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Current Status and Past Trends of the Global Carbon Cycle

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In a global, long-term perspective, the record of atmospheric CO_2 content documents the magnitude and speed of climate-driven variations, such as the glacial-interglacial cycles (which drove CO_2 variations of ~100 parts per million [ppm] over 420,000 years). These observations also, however, document a remarkable stability, with variations in atmospheric CO_2 of <20 ppm during at least the last 11,000 years before the Industrial Era (Joos and Prentice, Chapter 7, this volume). In this longer-term context, the anthropogenic increase of ~100 ppm during the past 200 years is a dramatic alteration of the global carbon cycle. This atmospheric increase is also a graphic documentation of profound changes in human activity. The atmospheric record documents the Earth system's response to fossil-fuel releases that increased by more than 1,200 percent between 1900 and 1999 (Nakicenovic, Chapter 11, this volume).

To understand and predict future changes in the global carbon cycle, we must first understand how the system is operating today. In many cases the current fluxes of carbon are a direct result of past processes affecting these fluxes (Nabuurs, Chapter 16, this volume). Thus, it is important to understand the current carbon cycle in the context of how the system has evolved over time. The Third Assessment Report (TAR) of the Intergovernmental Panel on Climate Change (IPCC) recently compiled a global carbon budget (Prentice et al. 2001). While that budget reflected the state of the art at that time, this chapter presents a revised budget based on new information from model studies and oceanographic observations. The IPCC-TAR budget focused on the overall carbon bal-

ance between the major active reservoirs of land, atmosphere, and ocean. In this chapter we present a somewhat more comprehensive representation of the connections between the reservoirs, together with our current understanding of the biogeochemical processes and human driving forces controlling these exchanges. We also introduce the key processes involved in controlling atmospheric concentrations of CO_2 and relevant non-CO_2 gases (e.g., CH_4, N_2O) that may be susceptible to changes in the future, either through deliberate management or as direct and indirect consequences of global change.

The Global Carbon Budget

Over the past 200 years humans have introduced ~ 400 petagrams of carbon (PgC) to the atmosphere through deforestation and the burning of fossil fuels. Part of this carbon was absorbed by the oceans and terrestrial biosphere. Table 2.1, section 1, shows the global budget recently compiled by the IPCC Third Assessment Report (Prentice et al. 2001). The global carbon budget quantifies the relative importance of these two reservoirs today, and the budget uncertainty reflects our understanding of the exchanges between these reservoirs. The IPCC assessment partitions the uptake into net terrestrial and oceanic components based primarily on observations of the concurrent global trends of atmospheric CO_2 and oxygen. Since the compilation of the IPCC report, new evidence from model studies and oceanographic observations shows that this budget should be slightly revised to account for a previously ignored oceanic oxygen flux. This flux is the result of enhanced oceanic mixing, as inferred from observed changes in oceanic heat content (Le Quéré et al. 2003). The revised values are presented in section 2 of Table 2.1.

Section 3 of Table 2.1 shows the breakdown of the net land-atmosphere flux through the 1990s into emissions from changes in land use and a residual terrestrial sink, based on the updated land use change emissions of Houghton (2003). This breakdown has recently been challenged based on new estimates of land use change determined from remote sensing data (Table 2.1, sections 4 and 5). These new estimates lie at the lower end of the range of estimates based on data reported by individual countries (Houghton 2003). If correct, they imply a residual terrestrial sink in the 1990s that is about 40 percent smaller than previous estimates.

This budget, of course, does not attempt to represent the richness of the global carbon cycle. The land-atmosphere-ocean system is connected by a multitude of exchange fluxes. The dynamical behavior of this system is determined by the relative sizes of the different reservoirs and fluxes, together with the biogeochemical processes and human driving forces controlling these exchanges. Colorplate 1 shows the globally aggregated layout of the carbon cycle, together with the pools and exchange fluxes that are relevant on timescales of up to a few millennia. Panel a in Colorplate 1 presents a basic picture of the global carbon cycle, including the preindustrial (thin) and anthropogenic (bold) ocean-atmosphere and land-atmosphere exchange fluxes. The anthropogenic fluxes are average values for the 1980s and 1990s. Panel a also shows components of the long-term

Table 2.1. The global carbon budget (PgC y^{-1})

	1980s	1990s
1. Prentice et al. 2001		
Atmospheric increase	+3.3 ± 0.1	+3.2 ± 0.1
Emissions (fossil fuel, cement)	+5.4 ± 0.3	+6.3 ± 0.4
Ocean-atmosphere flux	−1.9 ± 0.6	−1.7 ± 0.5
Net land-atmosphere flux	−0.2 ± 0.7	−1.4 ± 0.7
Land use change	+1.7 (+0.6 to +2.5)	-
Residual terrestrial sink	−1.9 (−3.8 to +0.3)	-
2. Le Quéré et al. 2003		
Ocean corrected	−1.8 ± 0.8	−1.9 ± 0.7
Net land-atmosphere flux	−0.3 ± 0.9	−1.2 ± 0.8
3. Houghton 2003		
Land use change	+2.0 (+0.9 to +2.8)	+2.2 (+1.4 to +3.0)
Residual terrestrial sink	−2.3 (−4.0 to −0.3)	−3.4 (−5.0 to −1.8)
4. DeFries et al. 2002		
Land use change	+0.6 (+0.3 to +0.8)	+0.9 (+0.5 to +1.4)
Residual terrestrial sink	−0.9 (−3.0 to 0)	−2.1 (−3.4 to −0.9)
5. Achard et al. 2002		
Land use change		+1.0 ± 0.2
Residual terrestrial sink		−2.2 (−3.2 to −1.2)

Note: Positive values represent atmospheric increase (or ocean/land sources); negative numbers represent atmospheric decrease (sinks). Residual terrestrial sink determined by difference (net land/atmosphere flux minus land use change).

geological cycle and the composite estimates of CO_2 emissions from geological reservoirs (i.e., fossil fuels and the production of lime for cement). Panels *b* and *c* provide more detailed pictures of the ocean and terrestrial fluxes, respectively. Individual component pools and fluxes, including key climatic and anthropogenic drivers, are discussed later in this chapter and in subsequent chapters.

Although Colorplate 1 focuses primarily on fluxes directly related to CO_2, a number of non-CO_2 trace gases also play significant roles in the global carbon cycle (e.g., CO, CH_4, non-methane hydrocarbons) and/or climate forcing (e.g., CH_4, N_2O, chlorofluorocarbons). The global cycles of these trace gases, which share many of the processes driving the CO_2 cycle, are briefly discussed later in this chapter.

Reservoir Connections

The background chapters in this volume discuss the various carbon reservoirs and the processes relevant to controlling atmospheric CO_2 and related trace gas concentrations. To appreciate the Earth's carbon cycle and its evolution, it is necessary to exam-

ine the connections among the various carbon pools. This analysis must be done within a framework that provides an integrated perspective across both disciplinary and geographic boundaries, with particular emphasis on the carbon cycle as an integral part of the human-environment system.

Fossil Fuel–Atmosphere Connections

The world energy system delivered approximately 380 exajoules (EJ [10^{18} J]) of primary energy in 2002 (BP 2003). Of this, 81 percent was derived from fossil fuels, with the remainder derived from nuclear, hydroelectric, biomass, wind, solar, and geothermal energy sources (Figure 2.1). The fossil-fuel component released 5.2 PgC in 1980 and 6.3 PgC in 2002 (CDIAC 2003). Cement production is the other major industrial source of carbon, and its release increased to 0.22 PgC in 1999. The combined release of 5.9 PgC shown in Colorplate 1a represents an average emission for the 1980s and 1990s. Underground coal fires, which are poorly known and only partly industrial, may be an additional as yet unaccounted for source of carbon to the atmosphere as large as cement manufacturing (Zhang et al. 1998). In terms of energy released, the current mix of fossil fuels is approximately 44 percent oil, 28 percent coal, and 27 percent natural gas (Figure 2.1). At current rates of consumption, conventional reserves of coal, oil, and gas (those that can be economically produced with current technology; see Colorplate 1c) are sufficient to last 216, 40, and 62 years, respectively (BP 2003). If estimates of undiscovered oil and gas fields are included with the conventional reserves, oil and gas lifetimes increase to 101 and 142 years, respectively (Ahlbrandt et al. 2000).

