Chapter 13

Climate and Air Quality

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*This appears in Appendix A at the end of this volume.

Main Messages

Ecosystems, both natural and managed, exert a strong influence on climate and air quality. Ecosystems are both sources and sinks of greenhouse gases, aerosol precursors, and pollutants. Their physical properties affect heat and water fluxes, influencing temperature and precipitation—altering, for example, the reflection of solar radiation (albedo) and the flow of water through plants to the atmosphere (evapotranspiration), where it becomes available for rainfall. Thus ecosystems provide the following atmospheric "services":

- warming (for example, sources of greenhouse gases and reduction of albedo by boreal forests compared with bare soil and snow);
- cooling (for example, sinks of greenhouse gases, sources of aerosols that reflect solar radiation, and evapotranspiration);
- water recycling and regional rainfall patterns (for example, evapotranspiration and sources of cloud condensation nuclei);
- atmospheric cleansing (for example, sinks for pollutants such as tropospheric ozone, ammonia, NO_x, sulfur dioxide, and methane);
- pollution sources (for example, particulates from biomass burning, NO_x, carbon monoxide, and precursors of tropospheric ozone); and
- nutrient redistribution (for example, source of nitrogen deposited elsewhere and reduction of erosion and nutrient-rich airborne dust compared with bare soil).

Changes in ecosystems have made a large contribution to the changes in radiative forcing (the cause of global warming) between 1750 and the present. The main drivers are deforestation, fertilizer use, and agricultural practices. Ecosystem changes account for about 10–30% of the radiative forcing of carbon dioxide from 1750 to present and a large proportion of the radiative forcing due to methane and nitrous oxide. Ecosystems are currently a net sink for carbon dioxide and tropospheric ozone, while they remain a net source of methane and nitrous oxide. Management of ecosystems has the potential to significantly modify concentrations of a number of greenhouse gases, although this potential is small in comparison to IPCC scenarios of fossil fuel emissions over the next century (*high certainty*). Ecosystems influence the main anthropogenic greenhouse gases in several ways:

- **Carbon dioxide**—Preindustrial concentration, 280 parts per million; concentration in 2000, 370 ppm. About 40% of the emissions over the last two centuries and about 20% of the CO_2 emissions during the 1990s originated from changes in land use and land management, primarily deforestation. Terrestrial ecosystems have been a sink for about a third of cumulative historical emissions and a third of the 1990s total (energy plus land use) emissions. The sink may be explained partially by afforestation/reforestation/forest management in North America, Europe, China, and other regions, and partially by the fertilizing effects of nitrogen deposition and increasing atmospheric CO_2. Ecosystems were on average a net source of CO_2 during the nineteenth and early twentieth century and became a net sink sometime around the middle of the last century (*high certainty*).
- **Methane**—Preindustrial concentration, 700 parts per billion; concentration in late 1990s, 1750 ppb. Natural processes in wetland ecosystems account for 25–30% of current CH_4 emissions, and about 30% of emissions are due to agriculture (ruminant animals and rice paddies).
- **Nitrous oxide**—Preindustrial concentration, 270 ppb; concentration in late 1990s, 314 ppb. Ecosystem sources account for about 90% of current N_2O emissions, with 35% of emissions from agricultural systems, primarily driven by fertilizer use.
- **Tropospheric ozone**—Preindustrial, 25 Dobson Units; late 1990s, 34 DU. Several gases emitted by ecosystems, primarily due to biomass burning, act as precursors for tropospheric ozone. Dry deposition in ecosystems accounts for about half the tropospheric ozone sink. The net global effect of ecosystems is a sink for tropospheric ozone.

Land cover changes between 1750 and the present have increased the reflectivity to solar radiation (albedo) of the land surface (*medium certainty*), partially offsetting the warming effect of associated CO_2 emissions. Deforestation and desertification in the tropics and sub-tropics leads to a reduction in regional rainfall (*high certainty*). The biophysical effects of ecosystem changes on climate depend on geographical location and season. With *high certainty*:

- Deforestation in seasonally snow-covered regions leads to regional cooling during the snow season due to an increase in surface albedo and leads to warming during summer due to reduction in evapotranspiration.
- Large-scale tropical deforestation (hundreds of kilometers) reduces regional rainfall, primarily due to decreased evapotranspiration.
- Desertification in tropical and sub-tropical drylands leads to decrease in regional rainfall due to reduced evapotranspiration and increased surface albedo.

Biophysical effects such as this need to be accounted for in the assessment of options for mitigating climate change. For example, the warming effect of reforestation in seasonally snow-covered regions due to albedo decrease is likely to exceed the cooling effect of additional carbon storage in biomass.

Ecosystems are currently a net sink for several atmospheric pollutants, including tropospheric ozone (which causes respiratory problems and plant damage), CO_2 (which leads to ocean acidification, with negative effects on calcifying organisms such as corals and coccoliths), and ammonia (which contributes to health problems, eutrophication of lakes, and acidification of N-saturated ecosystems). Fertilizer use has led to increased ecosystem emissions of N gases (which contribute to acid rain and eutrophication of lakes). The net effect of ecosystems on acid rain and stratospheric ozone depletion are small compared with industrial emissions.

Vegetation burning, both natural and human-induced, is a major cause of air pollution. Particulates, tropospheric ozone, and carbon monoxide are toxic to humans at levels reached as a result of biomass burning. In the 1990s, biomass burning was responsible for about a quarter of global carbon monoxide emissions, just under half of particulate aerosol emissions, and a large but poorly quantified fraction of tropospheric ozone precursor emissions. Smoke plumes cause changes in plant productivity (generally decreases), changes in rainfall (generally decreases), and economic losses due reduced visibility (affecting transport, for example).

The self-cleansing ability of the atmosphere is fundamental to the removal of many pollutants and is affected by ecosystem sources and sinks of various gases. Removal of pollutants involves chemical reactions with the hydroxyl radical. OH concentration and hence atmospheric cleansing capacity has declined since preindustrial times but likely not by more than 10%. The net contribution of ecosystem changes to this decline is currently unknown. The reactions are complex, but generally emissions of NO_x and hydrocarbons from biomass burning increase tropospheric ozone and OH concentrations, and emissions of CH_4 and carbon monoxide from wetlands, agricultural practices, and biomass burning decrease OH concentration.

The most important ecosystem drivers of change in climate and air quality in the past two centuries have been deforestation (net CO_2 emissions, net increase in surface albedo, and rainfall reduction), agricultural practices (increasing emissions of CH_4, N_2O, and other N gases), and biomass burning (emissions of toxic pollutants). Wetland draining has decreased CH_4 emissions but increased emissions of CO_2 and N_2O. The net short-term (20–100 year) effect on radiative forcing is cooling (*medium certainty*), while the long-term effect is probably warming (*low certainty*). Management of drylands to increase vegetation cover reduces soil carbon loss, reduces dust emissions, and increases water recycling. Loss of species richness has probably not had significant impacts on climate and air quality in the recent past, but shifts in functional types—such as trees versus grasses, deciduous versus evergreen trees, or calcifying versus non-calcifying plankton—could alter the biological storage of carbon and trace gas emissions in the future.

Ecosystem interactions with the atmosphere are highly nonlinear, with many feedbacks and thresholds that, if passed, may lead to abrupt changes in climate and land cover. Human-induced land cover changes may become irreversible due to ecosystem-climate feedbacks. For example, in the Sahara-Sahel region, two alternative land cover types are theoretically sustainable: savanna and desert. If a threshold in loss of vegetation cover contributing to rainfall reduction is crossed, the desert state becomes self-sustaining (*low certainty*). The complexity and incomplete understanding of the feedbacks make it hard to predict thresholds and their future changes.

13.1 Introduction

> *Living matter builds bodies of organisms out of atmospheric gases such as oxygen, carbon dioxide and water, together with compounds of nitrogen and sulphur, converting these gases into liquid and solid combustibles that collect the cosmic energy of the sun. After death, it restores these elements to the atmosphere by means of life's processes . . . Such a close correspondence between terrestrial gases and life strongly suggest that the breathing of organisms has primary importance in the gaseous system of the biosphere; in other words, it must be a planetary phenomenon.*
>
> (Vernadsky 1926)

The composition of the atmosphere and the climate we experience are products of the co-evolution of the biosphere, atmosphere, and geosphere over billions of years (Vernadsky 1926; Zavarzin 2001). The climate and the concentration of various gases in the atmosphere are determined by the flow of energy (radiation, heat) and materials (such as water, carbon, nitrogen, trace gases, aerosols) between the atmosphere, ocean, soils, and vegetation. These interlinking components are referred to collectively as the Earth System to stress their inter-dependence. (See Box 13.1.) Lovelock and Margulis (1974) proposed the Gaia Hypothesis: that biospheric feedbacks regulate the climate within a range suitable for life. Although not universally accepted, Gaia remains an inspirational idea in Earth System science (Lenton 1998; Watson 1999; Kirchner 2003).

Ecosystems alter atmospheric chemistry, providing both sources and sinks for many atmospheric constituents that affect air quality or that affect climate by changing radiative forcing. In this chapter we refer to these as "biogeochemical effects." Ecosystems further influence climate through the effects of their physical properties on water fluxes (such as rainfall) and energy balance (such as temperature). (See Figure 13.1.) We refer to these as "biophysical effects."

The ability of ecosystems to modify climate and air quality and thereby provide a service to humans occurs both through natural processes and as a result of ecosystem management. For example, the conversion of carbon dioxide and water to oxygen by ecosystems billions of years ago could be considered the fundamental ecosystem service, enabling evolution and maintenance of a breathable atmosphere. Not all impacts of ecosystems on the atmosphere and climate are beneficial to human well-being, and the effects often depend on the location and magnitude of the impact. A change in magnitude can change the sign—for example, a small temperature increase may help some people at some locations by, say, extending the crop-growing season or potential area, but a large temperature increase is detrimental to the majority of people in the majority of locations through, for instance, damage to crops and human health (IPCC 2001d).

This chapter assesses all major effects of ecosystems on climate and air quality, be it "good" or "bad" for human well-being. The impacts of climate and air quality on ecosystems and human well-being, in contrast, are dealt with in detail by other assessments (e.g., IPCC 2001b, 2001d; WHO 2002; WMO 2003; Brasseur et al. 2003a; Emberson et al. 2003) and are not the focus of the MA other than as drivers of ecosystem change (which are summarized in Chapter 3 and in MA *Scenarios,* Chapter 7, in several sections later in this chapter, and in relevant sections of other chapters). For more detailed reviews of the science behind global climate change see IPCC (2001a); see also Kabat et al. (2004) and Kedziora and Olejnik (2002) on biophysical mechanisms and impacts and Brasseur et al. (2003a) on atmospheric chemistry.

During the Quaternary period (approximately the past 2.5 million years), the Earth System has shown a persistent pattern of glacial-interglacial cycles during which the climate and atmospheric composition varied between fairly consistent bounds, as shown by ice core measurements (Petit et al. 1999, EPICA community members 2004). These quasi-periodic cycles are triggered primarily by variations in Earth's orbit. The associated changes in climate and in carbon dioxide, methane, and other atmospheric constituents are controlled by mechanisms involving both terrestrial and ocean ecosystems (IPCC 2001a; Prentice and Raynaud 2001; Steffen et al. 2004; Joos and Prentice 2004). However, the balance of these mechanisms is not well understood, and this implies uncertainties in predicting future changes, especially on time scales of centuries or longer.

Burning fossil fuels, changes in land cover, increasing fertilizer use, and industrial emissions over the past two centuries have propelled the Earth System outside the boundaries of the natural system dynamics of the Quaternary period (*high certainty*). The current concentration of carbon dioxide and methane are unprecedented in the last 420,000 years and possibly in the last 20 million years, and the rate of increase is unprecedented in at least the last 20,000 years (IPCC 2001a, 2001d; see also MA *Scenarios,* Chapter 7). The increase in temperature in the twentieth century was the largest of any 100 years in the last 1,000 years (IPCC 2001a, 2001d). The Intergovernmental Panel on Climate Change concluded that "there is newer and stronger evidence that most of the warming observed over the last 50 years is attributable to human activities" (IPCC 2001a, 2001d).

Human intervention in global biogeochemical cycles has triggered a chain of biogeochemical and biophysical mechanisms that will continue to affect both atmospheric chemistry and climate on time scales from years to millennia (*high certainty*). Even if emissions ceased today, past emissions would continue to have an impact in the future related to the lifetime of the emitted gas in the

BOX 13.1

The Earth System, Thresholds, and Feedbacks

The "Earth System" has several interacting components: the atmosphere, ocean, terrestrial and marine biosphere, cryosphere (ice, including permafrost), the pedosphere (soils), and humans. These components are tightly linked with each other. Ecosystems are an integral part of the Earth System; they provide different services to the climate system via numerous physical and chemical mechanisms that control fluxes of energy (radiation, heat), water, and atmospheric constituents.

The Earth System is highly nonlinear: climate, air quality, and ecosystem distribution across the planet may change quite abruptly in response to smooth changes in external forcing (as occurred, for example, during the last deglaciation about 15,000 years ago). Current theories support the possibility of multiple stable states (regimes of a particular balance of components that are resistant to change) and abrupt transitions between these different states or regimes (as with desert and vegetation in the Sahara/Sahel region). These transitions are reinforced by positive (amplifying) feedback loops between components of the Earth System, whereby a small change in one component can cause changes in other components that continue to push the system away from its previous state and toward a new one. Conversely, negative (stabilizing) feedbacks can maintain stable states by preventing the system moving beyond certain thresholds.

Numerous examples of feedbacks between ecosystems, climate, and atmospheric constituents are mentioned throughout this chapter.

- **Climate–greenhouse gases, positive feedback.** Warming enhances emissions of CH_4, N_2O, and tropospheric ozone precursors (NO_x and VOCs) (*very certain*). Warming reduces inorganic ocean uptake of CO_2, increases soil emissions of CO_2, and has been predicted to reduce carbon storage in terrestrial and ocean ecosystems; the net result is an increase in atmospheric CO_2 (*high certainty*). Increasing concentration of greenhouse gases causes further warming.
- **CO_2 fertilization–CO_2 uptake, negative feedback.** Increased atmospheric concentration of CO_2 has a fertilizing effect on plants, increasing uptake of CO_2 and reducing atmospheric concentrations (*high certainty*).
- **Taiga-tundra albedo-temperature, positive feedback.** Afforestation of snow-covered regions due to northward movement of forest boundaries in a warmer world, or due to tree planting, reduces albedo, leading to further warming, less snow, and further reduced albedo (*high certainty*).
- **Tropical rainforest–precipitation, positive feedback.** Large-scale reduction in tropical rainforest cover reduces regional precipitation, potentially causing further forest loss and precipitation reduction (*high certainty*).
- **Sahara-Sahel vegetation–precipitation, positive feedback.** Decreased vegetation cover increases albedo, reduces soil-atmosphere water recycling, and reduces monsoon circulation, decreasing precipitation, all of which further suppress vegetation cover (*high certainty*).
- **DMS-cloudiness, negative feedback.** Emissions of DMS by ocean ecosystems (and of VOCs by terrestrial ecosystems) increase cloud condensation nuclei, cooling Earth, reducing photosynthesis and emissions of DMS (and VOCs), and increasing thermal stability, reducing cloud formation (*medium certainty*).
- **Tropical forest–tropospheric ozone, positive feedback.** High levels of tropospheric ozone have a deleterious effect on vegetation, compromising further uptake of tropospheric ozone (*medium certainty*).
- **Pollution–reduction in cleansing capacity, positive feedback.** For example, CH_4 in the atmosphere reduces OH concentration and atmospheric cleansing capacity, increasing lifetime and atmospheric concentrations of CH_4 (*high certainty*).

atmosphere, atmospheric chemistry, and inertia in different parts of the Earth System (such as the uptake and mixing of carbon dioxide in the ocean and the response of sea level rise to temperature and ice melting) (IPCC 2001d Figure 5.2).

A summary of the main ecosystem effects on climate and air quality, the drivers, and the impacts on human well-being that are discussed in this chapter is presented in Figure 13.2. Changes in climate or air quality are often simultaneously affected by several atmospheric constituents. Likewise, a particular atmospheric constituent can affect both climate and air quality. Furthermore, ecosystem drivers (such as deforestation, biomass burning, and agricultural practices) often simultaneously affect biogeochemical and biophysical properties, and their effects can work in the same or opposite directions. Thus it is often not possible to quantify cause and effect. Each atmospheric constituent and vegetation property considered in this chapter is summarized in Table 13.1, along with its magnitude and distribution, main drivers, and impacts.

13.2 Biogeochemical Effects of Ecosystems on Climate: Greenhouse Gases and Aerosols

Many atmospheric constituents determine the radiative forcing of Earth's climate (IPCC 2001a, 2001d). (Radiative forcing is the change in net vertical irradiance (radiation or energy) of the tropopause (upper troposphere), with an increase in radiative forcing implying an increase in global temperature. Global warming potential is an index, relative to CO_2, describing the radiative properties of greenhouse gases based on their effectiveness at absorbing long-wave radiation, and the time they remain in the atmosphere. For more detailed explanations of these terms, see IPCC (2001a, 2001d).)

Key atmospheric compounds that have an ecosystem source or sink include:

- greenhouse gases that absorb long-wave radiation from Earth's surface, leading to warming—carbon dioxide, methane, tropospheric ozone (formed from precursors NO_x, methane, and volatile organic compounds), and nitrous oxide; and
- aerosols, of which some types (such as sulfate aerosols) reflect solar radiation leading to cooling, while others (such as black carbon) trap radiation leading to warming.

Ecosystems have played a significant role in past radiative forcing (see Figure 13.3A) and in current sources and sinks of greenhouse gases and aerosols (see Figure 13.3B). The net biochemical contribution of ecosystems to historical radiative forcing has been to increase global warming, accounting for about 10–30% of the radiative forcing of CO_2 from 1750 to the present (Brovkin et al. 2004) and a large proportion of the warming due to CH_4 and N_2O, while reducing tropospheric ozone forcing. Ecosystems are currently a net sink for CO_2 and tropospheric ozone, while they remain a net source of CH_4 and N_2O.

