reduced forest growth in North America (Mclaughlin and Percy, 2000) and Europe (Braun *et al*., 2000). Acidification of soil as a result of deposition of NO_3^- and SO_4^{2-} in precipitation depletes the soils of available plant nutrients (Ca^{2+}, Mg^{2+}, K^+), increases the mobility and toxicity of aluminum, and increases the amount of nitrogen and sulfur stored in forest soils (Driscoll *et al*., 2001). The loss of plant nutrients raises concerns about the long-term health and productivity of forests in the northeastern US, Europe, and southern China.

Although the effects of tropospheric ozone and sulfur generally reduce NPP, their actual or potential effects on carbon stocks are not known. The pollutants could potentially increase carbon stocks if they reduce decomposition of organic matter more than they reduce NPP.

Climatic variability and climatic change. Year-to-year differences in the growth rate of CO_2 in the atmosphere are large (Figure 3). The annual rate of increase ranged from 1.9 Pg C in 1992 to 6.0 Pg C in 1998 (Prentice *et al*., 2001; see also Conway *et al*., 1994). In 1998 the net global sink (ocean and land) was nearly 0 Pg C, while the average combined sink in the previous eight years was $\sim$3.5 Pg C yr^{-1} (Tans *et al*., 2001). The terrestrial sink is generally twice as variable as the oceanic sink (Bousquet *et al*., 2000). This temporal variability in terrestrial fluxes is probably caused by the effect of climate on carbon pools with short lifetimes (foliage, plant litter, soil microbes) through variations in photosynthesis, respiration, and possibly fire (Schimel *et al*., 2001). Measurements in terrestrial ecosystems suggest that respiration, rather than photosynthesis, is the major contributor to variability (Valentini *et al*., 2000). Annual respiration was almost twice as variable as photosynthesis over a five-year period in the Harvard Forest (Goulden *et al*., 1996). Respiration is also more sensitive than photosynthesis to changes in both temperature and moisture. For example, during a dry year at the Harvard Forest, both photosynthesis and respiration were reduced, but the reduction in respiration was greater, yielding a greater than

average net uptake of carbon for the year (Goulden *et al*., 1996). A tropical forest in the Brazilian Amazon behaved similarly (Saleska *et al*., in press).

The greater sensitivity of respiration to climatic variations is also observed at the global scale. An analysis of satellite data over the US, together with an ecosystem model, shows that the variability in NPP is considerably less than the variability in the growth rate of atmospheric CO_2 inferred from inverse modeling, suggesting that the cause of the year-to-year variability in carbon fluxes is largely from varying rates of respiration rather than photosynthesis (Hicke *et al*., 2002). Also, global NPP was remarkably constant over the three-year transition from El Niño to La Niña (Behrenfeld *et al*., 2001). Myneni *et al*. (1995) found a positive correlation between annual "greenness," derived from satellites, and the growth rate of CO_2. Greener years presumably had more photosynthesis and higher GPP, but they also had proportionately more respiration, thus yielding a net release of carbon from land (or reduced uptake) despite increased greenness.

Climatic factors influence terrestrial carbon storage through effects on photosynthesis, respiration, growth, and decay. However, prediction of future terrestrial sinks resulting from climate change requires an understanding of not only plant and microbial physiology, but the regional aspects of future climate change, as well. The important aspects of climate are: (i) temperature, including the length of the growing season; (ii) moisture; and (iii) solar radiation and clouds. Although year-to-year variations in the growth rate of CO_2 are probably the result of terrestrial responses to climatic variability, longer-term changes in carbon storage involve acclimation and other physiological adjustments that generally reduce short-term responses.

In cold ecosystems, such as those in high latitudes (tundra and taiga), an increase in temperature might be expected to increase NPP and, perhaps, carbon storage (although the effects might be indirect through increased rates of nitrogen mineralization; Jarvis and Linder, 2000). Satellite records of "greenness" over the boreal zone and temperate Europe show a lengthening of the growing season (Myneni *et al*., 1997), suggesting greater growth and carbon storage. Measurements of CO_2 flux in these ecosystems do not consistently show a net uptake of carbon in response to warm temperatures (Oechel *et al*., 1993; Goulden *et al*., 1998), however, presumably because warmer soils release more carbon than plants take up. Increased temperatures in boreal forests may also reduce plant growth if the higher temperatures are associated with drier conditions (Barber *et al*., 2000; Lloyd and Fastie, 2002). The same is true in the tropics, especially as the risk of fires increases with drought (Nepstad *et al*., 1999; Page *et al*., 2002). A warming-enhanced increase in rates of respiration and decay may already have begun to release carbon to the atmosphere (Woodwell, 1983; Raich and Schlesinger, 1992; Houghton *et al*., 1998).