Conventional reserves represent only a fraction of the total fossil carbon in the Earth's crust. A much larger quantity of fossil reserves is in tar and heavy oil. These reserves cannot be economically produced with existing technology but are likely to become accessible in the future. The best estimates for total fossil resources that might ultimately be recovered are in the range of 6,000 Pg (Nakicenovic, Chapter 11, this volume), or about five times the conventional reserves. In addition, vast quantities of methane, exceeding all known fossil-fuel reserves, exist in the form of methane hydrates under continental shelf sediments around the world, in the Arctic permafrost, and in various marginal seas. With current technology, however, these reserves do not appear viable as a future energy source.

Although not included in Table 2.1 or Colorplate 1, combustion of fossil fuels also releases a number of non-CO_2 carbon gases. In particular, carbon monoxide (CO) can be used as an effective tracer of fossil-fuel combustion in atmospheric gas measurements. The relative impact of these gases is discussed in a later section.

Land-Atmosphere Connections

The exchange of carbon between the terrestrial biosphere and the atmosphere is a key driver of the current carbon cycle. Global net primary production (NPP) by land plants
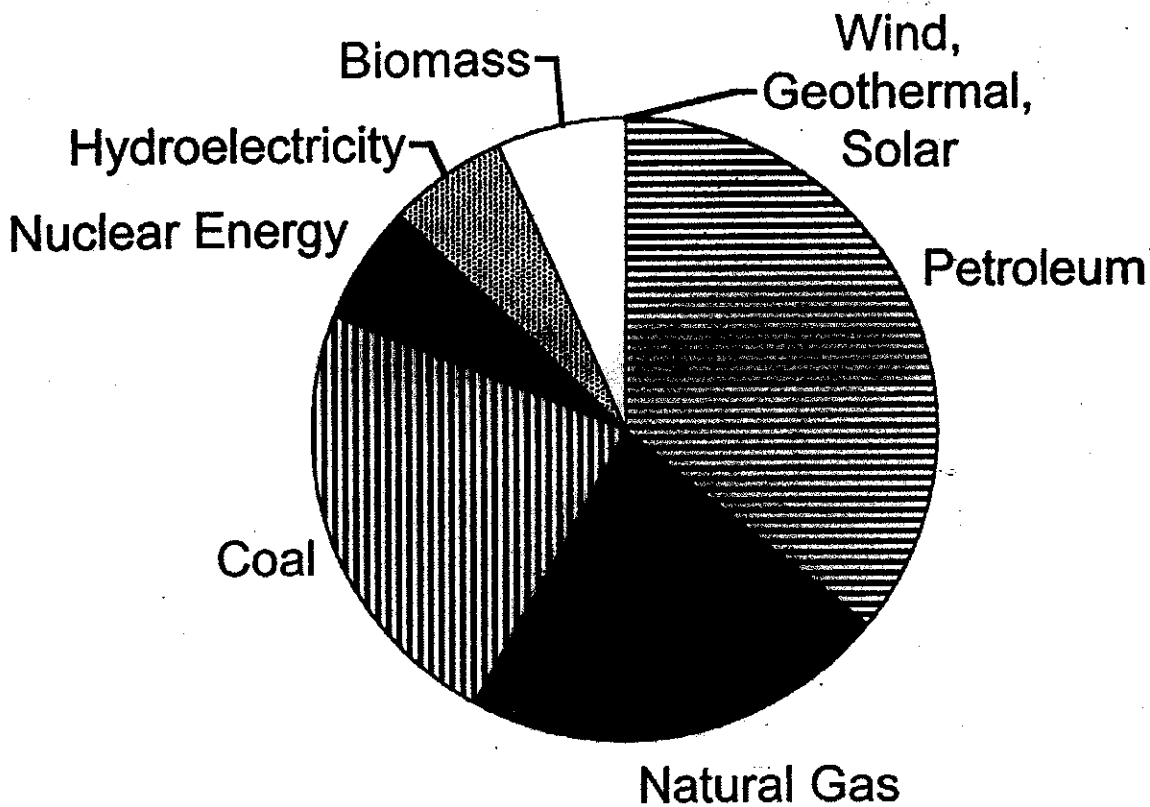
Figure 2.1. The distribution of sources for the world energy system in 2000. The values for all sources except biomass are from BP (2003). The value for biomass is from the U.S. Department of Energy (http://www.eere.energy.gov).

is about 57 PgC per year (y^{-1}) (Colorplate 1). Of this, about 4 PgC y^{-1} is in crops. Humans co-opt a much larger fraction of terrestrial NPP, probably about 40 percent, where co-opting is defined as consuming, removing some products from, or altering natural states of the terrestrial biosphere through ecosystem changes (Vitousek et al. 1986). Total NPP is approximately 40 percent of gross primary production (GPP), with the remainder returned to the atmosphere through plant respiration. For many purposes, NPP is the most useful summary of terrestrial plant activity. NPP can be assessed with inventories and harvest techniques, and it represents the organic matter passed to other trophic levels (Lindemann 1942). For other purposes, including isotope studies and scaling from eddy flux, GPP is a more useful index.

Most of the annual flux in NPP is returned to the atmosphere through the respiration of heterotrophs, including microorganisms, saprophytes, and animals. A smaller fraction, 5–10 percent, is released to the atmosphere through combustion (Colorplate 1). Approximately 5 percent of NPP leaves land ecosystems in organic form, as CH_4 or volatile organic carbon (Prinn, Chapter 9, this volume). In recent decades the land was close to carbon neutral (in the 1980s) or was a net sink (in the 1990s) for atmospheric CO_2 (Table 2.1). This net flux represents a balance between substantial emissions from biomass clearing or fires and enhanced uptake as a physiological response to rising CO_2 or the regrowth of previously cleared areas.

Land plants contain slightly less carbon than the atmosphere; soils contain substan-

tially more (Colorplate 1). The estimates for soil carbon in Colorplate 1 are higher than shown in previous budgets for two reasons. First, previous budgets estimated soil carbon to a depth of 1 meter (m). Jobaggy and Jackson (2000) extended these to 3 m, adding about 55 percent to the known stock. Second, previous budgets ignored carbon in wetland and permanently frozen soils. The exact magnitudes of these stocks are very uncertain. The potential for substantial carbon losses from these stocks in coming decades, however, is a strong incentive for careful quantification and further analysis.

Tropical forests contain the largest carbon pool of terrestrial biota (see Table 2.2) and also the largest NPP (Saugier et al. 2001). Averaged over several years, tropical forests traditionally have been believed to be close to carbon neutral, with uptake from NPP balanced by releases from decomposition and fire. The difference between NPP and the sum of all of the processes that release carbon from the land is net ecosystem production (NEP). Recent studies based on in situ flux measurements and forest inventory techniques indicate net carbon uptake or positive NEP of 100–700 gC per square meter (m^{-2}) y^{-1}, corresponding to ~1 PgC y^{-1} across the tropical evergreen forest biome. The duration and spatial scale of these studies is not yet extensive enough for solid extrapolation. Still, the estimated uptake is in the proper range to compensate for emissions from tropical land use (Achard et al. 2002; DeFries et al. 2002; Foley and Ramankutty, Chapter 14, this volume). The tropical NEP estimates, however, are also in the range of recent estimates of CO_2 releases from tropical rivers, caused mainly by the decomposition of material transported from the land (Richey, Chapter 17, this volume). A full carbon accounting has yet to be accomplished, but these individual flux estimates need to be reconciled with top-down estimates from atmospheric inverse modeling studies, which show that the overall net carbon balance of the tropical land areas (30°S–30°N) must be close to zero, albeit with large uncertainty ranges (Gurney et al. 2002; Heimann et al., Chapter 8, this volume).

Temperate forests cover about 60 percent of the area of tropical forests and contain the second largest pool of plant carbon. Together, tropical and boreal forests account for approximately 75 percent of the world's plant carbon and for 40 percent of the non-wetland, nonfrozen soil carbon (Table 2.2).

Carbon emissions from land use and land management have increased dramatically over the past two centuries because of the expansion of cropland and pasture, infrastructure extension, and other effects driven by market growth, pro-deforestation policies, and demographic pressures (Geist and Lambin 2001). Before about 1950, carbon emissions from land use change were mainly from temperate regions. In recent decades, however, carbon releases from land use change have been concentrated in the tropics (Achard et al. 2002; DeFries et al. 2002; Houghton 2003). Cumulative emissions from land use, estimated at approximately 185 PgC, entail substantial fluxes from every continent except Antarctica (DeFries et al. 1999). Before about 1970, cumulative emissions from land use and land management were larger than cumulative emissions from fossil-fuel combustion.