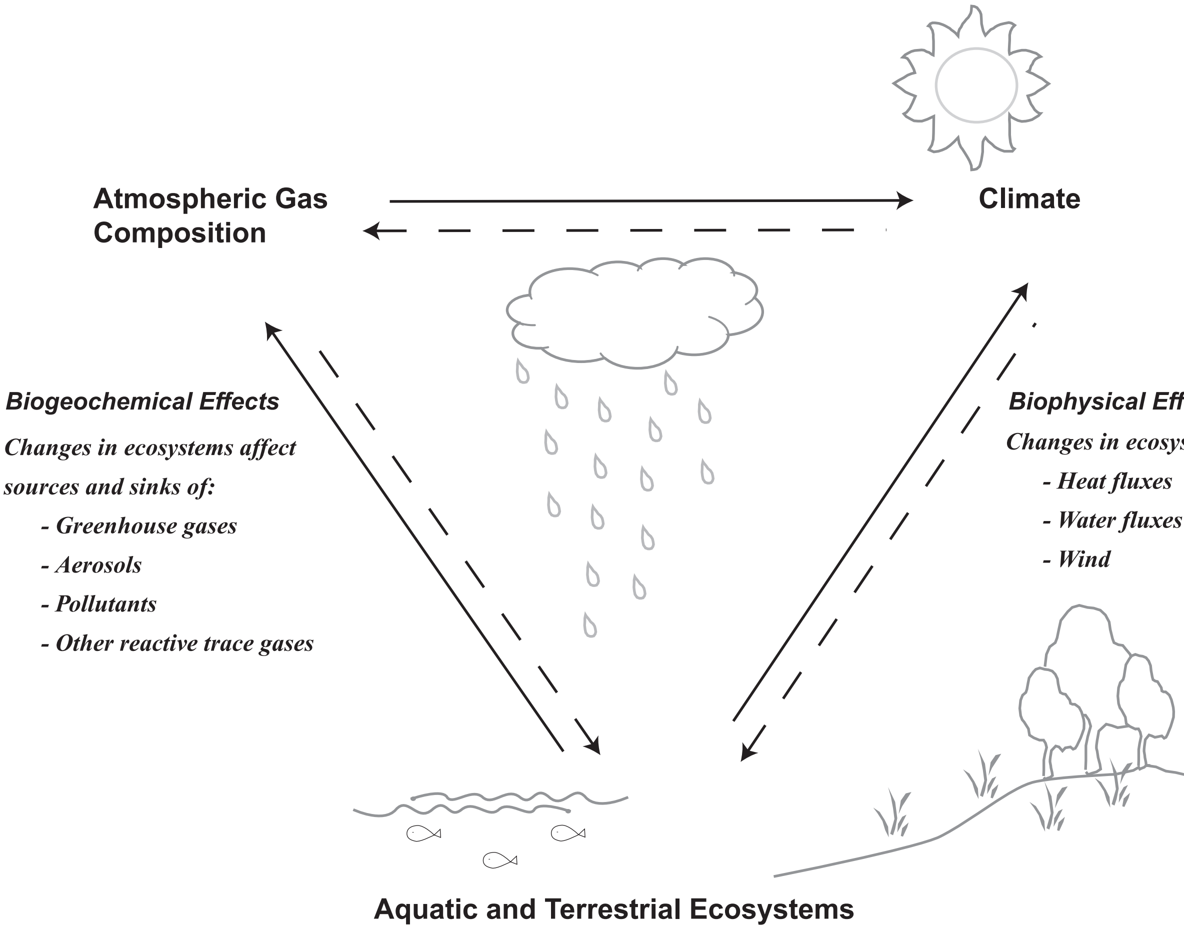

Figure 13.1. Ecosystem Effects on the Atmosphere and Climate. Ecosystems, the concentration of many atmospheric gases/species, and the climate all strongly interact. However, it is the effect of ecosystems on atmospheric air quality (as sources and sinks of pollutants) and on the climate (both directly due to biophysical properties of vegetation, and indirectly as a source and sink of greenhouse gases and aerosols) that is the focus of this chapter, as indicated by the solid arrows.

13.2.1 Carbon Dioxide

Increasing carbon dioxide concentration has had more impact on historical radiative forcing than any other greenhouse gas. In addition, CO_2 has a fertilizing effect on most land plants, while rapid injection of CO_2 into the atmosphere causes acidification of the global ocean, with negative implications for calcifying organisms. Ecosystems are both a source and sink for CO_2. Management of ecosystems for carbon storage is currently regarded as an important ecosystem service by policy-makers, therefore more information is included for this gas than for others. A summary of pertinent details of the carbon cycle is presented in this section; for more details, see Prentice et al. (2001), Kondratyev et al. (2003), and Field and Raupach (2004).

Carbon, the basic building block of all animal and plant cells, is converted to carbohydrates by the process of plant photosynthesis. Terrestrial plants capture CO_2 from the atmosphere; marine plants (phytoplankton) take up carbon from seawater, which exchanges CO_2 with the atmosphere. Plant, soil, and animal respiration returns carbon to the atmosphere, as does burning biomass. Burning fossilized biomass (fossil fuels) also returns carbon, captured by plants in Earth's geological history, to the atmosphere.

CO_2 is continuously exchanged between the atmosphere and the ocean; it dissolves in surface waters and is then transported into the deep ocean (the "solubility pump"). It takes roughly one year for CO_2 concentration in surface waters to equilibrate with the atmosphere, but subsequent mixing between surface waters and deep waters, which drives the ongoing uptake of increased atmospheric CO_2, takes decades to centuries. Some of the dissolved carbon that is taken up by marine plants sinks in the form of dead organisms, particles, and dissolved organic carbon (the "biological pump"). A small amount remains in ocean sediments; the rest is respired at depth and eventually recirculated to the surface. The biological pump acts as a net sink for CO_2 by increasing the concentration of CO_2 at depth, where it is isolated from the atmosphere for decades to centuries, causing the concentration of CO_2 in the atmosphere to be about 200 parts per million lower than it would be in the absence of marine life (Sarmiento and Toggweiler 1984; Maier-Reimer et al 1996).

It has been widely assumed that ocean ecosystems are at steady state at present, although there is now much evidence of large-scale trends and variations (Beaugrand et al. 2002; Chavez et al. 2003; Richardson and Schoeman 2004). Changes in marine ecosystems, such as increased phytoplankton growth rate due to the fertilizing effect of iron in dust (see section 13.4.4.3) and shifts in species composition due to ocean acidification (see section 13.4.2.1) or for other reasons, have the potential to alter the oceanic carbon sink. The net impact of changes in ocean biology on global CO_2 fluxes is unknown.

CO_2 concentration varied within consistent bounds of 180 to 300 ppm during glacial-interglacial cycles. Prior to the industrial

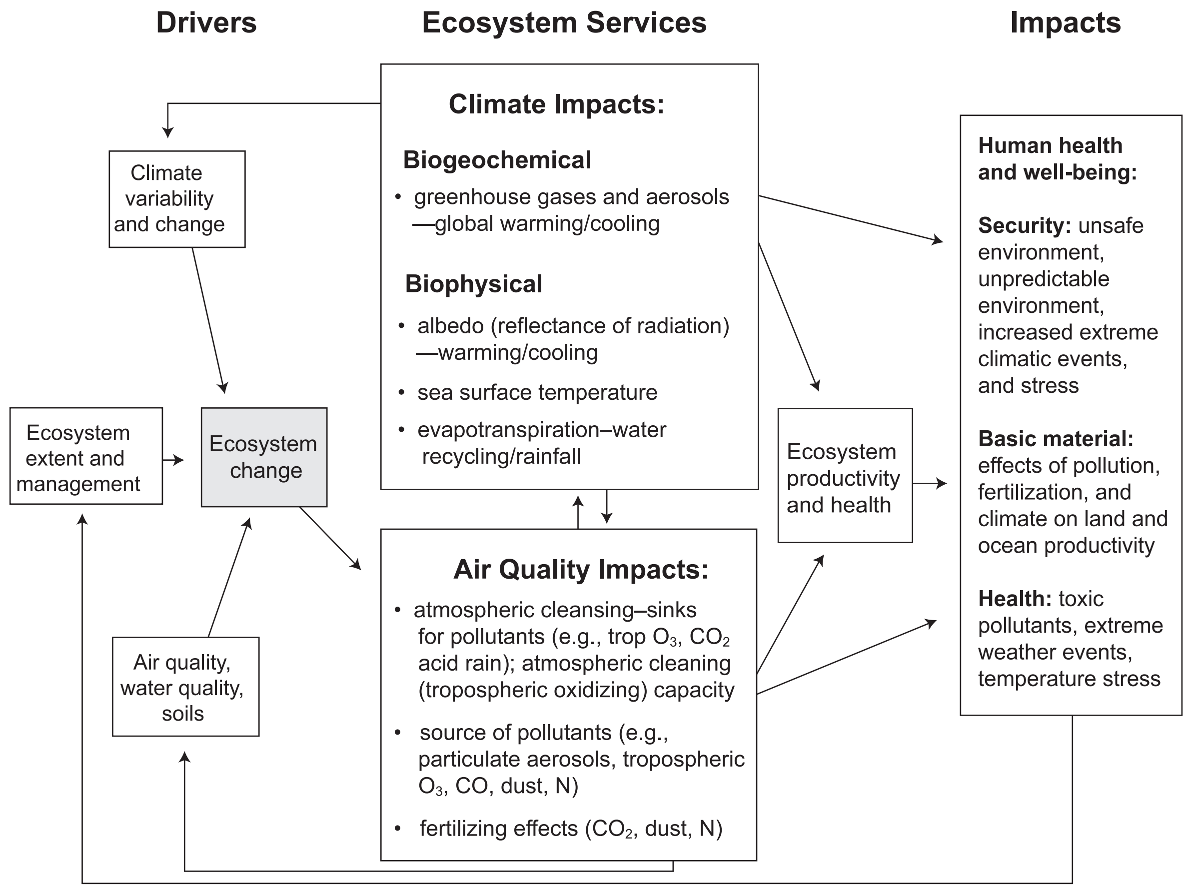

Figure 13.2. Ecosystem Effects on Climate and Air Quality: Services, Drivers, and Impacts on Human Well-being. This figure summarizes the services, drivers, and impacts discussed in this chapter. Arrows represent a direct impact on human health and well-being of the two ecosystems services of this chapter (climate regulation and regulation of atmospheric composition). Note: not all arrows are shown, for simplicity (e.g., direct impacts of climate change on atmospheric composition through changes in atmospheric chemistry).

revolution (that is, before 1750), CO_2 concentration was about 280 ppm, and since then it has risen rapidly, reaching 370 ppm in 2000 (MA *Scenarios,* Chapter 7). It has been estimated that about 40% of CO_2 emissions over the last two centuries came from land use change (primarily deforestation), while 60% came from fossil fuel burning (DeFries et al. 1999). (See Figure 13.4.) About 40% of total CO_2 emissions have remained in the atmosphere.

Oceans are estimated to have taken up approximately a quarter, an amount that can be fully accounted for by the solubility pump. This means that terrestrial ecosystems took up about a third of all emissions (Prentice et al. 2001, House et al. 2002) through a combination of ecosystem processes whose relative importance is still not firmly established but that probably include growth of replacement vegetation on cleared land (e.g., Dixon et al. 1994; Houghton et al. 1998; McGuire et al. 2001; Goodale et al. 2002); agricultural and forest management (e.g., Spiecker et al. 1996; Houghton et al. 1999); other land management practices, such as fire suppression leading to woody encroachment (e.g., Houghton et al. 1999); and fertilizing effects of elevated CO_2 and nitrogen deposition (e.g., Lloyd 1999; Holland et al. 1997).

Analyses of historical atmospheric CO_2 concentrations preserved in ice cores and more recent atmospheric measurements suggest that the land was a net source of CO_2 during the nineteenth and early twentieth centuries (that is, emissions exceeded uptake), and that land changed to a net sink around the 1940s (Bruno and Joos 1997; Joos et al. 1999). Model analyses with reconstructed land use and environmental data indicate a later change from source to sink (1960s to 1970s), due to decreasing deforestation in the tropics, forest regrowth in North America and Asia (Houghton 2003; McGuire et al. 2001; Ramankutty and Foley 1999; Brovkin et al. 2004), and increased uptake of CO_2 by extant ecosystems (McGuire et al. 2001). However, there are uncertainties in modeling the magnitude of changes in the terrestrial carbon budget resulting from several sources: differences in land cover data sets; the lack of systematic global inventory data for vegetation and soil carbon density; and poor quantification of N and CO_2 fertilization effects and climate impacts on ecosystems (Prentice et al. 2001; House et al. 2003).

Measured fluxes of CO_2 during the 1980s and 1990s are shown in Table 13.2. During this period, ecosystems were a net CO_2 sink. Model results indicate that, during the 1990s, terrestrial ecosystems accounted for about 20% of the total emissions (land plus fossil fuels) but were a sink for about a third of the total emissions. Figure 13.5 (in Appendix A) shows a reconstruction of the spatial distribution of ocean and terrestrial net fluxes in the latter half of the 1990s, based on atmospheric measurements (Rödenbeck et al. 2003). These net fluxes are not broken down into source/sink terms or their underlying drivers. Information on regional fluxes due to different drivers assessed by different methods is reviewed in House et al. (2003). Generally, areas of deforestation and forest degradation in the tropics are losing carbon, while areas of afforestation and forest growth in North America and Europe are gaining carbon (House et al. 2003).

Table 13.1. Summary Table of Atmospheric Constituents and Biophysical Factors Affected by Ecosystems: Trends, Drivers, and Impacts on Ecosystem and Human Well-being[a]

Atmospheric Constituents	Sources	Sinks	Trends	Ecosystem Drivers	Impacts
CO_2	*Ecosystem:* land use change (mainly deforestation) ≈1.6 (0.5 to 3.0) PgC/yr (net land flux uptake of 1.2 ± 0.9 PgC/yr) *Other:* fossil fuel 6.3 ± 0.4 PgC/yr *[note: numbers updated since IPCC 2001a, b; see Table 13.2]*	*Ecosystem:* terrestrial uptake (photosynthesis) ≈ 2.8 (0.9 to 5.0) PgC/yr *Other:* ocean uptake (dissolution and mixing) 1.9 ± 0.7 PgC/yr	*Atmospheric concentration:* Increased from preindustrial 280 ppm to 370 ppm (2000). Increased by 3.2 ± 0.1 PgC/yr during 1990s. Average annual rate of increase rising. Projected to continue rapid increase due to fossil fuel burning and long atmospheric lifetime: ≈ 250 years, but a small amount persists for much longer. *Ecosystem:* Terrestrial source until around middle of last century, then increasing sink. Sink likely to decline due to limited management opportunities, saturation of CO_2 fertilization effect, and climate impacts. Ocean ecosystems show evidence of large-scale trends and variations, but the net impact of these changes on CO_2 fluxes is unknown. Non-biological uptake by the ocean will continue, but the rate will decline with increasing CO_2 concentration and warmer climate.	– climate change – land use and land management: deforestation, afforestation, reforestation, forest management, agricultural management – biomass burning – N fertilization – Fe fertilization (dust) – CO_2 fertilization effects	– climate: positive radiative forcing (heating) – ocean acidification: reduced growth of oceanic calcifying organisms including corals, potential negative impacts on fish production – "fertilizing" effect on plants
CH_4	*Ecosystem:* peatlands/wetlands 92–237 TgC/yr ruminants 80–115 TgC/yr rice 25–100 TgC/yr termites 20 TgC/yr oceans 10–15 TgC/yr biomass burning 23–55 TgC/yr *Other:* energy 75–110 TgC/yr, landfills 35–73 TgC/yr waste treatment 14–25 TgC/yr methane hydrates 5–10 TgC/yr	*Ecosystem:* soil uptake 30 TgC/yr *Other:* tropospheric OH reactions 506 TgC/yr stratospheric loss 40 TgC/yr	*Atmospheric concentration:* Increased from preindustrial 700 ppb to 1,745 ppb in 1998. Increased 7.0 ppb/yr during 1990s. Atmospheric lifetime: 8.4 years. Growth rate peaked in 1981 at 17 ppb/yr but is highly variable from year to year. *Ecosystem:* Increasing terrestrial source; sink relatively small contribution to the overall trend; growth rate slowed 1990–96 partly due to decreased northern wetland emissions rates from anomalously low surface temperatures and reduction in OH from strat. O_3 depletion; removal rates increased 1990–2000 by +0.5%/yr.	– climate change – land use and land management: agricultural practices, wetland draining – biomass burning – flooding	– climate: positive radiative forcing (greenhouse gas, heating) – tropospheric ozone formation – stratospheric ozone formation – tropospheric oxidizing capacity reduction
CO	*Ecosystem:* mostly tropical sources vegetation 150 TgC/yr oceans 50 TgC/yr biomass burning 700 TgC/yr *Other:* oxidation of: – CH_4 800 TgC/yr – VOCs 430 TgC/yr fossil/domestic fuel 650 TgC/yr	*Ecosystem:* dry (surface) deposition 190 TgC/yr (Hauglustaine et al. 1998) *Other:* OH reaction 1920 TgC/yr (Hauglustaine et al. 1998)	*Atmospheric concentration:* Preindustrial concentration unknown, 1998 concentration 80 ppb. Increasing 6 ppb/yr during 1990s. Atmospheric Lifetime: 0.08–0.25 years. Slowly increasing till late 1980s, then decreased possibly due to catalytic converters decreasing automobile emissions, increased again late 1990s. Increase may be mostly in Northern Hemisphere (Haan et al. 1996), which already contains twice as much CO as Southern Hemisphere.	– biomass burning – ecosystem uptake – land use and land management: vegetation cover	– human health: hypoxia, neurological problems, cardiovascular disease – tropospheric ozone precursor – tropospheric oxidizing capacity reduction, removal of OH – indirect climate impacts: reacts with other greenhouse gases
N_2O	*Ecosystem:* tropical soils 4.0 TgN/yr temperate soils 2.0 TgN/yr agricultural soils 4.2 TgN/yr ocean 3 TgN/yr cattle/feedlots 2.1 TgN/yr biomass burning 0.5 TgN/yr *Other:* industrial 1.3 TgN/yr atmosphere (NH_3 oxid.) 0.6 TgN/yr	*Ecosystem:* N_2O uptake by soils and conversion to N_2 (relatively small) *Other:* stratospheric reactions that deplete ozone 12.3 TgN/yr	*Atmospheric concentration:* Increased from preindustrial 270 ppb to 314 ppb in 1998. Increased by 0.8ppb/yr during 1990s. Increase rate slower in 1990s than 1980s. Atmospheric lifetime: 120 years. *Ecosystem:* Terrestrial and oceanic sources, exponential rise in concentration since preindustrial. Agricultural emissions increased fourfold from 1900 to 1994 (Kroeze et al. 1999).	– climate change: emission higher in wetter soils – land use and management: acceleration of the global N cycle due to fertilizer use and agricultural N fixation, animal production – biomass burning – N deposition – atmospheric NO_x pollution	– climate: positive radiative forcing (greenhouse gas, heating) – stratospheric ozone depletion