The results of short-term experiments may be misleading, however, because of acclimation or because the more easily decomposed material is respired rapidly. The long-term, or equilibrium, effects of climate on carbon storage can be inferred from the fact that cool, wet habitats store more carbon in soils than hot, dry habitats (Post *et al*., 1982). The transient effects of climatic change on carbon storage, however, are difficult to predict, in large part because of uncertainty in predicting regional and temporal changes in temperature and moisture (extremes as well as means) and rates of climatic change, but also from incomplete understanding of how such changes affect fires, disease, pests, and species migration rates.

In the short term of seasons to a few years, variations in terrestrial carbon storage are most likely driven by variations in climate (temperature, moisture, light, length of growing season). Carbon dioxide fertilization and nitrogen deposition, in contrast, are unlikely to change abruptly. Interannual variations in the emissions of carbon from land-use change are also likely to be small (<0.2 Pg C yr^{-1}) because socioeconomic changes in different regions generally offset each other, and because the releases and uptake of carbon associated with a land-use change lag the change in land use itself and thus spread the emissions over time (Houghton, 2000). Figure 8 shows the annual net emissions of carbon from deforestation and reforestation in the Brazilian Amazon relative to the annual fluxes observed in the growth rates of trees and modeled on the basis of physiological responses to climatic variation. Clearly, metabolic responses to climatic variations are more important in the short term than interannual variations in rates of land-use change.

Understanding short-term variations in atmospheric CO_2 may not be adequate for predicting longer-term trends, however. Organisms and populations acclimate and adapt in ways that generally diminish short-term responses. Just as increased rates of photosynthesis in response to elevated levels of CO_2 often, but not always, decline within months or years (Tissue and Oechel, 1987), the same diminished response has been observed for higher temperatures (Luo *et al*., 2001). Thus, over decades and centuries the factors most important in influencing concentrations of atmospheric CO_2 (fossil fuel emissions, land-use change, oceanic uptake) are probably different from those factors important in determining the short-term variations in

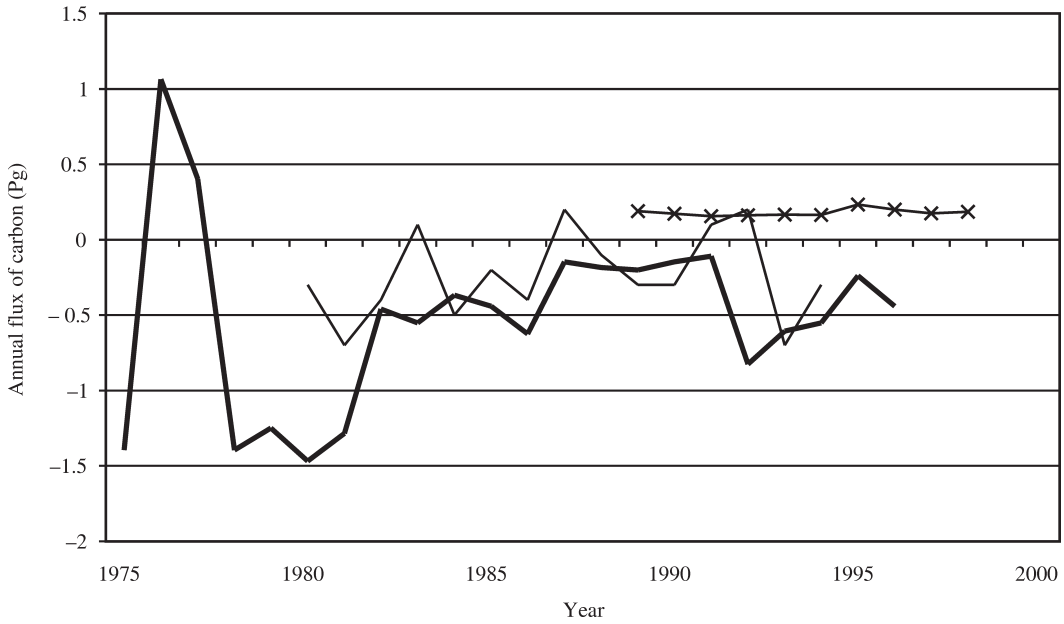

Figure 8 Net annual sources (+) and sinks (−) of carbon for the Brazilian Amazon, as determined by three different methods: (×) land-use change (Houghton *et al*., 2000); (—) tree growth (Phillips *et al*., 1998); and (—) modeled ecosystem metabolism (Tian *et al*., 1998) (reproduced by permission of the American Geophysical Union from *J. Geophys. Res.*, **2000**, *105*, 20121–20130).

atmospheric CO_2 (Houghton, 2000). Long-term changes in climate, as opposed to climatic variability, may eventually lead to long-term changes in carbon storage, but probably not at the rates suggested by short-term experiments.