Table 2.2. Plant carbon, soil carbon, and net primary production in the world's major biomes

Ecosystem	Area (10⁶ km²)	NPP (PgC y⁻¹)	Plant C (PgC)	Soil C (Pg C)
Tropical forests	17.5	20.1	340	692
Temperate forests	10.4	7.4	139	262
Boreal forests	13.7	2.4	57	150
Arctic tundra	5.6	0.5	2	144
Mediterranean shrublands	2.8	1.3	17	124
Crops	13.5	3.8	4	248
Tropical savannas and grasslands	27.6	13.7	79	345
Temperate grasslands	15	5.1	6	172
Deserts	27.7	3.2	10	208
Subtotal	*149.3*	*57.5*	*652*	*2,344*
Wetlands				450
Frozen soils	25.5			400
Total	*174.8*	*57.5*	*652*	*3,194*

Note: Plant carbon is from Saugier et al. (2001). NPP is from Saugier et al. (2001), scaled to give the terrestrial total as the satellite study of Behrenfeld et al. (2001). Soil carbon is from Jobbagy and Jackson (2000), to a depth of 3 m. Wetland soil carbon is from (Gorham 1991). Carbon in frozen (nonwetland) soils is from (Zimov et al. 1997).

As with fossil-fuel combustion, carbon is lost from terrestrial vegetation in a variety of non-CO_2 gases, including biogenic volatile organic compounds (VOCs). Globally VOC emissions are 0.2–1.4 PgC y⁻¹ (Prinn, Chapter 9, this volume). For grassland and cultivated land, large emissions of other greenhouse gases such as N_2O and CH_4 can also occur. These emissions are strongly affected by land management and land use change (Smith, Chapter 28, and Robertson, Chapter 29, both this volume).

Ocean-Atmosphere Connections

The oceans contain about fifty times more CO_2 than the atmosphere and ten times more than the latest estimates of the plant and soil carbon stores (Colorplate 1). CO_2 moves between the atmosphere and the ocean by molecular diffusion, when there is a difference between the CO_2 gas pressure (pCO_2) in the oceans and the atmosphere. For example, when the atmospheric pCO_2 is higher than the surface ocean pCO_2, CO_2 diffuses across the air-sea boundary into the seawater. Based on about 940,000 measurements of surface water pCO_2 obtained since the 1960s, the climatological, monthly distribution of pCO_2 in the global surface waters has been calculated with a spatial resolution of 4° × 5° (Takahashi et al. 2002). Although the published value for the net

air-sea flux indicates a net ocean uptake of 2.1 PgC for the reference year 1995, this estimate was based on inappropriate wind speed estimates (T. Takahashi, personal communication). The winds used for these estimates correspond to approximately 40 m above mean sea level, but the gas exchange coefficient formulas are typically related to winds at 10 m above sea level. The near-surface gradient in wind speed is such that the 10 m winds are about 1 m per second (s^{-1}) slower than the 40 m winds. Thus, Takahashi's corrected net global ocean uptake is 1.5 (-19 percent to +22 percent) PgC for the reference year 1995 (Colorplate 2). The asymmetrical error estimates reflect the paucity of data on the spatial and temporal variability in sea surface pCO_2 concentrations, plus limitations in the non-linear wind speed parameterizations for the gas exchange coefficient. Ocean models and observations suggest that the interannual variability in the global ocean CO_2 flux is around ±0.5 PgC y^{-1} (Greenblatt and Sarmiento, Chapter 13, this volume).

The gross exchanges of CO_2 across the air-sea interface, as shown in Colorplate 1, are much larger than the net flux. The global budget presented in this figure shows the preindustrial oceans as a net source of -0.6 PgC y^{-1} to the atmosphere, partially offsetting the addition of carbon to the oceans from rivers. The total net modern flux of Takahashi et al. (2002), after the wind speed correction discussed previously (ocean uptake = 1.5 PgC y^{-1}), is consistent with the modern balance implied by Colorplate 1 (21.9 − 20 + 70 − 70.6 = 1.3 PgC y^{-1}), despite the fact that the coastal zones are not well represented in the Takahashi et al. analysis. The coastal zone fluxes represent the largest unknown in the CO_2 balance of the oceans and are a topic of active research (Chen, Chapter 18, this volume).

During the past decade, significant advances have been made in separating the anthropogenic component from the large background of ocean dissolved inorganic carbon (DIC). Data-based approaches estimate a global inventory of anthropogenic CO_2 in the oceans to be -112 ± 17 PgC for a nominal year, 1994 (Colorplate 3) (Lee et al. 2003). Inventories are generally high in the mid-latitudes and lowest in the high-latitude Southern Ocean and near the Equator.

The high-inventory regions are convergence zones, where waters with relatively high anthropogenic concentrations are moving into the ocean's interior. Roughly 25 percent of the total inventory of anthropogenic carbon is in the North Atlantic, one of the main regions of deepwater formation. The low-inventory waters are generally regions of upwelling, where waters with low anthropogenic concentrations are brought near the surface. The high-latitude Southern Ocean generally has very low anthropogenic CO_2 inventories and very shallow penetration. The Southern Hemisphere mode and intermediate waters at around 40–50°S, on the other hand, contain some of the largest inventories of anthropogenic CO_2. More than 56 percent of the total anthropogenic CO_2 inventory is stored in the Southern Hemisphere.

Over the long term (millennial timescales), the ocean has the potential to take up approximately 85 percent of the anthropogenic CO_2 that is released to the atmosphere.

The reason for the long time constant is the relatively slow ventilation of the deep ocean. Most of the deep and intermediate waters have yet to be exposed to anthropogenic CO_2. As long as atmospheric CO_2 concentrations continue to rise, the oceans will continue to take up CO_2. This reaction, however, is reversible. If atmospheric CO_2 were to decrease in the future, the recently ventilated waters would start releasing part of the accumulated anthropogenic CO_2 back to the atmosphere, until a new equilibrium is reached.

Land-River-Ocean Connections

Carbon is transported from the land to the oceans via rivers and groundwater. The transfer of organic matter from the land to the oceans via fluvial systems is a key link in the global carbon cycle. Rivers also provide a key link in the geological-scale carbon cycle by moving weathering products to the ocean (Colorplate 1). The conventional perspective is that rivers are simply a conduit for transporting carbon to the ocean. Reevaluation of this model suggests that the overall transfer of terrestrial organic matter through fluvial systems may be more complex (Richey, Chapter 17, this volume). A robust evaluation of the role of rivers, however, is complicated by both the diverse dynamics and multiple time constants involved and by the fact that data are scarce, particularly in many of the most affected systems.

Humans have had a significant impact on the concentrations of carbon and nutrients in river systems. Intensifying agriculture has led to extensive erosion, mobilizing perhaps 10–100 times more sediment, and its associated organic carbon, than undisturbed systems. Sewage, fertilizers, and organic waste from domestic animals also contribute to the carbon and nutrient loads of rivers. Not all of these materials make it to the ocean. Much can be deposited near and along river channels. The retention of particulate material in aquatic systems has increased since preindustrial times because of the proliferation of dams (primarily in the 30–50°N regions), which has increased the average residence time of waters in rivers. If the carbon mobilized via erosion were subsequently replaced by newly fixed carbon in agriculture, then a sink on the order of 0.5–1.0 PgC y^{-1} would be created. But organic carbon does not move passively through river systems; even very old and presumably recalcitrant soil carbon may be at least partially remineralized in aquatic systems. Remineralization of organic carbon during transport leads to elevated levels of dissolved CO_2 in rivers, lakes, and estuaries worldwide. These high concentrations subsequently lead to outgassing to the atmosphere on the order of ~1 PgC y^{-1}, with the majority in the humid tropics (Richey, Chapter 17, this volume).

Despite the increased particulate retention in rivers, a significant amount of carbon escapes to the ocean. DIC, particulate inorganic carbon (PIC), dissolved organic carbon (DOC), and particulate organic carbon (POC) exported from rivers to the coastal ocean are 0.4, 0.2, 0.3, and 0.2 PgC y^{-1}, respectively (Chen, Chapter 18, this volume). These fluxes include a pronounced, but difficult to quantify, anthropogenic component. These values are poorly constrained by direct measurements and may represent minimum esti-

mates of inputs to the coastal ocean (Richey, Chapter 17, this volume). Groundwater discharges, which make up about 10 percent of the surface flow to the ocean, also contribute poorly known contributions of carbon and nutrients to the coastal oceans.