Table 13.1. *continued*

Atmospheric Constituents	Sources	Sinks	Trends	Ecosystem Drivers	Impacts
NO_x (NO and NO_2) (precursors of nitrate)	*Ecosystem:* soils (mostly tropical) 13–21 TgN/yr biomass burning 7.1 TgN/yr *Other:* fossil fuel 33.0 TgN/yr aircraft 0.7 TgN/yr lightning 5.0 TgN/yr stratosphere <0.5 TgN/yr	*Ecosystem:* canopy uptake of soil emissions 4.7–8TgN/yr and of wet and dry deposition *Other:* reaction with OH to form nitric acid (HNO_3), which collects on aerosols (dry deposition) or dissolves in precipitation (wet deposition)	*Atmospheric Concentration:* Difficult to quantify because of the tremendous spatial and vertical variability. Atmospheric Lifetime: <0.01–0.03 years. Nitrate concentrations declined recently due to emission controls. *Ecosystem:* Difficult to quantify trend, largely stable.	– climate change: warming increases emissions – land use and management: tropical deforestation reduces soil emissions, but reduces canopy uptake more so net emissions increase. Acceleration of global N cycle due to fertilizer use, etc. – biomass burning	– human health: direct respiratory effects and respiratory effects of aerosols – tropospheric ozone precursor – tropospheric oxidizing capacity increase – acid rain formation – fertilization of plants (deposition) – eutrophication of lakes (deposition and nitrate leaching)
NH_3	*Ecosystem:* domestic animals 22 TgN/yr fertilizer use 9 TgN/yr crops (+ decomposition) 4 TgN/yr natural soils 2 TgN/yr oceans 8 TgN/yr biomass burning 6 TgN/yr	*Ecosystem:* direct soil and plant uptake wet and dry deposition (affected by vegetation cover) *Other:* reaction with OH (very small percentage)	*Atmospheric Concentration:* Documentation of trends is challenging because of the relatively short atmospheric lifetime. Atmospheric lifetime: 1 day–1 week. *Ecosystem:* Rise in agricultural sources exponential, much of the growth occurring since 1950. Main source areas Europe and North America (fertilizer use) and India (cattle).	– land use and management: acceleration of global N cycle due to fertilizer use, agricultural intensification/management; land cover change – biomass burning	– human health: hypoxia, pfisteria, respiratory effects – eutrophication of lakes (deposition and nitrate leaching) – acid neutralization and production – aerosol/particulate formation
SO_2/SO_4/H_2S DMS (dimethylsulfide) (precursors of sulfate aerosols)	*Ecosystem:* biomass burning [SO_2] 2.2 Tg/yr land biota [H_2S] 1.0 Tg S/yr marine plankton [DMS] 24 Tg S/yr *Other:* fossil fuel emissions [SO_2] 76 Tg S/yr volcanoes [SO_2] 9.3 Tg S/yr	*Ecosystem:* direct soil and plant uptake wet and dry deposition (affected by vegetation cover) ecosystems are a sink for about 30% of SO_2 emissions and sulphate aerosols *Other:* reaction with OH	*Atmospheric concentration:* Sulfate concentration in 1960s four times that of preindustrial, but declined recently due to stringent emissions regulations. Patchy distribution around source areas — polluted regions North America, Europe, and China. SO_2 emissions declining in North America and Europe, rising in South and East Asia. Emissions are projected to decrease substantially over the next century. *Ecosystem:* Mostly stable.	– climate change – land use and management; land cover change – biomass burning	– human health: respiratory effects of aerosols – climate: negative radiative forcing (cooling) – indirect climate impacts: cloud condensation nuclei – acid rain – reduced NPP though reduced solar radiation
VOCs (volatile organic compounds)	*Ecosystem:* vegetation (mostly tropical): isoprene 220 TgC/yr terpene 127 TgC/yr acetone 50 TgC /yr methanol 70–350 Tg/yr biomass burning 33 TgC/yr *Other:* fossil fuel 161 TgC/yr	*Ecosystem:* direct soil and plant uptake wet and dry deposition (affected by vegetation cover) *Other:* reaction with OH	*Atmospheric concentration:* Difficult to quantify because of the tremendous spatial and vertical variability. Atmospheric lifetime: < 1 day to >1 week. Emissions probably increased due to increasing use of gasoline and other hydrocarbon products. *Ecosystem:* Deforestation has probably decreased natural emissions.	– climate change (warming increases emissions) – land use and management: forest cover change, agricultural management, use of fertilizers – biomass burning – N deposition	– human health: aerosol precursor (terpene), respiratory effects – indirect climate impacts: cloud condensation nuclei – tropospheric ozone formation (isoprene) – tropospheric oxidizing capacity (increase) – organic acid formation – acid rain
Aerosols: organic matter	*Ecosystem:* biomass burning 54 Tg/yr biogenic (VOC oxidation, plant debris, humic matter and microbial particles) 56 Tg/yr *Other:* fossil fuel 28 Tg/yr	wet and dry deposition (affected by vegetation cover)	*Ecosystem:* Biogenic aerosols increasing due to increase in oxidizing agents, e.g., NO_3 and O_3, possible three- to fourfold increase since preindustrial times (Kanakidou et al. 2000). Not much is known about emissions from plants, microbes, and humic matter, but likely to be strongly affected by land use change.	– land use and management – biomass burning	– human health: respiratory effects – climate impacts: negative radiative forcing (cooling) – indirect climate impacts: cloud condensation nuclei – reduced NPP though reduced solar radiation

(continues over)

Table 13.1. *continued*

Atmospheric Constituents	Sources	Sinks	Trends	Ecosystem Drivers	Impacts
Aerosols: black carbon (0–2 μm)	*Ecosystem:* biomass burning 5.7 Tg/yr *Other:* fossil fuel 6.6 Tg/yr	wet and dry deposition (affected by vegetation cover)	*Atmospheric concentration:* Concentration and trend uncertain.	– land use and management: vegetation cover – biomass burning	– climate: positive radiative forcing (heating) – human health: respiratory effects – reduced NPP though reduced solar radiation
Aerosols: dust	*Ecosystem:* mineral (soil) dust 2000 Tg/yr (1–2 μm 300 Tg/yr); mainly from desert/dryland areas, <10 % from disturbed soil surfaces (Tegen et al. 2004) *Other:* industrial dust (>1 μm) 100 Tg/yr	wet and dry deposition (affected by vegetation cover)	*Atmospheric concentration:* Concentration and trend uncertain. *Ecosystem:* Decrease in North American emissions since dust bowl years due to changes in management. Emissions from Sahara/Sahel increased significantly since 1960s, possibly due to changing wind patterns and desertification. Chinese desert areas and loess plateau variable, trend and causes not clear. Climate change impacts–model results inconclusive.	– climate variability – climate change – land use and management: reduction in land cover, agricultural management – desertification	– human health: respiratory effects, irritation – climate impacts: radiative forcing net sign and magnitude unclear as reflects incoming radiation and traps outgoing radiation – fertilizing effects of iron in ocean and phosphate on land in some regions, increasing productivity, indirect climate effect as CO_2 sink – reduced NPP though reduced solar radiation – visibility reduction
Tropospheric ozone	*Ecosystem:* ecosystem precursors VOCs (isoprene), NO_x, CH_4, CO primarily from biomass burning in the tropics *Other:* transport from stratosphere 475 Tg O_3/yr precursors in urban pollution	*Ecosystem:* dry deposition: 620–1178 Tg O_3/yr *Other:* stratosphere/troposphere exchange: 400–1440 Tg O_3/yr	*Atmospheric Concentration:* Increased from preindustrial conc. 25 DU (Dobson units) to 34 DU (370 Tg O_3) in late 1990s. Increased from 1970 to 1980 but no clear trend from 1980 to 1996. Difficult to quantify due to the high reactivity and spatial and temporal variability of sources, but satellite measurements may improve quantification. Models predict increasing tropospheric O_3 driven regionally by increasing emissions of pollutants. Atmospheric lifetime: 0.01–0.05 years. Concentrated over areas of urban pollution and biomass burning.	– emissions of key precursor trace species including VOCs, CH_4, NO_x, CO – climate change: warming increases concentration – biomass burning	– human health: UV exposure – climate: positive radiative forcing (heating) – troposphere oxidizing capacity – stratospheric ozone production
OH radical	reactions between tropospheric ozone, non-methane hydrocarbons, and NO_x in UV light	reactions with many reduced compounds, especially CO and CH_4	Probably declining since preindustrial but not by more than 10%.	– emissions of key precursors and sinks	– reduced OH leads to reduced tropospheric oxidizing capacity
Biophysical Surface Properties	**Non-Forests**	**Forests**	**Trends**	**Ecosystem Drivers**	**Impacts**
Surface albedo	snow-free: 0.16 (tall grasslands) to 0.6 (sand desert) snow-covered: 0.5 to 0.8	snow-free: 0.11 (tropical evergreen) to 0.2 (deciduous) snow-covered: 0.2 to 0.25	Increase in mid-latitudes due to deforestation until middle of twentieth century, now decrease due to regrowth in some mid-latitude areas. Increase in tropics.	– land use and management: primarily forest cover	– climate: radiative forcing (higher albedo = more reflection = cooling)
Water fluxes (evapotranspiration)	up to 5 mm/day	up to 10 mm/day	Evapotranspiration decrease, especially in tropics, due to deforestation.	– land use and management: primarily forest cover	– climate: direct impacts on radiative forcing and indirect impactsvia clouds – hydrological cycle
Surface roughness	up to 0.1m	1.0–2.5m	Decrease, especially in tropics, due to deforestation.	– land use and management: primarily forest cover	– climate: atmospheric circulation (wind)

[a] All numbers relate to the 1990s and are from IPPC 2001b unless otherwise stated.

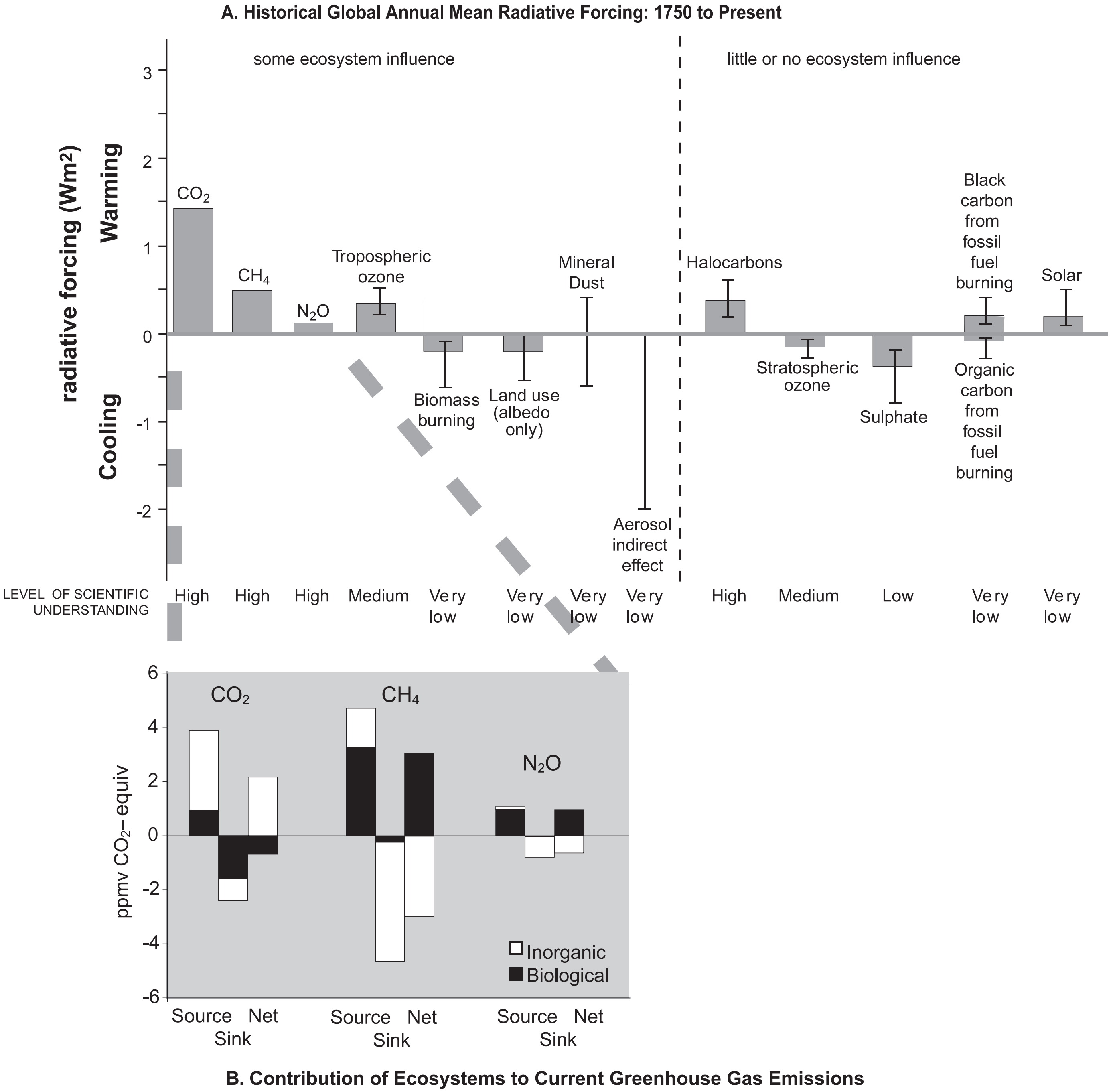

Figure 13.3. Contribution of Ecosystems to Historical Radiative Forcing and Current Greenhouse Gas Emissions (Adapted from IPCC 2001a, 2001b). Figure A is the radiative forcing caused by changes in atmospheric composition, alteration in land surface reflectance (albedo), and variation in the output of the sun for the year 2000 relative to the conditions in 1750. The height of the bar represents a best estimate, and the accompanying vertical line a likely range of values. We have separated factors with a significant ecosystem influence from those without. The indirect effect of aerosols shown is their effect on cloud droplet size and number, not cloud lifetime. Some of the radiative components are well mixed over the globe, such as CO_2, thereby perturbing the global heat balance. Others represent perturbations with stronger regional signatures because of their spatial distribution, such as aerosols. Radiative forcing continues to be a useful tool to estimate to a first order, the relative climate impacts such as the relative global mean surface temperature response due to radiatively induced perturbations, but these global mean forcing estimates are not necessarily indicators of the detailed aspects of the potential climate responses (e.g., regional climate change). Figure B is the relative contribution of ecosystems to sources, sinks, and net change of three of the main greenhouse gases. These can be compared by conversion into CO_2–equivalent values, based on the global warming potential (radiative impact times atmospheric lifetime) of the different gases. For CH_4 and N_2O, a 100-year time scale was assumed; a shorter time scale would increase the relative value compared with CO_2 and a longer time scale would reduce it. Ecosystems are also a net sink for tropospheric ozone, but it is difficult to calculate emissions in CO_2–equivalent values.

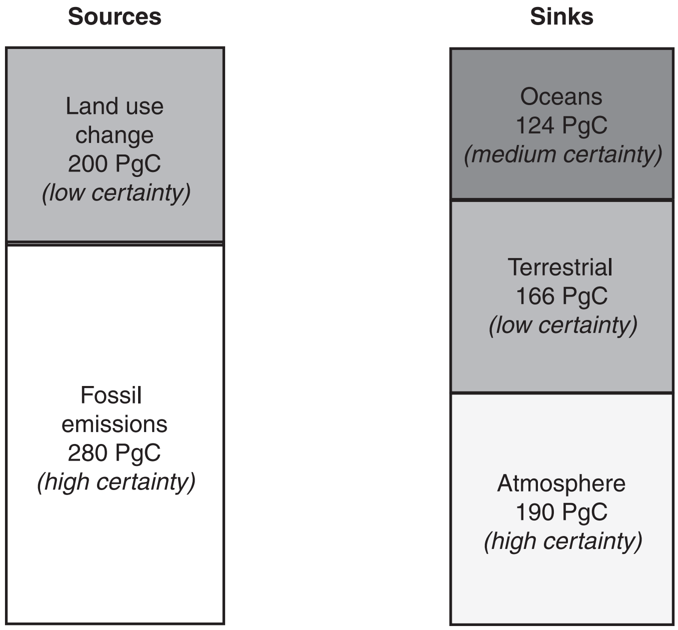

Figure 13.4. Carbon Sources and Sinks over the Last Two Centuries. Total carbon losses to the atmosphere due to historical land use change have been estimated at around 200 PgC (DeFries et al. 1999): two thirds to three quarters of this loss was due to conversion of forestland to cropland or other land uses; other carbon losses included degradation of grasslands and shrublands and the conversion of non-forestland to cropland. Fossil fuel emissions from pre-industrial times to 2000 are estimated as 280 PgC (Marland et al. 2000, update in Prentice et al. 2001), but the atmospheric increase during the same period was only 190 PgC. About 124 PgC, or ≈26% of the total emissions, were taken up by the oceans primarily as a result of chemical and physical processes (dissolution and mixing) (House et al. 2002 based on Gruber 1998; Sabine et al. 1999, 2002; Prentice et al. 2001; Langenfelds et al. 1999; Manning 2001). The remaining 166 PgC or ~34% was absorbed by the land biosphere. In this analysis, historical land use change is responsible for about 40% of the observed growth in atmospheric CO_2; Brovkin et al. (2004) estimate a range of 25–49%, with the lower end of this range being more likely.

Concern about global warming and the implementation of the Kyoto Protocol has led to carbon uptake for climate regulation being considered as an important ecosystem service (MA *Policy Responses,* Chapter 13). Forests can be managed as a sink (including preventing deforestation and promoting afforestation, reforestation, and improved forest management). This approach implies that once the forests stop growing, they must be protected to avoid loss of most of the carbon store (and to encourage long-term storage in soil carbon pools). Alternatively, forest biomass can be used to produce long-lived products that store carbon (such as furniture), or as a substitute for materials that are energy-intensive to produce (such as aluminum and plastics), or as biomass fuels that are used instead of fossil fuels.

In these ways, an area of forest can continue to offset CO_2 emissions indefinitely and may provide other services at the same time. The potential trade-offs with other environmental and socioeconomic values are also relevant—for example, biodiversity maintenance. (See Chapter 9 and MA *Policy Responses,* Chapter 13). The prospect of carbon trading under the Clean Development Mechanism, along with public and industry awareness of climate change issues, is already promoting small-scale forest activities with a view to carbon sequestration and the production of modern biomass fuels.

Forest degradation resulting from overexploitation can result in substantial carbon losses. For example, about 0.5 megagrams of carbon per hectare per year is being lost in Southeast Asia (Kim Phat et al. 2004) by this mechanism. Forest fragmentation leads to increased rates of big tree mortality, decomposition, and fire, causing carbon losses greater than deforestation in some areas (Nascimento and Laurance 2004). In central Amazonia, model estimates for the first half of this century suggest annual fragmentation losses of 4–5 megagrams of carbon per hectare per year. Timber removal from extant Amazonian forests is likely to account for a loss equal to about 10–20 megagrams of carbon per hectare per year, and fires for double that amount (Nepstad et al. 1999). Agroforestry systems have an annual sequestration capacity of 0.2–0.3 megagrams of carbon per hectare per year, and about 400 million hectares of degraded land are potentially suitable for agroforestry systems globally (IPCC 2000). One North Indian agroforestry system sequestered up to 19.6 megagrams of carbon per hectare per year (Singh et al. 2000).

Agricultural management alters the amount of carbon contained in soils and also affects other greenhouse gases such as nitrous oxide and methane (e.g., Rosenberg et al. 1998; Paustian et al. 2000; West and Post 2002; Witt et al. 2000). Soil carbon is lost when land is cleared and tilled, but it can be regained through low-tillage agriculture and other methods (Lal et al. 2004). It remains uncertain how much carbon contained in soil eroded from agricultural landscapes is delivered to the atmosphere in policy-relevant timescales (Renwick et al. 2004). Actions taken to promote carbon sequestration in soils would reduce dust emissions, which have a fertilizing effect on the ocean, thus reducing the uptake of CO_2 by the ocean and diminishing the effectiveness of this response strategy to an unknown degree (Ridgwell et al. 2002). Complex interactions of this type highlight the need to consider all implications of management options (see MA *Policy Responses,* Chapter 13). In the lowland tropics and sub-tropical areas where rice is now grown continuously throughout the year (continuous flooding), soil carbon has been seen to increase (Bronson et al. 1997; Cheng 1984). With intensive management typical of these high-yield continuous rice systems, net carbon sequestration rates of 0.7–1.0 megagrams of carbon per hectare per year have been measured (Witt et al. 2000). (See Chapter 26.)

Management of marine ecosystems to increase oceanic carbon sequestration (''ocean engineering,'' for instance via iron fertilization) has some theoretical potential, but very little is known about potential ecological and geochemical risks of such an endeavor. Due to its comparatively low estimated cost (probably in the range of a few dollars per ton of CO_2), the approach is financially attractive but the capacity is too small to slow down anthropogenic CO_2 increase significantly. Early upper limit calculations show that the potential for reducing atmospheric CO_2 is limited to a few tens of parts per million (Peng and Broecker 1991; Joos et al. 1991), in agreement with the most recent high-resolution ice core data of dust deposition and atmospheric CO_2 (Röthlisberger et al. 2004). More recent calculations put the maximum potential at around 1.0 petagrams (1.0 billion tons) of carbon per year for a maximum of 100–150 years, although it is likely to be much smaller, say less than 0.2 petagrams of carbon per year (Matear and Elliott 2004; Caldeira et al. 2004).

