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The Vulnerability of the Carbon Cycle in the 21st Century: An Assessment of Carbon-Climate-Human Interactions

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In most scenario calculations to date, emissions from fossil-fuel burning are prescribed, and a carbon cycle model computes the time evolution of atmospheric CO_2 as the residual between emissions and uptake by land and ocean, typically without considering feedbacks of climate on the carbon cycle (see, e.g., Schimel et al. 1996). The global carbon cycle is, however, intimately embedded in the physical climate system and tightly interconnected with human activities. As a consequence, climate, the carbon cycle, and humans are linked in a network of feedbacks, of which only those between the physical climate system and the carbon cycle have been explored so far (Friedlingstein, Chapter 10, this volume). One example of a carbon-climate feedback begins with the modification of climate through increasing atmospheric CO_2 concentration. This modification affects ocean circulation and consequently ocean CO_2 uptake (e.g., Sarmiento et al. 1998; Joos et al. 1999; Matear and Hirst 1999). Similar feedbacks occur on land. For example, rising temperatures lead to higher soil respiration rates, which lead to greater releases of carbon to the atmosphere (e.g., Cox et al. 2000; Friedlingstein et al. 2003). Human actions can also lead to feedbacks on climate. If climate change intensifies pressure to convert forests into pastures and cropland, then the climate change may be amplified by the human response (Raupach et al., Chapter 6, this volume). These positive feedbacks increase the fraction of the emitted CO_2 that stays in the atmosphere, increasing the growth rate of atmospheric CO_2 and accelerating climate change. Negative feedbacks are also possible. For example, a northward extension of forest or increased rates of plant growth in a warmer climate could increase rates of carbon storage, constraining further climate change.

A detailed quantitative assessment of a broad range of positive and negative feedbacks will require the use of fully coupled carbon-climate-human models. Such models do not exist yet. For some kinds of interactions (e.g., changes in institutional incentives and constraints on land use), predictive frameworks may be far in the future. Nevertheless, some experience has been gained already using coupled climate-carbon cycle models.

Two such coupled models that simulate the anthropogenic perturbation during the 21st century based on emission scenarios of the Intergovernmental Panel on Climate Change (IPCC) (Cox et al. 2000; Dufresne et al. 2002) show dramatically different magnitudes of climate-carbon cycle interactions (Figure 3.1). Both models simulate an accelerated increase of atmospheric CO_2 as a result of impacts of climate change on the carbon cycle. The magnitude of this feedback varies, however, by a factor of 4 between the two simulations. Without the carbon-climate interaction, both models reach an atmospheric concentration of 700 parts per million (ppm) by 2100. When the carbon-climate feedback is operating, the Hadley Centre model (Cox et al. 2000) reaches 980 ppm, leading to an average near-surface warming of +5K, while the IPSL model (Dufresne et al. 2002) attains only 780 ppm and a warming of +3K. This different behavior can be traced to the higher sensitivity of the land carbon cycle to warming in the Hadley Centre model, and to the larger ocean uptake in the IPSL model (Friedlingstein et al. 2003).

Although these pioneering model simulations represent a large step forward in scientists' ability to elucidate the interactions of the physical climate system with the global carbon cycle, they are also subject to important limitations. In both models, key processes are highly parameterized and poorly constrained. For example, emissions from land use changes are prescribed as an external input, and the associated changes in land cover are not explicitly modeled. In addition, a substantial number of processes and carbon pools are not included in such models. Several of these pools and processes could be vulnerable—that is, they are at risk of losing large amounts of carbon to the atmosphere as a result of a changing climate and/or human drivers (e.g., markets, policies, and demographic dynamics). Incorporating of all these pools and processes into coupled models is a difficult task, particularly because some processes operate through low-probability, high-consequence stochastic events. As a result, coupled climate-carbon models currently have limited capability for assessing the full magnitude and all risks associated with carbon-climate-human feedbacks, and their capability will continue to be limited for the foreseeable future.

An alternative and in many respects complementary approach is that of risk assessment, a method often used to identify and assess the risks associated with the operation of complex systems, such as nuclear power plants or chemical factories (Kammen and Hassenzahl 2001). In a risk analysis, the magnitude and likelihood of certain processes that may lead to catastrophic failures are determined and assessed in order to arrive at

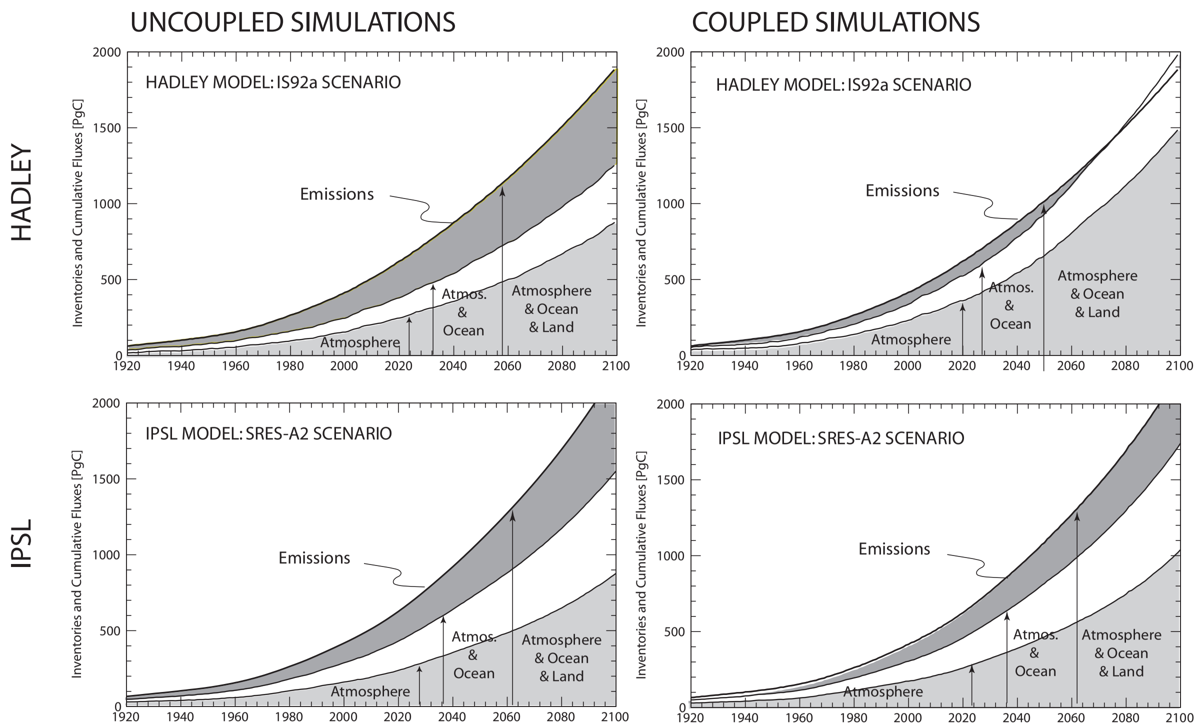

Figure 3.1. Cumulative fossil-fuel emissions and their redistribution among the three major reservoirs of the global carbon cycle over the period from 1920 to 2100 as modeled by two Earth system models (Cox et al. 2000; Dufresne et al. 2002). The left column represents the result of uncoupled simulations, in which changes in the physical climate system did not affect the carbon cycle, whereas results of the right column include such effects. The Hadley model (upper row; Cox et al. 2000) responds very sensitively to such feedbacks, whereas the carbon cycle within IPSL model (lower row; Dufresne et al. 2002) shows a much lower sensitivity. See also Friedlingstein, Chapter 10, this volume.

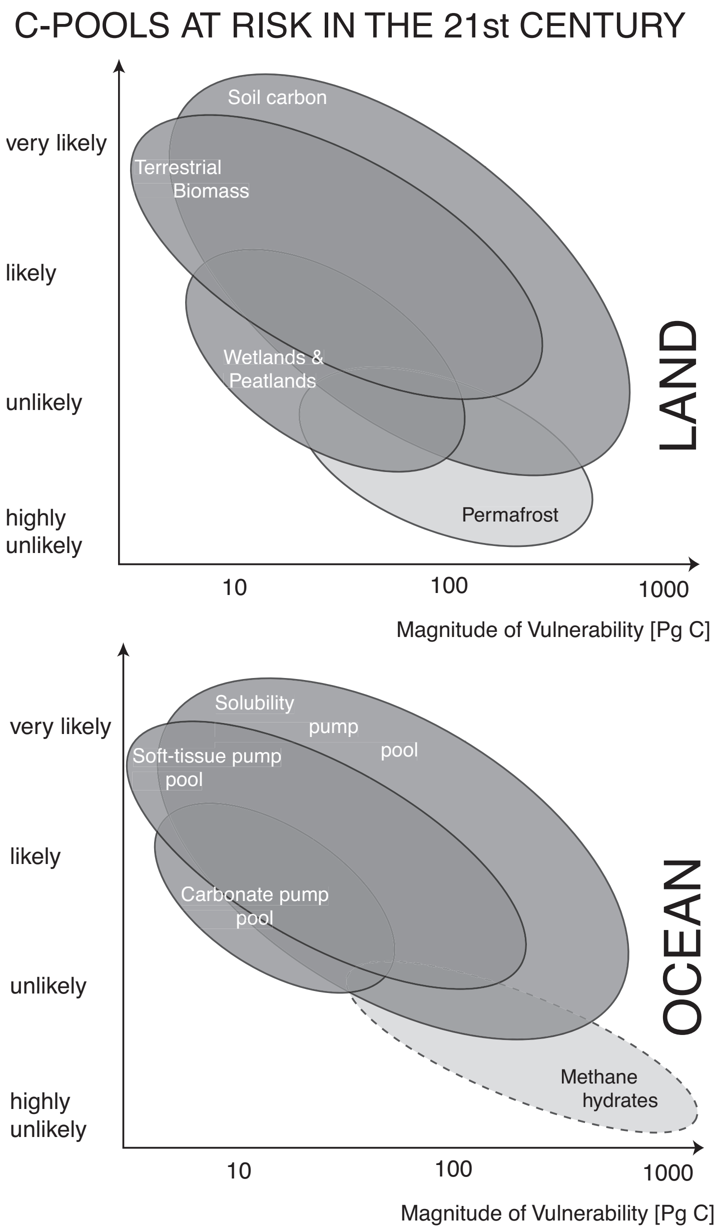

Figure 3.2. Vulnerability of the carbon pools in the 21st century. Shown are the estimated magnitudes and likelihood for these various pools to release carbon into the atmosphere. The exact size of the circles is not known and the boundaries therefore have to be viewed as being approximate only. See Table 3.1 for more details.

the outcomes and likelihood of a number of scenarios. Such approaches are most often used when dealing with systems whose dynamics cannot be captured with differential equations, and hence not encapsulated into numerical or analytical models.