One further observation is discussed here. Over the last decades the amplitude of the seasonal oscillation of CO_2 concentration increased by ~20% at Mauna Loa, Hawaii, and by ~40% at Point Barrow, Alaska (Keeling *et al*., 1996a). This winter–summer oscillation in concentrations seems to be largely the result of terrestrial metabolism in northern mid-latitudes. The increase in amplitude suggests that the rate of processing of carbon may be increasing. Increased rates of summer photosynthesis, increased rates of winter respiration, or both would increase the amplitude of the oscillation, but it is difficult to ascertain which has contributed most. Furthermore, the increase in the amplitude does not, by itself, indicate an increasing terrestrial sink. In fact, the increase in amplitude is too large to be attributed to CO_2 fertilization or to a temperature-caused increase in winter respiration (Houghton, 1987; Randerson *et al*., 1997). It is consistent with the observation that growing seasons have been starting earlier over the last decades (Randerson *et al*., 1999). The trend has been observed in the temperature data, in decreasing snow cover (Folland and Karl, 2001), and in the satellite record of vegetation activity (Myneni *et al*., 1997).

Synergies among physiological "mechanisms." The factors influencing carbon storage often interact nonadditively. For example, higher concentrations of CO_2 in air enable plants to acquire the same amount of carbon with a smaller loss of water through their leaves. This increased water-use efficiency reduces the effects of drought. Higher levels of CO_2 may also alleviate other stresses of plants, such as temperature and ozone. The observation that NPP is increased relatively more in "low productivity" years suggests that the indirect effects of CO_2 in ameliorating stress may be more important than the direct effects of CO_2 on photosynthesis (Luo *et al*., 1999).

Another example of synergistic effects is the observation that the combination of nitrogen fertilizer and elevated CO_2 concentration may have a greater effect on the growth of biomass in a growing forest than the expected additive effect (Oren *et al*., 2001). The relative increase was greater in a nutritionally poor site. The synergy between nitrogen and CO_2 was different in a grassland, however (Hu *et al*., 2001). There, elevated CO_2 increased plant uptake of nitrogen, increased NPP, and increased the carbon available for microbes; but it reduced microbial decomposition, presumably because the utilization of nitrogen by plants reduced its availability for microbes. The net effect of the reduced decomposition was an increase in the accumulation of carbon in soil.

Relatively few experiments have included more than one environmental variable at a time. A recent experiment involving combinations of four variables shows the importance of such work.

Shaw *et al*. (2003) exposed an annual grassland community in California to increased temperature, precipitation, nitrogen deposition, and atmospheric CO_2 concentration. Alone, each of the treatments increased NPP in the third year of treatment. Across all multifactor treatments, however, elevated CO_2 decreased the positive effects of the other treatments. That is, elevated CO_2 increased productivity under "poor" growing conditions, but reduced it under favorable growing conditions. The most likely explanation is that some soil nutrient became limiting, either because of increased microbial activity or decreased root allocation (Shaw *et al*., 2003).

The expense of such multifactor experiments has led scientists to use process-based ecosystem models (see the discussion of "terrestrial carbon models" below) to predict the response of terrestrial ecosystems to future climates. When predicting the effects of CO_2 alone, six global biogeochemical models showed a global terrestrial sink that began in the early part of the twentieth century and increased (with one exception) towards the year 2100 (Cramer *et al*., 2001). The maximum sink varied from $\sim$4 Pg C yr^{-1} to $\sim$10 Pg C yr^{-1}. Adding changes in climate (predicted by the Hadley Centre) to these models reduced the future sink (with one exception), and in one case reduced the sink to zero near the year 2100.

