PART III OVERVIEW

The Carbon Cycle in Land and Water Systems

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The six chapters (Chapters 10–15) in Part III consider the current and future carbon balance of terrestrial and aquatic ecosystems in North America. Although the amount of carbon exchanged between these ecosystems and the atmosphere each year through photosynthesis and plant and microbial respiration is large, the net balance for all of the ecosystems, combined, is currently a net sink of 472-592 Mt C yr^{-1}, and offsets only about 25-30% of current fossil fuel emissions from the region (1856 Mt C yr^{-1} in 2003) (see Chapter 3). If managed properly, these systems have the potential to become significantly larger sinks of carbon in the future; they may also become significant net sources of carbon if managed poorly or if the climate warms.

Much of the current North American carbon sink is the result of past changes in land use and management. The large sink in the forests of Canada and the United States, for example, is partly the result of continued forest growth following agricultural abandonment that occurred in the past, partly the result of current and past management practices (e.g., fire suppression), and partly the result of forest responses to a changing environment (climatic change, CO_2 fertilization, and the increased mobilization of nutrients). However, the relative importance of these three broad factors in accounting for the current sink is unknown. Estimates vary from attributing nearly 100% of the sink in United States forests to regrowth (Caspersen *et al.*, 2000; Hurtt *et al.*, 2002) to attributing nearly all of it to CO_2 fertilization (Schimel *et al.*, 2002). The attribution question is critical because the current sink may be expected to increase in the future if the important mechanism is CO_2 fertilization, for example, but may be expected to decline if the important mechanism is forest regrowth (forests accumulate carbon more slowly as they age). Understanding the history of land use, management, and disturbance is critical because disturbance and recovery are major determinants of the net terrestrial carbon flux.

Land-use change and management have been, and will be, important in the carbon balance of other ecosystems besides forests. The expansion of cultivated lands in Canada and the United States in the 19th century released large amounts of carbon to the atmosphere (Houghton *et al.*, 1999), leaving those lands with the potential for recovery (i.e., a future carbon sink), if managed properly. For example, recent

changes in farming practice may have begun to recover the carbon that was lost decades ago. Grazing lands, although not directly affected by cultivation, were, nevertheless, managed in the United States through fire suppression. The combined effects of grazing and fire suppression are believed to have promoted the invasion of woody vegetation, possibly a carbon sink at present. Wetlands are the second largest net carbon sink (after forests), but the magnitude of the sink was larger in the past than it is today, again, as a result of land-use change (draining of wetlands for agriculture and forestry). The only lands that seem to have escaped management are those lands overlying permafrost, and they are clearly subject to change in the future as a result of global warming. Settled lands, by definition, are managed and are dominated by fossil fuel emissions. Nevertheless, the accumulation of carbon in urban and suburban trees suggests a net sequestration of carbon in the biotic component of long-standing settled lands. Residential lands recently cleared from forests, on the other hand, are sources of carbon (Wienert and Hamburg, 2006).

From the perspective of carbon and climate, ecosystems are important if (1) they are currently large sources or sinks of carbon or (2) they have the potential to become large sources or sinks of carbon in the future through either management or environmental change, where "large" sources or sinks, in this context, are determined by the product of area (hectares) times flux per unit area (or flux density) (Mg C ha^{-1} yr^{-1}).

The largest carbon sink in North America (350 Mt C yr^{-1}) is associated with forests (Chapter 11) (Table 1). The sink includes the carbon accumulating in wood products (e.g., in increasing numbers of houses and landfills) as well as in the forests themselves. A sink is believed to exist in wetlands (Chapter 13), including the wetlands overlying permafrost (Chapter 12), although the magnitude of this sink is uncertain. More certain is the fact that the current sink is considerably smaller than it was before wetlands were drained for agriculture and forestry. The other important aspect of wetlands is that they hold nearly two thirds of the carbon in North America. Thus, despite the current net sink in these systems, their potential for future emissions is large.

Table 1. Ecosystems in North America: their areas, net annual fluxes of carbon, and their potential for sources (+) or sinks (–) in the future

Although management has the potential to increase the carbon sequestered in agricultural (cultivated) lands, these lands today are nearly in balance with respect to carbon (Chapter 10). The carbon lost to the atmosphere from cultivation of organic soils is approximately balanced by the carbon accumulated in mineral soils. In the past, before cultivation, these soils held considerably more carbon than they do today, but about 25% of that carbon was lost soon after the lands were initially cultivated. In large areas of

grazing lands, there is the possibility that the invasion and spread of woody vegetation (woody encroachment) is responsible for a significant net carbon sink at present (Chapter 10). The magnitude (and even sign) of this flux is uncertain, however, in part because some ecosystems lose carbon belowground (soils) as they accumulate it aboveground (woody vegetation), and in part because the invasion and spread of exotic grasses into semi-arid lands of the western United States are increasing the frequency of fires, reversing woody encroachment, and releasing carbon (Bradley *et al.*, in press).

The emissions of carbon from settled lands are largely considered in the chapters in Part II and in Chapter 14 of this report. Non-fossil carbon seems to be accumulating in trees in these lands, but the net changes in soil carbon are uncertain.

The only ecosystems that appear to release carbon to the atmosphere are the coastal waters. The estimated flux of carbon is close to zero (and difficult to determine) because the gross fluxes (from river transport, photosynthesis, and respiration) are large and variable in both space and time.

The average net fluxes of carbon expressed as Mg C ha^{-1} yr^{-1} in Table 1 are for comparative purposes. They show the relative flux density for different types of ecosystems. These annual fluxes of carbon are rarely determined with direct measurements of flux, however, because of the extreme variability of fluxes in time and space, even within a single ecosystem type. Extrapolating from a few isolated measurements to an estimate for the whole region's flux is difficult. Rather, the net changes are more often based on differences in measured stocks over intervals of 10 years, or longer (see Chapter 3), or are based on the large and rapid changes per hectare that are reasonably well documented for certain forms of management, such as the changes in carbon stocks that result from the conversion of forest to cultivated land. Thus, most of the flux estimates in the Table are long-term and large-area estimates.

Nevertheless, average flux density is one factor important in determining an ecosystem's role as a net source or sink for carbon. The other important factor is area. Permafrost wetlands, for example, are currently a small net sink for carbon. They cover a large area, however, hold large stocks of carbon, and thus have to potential to become a significant net source of carbon if the permafrost thaws with global warming (Smith *et al*,. 2005, Smith *et al.*, 2001, Osterkamp *et al.*, 1999, 2000). Forests clearly dominate the net sequestration of carbon in North America, although wetlands and settled lands have mean flux densities that are above average.

The two factors (flux density and area) demonstrate the level of management required to remove a significant amount of carbon from the atmosphere and keep it on land. Under current conditions, sequestration of 100 Mt C yr^{-1}, for example (about 5% of fossil fuel emissions from North America), requires management over hundreds of millions of hectares (e.g., the area presently in agriculture or forests) (Table 1). Enhancement of this terrestrial carbon sink through management would require considerable effort. Nevertheless, the cost (in $/metric ton CO_2) may be low relative to other options for

managing carbon. For example, forestry activities are estimated to have the potential to sequester 100–200 Mt C yr^{-1} in the United States at prices ranging from less than \$10/ton of CO_2 for improved forest management, to \$15/ton for afforestation, to \$30–50/ton for production of biofuels. Somewhat smaller sinks of 10–70 Mt C yr^{-1} might be sequestered in agricultural soils at low to moderate costs (\$3–30/ton CO_2). The maximum amounts of carbon that might be accumulated in forests and agricultural soils are not known, and thus the number of years these rates of sequestration might be expected to continue is also unknown. It seems unlikely that the amount of carbon currently held in forests and agricultural lands could double. Changes in climate will also affect carbon storage, but the net effect of management and climate is uncertain.

Despite the limited nature of carbon sequestration in offsetting the global emissions of carbon from fossil fuels, local and regional activities may, nevertheless, offset local and regional emissions of fossil carbon. This offset, as well as other co-benefits, may be particularly successful in urban and suburban systems (Chapter 14).

The effects and cost of managing aquatic systems are less clear. Increasing the area of wetlands, for example, would presumably sequester carbon; but it would also increase emissions of CH_4, countering the desired effect. Fertilization of coastal waters with iron has been proposed as a method for increasing oceanic uptake of CO_2, but neither the amount of carbon that might be sequestered nor the side effects are known (Chapter 15).

A few studies have estimated the potential magnitudes of future carbon sinks as a result of management (Chapters 10, 11). However, the contribution of management, as opposed to the environment, in today's sink is unclear (see Chapter 3), and for the future the relative roles of management and environmental change are even less clear. The two drivers might work together to enhance terrestrial carbon sinks, as seems to have been the case during recent decades (Prentice *et al.*, 2001) (Chapter 2). On the other hand, they might work in opposing directions. A worst-case scenario, quite possible, is one in which management will become ineffective in the face of large natural sources of carbon not previously experienced in the modern world. In other words, while management is likely to be essential for sequestering carbon, it may not be sufficient to preserve the current terrestrial carbon sink over North America, let alone to offset fossil fuel emissions.

At least one other observation about sequestering carbon in terrestrial and aquatic ecosystems should be mentioned. In contrast to the hundreds of millions of hectares that must be managed to sequester 100 Mt C annually, a few million hectares of forest fires can release an equivalent amount of carbon in a single year. This disparity in flux densities underscores the fact that a few million hectares are disturbed each year, while hundreds of millions of hectares are recovering from past disturbances. The natural cycling of carbon is large in comparison to net fluxes. The observation is relevant for carbon

management, because the cumulative effects of small managed net sinks to mitigate fossil fuel emissions will have to be understood, analyzed, monitored and evaluated in the context of larger, highly variable and uncertain sources and sinks in the natural cycle.

The major challenge for future research is quantification of the mechanisms responsible for current (and future) fluxes of carbon. In particular, what are the relative effects of management (including land-use change), environmental change, and natural disturbance in determining today's and tomorrow's sources and sinks of carbon? Will the current natural sinks continue, grow in magnitude, or reverse to become net sources? What is the role of soils in the current (and future) carbon balance (Davidson and Janssens, 2006)? What are the most cost-effective means of managing carbon?

Answering these questions will require two scales of measurement: (1) an expanded network of intensive research sites dedicated to understanding basic processes (e.g., the effects of management and environmental effects on carbon stocks), and (2) extensive national-level networks of monitoring sites, through which uncertainties in carbon stocks (inventories) would be reduced and changes, directly measured. Elements of these measurements are underway, but the effort has not yet been adequate for resolving these questions.

KEY UNCERTAINTIES AND GAPS IN UNDERSTANDING THE CARBON CYCLE OF NORTH AMERICA

- As mentioned above, the net flux of carbon resulting from woody encroachment and its inverse, woody elimination, is highly uncertain. Even the sign of the flux is in question.
- Rivers, lakes, dams, and other inland waters are mentioned in Chapter 15 as being a source of carbon, but they are claimed elsewhere to be a sink (Chapter 3). The sign of the net carbon flux attributable to erosion, transport, deposition, accumulation and decomposition is uncertain (e.g., Stallard, 1998; Lal, 2001; Smith *et al*., 2005).
- Several chapters cite studies that have attempted to quantify potential future carbon sinks in countries in North America, but no reference is made to estimates of future sources of carbon. Clearly, there are modeling studies that project large future carbon emissions, although these studies are largely global in scope (e.g., Cox *et al*., 2000; Jones *et al*., 2005). Are there no studies of future carbon sources and sinks for North America? Melting permafrost, in particular, is likely to increase emissions of carbon to the atmosphere, CH_4 as well as CO_2.
- The sum of land areas reported in these chapters is about 330 million ha larger than the area of North America (Table 1). The reason for this double-counting is unclear, but it implies a double counting of carbon stocks and, perhaps, current sinks, as well.

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Table 1. Ecosystems in North America: their areas, net annual fluxes of carbon, and their potential for sources (+) or sinks (–) in the future

Type of ecosystem	Area (10^6 ha)	Current mean flux density (Mg C ha^{-1} yr^{-1})	Current flux (Mt C yr^{-1})	Carbon stocks (Mt C)	Future potential flux (Mt C yr^{-1})
Agriculture	231	0.0	0±15[1]	18,500	–(50 to100) to +??
Grass, shrub and arid	558	–0.01	–6[2]	59,950	–34
Forests	771	–0.45	–350[3]	171,475	–(100 to 200) to+??
Permafrost wetlands	621[4]	⁻0.02	–14[5]	213,320	
Wetlands	246	–0.28	–70	220,000	
Settled lands	104	–0.31[6]	–32[6]	~1,000[6]	
Coastal waters	384	0.05	19		
Sum	2531[7]	–0.18[8]	–472[9]	684,245	
Total	2126[10.]				

1. Fossil fuel inputs to crop management are not included. Some of the C sequestration is occurring on grasslands as well as croplands, but the inventories do not separate these fluxes. The near-zero flux is for Canada and the United States only. Including Mexican croplands would likely change the flux to a net source because croplands are expanding in Mexico, and the carbon in biomass and soil is released to the atmosphere as native ecosystems are cultivated.
2. Fossil fuels are not included. The small net sink results from the Conservation Reserve Program in the United States Including Mexico is likely to change the net sink to a source because forests are being converted to grazing lands. Neither woody encroachment nor woody elimination (Bradley *et al.*, in press) is included in this estimate of flux because the uncertainties are so large.
3. Includes an annual sink of 67 Mt C yr^{-1} in wood products as well as a sink of 283 Mt C yr^{-1} in forested ecosystems.
4. Includes zones with isolated and sporadic permafrost.
5. This estimate is for peatlands (not mineral soils) in permafrost regions. The net flux for mineral soil permafrost areas is unknown. This estimate of flux may be high because it does not include the losses resulting from fires, but it may be low if mineral soils are also accumulating carbon in permafrost regions.
6. Urban trees only (does not include soil carbon).
7. Sum does not include coastal waters. The summed area is too high because an estimated 75×10^6 ha of permafrost peatlands in Canada are treed (and may be included in forest area as well as permafrost area). Nevertheless, another ~330×10^6 ha are double counted (United States forests on non-permafrost wetlands? Other wooded lands that are included as both forests and rangelands? Large areas of grasslands and shublands on non-permafrost lands within areas defined as sporadic or isolated permafrost? Inland waters?).
8. Weighted average; does not include coastal waters.
9. Does not include coastal waters. The total annual sink of 472 Mt C is lower than the estimate of 592 Mt C presented in Chapter 3 (Table 3-1). The largest difference results from the flux of carbon attributed to woody encroachment. Chapter 3 includes a sink of 120 Mt C yr^{-1}; Table 1, above, presents a net flux of zero (see note 2). Other differences between the two estimates include: (1) an additional sink in Table 1 of 14 Mt C yr^{-1} in permafrost wetlands; (2) an additional sink in Table 1 of 32 Mt C yr^{-1} in settled lands; and (3) a sink of 25 Mt C yr^{-1} in rivers and reservoirs that is included in Table 3-1 but not in Table 1. In addition, there are small differences in the estimates for agricultural lands and grasslands.
10. Areas (10^6 ha) (*The Times Atlas of the World*, 1990)

Globe	North America	Canada	United States	Mexico
14,900	2,126	992	936	197

Chapter 10. Agricultural and Grazing Lands

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KEY FINDINGS

- Agricultural and grazing lands (cropland, pasture, rangeland, shrublands, and arid lands) occupy 789 million ha (47% of the land area of North America) and contain 78.5±19.5 Gt C (17% of North American terrestrial carbon) in the soil alone.
- The emissions and sequestration of carbon on agricultural lands are mainly determined by two conditions: management and changes in the environment. The effects of converting forest and grassland to agricultural lands and of agricultural management (e.g., cultivation, conservation tillage) are reasonably well known and have been responsible for historic losses of carbon in Canada and the United States (and for current losses in Mexico); the effects of climate change or of elevated concentrations of atmospheric CO_2 are uncertain.
- Conservation-oriented management of agricultural lands (e.g., use of conservation tillage, improved cropping and grazing systems, reduced bare fallow, set-asides of fragile lands, and restoration of degraded soils) can significantly increase soil carbon stocks.
- Agricultural and grazing lands in the United States and Canada are currently near neutral with respect to their soil carbon balance, but agricultural and grazing lands in Mexico are likely losing carbon due to land use change. Although agricultural soils are estimated to be sequestering currently 6.4-15.9 Mt C yr^{-1}, the cultivation of organic soils releases 5.1-10.1 Mt C yr^{-1}. On-farm fossil fuel use and (30.9 Mt C yr^{-1}) and manufacture of agricultural inputs including fertilizer (6.4 Mt C yr^{-1}) yields a net source from the agricultural sector of 27-41 Mt C yr^{-1}.
- As much as 120 Mt C yr^{-1} may be accumulating through woody encroachment of arid and semi-arid lands of North America; this value is highly uncertain. Woody encroachment is generally

accompanied by decreased forage production and ongoing efforts to reestablish forage species are likely to reverse biomass carbon accumulation.

- Projections of future trends in agricultural land area and soil carbon stocks are unavailable or highly uncertain because of uncertainty in future land-use change and agricultural management practice.
- Annualized prices of \$15/tonne CO_2, would yield mitigation amounts of 168 Mt CO_2 yr^{-1} through agricultural soil C sequestration and 53 Mt CO_2 yr^{-1} from fossil fuel use reduction. At lower prices of \$5/tonne CO_2, the corresponding values would be 123 Mt CO_2 yr^{-1} and 32 Mt CO_2 yr^{-1}, respectively.
- Policies designed to suppress emissions of one greenhouse gas need to consider complex interactions to ensure that *net* emissions of total greenhouse gases are reduced. For example, increased use of fertilizer or irrigation may increase crop residues and carbon sequestration, but may stimulate emissions of CH_4 or N_2O.
- Many of the practices that lead to carbon sequestration and reduced CO_2 and CH_4 emissions from agricultural lands not only increase production efficiencies, but lead to environmental co-benefits, for example, improved soil fertility, reduced erosion and pesticide immobilization.
- An expanded network of intensive research sites is needed to better understand the effects of management on carbon cycling and storage in agricultural systems. An extensive national-level network of soil monitoring sites in which changes in carbon stocks are directly measured is needed to reduce the uncertainty in the inventory of agricultural and grazing land carbon. Better information about the spatial extent of woody encroachment, the amount and growth of woody biomass, and variation in impacts on soil carbon stocks would help reduce the large uncertainty of the carbon impacts of woody encroachment.

INVENTORY

Background

Agricultural and grazing lands (cropland, pasture, rangeland, shrublands, and arid lands[1]) occupy 47% of the land area in North America (59% in the United States, 70% in Mexico, and 11% in Canada), and contain 17% of the terrestrial carbon. Most of the carbon in these ecosystems is held in soils. Live vegetation in cropland generally contains less than 5% of total carbon, whereas vegetation in grazing lands contains a greater proportion (5–30%), but still less than that in forested systems (30–65%). Agricultural and grazing lands in North America contain 78.5±19.5 (±1 standard error) Gt C in the soil (Table 10-1). Significant increases in vegetation carbon stocks in some grazing lands have been observed and, together with soil carbon stocks from croplands and grazing lands, likely contribute significantly to

[1]We refer collectively to pasture, rangeland, shrublands, and arid lands as grazing lands since grazing is their primary use, even though not all of these lands are grazed.

the large North American terrestrial carbon sink (Houghton *et al.*, 1999; Pacala *et al.*, 2001; Eve *et al.*, 2002; Ogle *et al.*, 2003). These lands also emit greenhouse gases: fossil fuel use for on-farm machinery and buildings, for manufacture of agricultural inputs, and for transportation account for 3–5% of total CO_2 emissions in developed countries (Enquete Commission, 1995); activities on agricultural and grazing lands, like livestock production, animal waste management, biomass burning, and rice cultivation, emit 35% of global anthropogenic CH_4 (27% of United States, 31% of Mexican, and 27% of Canadian CH_4 emissions) (Mosier *et al.*, 1998b; CISCC, 2001; Matin *et al.*, 2004; EPA, 2006); and agricultural and grazing lands are the largest anthropogenic source of N_2O emissions (CAST, 2004; see Text Box 1). However, agricultural and grazing lands are actively managed and have the capacity to take up and store carbon. Thus improving management could lead to substantial reductions in CO_2 and CH_4 emissions and could sequester carbon to offset emissions from other lands or sectors.

Table 10-1. Carbon pools in agricultural and grazing lands in Canada, Mexico, and the United States; the area (M ha) for each climatic zone are in parentheses.

Carbon Dioxide Fluxes from Agricultural and Grazing Land

The main processes governing the carbon balance of agricultural and grazing lands are the same as for other ecosystems: the photosynthetic uptake and assimilation of CO_2 into organic compounds and the release of gaseous carbon through respiration (primarily CO_2 but also CH_4) and fire. Like other terrestrial ecosystems in general, for which CO_2 emissions are approximately two orders of magnitude greater than CH_4 emissions, carbon cycling in most agricultural and grazing lands is dominated by fluxes of CO_2 rather than CH_4. In agricultural lands, carbon assimilation is directed towards production of food, fiber, and forage by manipulating species composition and growing conditions (soil fertility, irrigation, etc.). Biomass, being predominantly herbaceous (i.e., non-woody), is a small, transient carbon pool (compared to forests) and hence soils constitute the dominant carbon stock. Cropland systems can be among the most productive ecosystems, but in some cases restricted growing season length, fallow periods, and grazing-induced shifts in species composition or production can reduce carbon uptake relative to that in other ecosystems. These factors, along with tillage-induced soil disturbances and removal of plant carbon through harvest, have depleted soil carbon stocks by 20–40% or more from pre-cultivated conditions (Davidson and Ackerman, 1993; Houghton and Goodale, 2004). Soil organic carbon stocks in grazing lands (see Text Box 2 for information on inorganic soil carbon stocks) have been depleted to a lesser degree than for cropland (Ogle *et al.*, 2004), and in some regions biomass has increased due to suppression of disturbance and subsequent woody encroachment (see Text Box 3). Woody encroachment is potentially a significant sink for atmospheric CO_2, but the magnitude of the sink is poorly constrained

(Houghton *et al.*, 1999; Pacala *et al.*, 2001). Since woody encroachment leads to decreased forage production, management practices are aimed at reversing it, with consequent reductions in biomass carbon. Disturbance-induced increases in decomposition rates of aboveground litter and harvest removal of some (30–50% of forage in grazing systems, 40–50% in grain crops) or all (e.g., corn for silage) of the aboveground biomass, have drastically altered carbon cycling within agricultural lands and thus the sources and sinks of CO_2 to the atmosphere.

Much of the carbon lost from agricultural soil and biomass pools can be recovered with changes in management practices that increase carbon inputs, stabilize carbon within the system, or reduce carbon losses, while still maintaining outputs of food, fiber, and forage. Increased production, increased residue C inputs to the soil, and increased organic matter additions have reversed historic soil C losses in long-term experimental plots (e.g., Buyanovsky and Wagner, 1998). Across Canada and the United States, mineral soils have been sequestering 0.1and 6.5–16 Mt C yr^{-1} (Smith *et al.*, 1997; Smith *et al.*, 2001b; Ogle *et al.*, 2003), respectively, largely through increased production and improved management practices on annual cropland (Fig. 10-1, Table 10-2). Conversion of agricultural land to grassland, like under the Conservation Reserve Program in the United States (6 Mt C yr^{-1} on 14 M ha of land), and afforestation have also sequestered carbon in agricultural and grazing lands. In contrast, cultivation of organic soils (e.g., peat-derived soils) is releasing an estimated 0.1 and 5-10 Mt C yr^{-1} from soils in Canada and the United States (Matin *et al.*, 2004; Ogle *et al.*, 2003). Compared with other systems, the high productivity and management-induced disturbances of agricultural systems promote movement and redistribution (through erosion, runoff and leaching) of organic and inorganic carbon, sequestering potentially large amounts of carbon in sediments and water (Raymond and Cole, 2003; Smith *et al.* 2005; Yoo *et al.*, 2005). However, the net impact of soil erosion on carbon emissions to the atmosphere remains highly uncertain.

Figure 10-1. North American agricultural and grazing land CO_2 (left side) and methane (right side), adjusted for global warming potential.

Table 10-2. North American agricultural and grazing land carbon fluxes for the years around 2000

Production, delivery, and use of field equipment, fertilizer, seed, pesticides, irrigation water, and maintenance of animal production facilities contribute 3–5% of total fossil fuel CO_2 emissions in developed countries (Enquete Commission, 1995). On-farm fossil fuel emissions together with manufacture of fertilizers and pesticides contribute emissions of 32.7 Mt C yr^{-1} within the United States (Lal *et al.*, 1998) and 4.6 Mt C yr^{-1} in Canada (Sobool and Kulshreshtha, 2005) (Table 10-2). Energy

consumption for heating and cooling high intensity animal production facilities is among the largest CO_2 emitters within the agricultural sector (Enquete Commission, 1995).

Much of the ammonia production and urea application (U.S.: 4.3 Mt C yr^{-1}; Mexico: 0.4 Mt C yr^{-1}; Canada: 1.7 Mt C yr^{-1}) and phosphoric acid manufacture (U.S.: 0.4 Mt C yr^{-1}; Mexico: 0.2 Mt C yr^{-1}; Canada: not reported) are devoted to agricultural uses.

Methane Fluxes from Agricultural and Grazing Lands

Cropland and grazing land soils act as both sources and sinks for atmospheric CH_4. Methane formation is an anaerobic process and is most significant in waterlogged soils, like those under paddy rice cultivation (U.S.: 0.25 Mt CH_4-C yr^{-1}; Mexico: 0.01 Mt CH_4-C yr^{-1}; Canada: negligible, not reported; Table 10-2). Methane is also formed by incomplete biomass combustion of crop residues (U.S.: 0.03 Mt CH_4-C yr^{-1}; Mexico: <0.01 Mt CH_4-C yr^{-1}; Canada: negligible, not reported; Table 10-2). Methane oxidation in soils is a global sink for about 5% of CH_4 produced annually and is mainly limited by CH_4 diffusion into the soil. However, intensive cropland management tends to reduce soil methane consumption relative to forests and extensively grazing lands (CAST, 2004). Management-induced changes in CH_4-C fluxes have a smaller impact on terrestrial carbon cycling than changes in CO_2-C fluxes (Table 10-2), but relatively greater radiative forcing for CH_4 amplifies the impact of increasing atmospheric CH_4 concentrations on net radiative forcing (Fig. 10-1). Recent research has shown that live plant biomass and litter produce substantial amounts of CH_4, potentially making plants as large a source of CH_4 as livestock (Keppler *et al.*, 2006). If this is the case, activities that increase plant biomass—and sequester CO_2—may lead to increased CH_4 production (Keppler *et al.*, 2006).

Methane Fluxes from Livestock

Enteric fermentation (the process of organic matter breakdown by gut flora within the gastrointestinal tract of animals, particularly ruminants) allows for the digestion of fibrous materials by livestock, but the extensive fermentation of the ruminant diet requires 5–7% of the dietary gross energy to be belched out as CH_4 to sustain the anaerobic processes (Johnson and Johnson, 1995). Methane emissions from livestock contribute significantly to total CH_4 emissions in the United States (5.8 Mt CH_4-C yr^{-1}, 21% of total U.S. CH_4 emissions), Canada (0.6 Mt CH_4-C yr^{-1}, 22% of total) (Sobool and Kulshreshtha, 2005), and Mexico (3.7 Mt CH_4-C yr^{-1}, 27% of total) with the vast majority of enteric CH_4 emissions are from beef (72%) and dairy cattle (23%) (Table 10-2). Emissions from ruminants are tightly coupled to feed consumption, since CH_4 emission per unit of feed energy is relatively constant, except for feedlot cattle with diets high in cereal grain contents, for which the fractional loss falls to one-third to one-half of normal rates

(Johnson and Johnson, 1995). Between 1990 and 2002, CH_4 emissions from enteric fermentation fell 2% in the United States but increased by 20% in Canada (EPA, 2000; Matin *et al.*, 2004).

Methane emissions during manure storage (U.S.: 1.9 Mt CH_4 yr^{-1}; Mexico: 0.06 Mt CH_4 yr^{-1}; Canada: 0.3 Mt CH_4 yr^{-1}) are governed by the amount of degradable organic matter, degree of anoxia, storage temperature, and duration of storage. Unlike enteric CH_4, the major sources of manure CH_4 emissions in the United States are from swine (44%) and dairy cattle (39%). Manure CH_4 production is greater for production systems with anoxic lagoons, largely anoxic pits, or manure handled or stored as slurry. Between 1990 and 2002, CH_4 emissions from manure management increased 25% in the United States and 21% in Canada (EPA, 2000; Matin *et al.*, 2004).

DRIVERS AND TRENDS

The extent to which agricultural options will contribute to greenhouse gas mitigation will largely depend on government policy decisions, but mitigation opportunities will also be constrained by technological advances and changing environmental conditions (see discussion below). Estimates from national inventories suggest that U.S. and Canadian agricultural soils are currently near neutral or small net sinks for CO_2, which has occurred as a consequence of changing management (e.g., reduced tillage intensity) and government programs designed for purposes other than greenhouse gas mitigation (e.g., soil conservation, commodity regulation). However, to realize the much larger potential for soil carbon sequestration (see section below) and for significant reductions in CH_4 (and N_2O) emissions, specific policies targeted at greenhouse gas reductions are required. It is generally recognized that farmers (and other economic actors) are, as a group, 'profit-maximizers,' which implies that to change from current practices to ones that reduce net emissions, farmers will incur additional costs (termed 'opportunity cost'). Hence, where the incentives (e.g., carbon offset market payments, government subsidies) to adopt new practices exceed the opportunity costs, farmers will adopt new practices. Crop productivity, production input expenses, marketing costs, etc. (which determine profitability) vary widely within (and between) countries. Thus, the payment needed to achieve a unit of emission reduction will vary, among and within regions. In general, each successive increment of carbon sequestration or emission reduction comes at a progressively higher cost (this relationship is often shown in the form of an upward bending marginal cost curve).

The interaction of changes in technological and environmental conditions, including crop growth improvements, impacts of CO_2 increase, N deposition, and climate change, will shape future trends in greenhouse gas emissions and mitigation from agricultural and grazing lands. A continuation of the yield increases seen in the past several decades for agricultural crops (Reilly and Fuglie, 1996) would tend to enhance the potential for soil C sequestration (CAST, 2004). Similarly, increased plant growth due to

higher concentrations of CO_2 (and N deposition) has been projected to boost carbon uptake on agricultural (and other) lands, offsetting some or all of the climate-change induced reductions in productivity projected in some regions of North America (NAS, 2001). However, recent syntheses from field-scale FACE (Free-Air Carbon dioxide Enrichment) studies of croplands (Long *et al.*, 2006) and grasslands (Nowak *et al.*, 2004) suggest that the growth enhancement from CO_2 fertilization may be much less than previously thought. Feedbacks between temperature and soil carbon stocks could counteract efforts to reduce greenhouse gases via carbon sequestration within agricultural ecosystems. Increased temperatures tend to increase the rate of biological processes—including plant respiration and organic matter decay and CO_2 release by soil organisms—particularly in temperate climates that prevail across most of North America. Because soil carbon stocks, including those in agricultural lands, contain such large amounts of carbon, small percentage increases in rate of soil organic matter decomposition could lead to substantially increased emissions (Jenkinson *et al.*, 1991; Cox *et al.*, 2000). There is currently a scientific debate about the relative temperature sensitivity of the different constituents making up soil organic matter (e.g., Kätterer *et al.*, 1998; Giardina and Ryan, 2000; Ågren and Bosatta, 2002; Knorr *et al.*, 2005), reflecting uncertainty in the possible degree and magnitude of climate change feedbacks. Despite this uncertainty, the potential for climate and other environmental feedbacks to influence the carbon balance of agricultural systems by perturbing productivity (and carbon input rates) and organic matter turnover, and potentially soil N_2O and CH_4 fluxes, cannot be overlooked.

OPTIONS FOR MANAGEMENT

Carbon Sequestration

Agricultural and grazing land management practices capable of increasing carbon inputs or decreasing carbon outputs, while still maintaining yields, can be divided into two classes: those that impact carbon inputs, and those that affect carbon release through decomposition and disturbance. Reversion to native vegetation or setting agricultural land aside as grassland, such as in the Canadian Prairie Cover Program and the U.S. Conservation Reserve Program, can increase the proportion of photosynthesized carbon retained in the system and sequester carbon in the soil[2] (Post and Kwon, 2000; Follett *et al.*, 2001b) (Fig. 10-2). In annual cropland, improved crop rotations, yield enhancement measures, organic amendments, cover crops, improved fertilization and irrigation practices, and reduced bare fallow tend to increase productivity and carbon inputs, and thus soil carbon stocks (Lal *et al.*, 1998;

[2]The bulk of carbon sequestration potential in agricultural and grazing lands is restricted to soil carbon pools, though carbon can be sequestered in woody biomass in agroforestry systems (Sheinbaum and Masera, 2000). Woody encroachment on grasslands can also store substantial amounts of carbon in biomass, but the phenomenon is neither well-controlled nor desirable from the standpoint of livestock production, since it results in decreased forage productivity, and the impacts on soil carbon pools are highly variable and poorly understood.

Paustian *et al.*, 1998; VandenBygaart *et al.*, 2003) (Fig. 10-2). Tillage, traditionally used for soil preparation and weed control, disturbs the soil and stimulates decomposition and loss of soil carbon. Practices that substantially reduce (reduced-till) or eliminate (no-till) tillage-induced disturbances are being increasingly adopted and generally increase soil carbon stocks while maintaining or enhancing productivity levels (Paustian *et al,.* 1997; Ogle *et al.*, 2003) (Fig. 10-2). Estimates of the technical potential for annual cropland soil carbon sequestration are on the order of 50–100 Mt C yr^{-1} in the United States (Lal *et al.*, 2003; Sperow *et al.*, 2003) and approximately 5 Mt C yr^{-1} in Canada (Boehm *et al.*, 2004).

Figure 10-2. Relative soil carbon following implementation of new agricultural or grassland management practices.

Within grazing lands, historical overgrazing has substantially reduced productive capacity in many areas, leading to loss of soil carbon stocks (Conant and Paustian, 2002) (Fig. 10-2). Conversely, improved grazing management and production inputs—like fertilizer, adding (N-fixing) legumes, organic amendments, and irrigation—can increase productivity, carbon inputs, and soil carbon stocks, potentially storing 0.44 Mt C yr^{-1} in Canada (Lynch *et al.*, 2005) and as much as 33.2 Mt C yr^{-1}in the United States (Follett *et al.*, 2001a). Such improvements will carry a carbon cost, particularly fertilization and irrigation since their production and implementation require the use of fossil fuels.

Fossil Fuel-Derived Emission Reductions

The efficiency with which on-farm (from tractors and machinery) and off-farm (from production of agricultural input) energy inputs are converted to agricultural products varies several-fold (Lal, 2004). Where more energy-efficient practices can be substituted for less efficient ones, fossil fuel CO_2 emissions can be reduced (Lal, 2004). For example, converting from conventional plowing to no-tillage can reduce on-farm fossil fuel emissions by 25–80% (Frye, 1984; Robertson *et al.*, 2000) and total fossil fuel emissions by 14–25% (West and Marland, 2003). Substitution of legumes for mineral nitrogen can reduce energy input by 15% in cropping systems incorporating legumes (Pimentel *et al.*, 2005). More efficient heating and cooling (e.g., better building insulation) could reduce CO_2 emissions associated with housed animal (e.g., dairy) facilities. Substitution of crop-derived for fossil fuels could decrease net emissions.

Energy intensity (energy per unit product) for the U.S. agricultural sector has declined since the 1970s (Paustian *et al.*, 1998). Between 1990 and 2000, fossil fuel emissions on Canadian farms increased by 35% (Sobool and Kulshreshtha, 2005).

Methane Emission Reduction

Reducing flood duration and decreasing organic matter additions to paddy rice fields can reduce CH_4 emissions. Soil amendments such as ammonium sulfate and calcium carbide inhibit CH_4 formation. Coupled with adoption of new rice cultivars that favor lower CH_4 emissions, these management practices could reduce CH_4 emission from paddy rice systems by as much as 40% (Mosier *et al.*, 1998b).

Biomass burning is uncommon in most Canadian and U.S. crop production systems; less than 3% of crop residues are burned annually in the United States (EPA, 2006). Biomass burning in conjunction with land clearing and with subsistence agriculture still occurs in Mexico, but these practices are declining. The primary path for emission reduction is reducing residue burning (CAST, 2004).

Refinement of feed quality, feed rationing, additives, and livestock production efficiency chains can all reduce CH_4 emissions from ruminant livestock with minimal impacts on productivity or profits (CAST, 2004). Boadi *et al.* (2004) review several examples of increases in energy intensity. Wider adoption of more efficient practices could reduce CH_4 production from 5–8% to 2–3% of gross feed energy (Agriculture and Agri-Food Canada, 1999), reducing CH_4 emissions by 20–30% (Mosier *et al.*, 1998b).

Methane emissions from manure storage are proportional to duration of storage under anoxic conditions. Handling solid rather than liquid manure, storing manure for shorter periods of time, and keeping storage tanks cool can reduce emissions from stored manure (CAST, 2004). More important, capture of CH_4 produced during anaerobic decomposition of manure—in covered lagoons or small- or large-scale digesters—can reduce emissions by 70–80% (Mosier *et al.*, 1998b). Use of digester systems is spreading in the United States, with 50 digesters currently in operation and 60 systems in construction or planned (NRCS, 2005). Energy production using CH_4 captured during manure storage will reduce energy demands and associated CO_2 emissions.

Environmental Co-benefits from Carbon Sequestration and Emission Reduction Activities

Many of the practices that lead to carbon sequestration and reduced CO_2 and CH_4 emissions not only increase production efficiencies but also lead to environmental co-benefits. Practices that sequester carbon in agricultural and grazing land soils improve soil fertility, buffering capacity, and pesticide immobilization (Lal, 2002; CAST, 2004). Increasing soil carbon content makes the soil more easily workable and reduces energy requirements for field operations (CAST, 2004). Decreasing soil disturbance and retaining more surface crop residues enhance water infiltration and prevent wind and water erosion, improving air quality. Increased water retention plus improved fertilizer management reduces nitrogen losses and subsequent NO_3^- leaching and downstream eutrophication.

Economics and Policy Assessment

Policies for agricultural mitigation activities can range from transfer payments (as subsidies, tax credits, etc.), to encourage greenhouse gas mitigating practices (or taxes or penalties to discourage practices with high emissions), to emission offset trading in a free market-based system with governmental sanction. Currently the policy context of the North American three countries differs greatly. Canada and the United States are both Annex 1 (developed countries) within the UNFCCC, but Canada is obligated to mandatory emission reductions as a party to the Kyoto Protocol, while the United States currently maintains a national, voluntary emission reduction policy outside of Kyoto. Mexico is a non-Annex 1 (developing) country and thus is not currently subject to mandatory emission reductions under Kyoto.

At present there is relatively little practical experience upon which to judge the costs and effectiveness of agricultural mitigation activities—governments are still in the process of developing policies and, moreover, the economics of various mitigation activities will only be known when there is a significant economic incentive for emission reductions, e.g., through regulatory emission caps or government-sponsored bids and contracts. However, several economic analyses have been performed in the United States, using a variety of models (e.g., McCarl and Schneider, 2001; Antle *et al.*, 2003; Lewandrowski *et al.*, 2004). Most studies have focused on carbon sequestration, and less work has been done on the economics of reducing CH_4 and N_2O emissions. While results differ between models and for different parts of the country, some preliminary conclusions have been drawn (see Boehm *et al.*, 2004; CAST, 2004).

- Significant amounts (10–70 Mt yr^{-1}) of carbon sequestration in soils can be achieved at low to moderate costs ($10–100 per metric ton of carbon).
- Mitigation practices that maintain the primary income source (i.e., crop/livestock production), e.g., conservation tillage, pasture improvement, have a lower cost per ton sequestered carbon compared with practices where mitigation would be a primary income source (foregoing income from crop and/or livestock production), such as land set-asides, even if the latter have a higher biological sequestration potential.
- With higher energy prices, major shifts in land use in favor of energy crops and afforestation may occur at the expense of annual cropland and pasture.
- Policies based on per-ton payments (for carbon actually sequestered) are more economically efficient than per-hectare payments (for adopting specific practices – see Antle *et al*., 2003), although the

former have a higher verification cost (i.e., measuring actual carbon sequestered versus measuring adoption of specific farming practices on a given area of land).