In this chapter, we adopt an approach that is similar to a risk analysis and attempt to identify the magnitude and likelihood of key feedbacks in the carbon-climate-human system. We focus on the major carbon pools and quantify the maximum fraction of these pools that are potentially at risk over the next two decades and over the next century. We contrast these results with the maximum sink strengths of these pools over the same periods. Although we take advantage of model results, we mostly rely on data-based analyses. This approach to quantifying vulnerability is inherently subjective and uncertain. It offers, however, the possibility of going beyond the practical and sometimes fundamental restrictions associated with models and of including all pools and processes that are known to have the potential to influence atmospheric CO_2. We would like to emphasize that we restrict our analysis here to those changes that affect atmospheric CO_2. Other effects of human actions and climate change (e.g., loss of biodiversity and of economic activities currently and potentially related to it) may have impacts that are as large as or even larger than those on the carbon cycle, but they are beyond the scope of this discussion.

Figure 3.2 shows a summary of the vulnerable carbon pools, their magnitude, and the likelihood of the release of the carbon stored in these pools into the atmosphere over the 21st century. This figure is based on the detailed information in Table 3.1. Several key carbon pools may be vulnerable over the timescale of this century. The total amount of carbon stored in these pools is very large, two to five times as large as the total carbon in the current atmosphere. If a substantial fraction of the carbon in these pools were released to the atmosphere, the consequence would be a major increase in atmospheric CO_2 and hence a severe reduction of the anthropogenic CO_2 emissions permitted, if the atmosphere is to not exceed a certain target CO_2 concentration. As discussed in detail later in this chapter, we expect the magnitude of potential releases to increase with increasing stabilization targets, as the higher targets tend to be associated with a larger climate change. The potential release of carbon from these pools has to be viewed in the context of the sinks for atmospheric CO_2, which will probably continue to operate through the 21st century, although likely with gradually falling efficiency (Schimel et al. 2001). Because the potentially vulnerable pools are large, and because they can be rapidly mobilized for release, we believe that the majority of the feedbacks within the carbon-climate-human system are positive—that is, that they lead to an acceleration of the greenhouse gas–induced climate change.

This conclusion does not take into account the possibility of surprises. A small number of very large carbon pools may become vulnerable during the 21st century, especially the carbon stored in the very deep permafrost soils in Siberia and Alaska and the carbon stored in methane hydrates (Figure 3.2). Although we view the release of large quantities of carbon from these two pools as very unlikely, scientists' present understanding of the dynamics and controls of these two pools is insufficient.

Table 3.1. Summary of pools and their vulnerability over the next 20 years and over the course of the 21st century

	Stock size	*Sink capacity*		*Vulnerability in 20 yrs*[a]			*Vulnerability in 100 yrs*[a]				
Pools	*(PgC)*	*20 yrs*	*100 yrs*	*PgC*	*%*	*Confidence*[b]	*PgC*	*%*	*Confidence*[b]	*Processes*	*Direct human drivers*
Land											
Total soils	3,150		300								Deforestation, cattle, management practices
Soils	2,300			10	0.4%	Medium	400	17%	Medium	Rh/erosion/fire	
Permafrost	400			5	1.3%	Low	100	25%	Low	Rh (T, moisture)	
Wetlands	450			10	2.2%	Low	100	22%	Low	Rh vs. methano-genesis/ T., water level	Rice paddies, cattle
Terrestrial living biomass	650	40	150	50	7.7%	Medium	200	31%	Medium		
Tropical	420									Logging/fire/NPP vs Ra/Rh	Logging, fire, land use change, reforestation
Boreal	57									Logging/fire/NPP vs Ra/Rh	Logging, fire, land use change, reforestation
Temperate	145									Pest/diseases/fire/ logging/NPP vs. R	Land use change, refor-estation, aforestation
TOTAL LAND (summed)	3,800	40	450								

Ocean											
Inorganic ocean carbon pool	38,000										
“Inert”	32,200	60–80	200–700	10	0.0%	Medium	400	1%	Medium	Circulation	None
Solubility pump associated	2,700			15	0.6%	Medium	150	6%		Circulation/salinity/T	None
Soft-tissue pump associated[c]	2,500			10	0.4%	Low	150	6%	Low	Circulation/nutrients/ internal dynamics	Coastal eutrophication/ deliberate fertilization/ fisheries
Carbonate pump associated[c]	600			2	0.3%	Low	20	3.3%	Low	Circulation/ nutrients/internal dynamics	Coastal eutrophication/ deliberate fertilization/ fisheries
DOC/POC	725										
TOTAL OCEAN (summed)	38,725	60–80	200–700								
Methane hydrates	>10,000			?			?		Extremely low	Ocean temperature	

Note: T = temperature; R = respiration; Ra = autotrophic respiration; Rh = heterotrophic respiration; NPP = net primary production

[a]We define vulnerability as the maximum part of the total carbon stock that could be released into the atmosphere over the time scale indicated.

[b]Subjective assessment of our confidence

[c]Only the fraction that causes a change in atmospheric CO_2 is given here. The true changes in the pool are about five times larger

While the past 10 years have seen a large increase in the understanding of the present-day global carbon cycle and its anthropogenic CO_2 perturbation (Sabine et al., Chapter 2, this volume), scientists' understanding of how the carbon cycle will evolve into the 21st century and how it will interact with human actions and the physical climate system is overall still poor. This chapter is intended as one step on a long road to accurately incorporating the full suite of carbon-climate-human feedbacks in numerical models.

In assessing the future dynamics of the global carbon cycle, the natural starting point is the current status of the carbon cycle. This aspect is extensively reviewed by Sabine et al. (Chapter 2). We, therefore, concentrate only on those aspects that may cause a change in the carbon cycle over the next 100 years. In the next sections, we first briefly review the expected future trends in the climate system under typical IPCC scenarios. From these changes, we attempt to determine the strengths and vulnerability of the land and ocean carbon sinks in the 21st century.

Setting the Stage: The Earth System in the 21st Century

The vulnerability of many carbon cycle processes and pools depends on the magnitude of future climate change. The magnitude of future climate change, in turn, depends on the vulnerability of the carbon cycle. Therefore there is no unique answer, but it is nevertheless instructive for our ensuing analysis to put some bounds on the expected climate change.

Cubasch et al. (2001) provide a comprehensive review of the projected climate change for the 21st century on the basis of coupled atmosphere-ocean-land climate models and for a large range of radiative forcing projections. For the IS92a emission scenario, they project a mean warming of 1.3°C for the mid-21st century (2021–2050) relative to the 1961–1990 average, with a range from +0.8°C to +1.7°C. This estimate includes the effect of sulphate aerosols, whereas the effect from greenhouse gases alone would be about 0.3°C higher. For the Special Report on Emissions Scenarios (SRES) A2 and B2 emission scenarios,[1] the mean and range are very similar for the same time period. For the end of the 21st century (2071–2100), the mean global surface temperature is projected to increase by 3.0°C, with a range from 1.3°C to 4.5°C, while for the B2 scenario the mean temperature change is +2.2°C and the range is +0.9°C to +3.4°C. These individual scenarios can be compared with the full set of scenarios, for which Cubasch et al. (2001) give a range of +1.4°C to +5.8°C. Unfortunately, no comprehensive coupled climate models have simulated climate change using WRE (Wigley, Richels, Edmonds 1996) stabilization trajectories. Instead, to estimate the temperature changes associated with the WRE scenarios, Cubasch et al. (2001) used a simple climate model (Raper and Cubasch 1996) that has been tuned to the results of a number of comprehensive climate models. They find for the WRE 450 scenario (assuming stabilization at an atmospheric CO_2 concentration of 450 ppm), a temperature change between +1.2°C and +2.3°C for the latter part of this century, increasing to +1.6°C to +2.9°C for WRE550, and +1.8°C to +3.2°C for WRE650. The general pattern of warming is relatively uniform across a large range of models and generally consists of maximum warming in the high latitudes of the Northern Hemisphere and a minimum in the Southern Ocean. There is also a general tendency for the land to warm more than the ocean.

In some cases, even more important for the global carbon cycle is the projected change in the hydrological cycle. While uncertainties and intermodel differences are substantially larger for the hydrological cycle than for the projected temperature changes, the following trends appear to be quite robust: a general increase in the intensity of the hydrological cycle—that is, an increase in mean water vapor, evaporation, and precipitation; and a tendency for an increase in precipitation in the tropics and in the high latitudes and a reduction in precipitation in the subtropical latitudes (Cubasch et al. 2001).

As the coupling between the carbon cycle and human systems is tight, particularly with regard to the terrestrial carbon cycle, common features and regional differences in the development of population, the economy, and social systems in the 21st century are crucial in determining whether certain pools become at risk (Romero Lankao, Chapter 19, this volume). The SRES story lines of Nakicenovic et al. (2000) provide a basis for evaluating these interactions. In the A1 and B1 scenarios, global population is expected to increase to 9 billion people in 2050 and to decrease thereafter to about 7 billion people in 2100. In the A2 and B2 scenarios, global population is expected to grow continuously in the 21st century. Raupach et al. (Chapter 6, this volume) assess some impacts of these scenarios on land use change. We expect that the A2 scenario will lead to highest pressures on local resources, because of the underlying assumption of regionally different impacts on land, and "low incomes in developing countries lessening their ability to meet food demand by improved management or technology" (Raupach et al., Chapter 6).

In summary, the expected stresses on the global carbon cycle might range from relatively small to very substantial, depending on greenhouse gas emission pathways, the magnitude of climate sensitivity, and the development of human systems with regard to their pressure on terrestrial carbon resources. Given this non-negligible risk, it is of great importance to assess the magnitude of the carbon system feedbacks in greater detail. We do not adopt any specific emission scenario or climate sensitivity but tend to use the upper-bound scenarios, consistent with quantifying the maximum expected vulnerability (note that these are nonintervention scenarios, i.e., scenarios that do not contain purposeful human action to curb emissions or increase sinks).

Land Sinks and Their Vulnerability

How will future anthropogenic emission pathways affect the terrestrial carbon cycle? As summarized in Sabine et al. (Chapter 2, this volume), the land biosphere has acted as a substantial net sink for carbon over the past 20 years, removing on average about 0.6

petagrams of carbon (PgC) per year (y^{-1}) from the atmosphere. This net sink is the residual of a land use change flux of the order of 1.2 PgC y^{-1} (DeFries et al. 2002) and an inferred uptake of the order of 1.8 PgC y^{-1} (Table 2.1 and Colorplate 1 in Sabine et al., Chapter 2). Assessing the responses of the terrestrial carbon cycle to the projected climate and human factors in the 21st century will require separating the processes that lead to sources from those that lead to sinks. This attribution task is complicated, because the carbon balance on land is, in general, a small difference between large fluxes in and out. As a consequence, modest proportional changes in fluxes in or out of a pool can have a large impact on the magnitude of the residual sink.