The final step in the land-to-ocean pathway is the marine fate of fluvial carbon. Previous estimates have suggested that the marginal seas are net heterotrophic. Chen (Chapter 18, this volume) has suggested that continental shelves are in fact net autotrophic, mostly because of production from coastal upwelling of nutrient-rich waters. If this is correct, biological production in the coastal zone may decrease the thermodynamic drive to outgas the terrestrial carbon delivered by rivers, resulting in greater preservation in the marine environment.

Trade and Commerce Connections

The lateral fluxes of carbon are poorly constrained by measurements and are frequently excluded from models. The river fluxes outlined in the previous section are only beginning to be quantified. The lateral transport of carbon through trade has not been constrained or quantified in global carbon budgets based on inventories of terrestrial carbon stocks and fluxes. Approximately one-third of the 4 PgC fixed annually through agriculture represents harvested products that are directed toward some form of human or domestic animal consumption (Tschirley and Servin, Chapter 21, this volume). In 2001, US$547 million in agricultural and forest-related products entered into international trade, amounting to 9 percent of total trade (excluding services). Imports and exports of cereals, wood, and paper products accounted for about 0.72 PgC of "embodied" carbon traded in 2000.

As we move from the basic carbon budget presented in Table 2.1 to a more detailed review of the fluxes shown in Colorplate 1 and to regional assessments of carbon budgets, it becomes increasingly important to recognize and quantify the lateral transports of carbon. The transport of carbon through trade and commerce represents a flux that is nearly as large as the net terrestrial and oceanic sinks. This carbon is assumed to be respired back to the atmosphere, but if the CO_2 is taken out of the atmosphere on one continent, then transported to another continent before being respired, it can have a significant impact on regional budget assessments.

Embodied carbon represents the carbon that resides directly within the product itself (for cereals, wood, and paper, about half of the final weight is carbon). In addition to embodied carbon, at least two other components are usually excluded from estimates of carbon content in international trade. One is the "production" carbon, which comprises the inputs required to produce a final product, primarily energy-related consumption to provide inputs (e.g., pesticides, fertilizers) and processing (e.g., machinery for harvesting, milling, sawing). A second element is the "transport" carbon, which represents the energy costs of transporting the processed products to a final destination. Countries that use intensive production systems and engage in significant exports (pri-

marily in the developed countries) would have higher ratios of carbon content to product mass exported, possibly up to 20 percent, and would therefore have higher carbon emissions to the atmosphere than currently stated, under a system of full carbon accounting. Other countries would have lower carbon release to the atmosphere under full carbon accounting. Trade and commerce carbon accounting has a political background and differs from budget studies by atmospheric inversion and inventory checks. For example, CO_2 emissions during production and transport of products across regional boundaries would fully and partly, respectively, be accounted for by the country of origin using inventory studies. Inverse models based on atmospheric measurements, however, would reflect the emissions as they were actually distributed over the regions. As we begin to move from the global carbon budget to regional carbon budgets, fluxes like those in trade and commerce will need to be better quantified.

Non-CO₂ Trace Gas Connections

Rising concentrations of a large number of potent anthropogenic greenhouse gases other than carbon dioxide have collectively contributed an amount of radiative forcing comparable to that of CO_2 since preindustrial times. Many of these non-CO_2 gases (e.g., CH_4, N_2O, CF_2Cl_2, SF_6) are emitted at the Earth's surface and contribute directly to this forcing (Prinn, Chapter 9, this volume). They are characterized by atmospheric lifetimes of decades to millennia (lifetime as used here is the amount of the gas in the global atmosphere divided by its global rate of removal). Other non-CO_2 gases (e.g., isoprene, terpenes, nitric oxide [NO], carbon monoxide [CO], sulphur dioxide [SO_2], dimethyl sulphide [$CH_3]_2S$), most of which are also emitted at the Earth's surface, contribute indirectly to this forcing through production of either tropospheric ozone (which is a powerful greenhouse gas) or tropospheric aerosols (which directly reflect sunlight back to space, absorb it, or indirectly change the reflection properties of clouds). This second group of climatically important non-CO_2 gases is characterized by much shorter lifetimes (hours to months).

The role of non-CO_2 carbon gases is not typically included in global carbon budgets, because the sources and sinks for these gases are not well understood. Figure 2.2 summarizes, albeit very simply, the basic cycles and fluxes of the major non-CO_2 trace gases relevant to climate. Note that the emissions of the carbon-containing gases alone contribute about 2.3 PgC y^{-1} to the carbon cycle (Prinn, Chapter 9, this volume). To aid the handling of the non-CO_2 gases in the policy processes under the United Nations Framework Convention on Climate Change (UNFCCC), scientists have calculated so-called global warming potentials (GWPs). These dimensionless GWPs, which range from 20 to 20,000 for the major non-CO_2 gases, are intended to relate the time-integrated radiative forcing of climate by an emitted unit mass of a non-CO_2 trace gas to the forcing caused by emission of a unit mass of CO_2. The GWP concept has difficulties, because the removal mechanisms for many
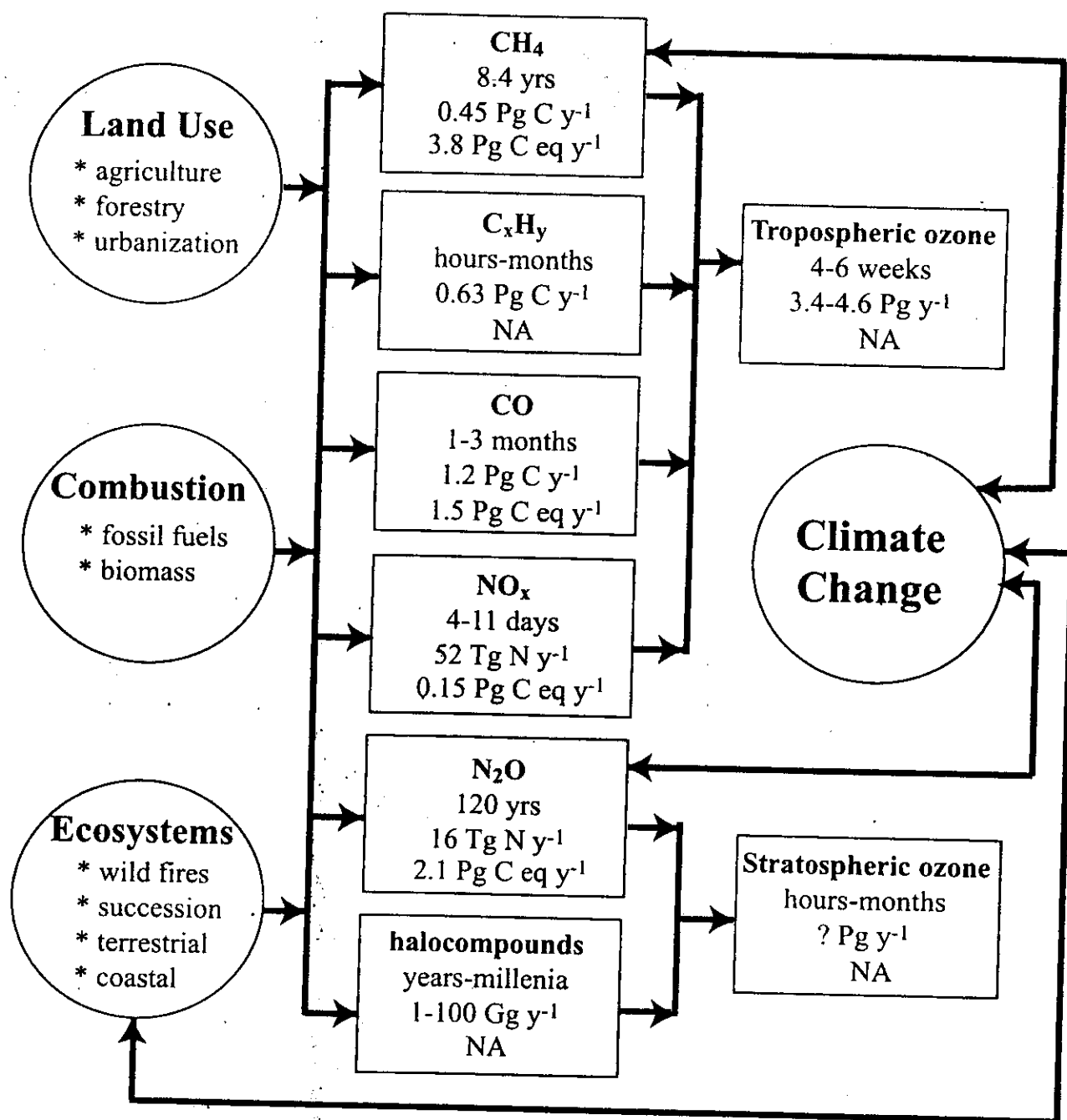
Figure 2.2. A summary of the sources, quantities, lifetimes, and consequences of carbon-containing non-CO_2 gases in the atmosphere. The data are from sources summarized in Prinn (Chapter 9, this volume). NA = not available.