Future trends in atmospheric CO_2 are likely to depend more on fossil fuel emissions than on ecosystem change. Although land use management can have a significant impact on CO_2 concentra-

Table 13.2. Annual Fluxes of Carbon Dioxide over the Last Two Decades. Positive values represent atmospheric increase (or ocean/land sources); negative numbers represent atmospheric decrease (ocean/land sinks). Land and ocean uptake of CO_2 can be separated using atmospheric measurements of oxygen (O_2) in addition to CO_2 because biological processes on land involve simultaneous exchange of O_2 with CO_2, while ocean uptake does not (ocean ecosystems are assumed to be in equilibrium for the purposes of this calculation). While this technique allows quantification of the net contribution of terrestrial ecosystems, breaking this down into sources and sinks requires modeling. The CO_2 released due to land use change during the 1980s has been estimated as the range across different modeling approaches. The difference between the modeled land use change emissions and the net land-atmosphere flux can then be interpreted as uptake by terrestrial ecosystems (the "residual terrestrial sink"). (IPCC data from Prentice et al. 2001)

	IPCC		Update[a]	
Source of Flux	**1980s**	**1990s**	**1980s**	**1990s**
		(gigatons of carbon equivalent per year)		
Atmospheric increase	+3.3 ± 0.1	+3.2 ± 0.1		
Anthropogenic emissions (fossil fuel, cement)	+5.4 ± 0.3	+6.3 ± 0.4		
Ocean-atmosphere flux	−1.9 ± 0.6	−1.7 ± 0.5	−1.8 ± 0.8	−1.9 ± 0.7
Land–atmosphere flux:	−0.2 ± 0.7	−1.4 ± 0.7	−0.3 ± 0.9	−1.2 ± 0.9
—Land use change[b]	+1.7 (+0.6 to +2.5)	incomplete	+1.3 (+0.3 to 2.8)	+1.6 (+0.5 to +3.0)
—Residual terrestrial sink	−1.9 (−3.8 to +0.3)	incomplete	−1.6 (−4.0 to −0.0)	−2.8 (−5.0 to −0.9)

[a] Same data as Prentice et al. 2001 but including a correction for the air-sea flux of oxygen caused by changes in ocean circulation (Le Quéré et al. 2003). Other estimates of the air-sea oxygen correction have given slightly different results for the 1990s, mostly due to the fact that direct observations of heat change in the ocean have not yet been compiled for after 1998 (Keeling and Garcia 2002; Plattner et al. 2002; Bopp et al. 2002).

[b] The IPCC estimated range for the land use change flux is based on the full range of Houghton's bookkeeping model approach (Houghton 1999) and the CCMLP ecosystem model intercomparison (McGuire et al. 2001). The update is based on the full range of Houghton (2003) and DeFries et al. (2002); the CCMCP analysis only extended to 1995.

tions in the short term (Prentice et al. 2001), the maximum feasible reforestation and afforestation activities over the next 50 years would result in a reduction in CO_2 concentration of only about 15–30 ppm by the end of the century (IPCC 2000). Even if all the carbon released so far by anthropogenic land use changes throughout history could be restored to the terrestrial biosphere, atmospheric CO_2 concentration at the end of the century would be about 40–70 ppm less than it would be if no such intervention had occurred (Prentice et al. 2001, House et al. 2002). Conversely, complete global deforestation over the same time frame would increase atmospheric concentrations by about 130–290 ppm (House et al. 2002).

This compares with the projected range of CO_2 concentrations in 2100, under emissions scenarios developed for the IPCC, of 170–600 ppm above 2000 levels, mostly due to fossil fuel emissions (Prentice et al. 2001). The ability of the land and ocean to take up additional increments of carbon decreases as the CO_2 concentration rises, primarily due to the finite buffering capacity and rate at which ocean water can take up CO_2, as well as the saturation of the CO_2 fertilization response of plant growth (Cox et al. 2000; Prentice et al. 2001). Global warming is predicted to have a strong positive feedback on the carbon cycle due, for example, to increases in soil organic matter decomposition and a reduction in inorganic ocean uptake due to reduced CO_2 solubility and ocean stratification at higher temperatures, as described later.

13.2.2 Methane

Methane is a greenhouse gas and an energy source (natural gas). It is involved in many atmospheric chemistry reactions, including formation of tropospheric and stratospheric ozone and the reduction of atmospheric cleansing capacity (see section 13.4.1). The major source of methane is microorganisms living in a variety of anaerobic environments such as flooded wetlands and rice paddies, the guts of termites and ruminant animals, the ocean, landfill sites, and waste treatment plants. Other soil bacteria re-oxidize CH_4, preventing much of the CH_4 produced in anaerobic (wet) soils from reaching the atmosphere and accounting for a small but significant sink of atmospheric CH_4 in soils remote from methane sources.

The current atmospheric concentration of CH_4 is more than twice that of preindustrial times, as Table 13.1 indicated (see also MA *Scenarios,* Chapter 7). The growth rate peaked in 1981 and has declined since (Prather et al. 2001), with no increase in the concentration between 1999 and 2002 (Dlugokencky et al. 2003). The observed values are subject to high interannual variability due to changes in sources, sinks, atmospheric transport, atmospheric chemistry, and climate variability. Emissions from ecosystems account for about 70% of total emissions (Prather et al. 2001), with about 30% from wetlands and 30% from agriculture. The spatial distribution of emissions from natural wetlands and agriculture (based on modeling) is shown in Figure 13.6. While northern wetlands are rather well studied with respect to magnitude and drivers of CH_4 emissions, the lack of data from tropical wetlands is a major knowledge gap.

Agriculture (ruminant animals and rice paddies) is the most important anthropogenic driver of CH_4 emissions. In pastoral countries such as Bolivia, Uruguay, and New Zealand, CH_4 emissions by ruminant livestock are responsible for over 40% of all greenhouse gas emissions (when expressed as CO_2 equivalents)

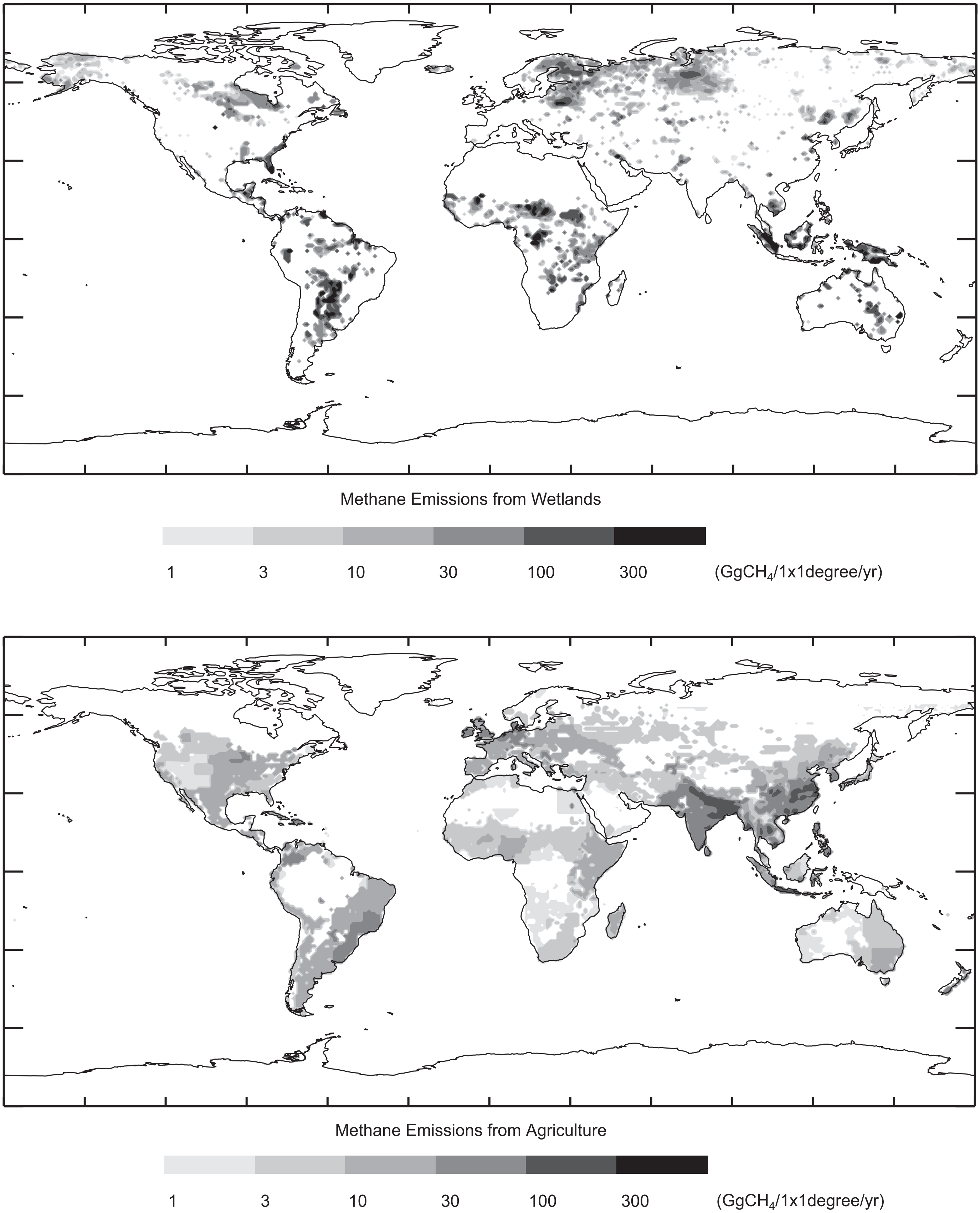

Figure 13.6. Map of Methane Fluxes from Wetlands and Agriculture Models. The top figure shows the spatial location of methane fluxes from natural wetlands according to Walter and Heimann (2000). The total emissions are probably too high; a maximum of about 160–170 $GgCH_4$/yr may be more realistic (Houweling et al. 2000). The bottom figure shows emissions from agriculture sources (rice paddies, animals and animal wastes) according to the model EDGAR (v3.2) (Olivier et al. 1994).

(UNFCCC at ghg.unfccc.int). Control of this methane source through changing diet is being examined as a potential control on global warming. Methane emissions from lowland, irrigated rice systems are affected by various management practices. For example, the use of green manures can substantially increase CH_4 emissions compared with the use of inorganic N fertilizers (Wassmann et al. 2000a, 2000b). (More detailed discussion of rice paddies can be found in Chapter 26.)

Wetland and peatland soils under waterlogged or seasonally frozen conditions tend to store carbon fixed by plants during photosynthesis. This is because in waterlogged soils, decomposition of plant material is slower than in aerated soils and is accompanied by a relatively slow release of CH_4 compared to the faster CO_2 efflux of aerobic soils. Wetland draining for agriculture, for forestation, or due to water extraction leads to a decrease in CH_4 production but a rapid increase in CO_2 release from soils that are often very rich in organic matter. The current consensus is that, while some areas are carbon sinks and some are carbon sources, most wetlands—in particular, the northern peat-forming ones—have a carbon balance close to zero (Callaghan et al. 2004).

CH_4 has a higher global warming potential than CO_2 (23:1 on a 100-year time horizon), although CO_2 is much longer-lived in the atmosphere. In the short term (20–100 years), CH_4 emissions have a more powerful effect on the climate per unit mass than CO_2. Thus as most wetlands are currently sources of CH_4, they are also net sources of radiative forcing (*medium certainty*) (Friborg et al. 2003). Draining of wetlands leads to a decrease in radiative forcing (cooling) in the short term, but in the long term the opposite may be true (*low certainty*) (IPCC 2001a; Christensen and Keller 2003).

The climate change impacts on peatlands have different feedbacks depending on the location and extent of global warming. Some regions, such as northern Alaska, are predicted to experience—and are indeed already experiencing—soil drying, leading to net losses of carbon as CO_2 and decreases in CH_4 emissions (Chapter 25). In the subarctic, however, where recent decadal warming has led to the loss of permafrost with thermokarst erosion and a wetting of the soils as a consequence, a significant increase in CH_4 emissions has been seen (Christensen et al. 2004), with the overall effect of an increase in radiative forcing. In tropical regions, the key issue is what changes will occur to seasonal flooding—draining will lower the impact on radiative forcing as the CH_4 emissions are very high, yet the carbon store in peat is insufficient for emissions of CO_2 to overwhelm this in the long term.

Preliminary model results of changes in wetland areas under future climate (e.g., Cox et al. 2000) suggest that the northern wetlands will increase their carbon storage and subsequent CH_4 emissions while tropical ones will lose significant amounts of carbon and decrease CH_4 emissions. Furthermore, rising soil temperature and enhanced microbial rates could increase methane emissions, amplifying global warming by 3.5–5% by the end of this century (Gedney and Cox 2003; Gedney et al. 2004).

13.2.3 Nitrous Oxide

Ecosystems are a source of nitrogen in various gaseous forms, each with different effects on climate and air quality. They are also a sink for atmospheric nitrogen, taking up N_2 directly from the atmosphere (nitrogen fixation) or reactive nitrogen after deposition (wet deposition–rained out, or dry deposition). For a detailed explanation of the nitrogen cycle and the role of ecosystems, see Chapter 12. The nitrogen cycle has been profoundly altered by use of synthetic fertilizers. (See Chapter 26.)

Nitrous oxide is a powerful, long-lived greenhouse gas (GWP 296:1 on a 100-year time horizon), which (unlike other N oxides) is unreactive in the troposphere. Atmospheric concentrations of N_2O have increased since preindustrial times from 270 ppb to 314 ppb (MA *Scenarios,* Chapter 7). Ecosystem sources—primarily soil microorganisms in an array of environments—account for about 90% of N_2O emissions and a small fraction of N_2O uptake (Prather et al. 2001). Enhanced ecosystem N_2O emissions are mainly driven by increased fertilizer use, agricultural nitrogen fixation, and atmospheric nitrogen deposition (Nevison and Holland 1997; Prather et al. 2001). Wetland draining also increases N_2O emissions. Fertilizer use and nitrogen deposition are projected to increase substantially in the tropics (Matson et al. 1999; Prather et al. 2001).

13.2.4 Tropospheric Ozone

Besides being a greenhouse gas, tropospheric ozone is a toxic pollutant. It is highly reactive in the atmosphere and also helps maintain the atmospheric cleansing capacity. It is formed in the atmosphere in the presence of light from precursors: mainly volatile organic compounds (the most important VOC being isoprene), nitrogen oxides (NO and NO_2, collectively denoted as NO_x), CH_4, and carbon monoxide. Biomass burning is an ecosystem source of precursors, but urban pollution sources dominate, with very high concentrations of ozone mostly appearing downwind of urban areas. In addition to being a source of tropospheric ozone precursors, ecosystems account for about half the total sink for tropospheric ozone, through dry deposition (Prather et al. 2001). Ecosystems are thus currently a net sink for tropospheric ozone. Deforestation reduces this sink (see Figure 13.7): NO_x soil emissions decline, but canopy uptake declines more. Where forests are replaced with agriculture, NO_x emissions increase further.

The concentration, sources, and sinks of tropospheric ozone are difficult to quantify due to its high reactivity and the spatial and temporal variability of sources and sinks. Most surface measuring stations show an increase from 1970 to 1980, but no clear trend from 1980 to 1996. Models predict increasing tropospheric ozone in the future, driven regionally by increasing emissions of its precursors (Prather et al. 2001).

13.2.5 Aerosols

Ecosystems are sources and sinks for a variety of aerosols (or aerosol precursors) that directly affect radiative forcing, causing warming or cooling depending on their properties (such as reflectivity) and location (such as height or the underlying surface) (Penner et al. 2001). Many aerosols affect cloud formation, which in turn affects radiative forcing in complex ways. The net effect of clouds on radiative forcing remains uncertain, but increasing the aerosol load probably, on average, causes cooling. The net effect of aerosols on climate is matter of intensive investigation, and while the field is progressing rapidly, more research is needed before firm conclusions can be drawn.

Sulfur compounds (sulfur dioxide, hydrogen sulfide, and dimethyl sulphide) contribute to the formation of sulfate aerosols with a negative radiative forcing (climate cooling) (MA *Scenarios,* Chapter 7). Industrial sources dominate, despite declines in some regions due to controls and legislation (Rodhe 1999; Penner et al. 2001). Ecosystems are a sink for about 30% of SO_2 emissions and for sulfate aerosols. Dimethyl sulfide, emitted by marine phytoplankton when they die or are eaten, contributes to cloud formation. DMS emissions are quite variable in relation to phytoplankton species and mode of release. Global mean DMS concentration in surface waters is fairly well known, but regional emissions are

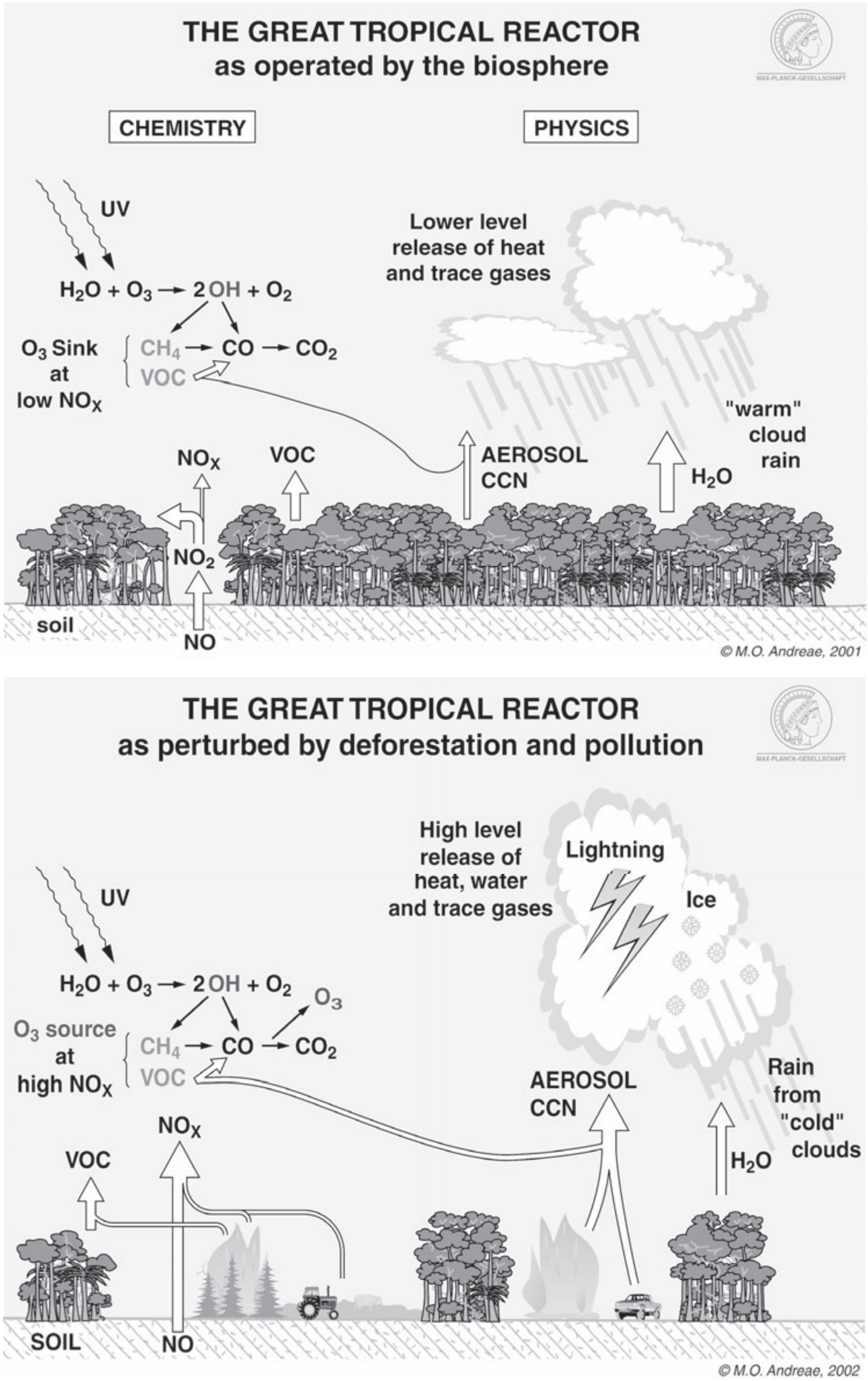

Figure 13.7. The Tropical Reactor—Biochemical and Biophysical Interactions in the Tropics (Andreae 2001). The top figure shows the natural biochemical and biophysical fluxes and interactions over intact forest. The bottom figure shows the fluxes when the natural forest area is subject to deforestation and pollution. The land changes from a net sink of tropospheric ozone to a net source when the canopy does not trap NO_x emissions from soil, and emissions of NO_x from agriculture increase. Aerosols from vegetation fires and pollutions sources (e.g., cars) act as cloud condensation nuclei leading to higher storm clouds. Overall rainfall is reduced due to a reduction or evapotranspiration (water recycling through vegetation).

more uncertain. Climate-related change in surface ocean stratification is bound to affect phytoplankton species distribution and succession, with likely consequences for DMS production and potentially also for cloud formation (Kiene et al. 1996).