A recent study commissioned by the U.S. Environmental Protection Agency (EPA 2005), estimated economic potential for some agricultural mitigation options, assuming constant price scenarios for 2010–2110, where the price represents the incentive required for the mitigation activity. Annualized prices of \$15/ton of CO_2 would yield mitigation amounts of 168 Mt CO_2 per year through agricultural soil carbon sequestration and 53 Mt CO_2 per year from fossil fuel use reduction (compare with estimated U.S. national ecosystem carbon sink of 1760 Mt CO_2 per year). At lower prices of \$5/ton CO_2, the corresponding values would be 123 Mt CO_2 per year (for soil sequestration) and 32 Mt CO_2 per year (for fossil fuel reduction), respectively, reflecting the effect of price on the supply of mitigation activities.

Other Policy Considerations

Agricultural mitigation of CO_2 through carbon sequestration and emission reductions for CH_4 (and N_2O), differ in ways that impact policy design and implementation. Direct emission reductions of CH_4 and CO_2 from fossil fuel use are considered 'permanent' reductions, while carbon sequestration is a 'non-permanent' reduction, in that carbon stored through conservation practices could potentially be re-emitted if management practices revert back to the previous state or otherwise change so that the stored carbon is lost. This *permanence* issue applies to all forms of carbon sinks. In addition, a given change in management (e.g., tillage reduction, pasture improvement, afforestation) will stimulate carbon storage for a finite duration. For many practices, soil carbon storage will tend to level off at a new steady state level after 15–30 years, after which there is no further accumulation of carbon (West *et al*., 2004). Thus, to maintain these higher stocks, the management practices will need to be maintained. Key implications for policy are that the value of sequestered carbon will be discounted compared to direct emission reductions to compensate for the possibility of future emissions. Alternatively, long-term contracts will be needed to build and maintain C stocks, which will tend to increase the price per unit of sequestered carbon. However, even temporary storage of carbon has economic value (CAST, 2004), and various proposed concepts of leasing carbon storage or applying discount rates could accommodate carbon sequestration as part of a carbon offset trading system (CAST, 2004). In addition, switching to practices that increase soil carbon (and hence improve soil fertility) could be more profitable to farmers in the long-run, so that additional incentives to maintain the practices once they become well established may not be necessary (Paustian *et al.,* 2006).

Another policy issue relating to carbon sequestration is *leakage* (also termed 'slippage' in economics), whereby mitigation actions in one area (e.g., geographic region, production system) stimulate

additional emissions elsewhere. For forest carbon sequestration, leakage is a major concern—for example, reducing harvest rates in one area (thereby maintaining higher biomass carbon stocks) can stimulate increased cutting and reduction in stored carbon in other areas, as was seen with the reduction in harvesting in the Pacific Northwest during the 1990s (Murray *et al.*, 2004). Preliminary studies suggest that leakage is of minor concern for agricultural carbon sequestration, since most practices would have little or no effect on the supply and demand of agricultural commodities. However, there are uncertain and conflicting views on whether land-set asides—where land is taken out of agricultural production, such as the Conservation Reserve Program in the United States, might be subject to significant leakage.

A further question, relevant to policies for carbon sequestration, is how practices for conserving carbon affect emissions of other greenhouse gases. Of particular importance is the interaction of carbon sequestration with N_2O emission, because N_2O is such a potent greenhouse gas (Robertson and Grace, 2004; Six *et al.*, 2004; Gregorich *et al.*, 2005). (See Text Box 4). In some environs, carbon-sequestration practices, such as reduced tillage, can stimulate N_2O emissions thereby offsetting part of the benefit; elsewhere, carbon-conserving practices may suppress N_2O emissions, amplifying the net benefit (Smith *et al.*, 2001a; Smith and Conen, 2004; Conant *et al.*, 2005; Helgason *et al.*, 2005).

Similarly, carbon-sequestration practices might affect emissions of CH_4, if the practice, such as increased use of forages in rotations, leads to higher livestock numbers. These examples demonstrate that policies designed to suppress emission of one greenhouse gas need to also consider complex interactions to ensure that *net* emissions of total greenhouse gases are reduced.

A variety of other factors will affect the willingness of farmers to adopt greenhouse gas reducing practices and the efficacy of agricultural policies, including perceptions of risk, information and extension efforts, technological developments and social and ethical values (Paustian *et al.*, 2006) Many of these factors are difficult to incorporate into traditional economic analyses. Pilot mitigation projects, along with additional research using integrated ecosystem and economic assessment approaches (e.g., Antle *et al.*, 2001), will be needed to get a clearer picture of the actual potential of agriculture to contribute to greenhouse gas mitigation efforts.

RESEARCH AND DEVELOPMENT NEEDS

Expanding the network of intensive research sites dedicated to understanding basic processes, coupled with national-level networks of soil monitoring/validation sites could reduce inventory uncertainty and contribute to attributing changes in ecosystem carbon stocks to changes in land management (see Bellamy *et al.*, 2005). Expansion of both networks should be informed by knowledge about how different geographic areas and ecosystems contribute to uncertainty and the likelihood that reducing uncertainty could inform policy decisions. For example, changes in ecosystem carbon stocks due

to woody encroachment on grasslands constitute one of the largest, but least certain, aspects of terrestrial carbon cycling in North America (Houghton *et al.*, 1999; Pacala *et al.*, 2001). Better information about the spatial extent of woody encroachment, the amount and growth of woody biomass, and variation in impacts on soil carbon stocks would help reduce that uncertainty. Identifying location, cause, and size of this sink could help identify practices that may promote continued sequestration of carbon and would constrain estimates of carbon storage in other lands, possibly helping identify other policy options. Uncertainty in land use, land use change, soil carbon responses to management (e.g., tillage) on particular soils, and impacts of cultivation on soil carbon stocks (e.g., impacts of erosion) are the largest contributors to uncertainty in the Canadian and U.S. national agricultural greenhouse gas inventories (Ogle *et al.*, 2003; VandenBygaart *et al.*, 2004). Finally, if the goal of a policy instrument is to reduce greenhouse gas emissions, net impacts on CO_2, CH_4, and N_2O emissions, which are not as well understood, should be considered.

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[START OF TEXT BOX 1]

Nitrous oxide (N_2O) emissions from agricultural and grazing lands

Nitrous oxide (N_2O) is the most potent greenhouse gas in terms of global warming potential, with a radiative forcing 296 times that of CO_2 (IPCC, 2001). Agricultural activities that add mineral or organic nitrogen—fertilization, plant N_2 fixation, manure additions, etc.—augment naturally occurring N_2O emissions from nitrification and denitrification by 0.0125 kg N_2O per kg N applied (Mosier *et al.*, 1998a). Agriculture contributes significantly to total global N_2O fluxes through soil emissions (35% of total global emissions), animal waste handling (12%), nitrate leaching (7%), synthetic fertilizer application (5%), grazing animals (4%), and crop residue management (2%). Agriculture is the largest source of N_2O in the United States (78% of total N_2O emissions), Canada (59%), and Mexico (76%).

[END OF TEXT BOX 1]

[START OF TEXT BOX 2]

Inorganic soil carbon in agricultural and grazing ecosystems

Inorganic carbon in the soil is comprised of primary carbonate minerals, such as calcite ($CaCO_3$) or dolomite [$CaMg(CO_3)_2$], or secondary minerals formed when carbonate (CO_3^{2-}), derived from soil CO_2, combines with base cations (e.g., Ca^{2+}, Mg^{2+}) and precipitates within the soil profile in arid and semi-arid ecosystems. Weathering of primary carbonate minerals in humid regions is a source of CO_2, whereas formation of secondary carbonates in drier areas is a sink for CO_2; however, the magnitude of either flux is highly uncertain. Agricultural liming involves addition of primary carbonate minerals to the acid soils to increase the pH. In the United States, about 1 Mt C yr^{-1} is emitted from liming (EPA, 2006).

[END OF TEXT BOX 2]

[START OF TEXT BOX 3]

Impacts of woody encroachment into grasslands on ecosystem carbon stocks

Encroachment of woody species into grasslands—caused by overgrazing-induced reduction in grass biomass and subsequent reduction or elimination of grassland fires—is widespread in the United States and Mexico, decreases forage production, and is unlikely to be reversed without costly mechanical intervention (Van Auken, 2000). Encroachment of woody species into grassland tends to increase biomass carbon stocks by 1 Mg C ha^{-1} yr^{-1} (Pacala *et al.*, 2001), with estimated net sequestration of 0.12–0.13 Gt C yr^{-1} in encroaching woody biomass (Houghton *et al.*, 1999; Pacala *et al.*, 2001). In response to woody encroachment, soil carbon stocks can significantly increase or decrease, thus predicting impacts on soil carbon or ecosystem carbon stocks is very difficult (Jackson *et al.*, 2002).

[END OF TEXT BOX 3]

[START OF TEXT BOX 4]

Agricultural and grazing land N_2O emission reductions

When mineral soil nitrogen content is increased by nitrogen additions (i.e., fertilizer), a portion of that nitrogen can be transformed to N_2O as a byproduct of two microbiological processes (nitrification and denitrification) and lost to the atmosphere. Coincidental introduction of large amounts of easily decomposable organic matter and NO_3^- from either a plow down of cover crop or manure addition greatly stimulates denitrification under wet conditions (Peoples *et al.*, 2004). Some practices intended to sequester atmospheric carbon in soil could prompt increases in N_2O fluxes. For example, reducing tillage intensity tends to increase soil moisture, leading to increased N_2O fluxes, particularly in wetter environments (Six *et al.*, 2004). Synchronizing organic amendment applications with plant nitrogen uptake and minimizing manure storage under anoxic conditions can reduce N_2O emissions by 10–25% and will increase nitrogen use efficiency which can decrease indirect emissions (in waterways) by 5–20% (CAST, 2004).

[END OF TEXT BOX 4]

1

Table 10-1. Carbon pools in agricultural and grazing lands in Canada, Mexico, and the United States; the area (M ha) for each climatic zone are in parentheses. Current soil carbon stocks are secondary quantities derived from an initial starting point of undisturbed native ecosystems carbon stocks, which were quantified using the intersection of MODIS-IGBP[a] land cover types (Friedl *et al.*, 2002) and mean soil carbon contents to 1-m depth from Sombroek *et al.* (1993), spatially arrayed using Food and Agriculture Organization soil classes (ISRIC, 2002), and summed by climate zone. These undisturbed native ecosystem carbon stock values were then multiplied by soil carbon loss factors for tillage- and overgrazing-induced losses (Nabuurs *et al.*, 2004; Ogle *et al.*, 2004) to estimate current soil carbon stocks (see Fig. 10-2).

Practice	Temperate dry[b,c]	Temperate wet	Tropical dry	Tropical wet	Total
			Gt C		
			Agricultural lands		
Canada	1.79±0.35 (17.3)	1.77±0.36 (22.1)	–	–	3.60±0.77 (39.4)
Mexico	–	–	0.24±0.06 (3.9)	0.53±0.14 (10.2)	0.81±0.22 (14.1)
United States	3.31±0.74 (34.8)	8.66±2.18 (108.4)	0.35±0.08 (5.6)	1.53±0.33 (28.4)	14.05±3.20 (177.1)
Total	5.16±1.07 (52.1)	10.57±2.42 (130.5)	0.61±0.14 (9.5)	2.18±0.54 (38.6)	18.5±4.16 (230.6)
			Grazing lands		
Canada	2.17±0.55 (18.4)	9.49±1.27 (40.8)	–	–	11.66±4.88 (59.2)
Mexico	–	–	7.20±1.62 (99.1)	2.19±0.58 (20.3)	9.99±2.60 (119.4)
United States	16.89±3.62 (209.9)	5.67±1.39 (55.0)	4.26±0.98 (68.1)	4.30±0.89 (46.7)	32.88±7.18 (379.7)
Total	19.34±4.27 (228.3)	21.07±5.80 (95.8)	12.59±2.73 (167.1)	6.94±1.86 (67.0)	59.95±14.65 (558.2)

[a]Cropland area was derived from the IGBP cropland land cover class plus the area in the cropland/natural vegetation IGBP class in Mexico and one-half of the area in the cropland/natural vegetation IGBP class in Canada and the United States. Grazing land area includes IGBP woody savannas, savannas, and grasslands in all three countries, plus open shrubland in Mexico and open shrublands not in Alaska in the United States

[b]Temperate zones are those located above 30º latitude. Tropical zones (<30º latitude) include subtropical regions.

[c]Dry climates were defined as those where the ratio of mean annual precipitation (MAP) to potential evapotranspiration (PET) is less than 1; in wet areas, MAP/PET >1.

1

Table 10-2. North American agricultural and grazing land carbon fluxes for the years around 2000.
All units are in Mt C yr^{-1}. Negative numbers (in parentheses) indicate net flux from the atmosphere to soil and biomass carbon pools. Unless otherwise noted, data are from Canadian (Matin *et al.*, 2004) and U.S. (EPA, 2006) National Inventories and from the second Mexican National Communication (CISCC, 2001). Values are for 2003 for United States and Canada and 1998 for Mexico. A factor of 12/44 was used convert from CO_2 to carbon and a factor of 12/16 to convert CH_4 to carbon.

	Canada	Mexico	United States	Total
			Mt C yr^{-1}	
CO_2				
On-farm fossil fuel use	2.9[a]	ND	28[b]	30.9
Fertilizer manufacture	1.7	ND	4.7	6.4
Mineral soil carbon sequestration	(0.1)	ND	(6.5) – (16)	(6.4) – (15.9)
Organic soil cultivation	0.1	ND	5–10	5.1 – 10.1
Woody encroachment	ND	ND	(120)[c]	(120)
Total	4.6	ND	(98.3) – (83.8)	(93.7) – (79.2)
CH_4				
Rice production	0	0.011	0.25	0.26
Biomass burning	<0.01	<0.01	0.03	0.05
Livestock	0.62	1.48	3.67	5.77
Manure	0.18	0.05	1.28	1.51
Total	0.82	1.54	5.23	7.59

ND = no data reported.
[a]From Sobool and Kulshreshtha (2005).
[b] From Lal *et al.* (1998).
[c]From Houghton *et al.* (1999).

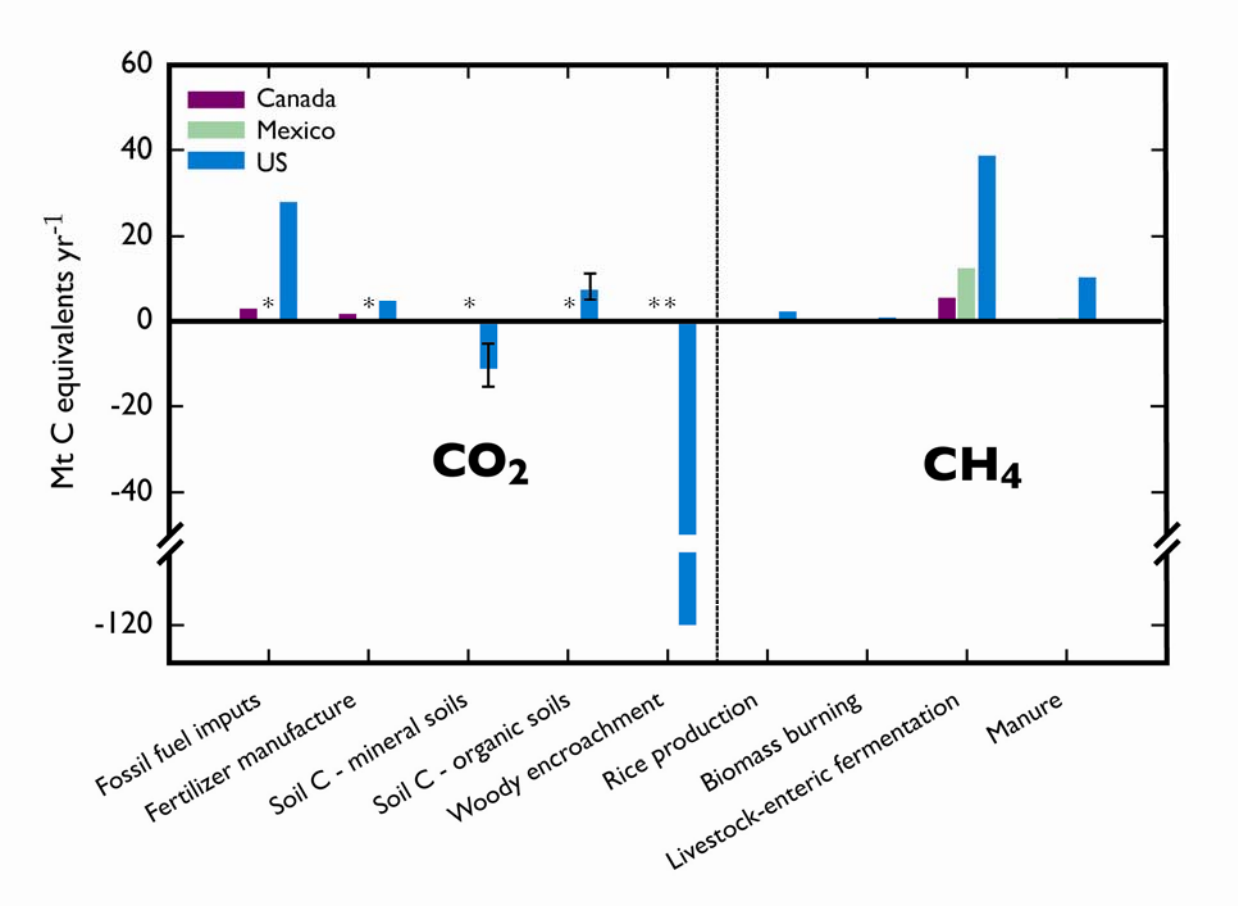

Fig. 10-1. North American agricultural and grazing land CO_2 (left side) and methane (right side), adjusted for global warming potential. All units are in Mt C-equivalent yr^{-1} for years around 2000. Negative values indicate net flux from the atmosphere to soil and biomass carbon pools (i.e., sequestration). All data are from Canadian (Matin *et al.*, 2004) and U.S. (EPA, 2006) National Inventories and from the second Mexican National Communication (CISCC, 2001), except for Canadian [from Kulshreshtha *et al.* (2000)] and U.S. fossil fuel inputs [from Lal *et al.* (1998)] and woody encroachment [from Houghton *et al.* (1999)]. Values are for 2003 for Canada, 1998 for Mexico, and 2004 for the United States. A global warming potential of 23 for methane was used to convert emissions of CH_4 to CO_2 equivalents (IPCC, 2001) and a factor of 12/44 to convert from CO_2 to carbon. Asterisks indicate unavailable data. Data ranges are indicated by error bars where available.

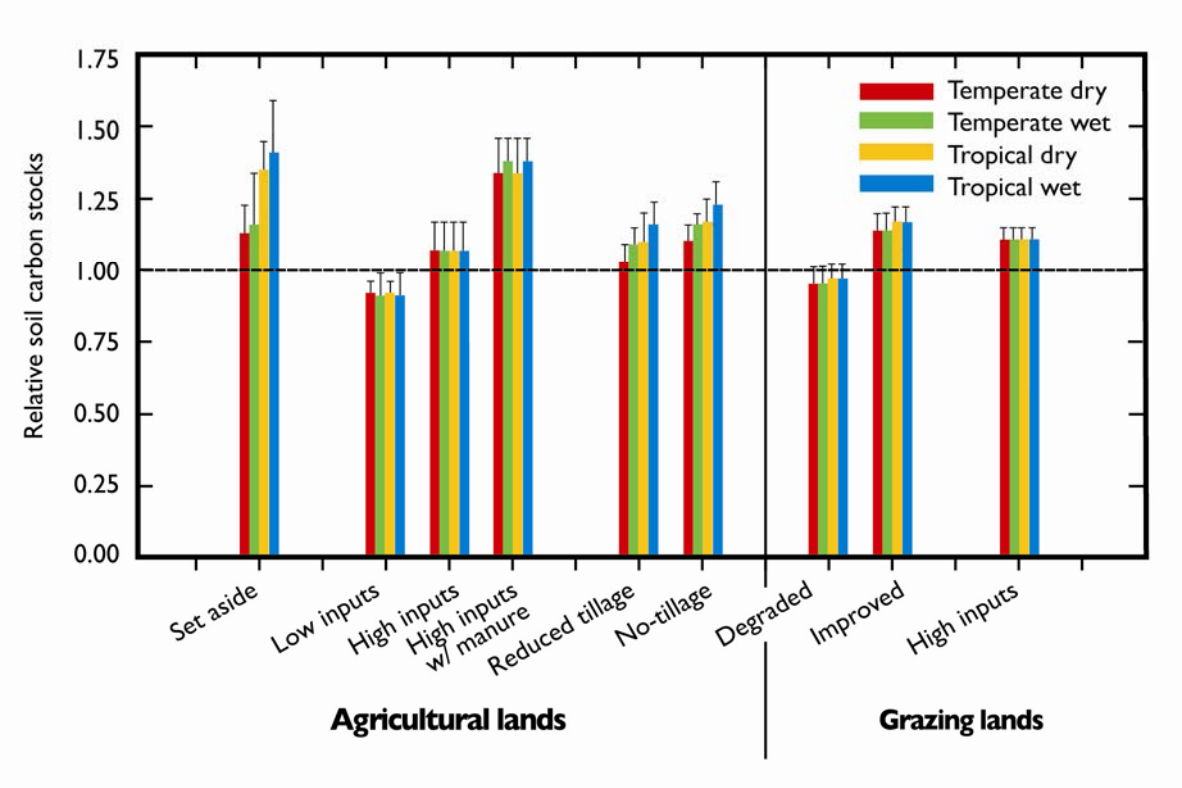

Fig. 10-2. Relative soil carbon following implementation of new agricultural or grassland management practices. Conventionally tilled, medium-input cultivated land and moderately grazed grasslands with moderate inputs are defaults for agricultural and grazing lands, respectively. Default soil carbon stocks (like those in Table 10-1) can be multiplied by one or more emission factors to estimate carbon sequestration rates. The dashed horizontal line indicates default soil carbon stocks (i.e., those under conventional-tillage cropland or undegraded garzingland, with medium inputs). Temperature/precipitation divisions are the same as those described in Table 10-1. Data are from Nabuurs e*t al.* (2004) and Ogle *et al.* (2004).

Chapter 11. North American Forests

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KEY FINDINGS

- North American forests contain more than 170 Gt of carbon, of which 28% is in live biomass and 72% is in dead organic matter.
- North American forests were a net carbon sink of approximately -269 Mt C yr^{-1} over the last 10 to 15 years. This estimate is highly uncertain.
- Deforestation continues in Mexico where forests are a source of CO_2 to the atmosphere. Forests of the United States and parts of Canada have become a carbon sink as a consequence of the recovery of forests following the abandonment of agricultural land.
- Carbon dioxide emissions from Canada's forests are highly variable because of interannual changes in area burned by wildfire.
- The size of the carbon sink in U.S. forests appears to be declining based on inventory data from 1952 to the present.
- Many factors that cause changes in carbon stocks of forests have been identified, including land-use change, timber harvesting, natural disturbance, increasing atmospheric CO_2, climate change, nitrogen deposition, and tropospheric ozone. There is a lack of consensus about how these different natural and anthropogenic factors contribute to the current sink, and the relative importance of factors varies geographically.
- There have been several continental- to subcontinental-scale assessments of future changes in carbon and vegetation distribution in North America, but the resulting projections of future trends for North American forests are highly uncertain. Some of this is due to uncertainty in future climate, but there is also considerable uncertainty in forest response to climate change and in the interaction of climate with other natural and anthropogenic factors.

- Forest management strategies can be adapted to manipulate the carbon sink strength of forest systems. The net effect of these management strategies will depend on the area of forests under management, management objectives for resources other than carbon, and the type of disturbance regime being considered.
- Decisions concerning carbon storage in North American forests and their management as carbon sources and sinks will be significantly improved by (1) filling gaps in inventories of carbon pools and fluxes, (2) a better understanding of how management practices affect carbon in forests, (3) better estimate of potential changes in forest carbon under climate change and other factors, and (4) the increased availability of decision support tools for carbon management in forests.

INTRODUCTION

The forest area of North America totals 771 million hectares, 36% of the land area of North America and about 20% of the world's forest area (Food and Agriculture Organization 2001) (see Table 11-1). About 45% of this forest area is classified as boreal, mostly in Canada and some in Alaska. Temperate and tropical forests constitute the remainder of the forest area.

Table 11-1. Area of forest land by biome and country, 2000 (1000 ha).

North American forests are critical components of the global carbon cycle, exchanging large amounts of CO_2 and other gases with the atmosphere and oceans. In this chapter we present the most recent estimates of the role of forests in the North American carbon balance, describe the main factors that affect forest carbon stocks and fluxes, describe how forests the carbon cycle through CO_2 sequestration and emissions, and discuss management options and research needs.

CARBON STOCKS AND FLUXES

Ecosystem Carbon Stocks And Pools

North American forests contain more than 170 Gt of carbon, of which 28% is in live biomass and 72% is in dead organic matter (Table 11-2). Among the three countries, Canada's forests contain the most carbon and Mexico's forests the least.

Table 11-2. Carbon stocks in forests by ecosystem carbon pool and country (Mt C).

Carbon density (the amount of carbon stored per unit of land area) is highly variable. In Canada, the majority of carbon storage occurs in boreal and cordilleran forests (Kurz and Apps, 1999). In the U.S., forests of the Northeast, Upper Midwest, Pacific Coast, and Alaska (with 14,000 Mt C) store the most carbon. In Mexico, temperate forests contain 4,500 Mt C, tropical forests contain 4,100 Mt C, and semiarid forests contain 5,000 Mt C.

Net North American Forest Carbon Fluxes

According to nearly all published studies, North American lands are a net carbon sink (Pacala *et al*., 2001). A summary of currently available data from greenhouse gas inventories and other sources suggests that the magnitude of the North American forest carbon sink was approximately –269 Mt C yr^{-1} over the last decade or so, with U.S. forests accounting for most of the sink (Table 11-3). This estimate is likely to be within 50% of the true value.

Table 11-3. Change in carbon stocks for forests and wood products by country (Mt C yr^{-1}).

Canadian forests were estimated to be a net sink of -17 Mt C yr^{-1} from 1990-2004 (Environment Canada, 2006) (Table 11-3). These estimates pertain to the area of forest considered to be "managed" under international reporting guidelines, which is 82% of the total area of Canada's forests. The estimates also include the carbon changes that result from land-use change. Changes in forest soil carbon are not included. High interannual variability is averaged into this estimate—the annual change varied from approximately -50 to +40 between 1990 and 2004. Years with net emissions were generally years with high forest fire activity (Environment Canada, 2006).

Most of the net sink in U.S. forests is in aboveground carbon pools, which account for –146 Mt C yr^{-1} (Smith and Heath, 2005). The net sink for the belowground carbon pool is estimated at –90 Mt C (Pacala *et al*., 2001). The size of the carbon sink in U.S. forest ecosystems appears to have declined slightly over the last decade (Smith and Heath, 2005). In contrast, a steady or increasing supply of timber products now and in the foreseeable future (Haynes, 2003) means that the rate of increase in the wood products carbon pool is likely to remain steady.

For Mexico, the most comprehensive available estimate for the forest sector suggests a source of +52 Mt C per year in the 1990s (Masera *et al*., 1997). This estimate does not include changes in the wood products carbon pool. The main cause of the estimated source is deforestation, which is offset to a much lesser degree by restoration and recovery of degraded forestland.

Landscape-scale estimates of ecosystem carbon fluxes reflect the dynamics of individual forest stands that respond to unique combinations of disturbance history, management intensity, vegetation, and site

characteristics. Extensive land-based measurements of forest/atmosphere carbon exchange for forest stands at various stages of recovery after disturbance reveal patterns and causes of sink or source strength, which is highly dependent on time since disturbance. Representative estimates for North America are summarized in Appendix 11.A.

TRENDS AND DRIVERS

Overview of Trends and Drivers of Change in Carbon Stocks

Many factors that cause changes in carbon stocks of forests and wood products have been identified, but the importance of each is still debated in the scientific literature (Barford *et al*., 2001; Caspersen *et al*., 2000; Goodale *et al*., 2002; Korner, 2000; Schimel *et al*., 2000). Land-use change, timber harvesting, natural disturbance, increasing atmospheric CO_2, climate change, nitrogen deposition, and tropospheric ozone all have effects on carbon stocks in forests, with their relative influence depending on geographic location, the type of forest, and specific site factors. It is important for policy implementation and management of forest carbon to separate the effects of direct human actions from natural factors.

The natural and anthropogenic factors that significantly influence forest carbon stocks are different for each country, and still debated in the scientific literature. Natural disturbances are significant in Canada, but estimates of the relative effects of different kinds of disturbance are uncertain. One study estimated that impacts of wildfire and insects caused emissions of about +40 Mt C yr^{-1} of carbon to the atmosphere over the two decades (Kurz and Apps, 1999). Another study concluded that the positive effects of climate, CO_2, and nitrogen deposition outweighed the effects of wildfire and insects, making Canada's forests a net carbon sink in the same period (Chen *et al*., 2003). In the United States, land use change and timber harvesting seem to be dominant factors according to repeated forest inventories from 1952 to 1997 that show forest carbon stocks (excluding soils) increasing by about 175 Mt C yr^{-1}. The most recent inventories show a decline in the rate of carbon uptake by forests, which appears to be mainly the result of changing growth and harvest rates following a long history of land-use change and management (Birdsey *et al*., 2006; Smith and Heath, 2005). The factors behind net emissions form Mexico's forests are deforestation, forest degradation, and forest fires that are not fully offset by forest regeneration (Masera *et al.*, 1997; de Jong *et al*., 2000).

Effects of Land-Use Change

Since 1990, approximately 549,000 ha of former cropland or grassland in Canada have been abandoned and are reverting to forest, while 71,000 ha of forest have been converted to cropland, grassland, or settlements, for a net increase in forest area of 478,000 ha (Environment Canada, 2005). In 2004, approximately 25,000 ha were converted from forest to cropland, 19,000 ha from forest to

settlements and approximately 3,000 ha converted to wetlands. These land use changes resulted in emissions of about 4 Mt C (Environment Canada 2006).

In the last century more than 130 million hectares of land in the conterminous United States were either afforested (62 million ha) or deforested (70 million ha) (Birdsey and Lewis 2003). Houghton *et al*. (1999) estimated that cumulative changes in forest carbon stocks for the period from 1700 to 1990 in the United States were about +25 Gt C, primarily from conversion of forestland to agricultural use and reduction of carbon stocks for wood products.

Emissions from Mexican forests to the atmosphere are primarily due to the impacts of deforestation to pasture and degradation of 720,000 to 880,000 ha per year (Masera *et al*., 1997; Palacio *et al.* 2000). The highest deforestation rates occur in the tropical deciduous forests (304,000 ha in 1990) and the lowest in temperate broadleaf forests (59,000 ha in 1990).

Effects of Forest Management

The direct human impact on North American forests ranges from very minimal for protected areas to very intense for plantations (Table 11-4). Between these extremes is the vast majority of forestland, which is impacted by a wide range of human activities and government policies that influence harvesting, wood products, and regeneration.

Table 11-4. Area of forestland by management class and country, 2000 (1000 ha).

Forests and other wooded land in Canada occupy about 402 Mha. Approximately 310 Mha is considered forest of which 255 Mha (83%) are under active forest management (Environment Canada, 2006). Managed forests are considered to be under the direct influence of human activity and not reserved. Less than 1% of the area under active management is harvested annually. Apps *et al*. (1999) used a carbon budget model to simulate carbon in harvested wood products (HWP) for Canada. Approximately 800 Mt C were stored in the Canadian HWP sector in 1989, of which 50 Mt C were in imported wood products, 550 Mt C in exported products, and 200 Mt C in wood products produced and consumed domestically.

Between 1990 and 2000, about 4 Mha yr^{-1} were harvested in the U.S., two-thirds by partial-cut harvest and one-third by clear-cut (Birdsey and Lewis, 2003). Between 1987 and 1997, about 1 Mha yr^{-1} were planted with trees, and about 800,000 ha were treated to improve the quality and/or quantity of timber produced (Birdsey and Lewis, 2003). Harvesting in U.S. forests accounts for substantially more tree mortality than natural causes such as wildfire and insect outbreaks (Smith *et al*., 2004). The

harvested wood resulted in -57 Mt C added to landfills and products in use, and an additional 88 Mt C were emitted from harvested wood burned for energy (Skog and Nicholson, 1998).

About 80% of the forested area in Mexico is socially owned by communal land grants (*ejidos*) and rural communities. About 95% of timber harvesting occurs in native temperate forests (SEMARNAP, 1996). Illegal harvesting involves 13.3 million m^3 of wood every year (Torres, 2004). The rural population is the controlling factor for changes in carbon stocks from wildfire, wood extraction, shifting agriculture practices, and conversion of land to crop and pasture use.

Effects of Climate and Atmospheric Chemistry

Environmental factors, including climate variability, nitrogen deposition, tropospheric ozone, and elevated CO_2, have been recognized as significant factors affecting the carbon cycle of forests (Aber *et al*., 2001; Ollinger *et al*., 2002). Some studies indicate that these effects are significantly smaller than the effects of land management and land-use change (Caspersen *et al*., 2000; Schimel *et al*., 2000). Recent reviews of ecosystem-scale studies known as Free Air CO_2 Exchange (FACE) experiments suggest that rising CO_2 increases net primary productivity by 12–23% over all species (Norby *et al*., 2005; Nowak *et al*., 2004). However, it is uncertain whether this effect results in a lasting increase in sequestered carbon or causes a more rapid cycling of carbon between the ecosystem and the atmosphere (Korner *et al*., 2005; Lichter, 2005). Experiments have also shown that the effects of rising CO_2 are significantly moderated by increasing tropospheric ozone (Karnosky *et al*., 2003; Loya *et al*., 2003). When nitrogen availability is also considered, reduced soil fertility limits the response to rising CO_2, but nitrogen deposition can increase soil fertility to counteract that effect (Finzi *et al.* 2006; Johnson *et al*., 1998; Oren *et al*., 2001). Observations of photosynthetic activity from satellites suggest that productivity changes due to lengthening of the growing season depend on whether areas were disturbed by fire (Goetz *et al*., 2005). Based on these conflicting and complicated results from different studies and approaches, a definitive assessment of the relative importance, and interactions, of natural and anthropogenic factors is a high priority for research (U.S. Climate Change Science Program, 2003).

Effects of Natural Disturbances

Wildfire, insects, diseases, and weather events are common natural disturbances in North America. These factors impact all forests but differ in magnitude by geographic region.

Wildfires were the largest disturbance in the twentieth century in Canada (Weber and Flannigan, 1997). In the 1980s and 1990s, the average total burned area was 2.6 Mha yr^{-1} in Canada's forests, with a maximum 7.6 Mha yr^{-1} in 1989. Carbon emissions from forest fires range from less than +1 Mt C yr^{-1} in the interior of British Columbia to more than +10 Mt C yr^{-1} in the western boreal forest. Total emissions

from forest fires in Canada averaged approximately +27 Mt C yr^{-1} between 1959 and 1999 (Amiro *et al*., 2001). Estimated carbon emissions from four major insect pests in Canadian forests (spruce budworm, jack pine budworm, hemlock looper, and mountain pine beetle) varied from +5 to 10 Mt C yr^{-1} in the 1970s to less than +2 Mt C yr^{-1} in the mid-1990s[1]. Much of the Canadian forest is expected to experience increases in fire severity (Parisien *et al*., 2005) and burn areas (Flannigan *et al*., 2005), and continued outbreaks of forest pests are also likely (Volney and Hirsch, 2005).

In U.S. forests insects, diseases, and wildfire combined affect more than 30 Mha per decade (Birdsey and Lewis, 2003). Damage from weather events (hurricanes, tornados, ice storms) may exceed 20 Mha per decade (Dale *et al*., 2001). Although forest inventory data reveal the extent of tree mortality attributed to all causes combined, estimates of the impacts of individual categories of natural disturbance on carbon pools of temperate forests are scarce. The impacts of fire are clearly significant. According to one estimate, the average annual carbon emissions from biomass burning in the contemporary United States ranges from 9 to 59 Mt C (Leenhouts, 1998). McNulty (2002) estimated that large hurricanes in the United States could convert 20 Mt C of live biomass into detrital carbon pools.

The number and area of sites affected by forest fires in Mexico have fluctuated considerably between 1970 and 2002 with a clear tendency of an increasing number of fire events (4,000–7,000 in the 1970s and 1,800–15,000 in the 1990s), and overall, larger areas are being affected (0.08–0.25 Mha in 1970s and 0.05–0.85 Mha in 1990s). During El Nino years, increasing drought increases fire frequencies (Torres, 2004). Between 1995 and 2000, an average 8,900 fire events occurred per year and affected about 327,000 ha of the forested area. Currently, no estimates are available on the contribution of these fires to CO_2 emissions. Pests and diseases are important natural disturbance agents in temperate forests of Mexico; however, no statistics exist on the extent of the affected land area.

Projections of Future Trends

Large portions of the Canadian and Alaskan forest are expected to be particularly sensitive to climate change (Hogg and Bernier, 2005). Climate change effects on forest growth could be positive (e.g., increased rates of photosynthesis and increased water use efficiency) or negative (decreased water availability, higher rates of respiration) (Baldocchi and Amthor, 2001). It is difficult to predict the direction of these changes and they will likely vary by species and local conditions of soils and topography (Johnston and Williamson, 2005). Because of the large area of boreal forests and expected

[1]These estimates are the product of regional carbon density values, the proportion of mortality in defoliated stands given in Kurz and Apps (1999), data on area affected taken from NFDP (2005), and the proportion of C in insect-killed stands that is emitted directly to the atmosphere (0.1) from the disturbance matrix for insects used in the CBM-CFS (Kurz *et al*., 1992).

high degree of warming in northern latitudes, Canada and Alaska require close monitoring over the next few decades as these areas will likely be critical to determining the carbon balance of North America.

Assessments of future changes in carbon and vegetation distribution in the U.S. suggest that under most future climate conditions, NPP would respond positively to changing climate but total carbon storage would remain relatively constant (VEMAP Members, 1995; Pan *et al*., 1998; Neilson *et al*., 1998; Joyce *et al*., 2001). Under most climate scenarios the West gets wetter; when coupled with higher CO_2 and longer growing seasons, simulations show woody expansion and increased sequestration of carbon as well as increases in fire (Bachelet *et al*., 2001). However, recent scenarios from the Hadley climate model show drying in the Northwest, which produces some forest decline (Price *et al*., 2004). Many simulations show continued growth in eastern forests through the end of the twenty-first century, but some show the opposite, especially in the Southeast. Eastern forests could experience a period of enhanced growth in the early stages of warming, due to elevated CO_2, increased precipitation, and a longer growing season. However, further warming could bring on increasing drought stress, reducing the carrying capacity of the ecosystem and causing carbon losses through drought-induced dieback and increased fire and insect disturbances.

For Mexican forests, deforestation will continue to cause large carbon emissions in the years to come. However, government programs (since 2001) are trying to reduce deforestation rates and forest degradation, implement sustainable forestry in native forests, promote commercial plantations and diverse agroforestry systems, and promote afforestation and protection of natural areas (Masera *et al*., 1997).

OPTIONS FOR MANAGEMENT

Forest management strategies can be adapted to increase the amount of carbon uptake by forest systems. Alternative strategies for wood products are also important in several ways: how long carbon is retained in use, how much wood is used for biofuel, and substitution of wood for other materials that use more energy to produce. The net effect of these management and production strategies on carbon stocks and emissions will depend on emerging government policies for greenhouse gas management, the area of forests under management, management objectives for resources other than carbon, and the type of management and production regime being considered.

The forest sector includes a variety of activities that can contribute to increasing carbon sequestration, including: afforestation, mine land reclamation, forest restoration, agroforestry, forest management, biomass energy, forest preservation, wood products management, and urban forestry (Birdsey *et al*., 2000). Although the science of managing forests specifically for carbon sequestration is not well developed, some ecological principles are emerging to guide management decisions (Appendix 11.B). The prospective role of forestry in helping to stabilize atmospheric CO_2 depends on government policy,

harvesting and disturbance rates, expectations of future forest productivity, the fate and longevity of forest products, and the ability to deploy technology and forest practices to increase the retention of sequestered CO_2. Market factors are also important in guiding the behavior of the private sector.

For Canada, Price *et al*. (1997) examined the effects of reducing natural disturbance, manipulating stand density, and changing rotation lengths for a forested landscape in northwest Alberta. By replacing natural disturbance (fire) with a simulated harvesting regime, they found that long-term equilibrium carbon storage increased from 105 to 130 Mt C. Controlling stand density following harvest had minimal impacts in the short term but increased landscape-level carbon storage by 13% after 150 years. Kurz *et al*. (1998) investigated the impacts on landscape-level carbon storage of the transition from natural to managed disturbance regimes. For a boreal landscape in northern Quebec, a simulated fire disturbance interval of 120 yr was replaced by a harvest cycle of 120 yr. The net impact was that the average age of forests in the landscape declined from 110 yr to 70 yr, and total carbon storage in forests declined from 16.3 to 14.8 Mt C (including both ecosystem and forest products pools).

