Importance and Roles of Soil Respiration

Soil respiration is a subject that is of concern not only to ecologists but also to scientists who study atmospheric dynamics and earth system functioning. As an integral part of the ecosystem carbon cycle, soil respiration is related to various components of ecosystem production. Soil respiration is also intimately associated with nutrient processes such as decomposition and mineralization. Moreover, soil respiration plays a critical role in regulating atmospheric CO_2 concentration and climate dynamics in the earth system. Thus, it becomes relevant to the mitigation of climate change and the implementation of international climate treaties in terms of carbon storage and trading. This chapter relates soil respiration to ecosystem carbon balance and production, nutrient cycling, regional and global carbon cycling, climate change, and carbon storage and trading.

2.1. SOIL RESPIRATION AND ECOSYSTEM CARBON BALANCE

The carbon cycle in an ecosystem usually initiates when plants fix CO_2 from the air and convert it to organic carbon compounds through photosynthesis (Fig. 2.1). Some of the organic carbon compounds are used to grow plant tissues. Some are broken down to supply the plants with energy. During this
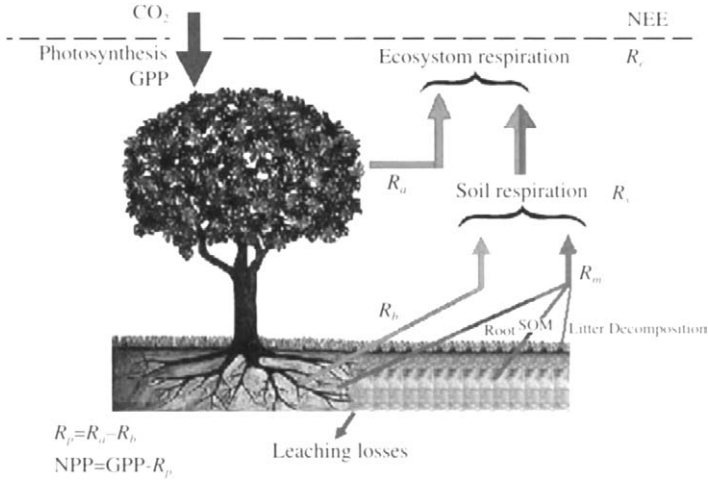
FIGURE 2.1   Schematic diagram of ecosystem carbon processes. Abbreviation see text.

process, CO_2 is released back into the atmosphere through plant respiration. The grown tissues include leaves, stems (e.g., wood for trees), and roots. Leaves and fine roots usually live for several months up to a few years before death, whereas woody tissues may grow for hundreds of years in forests. Dead plant materials (i.e., litter) are decomposed by microorganisms to provide energy for microbial biomass growth and other activities. At the same time, CO_2 is released back into the atmosphere through microbial respiration. The live microbial biomass is mixed with organic residuals of dead plants and dead microbes to form soil organic matter (SOM). SOM can store carbon in soil for hundreds and thousands of years before it is broken down to CO_2 through respiration by microbes.

Through the carbon cycle, CO_2 is produced by both plant respiration (R_p) and microbial respiration (R_m) that occurs during decomposition of litter and SOM. R_p is often called autotrophic respiration and can be separated into aboveground plant respiration (R_a) and belowground plant respiration (R_b). (The belowground plant respiration is often equivalent to root respiration.) Microbial respiration (R_m) during the decomposition of litter and SOM is called heterotrophic respiration. The efflux rate measured at the soil surface (R_s) is the sum of root respiration and microbial respiration:

$$R_s = R_b + R_m \tag{2.1}$$

The CO_2 efflux measured at the soil surface can be considered as soil respiration when CO_2 production and transport are at a steady state (see

Chapter 1). Thus, ecosystem respiration (R_e), the total CO_2 emission from an ecosystem, can be estimated by:

$$R_e = R_a + R_s \qquad (2.2)$$

The relationship of R_s with R_e, as seen in equation 2.2, is well illustrated by data collected from an aspen-dominated mixed hardwood forest in Michigan from 1999 to 2003 (Curtis et al. 2005). On average, over the five years R_s accounts for 71% of R_e, while leaves and aboveground live wood combined (R_a) contribute the rest of R_e (Table 2.1). The relative contribution of R_s to R_e varies considerably in a year. R_s contributes nearly 100% of R_e for most of the winter; the contribution drops to about 60% during the period of fast leaf expansion and then gradually increases during the growing season as soil warms, reaching about 75% at the time of leaf abscission in the autumn (Curtis et al. 2005). Typically, R_s contributes 30–80% of R_e in forests.