gases (including CO_2 itself) involve complex chemical and/or biological processes and because the time period (e.g., decade, century) over which one integrates the instantaneous radiative forcing of a gas to compute its GWP is somewhat arbitrary (Manne and Richels, Chapter 25, this volume). Nevertheless, by multiplying the emissions of each major non-CO_2 greenhouse gas by its GWP, we obtain equivalent amounts of CO_2 emissions, which are comparable to the total actual emissions of CO_2 (e.g., current total emissions of CH_4 and N_2O are equivalent to 3.8 and 2.1 PgC y^{-1}, respectively). For this reason, a "CO_2 emissions only" approach for global warming pol-

icy may lead to significant biases in the estimation of global warming abatement costs (Manne and Richels, Chapter 25, this volume). Thus, a multi-gas approach for studying the carbon cycle and how it relates to climate change needs to be implemented.

Carbon Cycle Processes

To fully understand the significance of the reservoir connections, as well as the possibility that these fluxes might change in the future, one must consider the processes responsible for controlling the fluxes within and between reservoirs. In some cases human activities have already significantly affected these processes. In other cases, the processes may be vulnerable to climate change in the future.

Terrestrial Processes

On seasonal timescales the gains and losses from the terrestrial biosphere are reflected in the cyclic variations in atmospheric CO_2 concentrations. Although there is significant interannual variability in the seasonal balance between the gains and losses, it is the trends in this net balance between NPP and carbon losses through decomposition, fires, and disturbances that can affect climate on timescales of decades to centuries. The scientific community currently believes that the terrestrial biosphere has been acting as a net sink for atmospheric carbon for the past few decades (Foley and Ramankutty, Chapter 14, this volume). For example, Pacala et al. (2001) recently estimated the carbon sink of the coterminous Unites States using a combination of inventory and atmospheric concentration inversions. They determined a consistent land- and atmosphere-based carbon sink for the United States in the range of 0.37–0.71 PgC y^{-1}. Similar studies have also shown a net sink in Europe (Janssens et al. 2003). The mechanisms underlying the sinks have been the subject of much recent research. The leading hypothesis in the 1970s and 1980s was that the sinks were mainly a result of more rapid plant growth from elevated CO_2 and climate change. This hypothesis has gradually been replaced with a multi-mechanism explanation, including contributions from changes in forest management, agriculture, long-lived products (e.g., wood), aquatic systems, and nitrogen deposition, in addition to CO_2 fertilization and changes in plant growth and soil carbon pools resulting from climate change (Schimel et al. 2001).

NPP is sensitive to a broad range of factors, including climate, soil fertility, atmospheric CO_2, atmospheric pollutants, and human management. An increase in NPP can lead to a carbon sink, but only if it is not matched by a corresponding increase in carbon losses. In general, processes that promote gradual increases in NPP can lead to carbon sinks, because increases in respiration tend to follow changes in biomass and soil carbon and not NPP directly (Field 1999). For the past 30 years, carbon cycle researchers have hypothesized that gradual increases in NPP in response to the 30 percent rise in atmospheric CO_2 since preindustrial times explain much or all of the terrestrial

sink inferred from atmospheric studies (Bacastow and Keeling 1973). Experimental studies at the ecosystem level often demonstrate accelerated plant growth in response to elevated atmospheric CO_2 (Mooney et al. 1999), but these growth rates only explain a small fraction of the sink required to balance the global atmospheric CO_2 budget (Friedlingstein et al. 1995). Other factors that can increase NPP, including warming, greater precipitation and humidity, nitrogen deposition, and changes in plant species composition, may also contribute to terrestrial carbon sinks.

Processes that retard carbon releases can also lead to terrestrial sinks. Evidence indicates that a substantial fraction of the forest sink in temperate forests is a result of changes in land use and land management. In essence, forests, cut in the past, are regrowing, with a growth rate that is faster than the rate of harvesting (Goodale et al. 2002). Some of this change is a result of shifts in land use, especially the abandonment of agriculture over large regions of North America and parts of Europe (Foley and Ramankutty, Chapter 14, this volume). Fire suppression and the thickening (more trees per unit of area) of marginal forests also contribute to increased forest biomass in some areas (Pacala et al. 2001).

Changes in agriculture can also promote terrestrial carbon sinks. Increases in yield, incorporation of crop residue, and areas in perennial crops, as well as a reduction in tillage, can all contribute to carbon sinks (Lal et al. 1998).

Recent terrestrial carbon sinks reflect a number of mechanisms (Pacala et al. 2001). The potential for sinks to persist varies from mechanism to mechanism. Sinks due to CO_2 fertilization are likely to persist until NPP is limited by another factor. In some settings, this limitation might occur very soon, and in others it might be far in the future. Sinks caused by forest regrowth saturate as the forests mature (Nabuurs, Chapter 16, this volume). In general, sinks saturate when increases in NPP are outpaced by increases in the sum of decomposition and combustion. The persistence of terrestrial sinks in the future is by no means assured. Many lines of evidence suggest that the prospect of increasing carbon sources in the terrestrial biosphere is a real possibility (Gruber et al., Chapter 3, this volume).

Ocean Processes

Air-sea gas exchange is a physico-chemical process, primarily controlled by the air-sea difference in gas concentrations and the exchange coefficient, which determines how quickly a molecule of gas can move across the ocean-atmosphere boundary (see Le Quéré and Metzl, Chapter 12, this volume). It takes about one year to equilibrate CO_2 in the surface ocean with atmospheric CO_2, so it is not unusual to observe large air-sea differences in CO_2 concentrations (Colorplate 2). Most of the differences are caused by variability in the oceans due to biology and ocean circulation. The oceans contain a large reservoir of carbon that can be exchanged with the atmosphere (see Colorplate 1). Since air-sea exchange can only occur at the surface, however, the rate at which carbon

exchanges between the surface and the ocean interior ultimately regulates how well the atmosphere equilibrates with the ocean as a whole.

Two basic mechanisms control the natural distribution of carbon in the ocean interior: the solubility pump and the biological pump. The solubility pump is driven by two principle factors. First, more CO_2 can dissolve into cold polar waters than in the warm equatorial waters. As major ocean currents (e.g., the Gulf Stream) move waters from the tropics to the poles, they are cooled and can take up more CO_2 from the atmosphere. Second, the high-latitude zones are places where deepwater is formed. As the water cools, it becomes denser and sinks into the ocean's interior, taking with it the CO_2 accumulated at the surface.

The biological pump also transports CO_2 from the surface to the deep ocean. Growth of phytoplankton uses CO_2 and other chemicals from the seawater to form plant tissue. Roughly 70 percent of the CO_2 taken up by phytoplankton is recycled near the surface, and the remaining 30 percent sinks into the deeper waters before being converted back into CO_2 by marine bacteria (Falkowski et al. 1998). Only about 0.1 percent of the organic carbon fixed at the surface reaches the seafloor to be buried in the sediments. The carbon that is recycled at depth is transported large distances by currents to areas where the waters return to the surface (upwelling regions). When the waters regain contact with the atmosphere, the CO_2 originally taken up by the phytoplankton is returned to the atmosphere. This exchange helps to control atmospheric CO_2 concentrations over decadal and longer time scales.

The amount of organic carbon that is formed and sinks out of the surface ocean is limited by the availability of light and nutrients (mainly nitrate, phosphate, silicate, and iron) and by temperature. The plankton types present in the water also play a role. Plankton that bloom create favorable conditions for the formation of fast-sinking particles, particularly when they have shells of calcium carbonate or silicate (Klaas and Archer 2002). The formation of calcium carbonate shells affects carbon chemistry in such a way that it works to counteract the drawdown of CO_2 by soft tissue production. The biological pump also removes inorganic nutrients from surface waters and releases them at depth. Since productivity is limited by the availability of these nutrients, the large-scale thermohaline circulation (THC) of the oceans has a strong impact on global ocean productivity by regulating the rate at which nutrients are returned to the surface.