Carbonaceous aerosols of many forms are emitted by ecosystems and can cause warming or cooling depending on their composition, size, shape, and location. IPCC (2001a) concluded that biomass-burning aerosols have a net cooling effect (indoor biomass fuel burning was underrepresented in this estimate). But these aerosols can also reduce cloudiness, which is thought to enhance climate warming (Penner et al. 2003). Black carbon (soot from incomplete combustion of biomass and fossil fuels) has been suggested to have a large warming effect on the climate, due in part to various feedbacks (Hansen and Sato 2001; Jacobson 2002), although the magnitude of this effect is disputed (Penner et al. 2003; Roberts and Jones 2004). Air quality impacts of biomass-burning aerosols are dealt with later in this chapter.

There are huge uncertainties regarding emissions and trends of biogenic aerosols (VOC oxidation products, plant debris, humic matter, and microbial particles), thus IPCC (2001a) did not estimate their net radiative warming impacts. Their contribution could be significant in densely vegetated regions of the tropics, and it is under threat as a result of land use change.

Mineral dust is entrained into the atmosphere from sparsely vegetated soils. Dust scatters solar radiation and absorbs terrestrial radiation. Its net impact on radiative forcing is uncertain, but updates since IPCC (2001a) infer that the net effect is cooling (Kaufman et al. 2001). Its trends and drivers are described later in this chapter.

13.3 Biophysical Effects of Ecosystems on Climate

13.3.1 Surface Properties and Climate Processes Affected by Ecosystems

Ecosystems affect climate through the alteration of energy and water fluxes in the lowest atmosphere, or planetary boundary layer (about 1–2.5 kilometers above Earth's surface) (Pitman et al. 2004). Within this layer, vertical profiles of temperature and humidity depend strongly on the partitioning of energy between sensible heat and latent heat. Over land, this partitioning is largely controlled by ecosystems. Over bare, dry land, the energy is transported via sensible heat, resulting in relatively high surface air temperatures. Vegetation canopies transpire water extracted from the root zone, increase the upward latent heat flux, and cool the surface air (Avissar et al. 2004). Modification of the fluxes of water and energy by ecosystems has significant regional effects on precipitation, temperature, and wind. Globally averaged impacts are small, complex, and hard to detect against the background of natural climate variability and anthropogenic climate change. Key physical properties and processes affected by ecosystems are summarized here:

- ***Surface albedo*** is the fraction of solar radiation reflected back into the atmosphere from Earth's surface. Higher albedo means that more energy leaves the planetary boundary layer (net cooling of the atmosphere). Vegetation traps radiation, generally reducing albedo compared with, for instance, snow cover or bare ground in dry lands. In agricultural regions, tillage usually decreases albedo since bare soil in moist climates is generally darker (less reflective) than plant canopies. Forests are very effective at trapping radiation by multiple reflection within the canopy; this effect is particularly strong in snow-covered regions where trees extend above the snow, while short vegetation such as crops and pastures are covered by snow. (See Figure 13.8 in Appendix A.) Phytoplankton modify ocean surface albedo, with different types either reducing (Frouin and Lacobellis 2002) or increasing it (Brown and Yoder 1994; Balch et al. 1991).
- ***Transpiration*** is the flux of water from the ground to the atmosphere through plants, controlled by the opening and closing of tiny pores in the leaf's surface called stomata. The volume of water transpired is determined by vegetation rooting depth, leaf area, soil moisture, temperature, wind, and stomatal conductance (which is biologically regulated). Transpiration drives the hydrological cycle—recycling rain water back to the atmosphere to be rained out elsewhere. Thus terrestrial ecosystems mediate the service of water recycling. Through transpiration and precipitation, water evaporated over the ocean is transported into the interior of continents. A part of the rainfall escapes immediate recycling and forms river runoff; thus the presence of vegetation reduces the fraction of rainfall going into runoff. (See Figure 13.9.) As runoff is part of the freshwater flux into the surface ocean, changes in terrestrial ecosystems can in principle affect ocean dynamics. Transpiration cools the surface during the daytime and increases air humidity in the near-surface atmospheric layer. Increased concentration of water vapor (a greenhouse gas) leads to reduced fluctuations in the diurnal temperature cycle by increasing the night temperatures. Photosynthesis is tightly coupled to transpiration, but while increased atmospheric CO_2 concentration in the future is likely to enhance photosynthesis, it may tend to reduce transpiration due to reduced stomatal conductance (*medium certainty*).
- ***Cloud formation*** has strong but complex effects on global and regional climate (Stocker et al. 2001). Evapotranspiration determines the availability of water vapor for the formation of clouds. Clouds alter the radiation balance (low clouds are cooling while high cirrus clouds are warming), air circulation, and precipitation. Vegetation also affects cloud formation via changes in surface albedo and roughness. Some of the atmospheric constituents with ecosystem sources act as cloud condensation nuclei: in particular DMS emitted by marine plankton, VOCs emitted by some types of vegetation, and some aerosols emitted during biomass burning. Increased concentrations of CCNs produce more and smaller cloud droplets, making clouds more reflective and persistent; this has a cooling effect on Earth. In addition, such clouds tend to rise in the atmosphere, delaying the onset of rain; increasing ice formation, rainfall intensity, and lightning; creating more violent convective storms; and altering energy balances and air circulation (Andreae et al. 2004). The net effect on the total rainfall within a given area is unknown.

Both marine and terrestrial biota naturally regulate CCN concentrations to remain at fairly low levels (Charlson et al. 1997; Williams et al. 2002). Increased DMS and VOC emissions increase CCNs, which reduces radiation and cools the planet; this in turn reduces photosynthesis and emissions of DMS and VOCs and increases thermal stability, thus reducing the probability of cloud formation in a negative feedback loop. The natural regulation mechanism is becoming overwhelmed by anthropogenic emissions of aerosols and deforestation. Some aerosols, such as soot particles, absorb sunlight, which cools the surface and heats the atmosphere, reducing cloud formation.

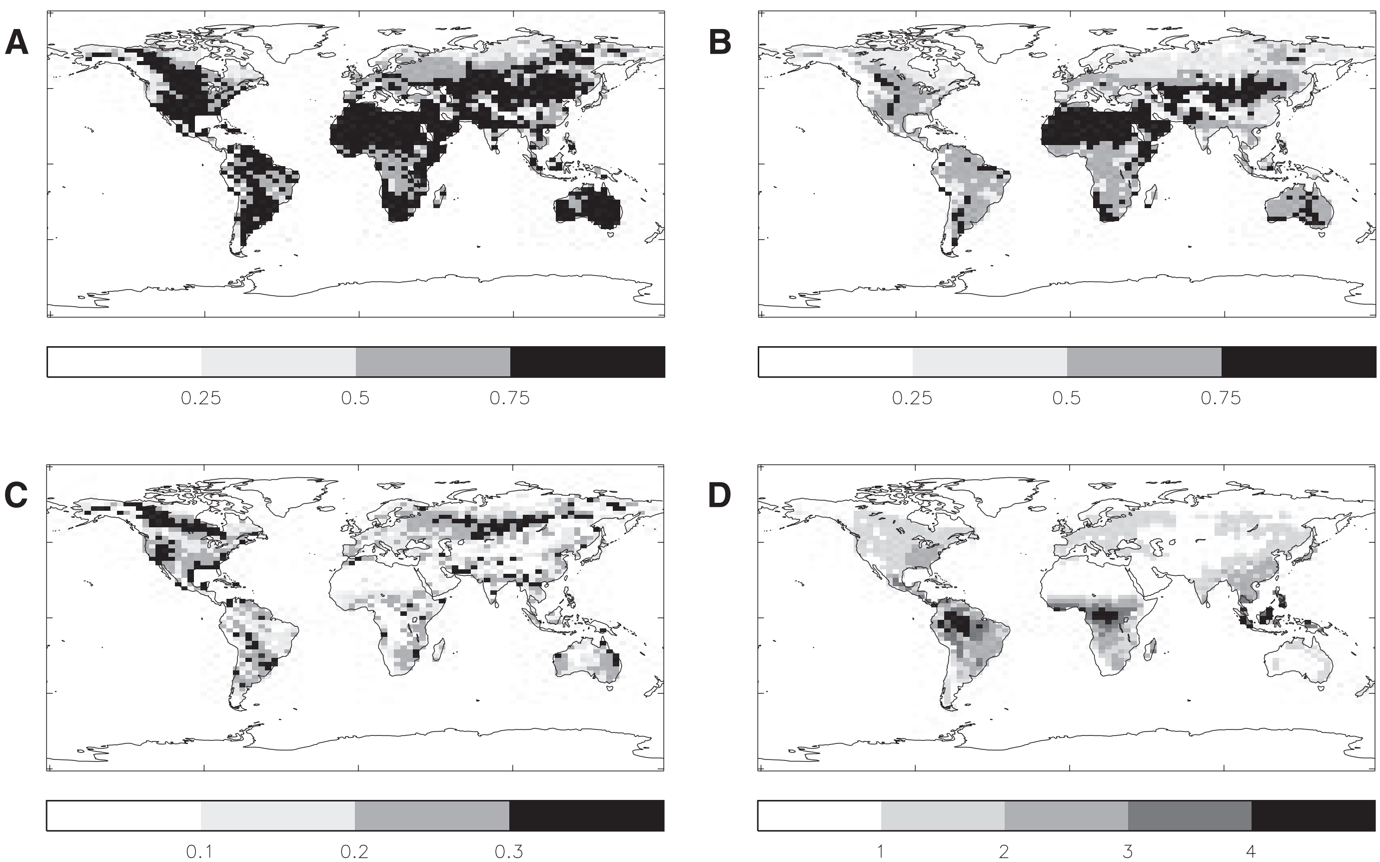

Figure 13.9. The Influence of Terrestrial Vegetation on Water Recycling (in mm/day). A general circulation model of climate has been used to simulate the ratio of land evapotranspiration to precipitation (Betts 1999) with (A) present-day vegetation and (B) all vegetation removed leaving bare soil. The difference between the two (C) illustrates the general increase in recycling of water back into the atmosphere via evapotranspiration when vegetation in present. For reference, (D) shows the absolute rate of evapotranspiration simulated with present-day vegetation.

- ***The aerodynamic properties of the surface*** (roughness length) modify the strength and direction of the surface wind. On land, the height and cover of surface vegetation are the main determinants of roughness length.
- ***Sea surface temperature*** is warmed by phytoplankton that trap radiation within the surface layer. A warmer surface reduces vertical mixing (ocean stratification) and ice cover, with potential feedbacks on regional circulation (Sathyendranath et al. 1991; Miller et al. 2003). Stratification reduces the flow of nutrients, feeding back on phytoplankton growth and composition, CO_2 uptake, and climate.

13.3.2 Biophysical Hotspots

The biophysical impacts of land use on climate are region- and season-specific and are often confounded by other climate drivers (for example, greenhouse gases and aerosols). On the global scale, biophysical effects of historical land cover changes are limited. Most model results indicate a slight biophysical cooling, partly offsetting the biogeochemical warming due to CO_2 emissions from land cover change (Betts 1999; Matthews et al. 2004; Brovkin et al. 2004). Regionally, the biophysical climate effects of ecosystem change can be substantial.

13.3.2.1 Tropical Forests

Tropical forests in South America, Africa, and Southeast Asia are being cleared to make land available for agriculture. (See Chapters 21 and 26.) The change in vegetation from forest to pasture or crops increases surface albedo (leading to cooling of the atmosphere), decreases roughness, and reduces evapotranspiration (reducing rainfall and leading to local warming) (Henderson-Sellers et al. 1993). Measurable impacts can be expected when the area of deforestation is on a scale of a few hundred kilometers. Trees can get access to deep soil water and have been observed to maintain evapotranspiration through the dry season at levels equal to the wet season. Thus extensive deforestation generally leads to decreased regional rainfall (see Box 13.2), although under some conditions rainfall can be higher over areas with partial deforestation compared with areas with no deforestation (Durieux et al. 2003; Avissar et al. 2002). Since forest existence crucially depends on rainfall, the relationship between tropical forests and precipitation forms a positive feedback, which, under certain conditions, theoretically leads to the existence of two steady states: rainforest and savanna (Sternberg 2001; Oyama and Nobre 2003), although some models suggest only one stable climate-vegetation state in the Amazon (Claussen 1998).

Biophysical impacts of tropical deforestation are different in the wet and dry seasons. In the dry season, non-forest areas become hot and dry during the daytime, and air flows from forests to non-forest areas, enhancing thermal turbulence. These conditions favor the formation of shallow rain clouds over non-forest areas. During the wet season, evaporation in forest and pasture are about the same but the forest reflects less radiation due to lower albedo, leading to warming of the planetary boundary layer.

BOX 13.2

Case Study: Deforestation and Rainfall Impacts in the Amazon (Maria Silva Dias, personal communication, 2002)

The state of Rondônia in the southwest Amazon was opened for colonization in the early 1970s by developers, following government incentives. Afonso Andrade was one of the first farmers, clearing an area of about 2,000 hectares for farming cattle and crops like rice, beans, and corn for local consumption. During the first five years, when Andrade's property was mostly surrounded by forest, he would plant a brown bean crop at the end of the wet season. Even if rain was scarce, the seeds would germinate and the crops would grow because dew was very abundant during the nighttime, and "the soil would be wet early in the morning." The flow of moisture from the neighboring forest during daytime would sustain the dew formation during nighttime. As deforestation proceeded, however, the forest was further away in successive years, the atmosphere and the soil became drier during the dry season, and the forest moisture was diluted into a larger area. Now it is very hard to get a crop going during the dry season without irrigation.

Warming generates more thermal turbulence, which favors the formation of clouds and rainfall over forest areas (Nobre et al. 2004). Cloud formation is further affected by aerosols from biomass burning and other sources (Williams et al. 2002).

DeFries et al. (2002) found that projected future land cover change, which is likely to occur mostly in the sub-tropics and tropics, will have a warming effect on climate, driven mostly by decrease in evapotranspiration. Increasing atmospheric CO_2 decreases stomatal conductance in many species. If this effect occurs on a large scale, it will reduce latent heat flux and therefore increase land surface temperature (Sellers 1996). In some cases, agricultural leaf area index (the area of green leaf per unit area of ground) is higher than forest leaf area index, which will reduce the effect of decreased stomatal conductance on large-scale transpiration (Betts et al. 1997). Decreased canopy conductance may contribute to a decrease in precipitation in regions where water recycling by vegetation is important—for example, Amazonia (Betts et al. 2004).

13.3.2.2 Boreal Forests

The presence of forests in boreal regions reduces the albedo of the land surface compared with short tundra vegetation. Solar radiation is trapped within the forest canopy, causing warming (Betts and Ball 1997). This effect is particularly accentuated during the snowy season, when short vegetation becomes fully covered with snow, which strongly reflects solar radiation back to the atmosphere (Harding and Pomeroy 1996; Hall et al. 2004). Increased air temperature leads to earlier snow melt. The treeline boundary is limited by temperature, so the relationship between forest and air temperature forms a positive taiga-tundra feedback. This biophysical mechanism plays a substantial role in Earth System dynamics; for example, a reduction of forest cover may have helped trigger the onset of the last glaciation (Gallimore and Kutzbach 1996; de Noblet et al. 1996), while enhanced forest cover has contributed to the regional warming during the mid-Holocene (Foley et al. 1994).

Boreal deforestation leads to spring cooling and extension of the snow season due to albedo changes. During the growing season, trees have a denser, more productive canopy than herbaceous plants, and therefore they transpire more water, cooling surface air (Pielke et al. 1998). This hydrological effect is of primary importance during the summer, so deforestation leads to summer warming. Model results suggest that for deforestation in most boreal forest areas, the cooling effect of albedo changes dominates over the hydrological warming effect on annual average surface temperature (Chalita and Le Treut 1994; Betts 1999; Brovkin et al. 1999). A sea ice–albedo feedback enhances the cooling effect of boreal deforestation (Bonan et al. 1992; Brovkin et al. 2003).

In the recent past, deforestation in the temperate and boreal regions has likely led to a biophysical cooling that partially offset the warming effect of associated CO_2 emissions. In the future, reforestation of regions permanently covered by snow in winter is likely to lead to an increase in global temperature, as the biophysical warming due to albedo changes outweighs the biogeochemical cooling due to uptake of CO_2 in forest stands (Betts 2000; Claussen et al. 2001). This would be counter to the aims of carbon sequestration schemes in these regions. It is also expected that warming at the high northern latitudes will be substantially amplified through the taiga-tundra feedback (Brovkin et al. 2003), although changes in permafrost, forest fire frequency, and outbreaks of pests complicate projections for vegetation cover dynamics in the boreal and polar regions. (See also Chapters 21 and 25.)

13.3.2.3 Sahel/Sahara

In the Sahel region of North Africa, vegetation cover is almost completely controlled by rainfall. The rainy season lasts between two and four months, during which rainfall is highly unpredictable: rainy days can be followed by weeks with no rainfall. When vegetation is present, it quickly recycles this water, as Figure 13.9 illustrated, generally increasing regional precipitation and, in turn, leading to a denser vegetation canopy (Dickinson 1992; Xue and Shukla 1993). This positive feedback between vegetation cover and precipitation amplifies rainfall variability in the Sahel (Zeng et al. 1999). Model results suggest that land degradation leads to a substantial reduction in water recycling and may have contributed to the observed trend in rainfall reduction in the region over the last 30 years (Xue et al. 2004). Combating degradation should maintain or restore the water recycling service, increasing precipitation and contributing to human well-being in the region. (See Chapter 22.)