Terrestrial Living Biomass

Living biomass on land contains about 650 PgC, which is slightly less than the amount of carbon in the atmosphere. With an average residence time of about 10 years, the turnover of this pool is relatively fast, making a substantial fraction of this pool potentially vulnerable on timescales of decades to centuries. It is also likely that most of the carbon gained by the terrestrial biosphere over the past few decades has entered this pool and has not progressed in any substantial manner into the longer-lived soil pools (Schlesinger and Lichter 2001). The biomass pool can expand only slowly as a result of increased plant growth or decreased disturbance, but this pool can be degraded quickly, especially with natural and human-induced disturbances.

Land Sinks in the 21st Century

As discussed in Sabine et al. (Chapter 2), scientists' perspective on the current terrestrial carbon balance has changed dramatically over the past decade. A decade ago, it was assumed that the entire land sink is a result of CO_2 fertilization, and perhaps warming (Houghton et al. 1990). This view has been supplanted by an appreciation of the fact that the current and future land sinks are most likely the result of many factors, including land use history and land cover management, and that CO_2 fertilization is likely smaller than previously assumed (e.g., Pacala et al. 2001).

These changes in knowledge are only beginning to be incorporated into terrestrial carbon cycle models (McGuire et al. 2001). In both the Hadley and IPSL models discussed earlier (Cox et al. 2000; Dufresne et al. 2002), CO_2 fertilization is assumed to be the main driver of the present and future terrestrial carbon sink. In the absence of climate change, the projection from these two models is that CO_2 fertilization would be responsible for a cumulative land uptake of 600 PgC for a doubling of atmospheric CO_2 (Friedlingstein et al. 2003). In the Hadley model, a substantial fraction of the carbon that is being released back into the atmosphere as a result of surface warming comes from soil pools that have grown significantly during the 20th and 21st centuries as a result of CO_2 fertilization. Projections of large sinks from CO_2 fertilization are difficult to reconcile with the present understanding of the current land carbon cycle, where a number of constraints (Field et al. 1992) are likely to limit potential land sinks from CO_2 fertilization to the order of 100 PgC over the next 100 years.

The contribution of land use history to the future land sink is rather uncertain. In general, forest sinks due to regrowth of previously harvested forests tend to saturate as a forest matures or are rapidly converted to large sources in forests disturbed by fire, storms, or disease. Key controllers of the future trajectory of sinks in Northern hemisphere forests include the forest age structure, changes in the disturbance regime, and changes in management (Nabuurs, Chapter 16, this volume). The land sink in the 21st century as a result of historical land use change is, at maximum, on the order of 170 PgC (House et al. 2002). This amount assumes that all carbon lost as a result of the past land use change could be stored back on land. A realistic estimate should be less than half this number, since humans still use most of the land deforested in the past and since the land not used for food production is often of poor quality.

From the perspective of the carbon budget of the 21st century, a slowing of deforestation has the same effect as a decrease in fossil emissions or an increase in a land or ocean sink. In addition, reforestation and afforestation may be increasingly important aspects of future carbon management. The extent to which new forest sinks can be realized must be consistent with other societal constraints on available land (Raupach et al., Chapter 6, this volume). The need for expanded areas for agriculture and for replacement of degraded agricultural lands, plus constraints of climate and water availability, are likely to be profoundly important. Given these constraints, reforestation and afforestation are likely to continue to generate carbon sinks throughout the 21st century, but their magnitudes are unlikely to be much more than a few Pg for the next 20 years and a few tens of Pg for the entire 21st century.

Woody encroachment, including the establishment of trees and shrubs on grassland, as well as the thickening of open forests, has been an important aspect of land dynamics on several continents over the last century (Sabine et al., Chapter 2). The driving factors for woody encroachment are too poorly known to support a quantitative estimate of future changes, particularly since its net contribution to atmospheric CO_2 can be positive or negative. We expect this process, however, to be a relatively small contributor, and one whose future carbon trajectory will be primarily determined by human actions.

Fires are a major factor in the terrestrial carbon cycle, annually returning more than 3 PgC to the atmosphere (see Colorplate 1 of Sabine et al., Chapter 2). Because fire has always been a part of the land biosphere and recovery from fires is slow, its role as a carbon source or sink depends on fire frequency and intensity trends over decades to centuries rather than on present burning. Over the past few decades, some regions have probably experienced carbon sinks as a consequence of the success of fire suppression (Sabine et al., Chapter 2). It is unclear whether this trend will continue, as fuels accumulate. We believe that the carbon sink that arises from fire suppression is going to be very small in this coming century but that an increase in fire frequency and intensity is a significant concern (see below).

In summary, the sink potential in living terrestrial biomass over the next 20 years is modest, on the order of a few tens of Pg. Terrestrial sinks may grow somewhat from current levels, but societal and natural limits from land use, fire, and nutrient availability will operate as constraints. Over the next 100 years, some mechanisms and societal drivers contributing to land sinks will continue to operate, but not all (see Raupach et al., Chapter 6, this volume) giving rise to a maximum potential sink of the order of 150 PgC.

Potential Vulnerability of the Land Pools

The carbon pools on land also have the potential to release substantial amounts of carbon into the atmosphere. Many of the mechanisms that are responsible for generating current sinks are closely related to mechanisms that can lead to future sources of carbon to the atmosphere.

Probably the largest future vulnerability of land pools arises from land use change and deforestation, particularly if the world develops along lines similar to those described in the A2 story line—that is, a world in which countries develop in a very heterogeneous pattern. The exact magnitude of the pools at risk is difficult to assess, but we expect that the carbon stored in the living biomass in tropical and subtropical ecosystems is the most vulnerable. Based on past trends, we estimate that on the order of 40 PgC carbon stored in the living biomass might be at risk over the next 20 years, and up to 100 PgC over the course of the 21st century. More subtle changes to land ecosystems, such as human-induced degradation and increases in mortality due to pests might increase this estimate, but probably not by a significant amount.

Deforestation has a number of other implications for climate and the carbon cycle. One set of effects concerns albedo. Boreal deforestation leads to large increases in albedo, with a cooling effect that may be larger than the warming related to the CO_2 release (Bonan et al. 1992; Betts et al. 1997). In the tropics, the albedo effect is much smaller, and effects of rooting depth on transpirable soil moisture may feed back to increase local temperature and decrease precipitation (Shukla et al. 1990). Finally, deforestation also removes the potential that growth of an established forest can serve as a carbon sink.

A second important mechanism that could lead to substantial losses of carbon from the terrestrial biosphere is the redistribution of species and biomes as a result of climate change. It is estimated that the entire terrestrial biosphere gained about 500 PgC between the last glacial maximum (about 20,000 years ago) and the ensuing Holocene (Sigman and Boyle 2000). Whereas the postglacial warming of about 5K occurred over 10,000 years, however, global warming over the next century might result in a similar temperature change, but over a timescale of 100 years. This climate perturbation is likely to occur at a rate exceeding the capacity of plant species to adapt, resulting in a total rearrangement of species and communities. This change will result in transient carbon losses that can eventually be recovered, but only over a much longer timescale (Smith and Shugart 1993).

In the Northern Hemisphere, a warming may lead to a forest dieback at the southern boundaries of presently forested regions, although increase in spring or fall temperature may extend the growing season length of temperate and boreal ecosystems. Invasive species could spread in these regions. Direct human action on species composition, however, including establishment of plantations of non-native "exotic" species, is also likely to continue. The net effect on carbon reservoirs is difficult to estimate but is more likely to be a source than a sink, and the total forest area is more likely to shrink than to expand. In the tropics, warming and increased aridity may lead to a size reduction of the tropical forests (Cox et al. 2000). Given that tropical forests contain the largest carbon pool of terrestrial biota and also the largest net primary production (see Table 2.2 in Sabine et al., Chapter 2), this reduction will likely lead to a release of carbon to the atmosphere.

In summary, we estimate that on the order of 40 PgC carbon stored in the living biomass might be at risk over the next 20 years, and up to 100 PgC over the course of the 21st century. This risk arises mostly from deforestation and other anthropogenic land use changes, An additional 10 PgC (20 years) and 100 PgC (100 years) is probably at risk as a result of climate change—that is, the inability of communities to adapt to a changed climate.

Soil Carbon

Soil carbon is by far the largest carbon pool in terrestrial ecosystems. As summarized in Colorplate 1 of Sabine et al. (Chapter 2), about 2300 PgC is stored in the upper 3 meters (m) of tropical, temperate, and boreal soils. Carbon stocks in wetland and permanently frozen soils are much less certain and are usually not included in carbon cycle models. These stocks are in the range of 200–800 PgC in wetland soils and 200–800 PgC in (nonwetland) permafrost soils. There is therefore more than five times more carbon stored in soils than in the living biomass on land. This large organic carbon reservoir is the result of a delicate balance between biomass accumulation by photosynthesis and oxidation by respiration. Most of this soil carbon is in long-lived pools and therefore relatively inert. Over the course of this century, however, a substantial fraction of this pool might be mobilized. Given the very different dynamics of the various soil pools, we review separately the wetland, the frozen, and the "other" soil pools.

Tropical, Temperate, and Boreal Soils: Potential Sinks and Vulnerability

Carbon sequestration in soil can be achieved through changes in agricultural practices (i.e., no tillage) or with land use changes, such as reforestation of degraded lands. The global potential for carbon sequestration in soils is estimated to be about 0.9 ± 0.3 PgC y^{-1}, with different patterns and rates in European cropland (Smith et al. 2000), U.S. cropland (Lal et al. 1998), and degraded lands (Lal 2003). These estimates of carbon sequestration potential do not account for limitations due to land use, erosion, climate change or economic, social, and political factors (Raupach et al., Chapter 6, this volume). Increased net primary production (NPP) resulting from fertilization of plant growth by elevated CO_2 and nitrogen deposition will tend to increase soil carbon. If these increases are not offset by increased decomposition in a warmer climate, the equilibrium increases in soil carbon should reach the same proportional size as the NPP stimulation. Assuming a fertilization effect of the order of 20 percent of NPP and equilibrium conditions for organic matter decomposition, the maximum possible increase in soil C from CO_2 and N effects is on the order of 300 PgC over the 21st century. This is considerably smaller than the more than 600 PgC accumulation simulated in the Hadley model.