Market approaches and incentive programs to manage greenhouse gases, particularly CO_2, are under development in the United States, the European Union, and elsewhere (Totten, 1999). Since forestry activities have highly variable costs because of site productivity and operational variability, most recent studies of forestry potential develop "cost curves", i.e., estimates of how much carbon will be sequestered by a given activity for various carbon prices (value in a market system) or payments (in an incentive system). There is also a temporal dimension to the analyses because the rate of change in forest carbon stocks is variable over time, with forestry activities tending to have a high initial rate of net carbon sequestration followed by a lower or even a negative rate as forests reach advanced age.

In the United States, a bundle of forestry activities could potentially increase carbon sequestration from -100 to -200 Mt C yr^{-1} according to several studies (Birdsey *et al*., 2000; Lewandrowski, 2004; Environmental Protection Agency, 2005; Stavins and Richards, 2005). The rate of annual mitigation would likely decline over time as low-cost forestry opportunities become scarcer, forestry sinks become saturated, and timber harvesting takes place. Economic analyses of the U.S. forestry potential have focused on three broad categories of activities: afforestation (conversion of agricultural land to forest), improved management of existing forests, and use of woody biomass for fuel. Improved management of existing forest lands may be attractive to landowners at a carbon prices below $10 per ton of CO_2; afforestation requires a moderate price of $15 per ton of CO_2 or more to induce landowners to participate; and biofuels become dominant at prices of $30-50 per ton of CO_2 (Lewandrowski, 2004; Stavins and Richards, 2005; Environmental Protection Agency, 2005). Table 11-5 shows a simple scenario of emissions reduction below baseline, annualized over the time period 2010-2110, for forestry activities as part of a bundle of reduction options for the land base.

Table 11-5. Illustrative emissions reduction potential of various forestry activities in the United States under a range of prices and sequestration rates.

Production of renewable materials that have lower life-cycle emissions of greenhouse gases than non-renewable alternatives is a promising strategy for reducing emissions. Lippke *et al*. (2004) found that wood components used in residential construction had lower emissions of CO_2 from energy inputs than either concrete or steel.

Co-benefits are vitally important for inducing good forest carbon management. For example, conversion of agricultural land to forest will generally have positive effects on water, air, and soil quality and on biodiversity. In practice, some forest carbon sequestration projects have already been initiated even though sequestered carbon has little current value (Winrock International, 2005). In many of the current projects, carbon is a secondary objective that supports other landowner interests, such as restoration of degraded habitat. But co-effects may not all be beneficial. Water quantity may decline because of increased transpiration by trees relative to other vegetation. And taking land out of crop production may affect food prices—at higher carbon prices, nearly 40 million ha may be converted from cropland to forest (Environmental Protection Agency, 2005). Implementation of a forest carbon management policy will need to carefully consider co-effects, both positive and negative.

DATA GAPS AND INFORMATION NEEDS FOR DECISION SUPPORT

Decisions concerning carbon storage in North American forests and their management as carbon sources and sinks will be significantly improved by (1) filling gaps in inventories of carbon pools and fluxes, (2) a better understanding of how management practices affect carbon in forests, and (3) the increased availability of decision support tools for carbon management in forests.

Major Data Gaps in Estimates of Carbon Pools and Fluxes

Effective carbon policy and management to increase carbon sequestration and/or reduce emissions requires thorough understanding of current carbon stock sizes and flux rates, and responses to disturbance. Data gaps complicate analyses of the potential for policies to influence natural, social and economic drivers that can change carbon stocks and fluxes. Forests in an area as large as North America are quite diverse, and comprehensive data sets that can be used to analyze forestry opportunities, such as spatially explicit historical management and disturbance rates and effects on the carbon cycle, would enable managers to change forest carbon stocks and fluxes.

In the United States, the range of estimates of the size of the land carbon sink is between 0.30 and 0.58 Mt C yr^{-1} (Pacala *et al*., 2001). Significant data gaps among carbon pools include carbon in wood products, soils, woody debris, and water transport (Birdsey, 2004; Pacala *et al*., 2001). Geographic areas that are poorly represented in the available data sets include much of the Intermountain Western United States and Alaska, where forests of low productivity have not been inventoried as intensively as more productive timberlands (Birdsey, 2004). Accurate quantification of the relative magnitude of various causal mechanisms at large spatial scales is not yet possible, although research is ongoing to combine various approaches and data sets: large-scale observations, process-based modeling, ecosystem experiments, and laboratory investigations (Foley and Ramankutty, 2004).

Data gaps exist for Canada, particularly regarding changes in forest soil carbon and forestlands that are considered "unmanaged" (17% of forest lands). Aboveground biomass is better represented in forest inventories; however, the information needs to be updated and made more consistent among provinces. The new Canadian National Forest Inventory, currently under way, will provide a uniform coverage at a 20×20 km grid that will be the basis for future forest carbon inventories. Data are also lacking on carbon fluxes, particularly those due to insect outbreaks and forest stand senescence. The ability to model forest carbon stock changes has considerably improved with the release of the CBM (Kurz *et al*., 2002); however the CBM does not consider climate change impacts (Price *et al*., 1999; Hogg and Bernier, 2005).

For Mexico, there is very little data about measured carbon stocks for all forest types. Information on forest ecosystem carbon fluxes is primarily based on deforestation rates, while fundamental knowledge of carbon exchange processes in almost all forest ecosystems is missing. That information is essential for understanding the effects of both natural and human-induced drivers (hurricanes, fires, insect outbreaks, climate change, migration, and forest management strategies), which all strongly impact the forest carbon cycle. Current carbon estimates are derived from studies in preferred sites in natural reserves with species-rich tropical forests. Therefore, inferences made from the studies on regional and national carbon stocks and fluxes probably give biased estimates on the carbon cycle.

Major Data Gaps in Knowledge of Forest Management Effects

There is insufficient information available to guide land managers in specific situations to change forest management practices to increase carbon sequestration, and there is some uncertainty about the longevity of effects (Caldeira *et al*., 2004). This reflects a gap in the availability of inexpensive techniques for measuring, monitoring, and predicting changes in ecosystem carbon pools at the smaller scales appropriate for managers. There is more information available about management effects on live biomass and woody debris, and less about effects on soils and wood products. This imbalance in data has

the potential to produce unintended consequences if predicted results are based on incomplete carbon accounting.

In the tropics, agroforestry systems offer a promising economic alternative to slash-and-burn agriculture, including highly effective soil conservation practices and mid-term and long-term carbon mitigation options (Soto-Pinto *et al*., 2001; Nelson and de Jong, 2003; Albrecht and Kandji, 2003). However, a detailed assessment of current implementations of agroforestry systems in different regions of Mexico is missing. Agroforestry also has potential in temperate agricultural landscapes, but as with forest management, there is a lack of data about how specific systems affect carbon storage (Nair and Nair, 2003).

Refining management of forests to realize significant carbon sequestration while at the same time continuing to satisfy the other needs and services of provided by forests (e.g., timber harvest, recreational value, watershed management) will require a multi-criteria decision support framework for a holistic and adaptive management program of the carbon cycle in North American forests. For example, methods should be developed for enhancing the efficiency of forest utilization as a renewable energy source, increasing the carbon storage per acre from existing forests, or even increasing the acreage devoted to forest systems that provide carbon sequestration. Currently there is little information about how appropriate incentives might be applied to accomplish these goals effectively, but given the importance of forests in the global carbon cycle, success in this endeavor could have important long-term and large-scale effects on global atmospheric carbon stocks.

Availability Of Decision-Support Tools

Few decision-support tools for land managers that include complete carbon accounting are available. Some are in development or have been used primarily in research studies (Proctor *et al*., 2005; Potter *et al*., 2003). As markets emerge for trading carbon credits, and if credits for forest management activities have value in those markets, then the demand for decision-support tools will encourage their development.

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Table 11-1. Area of forest land by biome and country, 2000 (1000 ha)[1]

Ecological zone:	Canada[2]	U.S.[3]	Mexico[4]	Total
Tropical/subtropical	0	115,200	30,700	145,900
Temperate	101,100	142,400	32,900	276,400
Boreal	303,000	45,500	0	348,500
Total	404,100	303,100	63,600	770,800

[1]There is 95% certainty that the actual values are within 10% of those reported in this table (e.g., for the United States see Bechtold and Patterson, 2005).
[2]Canadian Forest Service, 2005
[3]Smith *et al*., 2004
[4]Palacio *et al*., 2000

Table 11-2. Carbon stocks in forests by ecosystem carbon pool and country (Mt C)[1]

Ecosystem carbon pool:	Canada[2]	U.S.[3]	Mexico[4]	Total
Biomass	14,500	24,900	7,700	47,100
Dead organic matter[5]	71,300	41,700	11,400	124,400
Total	85,800	66,600	19,100	171,500

[1]There is 95% certainty that the actual values are within 25% of those reported in this table (Heath and Smith, 2000; Smith and Heath, 2000).
[2]Kurz and Apps, 1999
[3]Heath and Smith, 2004; Birdsey and Heath, 1995
[4]Masera *et al*., 2001
[5]Includes litter, coarse woody debris, and soil carbon

Table 11-3. Change in carbon stocks for forests and wood products by country (Mt C yr^{-1})

Carbon pool:	Canada[1]	U.S.[2]	Mexico[3]	Total
Forest Ecosystem	–17	–236	+52	–201
Wood Products	–11	–57	ND[4]	–68
Total	–28	–293	+52	–269

[1]Data for 1990-2004, taken from Environment Canada (2006), Goodale *et al*. (2002). There is 95% certainty that the actual values are within 100% of those reported for Canada.
[2]From Smith and Heath, 2005 (excluding soils), and Pacala *et al*., 2001 (soils). Estimates do not include urban forests. There is 95% certainty that the actual values are within 50% of those reported for the United States.
[3]From Masera, 1997. There is 95% certainty that the actual values are within 100% of those reported for Mexico.
[4]Estimates are not available.

Table 11-4. Area of forestland by management class and country, 2000 (1000 ha)[1]

Management class:	Canada	U.S.	Mexico	Total
Protected	19,300	66,700	6,000	92,000
Plantation	4,500	16,200	200	20,900
Other	380,300	220,200	57,400	657,900
Total	404,100	303,100	63,600	770,800

[1]From Food and Agriculture Organization 2001; Natural Resources Canada 2005. Estimates in this table are within 10% of the true value at the 95% confidence level (e.g. for the U.S. see Bechtold and Patterson 2005).

Table 11-5. Illustrative emissions reduction potential of various forestry activities in the United States under a range of prices and sequestration rates[1]

Forestry activity	Carbon sequestration rate ($t\ CO_2\ ha^{-1}\ yr^{-1}$)	Price range ($/t CO_2)	Emissions reduction potential ($Mt\ CO_2\ yr^{-1}$)
Afforestation	5.4–23.5	15–30	137–823
Forest management	5.2–7.7	1–30	25–314
Biofuels	11.8–13.6	30–50	375–561

[1]Adapted from Environmental Protection Agency (2005). Maximum price analyzed was $50/t CO_2.

APPENDIX 11A
ECOSYSTEM CARBON FLUXES

The recent history of disturbance largely determines whether a forest system will be a net source or sink of C. For example, net ecosystem productivity (NEP, gains due to biomass growth minus losses due to respiration in vegetation and soil) is being measured across a range of forest types in Canada using the eddy covariance technique. In mature forests, values range from –19.6 t C ha^{-1} yr^{-1} in a white pine plantation in southern Ontario (Arain and Restrepo-Coupe, 2005) to –3.2 t C ha^{-1} yr^{-1} in a jack pine forest in (Amiro *et al*., 2005; Griffis *et al*., 2003). In recently disturbed forests, NEP ranges from +58.0 t C ha^{-1} yr^{-1} in a harvested Douglas-fir forest (Humphreys *et al*., 2005) to +5.7 t C ha^{-1} yr^{-1} in a 7 year old harvested jack pine forest (Amiro *et al*., 2005). In general, forest stands recovering from disturbance are sources of carbon until uptake from growth becomes greater than losses due to respiration, usually within 10 years (Amiro *et al*., 2005).

In the United States, extensive land-based measurements of forest/atmosphere carbon exchange reveal patterns and causes of sink or source strength (Table 11A-1). Results show that net ecosystem exchange (NEE) of carbon in temperate forests ranges from a source of +12.7 t C ha^{-1} yr^{-1} to a sink of –5.9 t C ha^{-1} yr^{-1}. Forests identified as sources are primarily forests in the earliest stages of regeneration (up to about 8 years) following stand-replacing disturbances such as wildfire and logging (Law *et al*., 2002). Mature temperate deciduous broadleaf forests and mature evergreen coniferous forests were an average sink of –2.7 and –2.5 t C ha^{-1} yr^{-1}, respectively (12 sites, 54 site-years of data). Values ranged from a source of +0.3 for a mixed deciduous and evergreen forest to a sink of –5.8 for an aggrading deciduous forest, averaged over multiple years. Young temperate evergreen coniferous forests (8 to 20 years) ranged from a sink of –0.6 to –5.9 t C ha^{-1} yr^{-1} (mean 3.1). These forests are still rapidly growing and have not reached the capacity for carbon uptake.

Mature forests can have substantial stocks of sequestered carbon. Disturbances that damage or replace forests can result in the land being a net source of carbon dioxide for a few years in mild climates to 10–20 years in harsh climates while the forests are recovering (Law *et al*., 2004; Clark *et al*., 2004). Thus, the range of observed annual NEE of carbon dioxide ranges from a source of about +13 t C ha^{-1} yr^{-1} in a clearcut forest to a net sink of –6 t C ha^{-1} in mature temperate forests.

For Mexican forests, estimates of net ecosystem carbon exchange are unavailable, but estimates from other tropical forests may indicate rates for similar systems in Mexico. In Puerto Rico, aboveground NPP in tropical forests range from –9.2 to –11.0 t C ha^{-1} yr^{-1} (Lugo *et al*., 1999). Belowground NPP measurements exist for only one site with –19.5 t C ha^{-1} yr^{-1} (Lugo *et al*., 1999). In Hawaii, aboveground

and belowground NPP of native forests dominated by *Metreosideros polymorpha* vary depending on substrate age and precipitation regime. Aboveground NPP ranges between –4.0 to –14.0 t C ha^{-1} yr^{-1}, while belowground NPP ranges between –5.2 and –9.0 t C ha^{-1} yr^{-1} (Giardina *et al.*, 2004). Soil carbon emissions along the substrate age gradient range from +2.2 to +3.3 t C ha^{-1} yr^{-1}, and along the precipitation gradient from +4.0 to +9.7 t C ha^{-1} yr^{-1} (Osher *et al.*, 2003). NEP estimates are not available for these tropical forests, so their net impact on atmospheric carbon stocks cannot be calculated.

APPENDIX 11A REFERENCES

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Table 11A-1. Comparison of net ecosystem exchange (NEE) for different types and ages of temperate forests. Positive NEE means the forest is a sink for atmospheric CO_2. Eighty-one site years of data are from multiple published papers from each of the AmeriFlux network sites, and a network synthesis paper (Law *et al*. 2002). NEE was averaged by site, then the mean was determined by forest type and age class. SD is standard deviation among sites in the forest type and age class.

	NEE (t C ha^{-1} y^{-1})		
	Regenerating Clearcut (–1 ~ 3 years after disturbance) (1 site, 5 site-years)	Young forest (8 ~ 20 years old) (4 sites, 16 site-years)	Mature forest (>20 years old) (13 sites, 60 site-years)
Evergreen Coniferous Forests	–12.7 ~ 1.7, mean –7.1 (SD 4.7) (1 site, 5 site-years)	0.6 ~ 5.9, mean 3.1 (SD 2.6) (4 sites, 16 site-years)	0.6 ~ 4.5, mean 2.5 (SD 1.4) (6 sites, 20 site-years)
Mixed Evergreen and Deciduous Forests	NA	NA	0.3 ~ 2.1, mean –1.0 (SD 0.6) (1 site, 6 site-years)
Deciduous Broadleaf Forests	NA	NA	0.6 ~ 5.8, mean 2.7 (SD 1.8) (6 sites, 34 site-years)

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APPENDIX 11B
PRINCIPLES OF FOREST MANAGEMENT FOR ENHANCING CARBON SEQUESTRATION

The net rate of carbon accumulation has been generally understood (Woodwell and Whittaker, 1968) as the difference between gross primary production (gains) and respiration (losses), although this neglects important processes such as leaching of DOC, emission of methane (CH4), fire, harvests or erosion that may contribute substantially to carbon loss and gain in forest ecosystems (Schulze *et al*., 1999; Harmon, 2001; Chapin *et al*., in review). The net ecosystem carbon balance (NECB) in forests is therefore defined as net ecosystem production, or NEP, plus the non-physiological horizontal and vertical transfers into and out of the forest stand.

With respect to the impacts of forest management on the overall carbon balance, some general principles apply (Harmon, 2001; Harmon and Marks, 2002; Pregitzer *et al*., 2004). First, forest management can impact carbon pool sizes via:

- changing production rates (since NEP = NPP—heterotrophic respiration Rh);
- changing decomposition flows (Rh) (e.g., Fitzsimmons *et al*., 2004);
- changing the amount of material transferred between pools; or
- changing the period between disturbances/ management activities.

The instantaneous balance between production, decomposition, and horizontal or vertical transfers into and out of a forest stand determines whether the forest is a net source or a net sink. Given that these terms all change as forests age, the disturbance return interval is a key driver of stand- and landscape-level carbon dynamics. Rh tends to be enhanced directly after disturbance, so as residue and other organic carbon pools decompose, a forest is often a net source immediately after disturbances such as management activity. NPP tends to increase as forests age, although in older forests it may decline (Ryan, 1997). Eventually, as stands age, NPP and Rh become similar in magnitude, although few managed stands are allowed to reach this age. The longer the average time interval between disturbances, the more carbon is stored. The nature of the disturbance is also important; the less severe the disturbance (e.g., less fire removal), the more carbon is stored.

Several less general principles can be applied to specific carbon pools, fluxes, or situations:

- Management activities that move live carbon to dead pools (such as CWD or soil C) over short periods of time will often dramatically enhance decomposition (Rh), although considerable carbon can be stored in decomposing pools (Harmon and Marks, 2002). Regimes seeking to reduce the decomposition-related flows from residue following harvest may enhance overall sink capacity of these forests if these materials are used for energy generation or placed into forest products that last longer than the residue.
- Despite the importance of decomposition rates to the overall stand-level forest carbon balance, management of CWD pools is mostly impacted by recruitment of new CWD rather than by changing decomposition rates (Janisch and Harmon, 2002; Pregitzer and Euskirchen, 2004). Decreasing the interval between harvests can significantly decrease the store in this pool.
- Live coarse root biomass accounts for approximately 20–25% of aboveground forest biomass (Jenkins *et al*., 2003), and there is additional biomass in fine roots. Following harvest, this pool of live root biomass is transferred to the dead biomass pool, which can form a significant carbon store. Note that roots of various size classes and existing under varying environmental conditions decompose at different rates.
- Some carbon can be sequestered in wood products from harvested wood, though due to manufacturing losses only about 60% of the carbon harvested is stored in products (Harmon, 1996). Clearly, longer-lived products will sequester carbon for longer periods of time.
- According to international convention, the replacement of fossil fuel by biomass fuel can be counted as an emissions offset if the wood is produced from sustainably managed forests (Schoene and Netto 2005).

Little published research has been aimed at quantifying the impacts of specific forest management activities on carbon storage, but examples of specific management activities can be given.

- Practices aimed at increasing NPP: fertilization; genetically improved trees that grow faster (Peterson *et al*., 1999); any management activity that enhances growth rate without causing a concomitant increase in decomposition (Stanturf *et al*., 2003; Stainback and Alavalapati, 2005).
- Practices aimed at reducing Rh (i.e., minimizing the time forests are a source to the atmosphere following disturbance): low impact harvesting (that does not promote soil respiration); utilization of logging residues (biomass energy and fuels); incorporation of logging residue into soil during site prep (but note that this could also speed up decomposition); thinning to capture mortality; fertilization.

Since NECB changes with time as forests age, if a landscape is composed of stands with different ages then carbon gains in one stand can be offset by losses from another stand. The net result of these stand-level changes determines overall landscape-level carbon stores. Note that disturbance-induced Rh losses are typically larger than annual gains, such that a landscape where forest area is increasing might still be neutral with respect to carbon stocks overall. Thus, at the landscape level practices designed to enhance carbon sequestration must, on balance, replace lower-C-density systems with higher-C-density systems. Examples of these practices include: reducing fire losses; emphasizing very long-lived forest products; increasing the interval between disturbances; or reducing decomposability of dead material.

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Chapter 12. Carbon Cycles in the Permafrost Region of North America

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KEY FINDINGS

- Much of northern North America (more than 6 million km^2) is characterized by the presence of permafrost, soils or rocks that remain frozen for at least two consecutive years. This permafrost region contains approximately 25% of the world's total soil organic carbon, a massive pool of carbon that is vulnerable to release to the atmosphere as CO_2 in response to an already detectable polar warming.
- The soils of the permafrost region of North America contain 213 Gt of organic carbon, approximately 61% of the carbon in all soils of North America.
- The soils of the permafrost region of North America are currently a net sink of approximately 11 Mt C yr^{-1}.
- The soils of the permafrost region of North America have been slowly accumulating carbon for the last 5-8 thousand years. More recently, increased human activity in the region has resulted in permafrost degradation and at least localized loss of soil carbon.
- Patterns of climate, especially the region's cool and cold temperatures and their interaction with soil hydrology to produce wet and frozen soils, are primarily responsible for the historical accumulation of carbon in the region. Non-climatic drivers of carbon change include human activities, including flooding associated with hydroelectric development, that degrade permafrost and lead to carbon loss. Fires, increasingly common in the region, also lead to carbon loss.
- Projections of future warming of the polar regions of North America lead to projections of carbon loss from the soils of the permafrost region, with upwards of 78% (34 Gt) and 41% (40 Gt) of carbon stored in soils of the Subarctic and Boreal regions, respectively, being severely or extremely severely affected by future climate change.
- Options for management of carbon in the permafrost region of North America, including construction methods that cause as little disturbance of the permafrost and surface as possible, are primarily those which avoid permafrost degradation and subsequent carbon losses.

- Most research needs for the permafrost region are focused on reducing uncertainties in knowing how much carbon is vulnerable to a warming climate and how sensitive that carbon loss is to climate change. Development and adoption of measures that reduce or avoid the negative impact of human activities on permafrost are also needed.

INTRODUCTION

It is especially important to understand the carbon cycle in the permafrost region of North America because the soils in this area contain large amounts of organic carbon, carbon that is vulnerable to release to the atmosphere as carbon dioxide and methane in response to climate warming. It is predicted that the average annual air temperature in the permafrost region will increase 3–4°C by 2020 and 5–10°C by 2050 (Hengeveld, 2000). The soils in this region contain approximately 61% of the organic carbon occurring in all soils in North America (Lacelle *et al*., 2000) even though the permafrost area covers only about 21% of the soil area of the continent. Release of even a fraction of this carbon in greenhouse gases could have global consequences.

Permafrost is defined, on the basis of temperature, as soils or rocks that remain below 0^{o}C for at least two consecutive years (van Everdingen, 1998 revised May 2005). Permafrost terrain often contains large quantities of ground ice in the upper section of the permafrost. If this terrain is well protected by forests or peat, this ground ice is generally in equilibrium with the current climate. If this insulating layer is not sufficient, however, even small temperature changes, especially in the southern part of the permafrost region, could cause degradation and result in severe thermal erosion (thawing). For example, some of the permafrost that formed in central Alaska during the Little Ice Age is now degrading in response to warming during the last 150 years (Jorgenson *et al*., 2001).

The permafrost region in North America is divided into four zones on the basis of the percentage of the land area underlain by permafrost (Fig. 12-1). These zones are the Continuous Permafrost Zone ($\geq$90 to 100%), the Discontinuous Permafrost Zone ($\geq$50 to <90%), the Sporadic Permafrost Zone ($\geq$10 to <50%), and the Isolated Patches Permafrost Zone (0 to <10%) (Brown *et al*., 1997).

Figure 12-1. Permafrost zones in North America (Brown *et al*., 1997).

These permafrost zones encompass three major ecoclimatic provinces (ecological regions) (Fig. 12-2): the Arctic (north of the arctic tree line), the Subarctic (open canopy coniferous forest), and the Boreal (closed canopy forest, either coniferous or mixed coniferous and deciduous). Peatlands (organic

wetlands characterized by more than 40 cm of peat accumulation) cover large areas in the Boreal, Subarctic, and southern part of the Arctic ecoclimatic provinces.

Figure 12-2. Arctic, Subarctic, and Boreal ecoclimatic provinces (ecological regions) in North America (Ecoregions Working Group, 1989; Baily and Cushwa, 1981).

Although northern ecosystems (Arctic, Subarctic, and Boreal) in North America cover approximately14% of the global land area, they contain approximately 25% of the world's total soil organic carbon (Oechel and Vourlitis, 1994). In addition, Oechel and Vourlitis (1994) indicate that the tundra (Arctic) ecosystems alone contain approximately 12% of the global soil carbon pool, even though they account for only 6% of the total global land area. The soils of the permafrost region of North America are currently a carbon sink and are unique because they are able to actively sequester carbon and store it for thousands of years.

The objectives of this chapter are to give the below-ground carbon stocks and to explain the mechanisms associated with the carbon cycle (sources and sinks) in the soils of the permafrost region of North America.

PROCESSES AFFECTING THE CARBON CYCLE IN A PERMAFROST ENVIRONMENT

Soils of the Permafrost Region

Soils cover approximately 6,211,340 km^2 of the area of the North American permafrost region (Tables 12-1 and 12-2), with approximately 58% of the soil area being occupied by permafrost-affected (perennially frozen) soils (Cryosols/Gelisols) and the remainder by non-permafrost soils. Approximately 17% of this area is associated with organic soils (peatlands), the remainder with mineral soils. It is important to distinguish between mineral soils and organic soils in the region because different processes are responsible for the carbon cycle in these two types of soils.

Table 12-1. Areas of mineral soils in the various permafrost zones.

Table 12-2. Areas of peatlands (organic soils) in the various permafrost zones.

Mineral Soils

The schematic diagram in Fig. 12-3 provides general information about the carbon sinks and sources in mineral soils. Most of the permafrost-affected mineral soils are carbon sinks because of the process of

cryoturbation, which moves organic matter into the deeper soil layers. Other processes, such as decomposition, wildfires, and thermal degradation, release carbon into the atmosphere and, thus, act as carbon sources.

Figure 12-3. Carbon cycle in permafrost-affected upland (mineral) soils, showing below-ground organic carbon sinks and sources.

For unfrozen soils and noncryoturbated frozen soils in the permafrost region, the carbon cycle is similar to that in soils occurring in temperate regions. In these soils, organic matter is deposited on the soil surface. Some soluble organic matter may move downward, but because these soils are not affected by cryoturbation, they have no mechanism for moving organic matter from the surface into the deeper soil layers and preserving it from decomposition and wildfires. Most of their below-ground carbon originates from roots and its residence time is relatively short.

The role of cryoturbation: Although permafrost-affected ecosystems produce much less biomass than do temperate ecosystems, permafrost-affected soils that are subject to cryoturbation (frost-churning), a cryogenic process, have a unique ability to sequester a portion of this organic matter and store it for thousands of years. A number of models have been developed to explain the mechanisms involved in cryoturbation (Mackay, 1980; Van Vliet-Lanoë, 1991; Vandenberghe, 1992). The most recent model involves the process of differential frost heave (heave–subsidence), which produces downward and lateral movement of materials (Walker *et al*., 2002; Peterson and Krantz, 2003).

Part of the organic matter produced annually by the vegetation is deposited as litter on the soil surface, with some decomposing as a result of biological activity. A large portion of this litter, however, builds up on the soil surface, forming an organic soil horizon. Cryoturbation causes some of this organic material to move down into the deeper soil layers (Bockheim and Tarnocai, 1998). Soluble organic materials move downward because of the effect of gravity and the movement of water along the thermal gradient toward the freezing front (Kokelj and Burn, 2005). Once the organic material has moved down to the cold, deeper soil layers where very little or no biological decomposition takes place, it may be preserved for many thousands of years. Radiocarbon dates from cryoturbated soil materials ranged between 490 and 11,200 yr BP (Zoltai *et al*., 1978). These dates were randomly distributed within the soil and did not appear in chronological sequence by depth (the deepest material was not necessarily the oldest), indicating that cryoturbation is an ongoing process.

The permafrost table (top of the permafrost) is very dynamic and is subject to deepening due to factors such as removal of vegetation and/or the insulating surface organic layer, wildfires, global climate change, and other natural or human activities. When this occurs, the seasonally thawed layer (active layer)

becomes deeper and the organic material is able to move even deeper into the soil (translocation). However, if such factors cause thawing of the soil and melting of the ground ice, some or all of the organic materials locked in the system could be exposed to the atmosphere. This change in soil environment gives rise to both aerobic and anaerobic decomposition, releasing carbon into the atmosphere as carbon dioxide and methane, respectively (Fig. 12-3). At this stage, the soil can become a major carbon source.

If, however, the permafrost table rises (and the active layer becomes shallower) because of reestablishment of the vegetation or buildup of the surface organic layer, this deep organic material becomes part of the permafrost and is, thus, more securely preserved. This is the main reason that permafrost-affected soils contain high amounts of organic carbon not only in the upper (0–100 cm) layer, but also in the deeper layers. These cryoturbated, permafrost-affected soils are effective carbon sinks.

Peatlands (Organic Soils)

The schematic diagram in Fig. 12-4 provides general information about the processes driving the carbon sinks and sources in peatland soils. The water-saturated conditions, low soil temperatures, and acidic conditions of northern peatlands provide an environment in which very little decomposition occurs; hence, the litter is converted to peat and preserved. This gradual buildup process has been ongoing in peatlands during the last 5,000–8,000 years, resulting in peat deposits that are an average of 2–3 m thick and, in some cases, up to 10 m thick. At this stage, peatlands can act as very effective carbon sinks for many thousands of years (Fig. 12-4).

Figure 12-4. Carbon cycle in permafrost peatlands, showing below-ground organic carbon sinks and sources.

***Carbon dynamics*:** Data for carbon accumulation in various peatland types in the permafrost regions are given in Table 12-3. Although some values for the rate of peat accumulation are higher (associated with unfrozen peatlands), the values for frozen peatlands, which are more widespread, generally range around 13 g C m^{-2} yr^{-1}. Peat accumulations in the various ecological regions were calculated on the basis of the thickness of the deposit and the date of the basal peat. The rate of peat accumulation is generally highest in the Boreal region and decreases northward (Table 12-3). Note, however, that if the surface of the peat deposit has eroded, the calculated rate of accumulation (based on the age of the basal peat and a decreased deposit thickness) will appear to be higher than it should be. This is probably the reason for some of the high rates of peat accumulation found for the Arctic region, which likely experienced a rapid rate of accumulation during the Hypsithermal Maximum with subsequent erosion of the surface of some

of the deposits reducing their thicknesses. Wildfires, decomposition, and leaching of soluble organic compounds release approximately one-third of the carbon input, causing most of the carbon loss in these peatlands.

Table 12-3. Organic carbon accumulation and loss in various Canadian peatlands. Positive values indicate net flux into the atmosphere (source); negative values indicate carbon sequestration (land sinks).

BELOW-GROUND CARBON STOCKS

The carbon content of mineral soils to a 1-m depth is 49–61 kg m^{-2} for permafrost-affected soils and 12–17 kg m^{-2} for unfrozen soils (Tables 12-4 and 12-5). The carbon content of organic soils (peatlands) for the total depth of the deposit is 81–129 kg m^{-2} for permafrost-affected soils and 43–144 kg m^{-2} for unfrozen soils (Tables 12-4 and 12-5) (Tarnocai, 1998 and 2000).

Table 12-4. Soil carbon pools and fluxes for the permafrost areas of Canada. Positive flux numbers indicate net flux into the atmosphere (source); negative values indicate carbon sequestration (land sinks).

Table 12-5. Average organic carbon content for soils in the various ecological regions (Tarnocai 1998 and 2000).

Soils in the permafrost region of North America contain 213 Gt of organic carbon (Tables 12-6 and 12-7), which is approximately 61% of the organic carbon in all soils on this continent (Lacelle *et al*., 2000). Mineral soils contain approximately 99 Gt of organic carbon in the 0- to 100-cm depth (Table 12-6). Although peatlands (organic soils) cover a smaller area than mineral soils (17% vs 83%), they contain approximately 114 Gt of organic carbon in the total depth of the deposit, or more than half (54%) of the soil organic carbon of the region (Table 12-7).

Table 12-6. Organic carbon mass in mineral soils in the various permafrost zones.

Table 12-7. Organic carbon mass in peatlands (organic soils) in the various permafrost zones.

CARBON FLUXES

Mineral Soils

Very little information is available about carbon fluxes in both unfrozen and perennially frozen mineral soils in the permafrost regions. For unfrozen upland mineral soils, Trumbore and Harden (1997)

report a carbon accumulation of 60–100 g C m^{-2} yr^{-1} (Table 12-4). They further indicate that the slow decomposition results in rapid organic matter accumulation, but the turnover time due to wildfires (every 500–1000 years) eliminates the accumulated carbon except for the deep carbon derived from roots in the subsoil. The turnover time for this deep carbon is 100–1600 years. Therefore, the carbon stocks in these unfrozen soils are low, and the turnover time of this carbon is 100 to 1000 years.

As with unfrozen mineral soils, very little information has been published on the carbon cycle in perennially frozen mineral soils. The carbon cycle in these soils differs from that in unfrozen soils in that, because of cryogenic activities, these soils are able to move the organic matter deposited on the soil surface into the deeper soil layers. Assuming that cryoturbation was active in these soils during the last six thousand years (Zoltai *et al*., 1978), an average of 9 Mt C have been added annually to these soils. Most of this carbon has been cryoturbated into the deeper soil layers, but some of the carbon in the surface organic layer is released by decomposition and, periodically, by wildfires. The schematic diagram in Fig. 12-5 shows the carbon cycle in these soils.

Figure 12-5. Carbon cycle in perennially frozen mineral soils in the permafrost region.

Peatlands (Organic Soils)

Peatland vegetation deposits various amounts of organic material (litter) annually on the peatland surface. Reader and Stewart (1972) found that the amount of litter (dry biomass) deposited annually on the bog surface in Boreal peatlands in Manitoba, Canada was 489–1750 g m^{-2}. Approximately 25% of the original litter fall was found to have decomposed during the following year. In the course of the study, they found that the average annual accumulation rate was 10% of the annual net primary production. Robinson *et al*. (2003) found that, in the Sporadic Permafrost Zone, mean carbon accumulation rates over the past 100 years for unfrozen bogs and frost mounds were 88.6 and 78.5 g m^{-2} yr^{-1}, respectively. They also found that, in the Discontinuous Permafrost Zone, the mean carbon accumulation rate during the past 1200 years in frozen peat plateaus was 13.31 g m^{-2} yr^{-1}, while in unfrozen fens and bogs the comparable rates were 20.34 and 21.81 g m^{-2} yr^{-1}, respectively.

Because peatlands cover large areas in the permafrost region of North America, their contribution to the carbon stocks is significant (Table 12-5). Zoltai *et al*. (1988) estimated that the annual carbon accumulation capacity of Boreal peatlands is approximately 9.8 Mt. Gorham (1988), in contrast, estimated that Canadian peatlands accumulate approximately 30 Mt of carbon annually.

Currently, wildfires are probably the greatest natural force in converting peatlands to a carbon source. Ritchie (1987) found that the western Canadian Boreal forests have a fire return interval of 50–100 years, while Kuhry (1994) indicated that, for wetter Sphagnum bogs, the interval is 400–1700 years. For peat

plateau bogs, each fire resulted in an average decrease in carbon mass of 1.46 kg m^{-2} and an average decrease in height of 2.74 cm, which represents about 150 years of peat accumulation (Robinson and Moore, 2000). In recent years, the number of these wildfires has increased, as has the area burned, releasing increasing amounts of carbon into the atmosphere.

The schematic diagram presented in Fig. 12-6 summarizes the carbon cycle in peatlands in the permafrost region. Based on average values for the rate of peat accumulation, approximately 17 g C m^{-2} yr^{-1}, or 18 Mt C, is added annually to peatlands in this region of North America. Approximately 1.46 kg C m^{-2} is released to the atmosphere every 600 years by wildfires in the northern boreal peatlands. In addition, decomposition of unfrozen peatlands releases approximately 2.0 g C m^{-2} yr^{-1}, and a further 2.0 g C m^{-2} yr^{-1} is released by leaching of dissolved organic carbon (DOC), leading to a carbon decrease of approximately 4 Mt annually, not including that released by wildfires (Fig. 12-6). Note that these values are based on current measurements. However, rates of peat accumulation have varied during the past 6000–8000 years, with periods during which the rate of peat accumulation was much higher than at present.

Figure 12-6. Carbon cycle in peatlands in the permafrost region.

Total Flux

Based on the limited data available for this vast, and largely inaccessible, area of the continent, approximately 27 Mt C yr^{-1} is deposited on the surface of mineral soils and peatlands (organic soils) in the permafrost region of North America. Approximately 8 Mt yr^{-1} of surface carbon (excluding vegetation) is released by decomposition and wildfires, and by leaching into the water systems. Thus, the soils in the permafrost region of North America currently act as a sink for approximately 19 Mt C yr^{-1} and as a source for approximately 8 Mt C yr^{-1} and are, therefore, a net carbon sink (Figs. 12-5 and 12-6).

POSSIBLE EFFECTS OF GLOBAL CLIMATE CHANGE

The permafrost region is unique because the soils in this vast area contain large amounts of organic materials and much of the carbon has been actively sequestered by peat accumulation (organic soils) and cryoturbation (mineral soils) and stored in the permafrost for many thousands of years. Historical patterns of climate are responsible for the large amount of carbon found in the soils of the region today, but cryoturbation is a consequence of the region's current cool to cold climate and the effects of that climate on soil hydrology. As a result, patterns of climate and climate change are dominant drivers of carbon cycling in the region. Future climate change will determine the fate of that carbon and whether the region

will remain a slow but significant carbon sink, or whether it will reverse and become a source, rapidly releasing large amounts of CO_2 and methane to the atmosphere.

Peatlands

A model for estimating the sensitivity of peatlands to global climate change was developed using current climate (1x CO_2), vegetation, and permafrost data together with the changes in these variables expected in a 2x CO_2 environment (Kettles and Tarnocai, 1999). The data generated by this model were used to produce a peatland sensitivity map. Using GIS techniques, this map was overlaid on the peatland map of Canada to determine both the sensitivity ratings of the various peatland areas and the associated organic carbon masses. The sensitivity ratings, or classes, used are no change, very slight, slight, moderate, severe, and extremely severe. Because global climate change is expected to have the greatest impact on the ecological processes and permafrost distribution in peatlands in the severe and extremely severe categories (Kettles and Tarnocai, 1999), the areas and carbon masses of peatlands in these two sensitivity classes are considered to be most vulnerable to climate change. The sensitivity ratings are determined by the degree of change in the ecological zonation combined with the degree of change in the permafrost zonation, with the greater the change, the more severe the sensitivity rating. For example, if a portion of the Subarctic becomes Boreal in ecology and the associated sporadic permafrost disappears (no permafrost remains in the region), the sensitivity of this region is rated as extremely severe. If however, a portion of the Boreal remains Boreal in ecology, but the discontinuous permafrost disappears (no permafrost remains in the region), the sensitivity of this region is rated as severe.

The peatland sensitivity model indicates that the greatest effect of global climate change will occur in the Subarctic region, where about 85% (314,270 km^2) of the peatland area and 78% (33.96 Gt) of the organic carbon mass will be severely or extremely severely affected by climate change, with 66% of the area and 57% of the organic carbon mass being extremely severely affected (Fig. 12-7) (Tarnocai, in press). The second largest effect will occur in the Boreal region, where about 49% (353,100 km^2) of the peatland area and 41% (40.20 Gt) of the organic carbon mass will be severely or extremely severely affected, with 10% of both the area and organic carbon mass being extremely severely affected. These two regions contain almost all (99%) of the Canadian peatland area and organic carbon mass that is predicted to be severely or extremely severely affected (Fig. 12-7) (Tarnocai, in press).

Figure 12-7. The organic carbon mass in the various sensitivity classes for the Subarctic and Boreal Ecoclimatic Provinces (ecological regions) (Tarnocai, in press).