Terrestrial carbon models. A number of ecosystem models have been developed to calculate gross and net fluxes of carbon from environmentally induced changes in plant or microbial metabolism, such as photosynthesis, plant respiration, decomposition, and heterotrophic respiration (Cramer *et al*., 2001; McGuire *et al*., 2001). For example, six global models yielded net terrestrial sinks of carbon ranging between 1.5 Pg C yr^{-1} and 4.0 Pg C yr^{-1} for the year 2000 (Cramer *et al*., 2001). The differences among models became larger as environmental conditions departed from existing conditions. The magnitude of the terrestrial carbon flux projected for the year 2100 varied between a source of 0.5 Pg C yr^{-1} and a sink of 7 Pg C yr^{-1}. Other physiologically based models, including the effects of climate on plant distribution as well as growth, projected a net source from land as tropical forests were replaced with savannas (White *et al*., 1999; Cox *et al*., 2000).

The advantage of such models is that they allow the effects of different mechanisms to be distinguished. However, they may not include all of the important processes affecting changes in carbon stocks. To date, e.g., few process-based terrestrial models have included changes in land use.

Although some processes, such as photosynthesis, are well enough understood for predicting responses to multiple factors, other processes, such as biomass allocation, phenology, and the replacement of one species by another, are not. Even if the physiological mechanisms and their interactions were well understood and incorporated into the models, other nonphysiological factors that affect carbon storage (e.g., fires, storms, insects, and disease) are not considered in the present generation of models. Furthermore, the factors influencing short-term changes in terrestrial carbon storage may not be the ones responsible for long-term changes (Houghton, 2000) (see next section). The variability among model predictions suggests that they are not reliable enough to demonstrate the mechanisms responsible for the current modest terrestrial sink (Cramer *et al*., 2001; Knorr and Heimann, 2001).

8.10.4.1.2 Demographic or disturbance mechanisms

Terrestrial sinks also result from the recovery (growth) of ecosystems disturbed in the past. The processes responsible for regrowth include physiological and metabolic processes, but they also involve higher-order or more integrated processes, such as succession, growth, and aging. Forests accumulate carbon as they grow. Regrowth is initiated either by disturbances or by the planting of trees on open land. Disturbances may be either natural (insects, disease, some fires) or human induced (management and changes in land use, including fire management). Climatic effects—e.g., droughts, storms, or fires—thus affect terrestrial carbon storage not only through physiological or metabolic effects on plant growth and respiration, but also through effects on stand demography and growth.

In some regions of the world—e.g., the US and Europe—past changes in land use are responsible for an existing sink (Houghton *et al*., 1999; Caspersen *et al*., 2000; Houghton, 2003). Processes include the accumulation of carbon in forests as a result of fire suppression, the growth of forests on lands abandoned from agriculture, and the growth of forests earlier harvested. In tropical regions carbon accumulates in forests that are in the fallow period of shifting cultivation. All regions, even countries with high rates of deforestation, have sinks of carbon in recovering forests, but often these sinks are offset by large emissions (Table 9). The sinks in tropical regions as a result of logging are nearly the same in magnitude as those outside the tropics.

Sinks of carbon are not limited to forests. Some analyses of the US (Houghton *et al*., 1999; Pacala *et al*., 2001) show that a number of processes in nonforest ecosystems may also be responsible for carbon sinks. Processes include the encroachment of woody vegetation into formerly herbaceous ecosystems, the accumulation of carbon in

agricultural soils as a result of conservation tillage or other practices, exportation of wood and food, and the riverine export of carbon from land to the sea (Table 12). At least a portion of these last two processes (import/export of food and wood and river export) represents an export of carbon from the US (an apparent sink) but not a global sink because these exports presumably become sources somewhere else (either in ocean waters or in another country).

Which terrestrial mechanisms are important? Until recently, the most common explanations for the residual carbon sink in the 1980s and 1990s were factors that affect the physiology of plants and microbes: CO_2 fertilization, nitrogen deposition, and climatic variability (see Table 13). Several findings have started to shift the explanation to include management practices and disturbances that affect the age structure or demography of ecosystems. For example, the suggestion that CO_2 fertilization may be less important in forests than in short-term greenhouse experiments (Oren *et al*., 2001) was discussed above. Second, physiological models quantifying the effects of CO_2 fertilization and climate change on the growth of US forests could account for only a small fraction of the carbon accumulation observed in those forests (Schimel *et al*., 2000). The authors acknowledged that past changes in land use were likely to be important. Third, and most importantly, 98% of recent accumulations of carbon in US forests can be explained on the basis of the age structure of trees without requiring growth enhancement due to CO_2 or nitrogen fertilization (Caspersen *et al*., 2000). Either the physiological effects of CO_2, nitrogen, and climate have been unimportant or their effects have been offset by unknown influences. Finally, the estimates of sinks in the US (Houghton *et al*., 1999; Pacala *et al*., 2001; Table 12) are based, to a large extent, on changes in land use and management, and not on physiological models of plant and soil metabolism.