Up to now, humans have had a relatively small direct impact on the global-scale ocean carbon cycle. This is primarily because humans generally only transit across the ocean and because the ocean naturally contains orders of magnitude more carbon than the atmosphere and the terrestrial biosphere. The ocean does, however, act as a significant sink for CO_2 ultimately derived from anthropogenic activities (Table 2.1). Because biology is not limited by carbon in the oceans, it is thought that increasing CO_2 levels have not significantly affected ocean biology. The current distribution of anthropogenic CO_2 is assumed to result from physico-chemical equilibration of the surface ocean with rising atmospheric CO_2 and slow mixing of the anthropogenic CO_2 into the

ocean's interior. The long residence time for the deep oceans means that most of the deep ocean waters have not been exposed to the rising atmospheric CO_2 concentrations observed over the past couple of centuries. Although the oceans have the potential to absorb 85 percent of the anthropogenic CO_2 released to the atmosphere, today's oceans are only at about 15 percent capacity (Le Quéré and Metzl, Chapter 12, this volume). Average penetration depth for anthropogenic CO_2 in the global ocean is only about 800 m (Sabine et al. 2002). There is growing evidence, however, that changes in ocean mixing and biology may be occurring as a result of climate change.

Since the solubility of CO_2 is a function of temperature, warming of the ocean will decrease its ability to absorb CO_2. Furthermore, changes in temperature and precipitation may lead to significant alterations of ocean circulation and the transport of carbon and nutrients to and from the surface. Because of its effect on ocean carbon distributions and biological productivity, changes in the THC have been used to help explain past excursions in climate and atmospheric CO_2, including glacial-interglacial and Dansgard-Oeschger events (Joos and Prentice, Chapter 7, this volume).

Ocean productivity can also be affected by atmospheric inputs that may change as a result of human activity. Iron in oceanic surface waters originates from terrestrial dust deposited over the ocean, deep ocean waters, continental shelves, and to a lesser extent river inflow. Because of the spatial distribution of dust deposition and other iron sources, large regions of the ocean show a deficit in iron (and to a smaller extent in silicate), although other nutrients are plentiful. These are called high-nutrient low-chlorophyll (HNLC) regions. There is a potential for enhanced biological productivity in these regions if the ocean can be "fertilized" by iron. The potential for CO_2 reduction in surface waters and in the atmosphere through this artificial sequestration, however, depends on factors like the composition of plankton types and oceanic circulation, and can lead to undesirable side effects (Bakker, Chapter 26, this volume).

There is also increasing evidence that rising CO_2 levels may directly affect ocean productivity and ecosystem structure. For example, Riebesell et al. (2000) showed a significant reduction in the ability of two different species of coccoliths to secrete calcium carbonate shells under elevated CO_2 conditions. Similar reductions in calcification have been observed in corals and coralline algae. As atmospheric CO_2 concentrations continue to rise, the potential for significantly altering the current balance between the amount of carbon moved into the ocean's interior by the biological pump versus the solubility pump in the ocean increases. The net effect on the ability of the ocean to act as a sink for anthropogenic CO_2 is not clear.

Coastal Ocean

Although the coastal zones, consisting of the continental shelves with depths less than 200 m including bays and estuaries, occupy only 10 percent of the total ocean area, they play a crucial role in the global carbon cycle. Carbon is transported to the coastal zone

by riverine inputs and transport of inorganic carbon from the open ocean. Estuaries and proximal coastal seas are believed to be sources of CO_2 because of the decay of terrestrial organic carbon. Since the riverine flux of nutrients has risen continually over the past few decades, however, these areas may now have enhanced biological productivity and hence may be releasing less CO_2. At present a large fraction (~80 percent) of the land-derived organic and inorganic materials that are transported to the ocean is trapped on the proximal continental shelves (Mackenzie and Ver 2001). The much wider open shelves, on the other hand, probably serve as sinks for atmospheric carbon. A recent overview suggests that the global coastal waters and marginal seas (extending to a water depth of 200 m) are now absorbing about 0.36 PgC y^{-1} from the atmosphere (Chen, Chapter 18, this volume).

Across most of the coastal seas and continental margins, surface waters are transported offshore because of fresh water inputs from land. This surface transport draws nutrient-rich subsurface waters from the open ocean onto the shelves. Such external sources of nutrients support high primary productivity. Most of the organic material produced is respired and recycled on the shelves. The organic matter that is not recycled either accumulates in the sediments or is exported to the slopes and open oceans. The coastal zone may account for 30–50 percent of the total calcium carbonate accumulation and up to 80 percent of the organic carbon accumulation in ocean sediments (Mackenzie and Ver 2001). Globally, the shelf seas are estimated to transport 0.6 PgC y^{-1} of DOC, 0.5 PgC y^{-1} of POC, and 0.2 PgC y^{-1} of PIC to the open oceans (Chen, Chapter 18, this volume). Although these transports have large uncertainties because of high variability and inadequate data coverage, they represent an important and often neglected link in the global carbon cycle. Shelves and estuaries are also important sources of other greenhouse or reactive gases, such as methane and dimethyl sulfide.

Finally, although humans appear to have had only a small direct impact on the open ocean, they have had a profound and poorly understood impact on the coastal oceans. Furthermore, direct and indirect human perturbations vis-à-vis the continental margins (e.g., pollution, eutrophication) are likely to have large and dire consequences on marine ecosystems in the future. As much as 40–60 percent of the world population lives in coastal areas, depending on the definitions and methodologies applied. The coastal regions have the most rapidly growing populations, because of migration from rural areas. Most megacities in this century will develop in coastal zones, approximately half of them in Asia (IHDP 2000). Such patterns of urbanization have affected and will continue to affect coastal and marine systems, through processes such as land use change, pressure on infrastructure (water, sewage, and transportation), coastal resource depletion and degradation, eutrophication, and other carbon-relevant impacts. Coastal zones support more than 60 percent of the global commercial fish production (World Resources Institute 1996). It is not clear how these fisheries will be affected by the increasing human pressures on the coastal zone. Additional studies of the coastal zone and the interaction between humans and the coastal biogeochemical systems are needed.

Human Systems

Humans have had a profound impact on carbon cycling in the atmosphere, the terrestrial biosphere, and, to a lesser extent, the oceans. To understand how humans will continue to interact with these reservoirs in the future, one must understand the drivers responsible for how humans interact with the environment.

Energy use has historically been viewed as an essential commodity for economic growth. The paradigm of "grow or die" historically meant increasing demands for additional resources and a commensurate increase in the amount of pollutants released. Over the past 30 years, however, a complex debate has ensued about how to decouple economic growth and resource consumption. Since this debate started, many scholars, policy makers, and nongovernmental organizations have asked themselves how to reshape this paradigm of development. Answers have ranged from doing nothing (business-as-usual approach) to slowing and even stalling economic growth, so as to not exceed the Earth's carrying capacity. Carrying capacity here refers to the ability of natural resources and ecosystems to cope with anthropogenic pressures, such as use of renewable natural resources, emission of pollutants, and modification of ecosystem structures without crossing "critical thresholds of damage beyond which [these resources] lose their ability for self-renewal and slide inexorably into deeper degradation" (Board on Sustainable Development Policy Division 1999).

Another more tenuous response is based on the idea that one could keep economic growth as a development goal and at the same time find mechanisms aimed at reducing the amount of material input and emissions or aimed at finding replacements for nonrenewable or dangerous resources (e.g., fossil fuels). This proposal has resulted in environmental policies and instruments promoting technological innovations that seek to increase the efficiency of economic activities and to decarbonize economies.

Energy intensity (the ratio of total domestic energy primary consumption to gross domestic product or physical output) in the United States, Japan, and other developed countries has increased far slower than economic growth. China, too, has made significant gains, and the growth rate of its energy consumption has been half that of its gross domestic product (GDP) since the early 1980s. While the energy use per unit of GDP has fallen or stabilized since 1970, energy use per capita has increased in most of the developed countries (Sathaye, Chapter 22, this volume). This statistic may be misleading however, because there has also been a change in consumption and production patterns away from materially intensive commodities toward less-intensive services, and production of many of the materially intensive commodities has moved out of the developed countries. This does not necessarily mean that the developed countries are using fewer resources, but rather that the regional distribution of this consumption has changed. Countries belonging to the former Soviet Union (FSU) have also experienced decreases in energy intensity but for very different reasons. Because of sociopolitical changes, economic activity declined substantially in these countries, resulting in a dramatic decline in energy use and associated carbon emissions.