In the Subarctic region and the northern part of the Boreal region, where most of the perennially frozen peatlands occur, the increased temperatures are expected to cause increased thawing of the perennially frozen peat. Thawing of the ice-rich peat and the underlying mineral soil will initially result in water-saturated conditions. These water-saturated conditions, together with the higher temperatures, result in anaerobic decomposition, leading to the production of CH_4.

In the southern part of the Boreal region, where the peatlands are generally unfrozen, the main impact is expected to be drought conditions resulting from higher summer temperatures and higher evapotranspiration. Under such conditions, peatlands become a net source of CO_2 because the oxygenated conditions lead to aerobic decomposition (Melillo *et al.*, 1990; Christensen, 1991). These dry conditions will likely also increase wildfires and, eventually, burning of peat, leading to the release of CO_2 to the atmosphere.

Permafrost-Affected Mineral Soils

The same model described above was used to determine the effect of climate change on mineral permafrost-affected soils. The model suggests that approximately 21% (11.9 Gt) of the total organic carbon in these soils could be severely or extremely severely affected by climate warming (Tarnocai, 1999). The model also suggests that the permafrost will probably disappear from the soils (the soils will become unfrozen) in the Sporadic and Isolated Patches permafrost zones. The main reason for the high sensitivity of mineral soils in these zones is that soil temperatures at both the 100- and 150-cm depths are only slightly below freezing (–0.3°C). The slightest disturbance or climate warming could initiate rapid thawing in these soils, with resultant loss of carbon (Tarnocai, 1999).

NON-CLIMATIC DRIVERS

Wildfires are an important part of the ecology of Boreal and Subarctic forests and are probably the major non-climatic drivers of carbon change in the permafrost region. There has been a rapid increase in both the frequency of fires and the area burned as a result of warmer and drier summers and increased human activity in the region. According to observations of natives, not only has the frequency of lightning strikes increased in the more southerly areas, but they have now appeared in more northerly areas where they were previously unknown. Because lightning is the major cause of wildfires in areas of little habitation, it is likely largely responsible for the increase in wildfires now being observed.

Increased human activity as a result of the construction of pipelines, roads, airstrips, and mines, expansion of agriculture, and development and expansion of town sites has disturbed the natural soil cover and exposed the organic-rich soil layers, leading to increased soil temperatures and, hence, decomposition of the exposed organic materials. Burgess and Tarnocai (1997), studying the Norman

Wells Pipeline, provide some examples of the effect of pipeline construction on frozen peatlands and permafrost in Canada.

Shoreline erosion along rivers, lakes, and oceans and thermal erosion (thermokarst) are also common processes in the permafrost region, exposing the carbon-rich frozen soil layers to the atmosphere and making the organic materials available for decomposition. As a result, carbon is released into the atmosphere as either CO_2 or methane, or it enters the water system as dissolved organic carbon.

Large hydroelectric projects in northern areas, such as Southern Indian Lake in Manitoba and the James Bay region of Quebec, have flooded vast areas of peatlands and initiated permafrost degradation and decomposition of organic carbon, some of which is released into the atmosphere as methane. Of greater immediate concern, however, is the carbon that has entered the water system as dissolved organic carbon. These compounds include contaminants such as persistent organic pollutants [e.g., PCBs, DDT, HCH, and chlorobenzene (AMAP, 2004)] that have been widely distributed in northern ecosystems over many years, much of it deposited by snowfalls, concentrated by cryoturbation, and stored in the organic soils. Of particular concern is the release of methylmercury because peatlands are net producers of this compound (Driscoll *et al*., 1998; Suchanek *et al*., 2000), which is a much greater health hazard than inorganic or elemental mercury. Natives in the regions where these hydroelectric developments have taken place have developed mercury poisoning after ingesting fish contaminated by this mercury, leading to serious health problems for many of the people. This is an example of what can happen when permafrost degrades as a result of human activities. When climate warming occurs, the widespread degradation of permafrost, with the resulting release of such dangerous pollutants into the water systems, could cause serious health problems for fish, animals, and humans that rely on such waters.

OPTIONS FOR MANAGEMENT OF CARBON IN THE PERMAFROST REGION

Although wildfires are the most effective mechanism for releasing carbon into the atmosphere, they are also an important factor in maintaining the integrity of northern ecosystems. Therefore, such fires are allowed to burn naturally and are controlled only if they are close to settlements or other manmade structures.

The construction methods currently used in permafrost terrain are designed to cause as little surface disturbance as possible and to preserve the permafrost. Thus, the construction of pipelines, airstrips, and highways is commonly carried out in the winter so that the heavy equipment used will cause minimal surface disturbance.

The greatest threat to the region is a warmer (and possibly drier) climate, which would drastically affect not only the carbon cycle, but also the biological systems, including human life. Unfortunately, we know very little about how to manage the natural systems in this new environment.

DATA GAPS AND UNCERTAINTIES

The permafrost environment is a very complex system, and the data available for it are very limited with numerous gaps and uncertainties. Information on the distribution of soils in the permafrost region is based on small-scale maps, and the carbon stocks calculated for these soils are derived from a relatively small number of datasets. Although there is some understanding of the carbon sinks and sources in these soils, the limited amount of data available make it very difficult, or impossible, to assign reliable values. Only limited amounts of flux data have been collected for the permafrost-affected soils and, in some cases, it has been collected on sites that are not representative of the overall landscape. This makes it very difficult to scale this information up for a larger area. As Davidson and Janssens (2006) state:

> "…the unresolved question regarding peatlands and permafrost is not the degree to which the currently constrained decomposition rates are temperature sensitive, but rather how much permafrost is likely to melt and how much of the peatland area is likely to dry significantly. Such regional changes in temperature, precipitation, and drainage are still difficult to predict in global circulation models. Hence, the climate change predictions, as much as our understanding of carbon dynamics, limit our ability to predict the magnitude of likely vulnerability of peat and permafrost carbon to climate change."

To obtain more reliable estimates of the carbon sinks and sources in permafrost-affected soils, we need much more detailed data on the distribution and characteristics of these soils. Carbon stock estimates currently exist only for the upper 1 m of the soil. Limited data from the Mackenzie River Valley in Canada indicate that a considerable amount of soil organic carbon occurs below the 1-m depth, even at the 3-m depth. Future estimates of carbon stocks should be extended to cover a depth of 0–2 m or, in some cases, even greater depths. More measurements of carbon fluxes and inputs are also needed if we are to understand the carbon sequestration process in these soils in the various permafrost zones. Our understanding of the effect that rapid climate warming will have on the carbon sinks and sources in these soils is also very limited. Future research should focus in greater detail on how the interactions of climate with the biological and physical environments will affect the carbon balance in permafrost-affected soils.

The changes that are occurring, and will occur, in the permafrost region are almost totally driven by natural forces and so are almost impossible for humans to manage on a large scale. Human activities, such as they are, are aimed at protecting the permafrost and, thus, preserving the carbon. Perhaps we humans should realize that there are systems (e.g., glaciers, ocean currents, droughts, and rainfall) that will be impossible for us to manage. We simply must learn to accept them and, if possible, adapt.

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Table 12-1. Areas of mineral soils in the various permafrost zones

Permafrost zones	Area ($10^3 \times km^2$)		
	Canada[a]	Alaska[b]	Total
Continuous	2001.80	353.46	2355.26
Discontinuous	636.63	479.15	1115.78
Sporadic	717.63	110.98	828.61
Isolated Patches	868.08	0.73	868.81
Total	4224.14	944.32	5168.46

[a]Calculated using the Soil Carbon of Canada Database (Soil Carbon Database Working Group, 1993).
[b]Calculated using the Northern and Mid Latitudes Soil Database (Cryosol Working Group, 2001).

Table 12-2. Areas of peatlands (organic soils) in the various permafrost zones

Permafrost zones	Area ($10^3 \times km^2$)		
	Canada[a]	Alaska[b]	Total
Continuous	176.70	51.31	228.01
Discontinuous	243.51	28.74	272.25
Sporadic	307.72	0.62	308.34
Isolated Patches	221.23	13.05	234.28
Total	949.16	93.72	1042.88

[a]Calculated using the Peatlands of Canada Database (Tarnocai *et al*., 2005).
[b]Calculated using the Northern and Mid Latitudes Soil Database (Cryosol Working Group, 2001).

Table 12-3. Organic carbon accumulation and loss in various Canadian peatlands. Positive values indicate net flux into the atmosphere (source); negative values indicate carbon sequestration (land sinks)

Peatlands	Amount of carbon
Boreal peatlands	–9.8 Mt yr^{-1} [a]
All Canadian peatlands	–30 Mt yr^{-1} [b]
All mineral and organic soils	–18 mg m^{-2} yr^{-1} [c]
Rich fens	–13.58 g m^{-2} yr^{-1} [d]
Poor fens (unfrozen, Discontinuous Permafrost Zone)	–20.34 g m^{-2} yr^{-1} [d]
Peat plateaus (frozen, Discontinuous Permafrost Zone)	–13.31 g m^{-2} yr^{-1} [d]
Collapse fens	–13.54 g m^{-2} yr^{-1} [d]
Bogs (unfrozen, Discontinuous Permafrost Zone)	–21.81 g m^{-2} yr^{-1} [d]
Dissolved organic carbon (DOC)	+2 g m^{-2} yr^{-1} [e]
Arctic peatlands	–0 to –16 cm/100 yr [f]
Subarctic peatlands	–2 to –5 cm/100 yr [f]
Boreal peatlands	–2 to –11 cm/100 yr [f]
Carbon release by each fire in northern boreal peatlands	+1.46 kg C m^{-2} [g]
Carbon release by fires in all terrain	+27 Mt yr^{-1} [h]
Carbon release by fires in Western Canadian peatlands	+5.9 Mt yr^{-1} [h]

[a] Zoltai *et al*., 1988.
[b] Gorham, 1988.
[c] Liblik *et al*., 1997.
[d] Robinson and Moore, 1999.
[e] Moore, 1997.
[f] Calculated based on the thickness of the deposit and the date of the basal peat (National Wetlands Working Group, 1988).
[g] Robinson and Moore, 2000.
[h] Turetsky *et al*., 2004.

Table 12-4. Soil carbon pools and fluxes for the permafrost areas of Canada. Positive flux numbers indicate net flux into the atmosphere (source); negative values indicate carbon sequestration (land sinks)

	Peatlands		Mineral soils		
Type	**Perennially frozen**	**Unfrozen**	**Perennially frozen**	**Unfrozen**	**Total**
Current area ($\times 10^3$ km^2)	422[a]	527[a]	2088[b]	2136[b]	5173
Current pool (Gt)	47[c]	65[a]	56[c]	28[b]	196
Current atm. flux (g m^{-2} yr^{-1})	–5.7[d]	–15.2[e]			
Carbon accumulation (g m^{-2} yr^{-1})	–13.3[f]	–20.3 to –21.8[f]		–60 to –100[g]	
Carbon release by fires (g m^{-2} yr^{-1})[h]	+7.57[i]				
Methane flux (g m^{-2} yr^{-1})		+2.0[j]			

[a]Calculated using the Peatlands of Canada Database (Tarnocai *et al*., 2005).
[b]Calculated using the Soil Carbon of Canada Database (Soil Carbon Database Working Group, 1993).
[c]Tarnocai, 1998.
[d]Using C accumulation rate of 0.13 mg ha^{-1} yr^{-1} (this report).
[e]Using C accumulation rate of 0.194 mg ha^{-1} yr^{-1} (Vitt *et al*., 2000).
[f]Robinson and Moore, 1999.
[g]Trumbore and Harden, 1997.
[h]Fires recur every 150–190 years (Kuhry, 1994; Robinson and Moore, 2000).
[i]Robinson and Moore, 2000.
[j]Moore and Roulet, 1995.

Table 12-5. Average organic carbon content for soils in the various ecological regions (Tarnocai, 1998 and 2000)

Ecological regions	Average carbon content (kg m^{-2})			
	Mineral soils[a]		Organic soils (peatlands)[b]	
	Frozen	Unfrozen	Frozen	Unfrozen
Arctic	49	12	86	43
Subarctic	61	17	129	144
Boreal	50	16	81	134

[a]For the 1-m depth.
[b]For the total depth of the peat deposit.

Table 12-6. Organic carbon mass in mineral soils in the various permafrost zones

Permafrost zones	Carbon mass[a] (Gt)		
	Canada[b]	Alaska[c]	Total
Continuous	51.10	9.04	60.14
Discontinuous	10.33	4.82	15.15
Sporadic	9.15	0.75	9.90
Isolated Patches	13.59	0	13.59
Total	84.17	14.61	98.78

[a]Calculated for the 0–100 cm depth.
[b]Calculated using the Soil Carbon of Canada Database (Soil Carbon Database Working Group, 1993).
[c]Calculated using the Northern and Mid Latitudes Soil Database (Cryosol Working Group, 2001).

Table 12-7. Organic carbon mass in peatlands (organic soils) in the various permafrost zones

Permafrost zones	Carbon mass[a] (Gt)		
	Canada[b]	Alaska[c]	Total
Continuous	21.82	1.46	23.28
Discontinuous	26.54	0.84	27.38
Sporadic	30.66	0.27	30.93
Isolated Patches	32.95	0	32.95
Total	111.97	2.57	114.54

[a]Calculated for the total depth of the peat deposit.
[b]Calculated using the Peatlands of Canada Database (Tarnocai *et al*., 2005).
[c]Calculated using the Northern and Mid Latitudes Soil Database (Cryosol Working Group, 2001).

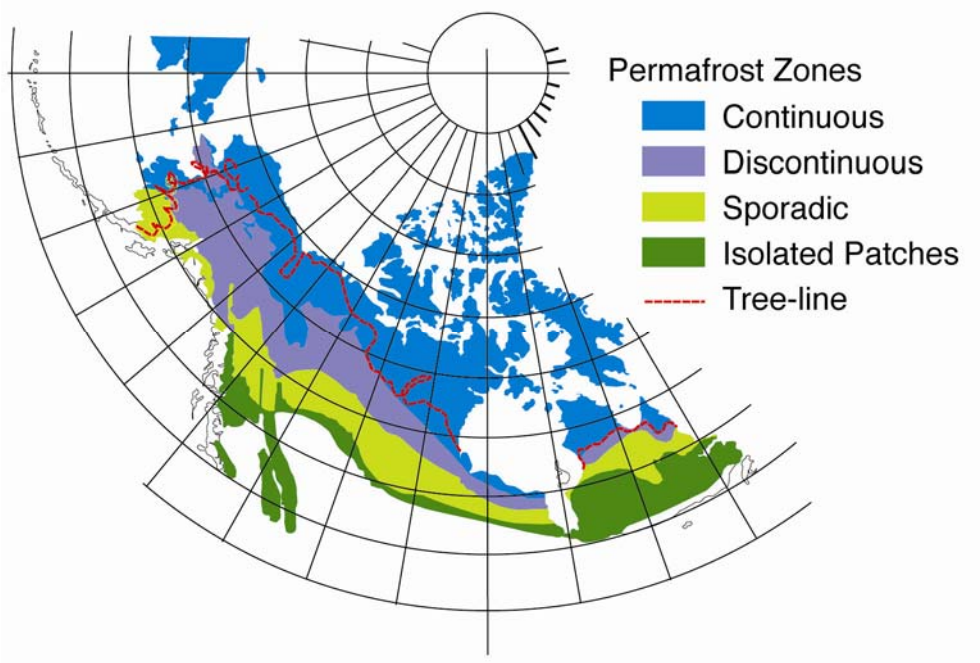


Fig. 12-1. Permafrost zones in North America (Brown *et al.*, 1997).

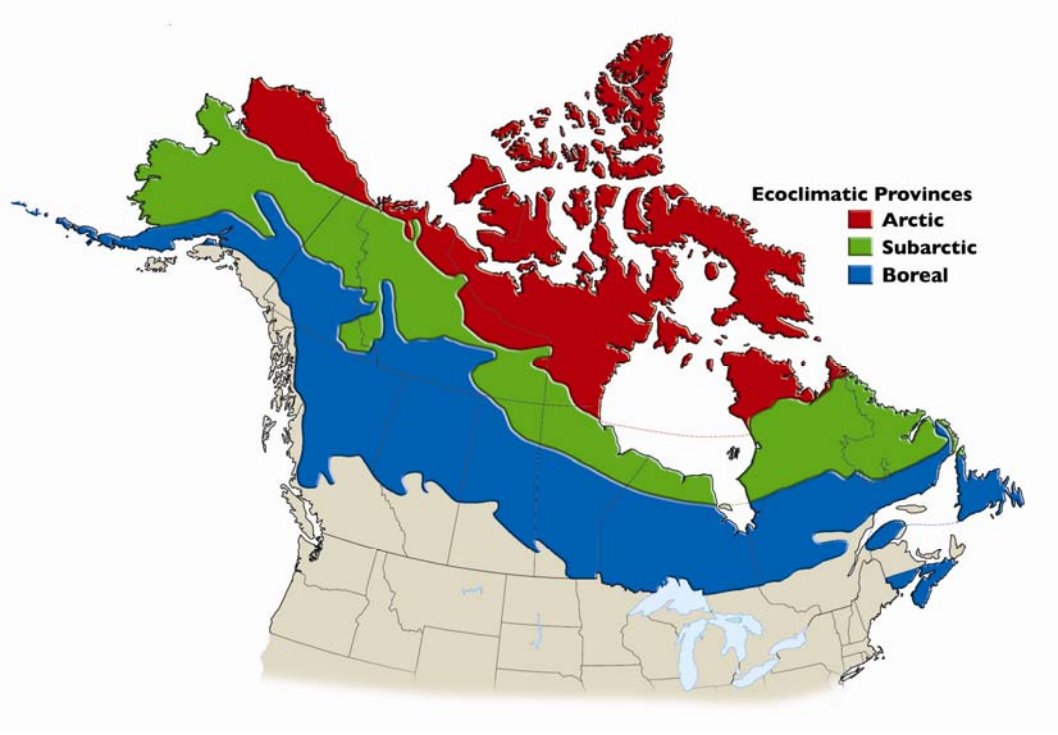


Fig. 12-2. Arctic, Subarctic, and Boreal ecoclimatic provinces (ecological regions) in North America (Ecoregions Working Group, 1989; Baily and Cushwa, 1981).

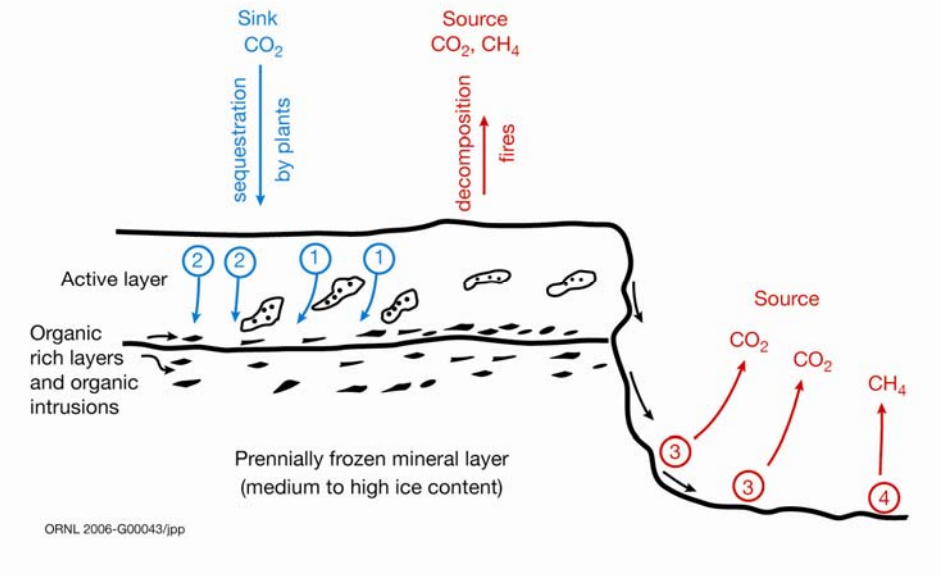
1

Carbon sinks

Permafrost-affected soil with a thick surface organic layer, dark-colored organic intrusions in the brown soil layer, and an underlying frozen, high-ice-content layer. The organic intrusions were translocated from the surface by cryoturbation. (Mackenzie Valley, Canada)

Carbon sources

Eroding high-ice-content permafrost soil composed of a dark frozen soil layer with an almost pure ice layer below. The thawing process generated a flow slide in which high-organic- content soil materials slumped into the water-saturated environment. (Mackenzie Delta area, Canada)



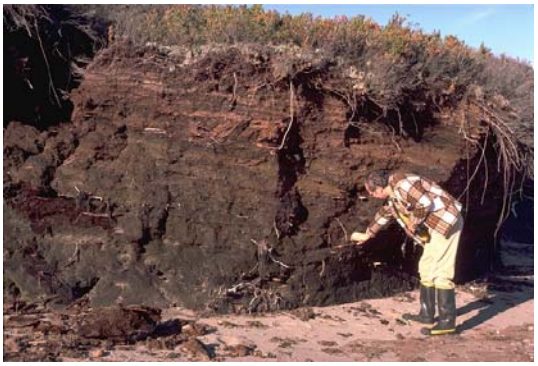
Perennially frozen deposit composed of an active layer that freezes and thaws annually and an underlying perennially frozen layer that has a high ice content.

Organic material deposited annually on the soil surface builds up as an organic soil layer. Some of this surface organic material is translocated into the deeper soil layers by cryoturbation (1). In addition, soluble organic matter is translocated into the deeper soil layers by movement of water to the freezing front and by gravity (2). Because these deeper soil layers have low temperatures (0 to -15°C), the organic material decomposes very slowly. Thus more organic material accumulates as long as the soil is frozen. In this state, the permafrost soil acts as a carbon sink.

Thermal erosion initiated by climate warming, wildfires or human activity causes the high-ice-content mineral soils to thaw, releasing the organic materials locked in the system. In this environment aerobic (3) and anaerobic (4) decomposition occurs releasing carbon dioxide and methane. In this state, the soil is a source of carbon.

2 **Fig. 12-3. Carbon cycle in permafrost-affected upland (mineral) soils, showing below-ground organic**
3 **carbon sinks and sources.**

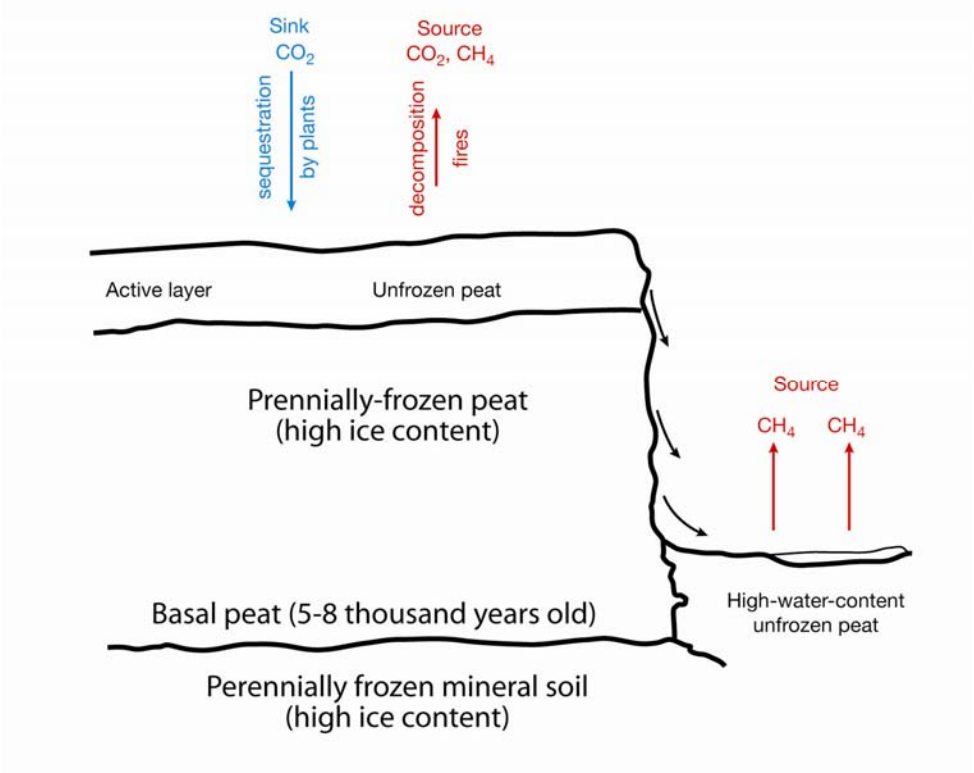
Carbon sinks



Perennially frozen peat deposit with multiple dark-colored peat layers. (Mackenzie River Delta area, Canada)

Carbon sources

Eroding perennially frozen peat deposit, showing the large blocks of peat slumping into the water- saturated collapsed area. (Fort Simpson area, Canada)



Perennially frozen peat deposits consist of an active layer that freezes and thaws annually and an underlying perennially frozen layer composed of ice-rich frozen peat and mineral materials.

Organic material is deposited annually on the peatland surface. Although a large portion (≥90%) of this organic material decomposes, the remainder is added to the peat deposit, producing an annual peat accumulation. The low soil temperatures (0 to –15°C) and the water-saturated and acid conditions cause this added organic carbon to be preserved and stored. This has been occurring for the last 5–8 thousand years. In this state, the peatland is a carbon sink.

Thermal erosion (thawing) of frozen peat deposits occurs as a result of climate change, wildfires, or human disturbances, releasing large amounts of water from the melting ice. This is mixed with the slumped peat material, initiating anaerobic decomposition in the much warmer environment. Anaerobic decomposition produces methane, which is expelled into the atmosphere. In this state, the peatland is a source of carbon.

Fig. 12-4. Carbon cycle in permafrost peatlands, showing below-ground organic carbon sinks and sources.

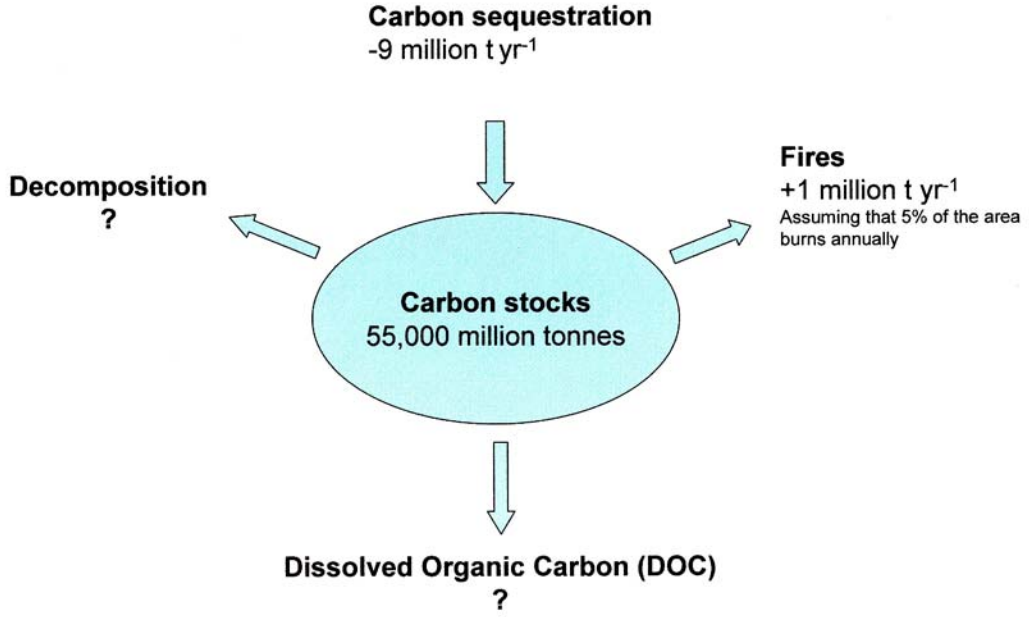


Fig. 12-5. Carbon cycle in perennially frozen mineral soils in the permafrost region.

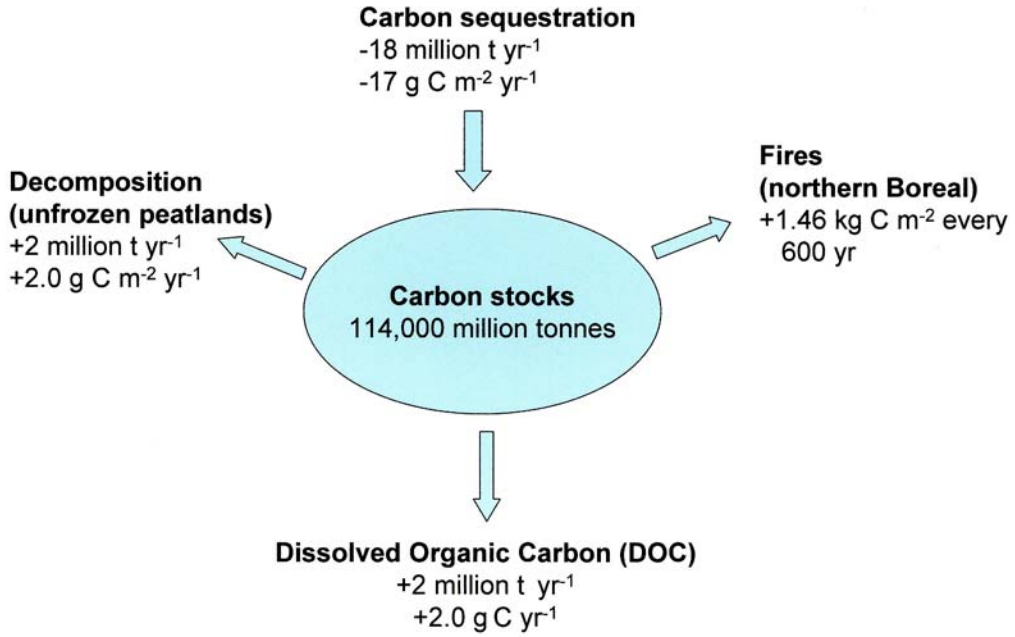


Fig. 12-6. Carbon cycle in peatlands in the permafrost region.

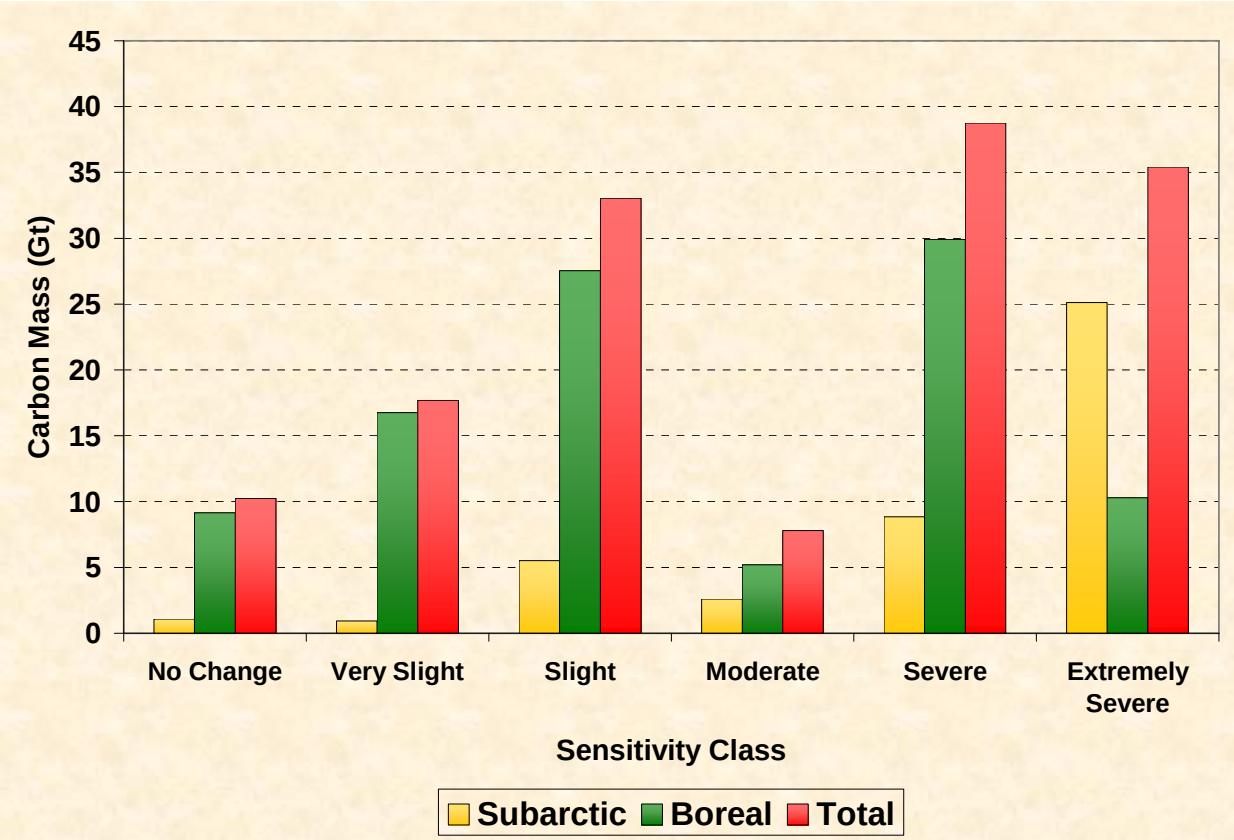


Fig. 12-7. The organic carbon mass in the various sensitivity classes for the Subarctic and Boreal Ecoclimatic Provinces (ecological regions) (Tarnocai, in press).

Chapter 13. Wetlands

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KEY FINDINGS

- North America is home to approximately 41% of the global wetland area, encompassing about 2.5 million km^2 with a carbon pool of approximately 220 Gt, mostly in peatland soils.
- North American wetlands currently are a CO_2 sink of approximately 70 Mt C yr^{-1}, but that estimate has an uncertainty of greater than 100%. North American wetlands are also a source of approximately 26 Mt yr^{-1} of methane, a more potent atmospheric heat-trapping gas. The uncertainty in that flux is also greater than 100%.
- Historically, the destruction of North American wetlands through land-use change has reduced carbon storage in wetlands by 43 Mt C yr^{-1}, primarily through the oxidation of carbon in peatland soils as they are drained and a more general reduction in carbon sequestration capacity of wetlands converted to other land uses. Methane emissions have also declined with the loss of wetland area.
- Projections of future carbon storage and methane emissions of North American wetlands are highly uncertain and complex, but the large carbon pools in peatlands may be at risk for oxidation and release to the atmosphere as CO_2 if they become substantially warmer and drier. Methane emissions may increase with warming, but the response will likely vary with wetland type and with changes in precipitation.
- Because of the potentially significant role of North American wetlands in methane production, the activities associated with the restoration, creation and protection of wetlands are likely to focus on the ecosystem services that wetlands provide, such as filtering of toxics, coastal erosion protection, wildlife habitat, and havens of biodiversity, rather than on carbon sequestration per se.
- Research needs to reduce the uncertainties in carbon storage and fluxes in wetlands to provide information about management options in terms of carbon sequestration and trace gas fluxes.

INTRODUCTION

While there are a variety of legal and scientific definitions of a wetland (National Research Council, 1995; National Wetlands Working Group, 1997), most emphasize the presence of waterlogged conditions in the upper soil profile during at least part of the growing season, and plant species and soil conditions that reflect these hydrologic conditions. Waterlogging tends to suppress microbial decomposition more than plant productivity, so wetlands are known for their ability to accumulate large amounts of soil carbon, most spectacularly seen in large peat deposits that are often many meters deep. Thus, when examining carbon dynamics, it is important to distinguish between freshwater wetlands with surface soil organic matter deposits >40 cm thick (i.e., peatlands) and those with lesser amounts of soil organic matter (i.e., freshwater mineral-soil wetlands, FWMS). Some wetlands have permafrost; fluxes and pools in wetlands with and without permafrost are discussed separately in Appendix 13A. We also differentiate between freshwater wetlands and estuarine wetlands (salt marshes, mangroves, and mud flats) with marine-derived salinity.

Peatlands occupy about 3% of the terrestrial global surface, yet they contain 16–33% of the total soil carbon pool (Gorham, 1991; Maltby and Immirzi, 1993). Most peatlands occur between 50 and 70º N, although significant areas occur at lower latitudes (Matthews and Fung, 1987; Aselmann and Crutzen, 1989; Maltby and Immirzi, 1993). Large areas of peatlands exist in Alaska, Canada, and in the northern midwestern, northeastern, and southeastern United States (Bridgham *et al.*, 2000). Because this peat formed over thousands of years, these areas represent a large carbon pool but with relatively slow rates of accumulation. By comparison, estuarine wetlands and some freshwater mineral-soil wetlands rapidly sequester carbon as soil organic matter due to rapid burial in sediments. Large areas of wetlands have been converted to other land uses globally and in North America (Dugan, 1993; OECD, 1996), which may have resulted in a net flux of carbon to the atmosphere (Armentano and Menges, 1986; Maltby and Immirzi, 1993). Additionally, wetlands emit 92–237 Mt methane (CH_4) yr^{-1}, which is a large fraction of the total annual global flux of about 600 Mt CH_4 yr^{-1} (Ehhalt *et al.*, 2001). This is important because methane is a potent greenhouse gas, second in importance to only carbon dioxide (Ehhalt *et al.*, 2001). A number of previous studies have examined the role of peatlands in the global carbon balance (reviewed in Mitra *et al.*, 2005). Roulet (2000) focused on the role of Canadian peatlands in the Kyoto process. Here we augment these previous studies by considering all types of wetlands (not just peatlands) and integrate new data to examine the carbon balance in the wetlands of Canada, the United States, and Mexico. We also briefly compare these values to those from global wetlands.

Given that many undisturbed wetlands are a natural sink for carbon dioxide and a source of methane, a note of caution in interpretation of our data is important. Using the International Panel on Climate Change (IPCC) terminology, a radiative forcing denotes “an externally imposed perturbation in the

radiative energy budget of the Earth's climate system" (Ramaswamy *et al.*, 2001). Thus, it is the change from a baseline condition in greenhouse gas fluxes in wetlands that constitute a radiative forcing that will impact climate change, and carbon fluxes in unperturbed wetlands are important only in establishing a baseline condition. For example, historical steady state rates of methane emissions from wetlands have zero net radiative forcing, but an increase in methane emissions due to climatic warming would constitute a positive radiative forcing. Similarly, steady state rates of soil carbon sequestration in wetlands have zero net radiative forcing, but the lost sequestration capacity and the oxidation of the extant soil carbon pool in drained wetlands are both positive radiative forcings. Here we consider changes from a historical baseline of about 1800 A.D. to present and future emissions of greenhouse gas fluxes in North American wetlands.

INVENTORIES

Current Wetland Area and Rates of Loss

The current and historical wetland area and rates of loss are the basis for all further estimates of pools and fluxes in this chapter. The loss of wetlands has caused the oxidation of their soil carbon, particularly in peatlands, reduced their ability to sequester carbon, and reduced their emissions of methane. The strengths and weakness of the wetland inventories of Canada, the United States, and Mexico are discussed in Appendix 13A.

The conterminous United States has 312,000 km^2 of FWMS wetlands, 93,000 km^2 of peatlands, and 23,000 km^2 of estuarine wetlands, which encompass 5.5% of the land area (Table 13-1). This represents just 48% of the original wetland area in the conterminous United States (Table 13A-1 in Appendix 13A). However, wetland losses in the United States have declined from 1,855 $km^2\ yr^{-1}$ in the 1950s–1970s to 237 $km^2\ yr^{-1}$ in the 1980s–1990s (Dahl, 2000). Such data mask large differences in loss rates among wetland classes and conversion of wetlands to other classes, with potentially large effects on carbon stocks and fluxes (Dahl, 2000). For example, the majority of wetland losses in the United States have occurred in FWMS wetlands. As of the early 1980s, 84% of U.S. peatlands were unaltered (Armentano and Menges, 1986; Maltby and Immirzi, 1993; Rubec, 1996), and, given the current regulatory environment in the United States, recent rates of loss are likely small.

Table 13-1. The area, carbon pool, net carbon balance, and methane flux from wetlands in North America and the world.

Canada has 1,301,000 km^2 of wetlands, covering 14% of its land area, of which 87% are peatlands (Table 13-1). Canada has lost about 14% of its wetlands, mainly due to agricultural development of

FWMS wetlands (Rubec, 1996), although the ability to estimate wetland losses in Canada is limited by the lack of a regular wetland inventory.

The wetland area in Mexico is estimated at 36,000 km^2 (Table 13-1), with an estimated historical loss of 16,000 km^2 (Table 13A-1 in Appendix 13A). However, given the lack of a nationwide wetland inventory and a general paucity of data, this number is highly uncertain.