To date, investigations of these two different classes of mechanisms have been largely independent. The effects of changing environmental conditions have been largely ignored in analyses of land-use change (see Section 8.10.3.1.5), and physiological models have generally ignored changes in land use (see Section 8.10.4.1.1).

As of early 2000s, the importance of different mechanisms in explaining known terrestrial carbon sinks remains unclear. Management and past disturbances seem to be the dominant mechanisms for a sink in mid-latitudes, but they are unlikely to explain a large carbon sink in the tropics (if one exists). Recovery from past disturbances is unlikely to explain a large carbon sink in the tropics, because both the area of forests and the stocks of carbon within forests have been declining. Rates of human-caused disturbance have been accelerating. Clearly there are tropical forests recovering from natural disturbances, but there is no evidence that the frequency of disturbances changed during the last century, and thus no evidence to suggest that the sink in recovering forests is larger or smaller today than in previous centuries. The lack of systematic forest inventories over large areas in the tropics precludes a more definitive test of where forests are accumulating carbon and where they are losing it.

Enhanced rates of plant growth cannot be ruled out as an explanation for apparent sinks in either the tropics or mid-latitude lands, but it is possible that the current sink is entirely the result of recovery from earlier disturbances, anthropogenic and natural.

How will the magnitude of the current terrestrial sink change in the future? Identifying the mechanisms responsible for past and current carbon sinks is important because some mechanisms are more likely than others to persist into the future. As discussed above, physiologically based models predict that CO_2 fertilization will increase the global terrestrial sink over the next 100 years (Cramer *et al*., 2001). Including the effects of projected climate change reduces the magnitude of projected sinks in many models but turns the current sink into a future global source in models that include the longer-term effects of climate on plant distribution (White *et al*., 1999; Cox *et al*., 2000). Thus, although increased levels of CO_2 are thought to increase carbon storage in forests, the effect of warmer temperatures may replace forests with savannas and grasslands, and, in the process, release carbon to the atmosphere. Future changes in natural systems are difficult to predict.

To the extent the current terrestrial sink is a result of regrowth (changes in age structure), the future terrestrial sink is more constrained. First, the net effect of continued land-use change is likely to release carbon, rather than store it. Second, forests that might have accumulated carbon in recent decades (whatever the cause) will cease to function as sinks if they are turned into croplands. Third, the current sink in regrowing forests will diminish as forests mature (Hurtt *et al*., 2002).

Despite the recent evidence that changes in land use are more important in explaining the current terrestrial carbon sink than physiological responses to environmental changes in CO_2, nitrogen, or climate, most projections of future rates of climatic change are based on the assumption that the current terrestrial sink will not only continue, but grow in proportion to concentrations of CO_2. Positive biotic feedbacks and changes in land use are not included in the general circulation models (GCMs) used to predict future rates of climate change. The GCMs

include physical feedbacks such as water vapor, clouds, snow, and polar ice, but not biotic feedbacks (Woodwell and Mackenzie, 1995). Thus, unless negative feedbacks in the biosphere become more important in the future, through physiological or other processes, these climate projections underestimate the rate and extent of climatic change. If the terrestrial sink were to diminish in the next decades, concentrations of CO_2 by the year 2100 might be hundreds of ppm higher than commonly projected.

8.10.4.2 Oceanic Mechanisms

8.10.4.2.1 Physical and chemical mechanisms

Increasing the concentration of CO_2 in the atmosphere is expected to affect the rate of oceanic uptake of carbon through at least eight mechanisms, half of them physical or chemical, and half of them biological. Most of the mechanisms reduce the short-term uptake of carbon by the oceans.