Losses of soil carbon can be caused by decreased inputs from NPP, accelerated decomposition, and increased losses from erosion and combustion. Each of these mechanisms has human and natural drivers (e.g., deforestation, increasing energy consumption, diversion of water bodies). Boreal soils contain huge stocks of carbon. These stocks could increase if tree growth increases in a warmer climate (Dufresne et al. 2002), but they could also decrease, especially if increases in decomposition are larger than increases in NPP or if an increase in fire frequency and magnitude reduces the input of carbon to the soil. Increasing temperature and more frequent droughts (due to increase of climate variability) can enhance desertification in subtropical and tropical regions, stimulating losses of soil carbon. Soil erosion due to changes in the hydrological cycle is also a major disturbance to soil carbon in the semi-arid regions. Climate regulates carbon release by soils through effects of temperature and soil moisture changes. The way in which these two factors interact, however, is highly nonlinear and often opposite (i.e., a temperature increase will increase soil carbon losses while a soil moisture decrease will reduce decomposition but increase soil temperature). In areas where increased temperature and associated drought are expected, the predictions of increased losses of soil carbon losses are debatable. The two coupled climate-carbon simulations from Hadley and IPSL simulate a total loss of soil carbon as high as 600 PgC by the end of the 21st century. As stated earlier, however, these estimates should be seen as unlikely upper estimates as the carbon release is fueled by a large accumulation of soil carbon, resulting from CO_2 fertilization. We estimate from FLUXNET data (Sanderman et al. 2003) that about 8 kilograms (Kg) C per square meter (m^{-2}) in 100 years could be potentially released by soils as a result of changes in climate, assuming a temperature change of 4°C. This amount reflects a potential loss of about 280 PgC in the next century.

Land use change, however, is the major driver of carbon losses in the soil. Since 1850 between 45 and 90 PgC (Houghton et al. 1999; Lal 1999), have been lost through cultivation and disturbances. One of the main characteristics of land use effects is that carbon fluxes are pulsed in response to disturbance frequency and intensity. The effects of such pulses can persist for several years after disturbances and are usually one of the major uncertainties in current models.

Land use change pulses of carbon are also directly influenced by climate-human interactions (i.e., desertification, which is the result of human pressure and aggravated temperature and water impacts). Thus the interaction of direct human effects with climate impacts can amplify the carbon losses, particularly in developing regions. An analysis of recent studies on the impact of land use conversion on soil carbon stocks suggests that up to 50 percent of the total pool of soil carbon can be lost in 50–100 years after conversion (Schulze et al. 2003). Of course not all the land use changes occur in organic rich soils. We therefore estimate that about 25 percent of the total soil carbon pool is vulnerable over the course of this century, equivalent to about 400 PgC.

Wetlands and Peatlands: Potential Sinks and Vulnerability

Wetland carbon fluxes are controlled by climate and land use change. Climate drivers include temperature effects on carbon and other greenhouse gas (mainly CH_4) fluxes. In addition, changes in hydrology can have pronounced effects on both carbon and other greenhouse gas fluxes. Temperature increases stimulate both CO_2 and methane emissions, while changes in the water table have contrasting effects on those two fluxes. Drainage stimulates oxidation of organic material and release of CO_2; in contrast, increasing water levels reduces CO_2 emission but stimulates anaerobic decomposition and methane fluxes to the atmosphere.

To estimate the vulnerability of wetland pools of carbon to climate effects requires a detailed analysis of the interaction between CO_2 and other trace gases. The global warming potential (GWP) of methane is about 20 times larger than that of CO_2. Therefore, a change in the ratio of methane to CO_2 respiration will have a climate impact. Christensen et al. (1996) showed that large northern wetlands, for example, are leading to a net increase in radiative forcing because of their large methane emissions even though they are a net carbon sink. On a global scale, this net source of radiative forcing is equivalent to about 0.5 PgC y^{-1}. Assessing the future evolution of the wetland and peatland emissions of CO_2 and CH_4 is very difficult, as this depends critically on temperature and water levels. If we accept the climate projections that regions that are already receiving substantial rainfall will receive even more in a future warmer climate, then we would expect an increase in the water level, stimulating the production of methane. In addition, higher temperatures would lead to an increase in respiration rates. Both factors conspire to give a relatively high sensitivity of the carbon pools in wetlands and peatlands. A rough estimate is that a CO_2 equivalent of about 100 PgC will become vulnerable in the next 100 years.

An interesting aspect of the interaction between CO_2 and CH_4 emissions in both natural and human-managed wetlands and peatlands is that drainage control provides the possibility of generating a sink potential, as a lowered water table reduces CH_4 emissions at the expense of an increase in the CO_2 emissions. Other environmental concerns (loss of biodiversity, habitats, etc.) should also be considered, however.

Permafrost: Potential Vulnerability

Permanently frozen soils contain a very large pool of carbon, the result of a gradual accumulation over thousands of years. As long as these soils are frozen, the carbon they contain is essentially locked away from the active part of the carbon cycle. When they thaw, however, the carbon becomes vulnerable to rapid loss. Using results from several climate models, all scaled to a global mean warming of 2°C, Anisimov et al. (1999) estimate that the area of permafrost could shrink by about 25 percent. They also estimate an increased summer thaw depth in the areas that maintain permafrost. Current knowledge is too incomplete to convert this loss of area into a loss of carbon, but the pool of carbon vulnerable to this relatively modest warming could total more than 100 PgC. If a substantial fraction of the C loss occurs as methane emissions, as is typical of wetlands, the impact on radiative forcing will be amplified. Assuming that total permafrost carbon is 400 Pg and 20 percent is lost in the next century, the potential release is 80 PgC.

The Ocean Sink and Its Vulnerability

The ocean represents the largest reservoir of carbon in the global carbon cycle (see Colorplate 1 of Sabine et al., Chapter 2) and has taken up approximately 30 percent of the total anthropogenic emissions since the beginning of the industrial period, constituting the second-largest sink of anthropogenic carbon after the atmosphere itself (Le Quéré and Metzl, Chapter 12; Sabine et al., Chapter 2). Therefore, relatively small changes in the oceanic reservoir and its current sink activity may have a large impact on the future trajectory of atmospheric CO_2.

The Ocean Sink Potential in the 21st Century

On millennial timescales, more than 95 percent of the anthropogenic CO_2 will be taken up by the ocean (Archer et al. 1997). On timescales of decades to centuries, however, this large sink potential can be realized only to a limited degree, with the rate-limiting step being the transport of the anthropogenic carbon from the surface into the interior of the ocean (Sarmiento et al. 1992). Because the primary driving force for the oceanic uptake of anthropogenic CO_2 is the atmospheric CO_2 concentration, the magnitude of the oceanic sink in the 21st century depends on the magnitude of the atmospheric perturbation. In the absence of major feedbacks, paradoxically, the higher the anthropogenic CO_2 burden in the atmosphere, the higher the oceanic sink is going to be.

Throughout the anthropocene, the oceanic uptake of anthropogenic CO_2 has scaled nearly linearly with the anthropogenic perturbation of atmospheric CO_2 (Gloor et al. 2003) (0.027 ± 0.008 PgC y^{-1} ppm^{-1}). This is because the atmospheric CO_2 so far has grown quasi exponentially, and the rate-limiting process is linear. This scaling cannot be extrapolated far into the future, as the atmospheric growth rate may change substantially. More important, this scaling is expected to decrease over time as the continued uptake of CO_2 by the ocean decreases the oceanic uptake capacity for a given change in atmospheric CO_2 (Greenblatt and Sarmiento, Chapter 13, this volume). Using this linear scaling and projected atmospheric CO_2, we estimate the maximum oceanic sink potential for the next 20 years in the absence of climate change to be on the order of 60–80 PgC. This estimate compares well with estimates from ocean models that were forced using the IS92a scenario (Watson and Orr 2003). The maximum oceanic sink for the next 100 years is more difficult to estimate as it depends strongly on the atmospheric CO_2 level attained. If we adopt 1,000 ppm as the upper envelope, the cumulative ocean uptake over the next century, in the absence of climate change, might approach 600–700 PgC. In the case of the stabilization scenarios, the oceanic sink is going to be substantially smaller, as the slower growth rate and the smaller atmospheric CO_2 burden both lead to a reduction in the uptake.

Oceanic Carbon Pools and Processes at Risk

How sensitive is this oceanic sink to natural and human-induced changes over the next 20 to 100 years? Greenblatt and Sarmiento (Chapter 13, this volume) review and discuss several detailed studies on the basis of coupled atmosphere-ocean models. Plattner et al. (2001) provide a very detailed analysis of the nature and magnitude of the various feedbacks using a model of intermediate complexity. The goal here is to put these results in perspective by analyzing and assessing the pools and processes that might become at risk in the 21st century from a systems perspective. It is thereby instructive to differentiate between the factors that affect the physical/chemical uptake of anthropogenic CO_2 and those that change the natural carbon cycle in the ocean. The former feedbacks are associated with the anthropogenic CO_2 perturbation in the atmosphere, whereas the latter feedbacks are independent of the anthropogenic CO_2 changes in the atmosphere and arise because of changes in climate and other factors. Table 3.2 gives a summary of the six feedbacks that can either accelerate or decelerate the flux of carbon from the atmosphere into the ocean.

Chemistry Feedback

The feedback that is understood and quantified best is the reduction of the oceanic uptake capacity as a consequence of the uptake of CO_2 from the atmosphere (see Figure 13.3 in Greenblatt and Sarmiento, Chapter 13, this volume). In addition to lowering the uptake capacity for anthropogenic CO_2, this reaction also lowers the pH of the ocean, which could have potentially important consequences for ocean biology and coral reefs (discussed later in this chapter). Sarmiento et al. (1995) demonstrated that the magnitude of the chemistry feedback for the next 20 years remains small but could lead to reduction of more than 30 percent in the cumulative uptake over the next 100 years, with the exact magnitude depending on the size of the anthropogenic perturba-

Table 3.2. Summary of marine carbon cycle feedbacks

Process	*Feedback*	*Estimate for 20 years (PgC)*	*Estimate for 100 years (PgC)*	*Uncertainty/ understanding*
Anthropogenic CO_2 uptake feedbacks				
Chemical feedback	Positive	<5	300	Low/high
Circulation feedback (anthropogenic CO_2)	Positive	6–8	400	Medium/medium
Natural carbon cycle feedbacks				
Temperature/salinity feedback	Positive	15	150	Low/medium to high
Ocean biota feedback	Positive/ negative	10–15	150	High/low
Circulation feedback (natural)	Negative	–20	–400	Medium/medium
Methane feedback	Positive	0	?	Extremely high/low

tion (see also Yi et al. 2001). This positive feedback is fully implemented in ocean carbon cycle models and therefore seldom explicitly discussed.