The Sahara Desert, to the north of the Sahel, is another important example of ecosystem-climate interactions. While rainfall there is too low to support much vegetation at present, it was not so in the past. During the mid-Holocene, about 9,000–6,000 years ago, Sahelian vegetation was greatly extended to the north (Prentice and Jolly 2000). Changes in Earth's orbit increased summer insolation, enhancing monsoon circulation and increasing moisture inflow into the region, and this effect was greatly enhanced by the reduced albedo of the vegetation itself (Kutzbach et al. 1996; Braconnot et al. 1999; Claussen et al. 1999; Joussaume et al. 1999). A rather abrupt collapse in west Saharan rainfall and vegetation cover occurred about 5,500 years ago (deMenocal et al. 2000). This abrupt change is consistent with the existence of alternative stable states in the climate-vegetation system (Claussen 1998; Brovkin et al. 1998). (See Figure 13.10.)

The Sahara desert today differs from other sub-tropical deserts in its exceptionally high albedo. Net cooling of the atmosphere leads to a horizontal temperature gradient and induces a sinking motion of dry air, suppressing rainfall over the region (Charney 1975). Low precipitation reduces the vegetation cover, increasing bare ground with high albedo. This positive feedback maintains desert conditions. On the other hand, if precipitation increases there is more vegetation, the albedo is lower, surface temperature

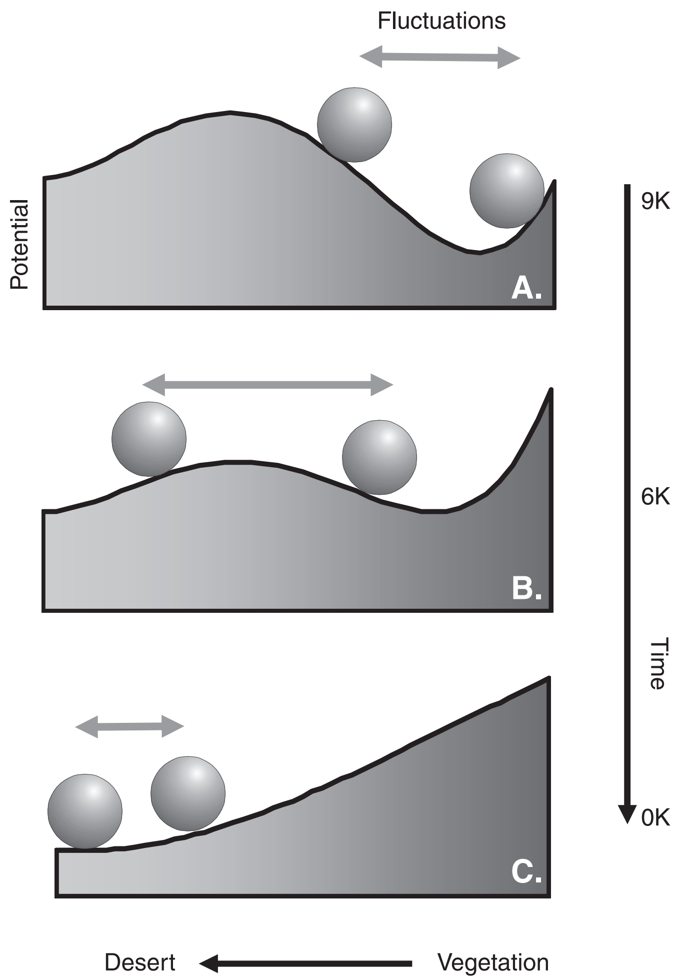

Figure 13.10. Changes in Stability of the Climate-Vegetation System in the Sahara/Sahel Region According to Climate Model Simulations of the Last 9,000 Years (Renssen et al. 2003). Hypothetically, strong positive feedbacks between precipitation and vegetation in the Sahara/Sahel region can lead to the existence of two steady states, desert and "green" (savanna-like). Balls and arrows in the figure indicate the maximum range of the fluctuations induced by large-scale atmospheric and oceanic variability. A) About 9,000 years ago, the system fluctuated in the vicinity of the green state, while the desert state with a shallow potential minimum was also stable. B) About 6,000 years ago, the potential became equal for both states and the system fluctuated between desert and green states. C) Later, the green state lost stability. Desert is the only steady state at present, and precipitation variability is reduced in comparison with the two-state system (A and B).

is higher, and the gradient in temperature between the land and ocean increases—amplifying monsoon circulation and upward air motion over the desert, resulting in increased summer rainfall.

A shift from "green" to "desert" state and vice versa is strongly influenced by externally induced fluctuations in rainfall (Wang and Eltahir 2000; Renssen et al. 2003). In the future, global warming may increase moisture in the Sahara/Sahel region (Brovkin et al. 1998), but it is unlikely that Sahara greening will reach mid-Holocene levels (Claussen et al. 2003).

13.3.2.4 Wetlands

Evapotranspiration of water from wetlands during the day leaves the air above the surface heavy with water vapor, preventing a loss of energy. In sub-tropical regions this is often sufficient to hold night temperatures above freezing point. When wetlands are drained, this ecosystem service is lost. Marshall et al. (2003) showed that the likelihood of agriculturally damaging freezes in southern Florida has increased as a result of the conversion of natural wetlands to agriculture. In January 1997, a rare freeze in southern Florida caused losses to vegetable and sugarcane crops that exceeded $300 million.

13.3.2.5 Cultivated Systems

Intensified agriculture generally increases leaf area index because crops are bred for maximum ability to intercept light. Therefore the effects of past extension of agriculture may, to a certain extent, be reversed by intensification, with increased leaf area during the summer season being a move back toward denser vegetation with characteristics closer to a natural forested state (Gregory et al. 2002). Where seasonal crops replace evergreen vegetation, this will not be the case.

13.4 Effects of Ecosystem on Air Quality

Ecosystems effect the concentrations of many atmospheric compounds that have a direct deleterious effect (for example, pollution) or a beneficial effect (for example, fertilization) on human well-being. Ecosystems are often both sources and sinks for various trace gases that undergo complex atmospheric reactions, simultaneously affecting several aspects of air quality in different ways. It is therefore often hard to quantify the current net effect of ecosystems or of ecosystem change on a particular aspect of air quality. This section concentrates on main net effects, complementing the summary information for each atmospheric constituent that was presented in Table 13.1.

13.4.1 OH and Atmospheric Cleansing Capacity

Some reactive gases released or absorbed by ecosystems are involved in maintaining the ability of the atmosphere to cleanse itself of pollutants through oxidation reactions that involve the hydroxyl radical. (See Box 13.3.) The reactions are complex but, generally, emissions of NO_x and hydrocarbons from biomass burning increase tropospheric ozone and OH concentrations. CH_4 and CO are removed by OH, so emissions of these gases from wetlands, agriculture, and biomass burning decrease OH concentration. Net effects of deforestation are uncertain.

A proxy for estimating atmospheric cleansing capacity (or tropospheric oxidizing capacity) is the concentration of OH, but this is hard to measure directly because of its short lifetime (on the order of seconds). Model estimates, based on measurements of compounds destroyed by OH, indicate a decline in OH concentration since preindustrial times, but the change has probably been less than 10% (Prinn et al. 2001; Jöckel et al. 2003). There is some concern that the pursuit of a hydrogen energy economy could further reduce oxidizing capacity (Schulze et al. 2003; Warwick et al. 2004). A threshold-dependent collapse in atmospheric cleansing capacity would have major implications for air quality (Brasseur et al. 1999). The fundamental importance of tropospheric oxidizing capacity to air quality and atmospheric chemistry means that improving the understanding of OH in the atmosphere, and of the role of ecosystems in regulating OH, is a focus for intensive scientific research.

13.4.2 Pollution Sinks

13.4.2.1 CO_2 and Ocean Acidification

Increased atmospheric CO_2 concentrations are having adverse effects on certain ocean ecosystems, and since terrestrial ecosystems

BOX 13.3

Atmospheric Cleansing (Tropospheric Oxidizing Capacity)

The ability of the atmosphere to cleanse itself of the many compounds emitted from biological and anthropogenic sources is becoming a key issue. Conversion and removal of a large number of environmentally important atmospheric compounds requires one or more chemical reactions involving oxidation taking place in the troposphere (the lower 9–16 kilometers of the atmosphere). The hydroxyl radical is the main oxidizing sink and is often referred to as the "atmospheric detergent." Compounds affected include:

- CH_4 converted to CO;
- removal of NO_x by direct uptake and conversion to HNO_3;
- conversion of CO to CO_2;
- conversion of SO_2 to H_2SO_4;
- removal of some gases responsible for stratospheric ozone depletion, such as methyl bromide; and
- conversion of DMS to SO_2

OH is formed when tropospheric ozone is broken down by UV light to release oxygen atoms that then react with water vapor. Ecosystem emissions of tropospheric O_3 precursors (NO_x, VOCs, CH_4, CO) contribute to OH formation. Ecosystem emissions of compounds oxidized by OH (such as CH_4, CO, hydrocarbons) contribute to the destruction of OH.

are currently a net sink for CO_2 (as described earlier), this could be considered an ecosystem service of benefit to human well-being. The component of anthropogenic CO_2 from fossil fuels and land use change that has dissolved in the ocean has increased the acidity of the ocean to a degree that is unprecedented in recent geological history (Caldeira and Wickett 2003). The global mean surface ocean pH has decreased by 0.12 units since preindustrial times. A further decrease of 0.25 units will occur during this century if atmospheric CO_2 concentration rises to 750 ppmv (Wolf-Gladrow et al. 1999).

Increased acidity changes carbonate chemistry in the surface ocean, with negative impacts on ocean organisms such as corals, coccolithophores, and foraminifera that make their bodies from calcium carbonate. The rate of calcification (growth) of the organisms will decrease (Riebesell et al. 2001), with possible consequences for ecosystem services such as tourism and fish production (Guinotte et al. 2003). A rise in the atmospheric CO_2 concentration to double the preindustrial level may reduce global calcification by as much as 30% (Kleypas et al. 1999). The negative feedback of declining growth of calcifying organisms on the atmospheric CO_2 increase is likely very small (Heinze 2004).

13.4.2.2 Tropospheric Ozone

Sources, sinks, and trends in tropospheric ozone concentration were described earlier in this chapter. Tropospheric ozone can have adverse effects even at relatively low concentrations. Harmful concentrations occur in urban areas but also in the region of vegetation fire events and of high NO_x emissions from fertilizer use, particularly when atmospheric conditions trap air, such as in valleys or temperature inversions. While ecosystems in some regions are a source of ozone precursors, globally they are a net sink. Ozone is destroyed by reacting with plant tissues.

Adverse human health effects of tropospheric ozone include impacts on pulmonary and respiratory function and the aggravation of pre-existing respiratory diseases such as asthma, resulting in increased excess mortality (Thurston and Ito 1999; WHO 2000). Current tropospheric ozone concentrations in Europe and North America cause visible leaf injury and reduced yield of some crops and trees (Mclaughlin and Percy 1999; Braun et al. 1999; Mclaughlin and Downing 1995; Ollinger et al. 1997). It is estimated that for wheat there is a 30% yield reduction for a seasonal seven-hour ozone daily mean of 80 nanomoles per mole, a concentration level that has been found in parts of the United States and Europe (Fuehrer 1996). Economic losses in the Untied States alone may amount to several billion dollars annually. Fowler et al. (1999) estimate that the proportion of global forests exposed to potentially damaging ozone concentrations will increase from about 25% in 1990 to about 50% in 2100.

13.4.2.3 Acid Rain and Acid Regulation

Acid deposition of SO_2, sulfate, NO_x, and nitrate has increased mainly due to industrial emissions (Satake et al. 2000; Rodhe et al. 1995). Biomass burning emissions also affect the chemical composition of rainfall over large areas, while soil emissions of NO_x are as important as industrial emissions in tropical areas, with large increases driven by increasing fertilizer use. Ecosystems are also a sink for NO_x and sulfur compounds. The net effect of ecosystems is probably a sink for compounds that contribute to acid rain. Industrial sources of acid deposition have caused damage to sensitive ecosystems, especially in northern Europe and parts of northeastern North America (Emberson et al. 2003; see also MA *Scenarios,* Chapter 7). Industrial emissions are now declining in these areas due to pollution control, but they continue to rise in other areas, such as Southeast Asia.

Ammonia is the only important gaseous alkaline component in the atmosphere. NH_3 neutralizes many of the acid compounds emitted, forming sulfate and nitrate aerosols. NH_3 in high concentrations is harmful to human health, causes eutrophication of lakes, and can also contribute to acidification in N-saturated ecosystems. Although the quantity of NH_3 volatilized from fertilized fields and animal feedlots can be extremely large (Bouwman et al. 1997), ecosystems as a whole are a net sink for NH_3 (Dentener and Crutzen 1994; Holland et al. 1997).

13.4.2.4 Stratospheric Ozone

Decreases in stratospheric ozone and subsequent increases in ultraviolet (UV-B) radiation have adverse effects on human and animal health, plant growth and mortality, and marine organisms. The net effect of ecosystems on stratospheric ozone is very small compared with industrial emissions, but the effect is probably to reduce the quantity of stratospheric ozone-destroying compounds. As with acid rain, compounds that destroy stratospheric ozone have greatly increased as a result of industrial activities. There are natural ecosystem sources of methyl halides (marine ecosystems and biomass burning), and ecosystems are the primary source of N_2O, which also destroys stratospheric ozone when oxidized in the atmosphere. However, ecosystems are also sinks for halocarbons. As industrial emissions have been controlled by the Montreal Protocol and fallen dramatically, concentrations of most ozone-depleting gases in the atmosphere are now at or near a peak (WMO 2003), while concentrations of N_2O continue to rise.

13.4.3 Pollution Sources

13.4.3.1 Biomass Burning—Vegetation Fires

Vegetation fires are a common natural phenomenon in many regions and vegetation types. Fires maintain vegetation diversity, productivity, and nutrient cycling (although too-frequent fire can lead to impoverishment of nutrients). Fire is also a common land management tool, particularly in the tropics, where it is used to clear land (shifting agriculture, for example, or disposing of residues) (see Box 13.4), to maintain grasslands for cattle grazing, to prevent encroachment of weeds, and to prevent destructive canopy fires and catastrophic wildfires. Prevention of uncontrolled wildfire is an integral component of land use policies: balancing the benefits of controlled use of fire with minimizing the many adverse effects of uncontrolled fires. Several studies have suggested increased fire risk during the twenty-first century (e.g., Flannigan and van Wagner 1991; Price and Rind 1994; Stocks et al. 1998; Mouillot et al. 2002).

Combustion of plant biomass in vegetation fires (or as an energy source, see Chapter 9) produces a mixture of compounds, including greenhouse gases, toxic pollutants, and reactive gases. Toxic pollutants are mostly the result of incomplete combustion; biomass burning is often inefficient, varying with biomass type and load, fire intensity, and weather conditions. Air pollution from biomass burning is associated with a broad spectrum of acute and chronic health effects (Schwela et al. 1999; WHO 2002; Brasseur et al. 2003b). Emissions associated with burning biomass for energy (mostly fuelwood and dung) have been linked to high levels of indoor pollution and major health effects (see Table 13.3); we do not consider this an ecosystem source, however, but rather one arising from use of ecosystem resources. Estimates of the emissions of biomass combustion products in vegetation fires are subject to large uncertainties because of the difficulties inherent in estimating the amount of biomass burned (Andreae and Merlet 2001).

13.4.3.1.1 Particulates

Particles small enough to be inhaled into the lungs, typically less than 10 micrometers (PM10), are associated with the most serious effects on humans, including respiratory disease, bronchitis, reduced lung function, lung cancer, and other cardiopulmonary sources of mortality and morbidity. Studies do not show threshold concentrations below which effects are not observed (WHO 2000, 2002). Particulate levels in plumes associated with large-scale tropical fires can be 2 to 15 times those observed in urban areas (Brauer and Hisham-Hashim 1998), although exposure levels are generally less than those from indoor air pollution.

During the peak of the burning season, the number of particles in the air is an order of magnitude higher than during the rest of the year. Solar radiation reaching the surface is reduced by about 10–30%, lowering both the surface temperature and the light available for plant growth. In China, the effect of atmospheric aerosols and regional haze from all pollution sources is reducing wheat yields by 5–30% (Chameides et al. 1999). In some circumstances, for example in cloudless conditions in areas with high solar radiation (Cohan et al. 2002), plant production could be increased by haze cover. Particulates tend to reduce rainfall but increase the likelihood of intense storms (Andreae et al. 2004). Rainfall reduction has a positive feedback effect, making further fires more likely in, for example, the Amazon (Koren and Kaufman, 2004), but intense rains storms can contribute to putting out fires.

> **BOX 13.4**
>
> **Case Study: Biomass Burning in and around the Amazon Basin during the Dry Season**
>
> Farmers prepare their crop and pastureland by burning to clear weeds; developers burn the forest to open new farmland. The alternative to burning crop and pasture land is to use herbicides or machinery to cut and mulch the weeds. Herbicides are expensive for the local farmer. The machinery is expensive to buy and operate (but could be bought by cooperatives). Burning is cheap and effective in the short term (two to three years), although it becomes costly to the crop yield in the medium term (three to five years). In areas with poor soils, open areas have been abandoned and a secondary forest is growing back. In areas with less poor soils, fertilizers have been introduced to compensate for the lack of nutrients. In a few places the culture of not burning is beginning to be adopted by more educated farmers.

13.4.3.1.2 Tropospheric ozone

Biomass burning emits tropospheric ozone precursors (VOCs, NO_x, CH_4, and CO). The impacts of tropospheric ozone were described earlier. High levels of tropospheric ozone can accumulate at a regional scale during the biomass-burning season (e.g., Swap et al. 2003). The interannual trends and seasonal cycle of tropospheric ozone concentrations correspond to the seasonal cycle and extent of biomass burning in tropical Africa, Latin America, and Asia (Thompson and Hudson 1999). Ozone concentrations can reach values that are larger than those observed in the most polluted urban areas of the world. Values of 100–120 ppb have been measured during the dry season in the southwest Amazon Basin; background values during the rainy season are around 10–15 ppb (Kirchhoff et al. 1996; Cordova et al. 2004).

13.4.3.1.3 Carbon monoxide

Potential health effects of CO include hypoxia, neurological deficits, neurobehavioral changes, and increases in daily mortality and cardiovascular diseases. CO toxicity is mostly associated with indoor pollution from biomass fuel burning, but some studies show effects of CO even at very low concentrations (WHO 2000; Schwela et al. 1999). Dangerous levels are only occasionally observed during vegetation fires, although fatalities caused by excessive carbon monoxide concentrations alone or in combination with other pollutants have been reported, as in China in 1987 (Schwela et al. 1999).

13.4.3.1.4 Other compounds

Volatile organic compounds, including benzene, toluene and xylene, and polynuclear aromatic hydrocarbons, have been identified in fire smoke plumes (Muraleedharan et al. 2000; Radojevic 2003) and are known or suspected carcinogens, mutagens, and teratogens with the potential to cause serious long-term effects. Volatized heavy metals can also pollute the environment. In 1992, severe wildfires spread into the 30-kilometer exclusion zone around the Chernobyl Power Plant in Belarus, burning radioactively contaminated vegetation and increasing the level of radioactive cesium in aerosols 10 times (Dusha-Gudym 1996).