The buffer factor. The oceanic buffer factor (or Revelle factor), by which the concentration of CO_2 in the atmosphere is determined, increases as the concentration of CO_2 increases. The buffer factor is discussed above in Section 8.10.3.1.3. Here, it is sufficient to describe the chemical equation for the dissolution of CO_2 in seawater.

$$2HCO_3 + Ca \rightleftharpoons CaCO_3 + CO_2 + H_2O$$

Every molecule of CO_2 entering the oceans consumes a molecule of carbonate as the CO_2 is converted to bicarbonate. Thus, as CO_2 enters the ocean, the concentration of carbonate ions decreases, and further additions of CO_2 remain as dissolved CO_2 rather than being converted to HCO_3^-. The ocean becomes less effective in taking up additional CO_2. The effect is large. The change in DIC for a 100 ppm increase above 280 ppm (pre-industrial) was 40% larger than a 100 ppm increase would be today. The change in DIC for a 100 ppm increase above 750 ppm will be 60% lower than it would be today (Prentice *et al*., 2001). Thus, the fraction of added CO_2 going into the ocean decreases and the fraction remaining in the atmosphere increases as concentrations continue to increase.

Warming. The solubility of CO_2 in seawater decreases with temperature. Raising the ocean temperature 1 °C increases the equilibrium p_{CO_2} in seawater by 10–20 ppm, thus increasing the atmospheric concentration by that much as well. This mechanism is a positive feedback to a global warming.

Vertical mixing and stratification. If the warming of the oceans takes place in the surface layers first, the warming would be expected to increase the stability of the water column. As discussed in Section 8.10.3.1.3, the bottleneck for oceanic uptake of CO_2 is largely the rate at which the surface oceans exchange CO_2 with the intermediate and deeper waters. Greater stability of the water column, as a result of warming, might constrict this bottleneck further. Similarly, if the warming of the Earth's surface is greater at the poles than at the equator, the latitudinal gradient in surface ocean temperature will be reduced; and because that thermal gradient plays a role in the intensity of atmospheric mixing, a smaller gradient might be expected to subdue mixing and increase stagnation. Alternatively, the increased intensity of the hydrologic cycle expected for a warmer Earth will probably increase the intensity of storms and might, thereby, increase oceanic mixing. Interactions between oceanic stability and biological production might also change the ocean's carbon cycle, with consequences for the oceanic uptake of carbon that are difficult to predict (Sarmiento *et al*., 1998; Matear and Hirst, 1999).

One aspect of the ocean's circulation that seems particularly vulnerable to climate change is the thermohaline circulation, which is related to the formation of North Atlantic Deep Water (NADW). Increased warming of surface waters may intensify the hydrologic cycle, leading to a reduced salinity in the sea surface at high latitudes, a reduction (even collapse) of NADW formation, reduction in the surface-to-deep transport of anthropogenic carbon, and thus a higher rate of CO_2 growth in the atmosphere. In a model simulation, modest rates of warming reduced the rate of oceanic uptake of carbon, but the reduced uptake was largely compensated by changes in the marine biological cycle (Joos *et al*., 1999a). For higher rates of global warming, however, the NADW formation collapsed and the concentration of CO_2 in the atmosphere was 22% (and global temperature 0.6 °C) higher than expected in the absence of this feedback.

Rate of CO_2 emission. High rates of CO_2 emissions will increase the atmosphere–ocean gradient in CO_2 concentrations. Although this gradient drives the uptake of carbon by surface waters, if the rate of CO_2 emissions is greater than the rate of CO_2 uptake, the fraction of emitted CO_2 remaining in the atmosphere will be higher. Under the business-as-usual scenario for future CO_2 emissions, rates of emissions increase by more than a factor of 3, from approximately 6 Pg C yr^{-1} in the 1990s to 20 Pg C yr^{-1} by the end of the twenty-first century.

8.10.4.2.2 Biological feedback/processes

Changes in biological processes may offset some of the physical and chemical effects

described above (Sarmiento *et al*., 1998; Joos *et al*., 1999a), but the understanding of these processes is incomplete, and the net effects far from predictable. Potential effects fall into four categories (Falkowski *et al*., 1998).

(i) Addition of nutrients limiting primary production. Nutrient enrichment experiments and observations of nutrient distributions throughout the oceans suggest that marine primary productivity is often limited by the availability of fixed inorganic nitrogen. As most of the nitrogen for marine production comes from upwelling, physical changes in ocean circulation might also affect oceanic primary production and, hence, the biological pump. Some nitrogen is made available through nitrogen fixation, however, and some is lost through denitrification, both of which are biological processes, limited by trace nutrients and the concentration of oxygen. The two processes are not coupled, however, and differential changes in either one would affect the inventory of fixed nitrogen in the ocean.