Ocean Circulation Feedback on Anthropogenic CO_2

Still relatively well understood but more difficult to quantify is the potential impact of ocean circulation changes on the uptake of anthropogenic CO_2. Current climate models tend to show that the warming of the surface ocean, together with a decrease in high latitude salinity as a result of increased precipitation, will lead to a reduction in the surface ocean density relative to that of the underlying waters, thereby increasing vertical stratification. Such an increase in stratification will lead to a reduction of the exchange of surface waters with deeper layers, reducing the downward transport of anthropogenic CO_2 and hence reduce the oceanic uptake of anthropogenic CO_2 from the atmosphere (Sarmiento et al. 1998). Less well established is the impact of climate change on deepwater formation rates and the meridional overturning circulation (Greenblatt and Sarmiento, Chapter 13, this volume).

To estimate the magnitude of the ocean circulation feedback on the oceanic uptake of anthropogenic CO_2, we assume that over the next 20 years the surface to mid-thermocline density gradient changes by about 10 percent (equivalent to a temperature change of 1.5°C). If we further assume that the reduction of the anthropogenic CO_2 uptake scales inversely with the increase of the vertical density gradient, we estimate that these changes in upper ocean stratification lead to a reduction in the cumulative uptake over the next 20 years on the order of 10 percent, or about 6–8 PgC. Over the next 100 years, the ocean circulation feedback could probably be up to three to four times as large, reducing the oceanic uptake of anthropogenic CO_2 by up to 40 percent, with the absolute magnitude depending on the size of oceanic sink. Our estimate here is quite a bit larger than the model-based estimates of 3 percent to 21 percent (Sarmiento and Le Quéré 1996; Sarmiento et al. 1998, Joos et al. 1999; Matear and Hirst 1999; Plattner et al. 2001; see summary by Greenblatt and Sarmiento, Chapter 13) mainly because we assumed a larger change in stratification than was simulated by these models.

The temperature, salinity, and circulation changes already discussed not only influence the uptake of anthropogenic CO_2, but also affect the natural carbon cycle within the ocean, as well as the vast quantities of methane hydrates that are stored along the continental shelves. We look at these changes in turn, focusing first on the individual effects and thereafter at their interactions.

Temperature and Salinity Feedbacks

The amount of dissolved inorganic carbon (DIC) in seawater for a given atmospheric partial pressure and ocean total alkalinity is highly temperature dependent, with a reduction of about 8 millimoles (mmol) per cubic meter (m^{-3}) in DIC for each degree of warming. Therefore, assuming a maximum surface ocean warming of 2°C and a 0.5°C warming over the upper 300 m in the next 20 years would lead to an equilibrium loss of about 15 PgC. Similarly, we estimate that a maximum surface warming of 5°C and a 2°C warming over the upper 1,000 m over the next 100 years would lead to an equilibrium loss of carbon into the atmosphere of more than 150 PgC. These warming-induced losses represent about a 20 percent reduction in the net oceanic uptake for atmospheric CO_2. Models generally find a smaller magnitude of the temperature feedback with a reduction of the order of 50–70 PgC (10–14 percent) over the 21st century (Greenblatt and Sarmiento, Chapter 13, this volume), mostly because their warming is smaller than the magnitude we assumed. Changes in surface salinity also have the potential to alter the ocean atmosphere distribution of inorganic carbon, on the one hand through changes in the solubility of CO_2 and on the other hand through changes in surface ocean alkalinity (dilution effect). We consider the salinity feedback to be a minor factor on the global scale for the 21st century, as we expect relatively small changes in global mean surface salinity, an assessment shared by the model simulations of Plattner et al. (2001).

Circulation Feedbacks (Ocean Biota Constant)

The increase in vertical stratification and the other circulation changes not only reduce the uptake of anthropogenic CO_2 from the atmosphere, but also affect the natural carbon cycling substantially, even in the absence of any changes in ocean productivity. An important distinction between this circulation feedback and that associated with the anthropogenic CO_2 uptake is that the latter feedback affects only the rate at which the atmospheric CO_2 perturbation is equilibrated with the ocean and does not change the long-term equilibrium, whereas the former feedback changes the equilibrium distribution of carbon between the ocean and atmosphere.

As it turns out, in the presence of the biological pumps, these circulation changes appear to lead to a negative feedback—that is, an increase in the uptake of atmospheric CO_2 from the atmosphere. The main reason for this somewhat surprising result is that a slowdown of the surface to deep mixing also reduces the upward transport of remineralized DIC from the thermocline into the upper ocean, while the downward transport of biologically produced organic carbon remains nearly unchanged.

To estimate the magnitude of this feedback, we assume that global export production by particles remains at about 10 PgC y^{-1} and that the upward supply of DIC to compensate in the steady state for this downward organic carbon transport changes inversely proportionally to the increase in vertical stratification. Adopting our previously estimated changes in surface stratification (10 percent over the next 20 years and 40 percent over the next 100 years), we estimate the cumulative reduction in the upward supply of DIC to be 20 PgC over the next 20 years and 400 PgC over the next 100 years. Not all of this reduced supply of DIC will be compensated for by an additional uptake of CO_2 from the atmosphere. Model simulations suggest that this ratio is about half, leading to an estimate of this negative feedback of about 10 PgC over the next 20 years and about 200 PgC over the next 100 years. The model simulations summarized by Greenblatt and Sarmiento (Chapter 13) suggest a range from 33 to more than 110 PgC over the anthropocene. They also show that the models that have a larger positive circulation feedback associated with the anthropogenic CO_2 uptake tend to have a large negative circulation feedback associated with the natural carbon cycle. This result is not unexpected, as the two feedbacks are tightly linked with each other.

Marine Biotic Feedbacks (at Constant Circulation)

The marine biota affects the cycling of carbon in the ocean in two fundamentally different ways. One is through the formation of organic matter in the surface ocean by photosynthesis and the subsequent export of this material into the ocean interior (soft-tissue pump), and the other is through the biogenic formation of $CaCO_3$ shells in the upper ocean, which also tend to sink and dissolve at depth (carbonate pump) (see also Sabine et al., Chapter 2). We will first focus on the soft-tissue pump, because it is quantitatively more important, and discuss the carbonate pump thereafter.

The pool of organic matter in the ocean is small, but most of it turns over on timescales shorter than a year. As a consequence, when considering the impact of climate change on the natural carbon cycle and in particular ocean biota, it is insufficient to look at the pool size of organic matter in the ocean. Instead we have to focus on the time-integrated effect of ocean biology on the oceanic DIC distribution, in particular, the fraction of the surface-to-deep gradient in DIC in the ocean that is induced by the soft-tissue pump. Gruber and Sarmiento (2002) estimated that about 50 percent of the surface-to-deep gradient in oceanic DIC is due to the soft-tissue pump (about 150 micromoles [μmol] kg^{-1}). Integrated over the global ocean, this inorganic carbon pool of biological origin amounts to about 2,500 PgC. Because the ocean CO_2 system is strongly buffered, we estimate that only about 10–20 percent of this "biological" carbon comes from the atmosphere, with the remainder coming from the oceanic DIC pool. As a consequence, only a fraction of any loss from this biological DIC pool ends up changing atmospheric CO_2. Model simulations suggest, for example, that a complete die-off of ocean life would lead to an atmospheric CO_2 increase of about 150–200 ppm only (300–400 PgC) (Gruber and Sarmiento 2002).

The biological pump in the ocean does not operate at full strength, however, as there are many regions where surface nutrients are not entirely used. Only about half of the global nitrate pool is currently associated with the biological carbon pool, offering the possibility of almost doubling the biological DIC pool to about 5,000 PgC. If we assume again that about 10 percent of this carbon comes out of the atmosphere, we find that the absolute upper and lower bounds of how changes in organic matter export can influence atmospheric CO_2 are on the order of ± 250 PgC, an estimate consistent with the recent model simulations by Archer et al. (2000). The fraction of the change in the biological DIC pool that shows up in the atmospheric CO_2 pool appears to be highly model dependent, for reasons not fully understood (Archer et al. 2000).

As very little is known about the sensitivity of marine export production to climate change, estimating how it might change over the next 20–100 years is by necessity very uncertain. Our estimates, therefore, must be viewed with caution. It is nevertheless instructive to determine possible upper bounds and to put them into perspective with the other pools and processes that might change atmospheric CO_2 in the 21st century.

Over the next 20 years, we estimate that global export production will not change by more than about about 3 PgC y^{-1} (25 percent). This estimate would lead to a maximum change in the biological carbon pool of about 60 PgC (4 percent), which would lead to a change in the atmospheric CO_2 pool of only about 6–12 PgC. There are many possible reasons for changes in global marine export production, including changes in surface ocean physical properties, changes in the delivery of nutrients from land by rivers and atmosphere, and internal dynamics of the ocean biota, including complex predator-prey interactions and fisheries-induced pressures on marine predators (see Boyd and Doney 2003 for a comprehensive review). Over the next 100 years, we estimate that the maximum change of the biological carbon in the ocean is likely less than about half of the total pool or maximally about 800 PgC. Taking into account the century timescales for this perturbation in the biological pool to equilibrate with the atmosphere, we believe that the maximum change in the atmospheric CO_2 pool over the next 100 years due to this effect is about 100 to 150 PgC, or of a magnitude similar to the glacial-interglacial CO_2 changes.

There exist possibilities for surprises, though, as our understanding of the mechanisms controlling marine productivity and the subsequent cycling of organic matter in the ocean is limited. For example, while marine export production is generally believed to be controlled primarily by the supply of nutrients and the availability of light (bottom-up control), grazing pressure (top-down control) also plays a major role. With the enormous fishing pressure on the top predators in the ocean, marine foodwebs have already been altered fundamentally in many regions (Jackson et al. 2001) and are expected to change further. On the basis of our current understanding, this human-induced change in foodweb structure should have a relatively small impact on global marine export production, but not enough is known today to exclude the possibility for a bigger role.

We next turn to the impact of climate change on the carbonate pump. Gruber and Sarmiento (2002) estimated that the surface-to-deep gradient generated by this pump amounts to about 60 μmol/kg, giving rise to a calcium carbonate ($CaCO_3$) pump-induced DIC pool in the ocean of about 600 Pg. The effect of changes in this pool on atmospheric CO_2 depends on the timescale considered. On timescales less than a couple of hundred years, an increase in this pool size tends to increase atmospheric CO_2. This increase is because the formation of $CaCO_3$ lowers the alkalinity of the surface ocean more than it lowers DIC, thereby reducing the buffer capacity of the ocean. On very long timescales, this effect is counteracted by interactions with the ocean sediments (Archer et al. 1997).