Soil respiration is not only an important component of ecosystem respiration but also closely related to ecosystem production such as gross primary production (GPP), net primary production (NPP), and net ecosystem production (NEP). GPP is annual carbon assimilation by photosynthesis ignoring photorespiration. In the Michigan forest, for example, soil respiration is approximately 63% of GPP (Table 2.1). NEP is GPP minus R_e and also related to soil respiration by:

$$NEP = GPP - R_a - R_s \qquad (2.3a)$$

or R_s is related to NEP though NPP, which is GPP minus autotrophic plant respiration, by:

$$\begin{aligned} NEP &= NPP - R_m \\ &= NPP + R_b - R_s \end{aligned} \qquad (2.3b)$$

TABLE 2.1 Various components of ecosystem carbon fluxes in a mixed hardwood forest from 1999 to 2003

Year	R_s	R_e	R_s/R_e	GPP*	R_s/GPP	NPP	NEP
1999	1116	1538	0.73	1637	0.68	656	99
2000	987	1396	0.71	1580	0.62	678	184
2001	1005	1412	0.71	1615	0.62	704	203
2002	946	1404	0.67	1549	0.61	618	145
2003	960	1375	0.70	1545	0.62	650	170
Mean	1003	1425	0.71	1585	0.63	661	160

Note: GPP was estimated by a biometrical approach that sums up different components. The biometrically estimated GPP was higher than that estimated by eddy-flux measurements by nearly 30%. Units are g C m^{-2} yr^{-2}. Modified with permission from New Phytologist: Curtis et al. (2005)

Equation 2.3 is a quantitative basis of the biometrical approach to estimation of net carbon storage in an ecosystem (i.e., *NEP*). *NPP* can be estimated by measuring yearly increments in plant biomass. R_a is often estimated from measured respiration rates of aboveground plant parts (i.e., leaves and live wood in forest). R_b is estimated either from measured respiration rates of roots or indirectly from R_s through partitioning techniques (see Chapter 9). With measured soil respiration, *NEP* can be estimated from Equation 2.3. In the Michigan hardwood forest, the estimated *NEP* by the biometrical method ranged from 100 to $200\,g\,C\,m^{-2}yr^{-1}$ (Table 2.1) (Curtis *et al.* 2005).

Another rate of flux in the ecosystem carbon cycle that can be relatively easily measured, especially in forests, is aboveground litterfall. For a long time scientists have sought a relationship between measured litterfall and soil respiration (e.g., Reiners 1968). By synthesizing experimental results from many forests in different regions with various types and ages of forests, Raich and Naderhoffer (1989) generalized the relationship (Fig. 2.2) as:

$$R_s = aL_a + b \qquad\qquad (2.4)$$

where L_a is aboveground litterfall and *a* and *b* are coefficients. Both R_s and L_a are expressed in units of $g\,C\,m^{-2}yr^{-1}$. The regression coefficient *a* is usually about 3 (Raich and Naderhoffer 1989, Davidson *et al.* 2002a), suggesting that carbon release from soil respiration is nearly three times the carbon input
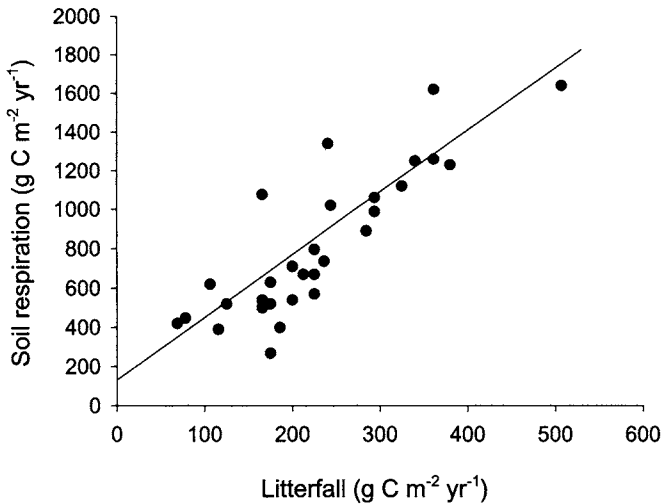

FIGURE 2.2 Correlation of soil respiration with the amount of aboveground litterfall across many forest ecosystems (Redrawn with permission from Ecology: Raich and Naderhoffer 1989).

TABLE 2.2 Annual carbon fluxes for mid-rotation loblolly pine plantations

Component	Control	Irrigated	Fertilized	Fertilized and Irrigated
Soil CO_2 release	1263	1489	1293	1576
Root respiration	663	745	942	1062
Microbial respiration	600	744	351	514
NPP	500	635	1020	1235
NEP	−100	−109	669	721

Note: Units are $gCm^{-2}yr^{-1}$. Modified with permission from Canadian Journal of Forest Research Maier and Kress (2000).

from aboveground litter. Indeed, soil respiration releases carbon from sources of root litter, root exudates, and root respiration in addition to the above-ground litterfall. The correlation was poor, however, among years at a single site (Davidson *et al.* 2002a).

The relationships of R_s with other fluxes can also be used to examine responses of an ecosystem to perturbations. Table 2.2, for example, presents annual carbon fluxes in mid-rotation loblolly pine plantations as affected by fertilization and irrigation (Maier and Kress 2000). Annual R_s is mainly affected by irrigation, ranging from 1263 to $1576gCm^{-2}yr^{-1}$ among the four treatments. Belowground root respiration (R_b) is much more responsive to fertilization than to irrigation, whereas R_m is considerably depressed by fertilization. As a consequence, the relative contribution of R_b to R_s increases from 52% under control to 73% under fertilization. Fertilization substantially increased NPP, resulting in net carbon storage in the forest. NEP is negative by $100gCm^{-2}yr^{-1}$ without fertilization and becomes positive to $700gCm^{-2}yr^{-1}$ with fertilization (Table 2.2).