(ii) Enhanced utilization of nutrients. One of the mysteries of ocean biology today is the observation of "high nutrient, low chlorophyll (HNLC) regions." That is, why does primary production in major regions of the surface ocean stop before all of the available nitrogen and phosphorous have been used up? It is possible that grazing pressures keep phytoplankton populations from consuming the available nitrogen and phosphorous, and any reduction in grazing pressures might increase the export of organic matter from the surface. Another possibility that has received considerable attention is that iron may limit production (Martin, 1990). In fact, deliberate iron fertilization of the ocean has received serious attention as a way of reducing atmospheric CO_2 (see Section 8.10.5.2, below). Iron might also become more available naturally as a result of increased human eutrophication of coastal waters, or it might be less available as a result of a warmer (more strongly stratified) ocean or reduced transport of dust (Falkowski *et al*., 1998). The aeolian transport of iron in dust is a major source of iron for the open ocean, and dust could either increase or decrease in the future, depending on changes in the distribution of precipitation.

(iii) Changes in the elemental ratios of organic matter in the ocean. The elemental ratio of C : N : P in marine organic particles has long been recognized as conservative (Falkowski *et al*., 1998). The extent to which the ratios can depart from observed concentrations is not known, yet variations could reduce the limitation of nitrogen and thus act in the same manner as the addition of nitrogen in affecting production, export, and thus oceanic uptake of CO_2.

(iv) Increases in the organic carbon/carbonate ratio of export production. The biological and carbonate pumps are described above (Section 8.10.2.2.2). Both pumps transport carbon out of the surface waters, and the subsequent decay at depth is responsible for the higher concentration of carbon in the intermediate and deep ocean. The formation of carbonate shells in the surface waters has the additional effect of increasing the p_{CO_2} in these waters, thus negating the export of the carbonate shells out of the surface. Any increase in the organic carbon/carbonate ratio of export production would enhance the efficiency of the biological pump.

8.10.5 THE FUTURE: DELIBERATE SEQUESTERING OF CARBON (OR REDUCTION OF SOURCES)

Section 8.10.4 addressed the factors thought to be influencing current terrestrial and oceanic sinks, and how they might change in the future. It is possible, of course, that CO_2 fertilization will become more important in the future as concentrations of CO_2 increase. Multiyear, whole ecosystems experiments with elevated CO_2 do not uniformly support this possibility, but higher concentrations of CO_2, together with nitrogen deposition or increases in moisture, might yet be important. Rather than wait for a more definitive answer, a more cautious approach to the future, besides reducing emissions of CO_2, would consider strategies for withdrawing carbon from the atmosphere through management. Three general options for sequestering carbon have received attention: terrestrial, oceanic, and geological management.

8.10.5.1 Terrestrial

Even if CO_2 fertilization and other environment effects turn out to be unimportant in enhancing terrestrial carbon storage, terrestrial sinks can still be counted on to offset carbon emissions or to reduce atmospheric concentrations of CO_2. Increasing the amount of carbon held on land might be achieved through at least six management options (Houghton, 1996; Kohlmaier *et al*., 1998): (i) a reduction in the rate of deforestation (current rates of deforestation in the tropics are responsible for an annual release of 1–2 Pg C (Section 8.10.3.2.3); (ii) an increase in the area of forests (afforestation); (iii) an increase in the stocks of carbon within existing forests; (iv) an increase in the use of wood (including increased efficiency of wood harvest and use); (v) the substitution of wood fuels for fossil fuels; and (vi) the substitution of wood for more energy-intensive materials, such as aluminum, concrete, and steel. Estimates of the amount of carbon that might be sequestered on land over the 55-year period 1995–2050 range between 60 Pg C and 87 Pg C (1–2 Pg C yr^{-1} on average)

(Brown, 1996). Additional carbon might also be sequestered in agricultural soils through conservation tillage and other agricultural management practices, and in grassland soils (Sampson and Scholes, 2000). An optimistic assessment, considering all types of ecosystems over the Earth, estimated a potential for storing 5–10 Pg C yr^{-1} over a period of 25–50 years (DOE, 1999).

The amount of carbon potentially sequestered is small relative to projected emissions of CO_2 from business-as-usual energy practices, and thus the terrestrial options for sequestering carbon should be viewed as temporary, "buying time" for the development and implementation of longer-lasting measures for reducing fossil fuel emissions (Watson *et al*., 2000).