Problems with inadequate wetland inventories are even more prevalent in lesser developed countries (Finlayson *et al.*, 1999). We estimate a global wetland area of 6.0×10^6 km^2 (Table 13-1); thus, North America currently has about 43% of the global wetland area. It has been estimated that about 50% of the world's historical wetlands have been converted to other uses (Moser *et al.*, 1996).

Carbon Pools

We estimate that North American wetlands have a current soil and plant carbon pool of 220 Gt, of which approximately 98% is in the soil (Table 13-1). The majority of this carbon is in peatlands, with FWMS wetlands contributing about 18% of the carbon pool. The large amount of soil carbon (27 Gt) in Alaskan FWMS wetlands had not been identified in previous studies (see Appendix 13A).

Soil Carbon Fluxes

North American peatlands currently have a net carbon balance of about -18 Mt C yr^{-1} (Table 13-1), but several large fluxes are incorporated into this estimate. **(Negative numbers indicate net fluxes into the ecosystem, whereas positive numbers indicate next fluxes into the atmosphere.)** Peatlands sequester -34 Mt C yr^{-1} (Table 13A-2 in Appendix 13A), but peatlands in the conterminous United States that have been drained for agriculture and forestry had a net oxidative flux of 18 Mt C yr^{-1} as of the early 1980s (Armentano and Menges, 1986). Despite a substantial reduction in the rate of wetland loss since the 1980s (Dahl, 2000), drained organic soils continue to lose carbon over many decades, so the actual flux to the atmosphere is probably close to the 1980s estimate. There has also been a loss in sequestration capacity in drained peatlands of 2.4 Mt C yr^{-1} (Table 13-1), so the overall soil carbon sink of North American peatlands is about 20 Mt C yr^{-1} smaller than it would have been in the absence of disturbance.

Very little attention has been given to the role of FWMS wetlands in North American or global carbon balance estimates, with the exception of methane emissions. Carbon sequestration associated with sediment deposition is a potentially large, but poorly quantified, flux in wetlands (Stallard, 1998). Using a review by Johnston (1991), we calculate a substantial carbon accumulation rate in sedimentation in FWMS wetlands of -129 g C m^{-2} yr^{-1} (see Appendix 13A). However, it is unlikely that the actual sequestration rate is this high. Researchers may have preferentially chosen wetlands with high sedimentation rates to study this process, providing a bias towards greater carbon sequestration. More

fundamentally, it is important to distinguish between autochthonous carbon (derived from on-site plant production) and allochthonous carbon (imported from outside the wetland) in soil carbon storage. Almost all of the soil carbon stored in peatlands is of autochthonous origin and represents sequestration of atmospheric carbon dioxide at the landscape scale. In contrast, much of the soil carbon that is stored in FWMS wetlands is likely of allochthonous origin. At a landscape scale, redistribution of sediments from uplands to wetlands does not represent net carbon sequestration if the decomposition rate of carbon is the same in both environments. Carbon exported from upland source areas is likely to be relatively recalcitrant and physically protected from decomposers by association with mineral soil. Thus, despite the anaerobic conditions in wetlands, decomposition rates in deposited sediments may not be substantially lower than in the uplands from which those sediments were eroded. There are no data to our knowledge to evaluate these important caveats. Because of this reasoning, we somewhat arbitrarily assumed that sediment carbon sequestered in FWMS wetlands is of allochthonous origin and decomposed 25% slower than in the uplands from which the sediment was derived. Accordingly, we reduced our calculated rates of *landscape-level* carbon sequestration in FWMS wetlands by 75% to -34 g C m^{-2} yr^{-1} (Table 13A-2 in Appendix 13A). Nevertheless, this still represents a substantial carbon sink. For example, Stallard (1998) estimated that global wetlands are a large sediment sink, with a flux on the order of -1 Gt C yr^{-1}. However, this analysis was based on many assumptions and was acknowledged by the author to be a first guess at best.

Decomposition of soil carbon in FWMS wetlands that have been converted to other land uses appears to be responsible for only a negligible loss of soil carbon currently (Table 13A-2 in Appendix 13A). However, due to the historical loss of FWMS wetland area, we estimate that they currently sequester 21 Mt C yr^{-1} less than they did prior to disturbance (Table 13-1). This estimate has the same unknowns described in the previous paragraph on current sediment carbon sequestration in FWMS wetlands.

We estimate that estuarine wetlands currently sequester -9.7 Mt C yr^{-1}, with a historical reduction in sequestration capacity of 1.6 Mt C yr^{-1} due to loss of area (Table 13-1). However, the reduction is almost certainly greater because our 'historical' area is only from the 1950s. Despite the relatively small area of estuarine wetlands, they currently contribute about 26% of total wetland carbon sequestration in the conterminous United States and about 14% of the North American total. Estuarine wetlands sequester carbon at a rate about 10 times higher on an area basis than other wetland ecosystems due to high sedimentation rates, high soil carbon content, and constant burial due to sea level rise. Estimates of sediment deposition rates in estuarine wetlands are robust, but it is unknown to what extent soil carbon sequestration is due to allochthonous versus autochthonous carbon. As with FWMS wetlands, the contribution of soil carbon sequestration in estuarine wetlands to the North American carbon budget is overestimated to the extent that allochthonous carbon simply represents redistribution of carbon in the

landscape. There is also large uncertainty in the area and carbon content of mud flats, particularly in Canada and Mexico.

Overall, North American wetland soils appear to be a substantial carbon sink with a net flux of -70 Mt C yr^{-1} (with very large error bounds because of FWMS wetlands) (Table 13-1). The large-scale conversion of wetlands to upland uses has led to a reduction in the wetland soil carbon sequestration capacity of 25 Mt C yr^{-1} from the likely historical rate (Table 13-1), but this estimate is driven by large losses of FWMS wetlands with their highly uncertain sedimentation carbon sink. Adding in the current net oxidative flux of 18 Mt C yr^{-1} from conterminous U.S. peatlands, we estimate that North American wetlands currently sequester 43 Mt C yr^{-1} less than they did historically (Table 13A-2 in Appendix 13A). Furthermore, North American peatlands and FWMS wetlands have lost 2.6 Gt and 4.9 Gt of soil carbon, respectively, and collectively they have lost 2.4 Gt of plant carbon since approximately 1800. Very little data exist to estimate carbon fluxes for freshwater Mexican wetlands, but because of their small area, they will not likely have a large impact on the overall North American estimates.

The global wetland soil carbon balance has only been examined in peatlands. The current change in soil carbon flux in peatlands is about 176 to 266 Mt C yr^{-1} (Table 13A-2 in Appendix 13A), largely due to the oxidation of peat drained for agriculture and forestry and secondarily due to peat combustion for fuel (Armentano and Menges, 1986; Maltby and Immirzi, 1993). Thus, globally peatlands are a moderate atmospheric source of carbon. The cumulative historical shift in soil carbon stocks has been estimated to be 5.5 to 7.1 Gt C (Maltby and Immirzi, 1993).

Methane and Nitrous Oxide Emissions

We estimate that North American wetlands emit 26 Mt CH_4 yr^{-1} (Table 13-1), a value that is substantially higher than the previous estimate by Bartlett and Harriss (1993) (see Appendix 13A). A mechanistic methane model yielded similar rates of 3.8 and 7.1 Mt CH_4 yr^{-1} for Alaska and Canada, respectively (Zhuang *et al.*, 2004). For comparison, a regional inverse atmospheric modeling approach estimated total methane emissions (from all sources) of 16 and 54 Mt CH_4 yr^{-1} for boreal and temperate North America, respectively (Fletcher *et al.*, 2004b).

Methane emissions are currently about 24 Mt CH_4 yr^{-1} less than they were historically in North American wetlands (see Table 13A-4 in Appendix 13A) because of the loss of wetland area. We do not consider the effects of conversion of wetlands from one type to another (Dahl, 2000), which may have a significant impact on methane emissions. Similarly, we estimate that global methane emissions from natural wetlands are only about half of what they were historically due to loss of area (Table 13A-4 in Appendix 13A). However, this may be an overestimate because wetland losses have been higher in more

developed countries than less developed countries (Moser *et al.*, 1996), and wetlands at lower latitudes have higher emissions on average (Bartlett and Harriss, 1993).

When we multiplied the very low published estimates of nitrous oxide emissions from natural and disturbed wetlands (Joosten and Clarke, 2002) by North American wetland area, the flux was insignificant (data not shown). However, nitrous oxide emissions have been measured in few wetlands, particularly in FWMS wetlands and wetlands with high nitrogen inputs (e.g., from agricultural run-off), where emissions might be expected to be higher.

We use global warming potentials (GWPs) as a convenient way to compare the relative contributions of carbon dioxide and methane fluxes in North American wetlands to the Earth's radiative balance. The GWP is the radiative effect of a pulse of a substance into the atmosphere relative to carbon dioxide over a particular time horizon (Ramaswamy *et al.*, 2001). However, it is important to distinguish between *radiative balance*, which refers to the static radiative effect of a substance, and *radiative forcing* which refers to an externally imposed perturbation on the Earth's radiative energy budget (Ramaswamy *et al.*, 2001). Thus, changes in radiative balance lead to a radiative forcing, which subsequently leads to a change in the Earth's surface temperature. For example, wetlands have a large effect on the Earth's radiative balance through high methane emissions, but, it is only to the extent that emissions change through time that they represent a positive or negative radiative forcing and impact climate change.

Methane has GWPs of 1.9, 6.3, and 16.9 CO_2-carbon equivalents on a mass basis across 500-year, 100-year, and 20-year time frames, respectively (Ramaswamy *et al.*, 2001)[1]. Depending upon the time frame and within the large confidence limits of many of our estimates in Table 13-1, the *net radiative balance* of North American wetlands as a whole currently are in a range between approximately neutral and a large source of net CO_2-carbon equivalents to the atmosphere (note that we discuss *net radiative forcing* in *Trends and Drivers of Wetland Carbon Fluxes*). It is likely that FWMS wetlands, with their high methane emissions, are a net source of CO_2-carbon equivalents to the atmosphere. In contrast, estuarine wetlands are a net sink for CO_2-carbon equivalents because they support both rapid rates of carbon sequestration and low methane emissions. However, caution should be exercised in using GWPs to draw conclusions about changes in the net flux of CO_2-carbon equivalents because GWPs are based upon a pulse of a gas into the atmosphere, whereas carbon sequestration is more or less continuous. For example, if one considers continuous methane emissions and carbon sequestration in peat over time, most peatlands are a net sink for CO_2-carbon equivalents because of the long lifetime of carbon dioxide sequestered as peat (Frolking *et al*., 2006).

[1]GWPs in Ramaswamy *et al.* (2001) were originally reported in CO_2-mass equivalents. We have converted them into CO_2-carbon equivalents so that the net carbon balance and methane flux columns in Table 13-1 can be directly compared by multiplying methane fluxes by the GWPs given here.

Plant Carbon Fluxes

We estimate that wetland forests in the conterminous United States currently sequester -10.3 Mt C yr^{-1} as increased plant biomass (see Table 13A-3 in Appendix 13A). Sequestration in plants in undisturbed wetland forests in Alaska, many peatlands, and estuarine wetlands is probably minimal, although there may be substantial logging of Canadian forested peatlands that we do not have the data to account for.

TRENDS AND DRIVERS OF WETLAND CARBON FLUXES

While extensive research has been done on carbon cycling and pools in North American wetlands, to our knowledge, this is the first attempt at an overall carbon budget for all of the wetlands of North America, although others have examined the carbon budget for North American peatlands as part of global assessments (Armentano and Menges, 1986; Maltby and Immirzi, 1993; Joosten and Clarke, 2002). Historically, the destruction of wetlands through land-use changes has had the largest effect on the carbon fluxes and, consequently, the radiative forcing of North American wetlands. The primary effects have been a reduction in their ability to sequester carbon (a small to moderate increase in radiative forcing depending on carbon sequestration by sedimentation in FWMS and estuarine wetlands), oxidation of their soil carbon reserves upon drainage (a small increase in radiative forcing), and a reduction in the emission of methane to the atmosphere (a large decrease in radiative forcing) (Table 13A-1 and Appendix 13A). Globally, the disturbance of peatlands appears to have shifted them into a net source of carbon to the atmosphere. Any positive effect of wetland loss due to a reduction in their methane emissions, and hence radiative forcing, will be more than negated by the loss of the many ecosystem services they provide such as havens for biodiversity, recharge of groundwater, reduction in flooding, fish nurseries, etc. (Zedler and Kercher, 2005).

A majority of the effort in examining future global change impacts on wetlands has focused on northern peatlands because of their large soil carbon reserves, although under current climate conditions they have modest methane emissions (Moore and Roulet, 1995; Roulet, 2000; Joosten and Clarke, 2002, and references therein). The effects of global change on carbon sequestration in peatlands are probably of minor importance as a global flux because of the relatively low rate of peat accumulation. However, losses of soil carbon stocks in peatlands drained for agriculture and forestry (Table 13A-2 in Appendix 13A) attest to the possibility of large losses from the massive soil carbon deposits in northern peatlands if they become substantially drier in a future climate. Furthermore, Turetsky *et al.* (2004) estimated that up to 5.9 Mt C yr^{-1} are released from western Canadian peatlands by fire and predicted that increases in fire frequency may cause these systems to become net atmospheric carbon sources.

Our compilation shows that attention needs to be directed toward understanding climate change impacts to FWMS wetlands, which collectively emit over 3-times more methane than North American peatlands and potentially sequester an equivalent amount of carbon. The effects of changing water table depths are somewhat more tractable in FWMS wetlands than peatlands because FWMS wetlands have less potential for oxidation of soil organic matter. In forested FWMS wetlands, increased precipitation and runoff may increase radiative forcing by simultaneously decreasing wood production and increasing methanogenesis (Megonigal *et al*., 2005). The influence of changes in hydrology on methane emissions, plant productivity, soil carbon preservation, and sedimentation will need to be addressed in order to fully anticipate climate change impacts on radiative forcing in these systems.

The effects of global change on estuarine wetlands is of concern because sequestration rates are rapid, and they can be expected to increase in proportion to the rate of sea level rise provided estuarine wetland area does not decline. Because methane emissions from estuarine wetlands are low, this increase in sequestration capacity could represent a net decrease in radiative forcing, depending on how much of the sequestered carbon is autochthonous. The rate of loss of tidal wetland area has declined in past decades due to regulations on draining and filling activities (Dahl, 2000). However, rapid conversion to open water is occurring in coastal Louisiana (Bourne, 2000) and Maryland (Kearney and Stevenson, 1991), suggesting that marsh area will decline with increased rates of sea level rise (Kearney *et al*., 2002). A multitude of human and climate factors are contributing to the current losses (Turner, 1997; Day Jr. *et al*., 2000; Day Jr. *et al*., 2001). Although it is uncertain how global changes in climate, eutrophication, and other factors will interact with sea level rise (Najjar *et al*., 2000), it is likely that increased rates of sea level rise will cause an overall decline in estuarine marsh area and soil carbon sequestration.

One of the greatest concerns is how climate change will affect future methane emissions from wetlands because of their large GWP. Wetlands emit about 107 Mt CH_4 yr^{-1} (Table 4), or 20% of the global total. Increases in atmospheric methane concentrations over the past century have had the second largest radiative forcing (after carbon dioxide) in human-induced climate change (Ehhalt *et al*., 2001). Moreover, methane fluxes from wetlands have provided an important radiative feedback on climate over the geologic past (Chappellaz *et al*., 1993; Blunier *et al*., 1995; Petit *et al*., 1999). The large global warming observed since the 1990s may have resulted in increased methane emissions from wetlands (Fletcher *et al*., 2004a; Wang *et al*., 2004; Zhuang *et al*., 2004).

Data (Bartlett and Harriss, 1993; Moore *et al*., 1998; Updegraff *et al*., 2001) and modeling (Gedney *et al*., 2004; Zhuang *et al*., 2004) strongly support the contention that water table position and temperature are the primary environmental controls over methane emissions. How this generalization plays out with future climate change is, however, more complex. For example, most climate models predict much of Canada will be warmer and drier in the future. Based upon this prediction, Moore *et al.* (1998) proposed a

variety of responses to climate change in the carbon fluxes from different types of Canadian peatlands. Methane emissions may increase in collapsed former-permafrost bogs (which will be warmer and wetter) but decrease in fens and other types of bogs (warmer and drier). A methane-process model predicted that modest warming will increase global wetland emissions, but larger increases in temperature will decrease emissions because of drier conditions (Cao *et al*., 1998).

The direct, non-climatic effects of increasing atmospheric CO_2 on carbon cycling in wetland ecosystems has received far less attention than upland systems. Field studies have been done in tussock tundra (Tissue and Oechel, 1987; Oechel *et al*. 1994), bog-type peatlands (Hoosbeek *et al*., 2001), rice paddies (Kim *et al*., 2001), and a salt marsh (Rasse *et al*., 2005); and a somewhat wider variety of wetlands have been studied in small scale glasshouse systems. Temperate and tropical wetland ecosystems consistently respond to elevated CO_2 with an increase in photosynthesis and/or biomass (Vann and Megonigal, 2003). By comparison, the response of northern peatland plant communities has been inconsistent. A hypothesis that remains untested is that the elevated CO_2 response of northern peatlands will be limited by nitrogen availability. In an *in situ* study of tussock tundra, complete photosynthetic acclimation occurred when CO_2 was elevated, but acclimation was far less severe with both elevated CO_2 and a 4^{o}C increase in air temperature (Oechel *et al*., 1994). It was hypothesized that soil warming relieved a severe nutrient limitation on photosynthesis by increasing nitrogen mineralization.

A consistent response to elevated CO_2-enhanced photosynthesis in wetlands is an increase in CH_4 emissions ranging from 50 to 350% (Megonigal and Schlesinger, 1997; Vann and Megonigal, 2003). It is generally assumed that the increased supply of plant photosynthate stimulates anaerobic microbial carbon metabolism, of which CH_4 is a primary end product. A doubling of CH_4 emissions from wetlands due to elevated CO_2 constitutes a positive feedback on radiative forcing because CO_2 is rapidly converted to a more effective greenhouse gas (CH_4).

An elevated CO_2-induced increase in CH_4 emissions may be offset by an increase in carbon sequestration in soil organic matter or wood. Although there are very little data to evaluate this hypothesis, a study on seedlings of a wetland-adapted tree species reported that elevated CO_2 stimulated photosynthesis and CH_4 emissions, but not growth, under flooded conditions (Megonigal *et al*., 2005). It is possible that elevated CO_2 will stimulate soil carbon sequestration, particularly in tidal wetlands experiencing sea level rise, but a net loss of soil carbon is also possible due to priming effects (Hoosbeek and VanKessel, 2004; Lichter *et al*., 2005). Elevated CO_2 has the potential to influence the carbon budgets of adjacent aquatic ecosystems by increasing export of DOC (Freeman *et al*., 2004) and DIC (Marsh *et al*., 2005).

Other important anthropogenic forcing factors that will affect future methane emissions include atmospheric sulfate deposition (Vile *et al.*, 2003; Gauci *et al.*, 2004) and nutrient additions (Keller *et al.*, 2005). These external forcing factors in turn will interact with internal ecosystem constraints such as pH and carbon quality (Moore and Roulet, 1995; Bridgham *et al.*, 1998), anaerobic carbon flow (Hines and Duddleston, 2001), and net ecosystem productivity and plant community composition (Whiting and Chanton, 1993; Updegraff *et al.*, 2001; Strack *et al.*, 2004) to determine the actual response.

OPTIONS AND MEASURES

Wetland policies in the United States and Canada are driven by a variety of federal, state or provincial, and local laws and regulations in recognition of the many wetland ecosystem services and large historical loss rates (Lynch-Stewart *et al.*, 1999; National Research Council, 2001; Zedler and Kercher, 2005). Thus, any actions to enhance the ability of wetlands to sequester carbon, or reduce their methane emissions, must be implemented within the context of the existing regulatory framework. The most important option in the United States has already been largely achieved, and that is to reduce the historical rate of peatland losses with their accompanying large oxidative losses of the stored soil carbon.

There has been strong interest expressed in using carbon sequestration as a rationale for wetland restoration and creation in the United States, Canada, and elsewhere (Wylynko, 1999; Watson *et al.*, 2000). However, high methane emissions from conterminous U.S. wetlands suggest that creating and restoring wetlands may increase net radiative forcing, although adequate data do not exist to fully evaluate this possibility. Roulet (2000) came to a similar conclusion concerning the restoration of Canadian wetlands. Net radiative forcing from restoration will likely vary among different kinds of wetlands and the specifics of their carbon budgets. The possibility of increasing radiative forcing by creating or restoring wetlands does not apply to estuarine wetlands, which emit relatively little methane compared to the carbon they sequester. Restoration of drained peatlands may stop the rapid loss of their soil carbon, which may compensate for increased methane emissions. However, Canadian peatlands restored from peat extraction operations increased their net emissions of carbon because of straw addition during the restoration process, although it was assumed that they would eventually become a net sink (Cleary *et al.*, 2005).

Regardless of their internal carbon balance, the area of restored wetlands is currently too small to form a significant carbon sink at the continental scale. Between 1986 and 1997, only 4,157 km^2 of uplands were converted into wetlands in the conterminous United States (Dahl, 2000). Using the soil carbon sequestration rate of 305 g C m^{-2} yr^{-1} found by Euliss *et al.* (2006) for restored prairie pothole

wetlands[2], we estimate that wetland restoration in the U.S. would have sequestered 1.3 Tg C over this 11-year period. However, larger areas of wetland restoration may have a significant impact on carbon sequestration. A simulation model of planting 20,000 km^2 into bottomland hardwood trees as part of the Wetland Reserve Program in the United States showed a sequestration of 4 Mt C yr^{-1} through 2045 (Barker *et al*., 1996). Euliss *et al.* (2006) estimated that if all cropland on former prairie pothole wetlands in the U.S. and Canada (162,244 km^2) were restored that 378 Tg C would be sequestered over 10 years in soils and plants. However, neither study accounted for the GWP of increased methane emissions.

Potentially more significant is the conversion of wetlands from one type to another; for example, 8.7% (37,200 km^2) of the wetlands in the conterminous United States in 1997 were in a previous wetland category in 1986 (Dahl, 2000). The net effect of these conversions on wetland carbon fluxes is unknown. Similarly, Roulet (2000) argued that too many uncertainties exist to include Canadian wetlands in the Kyoto Protocol.

In summary, North American wetlands form a very large carbon pool because of storage as peat and are a small-to-moderate carbon sink (excluding methane effects). The largest unknown in the wetland carbon budget is the amount and significance of sedimentation in FWMS wetlands. With the exception of estuarine wetlands, methane emissions from wetlands may largely offset any positive benefits of carbon sequestration in soils and plants. Given these conclusions, it is probably unwarranted to use carbon sequestration as a rationale for the protection and restoration of FWMS wetlands, although the many other ecosystem services that they provide justify these actions. However, protecting and restoring peatlands will stop the loss of their soil carbon (at least over the long term), and estuarine wetlands are an important carbon sink given their limited areal extent and low methane emissions.

The most important areas for further scientific research in terms of current carbon fluxes in the United States are to establish an unbiased, landscape-level sampling scheme to determine sediment carbon sequestration in FWMS and estuarine wetlands and to take additional measurements of annual methane emissions to better constrain these important fluxes. It would also be beneficial if the approximately decadal National Wetland Inventory (NWI) status and trends data were collected in sufficient detail with respect to the Cowardin *et al.* (1979) classification scheme to determine changes among mineral-soil wetlands and peatlands.

Canada lacks any regular inventory of its wetlands, and thus it is difficult to quantify land-use impacts upon their carbon fluxes and pools. While excellent scientific data exists on most aspects of carbon cycling in Canadian peatlands, Canadian FWMS and estuarine wetlands have been relatively poorly studied, despite having suffered large proportional losses to land-use change. Wetland data for Mexico is

[2]Euliss *et al.* (2006) regressed surface soil carbon stores in 27 restored semi-permanent prairie pothole wetlands against years since restoration to derive this estimate ($r^2 = 0.31$, $P = 0.002$). However, there was no significant relationship in seasonal prairie pothole wetlands ($r^2 = 0.04$, $P = 0.241$).

almost entirely lacking. Thus, anything that can be done to improve upon this would be helpful. All wetland inventories should consider the area of estuarine mud flats, which have the potential to sequester considerable carbon, and are poorly understood with respect to carbon sequestration.

The greatest unknown is how global change will affect the carbon pools and fluxes of North American wetlands. We will not be able to accurately predict the role of North American wetlands as potential positive or negative feedbacks to anthropogenic climate change without knowing the integrative effects of changes in temperature, precipitation, atmospheric carbon dioxide concentrations, and atmospheric deposition of nitrogen and sulfur within the context of internal ecosystem drivers of wetlands. To our knowledge, no manipulative experiment has simultaneously measured more than two of these perturbations in any North American wetland, and few have been done at any site. Modeling expertise of the carbon dynamics of wetlands has rapidly improved in the last few years (Frolking *et al*., 2002; Zhuang *et al*., 2004, and references therein), but this needs even further development in the future, including for FWMS and estuarine wetlands.

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Table 13-1. The area, carbon pool, net carbon balance, and methane flux from wetlands in North America and the world. Positive fluxes indicate net fluxes to the atmosphere, whereas negative fluxes indicate net fluxes into an ecosystem. Citations and assumptions in calculations are in the text and in Appendix 13A.

	Area[a] (km^2)		Carbon Pool[b] (Gt C)		Net Carbon Balance[c] (Mt C yr^{-1})		Historical Loss in Sequestration Capacity (Mt C yr^{-1})		Methane Flux (Mt CH_4 yr^{-1})	
Canada										
Peatland	1,135,608	****	149	****	-19	***	0.3	*	3.2	**
Freshwater Mineral	158,720	**	4.9	**	-5.1	*	6.5	*	5.7	*
Estuarine	6,400	***	0.1	***	-1.3	**	0.5	*	0.0	***
Total	1,300,728	****	154	****	-25	**	7.2	*	8.9	*
Alaska										
Peatland	132,196	****	15.9	**	-2.0	**	0.0	****	0.3	*
Freshwater Mineral	555,629	****	27.1	**	-18	*	0.0	****	1.4	*
Estuarine	8,400	****	0.1	***	-1.9	**	0.0	****	0.1	***
Total	696,224	*****	43.2	**	-22	*	0.0	****	1.8	*
Conterminous United States										
Peatland	93,477	****	14.4	***	4	*	2.1	*	3.4	**
Freshwater Mineral	312,193	*****	6.2	***	-18	*	15	*	11.2	**
Estuarine	23,000	*****	0.6	*****	-4.9	**	0.4	*	0.1	***
Total	428,670	*****	21.2	***	-19	*	17	*	14.7	**
U.S. Total	1,124,895	*****	64	**	-41	*	17	*	17	**
Mexico										
Peatland	10,000	*	1.5	*	-1.6	*	ND[d]	*	0.4	*
Freshwater Mineral	20,685	*	0.4	*	-0.7	*	ND	*	0.7	*
Estuarine	5,000	*	0.2	*	-1.6	*	0.5	*	0.0	*
Total	35,685	*	2.1	*	-3.9	*	ND	*	1.1	*

North America										
Peatland	1,371,281	****	180	****	-18	*	2.4	*	7	**
Freshwater Mineral	1,047,227	****	39	***	-42	*	21	*	19	*
Estuarine	42,800	***	1.0	***	-9.7	**	1.4	*	0.2	**
Total	2,461,308		220		-70	*	25	*	26	*
Global										
Peatland	3,443,000	***	460	***	150	**	16	*	37	**
Freshwater Mineral	2,315,000	***	46	***	-75	*	87	*	68	**
Estuarine	203,000	*	5.4	*	-43	*	13.2	*	1.5	**
Total	5,961,000	***	511	***	32	*	116	*	107	**

[a]Estuarine includes salt marsh, mangrove, and mudflat, except for Mexico and global for which no mudflat estimates were available.

[b]Includes soil C and plant C, but overall soil C is 98% of the total pool.

[c]Includes soil C sequestration, plant C sequestration, and loss of C due to drainage of wetlands. Plant C sequestration and soil oxidative flux due to drainage are either unknown or negligible for North American wetlands except for the conterminous United States (see Appendix 13A).

[d]No data.

The error categories are as follows:

***** = 95% certain that the actual value is within 10% of the estimate reported.
**** = 95% certain that the actual value is within 25%.
*** = 95% certain that the actual value is within 50%.
** = 95% certain that the actual value is within 100%.
* = uncertainty > 100%

Appendix 13A
Wetlands – Supplemental Material

INVENTORIES

Current Wetland Area and Rates of Loss

The ability to estimate soil carbon pools and fluxes in North American wetlands is constrained by the national inventories (or lack thereof) for Canada, the United States, and Mexico (Davidson *et al*., 1999). The National Wetland Inventory (NWI) program of the United States has repeatedly sampled several thousand wetland sites using aerial photographs and more limited field verification. The data are summarized in a series of reports detailing changes in wetland area in the conterminous United States for the periods of the mid-1950s to mid-1970s (Frayer *et al*., 1983), mid-1970s to mid-1980s (Dahl and Johnson, 1991), and 1986 to 1997 (Dahl, 2000). We used these relatively high-quality data sets extensively for estimating wetland area and loss rates in the conterminous United States, including mud flats. However, the usefulness of the NWI inventory reports for carbon budgeting is limited by the level of classification used to define wetland categories within the Cowardin *et al.* (1979) wetland classification system. At the level used in the national status and trend reports, vegetated freshwater wetlands are classified by dominant physiognomic vegetation type, and it is impossible to make the important distinction between wetlands with deep organic soils (i.e., peatlands) and wetlands with mineral soils. The data are not at an adequate spatial resolution to combine with U.S. Department of Agriculture (USDA) National Resources Conservation Service (NRCS) soil maps to discriminate between the two types of wetlands (T. Dahl, personal comm.). Because of these data limitations, we used the NRCS soil inventory of peatlands (i.e., Histosols and Histels, or peatlands with and without permafrost, respectively) to estimate historical peatland area (Bridgham *et al*., 2000) and combined these data with regional estimates of loss (Armentano and Menges, 1986) to estimate current peatland area in the conterminous United States. We calculated the current area of freshwater mineral-soil (FWMS) wetlands in the conterminous United States by subtracting peatland area from total wetland area (Dahl, 2000). This approach was limited by the Armentano and Menges peatland area data being current only up to the early 1980s, although large losses of peatlands since then are unlikely due to the institution of wetland protection laws.

We used a similar approach for Alaskan peatlands: peatland area was determined by the NRCS soil inventory [N. Bliss, query of the NRCS State Soil Geographic (STATSGO) database, February 2006] and overall wetland inventory was determined by standard NWI methods (Hall *et al*., 1994). However, our peatland estimate of 132,000 km^2 (Table 13A-1) is 22% of the often cited value by Kivinen and Pakarinen (1981) of 596,000 km^2.

Table 13A-1. Current and historical area of wetlands in North America and the world ($\times 10^3$ km^2).

Kivinen and Pakarinen also used NRCS soils data (Rieger *et al*., 1979) for their peatland estimates, but they defined a peatland as having a minimum organic layer thickness of 30 cm, whereas the current U.S. and Canadian soil taxonomies require a 40-cm thickness. The original 1979 Alaska soil inventory has been reclassified with current U.S. soil taxonomy (J. Moore, Alaska State Soil Scientist, personal comm.). Using the reclassified soil inventory, Alaska has 417,000 km^2 of wetlands with a histic modifier that are not Histosols or Histels, indicating significant carbon accumulation in the surface horizons of FWMS wetlands. Thus, we conclude that Kivinen and Pakarinen's Alaska peatland area estimate is higher because many Alaskan wetlands have a thin organic horizon that is not deep enough to qualify as a peatland under current soil taxonomy. Our smaller peatland area significantly lowers our estimate of carbon pools and fluxes in Alaskan peatlands compared to earlier studies (see *Carbon Pools* below).

The area of salt marsh in the conterminous U.S. and Alaska were taken from Alexander *et al*. (1986) and Hall (1994), respectively, as reported in Mendelssohn and McKee (2000). Because these estimates include brackish tidal marshes, they cannot be compared directly to the area of Canadian salt marsh. The historical area of tidal wetlands in the conterminous U.S. was based on the NWI (Dahl, 2000), but 'historical' here only refers to the 1950s as we could not find earlier estimates. It is almost certain that historical salt marsh area in the conterminous U.S. was larger than our estimate. We made the reasonable assumption that the historical area of Alaskan tidal wetlands was similar to the current area. The area of freshwater tidal marshes was not included.

A regular national inventory of Canada's wetlands has not been undertaken, although wetland area has been mapped by ecoregion (National Wetlands Working Group, 1988). Extensive recent effort has gone into mapping Canadian peatlands (Tarnocai, 1998; Tarnocai *et al*., 2005). We calculated the current area of mineral-soil wetlands as the difference between total wetland area and peatland area in National Wetland Working Group (1988). Historical FWMS wetland area was obtained from Rubec (1996). Canadian salt marsh estimates were taken from a compilation by Mendelssohn and McKee (2000). The compilation does not include brackish or freshwater tidal marshes, and we were unable to locate other estimates of Canadian brackish marsh area. The historical area of these marshes was estimated from the National Wetland Working Group (1988), but it is highly uncertain. There are no reliable country-wide estimates of mud flat area for Canada, but a highly uncertain extrapolation from a limited number of regional estimates was possible.

No national wetland inventories have been done for Mexico. Current freshwater wetland estimates for Mexico were taken from Davidson *et al*. (1999) and Spiers (1999), who used inventories of discrete

wetland regions performed by a variety of organizations. Thus, freshwater wetland area estimates for Mexico are highly unreliable and are possibly a large underestimate. For mangrove area in Mexico, we used the estimates compiled by Mendelssohn and McKee (2000), which are similar to estimates reported in Davidson *et al.* (1999) and Spalding *et al.* (1997). We could find no estimates of tidal marsh or mud flat area for Mexico. Since most vegetated Mexican tidal wetlands are dominated by mangroves (Olmsted, 1993; Mendelssohn and McKee, 2000), the omission of Mexican tidal marshes should not significantly affect our carbon budget. However, there may be large areas of mud flat that would significantly increase our estimate of carbon pools and sequestration in this country. We arbitrarily estimated that 25% of the mangrove area was lost since the late 1800s, which is less than the rough worldwide estimate of 50% wetland loss that is often cited (see Zedler and Kercher, 2005). A lower estimate is reasonable because wetland losses are lower in coastal systems than freshwater systems (Zedler and Kercher, 2005).

CARBON POOLS

Freshwater Mineral-Soil (Gleysol) Carbon Pools

Gleysol is a soil classification used by the Food and Agriculture Organization (FAO) and many countries that denotes mineral soils formed under waterlogged conditions (FAO-UNESCO, 1974). Tarnocai (1998) reported a soil carbon density of 200 Mg C ha^{-1} for Canadian Gleysols but did not indicate to what depth this extended. Batjes (1996) determined soil carbon content globally from the *Soil Map of the World* (FAO, 1991) and a large database of soil pedons. He gave a very similar average value for soil carbon density of 199 Mg C ha^{-1} (CV^3 = 212%, n = 14 pedons) for Gleysols of the world to 2-m depth; to 1-m depth, he reported a soil carbon density of 131 Mg C ha^{-1} (CV = 109%, n =142 pedons).

Gleysols are not part of the U.S. soil taxonomy scheme, and mineral soils with attributes reflecting waterlogged conditions are distributed among numerous soil groups. We used the NRCS State Soil Geographic (STATSGO) soils database to query for soil carbon density in "wet" mineral soils of the conterminous United States (all soils that had a surface texture described as peat, muck, or mucky peat, or appeared on the 1993 list of hydric soils, which were not classified as Histosols) (N. Bliss, query of NRCS STATSGO database, Dec. 2005). We used the average soil carbon densities of 162 Mg C ha^{-1} from this query for FWMS wetlands in the conterminous United States and Mexico.

Some caution is necessary regarding the use of Gleysol or 'wet' mineral soil carbon densities because apparently they include large areas of seasonally wet soils that are not considered wetlands by the more conservative definition of wetlands used by the United States and many other countries and organizations.

[3]CV is the "coefficient of variation," or 100 times the standard deviation divided by the mean.

For example, Eswaran *et al.* (1995) estimated that global wet mineral-soil area was 8,808,000 km^2, which is substantially higher than the commonly accepted mineral-soil wetland area estimated by Matthews and Fung (1987) of 2,289,000 km^2 and Aselmann and Crutzen (1989) of 2,341,000 km^2, even accounting for substantial global wetland loss. In our query of the NRCS STATSGO database for the United States, we found 1,258,000 km^2 of wet soils in the conterminous United States versus our estimate of 312,000 km^2 of FWMS wetlands currently and 762,000 km^2 historically (Table 13A-1). We assume that including these wet-but-not-wetland soils will decrease the estimated soil carbon density, but to what degree we do not know. However, just considering the differences in area will give large differences in the soil carbon pool. For example, Eswaran *et al.* (1995) estimated that wet mineral soils globally contain 108 Gt C to 1-m depth, whereas our estimate is 46 Gt C to 2-m depth (Table 13A-2).

For Alaska, many soil investigations have been conducted since the STATSGO soil data was coded. We updated STATSGO by calculating soil carbon densities from data obtained from the NRCS on 479 pedons collected in Alaska, and then we used this data for both FWMS wetlands and peatlands. For some of the Histosols, missing bulk densities were calculated using averages of measured bulk densities for the closest matching class in the USDA Soil Taxonomy (NRCS, 1999). A matching procedure was developed for relating sets of pedons to sets of STATSGO components. If there were multiple components for each map unit in STATSGO, the percentage of the component was used to scale area and carbon data. We compared matching sets of pedons to sets of components at the four top levels of the U.S. Soil Taxonomy: Orders, Suborders, Great Groups, and Subgroups. For example, the soil carbon for all pedons having the same soil order were averaged, and the carbon content was applied to all of the soil components of the same order (e.g., Histosol pedons are used to characterize Histosol components). At the Order level, all components were matched with pedon data. At the suborder level, pedon data were not available to match approximately 20,000 km^2 (compared to the nearly 1,500,000-km^2 area of soil in the state), but the soil characteristics were more closely associated with the appropriate land areas than at the Order level. At the Great Group and Subgroup levels, pedon data were unavailable for much larger areas, even though the quality of the data when available became better. For this study, we used the Suborder-level matching. The resulting soil carbon density for Alaskan FWMS wetlands was 469 Mg C ha^{-1}, reflecting large areas of wetlands with a histic epipedon as noted above.

Peatland Soil Carbon Pools

The carbon pool of permafrost and non-permafrost peatlands in Canada had been previously estimated by Tarnocai *et al.* (2005) based upon an extensive database. Good soil-carbon density data are unavailable for peatlands in the United States, as the NRCS soil pedon information typically only goes to a maximum depth of between 1.5 to 2 m, and many peatlands are deeper than this. Therefore, we used the

carbon density estimates of Tarnocai *et al.* (2005) of 1,441 Mg C ha^{-1} for Histosols and 1,048 Mg C ha^{-1} for Histels to estimate the soil carbon pool in Alaskan peatlands.

The importance of our using a smaller area of Alaskan peatlands becomes obvious here. Using the larger area from Kivinen and Pakarinen (1981), Halsey *et al.* (2000) estimated that Alaskan peatlands have a soil carbon pool of 71.5 Gt, almost 5-fold higher than our estimate. However, some of the difference in soil carbon between the two estimates can be accounted for by the 26 Gt C that we calculated resides in Alaskan FWMS wetlands (Table 13A-2).

Table 13A-2. Soil carbon pools (Gt) and fluxes (Mt yr^{-1}) of wetlands in North America and the world.

The peatlands of the conterminous United States are different in texture, and probably depth, from those in Canada and Alaska, so it is probably inappropriate to use the soil carbon densities for Canadian peatlands for those in the conterminous United States. For example, we compared the relative percentage of the Histosol suborders (excluding the small area of Folists, as they are predominantly upland soils) for Canada (Tarnocai, 1998), Alaska (updated STATSGO data, J. Moore, personal comm.), and the conterminous U.S. (NRCS, 1999). The relative percentage of Fibrists, Hemists, and Saprists, respectively, in Canada are 37%, 62%, and 1%, in Alaska are 53%, 27%, and 20%, and in the conterminous United States are 1%, 19%, and 80%. Using the STATSGO database (N. Bliss, query of NRCS STATSGO database, December 2005), the average soil carbon density for Histosols in the conterminous United States is 1,089 Mg C ha^{-1}, but this is an underestimate as many peatlands were not sampled to their maximum depth. Armentano and Menges (1986) reported average carbon density of conterminous U.S. peatlands to 1-m depth of 1,147 to 1,125 Mg C ha^{-1}. Malterer (1996) gave soil carbon densities of conterminous U.S. peatlands of 2,902 Mg C ha^{-1} for Fibrist, 1,874 Mg C ha^{-1} for Hemists, and 2,740 Mg C ha^{-1} for Saprists, but it is unclear how he derived these estimates. Batjes (1996) and Eswaran *et al.* (1995) gave average soil carbon densities to 1-m depth for global peatlands of 776 and 2,235 Mg C ha^{-1}, respectively. We chose to use an average carbon density of 1,500 Mg C ha^{-1}, which is in the middle of the reported range.