Over the past century and a half, the use of biomass has successively given way to expanded use of coal, oil, and natural gas as the primary fuels to supply energy (Figure 2.3). Over the past two decades, the use and share of nuclear and other forms of renewable energy have increased. The carbon intensity (PgC/gigajoule [GJ]) of each successive fuel, beginning with coal, is lower, and this decline has led to the decarbonization of the global fuel mix over the past century. Critical inventions, such as the steam and internal combustion engines, vacuum tubes, and airplanes, have accompanied and fostered the use of successive fuels and electricity.

The speed of energy consumption and land use changes has increased during the past two centuries. Three spatial and historical variations of carbon-relevant social tendencies need to be considered in developing an understanding of where we are today.

- Historically, industrialized countries have been the main releasers of carbon from combustion of fossil fuels. Although developing countries are expected to increase their share of emissions, the relocation strategies of corporations based in industrialized regions may actually significantly contribute to increased emissions in these developing countries. In 1925, for example, Australia, Japan, the United States, and Western Europe were responsible for about 88 percent of the world's fossil-fuel carbon dioxide emissions (Houghton and Skole 1990). Data from 1950–1995 indicate that 26 countries fell above the average cumulative emissions figure. These countries include almost all developed, or Annex I countries, plus Brazil, China, India, and several other larger developing countries that have seen significant growth in foreign investments (Claussen and McNeilly 1998).

- Until the 20th century, most of the conversion from forest area to cultivated land occurred in the developed countries. During the past few decades, however, most of the deforestation has occurred in tropical forests.

- Urbanization, a key driver of energy and land use, was primarily a feature of industrialized countries until the middle of the 20th century. The largest and demographically most dynamic urban agglomerations, like London, New York, and Tokyo, were in developed countries. As of the beginning of the 21st century, however, most urban agglomerations are situated in developing countries. These developing-country urban agglomerations have different production systems and living standards than the urban centers in developed countries. These differences affect the carbon emissions of various urban areas (Romero Lankao, Chapter 19, this volume).

Although technology is perceived as the answer to decarbonizing economies, other societal factors work as constraints and windows of opportunity for that purpose (e.g., institutional settings and economic dynamics; see Raupach et al., Chapter 6, this volume). In addition, new technological paradigms of production and consumption patterns emerge only over decades. The main energy and production components of the *engineering* epoch, for instance (1850–1940), took at least 20 years to develop. Hence, it will likely take decades to set up alternative energy sources and materials aimed at decarbonizing industrial and agricultural activities. Despite efforts aimed at decoupling
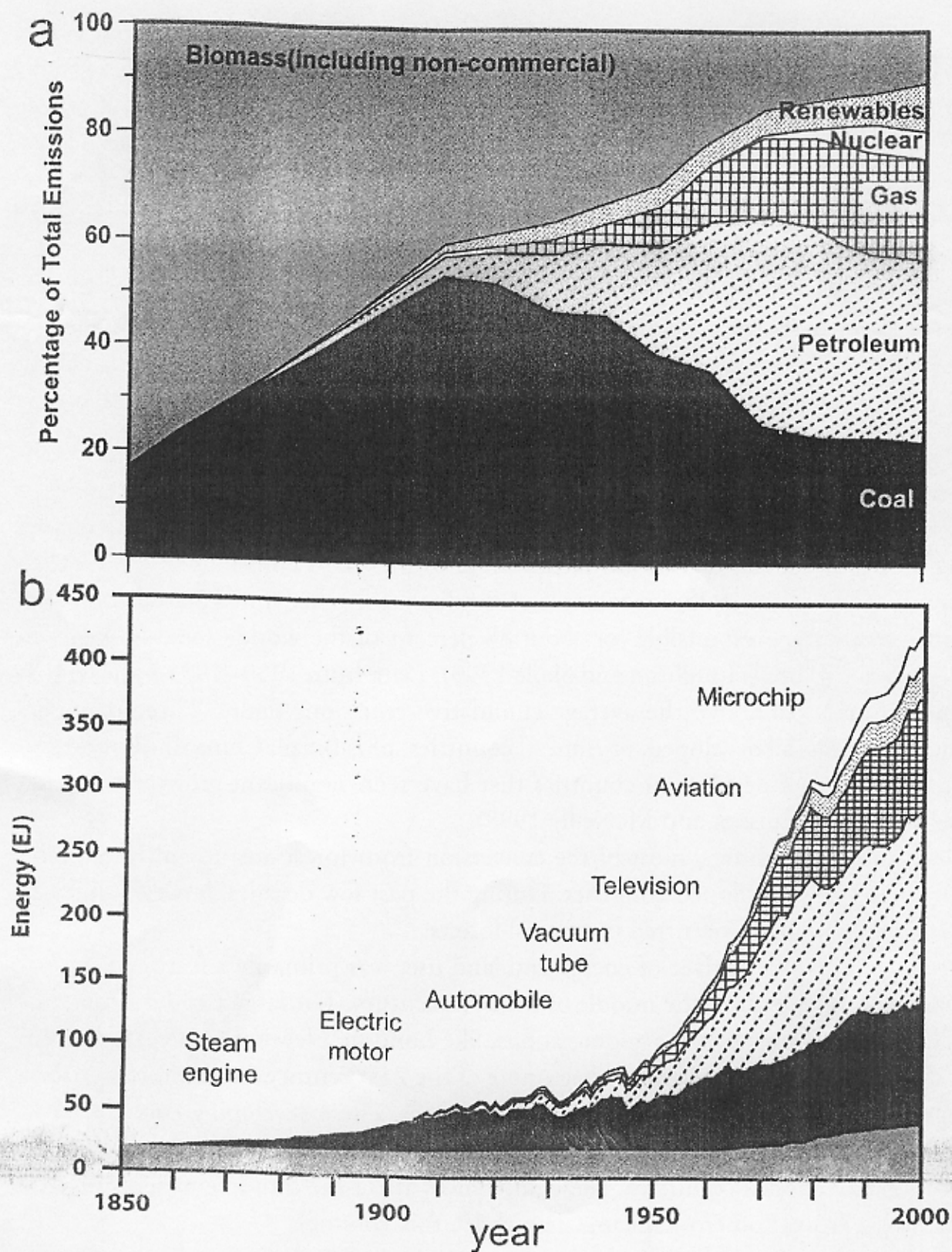
Figure 2.3. Historical composition of the world energy system, in (a) percent contributions and (b) energy contributions in EJ (1 EJ = 10^{18} J = 1.05 Quad = $1.05*10^{15}$ BTU). In 2000 the global consumption of primary energy was about 400 EJ.

economic growth from its carbon impacts, CO_2 emissions from the combustion of fossil fuels and from land use change have been increasing. Many features of the development trends need to be studied and better understood before opportunities for modifications in life styles, technologies, institutions, and other drivers of carbon emissions can be fully addressed.

Geological Processes

The discussion thus far has focused on reservoirs and processes relevant to human timescales. With a residence time of about 300 million years, the huge reserves of carbon stored in the sedimentary rocks are not expected to play a large role in the short-term carbon budget. On timescales greater than 500,000 years, about 80 percent of the CO_2 exchange between the solid earth and the atmosphere is controlled by the carbonate-silicate cycle (Kasting et al. 1988). In this cycle, atmospheric CO_2 is used to weather calcium-silicate rock minerals on land, which are then transported to the ocean via rivers as calcium and bicarbonate ions. In the oceans, plankton and other organisms incorporate the ions into calcium carbonate shells. A portion of the calcium carbonate is deposited onto the ocean floor, and eventually CO_2 is returned to the atmosphere through volcanic and diagenetic processes.

Vast quantities of carbon stored are in ocean sediments as methane hydrates and as calcium carbonate. The methane hydrates are relatively stable but could be released if ocean temperatures increase sufficiently through global warming (Harvey and Huang 1995). The carbonate sediments are likely to be a significant sink for fossil fuel CO_2 on millennial timescales (Archer et al. 1999). As the oceans continue to take up anthropogenic CO_2, the CO_2 will penetrate deeper into the water column, lowering the pH and making the waters more corrosive to calcium carbonate. Dissolution of sedimentary carbonates binds the carbon in a dissolved form that is not easily converted back into atmospheric CO_2. Carbonate dissolution is typically thought to occur in the deep ocean, well removed from the anthropogenic CO_2 taken up in the surface waters. In portions of the North Atlantic and North Pacific Oceans, however, anthropogenic CO_2 may have already penetrated deep enough to influence the dissolution of calcium carbonate in the water column and shallow sediments (Feely et al. 2002).