8.10.5.2 Oceanic

Schemes for increasing the storage of carbon in the oceans include stimulation of primary production with iron fertilization and direct injection of CO_2 at depth. As pointed out in Section 8.10.4.2.2., there are large areas of the ocean with high nutrient, low chlorophyll, concentrations. One explanation is that marine production is limited by the micronutrient iron. Adding iron to these regions might thus increase the ocean's biological pump, thereby reducing atmospheric CO_2 (Martin, 1990; Falkowski *et al*., 1998). Mesoscale fertilization experiments have been carried out (Boyd *et al*., 2000), but the effects of large-scale iron fertilization of the ocean are not known (Chisholm, 2000).

The direct injection of concentrated CO_2 (probably in liquid form) below the thermocline or on the seafloor might sequester carbon for hundreds of years (Herzog *et al*., 2000). The gas might be dissolved within the water column or held in solid, ice-like CO_2 hydrates. The possibility is receiving attention in several national and international experiments (DOE, 1999). Large uncertainties exist in understanding the formation and stability of CO_2 hydrates, the effect of the concentrated CO_2 on ocean ecosystems, and the permanence of the sequestration.

8.10.5.3 Geologic

CO_2 may be able to be sequestered in geological formations, such as active and depleted oil and gas reservoirs, coalbeds, and deep saline aquifers. Such formations are widespread and have the potential to sequester large amounts of CO_2 (Herzog *et al*., 2000). A model project is underway in the North Sea off the coast of Norway. The Sleipner offshore oil and natural gas field contains a gas mixture of natural gas and CO_2 (9%). Because the Norwegian government taxes emissions of CO_2 in excess of 2.5%, companies have the incentive to separate CO_2 from the natural gas and pump it into an aquifer 1,000 m under the sea. Although the potential for sequestering carbon in geological formations is large, technical and economic aspects of an operational program require considerable research.

8.10.6 CONCLUSION

We are conducting a great geochemical experiment, unlike anything in human history and unlikely to be repeated again on Earth. "Within a few centuries we are returning to the atmosphere and oceans the concentrated organic carbon stored in sedimentary rocks over hundreds of millions of years" (Revelle and Suess, 1957). During the last 150 years (~1850–2000), there has been a 30% increase in the amount of carbon in the atmosphere. Although most of this carbon has come from the combustion of fossil fuels, an estimated 150–160 Pg C have been lost during this time from terrestrial ecosystems as a result of human management (another 58–75 Pg C were lost before 1850). The global carbon balance suggests that other terrestrial ecosystems have accumulated ~115 Pg C since about 1930, at a steadily increasing rate. The annual net fluxes of carbon appear small relative to the sizes of the reservoirs, but the fluxes have been accelerating. Fifty percent of the carbon mobilized over the last 300 years (~1700–2000) was mobilized in the last 30–40 of these years (Houghton and Skole, 1990) (Figure 3). The major drivers of the geochemical experiment are reasonably well known. However, the results are uncertain, and there is no control. Furthermore, the experiment would take a long time to stop (or reverse) if the results turned out to be deleterious.

In an attempt to put some bounds on the experiment, in 1992 the nations of the world adopted the United Nations Framework Convention on Climate Change, which has as its objective "stabilization of greenhouse gas concentrations in the atmosphere at a level that would prevent dangerous anthropogenic interference with the climate system" (UNFCCC, 1992). The Convention's soft commitment suggested that the emissions of greenhouse gases from industrial nations in 2000 be no higher than the emissions in 1990. This commitment has been achieved, although more by accident than as a result of deliberate changes in policy. The "stabilization" resulted from reduced emissions from Russia, as a result of economic downturn, balanced by increased emissions almost everywhere else. In the US, e.g., emissions were 18% higher in the year 2000 than they had been in 1990. The near-zero increase in industrial nations' emissions between 1990 and 2000 does not suggest that the stabilization will last.

Ironically, even if the annual rate of global emissions were to be stabilized, concentrations of the gases would continue to increase. Stabilization of concentrations at early 2000's levels, e.g., would require reductions of 60% or more in the emission of long-lived gases, such as CO_2. The 5% average reduction in 1990 emissions by 2010, agreed to by the industrialized countries in the Kyoto Protocol (higher than 5% for the participating countries now that the US is no longer participating), falls far short of stabilizing atmospheric concentrations. Such a stabilization will require nothing less than a switch from fossil fuels to renewable forms of energy (solar, wind, hydropower, biomass), a switch that would have salubrious economic, political, security, and health consequences quite apart from limiting climatic change. Nevertheless, the geophysical experiment seems likely to continue for at least the near future, matched by a sociopolitical experiment of similar proportions, dealing with the consequences of either mitigation or not enough mitigation.

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