Estuarine Soil Carbon Pools

Tidal wetland soil carbon density was based on a country-specific analysis of data reported in an extensive compilation by Chmura *et al.* (2003). There were more observations for the United States (n = 75) than Canada (n = 34) or Mexico (n = 4), and consequently there were more observations of marshes than mangroves. The Canadian salt marsh estimate was used for Alaskan salt marshes and mud flats. In the conterminous United States and Mexico, country-specific marsh or mangrove estimates were

used for mudflats. Although Chmura *et al.* (2003) reported some significant correlations between soil carbon density and mean annual temperature, scatter plots suggest the relationships are weak or driven by a few sites. Thus, we did not separate the data by region or latitude and used mean values for scaling. Chmura *et al.* (2003) assumed a 50-cm-deep profile for the soil carbon pool, which may be an underestimate.

Plant Carbon Pools

While extensive data on plant biomass in individual wetlands have been published, no systematic inventory of wetland plant biomass has been undertaken in North America. Nationally, the forest carbon biomass pool (including aboveground and belowground biomass) has been estimated to be 5.49 kg C m^{-2} (Birdsey, 1992), which we used for forested wetlands in the United States and Canada. This approach assumes that wetland forests do not have substantially different biomass carbon densities from upland forests. There is one regional assessment of forested wetlands in the southeastern United States, which comprise approximately 35% of the total forested wetland area in the conterminous United States. We utilized the southeastern U.S. regional inventory to evaluate this assumption; aboveground tree biomass averaged 125.2 m^3 ha^{-1} for softwood stands and 116.1 m^3 ha^{-1} for hardwood stands. Using an average wood density and carbon content, the carbon density for these forests would be 3.3 kg C m^{-2} for softwood stands and 4.2 kg C m^{-2} for hardwood stands. However, these estimates do not include understory vegetation, belowground biomass, or dead trees, which account for 49% of the total forest biomass (Birdsey, 1992). Using that factor to make an adjustment for total forest biomass, the range would be 4.9 to 6.6 kg C m^{-2} for the softwood and hardwood stands, respectively. Accordingly, the assumption of using 5.49 kg C m^{-2} seems reasonable for a national-level estimate.

The area of forested wetlands in Canada came from Tarnocai *et al.* (2005), for Alaska from Hall *et al.* (1994), and for the conterminous United States from Dahl (2000).

Since Tarnocai *et al.* (2005) divided Canadian peatland area into bog and fen, we used aboveground biomass for each community type from Vitt *et al.* (2000), and assumed that 50% of biomass is belowground. We used the average bog and fen plant biomass from Vitt *et al.* (2000) for Alaskan peatlands. For other wetland areas, we used an average value of 2,000 g C m^{-2} for non-forested wetland biomass carbon density (Gorham, 1991).

Tidal marsh root and shoot biomass data were estimated from a compilation in Table 8-7 in Mitsch and Gosselink (1993). There was no clear latitudinal or regional pattern in biomass, so we used mean values for each. Mangrove biomass has been shown to vary with latitude, so we used the empirical relationship from Twilley *et al*. (1992), for this relationship. We made a simple estimate using a single latitude that visually bisected the distribution of mangroves either in the United States (26.9^o) or Mexico

(23.5°). Total biomass was estimated using a root-to-shoot ratio of 0.82 and a carbon-mass-to-biomass ratio of 0.45, both from Twilley *et al.* (1992).

Plant biomass carbon data are presented in Table 13A-3.

Table 13A-3. Plant carbon pools (Gt) and fluxes (Mt yr^{-1}) of wetlands in North America and the world.

CARBON FLUXES

Peatland Soil Carbon Accumulation Rates

Most studies report the long-term apparent rate of carbon accumulation (LORCA) in peatlands based upon basal peat dates, but this assumes a linear accumulation rate through time. However, due to the slow decay of the accumulated peat, the true rate of carbon accumulation will always be less than the LORCA (Clymo *et al*., 1998), so most reported rates are inherently biased upwards. Tolonen and Turunen (1996) found that the true rate of peat accumulation was about 67% of the LORCA.

For estimates of soil carbon sequestration in conterminous U.S. peatlands, we used the data from 82 sites and 215 cores throughout eastern North America (Webb and Webb III, 1988). They reported a median accumulation rate of 0.066 cm yr^{-1} (mean = 0.092, sd = 0.085). We converted this value into a carbon accumulation rate of -1.2 Mg C ha^{-1} yr^{-1} by assuming 58% C (see NRCS Soil Survey Laboratory Information Manual, available on-line at http://soils.usda.gov/survey/nscd/lim/), a bulk density of 0.59 g cm^{-3}, and an organic matter content of 55%. **(Positive carbon fluxes indicate net fluxes to the atmosphere, whereas negative carbon fluxes indicate net fluxes into an ecosystem**.) The bulk density and organic matter content were the average from all Histosol soil map units greater than 202.5 ha (n = 5,483) in the conterminous United States from the National Soil Information System (NASIS) data base provided by S. Campbell (USDA NRCS, Portland, OR). For comparison, Armentano and Menges (1986) used soil carbon accumulation rates that ranged from -0.48 Mg C ha^{-1} yr^{-1} in northern conterminous U.S. peatlands to -2.25 Mg C ha^{-1} yr^{-1} in Florida peatlands.

Peatlands accumulate lesser amounts of soil carbon at higher latitudes, with especially low accumulation rates in permafrost peatlands (Ovenden, 1990; Robinson and Moore, 1999). The rates used in this report reflect this gradient, going from -0.13 to -0.19 to -1.2 Mg C ha^{-1} yr^{-1} in permafrost peatlands, non-permafrost Canadian and Alaskan peatlands, and peatlands in the conterminous United States and Mexico, respectively (Table 13A-2).

Freshwater Mineral-Soil Wetland Carbon Accumulation Rates

Many studies have estimated sediment deposition rates in FWMS wetlands, with an average rate of 1,680 g m^{-2} yr^{-1} (range 0 to 7,840) in a review by Johnston (1991). Assuming 7.7% carbon for FWMS wetlands (Batjes, 1996), this gives a substantial accumulation rate of -129 g C m^{-2} yr^{-1}. Johnston (1991) found many more studies that just reported vertical sediment accumulation rates, with an average of 0.69 cm yr^{-1} (range -0.6 to 2.6). If we assume a bulk density of 1.38 g cm^{-3} for FWMS wetlands (Batjes, 1996), this converts into an impressive accumulation rate of -733 g C m^{-2} yr^{-1}. For reasons discussed in the main chapter, we assumed a lower carbon sequestration rate in FWMS wetlands of -34 g C m^{-2} yr^{-1}.

Agriculture typically increases sedimentation rates by 10- to 100-fold, and 90% of sediments are stored within the watershed, or about 3 Gt yr^{-1} in the United States (Meade *et al*., 1990, as cited in Stallard, 1998), as cited in Stallard, 1998). Converting this to 1.5% C equates to -45 Mt C yr^{-1}, part of which will be stored in wetlands and is well within our estimated storage rate in FWMS wetlands (Table 13A-2).

Estuarine Carbon Accumulation Rates

Carbon accumulation in tidal wetlands was assumed to be entirely in the soil pool. This should provide a reasonable estimate because marshes are primarily herbaceous, and mangrove biomass should be in steady state unless the site was converted to another use. An important difference between soil carbon sequestration in tidal and non-tidal systems is that tidal sequestration occurs primarily through burial driven by sea level rise. For this reason, carbon accumulation rates can be estimated well with data on changes in soil surface elevation and carbon density. Rates of soil carbon accumulation were calculated from Chmura *et al.* (2003) as described for the soil carbon pool (above). These estimates are based on a variety of methods, such as ^{210}Pb dating and soil elevation tables, which integrate vertical soil accumulation rates over periods of time ranging from 1–100 yr. The soil carbon sequestered in estuarine wetland sediments is likely to be a mixture of both allochthonous and autochthonous sources. However, without better information, we assumed that in situ rates of soil carbon sequestration in estuarine wetlands is representative of the true landscape-level rate.

Extractive Uses of Peat

Use of peat for energy production is, and always has been, negligible in North America, as opposed to other parts of the world (WEC, 2001). However, Canada produces a greater volume of horticultural and agricultural peat than any other country in the world (WEC, 2001). Currently, 124 km^2 of Canadian peatlands have been under extraction now or in the past (Cleary *et al*., 2005). A life-cycle analysis by these authors estimated that as of 1990 Canada emitted 0.9 Mt yr^{-1} of CO_2-C equivalents through peat

extraction. The U.S. production of horticultural peat is about 19% of Canada's (Joosten and Clarke, 2002), which assuming a similar life-cycle as for Canada, suggests that the United States produces 0.2 Mt of CO_2-C equivalents through peat extraction.

Methane Fluxes

Moore *et al.* (1995) reported a range of methane fluxes from 0 to 130 g CH_4 m^{-2} yr^{-1} from 120 peatland sites in Canada, with the majority <10 g CH_4 m^{-2} yr^{-1}. They estimated a low average flux rate of 2 to 3 g CH_4 m^{-2} yr^{-1}, which equaled an emission of 2–3 Mt CH_4 yr^{-1} from Canadian peatlands. We used an estimate of 2.5 g CH_4 m^{-2} yr^{-1} for Canadian peatlands and Alaskan freshwater wetlands (Table 13A-4).

Table 13A-4. Methane fluxes (Mt yr^{-1}) from wetlands in North America and the world.

To our knowledge, the last synthesis of field measurements of methane emissions from wetlands was done by Bartlett and Harriss (1993). We supplemented their analysis with all other published field studies (using chamber or eddy covariance techniques) we could find that reported annual or average daily methane fluxes in the conterminous United States (Table 13A-5). We excluded a few studies that used cores or estimated diffusive fluxes.

Table 13A-5. Methane fluxes measured in the conterminous United States.

In cases where multiple years from the same site were presented, we took the average of those years. Similarly, when multiple sites of the same type were presented in the same paper, we took the average. Studies were separated into freshwater and estuarine systems.

In cases where papers presented both an annual flux and a mean daily flux, we calculated a conversion factor [annual flux/(average daily flux × 10^3)] to quantify the relationship between those two numbers (Table 13A-5). When we looked at all studies (n = 30), this conversion factor was 0.36, suggesting that there is a 360-day emission season. There was surprisingly little variation in this ratio, and it was similar in freshwater (0.36) and estuarine (0.34) wetlands. In contrast, previous syntheses used a 150-day emission season for temperate wetlands (Matthews and Fung, 1987; Bartlett and Harriss, 1993). While substantial winter methane emissions have been found in some studies, it is likely that flux data from most studies have a non-normal distribution with occasional periods of high flux rates that are better captured with annual measurements.

Using the conversion factors for freshwater and estuarine wetlands, we estimated average annual fluxes from the average daily fluxes. For freshwater wetlands, the calculated average annual flux rate was

38.6 g CH_4 m^{-2} yr^{-1} (n = 74), which is slightly larger than the average actual measured flux rate of 32.1 g CH_4 m^{-2} yr^{-1} (n = 32). For estuarine wetlands, the average calculated annual flux rate was 9.8 g CH_4 m^{-2} yr^{-1} (n = 25), which is smaller than the average measured flux rate of 16.9 g CH_4 m^{-2} yr^{-1} (n = 13). However, if we remove one outlier, the average measured flux rate is 10.2 g CH_4 m^{-2} yr^{-1}.

Finally, we combined both approaches. In cases where a paper presented an annual value, we used that number. In cases where only an average daily number was presented, we used that value corrected with the appropriate conversion factor. For conterminous U.S. wetlands, FWMS Canadian wetlands, and Mexican wetlands, we used an average flux of 36 g CH_4 m^{-2} $yr^{-1,}$, and for estuarine wetlands, we used an average flux of 10.3 g CH_4 m^{-2} yr^{-1}.

Plant Carbon Fluxes

For ecosystems at approximately steady state, plant biomass should be reasonably constant on average because plant production is roughly balanced by mortality and subsequent decomposition. We assumed insignificant plant biomass accumulation in freshwater and estuarine marshes because they are dominated by herbaceous plants that do not accumulate carbon in wood. Sequestration in plants in relatively undisturbed forested wetlands in Alaska and many parts of Canada is probably small, although there may be substantial logging of Canadian forested wetlands for which we do not have data. Similarly, no data was available to evaluate the effect of harvesting of woody biomass in Mexican mangroves on carbon fluxes.

Tree biomass carbon sequestration averages -140 g C m^2 yr^{-1} in U.S. forests across all forest types (Birdsey, 1992). Using the tree growth estimates from the southeastern U.S. regional assessment of wetland forests (Brown *et al*., 2001) yields an even lower estimate of sequestration in aboveground tree biomass (approx. -50.2 g C m^2 yr^{-1}). We used this lower value and area estimates from Dahl (2000) to estimate that forested wetlands in the conterminous U.S. currently sequester -10.3 Mt C yr^{-1}.

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Table 13A-1. Current and historical area of wetlands in North America and the world ($\times 10^3$ km^2). Historical refers to approximately 1800, unless otherwise specified.

	Permafrost peatlands	Non-permafrost peatlands	Mineral-soil freshwater	Salt marsh	Mangrove	Mudflat	**Total**
Canada							
Current	422[a]	714[a]	159[b]	0.4[c]	0	6[d]	1301
Historical	424[e]	726[f]	359[g]	1.3[b]	0	7[h]	1517
Alaska							
Current	89[i]	43[i]	556[j]	1.4[c]	0	7[k]	696
Historical	89	43	556	1.4	0	7	696
Conterminous United States							
Current	0	93[l]	312[m]	18[c]	3[c]	2[n]	428
Historical	0	111[i]	762[o]	20[p]	4[n]	3[n]	899
Mexico							
Current	0	10[p]	21[p]	0	5[c]	ND[q]	36
Historical	0		45[p]	0	7[h]	ND	52
North America							
Current	511	861	1,047	20	8	15	2,461
Historical	513	894[r]	1,706[r]	23	11	17	3,164
Global							
Current	3,443[s]		2,315[t]	22[u]	181[v]	ND	~6,000
Historical	4,000[w]		5,000[x]	26[y]	ND	ND	~9,000[x]

[a]Tarnocai *et al.* (2005).

[b]National Wetlands Working Group (National Wetlands Working Group, 1988).

[c]Mendelssohn and McKee (2000).

[d]Estimated from the area of Canadian salt marshes and the ratio of mudflat to salt marsh area reported by Hanson and Calkins (1996).

[e]Accounting for losses due to permafrost melting in western Canada (Vitt *et al*., 1994). This is an underestimate, as similar, but undocumented, losses have probably also occurred in eastern Canada and Alaska.

[f]9000 km^2 lost to reservoir flooding (Rubec, 1996), 250 km^2 to forestry drainage (Rubec, 1996), 124 km^2 to peat harvesting for horticulture (Cleary *et al*., 2005), and 16 km^2 to oil sands mining (Turetsky *et al*., 2002). See note e for permafrost melting estimate.

[g]Rubec (1996).

[h]Assumed same loss rate as the conterminous United States since 1954 (Dahl, 2000).

[i]Historical area from NRCS soil inventory (Bridgham *et al.*, 2000), except Alaska inventory updated by N. Bliss from a February 2006 query of the STATSGO database. Less than 1% wetland losses have occurred in Alaska (Dahl, 1990).

[j]Total freshwater wetland area from (Hall *et al*., 1994) minus peatland area.

[k]Hall (1994).

[l]Historical area from Bridgham *et al.* (2000) minus losses in Armentano and Menges (1986).

[m]Overall freshwater wetland area from Dahl (2000) minus peatland area.

[n]Dahl (2000). Historical area estimates are only from the 1950s.

[o]Total historical wetland area from Dahl (1990) minus historical peatland area minus historical estuarine area.

[p]Spiers (1999).

[q]ND indicates that no data are available.

[r]Assuming that historical proportion of peatlands to total wetlands in Mexico was the same as today.

[s]Bridgham *et al.* (2000) for the United States, Tarnocai *et al.* (2005) for Canada, Joosten and Clarke (2002) for the rest of world. Recent range in literature 2,974,000–3,985,000 km^2 (Matthews and Fung, 1987; Aselmann and Crutzen, 1989; Maltby and Immirzi, 1993; Bridgham *et al.*, 2000; Joosten and Clarke, 2002).

[t]Average of 2,289,000 km^2 from Matthews and Fung (1987) and 2,341,000 km^2 Aselmann and Crutzen (1989).

[u]Chmura *et al.* (2003). Underestimated because no inventories were available for the continents Asia, South America and Australia which are mangrove-dominated but also support salt marsh.

[v]Spalding (1997).

[w]Range from 3,880 to 4,086 in Maltby and Immirzi (1993).

[x]Approximately 50% loss from Moser *et al.* (1996).

[y]Assumed.

Table 13A-2. Soil carbon pools (Gt) and fluxes (Mt yr^{-1}) of wetlands in North America and the world. "Sequestration in current wetlands" refers to carbon sequestration in extant wetlands; "oxidation in former wetlands" refers to emissions from wetlands that have been converted to non-wetland uses or conversion among wetland types due to human influence; "historical loss in sequestration capacity" refers to the loss in the carbon sequestration function of wetlands that have been converted to non-wetland uses; "change in flux from wetland conversions" is the sum of the two previous fluxes. Positive flux numbers indicate a net flux into the atmosphere, whereas negative numbers indicate a net flux into the ecosystem**.**

	Permafrost peatlands	Non-permafrost peatlands	Mineral-soil freshwater	Salt marsh	Mangrove	Mudflat	**Total**
Canada							
Pool Size in Current Wetlands	44.2[a]	102.9[a]	4.6[b]	0.0[c]	0.0	0.1[d]	151.8
Sequestration in Current Wetlands	-5.5[e]	-13.6[e]	-5.1[f]	-0.1	0.0	-1.2[d]	-25.5
Oxidation in Former Wetlands	0.2[g]		0.0[h]	0.0[i]	0.0	0.0	0.2
Historical Loss in Sequestration Capacity	0.0[e]	0.2[e]	6.5[f]	0.2	0.0	0.3	7.2
Change in Flux From Wetland Conversions	0.4		6.5	0.2	0.0	0.3	7.4
Alaska							
Pool Size in Current Wetlands	9.3[j]	6.2[j]	26.0[k]	0.0	0.0	0.1	41.7
Sequestration in Current Wetlands	-1.1[e]	-0.8[e]	-18.0[f]	-0.3	0.0	-1.6	-21.9
Oxidation in Former Wetlands	0.0	0.0	0.0	0.0	0.0	0.0	0.0
Historical Loss in Sequestration Capacity	0.0	0.0	0.0	0.0	0.0	0.0	0.0
Change in Flux From Wetland Conversions	0.0	0.0	0.0	0.0	0.0	0.0	0.0
Conterminous United States							
Pool Size in Current Wetlands	0	14.0[l]	5.1[k]	0.4	0.1	0.1	19.7
Sequestration in Current Wetlands	0	-11.6[m]	-10.1[f]	-3.9	-0.5	-0.5	-26.6
Oxidation in Former Wetlands	0	18.0[n]	0.0[h]	0.0	0.0	0.0	18.0
Historical Loss in Sequestration Capacity	0	2.1[m]	14.5[f]	0.3	0.0	0.1	17.1
Change in Flux from Wetland Conversions	0	20.1	14.6	0.3	0.0	0.1	35.2
Mexico							
Pool Size in Current Wetlands	0.0	1.5[l]	0.3[k]	0.0	0.1	ND*	1.9
Sequestration in Current Wetlands	0	-1.6[o]	-0.7[f]	0.0	-1.6	ND	-3.9

Oxidation in Former Wetlands	0	ND	ND	0.0	0.0	0.0	ND
Historical Loss in Sequestration Capacity	0	ND	ND	0.0	0.5	ND	0.5
Change in Flux from Wetland Conversions	0	ND	ND	0.0	0.5	ND	0.5
North America							
Pool Size in Current Wetlands	53.5	124.6	36.0	0.4	0.2	0.3	215.1
Sequestration in Current Wetlands	-6.6	-27.6	-33.9	-4.3	-2.1	-3.3	-77.8
Oxidation in Former Wetlands	18.2		0.0	0.0	0.0	0.0	18.2
Historical Loss in Sequestration Capacity	0	2.3	21.0	0.5	0.5	0.5	24.8
Change in Flux from Wetland Conversions	20.5		21.1	0.5	0.5	0.5	43.1
Global							
Pool Size in Current Wetlands	462[p]		46[q]	0.4[r]	5.0[r]	ND	513
Sequestration in Current Wetlands	-55[s]		-75[f]	-4.6[r]	-38.0[r]	ND	-173
Oxidation in Former Wetlands	205[t]		ND	0	0	0	205
Historical Loss in Sequestration Capacity	16[t]		87[f]	0.8[u]	12.7[v]	ND	116
Change in Flux From Wetland Conversions	221[t]		> 87[w]	0.8	12.7	ND	321

*ND indicates that no data are available.

[a]Tarnocai *et al.* (2005).

[b]Tarnocai (1998).

[c]Rates calculated from Chimura *et al.* (2003); areas from Mendelssohn and McKee (2000).

[d]Assumed the same carbon density and accumulation rates as the adjacent vegetated wetland ecosystem (mangrove data for Mexico and salt marsh data elsewhere).

[e]Assumed carbon accumulation rate of 0.13 Mg C ha^{-1} yr^{-1} for permafrost peatlands and 0.19 Mg C ha^{-1} yr^{-1} non-permafrost peatlands. Reported range of long-term apparent accumulation rates from 0.05-0.35 (Ovenden, 1990; Maltby and Immirzi, 1993; Trumbore and Harden, 1997; Vitt *et al*., 2000; Turunen *et al*., 2004).

[f]Potential rate calculated as the average sediment accumulation rate of 1680 g m^{-2} yr^{-1} (range 0–7840) from Johnston (1991) times 7.7% C (CV = 109) (Batjes, 1996). We assumed that all sequestered soil C was of allochthonous origin and decomposition was 25% slower in wetlands than in the uplands from which the sediment was eroded (see text).

[g]Sum of -0.24 Mt C yr^{-1} from horticulture removal of peat (Cleary *et al.*, 2005) and 0.10 Mt C yr^{-1} from increased peat sequestration due to permafrost melting (Turetsky *et al.*, 2002).

[h]Assumed that the net oxidation of 8.6% of the soil carbon pool (Euliss *et al*., 2006) over 50 yr after conversion to non-wetland use.

[i]Assumed that conversion of tidal systems is caused by fill and results in burial and preservation of SOM define SOM rather than oxidation.

[j]Soil carbon densities of 1,441 Mg C ha^{-1} for Histosols and 1,048 Mg C ha^{-1} for Histels (Tarnocai *et al.*, 2005).

[k]Soil carbon density of 162 Mg C ha^{-1} for the conterminous United States and Mexico and 468 Mg C ha^{-1} for Alaska based upon NRCS STATSGO database and soil pedon information.

[l]Assumed soil carbon density of 1,500 Mg C ha^{-1}.

[m]Webb and Webb (1988).

[n]Estimated loss rate as of early 1980s (Armentano and Menges,1986). Overall wetlands losses in the United States have declined dramatically since then (Dahl, 2000) and probably even more so for Histosols, so this number may still be representative.

[o]Using peat accumulation rate of 1.6 Mg C ha^{-1} (range 1.0–2.25) (Maltby and Immirzi, 1993).

[p]From Maltby and Immirzi (1993). Range of 234 to 679 Gt C (Gorham, 1991; Maltby and Immirzi, 1993; Eswaran *et al*., 1995; Batjes, 1996; Lappalainen, 1996; Joosten and Clarke, 2002).

[q]Soil carbon density of 199 Mg C ha^{-1} (Batjes, 1996).

[r]Chmura *et al.* (2003).

[s]Joosten and Clarke (2002) reported range of -40 to -70 Mt C yr^{-1} . Using the peatland estimate in Table 13A-1 and a C accumulation rate of 0.19 Mg C ha^{-1} yr^{-1}, we calculate a global flux of -65 Mt C yr^{-1} in peatlands.

[t]Current oxidative flux is the difference between the change in flux and the historical loss in sequestration capacity from this table. The change in flux is from Maltby and Immirzi (1993) (reported range 176 to 266 Mt C yr^{-1}) and the historical loss in sequestration capacity is from this table for North America, from Armentano and Menges (1986) for other northern peatlands, and from Maltby and Immirzi (1993) for tropical peatlands.

[u]Assumed that global rates approximate the North America rate because most salt marshes inventoried are in North America.

[v]Assumed 25% loss globally since the late 1800s.

[w]> sign indicates that this a minimal loss estimate.

Table 13A-3. Plant carbon pools (Gt) and fluxes (Mt yr^{-1}) of wetlands in North America and the world. Positive flux numbers indicate a net flux into the atmosphere, whereas negative numbers indicate a net flux into the ecosystem.

	Permafrost peatlands	Non-permafrost peatlands	Mineral-soil freshwater	Salt marsh	Mangrove	**Total**
Canada						
Pool Size in Current Wetlands	1.4[a]		0.3[b]	0.0[c]	0.0	1.7
Sequestration in Current Wetlands	0.0	ND*		0.0	0.0	0.0
Alaska						
Pool Size in Current Wetlands	0.4[a]		1.1[d]	0.0	0.0	1.5
Sequestration in Current Wetlands	0.0	0.0	0.0	0.0	0.0	0.0
Conterminous United States						
Pool Size in Current Wetlands	0.0	1.5[d]		0.0	0.0	1.5
Sequestration in Current Wetlands	0.0	-10.3[e]		0.0	0.0	-10.3
Mexico						
Pool Size in Current Wetlands	0.0	0.0[b]	0.0[b]	0.0	0.1	0.1
Sequestration in Current Wetlands	0.0	ND	ND	0.0	ND	0.0
North America						
Pool Size in Current Wetlands	4.8			0.0	0.1	4.9
Sequestration in Current Wetlands	0.0	-10.3		0.0	ND	-10.3
Global						
Pool Size in Current Wetlands	6.9[b]		4.6[b]	0.0[f]	4.0[g]	15.5
Sequestration in Current Wetlands	0.0	ND	ND	0.0	ND	ND

*ND indicates that no data are available.

[a]Biomass for non-forested peatlands from Vitt *et al.* (2000), assuming 50% of biomass is belowground. Forest biomass density from Birdsey (1992) and forested area from Tarnocai *et al.* (2005) for Canada and from Hall *et al.* (1994) for Alaska.

[b]Assumed 2000 g C m^{-2} in aboveground and belowground plant biomass (Gorham, 1991).

[c]Biomass data from Mitsch and Gosselink (1993).

[d]Biomass for non-forested wetlands from Gorham (1991). Forest biomass density from Birdsey (1992), and forested area from Hall *et al.* (1994) for Alaska and Dahl (2000) for the conterminous U.S..

[e]50 g C m^{-2} yr^{-1} sequestration from forest growth from a southeastern U.S. regional assessment of wetland forest growth (Brown *et al*., 2001).
[f]Assumed that global pools approximate those from North America because most salt marshes inventoried are in North America.
[g]Twilley *et al.* (1992).

Table 13A-4. Methane fluxes (Mt yr^{-1}) from wetlands in North America and the world

	Permafrost peatlands	Non-permafrost peatlands	Mineral-soil freshwater	Salt marsh	Mangrove	Mudflat	**Total**
Canada							
CH_4 Flux in Current Wetlands	1.1[a]	2.1[b]	5.7	0.0	0.0	0.0[c]	8.9
Historical change in CH_4 Flux	0.0	0.3	-7.2	0.0	0.0	0.0	-6.9
Alaska							
CH_4 Flux in Current Wetlands	0.2	0.1	1.4	0.0	0.0	0.1	1.8
Historical change in CH_4 Flux	0.0	0.0	0.0	0.0	0.0	0.0	0.0
Conterminous United States							
CH_4 Flux in Current Wetlands	0.0	3.4	11.2	0.1	0.0	0.0	14.7
Historical change in CH_4 Flux	0.0	-0.6	-16.2	0.0	0.0	0.0	-16.8
Mexico							
CH_4 Flux in Current Wetlands	0.0	0.4	0.7	0.0	0.0	ND*	1.1
Historical change in CH_4 Flux	0.0	-0.5		0.0	0.0	ND	-0.5
North America							
CH_4 Flux in Current Wetlands	1.3	5.9	19.1	0.1	0.1	0.1	26.5
Historical change in CH_4 Flux	0.0	-24.2		0.0	0.0	0.0	-24.2
Global							
CH_4 Flux in Current Wetlands	14.1[d]	22.5[d]	68.0[d]	0.1[e]	1.4	ND	164[f]
Historical change in CH_4 Flux	-3.6		-79	0.0[g]	-0.5	ND	-83

*ND indicates that no data are available.

[a]Used CH_4 flux of 2.5 g m^{-2} yr^{-1} (range 0 to 130, likely mean 2–3) (Moore and Roulet, 1995) for Canadian peatlands and all Alaskan freshwater wetlands. Used CH_4 flux of 36.0 g m^{-2} yr^{-1} for Canadian freshwater mineral-soil wetlands and all U.S. and Mexican freshwater wetlands and 10.3 g m^{-2} yr^{-1} for estuarine wetlands—from synthesis of published CH_4 fluxes for the United States (see Table 13A-5).

[b]Includes a 17-fold increase in CH_4 flux (Kelly *et al*., 1997) in the 9000 km^2 of reservoirs that have been formed on peatlands (Rubec, 1996) and an estimated CH_4 flux of 15 g m^{-2} yr^{-1} (Moore *et al*., 1998) from 2,630 km^2 of melted permafrost peatlands (Vitt *et al.*, 1994).

[c]Assumed trace gas fluxes from unvegetated estuarine wetlands (i.e., mudflats) was the same as adjacent wetlands.

[d]Bartlett and Harriss (1993).

[e]Assumed that global rates approximate the North America rate because most salt marshes area is in North America.

[f]Ehhalt *et al.* (2001), range of 92 to 237 Mt yr^{-1}.

[g]Assumed a conservative 25% loss since the late 1800s.

Table 13A-5. Methane fluxes measured in the conterminous United States. The conversion factor is the ratio of the daily average flux to the measured annual flux × 10^3. The calculated annual flux was determined based upon the average conversion factor for freshwater (FW) and saltwater wetlands (SW). The measured annual flux was used if that was available; otherwise, the calculated annual flux was used.

Habitat	State	Method[a]	Salt/ Fresh	Daily Average Flux (mg CH_4 m^{-2} d^{-1})	Measured Annual Flux (g CH_4 m^{-2} yr^{-1})	Conversion Factor	Calculated Annual Flux (g CH_4 m^{-2} yr^{-1})	Used Annual Flux (g CH_4 m^{-2} yr^{-1})	Reference
Fens	CO	C	FW		40.7			40.7	Chimner and Cooper (2003)
Wet Alpine Meadow	CO	C	FW	0.1			0.0	0.0	Neff *et al.* (1994)
Lake - Average	CO	C	FW	25.4			9.2	9.2	Smith and Lewis (1992)
Wetland - Average	CO	C	FW	28.3			10.3	10.3	Smith and Lewis (1992)
Nuphar Bed	CO	C	FW	202.1			73.6	73.6	Smith and Lewis (1992)
Tundra - Carex Meadow	CO	C	FW	2.8			1.0	1.0	West *et al.* (1999)
Tundra - Acomastylis Meadow	CO	C	FW	-0.5			-0.2	-0.2	West *et al.* (1999)
Tundra - Kobresia Meadow	CO	C	FW	-0.8			-0.3	-0.3	West *et al.* (1999)
Moist Grassy	CO	C	FW	6.1	1.9	0.32	2.2	1.9	Wickland *et al.* (1999)
Moist Mossy	CO	C	FW	1.5	0.5	0.33	0.5	0.5	Wickland *et al.* (1999)
Wetland	CO	C	FW		41.7			41.7	Wickland *et al.* (1999)
Hardwood Hammock	FL	C	FW	0.0			0.0	0.0	Bartlett *et al.* (1989)
Dwarf Cypress / Sawgrass	FL	C	FW	7.5			2.7	2.7	Bartlett *et al.* (1989)
Spikerush	FL	C	FW	29.4			10.7	10.7	Bartlett *et al.* (1989)
Sawgrass < 1m	FL	C	FW	38.8			14.1	14.1	Bartlett *et al.* (1989)
Sawgrass/Spkerush/Periphyton	FL	C	FW	45.1			16.4	16.4	Bartlett *et al.* (1989)
Swamp Forest	FL	C	FW	68.9			25.1	25.1	Bartlett *et al.* (1989)
Sawgrass > 1m	FL	C	FW	71.9			26.2	26.2	Bartlett *et al.* (1989)
Sawgrass	FL	C	FW	107.0			38.9	38.9	Burke *et al.* (1988)
Pond Open Water	FL	C	FW	624.0			227.1	227.1	Burke *et al.* (1988)
Everglades - Cladium	FL	C	FW	45.4			16.5	16.5	Chanton *et al.* (1993)
Everglades - Typha	FL	C	FW	142.9			52.0	52.0	Chanton *et al.* (1993)
Wet Prairie (Marl)	FL	C	FW	87.0			31.6	31.6	Happell *et al.* (1993)
Wet Prairie (Marl)	FL	C	FW	27.4			10.0	10.0	Happell *et al.* (1993)
Marsh (Marl)	FL	C	FW	30.0			10.9	10.9	Happell *et al.* (1993)
Marsh (Marl)	FL	C	FW	49.6			18.0	18.0	Happell *et al.* (1993)
Marsh (Peat)	FL	C	FW	45.4			16.5	16.5	Happell *et al.* (1993)

Marsh (Peat)	FL	C	FW	13.0			4.7	4.7	Happell *et al.* (1993)
Marsh (Peat)	FL	C	FW	163.6			59.6	59.6	Happell *et al.* (1993)
Marsh (Peat)	FL	C	FW	20.4			7.4	7.4	Happell *et al.* (1993)
Wet Prairie / Sawgrass	FL	C	FW	61.0			22.2	22.2	Harriss *et al.* (1988)
Wetland Forest	FL	C	FW	59.0			21.5	21.5	Harriss *et al.* (1988)
Cypress Swamp - Flowing Water	FL	C	FW	67.0			24.4	24.4	Harriss and Sebacher (1981)
Open Water Swamp	FL	C	FW	480.0			174.7	174.7	Schipper and Reddy (1994)
Waterlily Slough	FL	C	FW	91.0			33.1	33.1	Schipper and Reddy (1994)
Cypress Swamp - Deep Water	GA	C	FW	92.3			33.6	33.6	Harriss and Sebacher (1981)
Bottotmand Hardwoods/ Swamps	GA	C	FW		23.0			23.0	Pulliam (1993)
Swamp Forest	LA	C	FW	146.0			53.1	53.1	Alford *et al.* (1997)
Freshwater Marsh	LA	C	FW	251.0			91.4	91.4	Alford *et al.* (1997)
Fresh	LA	C	FW	587.0	213.0	0.36	213.6	213.0	DeLaune *et al.* (1983)
Fresh	LA	C	FW	49.0	18.7	0.38	17.8	18.7	DeLaune *et al.* (1983)
Sphagnum Bog	MD	C	FW	-1.1			-0.4	-0.4	Yavitt et al. (1990)
Bog	MI	C	FW	193.0			70.2	70.2	Shannon and White (1994)
Bog	MI	C	FW	28.0			10.2	10.2	Shannon and White (1994)
Beaver Meadow	MN	C	FW		2.3			2.3	Bridgham *et al.* (1995)
Open Bogs	MN	C	FW		0.0			0.0	Bridgham *et al.* (1995)
Bog (Forested Hummock)	MN	C	FW	10.0	3.5	0.35	3.6	3.5	Dise (1993)
Bog (Forested Hollow)	MN	C	FW	38.0	13.8	0.36	13.8	13.8	Dise (1993)
Fen Lagg	MN	C	FW	35.0	12.6	0.36	12.7	12.6	Dise (1993)
Bog (Open Bog)	MN	C	FW	118.0	43.1	0.37	42.9	43.1	Dise (1993)
Fen (Open Poor Fen)	MN	C	FW	180.0	65.7	0.37	65.5	65.7	Dise (1993)
Poor Fen	MN	C	FW	242.0			88.1	88.1	Dise and Verry (2001)
Sedge Meadow	MN	C	FW		11.7			11.7	Naiman *et al.* ((1991)
Submergent	MN	C	FW		14.4			14.4	Naiman *et al.* (1991)
Deep Water	MN	C	FW		0.5			0.5	Naiman *et al.* (1991)
Poor Fen	MN	T	FW		14.6			14.6	Shurpali and Verma (1998)
Submerged Tidal	NC	C, E	FW	144.8			52.7	52.7	Kelly *et al.* (1995)
Banks Tidal	NC	C, E	FW	20.1			7.3	7.3	Kelly *et al.* (1995)
Tidal Marsh	NC	C	FW	3.0	1.0	0.34	1.1	1.0	Megonigal and Schlesinger (2002)
Tidal Marsh	NC	C	FW	3.5	2.3	0.65	1.3	2.3	Megonigal and Schlesinger (2002)
Prairie Marsh	NE	T	FW		64.0			64.0	Kim *et al.* (1998)
Poor Fen	NH	C	FW	503.3	110.6	0.22	183.2	110.6	Carroll and Crill (1997)
Poor Fen	NH	C	FW		69.3			69.3	Frolking and Crill (1994)

Forested Peatland	NY	C	FW	0.6	0.2	0.37	0.2	0.2	Coles and Yavitt (2004)
Pools Forested Swamp	NY	C	FW	224.6	69.0	0.31	81.7	69.0	Miller *et al.* (1999)
Typha Marsh - Mineral Soils	NY	C	FW	344.4			125.3	125.3	Yavitt (1997)
Typha Marsh - Peat Soils	NY	C	FW	65.1			23.7	23.7	Yavitt (1997)
Typha Marsh - All soils	NY	C	FW	204.8			74.5	74.5	Yavitt (1997)
Cypress Swamp - Floodplain	SC	C	FW	9.9			3.6	3.6	Harriss and Sebacher (1981)
Swamp	VA	C	FW	470.3			171.2	171.2	Chanton *et al.* (1992)
Maple/gum Forested Swamp	VA	C	FW		0.5			0.5	Harriss *et al.* (1982)
Emergent Tidal Freshwater Marsh	VA	C	FW		96.2			96.2	Neubauer *et al.* (2000)
Oak Swamp (Bank Site)	VA	C	FW	117.0	43.7	0.37	42.6	43.7	Wilson *et al.* (1989)
Emergent Macrophytes (Peltandra)	VA	C	FW	155.0			56.4	56.4	Wilson *et al*. (1989)
Emergent Macrophytes (Smartweed)	VA	C	FW	83.0			30.2	30.2	Wilson *et al.* (1989)
Ash Tree Swamp	VA	C	FW	152.0			55.3	55.3	Wilson *et al.* (1989)
Bog	WA	C	FW	73.0			26.6	26.6	Lansdown *et al.* (1992)
Lowland Shrub and Forested Wetland	WI	T	FW		12.4			12.4	Werner *et al*. (2003)
Sphagnum Eriophorum (Poor Fen)	WV	C	FW	6.6			2.4	2.4	Yavitt *et al.* (1990)
Sphagnum Shrub (Fen)	WV	C	FW	0.1			0.0	0.0	Yavitt *et al.* (1990)
Polytrichum Shrub (Fen)	WV	C	FW	-0.1			0.0	0.0	Yavitt *et al.* (1990)
Sphagnum Forest	WV	C	FW	9.6			3.5	3.5	Yavitt *et al.* (1990)
Sedge Meadow	WV	C	FW	1.5			0.5	0.5	Yavitt *et al*. (1990)
Beaver Pond	WV	C	FW	250.0			91.0	91.0	Yavitt *et al.* (1990)
Low Gradient Headwater Stream	WV	C	FW	300.0			109.2	109.2	Yavitt *et al.* (1990)
Sphagnum-Eriophorum	WV	C	FW	52.1	19.0	0.37	18.9	19.0	Yavitt *et al*. (1993)
Polytrichum	WV	C	FW	41.1	15.0	0.37	15.0	15.0	Yavitt *et al*. (1993)
Sphagnum-Shurub	WV	C	FW	4.4	1.6	0.37	1.6	1.6	Yavitt *et al.* (1993)
Salt Marsh	DE	C	SW	0.5			0.2	0.2	Bartlett *et al.* (1985)
Red Mangroves	FL	C	SW	4.2			1.4	1.4	Bartlett *et al.* (1989)
Dwarf Red Mangrove	FL	C	SW	81.9			27.9	27.9	Bartlett *et al.* (1989)
High Marsh	FL	C	SW	3.9			1.3	1.3	Bartlett *et al.* (1985)
Salt Marsh	FL	C	SW	0.6			0.2	0.2	Bartlett *et al.* (1985)
Salt Water Mangroves	FL	C	SW	4.0			1.4	1.4	Harriss *et al.*(1988)
Salt Marsh	GA	C	SW	13.4			4.6	4.6	Bartlett *et al*. (1985)

Short Spartina Marsh - High Marsh	GA	C	SW	145.2	53.1	0.37	49.5	53.1	King and Wiebe (1978)
Mid Marsh	GA	C	SW	15.8	5.8	0.37	5.4	5.8	King and Wiebe (1978)
Tall Spartina Marsh - Low Marsh	GA	C	SW	1.2	0.4	0.34	0.4	0.4	King and Wiebe (1978)
Intermediate Marsh	LA	C	SW	**912**[b]					Alford *et al.* (1997)
Salt Marsh	LA	C	SW	15.7	5.7	0.36	5.4	5.7	DeLaune *et al.* (1983)
Brackish	LA	C	SW	267.0	97.0		91.1	97.0	DeLaune *et al.* (1983)
Salt Marsh	LA	C	SW	4.8	1.7	0.35	1.6	1.7	DeLaune *et al.* (1983)
Brackish	LA	C	SW	17.0	6.4	0.38	5.8	6.4	DeLaune *et al.* (1983)
Cypress Swamp - Floodplain	SC	C	SW	1.5			0.5	0.5	Bartlett *et al.* (1985)
Salt Marsh	SC	C	SW	0.4			0.1	0.1	Bartlett *et al.* (1985)
Salt Marsh	VA	C	SW	3.0	1.3	0.43	1.0	1.3	Bartlett *et al.* (1985)
Salt Marsh	VA	C	SW	5.0	1.2	0.24	1.7	1.2	Bartlett *et al.* (1985)
Salt Meadow	VA	C	SW	2.0	0.4	0.22	0.7	0.4	Bartlett *et al.* (1985)
Salt Marsh	VA	C	SW	-0.8			-0.3	-0.3	Bartlett *et al.* (1985)
Salt Marsh	VA	C	SW	1.5			0.5	0.5	Bartlett *et al.* (1985)
Salt Meadow	VA	C	SW	-1.9			-0.6	-0.6	Bartlett *et al.* (1985)
Tidal Salt Marsh	VA	C	SW	16.0	5.6	0.35	5.5	5.6	Bartlett *et al.* (1987)
Tidal Brackish Marsh	VA	C	SW	64.6	22.4	0.35	22.0	22.4	Bartlett *et al.* (1987)
Tidal Brackish/Fresh Marsh	VA	C	SW	53.5	18.2	0.34	18.2	18.2	Bartlett *et al.* (1987)
				FW Average =	32.1	0.36	38.6	36.0	
				FW n =	32	18	74	88	
				FW StError=	7.9	0.02	6.0	5.0	
				SW Average =	16.9	0.34	9.8	10.3	
				SW n =	13	12	25	25	

SW StError=	7.8	0.02	4.1	4.4	

[a]C = chamber, T = tower, eddy covariance, E = ebulition measured separately.