13.4.3.2 Nitrogen Pollution

Elevated deposition of nitrogen compounds is driven by emissions of various N gases. Ecosystem sources include increased fertilizer use, rice paddies, ruminant animals, and biomass burning. Ecosystem N deposition adds to the burden of nitrate leaching to groundwater due to fertilizer use in agricultural ecosystems, causing changes in the functioning and stability of many sensitive ecosystems (for example, heathlands and bogs), particularly in the most populated parts of the world. In heavily affected areas (the Netherlands, for instance), the drinking water standard for nitrate has been exceeded, and lakes and coastal waters suffer from eutrophication. Episodes of high wet deposition of nitrogen are suspected of occasionally affecting even remote marine ecosystems. Changes in the nitrogen cycle and its impacts are dealt with in detail in Chapter 12.

13.4.3.3 Dust Pollution

Agricultural intensification coupled with increasing population density and climate variability in many areas of Sahelian Africa have led to soil degradation and greater soil exposure to wind erosion, increasing the sources of atmospheric dust in recent years (N'Tchayi 1994; Nicholson 1998). An increase in local dust storms is widely considered to be related to ill health (fever, coughing, sore eyes) and has been implicated in meningococcal meningitis epidemics in the region (Molesworth 2002). Dust emanating from this region and the Sahara has been implicated in respiratory problems as far away as the United States (Prospero 2001). Dust storms cause a strong reduction in visibility, resulting in serious disruptions in ground and air traffic. These conditions not only occur in the dust source regions themselves, they can also be problematic downwind. For example, visibility in Beijing is often adversely affected by dust storms originating in the Gobi Desert in springtime (Sugimoto et al. 2003; Shimuzu et al. 2004).

The major dust source regions are deserts, but semiarid regions where vegetation is sparse and soil surfaces disturbed by human activities also contribute to the atmospheric burden of dust. Soil disturbance has been estimated to account for up to 10% of total dust emissions (Tegen et al. 2004). Long-term increases in dust over the Atlantic are possibly associated with desertification of northern Africa (Prospero and Lamb 2003). There are many land use and climate drivers and feedbacks that are likely to affect dust production in the future, with climate drivers dominating. While both the magnitude and the direction of change in dust are uncertain, some models suggest that dust production could decline in a warming world due to increased vegetation in arid and semiarid regions (Mahowald and Luo 2003; Tegen et al. 2004).

13.4.4 Fertilizing Effects

Carbon dioxide, nitrogen gases (NO_x, NH_3), and nutrients in dust particles can all have fertilizing effects on terrestrial plants, potentially increasing production of services such as food and timber. Fertilization from these sources is one possible contributory mechanism for the increasing terrestrial CO_2 sink in recent years. Estimates of the magnitude of this effect in the past and future are limited by incomplete understanding of soil carbon dynamics, plant nutrient relationships, and plant physiology (Oren 2001; Finzi et al. 2002; Hungate et al. 2003; Zak et al. 2003; Norby et al. 2004; Nowak et al. 2004). Nitrogen and dust deposition on the ocean also have the potential to increase phytoplankton production, and the supply of iron in dust is thought to be a major control on the strength of biological carbon uptake in the ocean (Martin et al. 1991). Ecosystems are currently a net sink for CO_2 and a net source of NO_x, while vegetation cover reduces dust emissions (as described earlier in this chapter).

13.4.4.1 Carbon Dioxide

Carbon dioxide in the atmosphere and ocean is necessary to support plant photosynthesis. Most marine plants are not limited by CO_2 but by nutrients and light. On the other hand, most terrestrial plants are limited to some extent by CO_2 supply, and thus rising atmospheric concentrations can have a fertilizing effect, enhancing productivity both directly (Farquhar et al. 1980) and indirectly though stomatal closure and improvements in water use efficiency (Drake et al. 1997; Farquhar 1997; Körner 2000). The strength of the response in terrestrial plants depends on the photosynthetic pathway. Theoretically, those with a C_3 pathway (trees, cold climate plants, most nontropical grasses, and most agricultural crops, including wheat and rice) respond more strongly than those with a C_4 pathway (most tropical grasses, some desert shrubs, and some crops, including maize and sugarcane), although field experiments suggest a more complex picture (Owensby et al. 1993; Polley et al. 1996; Porter and Navas 2003; Nowak et al. 2004).

Experimental doubling of CO_2 in Free-Air CO_2 Enrichment systems produces an average aboveground biomass increase of 17% for C_3 and C_4 agricultural crops and a 20% increase in agricultural yield under conditions of ample N and water, but there is a wide range of responses among individual studies (Kimball et al. 2002). Increases are generally greater in conditions of low water availability. Trees in open top chambers have shown an enhancement of the annual wood mass increment of about 27% (Norby et al. 1999). This strong response of trees to elevated CO_2 has been confirmed in Free-Air CO_2 Enrichment experiments in young forest plantations (DeLucia et al. 1999; Hamilton et al. 2002; Nowak et al. 2004). However, the response of mature forests may be different from that of young forests for various reasons (Norby et al. 1999; Curtis and Wang 1998).

13.4.4.2 Nitrogen Deposition

Nitrogen limitation to plant production is widespread. (See Chapter 12.) There has been a rapid increase in reactive N deposition over the past 50 years (Vitousek et al. 1997; Holland et al. 1999). There is much field evidence that N deposition increases NPP (e.g., Chapin 1980; Vitousek and Howarth 1991; Bergh et al. 1999; Spieker et al. 1996) and soil carbon storage (Fog 1988; Bryant et al. 1998). When the nitrogen saturation limit is reached, as is thought to have happened in highly polluted areas of Europe, plants can no longer process the additional nitrogen and may suffer from deleterious effects of associated pollution (Shulze et al. 1989; Aber et al. 1998; see also Chapter 7). N addition leads to changes in plant species composition and an overall reduction in diversity. In general, the impacts are most pronounced in nutrient-poor systems, where N deposition enhances growth of the most responsive species, which often outcompete and eliminate rare species that occupy N-deficient habitats (Mooney et al. 1999).

13.4.4.3 Dust as a Fertilizer

Nutrients in dust particles (especially phosphate and iron) can act as fertilizers when deposited on oceans and land (Piketh et al. 2000). Fertilization of the ocean from expanded desert sources, and the resulting increase in ocean ecosystem CO_2 uptake, is thought to be one of the drivers of change in glacial-interglacial atmospheric CO_2 concentrations (Watson et al. 2000; Ridgwell et al. 2002; Bopp et al. 2003). Ice core records indicate decreased dust input in the ocean may account for up to about a quarter of the 80-ppm atmospheric CO_2 increase at the last glacial-interglacial transition (Rothlisberger et al. 2004). A recent synthesis (Piketh et al. 2000) suggests that aerosols derived from the southern African continent are increasing carbon uptake downwind in the Indian Ocean. Changes in dust sources, transport, and deliberate iron fertilization of the ocean could affect marine productivity, future CO_2 uptake (Mahowald and Luo 2003; Tegen et al. 2004), and marine N fixation and N_2O release (Denman et al. 1996), but the magnitude and associated impacts are uncertain. (Sources and drivers of dust are described earlier; see also Chapter 22.)

13.5 Climate Variability and Climate Feedbacks

13.5.1 Interannual Variability

Biological activities are dependent on climatic conditions, and therefore biogenic emissions of gases often show a seasonal cycle and interannual climate variability linked to natural climate variations. In some cases the interannual variability is large—for example, when biological processes are limited by water availability. This variability in turn affects biological sources and sinks of atmospheric compounds and biophysical properties of vegetation. For instance, during recent El Niño events atmospheric CO_2 increase has doubled or tripled compared with other times, partly due to reductions in land uptake caused by the effects of high temperatures, drought, and fire on terrestrial ecosystems in the tropics (Tian et al. 1998; Clark et al. 2004).

Warming enhances emissions of VOCs (which increase by 20–30% per degree Celsius) (Guenther et al. 2000), CH_4, and N gases, which generally tend to increase tropospheric ozone and OH (although these reactions also depend on water and light availability). Changes in OH concentration further affect the variability in concentration of some atmospheric compounds such as CH_4. Effects of interannual variability on biogenic sources and sinks are one of a number of processes that affect CH_4 and tropospheric ozone (Dlugokencky et al. 1998; Warwick et al. 2002; Sudo and Takahashi 2001).

Enhanced biomass associated with the La Niña phase of the El Niño-Southern Oscillation, followed by droughts in the El Niño phase, produce above average biomass burning emissions from savannas in southern Africa (Swap et al. 2003). In 1997–98, fires associated with an exceptional drought caused by ENSO devastated large areas of tropical rain forests worldwide (Siegert et al. 2001). Emissions of associated gases such as CO_2, CO, CH_4, and other trace gases have been correlated with large biomass burning events in tropical and boreal regions (Langenfelds et al. 2002).

13.5.2 Climate Feedbacks on Ecosystems, Climate, and Air Quality

Changes in global and regional climate, partially bought about by ecosystem change, can in turn lead to further changes in ecosystem sources and sinks of gases and biophysical properties. Temperature and moisture changes will cause a variety of changes in sources, sinks, and chemical reactions in the atmosphere, the net effect of which is uncertain and may differ from place to place. The predominant climate feedbacks operate though changes in the carbon cycle and CO_2 emissions, with a strong positive feedback predicted under future climate change. Methane emissions from wetlands are expected to increase under some conditions (such as permafrost melting) and to decrease in others (such as the drying of northern and tropical soils), as described earlier. Nitrous oxide emissions are generally higher in wetter soils. Increased emissions of tropospheric ozone precursors, NO_x and VOCs, occur under warmer conditions.

Where climate change causes shifts in vegetation there will be regional biophysical effects—for example, due to the northward shift of boreal forests and enhanced vegetation cover in the Sahara (Brovkin et al. 2003; Claussen et al. 2003). Vegetation loss could lead to positive climate feedbacks though biophysical effects—for instance, feedback on local drying from Amazon dieback (Betts et al. 2004)—and could lead to increased dust emissions. Changes in water availability affect transpiration, with drought reducing water recycling and rainfall.

On the land, warming increases the rate of decomposition of soil organic matter, thereby reducing carbon storage in soil. Although soil warming experiments have shown an increased rate of decomposition for the first one to three years only (Jarvis and Linder 2000; Oechel et al. 2000; Luo et al. 2001; Rustad et al. 2001), this represents the burning-off of the labile (easily decomposed) component only. The larger pool of more chemically stable soil organic matter is potentially vulnerable to warming over longer time scales (Cramer et al. 2001; Joos et al. 2001; Knorr et al. 2005). The effect of global warming on vegetation cover is highly uncertain but likely also to affect atmospheric CO_2. One coupled climate-carbon cycle model has predicted a dieback of tropical forests in South America, which, along with increased soil organic matter decomposition and subsequent carbon loss, would lead to an additional 200 ppm increase in atmospheric CO_2 (Cox et al. 2000). Another model predicted a smaller feedback (Friedlingstein et al. 2001).

In the oceans, sea surface temperature increase and changes in the global water cycle tend to increase vertical stratification (layering) and to slow down global ocean circulation. Warming reduces the solubility of CO_2 in the ocean. Stratification slows the mixing into deep layers of excess carbon in the surface water. Stratification further reduces nutrient input into the surface zone and leads to a prolonged residence time of phytoplankton at the surface, near light. Models indicate the net effect is reduced phytoplankton productivity (Bopp et al. 2001; Joos et al 1999). Models estimate that the combined effect of warming and circulation changes on ocean physics and biology will reduce the oceanic CO_2 uptake by 6–25% in 1990–2050, thus providing a positive climate feedback (Maier-Reimer et al. 1996; Sarmiento et al. 1998; Matear and Hirst 1999; Joos et al. 1999, Bopp et al. 2001; Plattner et al. 2001).

Changes in ocean circulation, pH, and temperature are also likely to have additional effects on ocean biology that have not been quantified in these models and that may induce further CO_2 feedbacks. These include changes in the community structure, net production, and bio-calcification. The effect of bio-calcification is estimated to increase the ocean carbon sink by less than 2.5% (Riebesell et al. 2001). The quality and magnitude of biological changes will vary over space and time and is highly uncertain. While the combined inorganic and biological changes tend to reduce global uptake of anthropogenic carbon, the global net effect on carbon uptake of the ocean biological changes alone is unknown. Altered size and timing of phytoplankton blooms can also potentially reduce fish production (Chavez et al. 2003; Beaugrand et al. 2002; Platt et al. 2003).

13.6 Impacts of Changes in Climate and Air Quality on Human Well-being

13.6.1 Impacts of Changes in Climate on Human Well-being

According to the IPCC (2001d), "The earth's climate system has demonstrably changed on both global and regional scales since the pre-industrial era, with some of these changes attributable to human activities. . . . Projected climate change will have beneficial and adverse effects on both environmental and socio-economic systems, but the larger the change and rate of change in climate, the more the adverse effects dominate." Changes in climate are linked to all aspects of human well-being as defined by the Millennium Ecosystem Assessment (MA 2003). This section provides a summary of the detailed results presented by IPCC (2001a, 2001b, 2001c, 2001d) unless otherwise stated.

13.6.1.1 Security

An increase in frequency and severity of floods and droughts has been noted in some areas. A fourfold increases in economic losses for catastrophic weather events from the 1980s to the 1990s (average annual global loss $40 billion in the 1990s) has been partly linked to regional climatic factors and partly to socioeconomic factors. IPCC projections include increasing ecological shocks and stress as well as vulnerability to them, alongside a reduction in the ability to predict and plan for the weather.

13.6.1.2 Health

Many vector-, food-, and water-borne infectious diseases are known to be sensitive to changes in climatic conditions, as are production of spores and pollens and the climatically related production of photochemical air pollutants. Floods increase risk of drowning, diarrhea, respiratory diseases, water-contamination diseases, hunger, and malnutrition. Heat waves in Europe and America have been associated with a significant increase in urban mortality. For example, during the European heat wave of 2003, almost 15,000 additional deaths were estimated to have occurred in France, mostly in elderly people (WHO 2004). Warmer wintertime temperatures can also result in reduced wintertime mortality in cold climates.

Indirect climate effects on human health include changes in water quality, air quality, food availability and quality, population displacement, and economic disruption. Poor understanding of the role of socioeconomic and technological factors in shaping and mitigating health impacts, and the difficulty in separating climate variability impacts from climate change impacts, means that current estimates of the potential health impacts of global warming are based on models with *medium* to *low certainty*.

The World Health Organization (WHO 2002) has estimated that global warming was responsible in 2000 for approximately 2.4% of worldwide diarrhea, 6% of malaria in some middle-income countries, and 7% of dengue fever in some industrial countries. (See Table 13.3.) These factors contribute to the estimated mortality of 154,000 deaths and 5.5 million disability-adjusted life years, mostly in Southeast Asia and Africa. (Such estimates are of high uncertainty, however, due to the difficulties in establishing direct causality.)

Overall, global warming is projected to increase threats to human health, particularly in lower-income populations predominantly within tropical and sub-tropical countries: thermal stress effects amplified with higher projected temperature increases; expansion of areas of potential transmission of malaria and dengue; greater increases in deaths, injuries, and infections from floods and storms; and water quality degraded by higher temperatures and salinization, with changes modified by changes in water flow volume.

13.6.1.3 Basic Material for a Good Life

The impacts of global warming include changes in species distributions, population sizes, the timing of reproduction or life-cycle events, and the frequency of pest and disease outbreaks (IPCC 2001b, 2001d). Growing season has lengthened by one to four days in the Northern Hemisphere during the last 40 years, with earlier onset of life-cycle events (such as flowering, migration, and breeding). Coral reef bleaching has increased in frequency. (See Chapter 19.)

The productivity of ecological systems is highly sensitive to climate change, and projections of change in productivity range from increases to decreases. Models of cereal crops indicate that in some temperate areas, potential yields increase for small increases in temperature but decrease with larger temperature changes. In most tropical and sub-tropical regions, potential yields are projected to decrease for most projected increases in temperature. An increase in frequency of disturbance by fire and insect pests is projected.

Stratification of the ocean at warmer temperatures may reduce phytoplankton productivity and thus fish production (Platt et al. 2003). A further increase in frequency and extent of coral reef bleaching is projected, along with loss of coastal wetlands and erosion of shorelines. Diversity in ecological systems is expected to be affected by climate change and sea level rise, with an increased risk of extinction of some vulnerable species. Projected climate change would exacerbate water shortages and water-quality problems in many water-scarce areas of the world but would alleviate it in others. Some systems—including coral reefs, glaciers, mangroves, boreal and tropical forests, polar and alpine systems, prairie wetlands, and temperate native grasslands—are particularly vulnerable to climate change because of limited adaptive capacity and may undergo significant and irreversible damage (IPCC 2001b, 2001d).

13.6.1.4 Good Social Relations

The impacts just described may compound the risk of conflict over natural resources.

Tropical and dryland regions are likely to incur more detrimental impacts than temperate and cold regions (IPCC 2001b, 2001d). People in poor countries are most vulnerable due to lower adaptive capacity. Climate change is expected to have negative impacts on development, sustainability, and equity (IPCC 2001b, 2001d; Toth 1999). The aggregated market sector effects are estimated to be negative for many developing countries for all magnitudes of global mean temperature increase studied and are estimated to be mixed for industrial countries for up to a few degrees Celsius warming and negative for warming beyond that point.

The global value of the climate regulation services of ecosystems was estimated by Costanza et al. (1997) to be $2 trillion per year, of which $800 billion was attributed to the biological role of ecosystems, principally carbon storage in forests and changes in greenhouse gas emissions and albedo from converting grasslands to agriculture. The remainder was due to nonbiological oceanic uptake of CO_2. This global value is a synthesis of published estimates of ecosystem service values for several different biomes, using a range of valuation techniques. Extrapolating from the biome values to a global aggregate is likely to underestimate the true total value because these are partial valuations in several ways. First of all, not all biomes were represented in the available literature. Second, not all processes (biochemical and biophysical) and feedbacks that generate ecosystem climate services were considered. For example, increasing loss of forests might alter other ecosystems so dramatically as to change their function in the carbon cycle, such as altering temperature in the oceans and net ocean uptake of CO_2.

Damages from reductions in carbon sequestration capacity may be nonlinear, with damages increasing more than proportionally to forest loss. The unit demand for an ecosystem service is likely to increase rapidly as its supply diminishes; in other words, there is reason to expect that the marginal value of forests for climate control may increase with forest loss. In this case, ag-

Table 13.3. Attributable Mortality and Disability-Adjusted Life Years from Environmental Risk Factors, 2000.[a] The risk factors and measured adverse outcomes of exposure are as follows: unsafe water, sanitation, and hygiene–diarrhea; urban air pollution–cardiovascular mortality, respiratory mortality, lung cancer, mortality from acute respiratory infections in children; indoor smoke from solid fuels–acute respiratory infections in children, chronic obstructive pulmonary disease, lung cancer; climate change–diarrhea, flood injury, malaria, malnutrition, dengue fever, cardiovascular mortality, population movement. (WHO 2002)

	World	Africa	North America[b]	South and Central America	Eastern Mediterranean	Europe	Southeast Asia	Western Pacific
					(thousand)			
Mortality								
Unsafe water, sanitation, and hygiene	1,730	608	1	54	270	18	699	77
Urban air pollution	799	32	28	35	59	107	164	373
Indoor smoke from solid fuels	1,619	392	0	26	118	21	559	503
Climate change[c]	154	54	0	0	21	0	74	3
DALYs								
Unsafe water, sanitation, and hygiene	54,158	18,636	61	2,045	8,932	736	19,727	4,018
Urban air pollution	7,865	485	200	360	727	859	1,852	3,386
Indoor smoke from solid fuels	38,539	12318	6	773	3,572	544	15,227	6,097
Climate change[c]	5,517	1,893	3	94	768	17	2,572	170

[a] Uncertainty ranges (range of coefficient of variation): water and indoor air pollution 0 to 4.9; urban air pollution 10 to 14.9; climate change >15.