Therefore, if changes in the export of $CaCO_3$ from the surface ocean are proportional to the changes in the export of organic matter (a constant rain ratio) as expected—for example, if $CaCO_3$ plays an important role as a mineral ballast (Armstrong et al. 2002; Klaas and Archer 2002)—this process would have a counteracting effect on atmospheric CO_2 changes. Conversely, if the formation and export of $CaCO_3$ was a process that is very independent of the export of organic matter, then the two processes could reinforce each other, leading to a larger ocean biota feedback.

One example of how the production and export of $CaCO_3$ could be affected is through changes in pH. There is increasing evidence that the calcification rate by coccolithophorids (the dominant class of phytoplankton that produces $CaCO_3$ shells) might be significantly reduced in response to a lowering of the surface ocean pH (Riebesell et al. 2000). As a reduction of calcification leads to an increase in the ocean uptake capacity for atmospheric CO_2, this represents a negative feedback for climate change. Zondervan et al. (2001), however, estimated the magnitude of this effect to be on the order of 10–20 PgC only for the 21st century. A lower pH also affects the calcification rates of corals (Gattuso et al. 1998; Kleypas et al. 1999). This effect, together with the enhanced sea-surface temperatures, poses a significant threat to coral reefs. The impact will likely be relatively small on atmospheric CO_2 but could be very large on ecosystem services such as fish recruitment, biodiversity, and tourism.

Methane Hydrates

Vast quantities of methane, exceeding all known fossil-fuel reserves, exist in the form of methane hydrates. These occur primarily under continental shelf sediments around the world and in the Arctic permafrost. Hydrates are a crystalline solid of gas trapped in a frozen cage of six water molecules. Geologic evidence suggests massive releases of methane from the ancient sea floor, associated with episodes of global warming (Dickens et al. 1997). Such events can trigger very large undersea landslides and associated tsunamis. Could such a large-scale release happen over the next 100 years? The mechanisms associated with gas hydrate instabilities are very poorly understood, but calculations of heat penetration into sediments suggest that the probability is very low.

Interactions between the Various Feedbacks

Are the various feedbacks described here additive, or is it possible that some of these feedbacks interact with each other to create synergies? The ocean biota feedback and the ocean circulation feedbacks are tightly coupled, because ocean circulation is a prime determinant of marine productivity, mainly because circulation controls the resupply of nutrients from the deep ocean to the light-lit upper ocean. This resupply of nutrients is also coupled with the resupply of the carbon that is associated with the biological pump. Therefore, while a decrease in the supply of nutrients (such as caused by an increase in upper ocean stratification) tends to cause a decrease in marine export production (positive feedback), at the same time it decreases the resupply of the carbon associated with the biological pump (negative feedback). As a consequence, there is a strong tendency for the two processes to counteract each other, with the sign of the combined feedback being very uncertain (see Greenblatt and Sarmiento, Chapter 13, this volume).

In contrast to the nutrient-limited regions, the expected increase in vertical stratification would create a tendency to increase export production in light-limited regions such as the high latitudes. On the basis of nutrient-light-circulation interactions alone, one would expect a relatively small change in global export production but large regional changes, similar to the results of Bopp et al. (2001). Many other interactions, however, might occur (e.g., associated with the delivery of iron and interactions with community structure) that make it nearly impossible to make accurate projections now. Despite all these uncertainties, the circulation feedback and the ocean biota feedbacks appear to be generally additive.

Reviewing the currently available literature, we are not aware of any strong nonlinear interactions between the various feedbacks, so the net oceanic response to the anthropogenic CO_2 perturbation in the atmosphere can be understood from the sum of its parts. Table 3.2 summarizes the six feedbacks that we discussed.

Atmosphere-Land-Ocean Interactions

The carbon cycle on land, in the atmosphere, and in the ocean are tightly linked with each other. After having specified the vulnerabilities within the various subsystems, it is important to investigate the interactions of these feedbacks from a global carbon cycle perspective.

Trade-off between Land and Ocean Uptake through Atmospheric CO_2

Since atmospheric CO_2 is one of the main drivers controlling the oceanic uptake of anthropogenic CO_2 and probably has some control over the sink strength of the terrestrial biosphere, there exists a trade-off between these two sinks. For example, if a given amount of carbon is lost from one of the land carbon pools into the atmosphere, the resulting higher atmospheric CO_2 concentrations would in turn foster increased oceanic uptake, thereby reducing the overall radiative forcing associated with the initial loss. The magnitude of the oceanic compensation can be significant. Cox et al. (2000), for example, find a compensation of approximately 50 percent. The magnitude of this compensation approaches nearly 100 percent over millennial timescales and played an important role in buffering the large loss of carbon from the land biosphere during transitions from interglacial periods to glacial periods. In a similar manner, a decrease in ocean uptake by, for example, increased stratification would also be partly compensated by increased terrestrial uptake. The magnitude of this trade-off effect, however, in this case depends on the sensitivity of the terrestrial biosphere to atmospheric CO_2.

Potential Vulnerability of the Land-Ocean "Conveyor Belt"

The vulnerability of the land-river-marine transport and fate system ("conveyor belt") depends on the regional functioning of the hydrological cycle and on the way its dynamics are driven by patterns of water use, changes in land use practices affecting mobilization of carbon and nutrients to and through channels, and ultimately how the seas of the continental margins respond to these changes in forcing. Under scenarios of an increased hydrological cycle and changes in land use, river flow could increase, leading to increased ability of rivers to carry materials. Conversely, those areas subject to drier conditions would incur reduced flow and a reduced capacity of rivers to mobilize materials. Regardless of climate changes, an increased demand for water for agricultural and for urban and industrial practices would reduce water available to be routed down channels. Changes in forcing from upstream would affect coastal receiving waters. Decreases in river flow (from either drier climates or increased irrigation or retention in reservoirs) would reduce the cross-shelf water exchange because of a reduced buoyancy effect, resulting in a diminished onshore nutrient supply (Chen, Chapter 18, this volume). Primary production on the shelf would decrease proportionately, and with it the ability of continental shelves to act as a sink for carbon. If, on the other hand, there was a significant increase in nutrients (from increased fertilizers and urban use), higher levels of productivity, and even eutrophication, in estuaries or coastal regions could be maintained. Over a longer term, sea-level rises could have pronounced effects on carbon. We are currently not in a position to estimate the exact magnitudes of these effects, but it seems likely that these changes will have a much larger impact on regional issues than on atmospheric CO_2.

Vegetation Cover–Dust–Ocean Fertilization

A third possible feedback arising out of the interaction of the atmosphere-land-ocean system concerns the increasing evidence that micronutrients such as iron play an important role in regulating the strength of the ocean's biological pump, particularly in the high-nutrient low-chlorophyll (HNLC) regions. As a substantial fraction of the iron input into the surface ocean comes from the atmosphere, changes in this input could lead to changes in ocean productivity, export production, and eventually atmospheric CO_2. This mechanism has actually been proposed as a possible reason for the low atmospheric CO_2 concentration during the last glacial period (e.g., Martin 1990; see Joos and Prentice, Chapter 7, this volume). This feedback may also operate in the future. If degradation or desertification on land leads to increased dust inputs into oceanic HNLC regions, this dust input would stimulate oceanic productivity and thus increase oceanic CO_2 uptake. We estimate, however, that this effect will be relatively minor over the next 20 or 100 years in really reducing atmospheric CO_2, but this effect might lead to large changes on a regional scale.

It also has been proposed that atmospheric iron input plays a major role in controlling marine N_2-fixation and that changes in the iron input would therefore lead to changes in the total inventory of fixed nitrogen in the ocean, which could fuel higher levels of marine productivity throughout the ocean (Falkowski 1997; Broecker and Henderson 1998). It is unlikely, however, that this mechanism will lead to large changes in the ocean atmosphere distribution of inorganic carbon over the next 20 to 100 years.

Summary and Conclusions

Our analysis of the vulnerability of the carbon pool on land and in the ocean reveals that a substantial amount of carbon is at risk of becoming mobilized and being released into the atmosphere (Figure 3.2, Figure 3.3, and Table 3.1). Recognizing that our estimates are quite uncertain, we find that on the order of several tens of Pg could be lost from the land and ocean carbon pool over the next 20 years and several hundred Pg over the course of this century. On land, we estimate that the maximum potential losses are substantially larger than the maximum potential sinks, suggesting that the overall sign of the feedbacks arising from carbon-climate-human interactions is positive—that is, accelerating climate change. For the ocean, we must differentiate between the ocean uptake that exists in the absence of climate change and the ocean sink that is due to climate feedbacks. The ocean uptake in the absence of any feedback is very large, on the order of several hundred Pg C. This uptake could be substantially offset by the feedbacks, since their net sign is likely positive as well.

Our results imply that the fraction of the anthropogenic CO_2 emissions that will remain in the atmosphere and force a climate change will increase in the future. As a consequence, the permissible anthropogenic emissions in a stabilization pathway are smaller in the presence of interactions between the carbon-climate-human systems. The