[b]Outlier that was removed from further analysis.

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Chapter 14. Human Settlements and the North American Carbon Cycle

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KEY FINDINGS

- Human settlements occupy almost 5 % of the North American land area.
- There is currently insufficient information to determine the complete carbon balance of human settlements in North America. Fossil fuel emissions, however, very likely dominate carbon fluxes from settlements.
- An estimated 410 to 1679 Mt C are currently stored in the urban tree component of North American settlements. The growth of urban trees in North America produces a sink of approximately 16 to 49 Mt C yr^{-1}, which is 1 to 3% of the fossil fuel emissions from North America in 2003.
- Estimates of historical trends of the net carbon balance of North American settlements are not available. Fossil fuel emissions have likely gone up with the growth of urban lands but the net balance of carbon loss during conversion of natural to urban or suburban land cover and subsequent sequestration in lawns and urban trees is highly uncertain.
- The density and development patterns of human settlements are drivers of fossil fuel emissions, especially in the residential and transportation sectors. Biological carbon gains and losses are influenced by type of predevelopment land cover, post-development urban design and landscaping choices, soil and landscape management practices, and the time since land conversion.
- Projections of future trends in the net carbon balance of North American settlements are not available. However, the projected expansion of urban areas in North America will strongly impact the future North American carbon cycle as human settlements affect (1) the direct emission of CO_2 from fossil fuel combustion, (2) alter plant and soil carbon cycling in converting wild lands to residential and urban land cover.
- A number of municipalities in Canada, Mexico, and the U.S. have made commitments to voluntary GHG emission reductions under the Cities for Climate Protection program of International Governments for Local Sustainability [formerly the International Council for Local Environmental Initiatives (ICLEI)]. Reductions have in some cases been associated with improvements in air quality.

- Research is needed to improve comprehensive carbon inventories for settled areas, to improve understanding of how development processes relate to driving forces for the carbon cycle, and to improve linkages between understandings of human and environmental systems in settled areas.

Activities in human settlements form the basis for much of North America's contribution to global CO_2 emissions. Settlements such as cities, towns, and suburbs vary widely in density, form, and distribution. Urban settlements, as they have been defined by the census bureaus of the United States, Canada, and Mexico, make up approximately 75 to 80% of the population of the continent, and this proportion is projected to continue to increase (United Nations, 2004). The density and forms of new development will strongly impact the future trajectory of the North American carbon cycle as human settlements affect the carbon cycle by (1) direct emission of CO_2 from fossil fuel combustion, (2) alterations to plant and soil carbon cycles in conversion of wildlands to residential and urban land cover, and (3) indirect effects of residential and urban land cover on energy use and ecosystem carbon cycling.

CARBON INVENTORIES OF HUMAN SETTLEMENTS

Conversion of agricultural and wildlands to settlements of varying densities is occurring at a rapid rate in North America, faster, in fact, than the rate of population growth. For example, according to U.S. Census Bureau estimates, urban land in the coterminous United States increased by 23% in the 1990s (Nowak *et al.*, 2005) while the population increased by 13%. Given these trends, it is important to determine the carbon balance of different types of settlements and how future urban policy and planning may impact the magnitude of CO_2 sources and sinks at regional, continental, and global scales. However, unlike many other types of common land cover, complete carbon inventories including fossil fuel emissions and biological sources and sinks of carbon have been conducted only rarely for settlements as a whole. Assessing the carbon balance of settlements is challenging, as they are characterized by large CO_2 emissions from fuel combustion and decomposition of organic waste as well as transformations to vegetation and soil that affect carbon sources and sinks.

Determining the extent of human settlements across North America also presents a challenge, as definitions of "developed," "built-up," and "urban" land vary greatly, particularly among nations. The U.S., Canadian, and Mexican census definitions are not consistent; in addition, several other classification schemes for defining and mapping settlements have been developed, such as the U.S. Department of Agriculture's National Resource Inventory categorization of developed land, which uses a variety of methods based on satellite imagery and ground-based information. One method of classifying settled land cover that has been consistently applied at a continental scale is the Global Rural-Urban Mapping Project

conducted by a consortium of institutions, including Columbia University and the World Bank (CIESIN *et al.*, 2004). This estimate, which is based on nighttime lights satellite imagery, is 1,039,450 km^2, almost 5 % of the total continental land area (Fig. 14-1).

Fig. 14-1. North America urban extents.

Currently, there is insufficient information to determine the complete current or historical carbon balance of total continental land area. Fossil fuel emissions very likely dominate carbon fluxes from settlements, just as settlement-related emissions likely dominate total fossil fuel consumption in North America. However, specific estimates of the proportion of total fossil fuel emissions directly attributable to settlements are difficult to make given current inventory methods, which are often conducted on a state or province-wide basis. In addition, the biological component of the carbon balance of settlements is highly uncertain, particularly with regard to the influence of urbanization on soil carbon pools and biogenic greenhouse gas emissions.

For the urban tree component of the settlement carbon balance, carbon stocks and sequestration have been estimated for urban land cover (as defined by the U.S. Census Bureau) in the coterminous United States to be on the order of 700 Mt (335–980 Mt C) with sequestration rates of 22.8 Mt C yr^{-1} (13.7–25.9 Mt C yr^{-1}) (Nowak and Crane, 2002). These estimates encompass a great deal of regional variability and contain some uncertainty about differences in carbon allocation between urban and natural trees, as urban trees have been less studied. However, to a first approximation, these estimates can be used to infer a probable range of urban tree carbon stocks and gross sequestration on a continental basis. Nowak and Crane (2002) estimated that urban tree carbon storage in the Canadian border states (excluding semi-arid Montana, Idaho, and North Dakota) ranged from 24 to 45 t C ha^{-1}, and carbon sequestration ranged from 0.8 to 1.5 t C ha^{-1} yr^{-1}. Applying these values to a range of estimates of the extent of urban land in Canada (28,045 km^2 from the 1996 Canadian Census and 131,560 km^2 from CIESIN *et al.*, 2004), Canadian urban forest carbon stocks are between 67 and 592 Mt while carbon sequestration rates are between 2.2 and 19.7 Mt C yr^{-1}. Similarly, for Mexico, Nowak and Crane (2002) estimated that urban carbon storage and sequestration in the U.S. southwestern states varied from 4.4 to 10.5 t ha^{-1} and 0.1 to 0.3 t $ha^{-1}yr^{-1}$, respectively, leading to estimates of 10 to 107 Mt C stored in urban trees in Mexico and 0.2 to 3.1 Mt C yr^{-1} sequestered. Estimates of historical trends are not available.

While complete national or continental-scale estimates of the carbon budget of settlements including fossil fuels, vegetation, and soils are not available, several methods are available to assess the full carbon balance of individual settlements and can be applied in the next several years toward constructing larger-scale inventories. Atmospheric measurements can be used to determine the net losses of carbon from

settlements and urbanizing regions (Grimmond *et al.*, 2002; Grimmond *et al.*, 2004; Nemitz *et al.*, 2002; Soegaard and Moller-Jensen, 2003). Specific sources of CO_2 can be determined from unique isotopic signatures (Pataki *et al.*, 2003; Pataki *et al.*, 2006b) and from the relationship between CO_2 and carbon monoxide (Lin *et al.*, 2004). Many of these techniques have been commonly applied to natural ecosystems and may be easily adapted for settled regions. In addition, there have been several attempts to quantify the "metabolism" of human settlements in terms of their inputs and outputs of energy, materials, and wastes (Decker *et al.*, 2000) and the "footprint" of settlements in terms of the land area required to supply their consumption of resources and to offset CO_2 emissions (Folke *et al.*, 1997). Often these calculations include local flows and transformations of materials as well as upstream energy use and carbon appropriation, such as remote electrical power generation and food production.

To conduct metabolic and footprint analyses of specific settlements, energy and fuel use statistics are needed for individual municipalities, and these data are seldom made available at that scale. Consequently, metabolic and footprint analyses of carbon flows and conversions associated with metropolitan regions have been conducted for a relatively small number of cities. A metabolic analysis of the Toronto metropolitan region showed per capita net CO_2 emissions of 14 t CO_2 yr^{-1} (Sahely *et al.*, 2003), higher than analyses of other large metropolitan areas in developed countries (Newman, 1999; Pataki *et al.*, 2006a; Warren-Rhodes and Koenig, 2001). In contrast, an analysis of Mexico City estimated per capita CO_2 emissions of 3.4 t CO_2 yr^{-1} (Romero Lankao *et al.*, 2004). Local emissions inventories can provide useful supplements to national and global inventories in order to ensure that emissions reductions policies are applied effectively and equitably (Easterling *et al.*, 2003).

Current projections for urban land development in North America highlight the importance of improving carbon inventories of settlements and assessing patterns and impacts of future urban and rural development. Projections for increases in the extent of developed, nonfederal land cover in the United States in the next 25 years are as high as 79%, which would increase the proportion of developed land from 5.2% to 9.2% of total land cover (Alig *et al.*, 2004). The potential consequences of this increase for the carbon cycle are significant in terms of CO_2 emissions from an expanded housing stock and transportation network as well as from conversion of agricultural land, forest, rangeland, and other ecosystems to urban land cover. Because the dynamics of carbon cycling in settled areas encompass a range of physical, biological, social, and economic processes, studies of the potential impacts of future development on the carbon cycle must be interdisciplinary. Large-scale research on what has been called the study "of cities as ecosystems" (Pickett *et al.*, 2001) has begun only relatively recently, pioneered by interdisciplinary studies such as the National Science Foundation's Long-Term Ecological Research sites in the central Arizona-Phoenix area and in Baltimore (Grimm *et al.*, 2000). Although there is not yet sufficient data to construct a complete carbon inventory of settlements across North America, it is a

feasible research goal to do so in the next several years if additional studies in individual municipalities are conducted in a variety of urbanizing regions.

TRENDS AND DRIVERS

Drivers of change in the carbon cycle associated with human settlements include (1) factors that influence the rate of land conversion and urbanization, such as population growth and density, household size, economic growth, and transportation infrastructure; (2) additional factors that influence fossil fuel emissions, such as climate, residence and building characteristics, transit choices, and affluence; and (3) factors that influence biological carbon gains and losses, including the type of predevelopment land cover, post-development urban design and landscaping choices, soil and landscape management practices, and the time since land conversion.

Fossil Fuel Emissions

The density and patterns of development of human settlements (i.e., their "form") are drivers of the magnitude of the fossil fuel emissions component of the carbon cycle. The size and number of residences and households influence CO_2 emissions from the residential sector, and the spatial distribution of residences, commercial districts, and transportation networks is a key influence in the vehicular and transportation sectors. Many of the attributes of urban form that influence the magnitude of fossil fuel emissions are linked to historical patterns of economic development, which have differed in Canada, the United States, and Mexico. The future trajectory of development and associated levels of affluence and technological and social change will strongly influence key aspects of urban form such as residence size, vehicle miles traveled, and investment in urban infrastructure, along with associated fossil fuel emissions. Whereas emissions from the transportation and residential sectors are discussed in detail in Chapters 7 and 9, respectively, this chapter discusses specific aspects of the form of human settlements that affect the current continental carbon balance and its possible future trajectories.

Household size in terms of the number of occupants per household has been declining in North America (Table 14-1) while the average size of new residences has been increasing. For example, the average size of new, single family homes in the United States increased from 139 m^2 (1500 ft^2) to more than 214 m^2 (2300 ft^2) between 1970 and 2004 (NAHB, 2005). These trends have contributed to increases in per capita CO_2 emissions from the residential sector as well as increases in the consumption of land for residential and urban development (Alig *et al.*, 2003; Ironmonger *et al.*, 1995; Liu *et al.*, 2003; MacKellar *et al.*, 1995). In addition, when considering total emissions from settlements, the trajectory of the transportation and residential sectors may be linked. There have been a number of qualitative discussions of the role of "urban sprawl" in influencing fossil fuel and pollutant emissions from cities (CEC, 2001;

Gonzalez, 2005), although definitions of urban sprawl vary (Ewing *et al.*, 2003). Quantitative linkages between urban form and energy use have been attempted by comparing datasets for a variety of cities, but the results have been difficult to interpret due to the large number of factors that may affect transportation patterns and energy consumption (Anderson *et al.*, 1996). For example, in a seminal analysis of data from a variety of cities, Kenworthy and Newman (1990) found a negative correlation between population density and per capita energy use in the transportation sector. However, their data have been reanalyzed and reinterpreted in a number of subsequent studies that have highlighted other important driving variables, such as income levels, employment density, and transit choice (Gomez-Ibanez, 1991; Gordon and Richardson, 1989; Mindali *et al.*, 2004).

Table 14-1. Increases in number of households and the total population of the United States, Canada, and Mexico between 1985 and 2000. (United Nations, 2002; United Nations Habitat, 2003).

Quantifying the nature and extent of the linkage between development patterns of human settlements and greenhouse gas emissions is critical from the perspective of evaluating the potential impacts of land use policy. One way forward is to further the application of integrated land use and transportation models that have been developed to analyze future patterns of urban development in a variety of cities (Agarwal *et al.*, 2000; EPA, 2000; Hunt *et al.*, 2005). Only a handful have been applied to date for generating fossil fuel emissions scenarios from individual metropolitan areas (Jaccard *et al.*, 1997; Pataki *et al.*, 2006a), such that larger-scale national or continental projections for human settlements are not currently available. However, there is potential to add a carbon cycle component to these models that would assess the linkages between land use and land cover change, residential and commercial energy use and emissions, emissions from the transportation sector, and net carbon gains and losses in biological sinks following land conversion. A critical feature of these models is that they may be used to evaluate future scenarios and the potential impacts of policies to influence land use patterns and transportation networks in individual settlements and developing regions.

Vegetation and Soils in Human Settlements

Human settlements contain vegetation and soils that are often overlooked in national inventories, as they fall outside common classification schemes. Nevertheless, patterns of development affect the carbon balance of biological systems, both in the replacement of natural ecosystems with rural, residential, or urban land cover and in processes within settlements that affect constructed and managed land cover. In the United States, satellite data and ecosystem modeling for the mid-1990s suggested that urbanization

occurred largely on productive agricultural land and therefore caused a net loss of carbon fixed by photosynthesis of 40 Mt C yr^{-1} (Imhoff *et al.*, 2004).

Urban forests and vegetation sequester carbon directly as described under carbon inventories. In addition, urban trees influence the carbon balance of municipalities indirectly through their effects on energy use. Depending on their placement relative to buildings, trees may cause shading and windbreak effects, as well as evaporative cooling due to transpiration (Akbari, 2002; Oke, 1989; Taha, 1997). These effects have been estimated in a variety of studies, mostly involving model calculations that suggest that urban trees generally result in net reductions in energy use (Akbari, 2002; Akbari and Konopacki, 2005; Akbari *et al.*, 1997; Akbari and Taha, 1992; Huang *et al.*, 1987). Taking into account CO_2 emissions resulting from tree maintenance and decomposition of removed trees, "avoided" emissions from energy savings were responsible for approximately half of the total net reduction in CO_2 emissions from seven municipal urban forests, with the remainder attributable to direct sequestration of CO_2 (McPherson *et al.*, 2005). Direct measurements of the components of urban energy balance that quantify the contribution of vegetation are needed to validate these estimates.

Like natural ecosystems, soils in human settlements contain carbon, although rates of sequestration are much more uncertain in urban soils than in natural soils. In general, soil carbon is generally lost following disturbances associated with conversion from natural to urban or suburban land cover (Pouyat *et al.*, 2002). Soil carbon pools may subsequently increase at varying rates, depending on the soil and land cover type, local climate, and management intensity (Golubiewski, 2006; Pouyat *et al.*, 2002; Qian and Follet, 2002). In ecosystems with low rates of carbon sequestration in native soil such as arid and semiarid ecosystems, conversion to highly managed, settled land cover can result in higher rates of carbon sequestration and storage than pre-settlement due to large inputs of water, fertilizer, and organic matter (Golubiewski, 2006). Pouyat *et al.* (2006) used urban soil organic carbon measurements to estimate the total above- and below-ground carbon storage, including soil carbon, in U.S. urban land cover to be 2,640 Mt (1,890 to 3,300 Mt). This range does not include the uncertainty in classifying urban land cover, but applies the range of uncertainty in aboveground urban carbon stocks reported in Nowak and Crane (2002) and the standard deviation of urban soil carbon densities reported in Pouyat *et al*. (2006). In addition, irrigated and fertilized urban soils have been associated with higher emissions of CO_2 and the potent greenhouse gas N_2O relative to natural soils, offsetting some potential gains of sequestering carbon in urban soils (Kaye *et al.*, 2004; Kaye *et al.*, 2005; Koerner and Klopatek, 2002). Finally, full carbon accounting that incorporates fossil fuel emissions associated with soil management (e.g., irrigation and fertilizer production and transport) has not yet been conducted. In general, additional data on soil carbon balance in human settlements are required to assess the potential for managing urban and residential soils for carbon sequestration.

OPTIONS FOR MANAGEMENT

A number of municipalities in Canada, the United States, and Mexico have committed to voluntary programs of greenhouse gas emissions reductions. Under the Cities for Climate Protection program (CCP) of International Governments for Local Sustainability (ICLEI, formerly the International Council of Local Environmental Initiatives) 269 towns, cities, and counties in North America have committed to conducting emissions inventories, establishing a target for reductions, and monitoring the results of reductions initiatives (the current count of the number of municipalities participating in voluntary greenhouse gas reduction programs may be found on-line at http://www.iclei.org). Emissions reductions targets vary by municipality, as do the scope of reductions, which may apply to the municipality as a whole or only to government operations (i.e., emissions related to operation of government-owned buildings, facilities, and vehicle fleets).

Kousky and Schneider (2003) interviewed representatives from 23 participating CCP municipalities in the United States who indicated that cost savings and other co-benefits of greenhouse gas reductions in cities and towns were the most commonly cited reasons for participating in voluntary greenhouse gas reductions programs. Potential cost savings include reductions in energy and fuel costs from energy efficiency programs in buildings, street lights, and traffic lights; energy co-generation in landfills and sewage treatment plants; mass transit programs; and replacement of municipal vehicles and buses with alternative fuel or hybrid vehicles (ICLEI, 1993; 2000). Other perceived co-benefits include reductions in emissions of particulate and oxidant pollutants, alleviation of traffic congestion, and availability of lower-income housing in efforts to curb urban sprawl. These co-benefits are often "perceived" because many municipalities have not attempted to quantify them as part of their emissions reductions programs (Kousky and Schneider, 2003); however, it has been suggested that they play a key role in efforts to promote reductions of municipal-scale greenhouse gas emissions because local constituents regard them as an issue of interest (Betsill, 2001).

Of the co-benefits of municipal programs to reduce CO_2 emissions, improvements in air quality are perhaps the most well studied. Cifuentes (2001) analyzed the benefits of reductions in atmospheric particulate matter measuring less than 10 μm in diameter (PM10) and ozone concentrations in four cities in North and South America. Using a greenhouse gas reduction of 13% of 2000 levels by 2020 from energy efficiency and fuel substitution programs, Cifuentes (2001) estimated that PM10 and ozone concentrations would decline by 10% of 2000 levels. Estimated health benefits from such a reduction included avoidance of 64,000 (18,000–116,000) premature deaths associated with air quality-related heath problems as well as avoidance of 91,000 (28,000–153,000) hospital admissions and 787,000 (136,000–1,430,000) emergency room visits. However, using calculations for co-control of CO_2 and air pollutants

in Mexico City, West *et al.* (2004) found that in practice, if electrical energy is primarily generated in remote locations relative to the urban area, cost-effective energy efficiency programs may have a relatively small effect on air quality. In that case, options for reducing greenhouse gas emissions would have to be implemented primarily in the transportation sector to appreciably affect air quality.

RESEARCH NEEDS

Additional studies of the carbon balance of settlements of varying densities, geographical location, and patterns of development are needed to quantify the potential impacts of various policy and planning alternatives on net greenhouse gas emissions. While it may seem intuitive that policies to curb urban sprawl or enhance tree planting programs will result in emissions reductions, different aspects of urban form (e.g., housing density, availability of public transportation, type and location of forest cover) may have different net effects on carbon sources and sinks, depending on the location, affluence, economy, and geography of various settlements. It is possible to develop quantitative tools to take many of these factors into account. To facilitate development and application of integrated urban carbon cycle models and to extrapolate local studies to regional, national, and continental scales, useful additional data include:

- common land cover classifications appropriate for characterizing a variety of human settlements across North America,
- emissions inventories at small spatial scales such as individual neighborhoods and municipalities,
- expansion of the national carbon inventory and flux measurement networks to include land cover types within human settlements,
- comparative studies of processes and drivers of development in varying regions and nations, and
- interdisciplinary studies of land use change that evaluate socioeconomic as well as biophysical drivers of carbon sources and sinks.

In general, there has been a focus in carbon cycle science on measuring carbon stocks and fluxes in natural ecosystems, and consequently highly managed and human-dominated systems such as settlements have been underrepresented in many regional and national inventories. To assess the full carbon balance of settlements ranging from rural developments to large cities, a wide range of measurement techniques and scientific, economic, and social science disciplines are required to understand the dynamics of urban expansion, transportation, economic development, and biological sources and sinks. An advantage to an interdisciplinary focus on the study of human settlements from a carbon cycle perspective is that human activities and biological impacts in and surrounding settled areas encompass many aspects of perturbations to atmospheric CO_2, including a large proportion of national CO_2 emissions and changes in carbon sinks resulting from land use change.

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Table 14-1. Increases in number of households and the total population of the United States, Canada, and Mexico between 1985 and 2000. (United Nations, 2002; United Nations Habitat, 2003).

	Total population (%)	Households (%)
Canada	19	39
Mexico	33	60
United States	15	25

Figure 14-1. North America urban extents.

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Chapter 15. Coastal Oceans

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KEY FINDINGS

- The combustion of fossil fuels has increased CO_2 in the atmosphere, and by diffusion the oceans have absorbed an equivalent of 20-30% of the released CO_2 on an annual basis. The present annual uptake by the oceans of 1.3-2.3 Gt C is well constrained, has slightly acidified the oceans and may ultimately affect ocean ecosystems in unpredictable ways.
- The carbon budgets of ocean margins (coastal regions) are not as well-characterized due to lack of observations coupled with complexity and highly localized spatial variability. Existing data are insufficient, for example, to estimate the amount of anthropogenic carbon stored in the coastal regions of North America or to predict future scenarios.
- New air-sea flux observations reveal that on average, nearshore waters surrounding North America are neither a source nor a sink of CO_2 to the atmosphere. A small net source of CO_2 to the atmosphere of 19 Mt C yr^{-1} is estimated mostly from waters around the Gulf of Mexico and the Caribbean Sea, with a variation (standard deviation) around that number of $\pm$ 22 Mt C yr^{-1}. This equates to 1% of the global ocean uptake.
- With the exception of one or two time-series sites, almost nothing is known about historical trends in air-sea fluxes and the source-sink behavior of North America's coastal oceans.
- The Great Lakes and estuarine systems of North America may be net sources of CO_2 where terrestrially-derived organic material is decomposing, while reservoir systems may be storing carbon through sediment transport and burial.
- Options and measures for sequestration of carbon in the ocean include deep-sea injection of CO_2 and iron fertilization, although it is unresolved how important, feasible or acceptable any of these options might be for the North American region. Ocean carbon sequestration studies should be continued.

- Highly variable air-sea CO_2 fluxes in coastal areas may introduce errors in North American CO_2 fluxes calculated by atmospheric inversion methods. Reducing these errors will require ocean observatories utilizing fixed and mobile platforms with instrumentation to measure critical stocks and fluxes as part of coordinated national and international research programs. Ocean carbon sequestration studies should also be continued.

INVENTORIES (STOCKS AND FLUXES, QUANTIFICATION)

This chapter first introduces the role the oceans play in modulating atmospheric carbon dioxide (CO_2), then quantifies air-sea CO_2 fluxes in coastal waters surrounding North America and considers how the underlying processes affect the air-sea fluxes. Aquatic stocks of living carbon are small relative to stocks in the terrestrial environments, but turnover rates are very high. In addition aquatic stocks are not well characterized because of their spatial and temporal variability, the complexity of carbon compound transformations, and limited data on these processes. The oceans act as a huge reservoir for inorganic carbon, containing about 50 times as much CO_2 as the atmosphere. The ocean's biological pump converts CO_2 to organic particulate carbon by photosynthesis, transports the organic carbon from the surface by sinking, and therefore plays a critical role in removing atmospheric CO_2 in combination with physical and chemical processes (Gruber and Sarmiento, 2002; Sarmiento and Gruber, 2006). Atmospheric concentration of CO_2 would be much higher in the absence of current ocean processes implying that climate-driven changes in ocean circulation, chemical properties or biological rates could result in strong feedbacks to the atmosphere.

The release of CO_2 into the atmosphere by the combustion of fossil fuels has increased pre-industrial concentrations from around 280 ppm to present day levels of 380 ppm. This increase in atmospheric concentrations is driving more CO_2 into the ocean with the present net air-sea CO_2 flux well constrained to about 1,800 ± 500 Mt C [1 Mt = one million (10^6) metric tons] or 1.8 ± 0.5 Gt C yr^{-1} [1 Gt = one billion (10^9) metric tons] from the atmosphere into the ocean (Figure 15-1 and Table 15-1) (See Chapter 2 for a description of how ocean carbon fluxes relate to the global carbon cycle). The uptake of this anthropogenically-driven CO_2 by the oceans is on average turning them more acidic with negative and potentially catastrophic effects on some biota (Kleypas *et al.*, 2006). The atmosphere is well mixed and nearly homogenous so the large spatial variability in air-sea CO_2 fluxes shown in Figure 15-1 is driven by a combination of physical, chemical, and biological processes in the ocean. The flux over the coastal margins has neither been well characterized (Liu *et al.*, 2000) nor integrated into global calculations because there are large variations over small spatial and temporal scales, and observations have been limited. The need for higher spatial resolution to resolve the coastal variability has hampered modeling

efforts. In the following sections we review existing information on the coastal ocean carbon cycle and its relationship to the global ocean, and we present the results of a new analysis of about a half million observations of air-sea flux of CO_2 in coastal waters surrounding the North American continent.

Table 15-1. Climatological mean distribution of the net air-sea CO_2 flux (in Gt C yr^{-1}) over the global ocean (excluding coastal areas) in reference year 1995. Positive values indicate a source for atmospheric CO_2, and negative values indicate a sink. The fluxes are based on about 1.75 million partial pressure measurements for CO_2 in surface ocean waters, excluding the measurements made in the equatorial Pacific (10°N- 10°S) during El Niño periods (see Takahashi *et al.*, 2002). The NCAR/NCEP 42-year mean wind speeds and the (wind speed)^2 dependence for air-sea gas transfer rate are used (Wanninkhof, 1992) for calculating the air-sea flux. The flux, however, depends on the wind speed and air-sea gas transfer rate parameterizations used, and varies by about ± 30% (Takahashi *et al*., 2002). The ocean uptake has also been estimated on the basis of the following methods: temporal changes in atmospheric oxygen and CO_2 concentrations (Keeling and Garcia, 2002; Bender *et al.*, 2005), $^{13}C/^{12}C$ ratios in sea and air (Battle *et al.*, 2000; Quay *et al.*, 2003), ocean CO_2 inventories (Sabine *et al.*, 2004), and coupled carbon cycle and ocean general circulation models (Sarmiento *et al.*, 2000; Gruber and Sarmiento, 2002). The consensus is that the oceans take up 1.3 to 2.3 Gt C yr^{-1}

Figure 15-1. Global distribution of air-sea CO_2 flux. The map yields a total annual air-to-sea flux of 1.5 Gt C yr^{-1}. The white line represents zero flux and separates sources (yellow and red) and sinks (blue and purple). Negative values indicate that the ocean is a CO_2 sink for the atmosphere. The sources are primarily in the tropics (yellow and red) with a few areas of deep mixing at high latitudes. Updated from Takahashi *et al.* (2002).

Global Coastal Ocean Carbon Fluxes

The carbon cycle in coastal oceans involves a series of processes, including runoff from terrestrial environments, upwelling and mixing of high CO_2 water from below, photosynthesis at the sea surface, sinking of organic particles, respiration, production and consumption of dissolved organic carbon, and air-sea CO_2 fluxes (Figure 15-2). Although fluxes in the coastal oceans are large relative to surface area, there is disagreement as to whether these regions are a net sink or a net source of CO_2 to the atmosphere (Tsunogai *et al.*, 1999; Cai and Dai, 2004; Thomas *et al.*, 2004). Great uncertainties remain in coastal carbon fluxes, which are complex and dynamic, varying rapidly over short distances and at high frequencies. Only recently have new technologies allowed for the measurement of these rapidly changing fluxes (Friederich *et al.*, 1995 and 2002; Hales and Takahashi, 2004).

Figure 15-2. In the top panel, mean air/sea CO_2 flux is calculated from shipboard measurements on a line perpendicular to the central California coast. Flux within Monterey Bay (~0–20 km offshore) is into the ocean, flux across the active upwelling region (~20–75 km offshore) is from the ocean, and flux in the California Current (75–300 km) is on average into the ocean. These fluxes result from the processes shown in the bottom panel. California Undercurrent water, which has a high CO_2 partial pressure, upwells near shore, and is advected offshore towards the California Current and into Monterey Bay. Phytoplankton growth and photosynthesis draw down CO_2 in seawater to low levels in the upwelled water. Phytoplankton carbon eventually sinks or is subducted below the euphotic zone, where it decays, elevating the CO_2 levels of subsurface waters. Where the level of surface seawater CO_2 is higher than the atmosphere, CO_2 is driven into the atmosphere. Conversely, where the level of surface CO_2 is lower than that of atmospheric CO_2, CO_2 is driven from the atmosphere into the ocean. The net sea/air flux on this spatial scale is near zero. DIC = dissolved inorganic carbon; POC = particulate organic carbon. Updated from Pennington *et al.* (in press).

Carbon is transported from land to sea mostly by rivers in four components: CO_2 dissolved in water, organic carbon dissolved in water, particulate inorganic carbon (e. g. calcium carbonate, $CaCO_3$), and particulate organic carbon. The global rate of river input has been estimated to be 1,000 Mt C yr^{-1}, about 38% of it as dissolved CO_2 (or 384 Mt C yr^{-1}), 25% as dissolved organic matter, 21% as organic particles and 17% as $CaCO_3$ particles (Gattuso *et al.*, 1998). Estimates for the riverine dissolved CO_2 flux vary from 385 to 429 Mt C yr^{-1} (Sarmiento and Sundquist, 1992). The Mississippi River, the seventh-largest in freshwater discharge in the world, delivers about 13 Mt C yr^{-1} as dissolved CO_2 (Cai, 2003). Organic matter in continental shelf sediments exhibits only weak isotope and chemical signatures of terrestrial origin, suggesting that riverine organic matter is reprocessed in coastal environments on a time scale of 20 to 130 years (Hedges *et al.*, 1997; Benner and Opsahl, 2001). Of the organic carbon, about 30% is accumulating in estuaries, marshes, and deltas, and a large portion (20% to 60%) of the remaining 70% is readily and rapidly oxidized in coastal waters (Smith and Hollibaugh, 1997). Only about 10% is estimated to be contributed by human activities, such as agriculture and forest clearing (Gattuso *et al.*, 1998), and the rest is a part of the natural carbon cycle.

One of the major differences between coastal and open ocean systems is the activity of the biological pump. In coastal environments, the pump operates much more efficiently, leading to rapid reduction of surface CO_2 and thus complicating the accurate quantification of air-sea CO_2 fluxes. For example, Ducklow and McCallister (2004) constructed a carbon balance for the coastal oceans using the framework of the ocean carbon cycle of Gruber and Sarmiento (2002) and estimated a net CO_2 removal by primary productivity of 1,200 Mt C yr^{-1} and a large CO_2 sink of 900 Mt C yr^{-1} for the atmosphere. In contrast, Smith and Hollibaugh (1993) estimated a biological pump of about 200 Mt C yr^{-1} and concluded that the

coastal oceans are a weak CO_2 sink of 100 Mt C yr^{-1}, about one-ninth of the estimate by Ducklow and McCallister (2004). Since the estimated air-sea CO_2 flux depends on quantities that are not well constrained, the mass balance provides widely varying results. For this reason, in this chapter the net air-sea flux over coastal waters is estimated on the basis of direct measurements of the air-sea difference of partial pressure of CO_2 (pCO_2).

North American Coastal Carbon

Two important types of North American coastal ocean environments can be identified: (1) river-dominated coastal margins with large inputs of fresh water, organic matter, and nutrients from land (e.g., Mid- and South-Atlantic Bights) (Cai *et al*., 2003) and (2) coastal upwelling zones (e.g., the California-Oregon-Washington coasts, along the eastern boundary of the Pacific) where physical processes bring cool, high-nutrient and high-CO_2 waters to the surface. In both environments, the biological uptake of CO_2 plays an important role in determining whether an area becomes a sink or a source for the atmosphere.

High biological productivity fueled by nutrients added to coastal waters can lead to seawater becoming a CO_2 sink during the summer growing season, as observed in the Bering Sea Shelf (Codispoti and Friederich, 1986) and the northwest waters off Oregon and Washington (van Geen *et al.*, 2000; Hales *et al.*, 2005). Similar CO_2 draw-downs may occur in the coastal waters of the Gulf of Alaska and in the Gulf of Mexico near the Mississippi River outflow. Coastal upwelling results in a very high concentration of CO_2 for the surface water (as high as 1,000 μatm), and hence the surface water becomes a strong CO_2 source. This is followed by rapid biological uptake of CO_2, which causes the water to become a strong CO_2 sink (Friederich *et al.*, 2002; Hales *et al.*, 2005).

A review of North American coastal carbon fluxes has been carried out by Doney *et al.* (2004) (Table 15-2). The information reviewed was very limited in space (only 13 locations) and time, leading Doney *et al.* to conclude that it was unrealistic to reliably estimate an annual flux for North American coastal waters. Measurement programs have increased recently, and we have used the newly available data to calculate annual North American coastal air-sea fluxes for the first time.

Table 15-2. Variability of CO_2 distributions and fluxes in U.S. coastal waters from regional surveys and moored measurements (from Doney *et al.* 2004). Negative values indicate that the ocean is a CO_2 sink for the atmosphere.

Synthesis of Available North American Air-Sea Coastal CO_2 Fluxes

A large data set consisting of 550,000 measurements of the partial pressure of CO_2 (pCO_2) in surface waters has been assembled and analyzed (Figure 15-3; see Appendix 15A for details). pCO_2 is measured in a carrier gas equilibrated with seawater and, as such, it is a measure of the outflux/influx tendency of CO_2 from the atmosphere. CO_2 reacts with seawater and 99.5% of the total amount of CO_2 dissolved in seawater is in the form of bicarbonate (HCO_3^-) and carbonate ions ($CO_3^{=}$), which do not exchange with the overlying atmosphere. Only CO_2 molecules, which constitute about 0.5% of the total dissolved CO_2, exchange with the atmosphere. This is expressed as pCO_2, which is affected by physical and biological processes increasing with temperature and decreasing with photosynthesis. The data were obtained by the authors and collaborators, quality-controlled, and assembled in a uniform electronic format for analysis (available at www.ldeo.columbia.edu/res/ pi/CO2). Observations in each 1° × 1° pixel area were compiled into a single year and were analyzed for time-space variability. Seasonal and interannual variations were not well characterized except in a few locations (Friederich *et al.*, 2002). The annual mean air-sea pCO_2 difference (ΔpCO_2) was computed for 5°-wide zones along the North American continent and was plotted as a function of latitude for four regions (Figure 15-4): North Atlantic, Gulf of Mexico/Caribbean, North Pacific, and Bering/Chukchi Seas. Figure 15-4A shows the fluxes in the first nearshore band, and Figure 15-4B shows the fluxes for a band that is several hundred kilometers from shore. The average fluxes for them and for the intermediate bands are given in Table 15-3. The flux and area data are listed in Table 15-4. A full complement of seasonal observations are lacking in the Arctic Sea, including Hudson Bay, the northern Labrador Sea, and the Gulf of St. Lawrence; the northern Bering Sea; the Gulf of Alaska; the Gulf of California; and the Gulf of Mexico and the Caribbean Sea.

Figure 15-3. (A). Distribution of coastal CO_2 partial pressure measurements made between 1979 and 2004. (B). The distribution of the net air-sea CO_2 flux over 1° × 1° pixel areas (N-S 100 km, E-W 80 km) around North America. The flux (grams of carbon per square meter per year) represents the climatological mean over the 25-year period. The magenta-blue colors indicate that the ocean water is a sink for atmospheric CO_2, and the green-yellow-orange colors indicate that the sea is a CO_2 sink. The data were obtained by the authors and collaborators of this chapter and are archived at the Lamont-Doherty Earth Observatory (www.ldeo.columbia.edu/res/pi/CO2).

Figure 15-4. Estimated air-sea CO_2 fluxes (grams of carbon per square meter per year) from 550,000 seawater CO_2 partial pressure (pCO_2) observations made from 1979 to 2004 in ocean waters surrounding the North American continent. (A) Waters within one degree (about 80 km) of the coast and (B) open ocean waters between 300 and 900 km from the shore (see Figure 15-3B). The annual mean air-sea pCO_2 difference (ΔpCO_2) values were calculated from the weekly mean atmospheric CO_2

concentrations in the GLOBALVIEW-CO2 database (2004) over the same pixel area in the same week and year as the seawater pCO_2 was measured. The monthly net air-sea CO_2 flux was computed from the mean monthly wind speeds in the National Centers for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) database in the $(\text{wind speed})^2$ formulation for the air-sea gas transfer rate by Wanninkhof (1992). Negative values indicate that the ocean is a CO_2 sink for the atmosphere. The ± uncertainties represent one standard deviation.