[b] North America: United States, Canada, and Cuba.

[c] Climate change impacts are modeled effects on disease, flood risk, and food production for modeled climate in year 2000 compared with mean climate in 1961–90.

gregating the marginal valuation methods used may underestimate the economic value of total forest climate control services. While the direct use of the Costanza et al. (1997) service values is problematic in many policy spheres, which need the marginal values, the review by Balmford et al. (2002) of the relative values of intact and human-modified ecosystems suggests that in general terms, the loss of nonmarketed services associated with ecosystem loss or conversion frequently exceeds the (marketed) benefits.

13.6.2 Impacts of Changes in Air Quality on Human Well-being

Impacts of air pollution on human health can be dramatic, as exemplified by the ''Asian/atmospheric brown cloud'' and the smoke haze generated by 1997–98 fires in Indonesia. Health effects can also be more subtle and are increasingly widespread. Industrial pollution is not the concern of this chapter, but some of its effects were included in Table 13.3 for illustration. Ecosystem emissions, particularly those resulting from biomass burning, can add to the burden of industrial pollution and affect human well-being in nonurban areas, while ecosystem sinks can reduce the negative impacts of industrial air pollution. Some ecosystem air quality effects described earlier in this chapter are summarized below according to the Millennium Ecosystem Assessment (MA 2003) definition of well-being:

13.6.2.1 Security

Vegetation fires can cause damage to property and life, with effects of smoke on transport and effects of toxic pollutants on health. The health implications of changes in clean air are outlined in a following section.

13.6.2.2 Access to Resources

Some pollutants with ecosystem sources and sinks are deleterious to ecosystem health, affecting production of resources such as food and timber. For example, ecosystems are currently a net sink for tropospheric ozone and compounds that contribute to acid rain as well as for CO_2 (ocean acidification and impacts on marine organisms). Agricultural ecosystems are a net source of nitrogen compounds that contribute to acid rain and eutrophication of lakes, decreasing agricultural and fish production. On the other hand, some of these N compounds have a fertilizing effect, increasing plant productivity up to the point where the ecosystems become saturated with that nutrient.

13.6.2.3 Health: Clean Air

Biomass burning is a source of particulates, tropospheric, ozone and CO, all of which have harmful respiratory effects. Smoke pollution generated by vegetation fires occasionally reaches levels with major public health and economic impacts—usually when wildfires or land management fires get out of hand under extreme weather conditions. Vegetation fires particularly enhance the risk of acute respiratory infections in childhood, a major killer of young children in developing countries, and affect the health of women already exposed to high levels of indoor air pollution (Schwela et al. 1999).

Few epidemiological studies investigate short-term and long-term implications of vegetation fires for human health. The health impacts of burning biomass (mainly fuelwood or dung) as an energy source in indoor cookstoves has been studied in more detail. (See also Chapter 9.) Exposures are far more concentrated and chronic than for vegetation fires, but since many of the compounds emitted are the same, it is useful to note the impacts for comparison. WHO has estimated that 1.6 million deaths and 39 million DALYs worldwide were attributable to indoor biomass burning, with women and children particularly at risk (WHO 2002). Ecosystems are a net sink for tropospheric ozone, reducing the impacts of urban air pollution. Dust adversely affects respiration and is reduced by vegetation cover.

13.6.2.4 Good Social Relations

Ecosystem reduction of air pollution, such as acid rain and ocean acidification, can limit damage to ecosystems valuable for aesthetic, cultural, religious, recreational, or educational purposes. On the other hand, the detrimental impacts of some wildfires on economies, human health, and safety have consequences comparable in severity to other major natural hazards and could lead to transboundary conflicts. In addition to the air quality impacts of fire mentioned already, wildfire can lead to the destruction of ecologically or economically important resources (such as timber and biodiversity), adding to rapid environmental changes and degradation. Smoke plumes can cause visibility problems, resulting in accidents and economic loss including closure of airports and marine traffic.

Fires can be catastrophic in areas that have been long protected from them, allowing a buildup of fuel, and where human settlement has extended into forest areas. In the 1980s and 1990s, the most serious pollution problems were noted in the Amazon Basin and in Southeast Asia. Land use fires and uncontrolled wildfires in Indonesia and neighboring countries in 1991, 1994, and 1997 created regional smog layers that lasted for several weeks. Box 13.5 provides examples of impacts and losses of particular fire events. Advances in satellite data and atmospheric transport models are expected to improve monitoring, evaluation, and early warning systems to prevent fires or manage impacts. (See Chapter 16 for more information on fire impacts other than air quality.)

BOX 13.5

Recent Major Fire Episodes and Losses

- Regional haze episodes caused by forest fires occurred in SE Asia on several occasions during the 1990s (Radojevic 2003). Measurements in Brunei in 1998 during a particularly severe haze episode caused mainly by local fires recorded many compounds, including VOCs (such as benzene and toluene), aldehydes, cresol, phenol, acetic acid, polynuclear aromatic hydrocarbons, heavy metals, and levels of particulates exceeding air quality guidelines (Muraleedharan et al. 2000). In 1994, the fires burning in Indonesia caused the visibility to drop to as low as 500 meters in Singapore. During the 1997 South East Asian smog episode, when particle levels in some areas were up to 15 times higher than normal, the Malaysian government was close to evacuating the 300,000 inhabitants of the city of Kuching (Brauer and Hisham-Hashim 1998), and the loss of an aircraft and 234 human lives in Sumatra was partially attributed to air traffic control problems caused by the smog.
- On the Indonesian islands of Kalimantan and Sumatra during 1997–98, an estimated 9 million hectares of vegetation burned. Some 20 million people in Indonesia alone suffered from respiratory problems, mainly asthma, upper respiratory tract illness, and skin and eye irritation during the episode, with nearly four times as many acute respiratory illnesses as normal reported in South Sumatra (Heil and Goldammer 2001). A first assessment of costs of damages caused by the fire episode on 4–5 million hectares was $4.5 billion (short-term health damages; loss of industrial production, tourism, air, ground and maritime transportation; fishing decline; cloud seeding and fire-fighting costs; losses of agricultural products, and timber; and direct and indirect forest benefits) (EEPSEA 1998).
- The fires burning in Mexico during the 1998 episode forced the local government to shut down industrial production in order to decrease additional industrial pollution during the fire-generated smog. Daily production losses were about $8 million (Schwela et al. 1999).
- In 2002, forest and peat fires in the Moscow region resulted in the worst haze seen in Moscow in 30 years. This has caused severe cardiovascular and respiratory problems among the population of Moscow, especially among children (GFMC 2003).

13.7 Synthesis: Effects of Ecosystem Change and Management on Climate and Air Quality Services

Table 13.4 provides a synthesis of the different biochemical and biophysical effects of each MA ecosystem type on climate and air quality. This section addresses the most pertinent types of ecosystem change and management due to the scale of their impacts or their relevance to the MA.

13.7.1 Changes in Ecosystem Cover and Management

The largest effects of ecosystems on air quality and climate due to human-induced changes in land cover and management have been associated with deforestation, agricultural management (fertilizer use, cattle, and irrigation), and biomass burning. Deforestation and agricultural practices are mainly driven by population growth, urbanization, and economic development and are modified by policies and subsidies. Chapter 7 in the *Scenarios* volume (and Chapter 3 here, more briefly) describes many of these direct and indirect driver and linkages. While industrial countries have been responsible for most of the industrial impacts on climate and air quality in the past, management of tropical ecosystems in particular has played a role and will likely continue to have a significant impact in the future.

13.7.1.1 Forest Cover and Management

Change in forest cover has had a larger impact on global and regional climate than any other ecosystem driver. Deforestation has been a major source of CO_2, only partially offset by reforestation, afforestation, and forest management activities and by the fertilizing effects of N and CO_2. Immediately after deforestation, tropical soils are a source of N_2O, although emissions decline to original level or below after 15–20 years. Tropical forests are also an important source of VOCs; therefore deforestation reduces VOC emissions, although this will depend on the emission rate of the replacement vegetation. Deforestation reduces the sink for tropospheric ozone and N gases (Ganzeveld and Lelieveld 2004).

Forest cover affects albedo, particularly in boreal snow-covered regions. Deforestation increases albedo (cooling). Model results suggest that historical deforestation has led to a cooling of the land surface (Betts 2001; Govindasamy et al. 2001) comparable to the warming caused by CO_2 emissions resulting from the same deforestation (Brovkin et al. 2004), and that this biophysical effect of historical deforestation is necessary to explain the observed climate during the second half of the nineteenth century (Crowley 2000; Bauer et al. 2003). Loss of forest cover profoundly affects the water cycle, reducing water recycling and local rainfall, but the net hydrological effect of deforestation is less certain, especially on a global scale (Rind 1996).

Table 13.4. Summary of Important Ecosystem Fluxes and Biophysical Properties, by Ecosystem Type

Biome	Major Biochemical Impacts	Major Biophysical Impacts
Cultivated systems	CO_2 source: conversion to cropland, management sink: management (e.g., low tillage) CH_4 source: rice paddies, ruminant animals, termites sink: upland soils N_2O source: soils, cattle/feedlots, fertilizer use NO_x source: soils NH_3 source: cattle, feedlots, fertilizer, plants, soils VOCs source: oxygenated VOCs (e.g., methanol, ethanol, acetone) dust source: disturbed soil surfaces and reduced vegetation cover	albedo: increase when forest conversion to cropland, decrease in case of irrigation, decrease where leaf area index higher than natural vegetation transpiration: decrease in case of forest conversion to cropland, increase for irrigated systems
Dryland systems (including savannas and grasslands)	CO_2 source: biomass burning, devegetation, sink: woody encroachment CH_4 source: biomass burning, ruminants, termites sink: upland soils CO source: biomass burning N_2O source: soils NO_x source: soils NH_3 source: plants, animal waste, soils VOCs source: plants, biomass burning S source: biomass burning particulates source: biomass burning tropospheric O_3 source: biomass burning CO source: biomass burning dust source: devegetation, degradation, and erosion	albedo: increase in case of desertification surface runoff: increase in case of desertification

13.7.1.2 Agriculture

Agriculture is a significant source of greenhouse gases, accounting for about 5% of total CO_2 emissions (Rosenberg et al. 1998), about a quarter of CH_4 emissions (rice paddies and ruminant animals) (Praether et al. 2001), and about a third of N_2O emissions (agricultural soils and cattle/feedlots) (Praether et al. 2001). Agricultural management can reduce carbon loss or promote storage to some extent (Lal et al. 2004; Renwick et al. 2004). The use of nitrogen fertilizers profoundly alters the nitrogen cycle, leading to increased emissions of N gases that, in addition to contributing to global warming, contribute to acid rain and eutrophication of lakes, increase the atmospheric cleansing capacity, destroy stratospheric ozone, and may cause respiratory and other health problems. Dust is lost from cultivated and denuded soil surfaces. Agricultural crops often have a higher leaf area index than natural vegetation, reducing albedo. Irrigation increases water recycling, raising latent heat flux and cooling the surface.

13.7.1.3 Wetlands

Wetland draining for agriculture, forestry, or water extraction leads to a decrease in CH_4 production and an increase in CO_2 and N_2O emissions, probably with a net decrease in radiative forcing on short time-frames (20–100 years), but in the longer term the opposite may be true (*low certainty*) (IPCC 2001a; Christensen and Keller 2003). The same will be true for wetlands experiencing drying due to global warming, such as Northern Alaska (now and in the future), and for tropical seasonally flooded areas (if they dry in the future) (*low certainty*). Where climate change has led to loss of permafrost, the net effect is increased CH_4 emissions that will likely continue in the future.

13.7.1.4 Dryland Management and Degradation

Management of drylands to increase vegetation cover and reduce soil erosion increases carbon storage, reduces dust sources, and increases rainfall recycling. Potential impacts are significant given the very large areas involved. Drylands store more carbon in soils than in biomass and are thus more vulnerable to carbon loss through soil erosion, but with good potential for increasing belowground carbon storage (IPCC 2000). (See Chapter 22.)

13.7.1.5 Biomass Burning

Biomass burning is a major source of toxic pollutants, greenhouse gases, and reactive gases—causing major health and visibility problems and contributing to global warming. Greenhouse gases emitted during fires are CO_2, CH_4 (5–10% of all sources), N_2O, and tropospheric ozone precursors: NO_x (just over 10% of all sources), CO (a quarter of all sources), and VOCs. Aerosols from biomass burning have a net cooling effect. Fire suppression reduces emissions from burning and encourages woody plant biomass to increase and act as a carbon sink. (In the United States, for example, this may have amounted to a sink of 0.2 petagrams of carbon per year during the 1980s (Houghton et al. 1999).) However, fire suppression can also increase the risk of future, catastrophic wildfires (Schwela et al. 1999). Pollutants include particulates, precursors of tropospheric ozone, and CO, plus a number of trace gases and compounds (such as polynuclear aromatic com-

Table 13.4. ***continued***

Biome	Major Biochemical Impacts	Major Biophysical Impacts
Forest and woodland systems	CO_2 source: deforestation sink: afforestation, reforestation, forest management CH_4 source: biomass burning, termites sink: upland soils CO source: biomass burning, decomposition N_2O source: soils NO_x source: soils sink: canopy NH_3 source: plants, animal waste, soils VOCs source: plants, biomass burning S source: biomass burning particulates source: biomass burning tropospheric O_3 source: biomass burning CO source: biomass burning	albedo: increase in case of deforestation, decrease due to afforestation transpiration: decrease in case of deforestation, increase for afforestation
Urban systems	CO_2 source: biomass (fuel) burning? CH_4 source: landfill, biogas N_2O source: landfills VOCs source: landfills S source: landfills tropospheric O_3 source: indoor biomass fuel burning CO source: indoor biomass fuel burning particulates: indoor biomass fuel burning	"heat island" effect albedo: increase with expansion and vegetation replacement transpiration: decreases with expansion and vegetation replacement
Inland water systems	CH_4 source: intermittent flooding of vegetation (remineralization)	freshwater incursions to ocean and effects on ocean circulation
Coastal systems	CO_2 sink: biological pump source: upwelling net balance unknown CH_4 source: remineralization N_2O source: denitrification	
Marine systems	CO_2 sink: biological and solubility pumps N_2O source: remineralization CH_4 source: remineralization DMS source: plankton	phytoplankton blooms—reduced albedo (warming sea surface temperatures)
Polar systems	CO_2 source: permafrost melting CH_4 source: permafrost melting	reduced ice cover due to warmer surface and longer growing season—decreased albedo and further warming
Mountain systems	CO_2, CH_4 and N_2O production under snowpack can constitute a significant proportion of the annual trace gas budget	reduced ice cover due to warmer surface and longer growing season—decreased albedo and further warming, shift in treeline
Island systems	Depends on land cover as above, no specific impacts	deforestation changes in wind patterns alters ocean upwelling, warming sea surface temperatures
Wetlands	CH_4 source: anaerobic respiration—decreased by draining CO_2 source: peatland burning, aerobic respiration after draining CO source: peatland burning N_2O source: soils	reduced evapotranspiration in case of draining

pounds, aldehydes, organic acids, sulfur dioxide, and methyl halides (stratospheric ozone depletion)).

13.7.2 Changes in Biodiversity, Invasive Species, and Disease

Change in species diversity in the strict sense is thought not to have a large influence on climate and air quality, although climate, climate change, and air quality conditions have a large influence on biodiversity. (See Chapters 4 and 11.) Changes in the relative abundance of different functional types (such as needle-leaved versus deciduous trees, shrubs versus grasses, and diatoms versus cocolithophorids), however, may have substantial impacts on sources and sinks of gases and on other ecosystem properties (e.g., Riebesell et al. 2001; Scherer-Lorenzen et al. 2005). Loss of biodiversity could further affect the adaptability and resilience of ecosystems and their ability to migrate with changing climate (Schulze and Mooney 1993; Tilman et al. 1997; Nepstad et al. 1999; Loreau et al. 2001; Kim Phat et al. 2004).

Furthermore, the loss of particular species could have a substantial impact on ecosystem functioning. Such ''keystone species'' or ''ecosystem engineers'' (Jones et al. 1994) may not necessarily be identified in advance, which makes preventive mitigation policy difficult. For a review of climate-biodiversity interactions, see Gitay et al. (2002) as well as Chapters 4 and 11. Some examples of drivers of change in functional type abundance are provided here, including climate change, species invasions, and disease.

Encroachment of invasive woody species is generally an additional sink of CO_2, changes biophysical properties (increases LAI, increases transpiration, and reduces albedo), and may reduce biodiversity—for example, Mesquite (*Prosopis* sp.) invasions in Texas in the United States (Archer et al. 2001; Dugas et al. 1996; Gibbens 1996).

The ''fertilizing'' effect of increased CO_2 levels benefit some species (most trees) more than others (such as grasses) (Nowak et al. 2004), giving them a competitive edge. For example, Smith et al. (2000) showed that elevated CO_2 increased the success of an invasive C_4 grass species in the Mojave Desert, potentially reducing biodiversity and altering ecosystem function. As with invasive tree species, this functional shift from grass to trees will also affect biophysical properties, biodiversity, and sources/sinks of various trace gases. Trees are not always the winners; for example, lianas respond more strongly than trees to the fertilizing effects of increased atmospheric CO_2 concentration (Condon et al. 1992; Granados and Koner 2002; Phillips et al. 2002.). Lianas enhance tree mortality and suppress tree growth, which could ultimately reduce carbon storage in forests.

The chestnut blight in the United States around the 1900s caused a switch from chestnut trees, which do not emit isoprene (a VOC involved in tropospheric ozone formation), to oaks, which do, approximately doubling the biomass of isoprene-emitting species (Lerdau and Keller 1997).

Regime shifts of marine pelagic ecosystems, which have occurred in Arctic waters since the mid-1980s, have caused major breakdown in fishery production. Diatom-dominated phytoplankton communities were replaced by extensive coccolithophorid blooms in the Barents Sea and the eastern Bering Sea (Smyth et al. 2004), causing massive changes in ecosystem structure. With coccolithophores being predominant producers of DMS, the observed regime shifts are likely to have altered the sulfur cycle and cloud formation in these areas, affecting air quality and water recycling.

Certain types of marine organisms (calcifiers—coccolithophorids, foaminifera, and corals) form shells of calcium carbonate ($CaCO_3$). This process releases CO_2 to the ambient seawater, countering part of the photosynthetic CO_2 drawdown (that is, the drawdown of CO_2 by coccolithophorids is much smaller than that of non-calcifying phytoplankton). A shift in species composition away from coccolithophorids, for example due to changes in ocean acidity, would increase the ocean's CO_2 storage capacity. On the other hand, biogenic particles containing $CaCO_3$ and SiO_2 sink faster than other particles, which implies that the plankton (coccolithophorids and diatoms) producing these two minerals should increase the drawdown of carbon from surface to depth in the ocean. Thus shifts in composition of the marine ecosystems have the potential to influence the oceanic carbon sink (Francois et al. 2002; Klaas and Archer 2002), but at present we cannot quantify the probability, extent, or direction of the likely future changes or their consequences for climate.

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