Although the processes of CO_2 uptake through weathering and CO_2 release from volcanism and diagenesis appear to have a small net effect on the global carbon cycle on millennial timescales, short-term variability in one of these fluxes can affect the carbon cycle on timescales relevant to humans. For example, explosive volcanic eruptions result in the emission of CO_2, dust, ash, and sulfur components. These sulfur particles may rain out and promote acid rain. Atmospheric sulfur particles block sunlight and cool the regional climate. Particles that reach the stratosphere reduce global temperatures for several years. The incidental occurrence of cataclysmic volcanic eruptions or the coincidence of several large eruptions (e.g., the eruptions of 1783 or the 1991

Mount Pinatubo eruption) affect the global carbon cycle directly by their CO_2 emissions and indirectly by their impact on marine and terrestrial primary production (Hamblyn 2001; Sarmiento and Gruber 2002).

Understanding Today's Carbon Cycle

Although our understanding of the contemporary global carbon cycle has increased dramatically over the past few decades, many aspects are still not well understood. Several specific areas where additional studies are needed have been mentioned. There are two general areas of research, however, where our understanding is exceptionally weak and focused research is necessary. One area that needs improvement is understanding regional variability. A second, somewhat related topic is how changes in the carbon cycle may be linked to different modes of climate variability.

Regional Budgets

Although atmospheric CO_2 concentrations are changing on a global scale, the spatial scales of natural processes, as well as the scales of human interventions and the associated societal mechanisms, have a profound regional character. Industrialized countries, for instance, dominate fossil fuel CO_2 emissions by direct release and through trade, whereas developing countries have become the primary CO_2 emitters through land use changes (Romero Lankao, Chapter 19, this volume). Insufficient observations limit our ability to conduct regional-scale assessments globally. However, the potential benefits of such approaches in identifying the biogeochemical and human processes responsible for controlling fluxes makes these studies very important. A regional-scale carbon budget assessment also provides a unique opportunity to verify and bridge independent methods and observations made over a range of spatial scales (e.g., top-down atmospheric inversion estimates versus land-based or ocean-based bottom-up observations).

In the land-based bottom-up method, carbon sinks and sources from various ecosystems (forests, croplands, grasslands, and organic soil wetlands) are aggregated over all regions to provide a large-scale perspective. The land-based approach can therefore provide information about which ecosystems and regions are accumulating carbon and which are losing carbon to the atmosphere. The diversity of land mosaics, the complexity of human activities, and the lateral transports of carbon in different components make it difficult, however, to provide a comprehensive carbon budget.

Ocean-based bottom-up methods can be used to assess the complex and often competitive controls of heat flux, mixing, and biology on air-sea gas exchange, as well as to develop proxies for extrapolating limited observations to larger time and space scales. Monitoring changes in ocean interior properties can also provide valuable information on surface processes and fluxes.

Atmosphere-based approaches, in contrast, give no information about which ecosystems or processes are contributing to a sink or source but produce a more consistent large-scale assessment of the net carbon flux to the atmosphere. No one approach holds the key to understanding regional variability in the carbon cycle. A suite of approaches much be used. In addition to providing different, but complementary, information, a combination of the bottom-up and top-down methods puts independent constraints on the integrated carbon balance of regional budgets.

The use of multiple approaches can also be used to identify key fluxes that the inventory approach may have missed. For example, Janssens et al. (2003) used inventories to estimate that the European continent sink is on the order of 0.11 PgC y^{-1}, compared with a mean atmospheric inversion estimate of 0.29 PgC y^{-1}. The discrepancy between these estimates is attributed to the intercontinental displacement of organic matter via trade, emissions of non-CO_2 gases, and a slight overestimation of fossil-fuel emissions. This example illustrates not only the need for comprehensive assessments of the major vertical and lateral fluxes, but also the power of multiple approaches for highlighting areas where additional studies are needed.

Variability

Observing and quantifying variability in natural processes and the impact of human interventions is of primary importance for three reasons:

- It provides a key diagnostic of how climatic factors and societal dynamics affect exchange fluxes and provides information that is needed to develop and validate comprehensive process-based carbon cycle models that include human dynamics.
- Several terrestrial and oceanic carbon cycle components effectively contain a longer-term "memory"—that is, their present state is partly a result of past natural and human perturbations. For example, enhanced fire frequencies or management strategies during a drier past period will be reflected in the age structure of a present forest, or the carbon content in particular oceanic deepwater masses will reflect climate-driven variations in deepwater formation in the past.
- Climate variability often tends to mask the slow, longer-term signals in the carbon cycle that are of primary interest; for example, many studies have shown that the uptake rates of the terrestrial biosphere or the ocean are highly variable. In this context, climate variability constitutes "noise" from which the signals must be distinguished. Indeed, a significant fraction of the uncertainty in the global budget in Table 2.1 can be traced to incomplete quantification of possible climate perturbations in the pertinent observations.

On a global scale, climate-driven variability can be inferred from atmospheric time series of CO_2 and associated variables, such as $^{13}C/^{12}C$ and O_2/N_2 ratios. Continental or ocean basin–scale variations can be detected on timescales of up to several years by

means of top-down inversions of atmospheric CO_2 concentration measurements (Heimann et al., Chapter 8, this volume). It is difficult, however, to separate the terrestrial signals from the ocean signals using these large-scale approaches (Greenblatt and Sarmiento, Chapter 13, this volume). On land and in the oceans, local-scale direct observations of variability exist from in situ flux measurements at a few time-series stations during the past few years. On a regional scale, however, an observational gap exists. The present atmospheric network is not dense enough to resolve carbon sources and sinks on regional scales by atmospheric inversion, while the upscaling of measured in situ oceanic and terrestrial carbon fluxes is extremely difficult. On land, the large heterogeneity of terrestrial ecosystems and complex atmospheric transport patterns resulting from topography make it difficult to scale up local measurements. In the oceans, the complex interplay between physical and biological controls on sea surface pCO_2 and the dynamics of air-sea exchange complicate the extrapolation of the in situ observations.

Interannual variability in climate leads to large changes in atmospheric temperature and rainfall patterns, as well as changes in ocean surface temperatures and circulation. All of these changes can have dramatic effects on biological productivity and complex effects on CO_2 exchanges with the atmosphere on land and in the oceans. For example, during El Niño events, the warming of ocean surface waters and the reduction in biological productivity in the Equatorial Pacific Ocean should lead to enhanced outgassing of CO_2. Since the upwelling of carbon-rich deepwaters, which release CO_2 to the atmosphere, is reduced during El Niños, however, the net effect is a significant reduction in the outgassing of CO_2 in this region. On the other hand, warmer temperatures and anomalous rainfall patterns during El Niños can lead to increased terrestrial biosphere respiration, forest fires, and droughts. The timing of these effects and the teleconnections between them have a direct impact on atmospheric CO_2 concentrations that is still not completely understood.

Overall, the general consensus is that the interannual variability in air-sea CO_2 fluxes is smaller than that of terrestrial CO_2 fluxes, but the exact amplitude and spatial distributions remain uncertain (Greenblatt and Sarmiento, Chapter 13, this volume). To better determine these signals, an expanded network of time-series CO_2 measurements must be maintained for the atmosphere, oceans, and land systems. These measurements, together with intensive process studies, a better use of satellite data, atmospheric observations, and rigorously validated models, will help us better understand the current global carbon cycle and how it is evolving over time.

Conclusions

The current global carbon cycle is in a state of transition. Human activities over the past few centuries have had a profound impact on many aspects of the system. As we begin to assess ways to monitor and potentially manage the global carbon cycle, it is imperative that we better understand how the system, including human dynamics and the bio-

geochemical processes controlling CO_2 and other carbon gases in the atmosphere, has operated in the past and how it is operating today. How will the land-ocean-atmosphere processes respond to human activities, and what are the societal dynamics that will determine how humans will respond to changes in the land, ocean, and atmosphere systems? We have made tremendous progress over the past few decades at reducing the uncertainties in Table 2.1 and Colorplate 1, but until we can confidently explain and model the contemporary carbon cycle, our ability to predict future changes in atmospheric CO_2 concentrations will be limited.

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