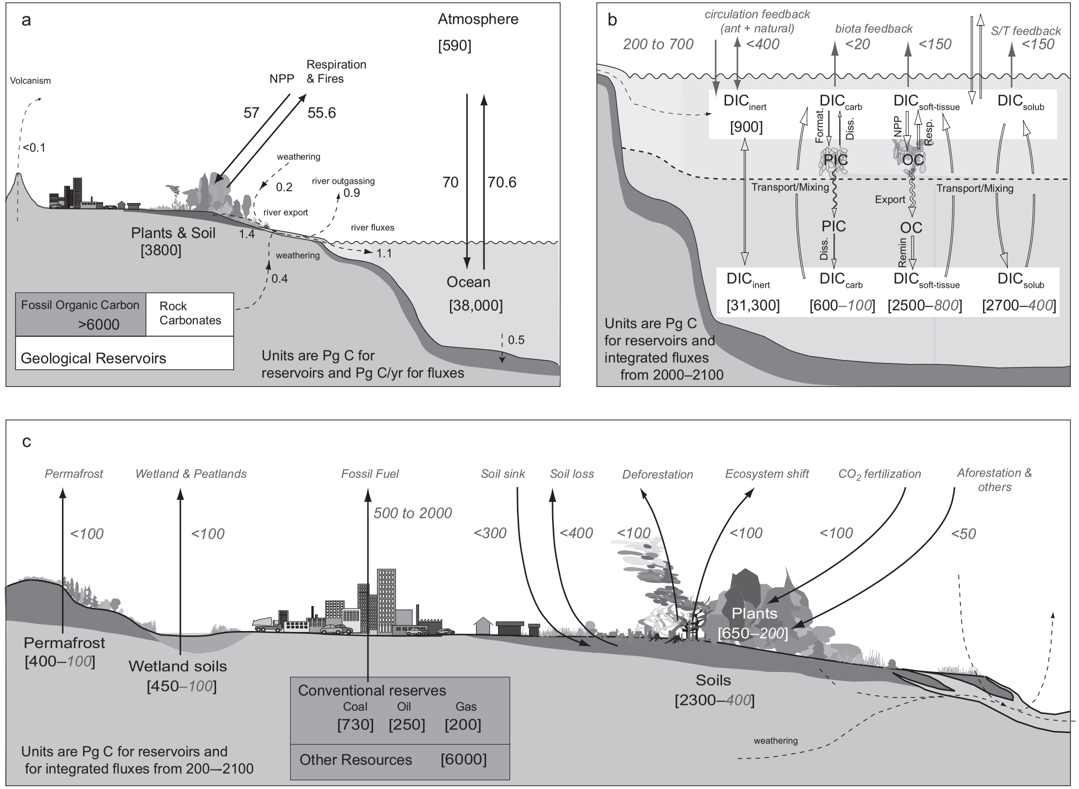

Figure 3.3. Global carbon cycle, its sink potential, and its vulnerability in the 21st century. (a) Preindustrial carbon cycle, (b) major ocean carbon pools and their vulnerability, and (c) major land pools and their vulnerability. The italic numbers shown in (b) and (c) represent our maximum estimates of the vulnerability. See also Table 3.1 and text. The oceanic changes in the pool sizes are larger than the induced air-sea CO_2 fluxes because of the buffering effect of seawater.

magnitudes of the feedbacks depend to a large degree on the state of the climate system. The feedbacks tend to get stronger with larger climate forcing, implying that the magnitude of the offsetting effect for the permissible emissions increases with the concentration of the atmospheric CO_2 stabilization target. A final observation that arises from the inspection of Table 3.1 is that the vulnerability of the global carbon cycle stems from the vulnerability of many different pools and processes. This finding highlights the need for a highly integrative and interdisciplinary approach for studying the global carbon cycle and the need to view Earth as a coupled system rather than as an entity that can be studied in parts.

Although we view our analysis as a first step toward identifying and quantifying these feedbacks, our estimates are preliminary at best. We hope, however, to encourage subsequent research that will improve our initial assessment.

Note

1. The SRES A2 and B2 emission scenarios were developed by Nakicenovic et al. (2000) and represent two representative samples out of a large family of scenarios. The A2 story line describes a very heterogeneous world. The underlying theme is self-reliance and preservation of local identities. Fertility patterns across regions converge very slowly, which results in continuously increasing population. Economic development is primarily regionally oriented, and per capita economic growth and technological change are more fragmented and slower than in other storylines. The B2 story line describes a world in which the emphasis is on local solutions to economic, social, and environmental sustainability. It is a world with continuously increasing global population, at a rate lower than A2, and intermediate levels of economic development and of technological change. Although the scenario is also oriented toward environmental protection and social equity, it focuses on local and regional levels. Because of higher population and slow technological advances, the A2 scenario results in one of the highest levels of CO_2 emissions, whereas B2 represents an intermediate-level emission scenario.

Literature Cited

Anisimov, O. A., F. E. Nelson, and A. V. Pavlov. 1999. Predictive scenarios of permafrost development under conditions of global climate change in the XXI century. *Earth Cryology* 3 (4): 15–25.

Archer, D., H. Kheshgi, and E. Maier-Reimer. 1997. Multiple timescales for neutralization of fossil fuel CO_2. *Geophysical Research Letters* 24 (4): 405–408.

Archer, D., G. Eshel, A. Winguth, and W. Broecker. 2000. Atmospheric CO_2 sensitivity to the biological pump in the ocean. *Global Biogeochemical Cycles* 14:1219–1230.

Armstrong, R. A., C. Lee, J. I. Hedges, S. Honjo and S. G. Wakeham. 2002. A new, mechanistic model for organic carbon fluxes in the ocean based on the quantitative association of POC with ballast minerals. *Deep-Sea Research* 49:219–236.

Bacastow, R., and C. D. Keeling. 1973. Atmospheric carbon dioxide and radiocarbon in the natural carbon cycle. II. Changes from A.D. 1700 to 2070 as deduced from a geochemical reservoir. Pp. 86–135 in *Carbon and the biosphere,* edited by G. M. Woodwell and E. V. Pecan. Springfield, VA: U.S. Department of Commerce.

Betts, R. A., P. M. Cox, S. E. Lee, and F. I. Woodward. 1997. Contrasting physiological and structural vegetation feedbacks in climate change simulations. *Nature* 387: 796–799.

Bonan, G. B., D. B. Pollard, and S. L. Thompson. 1992. Effects of boreal forest vegetation on global climate. *Nature* 359:716–718.

Bopp, L., P. Monfray, O. Aumont, J.-L. Dufresne, H. Le Treut, G. Madec, L. Terray, and J. C. Orr. 2001. Potential impact of climate change on marine export production. *Global Biogeochemical Cycles* 15 (1): 81–99.

Boyd, P. W., and S. C. Doney. 2003. The impact of climate change and feedback processes on the ocean carbon cycle. Pp. 157–193 in *Ocean biogeochemistry: The role of the ocean carbon cycle in global change,* edited by M. J. R. Fasham. Berlin: Springer Verlag.

Broecker, W. S., and G. M. Henderson. 1998. The sequence of events surrounding Termination II and their implications for the cause of glacial-interglacial CO_2 changes. *Paleoceanography* 13 (4): 352–364.

Caspersen, J. P., S. W. Pacala, J. C. Jenkins, G. C. Hurtt, P. R. Moorcroft, and R. A. Birdsey. 2000. Contributions of land-use history to carbon accumulation in U.S. forests. *Science* 290:1148–1151.

Christensen, T., I. Prentice, J. Kaplan, A. Haxeltine, and S. Sitch. 1996. Methane flux from northern wetlands and tundra: An ecosystem source modelling approach. *Tellus* 48B:652–661.

Collatz, G. J., J. A. Berry, and C. Grivet. 1992. Coupled photosynthesis-stomatal conductance model for leaves of C_4 plants. *Australian Journal of Plant Physiology* 19:519–538.

Cox, P. M., R. A. Betts, C. D. Jones, S. A. Spall and I. J. Totterdell. 2000. Acceleration of global warming due to carbon-cycle feedbacks in a coupled climate model. *Nature* 408:184–187.

Cubasch, U., G. A. Meehl, G. J. Boer, R. J. Stouffer, M. Dix, A. Noda, C. A. Senior, S. Raper, and K. S. Yap. 2001. Projections of future climate change. Pp. 525–582 in *Climate change 2001: The scientific basis (Contribution of Working Group I to the Third Assessment Report of the Intergovernmental Panel on Climate Change),* edited by J. T. Houghton, Y. Ding, D. J. Griggs, M. Noguer, P. J. van der Linden, X. Dai, K. Maskell, and C. A. Johnson. Cambridge: Cambridge University Press.

DeFries, R. S., L. Bounoua, and G. J. Collatz. 2002. Human modification of the landscape and surface climate in the next fifty years. *Global Change Biology* 8 (5): 438–458.

Dickens, G. R., M. M. Castillo, and J. C. G. Walker. 1997. A blast of gas in the latest Paleocene: Simulating the first-order effects of massive dissociation of oceanic methane hydrate. *Geology* 25:259–262

Dore, J. E, R. Lukas, D. W. Sadler, and D. M. Karl. 2003. Climate-driven changes to the atmospheric CO_2 sink in the subtropical North Pacific Ocean. *Nature* 424:754–757.

Dufresne, J.-L., P. Friedlingstein, M. Berthelot, L. Bopp, P. Ciais, L. Fairhead, H. LeTreut, and P. Monfray. 2002. Effects of climate change due to CO_2 increase on land and ocean carbon uptake. *Geophysical Research Letters* 29 (10), doi:10.1029/2001GL013777.

Falkowski, P. G.1997. Evolution of the nitrogen cycle and its influence on the biological sequestration of CO_2 in the ocean. *Nature* 387:272–275.

Falkowski, P. G., and C. Wilson. 1992. Phytoplankton productivity in the North Pacific Ocean since 1900 and implications for absorption of anthropogenic CO_2. *Nature* 358:741–743.

Farquhar, G. D., S. von Caemmerer, and J. A. Berry. 1980. A biochemical model of photosynthetic CO_2 assimilation in leaves of C_3 species. *Planta* 149:78–90.

Field, C. B., F. S. Chapin III, P. A. Matson, and H. A. Mooney. 1992. Responses of terrestrial ecosystems to the changing atmosphere: A resource-based approach. *Annual Review of Ecology and Systematics* 23:201–235.

Friedlingstein, P., J.-L. Dufresne, P. M. Cox, and P. Rayner. 2003. How important is the positive feedback between climate change and the carbon cycle? *Tellus* (in press).

Gattuso, J.-P., M. Frankignoulle, I. Bourge, S. Romaine, and R. W. Buddemeier. 1998. Effect of calcium carbonate saturation of seawater on coral calcification. *Global and Planetary Change* 18:37–46.

Gloor, M., N. Gruber, J. L. Sarmiento, C. S. Sabine, R. A. Feely, and C. Rödenbeck. 2003. A first estimate of present and pre-industrial air-sea CO2 flux patterns based on ocean interior carbon measurements and models. *Geophysical Research Letters* 30 (1), doi:10.1029/2002GL015594.

Gregg, W. W., and M. E. Conkright. 2002. Decadal changes in global ocean chlorophyll. *Geophysical Research Letters* 29 (15), doi:10.1029/2002GL014689.

Gruber, N., and J. L. Sarmiento. 2002. Biogeochemical/physical interactions in elemental cycles. Pp. 337–399 in *The sea,* Vol. 12, *Biological-physical interactions in the sea,* edited by A. R. Robinson, J. J. McCarthy, and B. J. Rothschild. New York: John Wiley and Sons.

Houghton, R. 1999. The annual net flux of carbon to the atmosphere from changes in land use 1850–1990. *Tellus* 51B:298–313.

Houghton, J. T., G. J. Jenkins, and J. J. Ephraums, eds. 1990. *Climate change: The IPCC scientific assessment (Contribution of Working Group I to the first assessment report of the Intergovernmental Panel on Climate Change).* Cambridge: Cambridge University Press.

Houghton, R. A., J. L. Hackler, and K. T. Lawrence. 1999. The U.S. carbon budget: Contributions from land-use change. *Science* 285:574–578.