Table 15-3. Climatological mean annual air-sea CO_2 flux (grams of carbon per square meter per year) over the oceans surrounding North America. Negative values indicate that the ocean is a CO_2 sink for the atmosphere. N is the number of seawater pCO_2 measurements. The ± uncertainty is given by one standard deviation of measurements used for analysis and represents primarily the seasonal variability.

The offshore patterns follow the same general trend found in the global open ocean data set shown in Figure 15-1. On an annual basis the lower latitudes tend to be a source of CO_2 to the atmosphere, whereas the higher latitudes tend to be sinks (Figures 15-3B and 15-4B). The major difference in the coastal waters is that the latitude where CO_2 starts to enter the ocean is further north than it is in the open ocean, particularly in the Atlantic. A more detailed region-by-region description follows.

Pacific Ocean

Observations made in waters along the Pacific coast of North America illustrate how widely coastal waters vary in space and time, in this case driven by upwelling and relaxation (Friederich *et al.*, 2002). Figure 15-5A shows a summertime quasi-synoptic distributions of temperature, salinity, and pCO_2 in surface waters based on measurements made in for July through September 2005. The effects of the Columbia River plume emanating from ~46°N are clearly seen (colder temperature, low salinity, and low pCO_2), as are coastal upwelling effects off Cape Mendocino (~40°N) (colder, high salinity, and very high pCO_2). These coastal features are confined to within 300 km from the coast. The 1997–2005 time-series data for surface water pCO_2 observed off Monterey Bay (Figure 15-5B) show the large, rapidly fluctuating air-sea CO_2 fluxes during the summer upwelling season in each year as well as the low-pCO_2 periods during the 1997–1998 and 2002–2003 El Niño events. In spite of the large seasonal variability, ranging from 200 to 750 μatm, the annual mean air-sea pCO_2 difference and the net CO_2 flux over the waters off Monterey Bay areas (~37°N) are close to zero (Pennington *et al.*, in press). The seasonal amplitude decreases away from the shore and in the open ocean bands, where the air-sea CO_2 flux changes seasonally in response to seawater temperature (out of the ocean in summer and into the ocean in winter).

Figure 15-5. Time-space variability of coastal waters off the west coast of North America. (A) Quasi-synoptic distribution of the temperature, salinity, and pCO_2 in surface waters during July–September 2005. The Columbia River plume (~46°N) and the upwelling of deep waters off the Cape Mendocino (~40°N) are clearly seen. (B) 1997–2005 time-series data for air-sea CO_2 flux from a mooring off Monterey Bay, California (the fluxes are reported in grams of carbon per square meter per year so they can be compared to values throughout the chapter). Seawater is a CO_2 source for the atmosphere during the summer upwelling events, but biological uptake reduces levels very rapidly. The rapid fluctuations seen in (B) can affect atmospheric CO_2 levels. For example, if CO_2 from the sea is mixed into a static column, a 500-m-thick planetary boundary layer over the course of one day, atmospheric CO_2 concentration would change by 2.5 μatm. If the column of air is mixed vertically through the troposphere to 500 mbar, a change of about 0.5 μatm would occur. The effects would be diluted as the column of air mixes laterally. However, this demonstrates that the large fluctuations of air-sea CO_2 flux observed over coastal waters could affect the concentration of CO_2 significantly enough to affect estimates of air-land flux based on the inversion of atmospheric CO_2 data. Air-sea CO_2 flux was low during the 1997–1998 and 2002–2003 El Niño periods.

The open ocean Pacific waters south of 30°N are on the annual average a CO_2 source to the atmosphere, whereas the area north of 40°N is a sink, and the zone between 30° and 40°N is neutral (Takahashi *et al.*, 2002). Coastal waters in the 40°N through 45°N zone (northern California-Oregon coasts) are even a stronger CO_2 sink, associated with nutrient input and stratification by fresh water from the Columbia River (Hales *et al.*, 2005). On the other hand, coastal pCO_2 values in the 15°N through 40°N zones have pCO_2 values similar to open ocean values and to the atmosphere. In the zones 15°N through 40°N, the annual mean values for the net air-sea CO_2 flux are nearly zero, consistent with the finding by Pennington *et al.* (in press).

Atlantic Ocean

With the exception of the 5°N–10°N zone, the open ocean areas are an annual net sink for atmospheric CO_2 with stronger sinks at high latitudes, especially north of 35°N (Figure 15-3B). In contrast the nearshore waters are a CO_2 source between 15°N and 45°N. Accordingly, in contrast to the Pacific coast, the latitude where Atlantic coastal waters become a CO_2 sink is located further north. In the areas north of 45°N, the open ocean waters are a strong CO_2 sink due primarily to the cold Labrador Sea waters.

In the coastal zone very high pCO_2 values (up to 2,600 μatm) are observed occasionally in areas within 10 km offshore of the barrier islands (see small red dots off the coasts of Georgia and Carolinas in Figures 15-3B). These waters which have salinities around 20 and high total CO_2 concentrations appear to represent outflow of estuarine/marsh waters rich in carbon (Cai *et al.*, 2003). The large contribution of

fresh water that is rich in organic matter relative to the Pacific contributes to this small coastal Atlantic source. Offshore fluxes are in phase with the seasonal cycle of warming and cooling; fluxes are out of the ocean in summer and fall and are the inverse in winter and spring.

Bering and Chukchi Seas

Although measurements in these high-latitude waters are limited, the relevant data for the Bering Sea (south of 65°N) and Chukchi Sea (north of 65°N) are plotted as a function of the latitude in Figure 15-4. The values for the areas north of 55°N are for the summer months only; CO_2 observations are not available during winter seasons. Although data scatter widely, the coastal and open ocean waters are a strong CO_2 sink during the summer months due to photosynthetic drawdown of CO_2. The data in the 70°–75°N zone are from the shallow shelf areas in the Chukchi Sea. These waters are a very strong CO_2 sink (air-sea pCO_2 differences ranging from −80 to −180 μatm) with little changes between the coastal and open ocean areas. The air-sea CO_2 flux during winter months is not known but the summer fluxes are shown in Figure 15-4 for comparison.

Gulf of Mexico and Caribbean Sea

Although observations are limited, available data suggest that these waters are a strong CO_2 source (Figure 15-4 and Table 15-3). A subsurface anoxic zone has been formed in the Texas-Louisiana coast as a result of the increased addition of anthropogenic nutrients and organic carbon by the Mississippi River (e.g., Lohrenz *et al.*, 1999). The carbon-nutrient cycle in the northern Gulf of Mexico is also being investigated (e.g., Cai, 2003), and the studies suggest that at times those waters are locally a strong CO_2 sink due to high biological production.

SYNTHESIS

An analysis of half a million measurements of air-sea flux of CO_2 shows that the nearshore (< 100 km) coastal waters surrounding North America are a net CO_2 source for the atmosphere on an annual average of about 19 ± 22 Mt C yr^{-1} (Table 15-4). Most of the flux (14 ± 9 Mt C yr^{-1}) occurs in the Gulf of Mexico and Caribbean Sea. The open oceans are a net CO_2 sink on an annual average (Table 15-4; Takahashi *et al.*, 2004). The reported uncertainties reflect the time-space variability but do not reflect uncertainties due to lack of observations in some portions of the Arctic Sea, Bering Sea, Gulf of Alaska, Gulf of Mexico, or Caribbean Sea. Observations in these areas will be needed to improve estimates. These results are consistent with recent global estimates that suggest that nearshore areas receiving terrestrial organic carbon input are sources of CO_2 to the atmosphere and that marginal seas are sinks (Borges, 2005; Borges *et al.*, in press). Hence, the net contribution from North American ocean margins is

small and difficult to distinguish from zero. It is not clear how much of the open ocean sink results from photosynthesis driven by nutrients of coastal origin.

Table 15-4. Areas (km^2) and mean annual air-sea CO_2 flux (Mt C yr^{-1}) over four ocean regions surrounding North America. Negative values indicate that the ocean is a CO_2 sink for the atmosphere. Since the observations in the areas north of 60°N in the Chukchi Sea were made only during the summer months, the fluxes from that area are not included. The ± uncertainty is given by one standard deviation of measurements used for analysis and represents primarily the seasonal variability.

TRENDS AND DRIVERS

The sea-to-air CO_2 flux from the coastal zone is small (about 1%) compared with the global ocean uptake flux, which is about 2,000 Mt C y^{-1} (or 2 Gt C yr^{-1}), and hence does not influence the global air-sea CO_2 budget. However, coastal waters undergo large variations in air-sea CO_2 flux on daily to seasonal time scales and on small spatial scales (Figure 15-5). Fluxes can change on the order of 250 g C m^{-2} yr^{-1} or 0.7 g C m^{-2} day^{-1} on a day to day basis (Figure 15-5). These large fluctuations can significantly modulate atmospheric CO_2 concentrations over the adjacent continent and need to be considered when using the distribution of CO_2 in calculations of continental fluxes.

Freshwater bodies have not been treated in this analysis except to note the large surface pCO_2 resulting from estuaries along the east coast. The Great Lakes and rivers also represent net sources of CO_2 as, in the same manner as the estuaries, organic material from the terrestrial environment is oxidized so that respiration exceeds photosynthesis. Interestingly, the effect of fresh water is opposite along the coast of the Pacific northwest, where increased stratification and iron inputs enhance photosynthetic activity (Ware and Thomson, 2005), resulting in a large sink for atmospheric CO_2 (Figure 15-3). A similar process may be at work at the mouth of the Amazon (Körtzinger, 2003). This emphasizes once again the important role of biological processes in controlling the air-sea fluxes of CO_2.

The air-sea fluxes and the underlying carbon cycle processes that determine them (Figure 15-2) vary seasonally, interannually, and on longer time scales. The eastern Pacific, including the U.S. west coast, is subject to changes associated with large-scale climate oscillations such as El Niño (Chavez *et al.*, 1999; Feely *et al.*, 2002; Feely *et al.*, 2006) and the Pacific Decadal Oscillation (PDO) (Chavez *et al.*, 2003; Hare and Mantua, 2000; Takahashi *et al.*, 2003). These climate patterns, and others like the North Atlantic Oscillation (NAO), alter the oceanic CO_2 sink/source conditions directly through seawater temperature changes as well as ecosystem variations that occur via complex physical-biological interactions (Hare and Mantua, 2000; Chavez *et al.*, 2003; Patra *et al.*, 2005). For example, during El Niño, upwelling of high CO_2 waters is dramatically reduced along central California (Figure 15-5) so that

flux out of the ocean is reduced. At the same time photosynthetic uptake of CO_2 is also reduced (Chavez *et al*. 2002) reducing ocean uptake. The net effect of climate variability on air-sea fluxes therefore remains uncertain and depends on the time-space integral of the processes.

OPTIONS AND MEASURES

Two options for ocean carbon sequestration have been considered: (1) deep-sea injection of CO_2 (Brewer, 2003) and (2) ocean iron fertilization (Martin, 1990). The first might be viable in North American coastal waters, although cost and potential biological side effects are unresolved issues. The largest potential for iron fertilization resides in the equatorial Pacific and the Southern Ocean, although it could be considered for the open ocean waters of the Gulf of Alaska and offshore waters of coastal upwelling systems. However, there is still disagreement over how much carbon would be sequestered (Bakker *et al.*, 2001; Boyd *et al.*, 2000; Coale *et al.*, 2004; Gervais *et al.*, 2002) and what the potential side effects would be (Chisholm *et al.*, 2001).

R&D NEEDS VIS A VIS OPTIONS

Waters with highly variable air-sea CO_2 fluxes are located primarily within 100 km of the coast (Figure 15-5). With the exception of a few areas, the available observations are grossly inadequate to resolve the high-frequency, small-spatial-scale variations. These high intensity air-sea CO_2 flux events may introduce errors in continental CO_2 fluxes calculated by atmospheric inversion methods. Achieving a comprehensive understanding of the carbon cycle in waters surrounding the North American continent will require development of advanced technologies, sustained and inter-disciplinary research efforts. Both of these seem to be on the horizon with (1) the advent of ocean observatories that include novel fixed and mobile platforms together with developing instrumentation to measure critical stocks and fluxes and (2) national and international research programs that include the Integrated Ocean Observing System (IOOS) and Ocean Carbon and Climate Change (OC^3). Given the importance of aquatic systems to atmospheric CO_2 concentrations, these developing efforts must be strongly encouraged. Ocean carbon sequestration studies should also be continued.

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Table 15-1. Climatological mean distribution of the net air-sea CO_2 flux (in Gt C yr^{-1}) over the global ocean regions (excluding coastal areas) in reference year 1995. The fluxes are based on about 1.75 million partial pressure measurements for CO_2 in surface ocean waters, excluding the measurements made in the equatorial Pacific (10°N- 10°S) during El Niño periods (see Takahashi *et al.*, 2002). The NCAR/NCEP 42-year mean wind speeds and the (wind speed)^2 dependence for air-sea gas transfer rate are used (Wanninkhof, 1992). Plus signs indicate that the ocean is a source for atmospheric CO_2, and negative signs indicate that ocean is a sink. The ocean uptake has also been estimated on the basis of the following methods: temporal changes in atmospheric oxygen and CO_2 concentrations (Keeling and Garcia, 2002; Bender *et al.*, 2005), $^{13}C/^{12}C$ ratios in sea and air (Battle *et al.*, 2000; Quay *et al.*, 2003), ocean CO_2 inventories (Sabine *et al.*, 2004), and coupled carbon cycle and ocean general circulation models (Sarmiento *et al.*, 2000; Gruber and Sarmiento, 2002). The consensus is that the oceans take up 1.3 to 2.3 Gt C yr^{-1}

Latitude bands	Pacific	Atlantic	Indian	Southern Ocean	Global
N of 50°N	+0.01	–0.31			–0.30
14°N-50°N	–0.49	–0.25	+0.05		–0.69
14°N-14°S	+0.65	+0.13	+0.13		+0.91
14°S-50°S	–0.39	–0.21	–0.52		–1.12
S of 50°S				–0.30	–0.30
Total flux	–0.23	–0.64	–0.34	–0.30	–-1.50
% of flux	15	42	23	20	100
Area (10^6 km^2)	152.0	74.6	53.0	41.1	320.7
% of area	47	23	17	13	100

Table 15-2. Variability of CO_2 distributions and fluxes in U.S. coastal waters from regional surveys and moored measurements (from Doney *et al.*, 2004)

Location	Surface seawater pCO_2 (µatm)	Instantaneous CO_2 flux (mol/m^{-2} yr^{-1})	Annual average (mol m^{-2} yr^{-1})	Sampling method	Reference
New Jersey Coast	211–658	–17 to +12	–0.65	Regional survey	Boehme *et al.* (1998)
Cape Hatteras, North Carolina	ND*	–1.0 to +1.2	ND	Moored meas.	DeGrandpre *et al.* (1997)
Middle Atlantic Bight, inner shelf	150–620	ND	–0.9	Regional survey	DeGrandpre *et al.* (2002)
Middle Atlantic Bight, middle shelf	220–480	ND	–1.6	Regional survey	DeGrandpre *et al.* (2002)
Middle Atlantic Bight, outer shelf	300–430	ND	–0.7	Regional survey	DeGrandpre *et al.* (2002)
Florida Bay, Florida	325–725	ND	ND	Regional survey	Millero *et al.* (2001)
Southern California Coastal Fronts	130–580	ND	ND	Regional survey	Simpson (1985)
Coastal Calif. (M-1; Monterey Bay)	245–550	–8 to +50	1997–98: –1.0 1998–99: +1.1	Moored meas.	Friederich *et al.* (2002)
Oregon Coast	250–640	ND	ND	Regional survey	van Geen *et al.* (2000)
Bering Sea Shelf in spring (April–June)	130–400	–8 to –12	–8	Regional survey	Codispoti *et al.* (1986)
South Atlantic Bight	300–1200	ND	2.5	Regional survey	Cai *et al.* (2003)
Miss. River Plume (summer)	80–800	ND	ND	Regional survey	Cai *et al.* (2003)
Bering Sea (Aug–Sep.)	192–400	ND	ND	Regional survey	Park *et al.* (1974)

* ND = no data available

Table 15-3. Climatological mean annual air-sea CO_2 flux (g C m^{-2} yr^{-1}) over the oceans surrounding North America. Negative values indicate that the ocean is a CO_2 sink for the atmosphere. N is the number of seawater pCO_2 measurements. The ± uncertainty is given by one standard deviation of measurements used for analysis and represents primarily the seasonal variability.

Ocean regions	Coastal boxes		First offshore		Second offshore		Third offshore		Open ocean	
	Flux	N	Flux	N	Flux	N	Flux	N	Flux	N
North Atlantic	3.2± 142	80,417	–1.4± 94	65,148	–7.3± 57	35,499	–10.4± 76.4	15,771	–26± 83	37,667
North Pacific	–0.2± 105	164,838	–6.0± 81	69,856	–4.3± 66	32,045	–5.3± 60	16,174	–1.2± 56	84,376
G. Mexico Caribbean	9.4± 24	75,496	8.4± 23	61,180	11.5± 17.0	8,410	13± 20	1,646		
Bering/ Chukchi	28.0± 110	892	–28± 128	868	–44± 104	3,399	–53± 110	1,465	–63± 130	1,848

Table 15-4. Areas (km^2) and mean annual air-sea CO_2 flux (Mt C yr^{-1}) over four ocean regions surrounding North America. Since the observations in the areas north of 60°N in the Chukchi Sea were made only during the summer months, the fluxes from that area are not included. The ± uncertainty is given by one standard deviation of measurements used for analysis and represents primarily the seasonal variability.

Ocean areas (km^2)					Mean air-sea CO_2 flux (10^{12} grams or Mt C yr^{-1})				
Coastal boxes	**First offshore**	**Second offshore**	**Third offshore**	**Open ocean**	**Coast box**	**First offshore**	**Second offshore**	***Third offshore***	**Open ocean**
North Atlantic coast (8° N to 45°N)									
625,577	651,906	581,652	572,969	3,388,500	2.7±9.5	-0.5±9.3	-4.0±4.9	-6.5±6.3	-41.5±28.1
North Pacific coast (8°N to 55°N)									
1,211,555	855,626	874,766	646,396	7,007,817	2.1±17.1	-7.0±14.1	-4.8±12.5	-3.7±5.3	-53.8±60.7
Gulf of Mexico and Caribbean Sea (8°N to 30°N)									
1,519,335	1,247,413	935,947	1,008,633		13.6±8.9	10.9±7.5	6.8±5.00	6.6±5.0	
Bering and Chukchi Seas (50°N to 70°N)									
481,872	311,243	261,974	117,704	227,609	0.8±3.1	-6.2±9.5	-5.3±7.5	-3.7±3.0	-9.8±3.7
Total ocean areas surrounding North America									
3,838,339	3,066,188	2,654,339	2,300,702	10,623,926	19.1±21.8	-2.8±20.7	-7.4±16.2	-7.3±10.1	-105.2±67.0

1

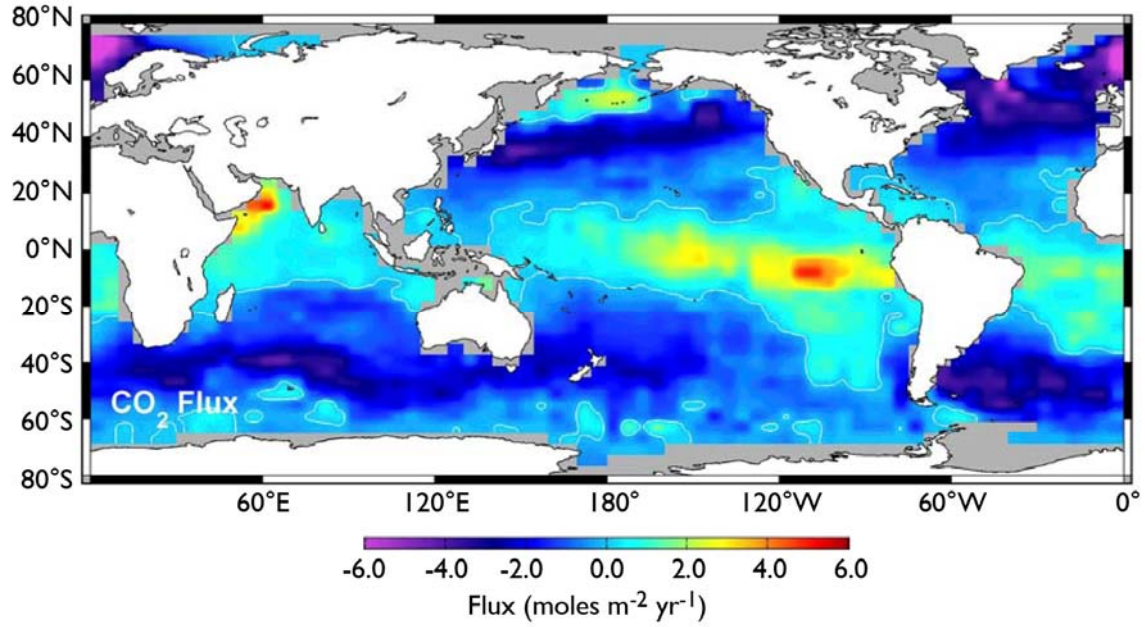


Figure 15-1. Global distribution of air-sea CO_2 flux. The white line represents zero flux and separates sources and sinks. The sources are primarily in the tropics (yellow and red) with a few areas of deep mixing at high latitudes. Updated from Takahashi *et al.* (2002).

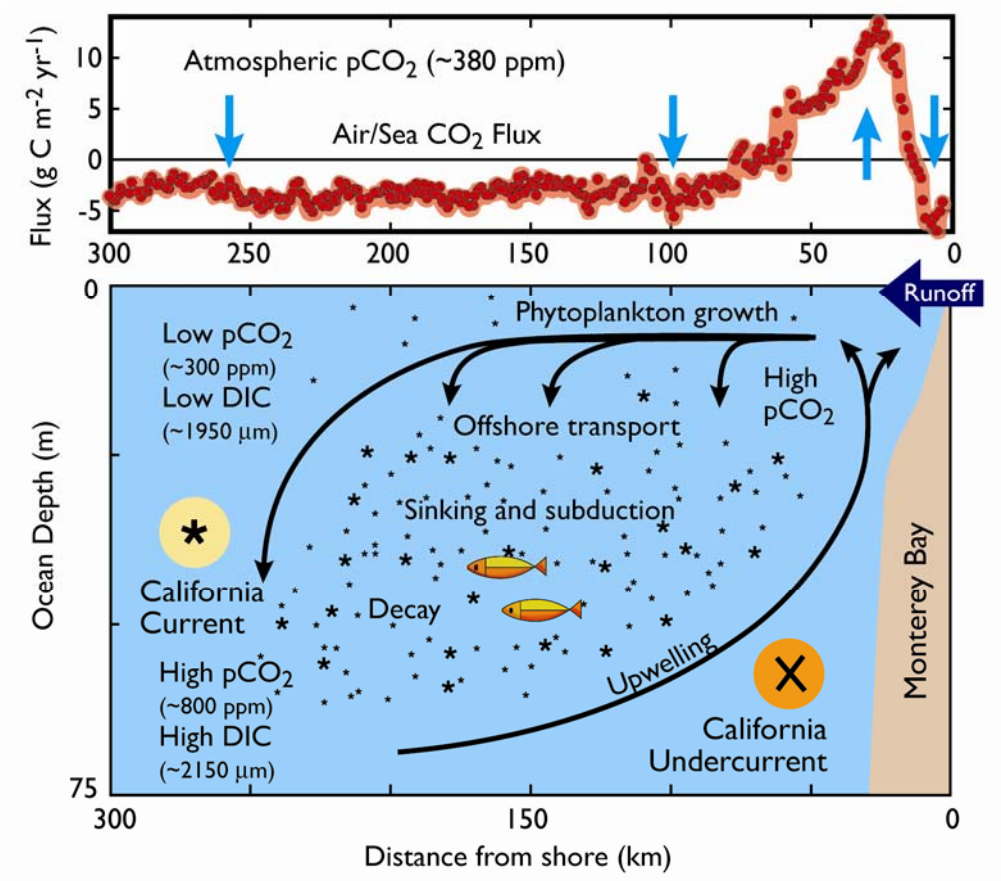
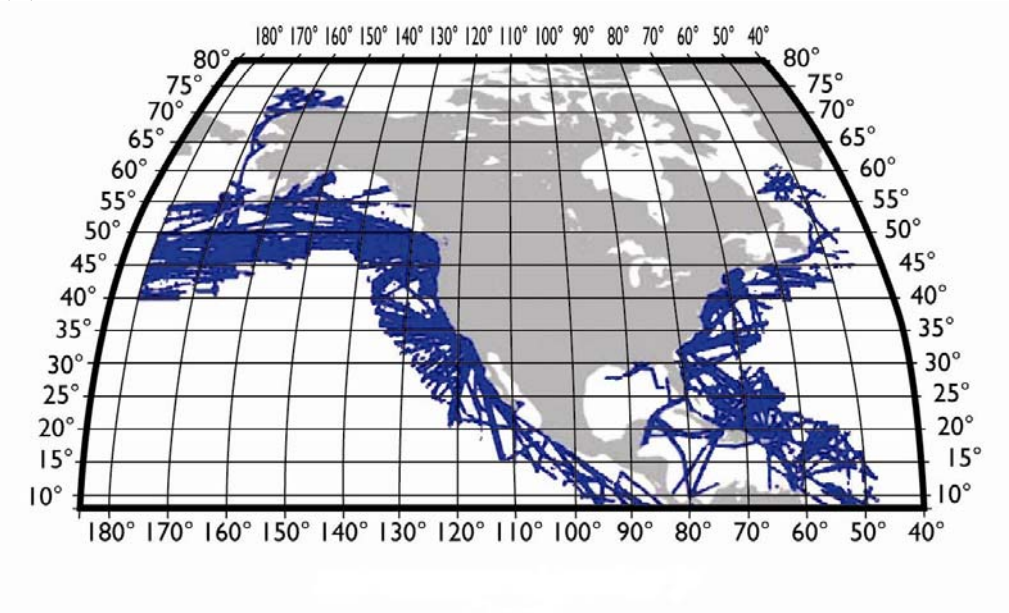


Figure 15-2. In the top panel, mean air-sea CO_2 flux is calculated from shipboard measurements on a line perpendicular to the central California coast. Flux within Monterey Bay (~0–20 km offshore) is into the ocean, flux across the active upwelling region (~20–75 km offshore) is from the ocean, and flux in the California Current (75–300 km) is on average into the ocean. These fluxes result from the processes shown in the bottom panel. California Undercurrent water, which has a high CO_2 partial pressure, upwells near shore, and is advected offshore into the California Current and into Monterey Bay. Phytoplankton growing in the upwelled water use CO_2 as a carbon source, and CO_2 is drawn to low levels in those areas. Phytoplankton carbon eventually sinks or is subducted below the euphotic zone, where it decays, elevating the CO_2 levels of subsurface waters. Where the level of surface CO_2 is higher than the level of atmospheric CO_2, diffusion drives CO_2 into the atmosphere. Conversely, where the level of surface CO_2 is lower than that of atmosphericCO_2, diffusion drives CO_2 into the ocean. The net air-sea flux on this spatial scale is near zero. DIC = dissolved inorganic carbon; POC = particulate organic carbon. Updated from Pennington et al. (in press).

1
2

(A)



(B)

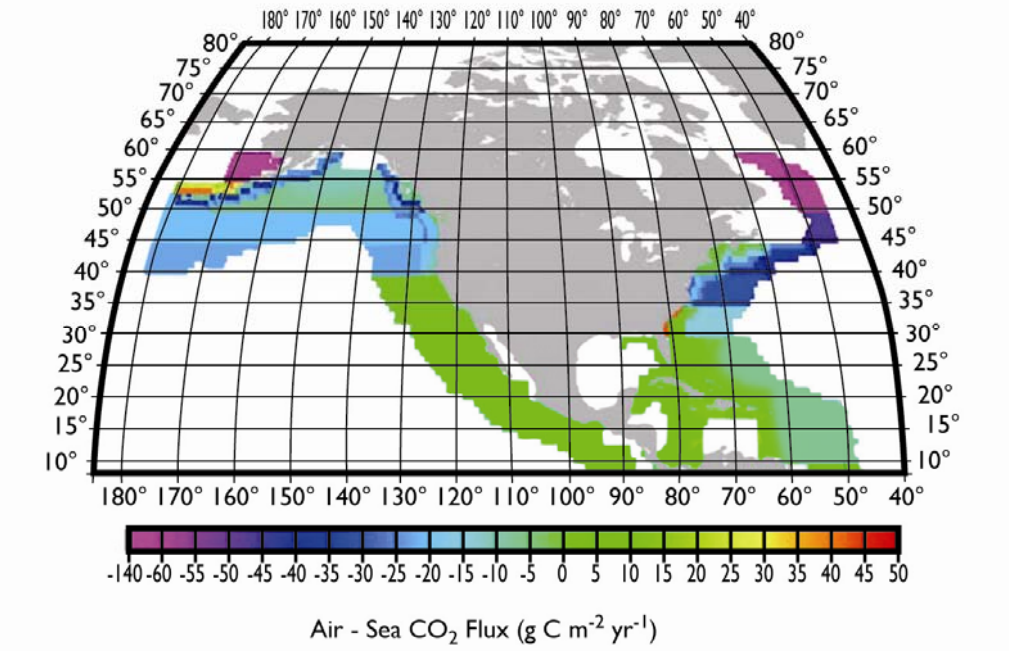
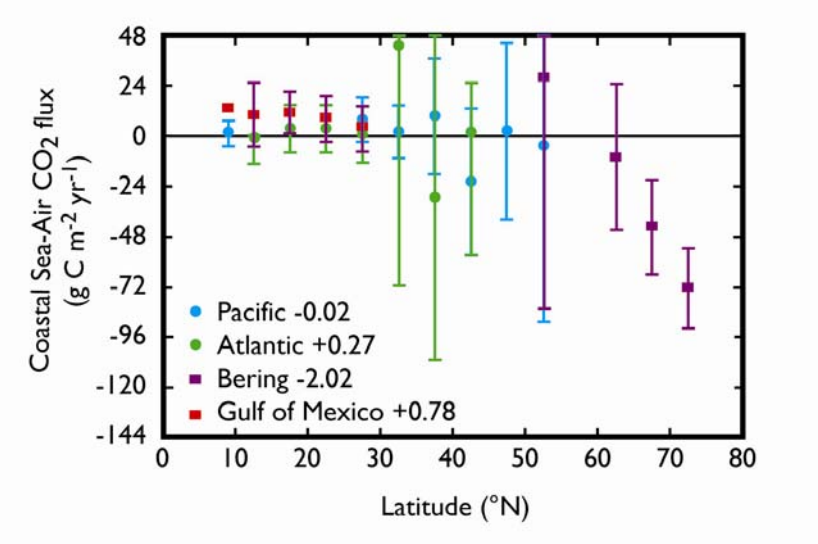


Figure 15-3. (A). Distribution of coastal CO_2 partial pressure measurements made between 1979 and 2004. (B). The distribution of the net air-sea CO_2 flux over 1° × 1° pixel areas (N-S 100 km, E-W 80 km) around North America. The flux (grams of carbon per square meter per year) represents the climatological mean over the 25-year period. The magenta-blue colors indicate that the ocean water is a sink for atmospheric CO_2, and the green-yellow-orange colors indicate that the sea is a CO_2 sink. The data were obtained by the authors and collaborators of this chapter and are archived at the Lamont-Doherty Earth Observatory (www.ldeo.columbia.edu/res/pi/CO2).

1

(A)



(B)

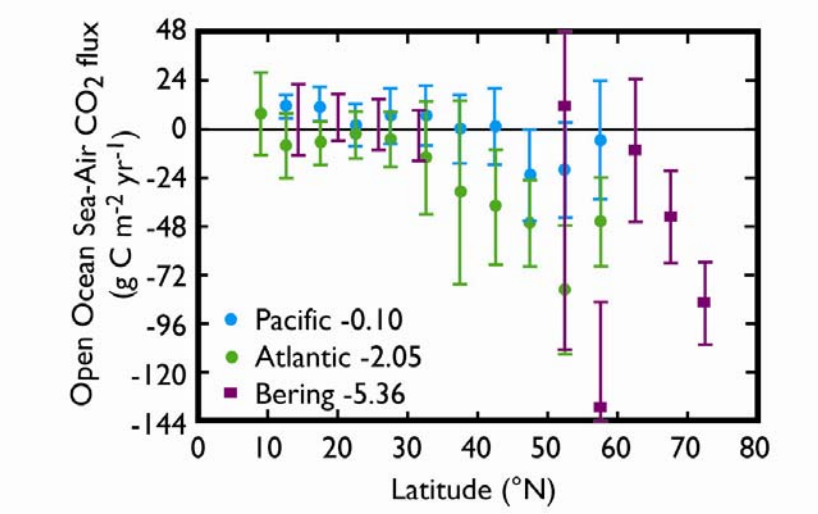
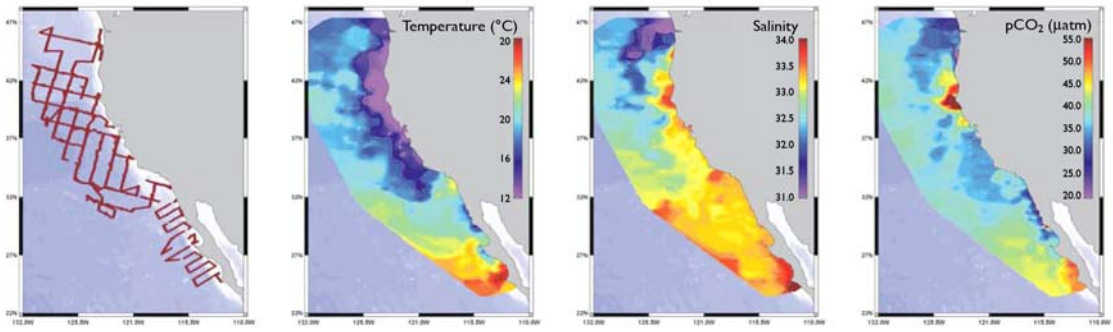


Figure 15-4. Estimated air-sea CO_2 fluxes (grams of carbon per square meter per year) from 550,000 seawater CO_2 partial pressure (pCO_2) observations made from 1979 to 2004 in ocean waters surrounding the North American continent. (A) Waters within one degree (about 80 km) of the coast and (B) open ocean waters between 300 and 900 km from the shore (see Figure 15-3B). The annual mean air-sea pCO_2 difference (delta pCO_2) values were calculated from the weekly mean atmospheric CO_2 concentrations in the GLOBALVIEW-CO2 database (2004) over the same pixel area in the same week and year as the seawater pCO_2 was measured. The monthly net air-sea CO_2 flux was computed from the mean monthly wind speeds in the National Centers for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) database in the (wind speed)^2 formulation for the air-sea gas transfer rate by Wanninkhof (1992). The ± uncertainties represent one standard deviation.

1

(A)



(B)

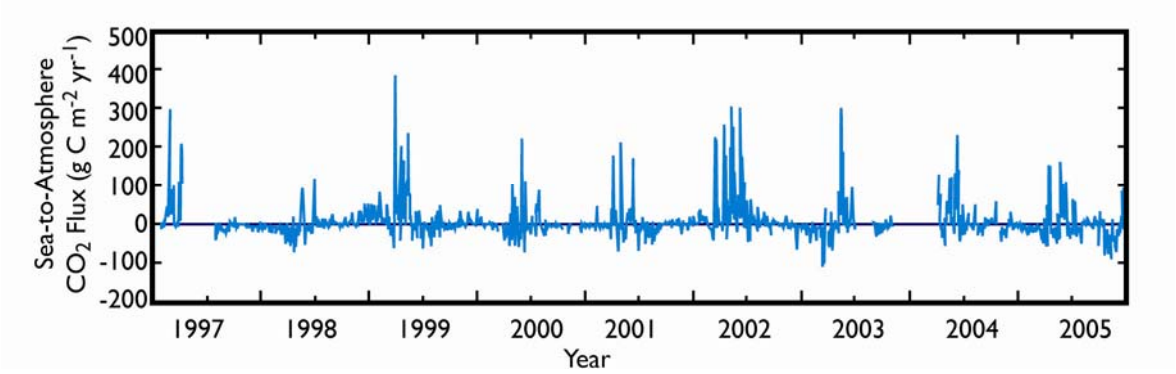


Figure 15-5. Time-space variability of coastal waters off the west coast of North America. (A) Quasi-synoptic distribution of the temperature, salinity, and pCO_2 in surface waters during July–September 2005. The Columbia River plume (~46°N) and the upwelling of deep waters off the Cape Mendocino (~40°N) are clearly seen. (B) 1997–2005 time-series data for air-sea CO_2 flux from a mooring off Monterey Bay, California. Seawater is a CO_2 source for the atmosphere during the summer upwelling events, but biological uptake reduces levels very rapidly. These rapid fluctuations can affect atmospheric CO_2 levels. For example, if CO_2 from the sea is mixed into a static column, a 500-m-thick planetary boundary layer over the course of one day, atmospheric CO_2 concentration would change by 2.5 µatm. If the column of air is mixed vertically through the troposphere to 500 mbar, a change of about 0.5 µatm would occur. The effects would be diluted as the column of air mixes laterally. However, this demonstrates that the large fluctuations of air-sea CO_2 flux observed over coastal waters could affect the concentration of CO_2 significantly enough to affect estimates of air-land flux based on the inversion of atmospheric CO_2 data. Air-sea CO_2 flux was low during the 1997–1998 and 2002–2003 El Niño periods.

2

Appendix 15A
Database and Methods

A database for pCO_2, temperature and salinity in surface waters within about 1,000 km from the shore of the North American continent has been assembled. About 550,000 seawater pCO_2 observations were made from 1979 to 2004 by the authors and collaborators of Chapter 15. The pCO_2 data have been obtained by a method using an infrared gas analyzer or gas-chromatograph for the determination of CO_2 concentrations in a carrier gas equilibrated with seawater at a known temperature and total pressure. The precision of pCO_2 measurements has been estimated to be about ± 0.7% on average. The quality-controlled data are archived at www.ldeo.columbia.edu/res/pi/CO2.

The zonal distribution of the surface water pCO_2, sea surface temperature (SST), and salinity data shows that the greatest variability is confined within 300 km from the shores of both the Atlantic and Pacific. Observations made in various years were combined into a single year and were averaged into 1° × 1° pixels (approximately N-S 100 km by E-W 80 km) for the analysis. Accordingly, the results represent a climatological mean condition over the past 25 years. Finer resolutions (10 × 10 km) may be desirable for some areas close to shore because of outflow of estuarine and river waters and upwelling. However, for this study, which is aimed at a broad picture of waters surrounding the continent, the fine scale measurements have been incorporated into the 1° × 1° pixels. In addition, data with salinities of less than 16.0 are considered to be inland waters and have been excluded from the analysis.

Climatological monthly and annual mean values for pCO_2 in each zone where computed first. Then the air-sea pCO_2 difference, which represents the thermodynamic driving potential for air-sea CO_2 gas transfer, was estimated using the atmospheric CO_2 concentration data. Finally, the net air-sea CO_2 flux was computed using transfer coefficients estimated on the basis of climatological mean monthly wind speeds using the $(\text{wind speed})^2$ formulation of Wanninkhof (1992). The transfer coefficient depends on the state of turbulence above and below the air-sea interface and is commonly parameterized as a function of wind speeds (corrected to 10 m above the sea surface). However, selection of wind data is problematic because wind speeds vary with the time scale (hourly, diurnal, or seasonal). For example, fluxes calculated for the South Atlantic Bight from 6-h mean wind speeds in the NCEP/NCAR version 2 file (1° × 1° mean) were lower than those estimated using the monthly mean. This discrepancy suggests that ships used commonly for coastal carbon studies tend to be small and hence are rarely at sea under high wind conditions, so observations are biased toward lower winds. Taking into account that the observations have been made infrequently over multiple years, the gas transfer coefficients estimated from climatological mean monthly wind speeds may be more representative. The Schmidt number is computed using

measured SST and climatological mean salinity (Da Silva *et al*. 1994). The flux values in a given month are then averaged to yield a climatological mean flux (and standard deviation) for each month. This procedure assumes implicitly that the seawater pCO_2 changes at much slower rates in space and time than the wind speed and that the seawater pCO_2 does not correlate with the wind speed.

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