House, J. I., C. Prentice, and C. Le Quéré. 2002. Maximum impacts of future reforestation or deforestation on atmospheric CO_2. *Global Change Biology* 8 (11): 1047–1052.

Jackson, J. B. C., M. X. Kirby, W. H. Berger, K. A. Bjorndal, L. W. Botsford, B. J. Bourque, R. H. Bradbury, R. Cooke, J. Erlandson, J. A. Estes, T. P. Hughes, S. Kidwell, C. B. Lange, H. S. Lenihan, J. M. Pandolfi, C. H. Peterson, R. S. Steneck, M. J. Tegner, and R. R. Warner. 2001 . Historical overfishing and the recent collapse of coastal ecosystems. *Science* 293:629–638

Jones, C. D, P. M. Cox, R. L. H. Essery, D. L. Roberts, and M. J. Woodage. 2003. Strong carbon cycle feedbacks in a climate model with interactive CO_2 and sulphate aerosols. *Geophysical Research Letters* 30 (9): 1479, doi:10.1029/2003GL016867.

Joos, F., G.-K. Plattner, T. F. Stocker, O. Marchal, and A. Schmittner. 1999. Global warming and marine carbon cycle feedbacks on future atmospheric CO_2. *Science* 284:464–467.

Kammen, D. M., and D. M. Hassenzahl. 2001. *Why should we risk it? Exploring environ-*

mental, health, and technological problem solving. Princeton, NJ: Princeton University Press.

Klaas, C., and D. E. Archer. 2002. Association of sinking organic matter with various types of mineral ballast in the deep sea: Implications for the rain ratio. *Global Biogeochemical Cycles* 16 (4), doi:10.1029/2001GB001765.

Kleypas, J. A., R. W. Buddemeier, D. Archer, J.-P. Gattuso, C. Langdon, and B. N. Opdyke. 1999. Geochemical consequences of increased atmospheric carbon dioxide on coral reefs. *Science* 284:118–120.

Lal, R. 1999. Soil management and restoration for C sequestration to mitigate the greenhouse effect. *Progress in Environmental Science* 1:307–326

———. 2003. Global potential of soil C sequestration to mitigate the greenhouse effect. *Critical Reviews in Plant Sciences* 22:151–184.

Lal, R., J. M. Kimble, R. F. Follett, and C. V. Cole. 1998. *The potential of U.S. cropland to sequester C and mitigate the greenhouse effect.* Chelsea, MI: Ann Arbor Press.

Laws, E. A., P. Falkowski, E. A. Carpenter, and H. Ducklow. 2000. Temperature effects on export production in the open ocean. *Global Biogeochemical Cycles* 14 (4): 1231–1246.

Martin, J. H. 1990. Glacial-interglacial CO_2 change: The iron hypothesis. *Paleoceanography* 5 (1): 1–13.

Matear, R. J., and A. C. Hirst. 1999. Climate change feedback on the future oceanic CO_2 uptake. *Tellus* 51B:722–733.

McGuire, A. D., S. Sitch, J. S. Clein, R. Dargaville, G. Esser, J. Foley, M. Heimann, F. Joos, J. Kaplan, D. W. Kicklighter, R. A. Meier, J. M. Melillo, B. Moore III, I. C. Prentice, N. Ramankutty, T. Reichenau, A. Schloss, H. Tian, L. J. Williams, and U. Wittenberg. 2001. Carbon balance of the terrestrial biosphere in the twentieth century: Analyses of CO_2, climate and land-use effects with four process-based ecosystem models. *Global Biogeochemical Cycles* 15:183–206.

Nakicenovic, N., J. Alcamo, G. Davis, B. de Vries, J. Fenhann, S. Gaffin, K. Gregory, A. Grübler, T. Y. Jung, T. Kram, E. L. La Rovere, L. Michaelis, S. Mori, T. Morita, W. Pepper, H. Pitcher, L. Price, K. Raihi, A. Roehrl, H.-H. Rogner, A. Sankovski, M. Schlesinger, P. Shukla, S. Smith, R. Swart, S. van Rooijen, N. Victor, Z. Dadi. 2000. *IPCC special report on emissions scenarios.* Cambridge: Cambridge University Press.

Pacala, S. W., G. C. Hurtt, R. A. Houghton, R. A. Birdsey, L. Heath, E. T. Sundquist, R. F. Stallard, D. Baker, P. Peylin, P. Ciais, P. Moorcroft, J. Caspersen, E. Shevliakova, B. Moore, G. Kohlmaier, E. Holland, M. Gloor, M. E. Harmon, S.-M. Fan, J. L. Sarmiento, C. Goodale, D. Schimel, and C. B. Field. 2001. Convergence of land- and atmosphere-based U.S. carbon sink estimates. *Science* 292:2316–2320.

Parton, W. J., D. S. Schimel, C. V. Cole, and D. S. Ojima. 1987. Analysis of factors controlling soil organic matter levels in Great Plains grasslands. *Soil Science Society of America Journal* 51:1173–1179.

Plattner, G.-K., F. Joos, T. F. Stocker, and O. Marchal. 2001. Feedback mechanisms and sensitivities of ocean carbon uptake under global warming. *Tellus* 53B:564–592.

Raich, J. W., E. B. Rastetter, J. M. Melillo, D. W. Kicklighter, P. A. Steudler, B. J. Peterson, A. L. Grace, B. Moore III, and C. J. Vörösmarty. 1991. Potential net primary production in South America. *Ecological Applications* 1:399–429.

Raper, S. C. B., and U. Cubasch. 1996. Emulation of the results from a coupled general circulation model using a simple climate model. *Geophysical Research Letters* 23:1107–1110.

Riebesell, U., I. Zondervan, B. Rost, P. D. Tortell, R. Zebe, and F. M. M. Morel. 2000. Increased CO_2 decreases marine planktonic calcification. *Nature* 407:364—367.

Salati, E., and P. B. Vose. 1984. Amazon basin: A system in equilibrium. *Science* 225:129.

Sanderman, J., R. G. Amundson, and D. D. Baldocchi. 2003. Application of eddy covariance measurements to the temperature dependence of soil organic matter mean residence time. *Global Biogeochemical Cycles* 17:1061, doi:10.1029/2001GB001833.

Sarmiento, J. L., and C. Le Quéré. 1996. Oceanic CO_2 uptake in a model of century-scale global warming. *Science* 274:1346–1350.

Sarmiento, J. L., J. C. Orr, and U. Siegenthaler. 1992. A perturbation simulation of CO_2 uptake in an ocean general circulation model. *Journal of Geophysical Research* 97 (C3): 3621–3645.

Sarmiento, J. L., C. Le Quéré, and S. W. Pacala. 1995. Limiting future atmospheric carbon dioxide. *Global Biogeochemical Cycles* 9:121–137.

Sarmiento, J. L., T. M. C. Hughes, R. J. Stouffer, and S. Manabe. 1998. Simulated response of the ocean carbon cycle to anthropogenic climate warming. *Nature* 393:245–249.

Schimel, D. S., D. Alves, I. Enting, M. Heimann, F. Joos, D. Raynaud, T. Wigley, M. Prather, R. Derwent, D. Ehhalt, P. Fraser, E. Sanhueza, X. Zhou, P. Jonas, R. Charlson, H. Rohd, K. P. Sadasivan, Y. Shine, Y. Fouquart, V. Ramaswamy, S. Solomon, J. Srinivasan, D. Albritton, I. Isaken, M. Lal, and D. Wuebbles. 1996. Radiative forcing of climate change. Pp. 65–131 in *Climate change 1995: The science of climate change.* Cambridge: Cambridge University Press.

Schimel, D. S., J. I. House, K. A. Hibbard, P. Bousquet, P. Ciais, P. Peylin, B. H. Braswell, M. J. Apps, D. Baker, A. Bondeau, J. Canadell, G. Churkina, W. Cramer, A. S. Denning, C. B. Field, P. Friedlingstein, C. Goodale, M. Heimann, R. A. Houghton, J. M. Melillo, B. Moore III, D. Murdiyarso, I. Noble, S. W. Pacala, I. C. Prentice, M. R. Raupach, P. J. Rayner, R. J. Scholes, W. L. Steffen, and C. Wirth. 2001. Recent patterns and mechanisms of carbon exchange by terrestrial ecosystems. *Nature* 414:169–172.

Schlesinger, W. H., and J. Lichter. 2001. Limited carbon storage in soil and litter of experimental forest plots under increased atmospheric CO_2. *Nature* 411:466–469.

Schulze, E. D., D. Mollicone, F. Achard, G. Matteucci, S. Frederici, H. D. Eva, and R. Valentini. 2003. Making deforestation pay under the Kyoto protocol? *Science* 299:1669.

Shaw, M. R., E. S. Zavaleta, N. Chiariello, H. A. Mooney, and C. B. Field. 2002. Grassland responses to global environmental changes suppressed by elevated CO_2. *Science* 298:1987–1990.

Shukla, J., C. Nobre, and P. Sellers. 1990. Amazon deforestation and climate change. *Science* 247:1322–1325.

Sigman, D., and E. Boyle. 2000. Glacial/interglacial variations in carbon dioxide: Searching for a cause. *Nature* 407:859–869.

Smith, P., D. Powlson, J. U. Smith, P. Falloon, and K. Coleman. 2000. Meeting Europe's climate change commitments: Quantitative estimates of the potential for carbon mitigation by agriculture. *Global Change Biology* 6:525–539.

Smith, T. M., and H. H. Shugart. 1993. The transient response of terrestrial carbon storage to a perturbed climate. *Nature* 361:523–526.

Tortell, P. D., G. R. DiTullio, D. M. Sigman, and F. M. M. Morel. 2002. CO_2 effects on species composition and nutrient utilization in an Equatorial Pacific phytoplankton community. *Marine Ecology Progress Series* 236:37–43.

Watson, A., and J. Orr. 2003. Carbon dioxide fluxes in the global ocean. Pp. 123–143 in *Ocean biogeochemistry: The role of the ocean carbon cycle in global change,* edited by M. J. R. Fasham. Berlin: Springer Verlag.

Wigley, T. M. L., R. Richels, and J. A. Edmonds. 1996. Economic and environmental choices in the stabilization of atmospheric CO_2 concentrations. *Nature* 379:240–243.

Yi, C., P. Gong, M. Xu, and Y. Qi. 2001. The effects of buffer and temperature feedback on the oceanic uptake of CO_2. *Geophysical Research Letters* 28 (5): 751–754.

Zondervan, I., R. E. Zeebe, B. Rost, and U. Riebesell. 2001. Decreasing marine biogenic calcification: A negative feedback on rising atmospheric pCO_2. *Global Biogeochemical Cycles* 15 (2): 507–516.