2.2. SOIL RESPIRATION AND NUTRIENT CYCLING

A major component of soil respiration is from microbial decomposition of litter and SOM that releases CO_2, meanwhile immobilizing or mineralizing nutrients (Coleman *et al.* 2004). During the initial phases of decomposition, nitrogen that is mineralized from litter substrate is simultaneously immobilized by microbes for their own growth, leading to an increased nitrogen concentration in the mixture of litter substrate and microbes. Since the litter substrate and microbes are not easily separated, in practice the mixture is also called litter. The nitrogen concentration of decomposing litter usually increases, while the absolute amount of nitrogen in the litter may or may not

increase during the decomposition. The absolute amount of nitrogen increases when nitrogen from exogenous sources in soil or from fixation is incorporated into microbial biomass growth. The release of carbon combined with nitrogen immobilization during the litter decomposition gradually decreases carbon-nitrogen ratio (C:N) until mineralized nitrogen from litter substrate is greater than required for microbial growth. After that point, litter decomposition leads to a net release of nitrogen. Similarly, phosphorus and sulfur may also increase in absolute amounts during initial phases of decomposition.

Decomposition of SOM usually results in net releases of nitrogen, since C:N of SOM is generally smaller than 20, much closer to C:N of microbes than litter (Paul and Clark 1996). Degradation of proteins and nucleic acids in SOM releases nitrogen in a mineral form (i.e., NH_4^+). The mineralized nitrogen from SOM is partly immobilized for growth of microorganisms and partly added to the mineral nitrogen pool in soil.

Due to the coupled carbon and nitrogen mineralization during microbial decomposition of litter and SOM, the rate of nitrogen mineralization often correlates with microbial respiration. For example, Zak et $al.$ (1993) studied carbon and nitrogen releases from labile organic matter within the forest floor and mineral soil of Jack pine, red pine, balsam fir, sugar maple, and quaking aspen forests in Michigan. Carbon released from microbial decomposition was correlated with mineralized nitrogen (N_{min}) by $R_m = 15.9 N_{min} + 27.4$ with $r = 0.853$ and $n = 154$ for litter and $R_m = 7.1 N_{min} + 159.9$ with $r = 0.616$ and $n = 154$ for SOM from a laboratory incubation. Similar relationships between net carbon and nitrogen mineralization were found in organic substrates with low C:N ratios (Gilmore et $al.$ 1985, Moorhead et $al.$ 1987, Ruess and Seagle 1994, Eriksen and Jensen 2001). Across different types of soils from three communities in an Alaskan boreal forest, rates of soil respiration were associated with rates of microbial turnover and nitrogen mineralization in a laboratory incubation study (Vance and Chapin 2001). In the field research, nitrogen mineralization may not be well correlated with soil respiration due to the nitrogen immobilization.

2.3. SOIL RESPIRATION AND REGIONAL AND GLOBAL CARBON CYCLING

Soil respiration plays a critical role in the regulation of carbon cycling on regional and global scales. The carbon cycle on the global scale involves exchanges of CO_2 among the land biosphere, the atmosphere, oceans, and the earth's crust (Fig. 2.3). Each year, photosynthesis of land plants takes up approximately $120 Pg (10^{15} g) C yr^{-1}$ from the atmosphere. A similar amount of carbon is released back to the atmosphere through ecosystem respiration.
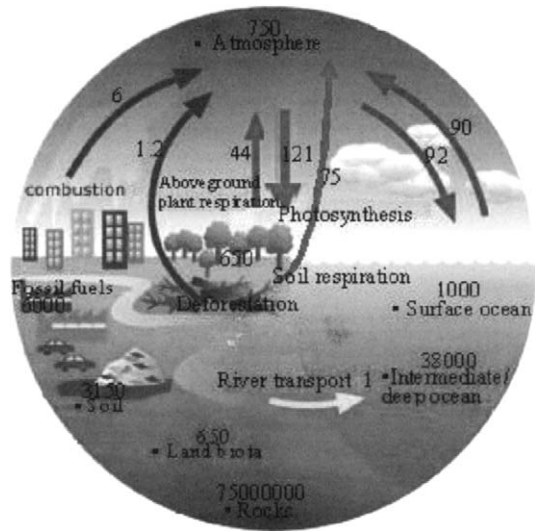
FIGURE 2.3   The global carbon cycle. Pools in Pg (= 10^{15} g) C and fluxes in PgCyr^{-1} as indicated by arrows.

Oceans absorb nearly 92 PgCyr^{-1} from the atmosphere and release 90.6 PgCyr^{-1} back to the atmosphere through physiochemical exchanges of CO_2 at the air-sea surface and through photosynthesis and respiration of marine organisms.

The global soils contain as high as 3150 PgC, including 450 PgC in wetlands, 400 PgC in permanently frozen soils, and 2300 PgC in other ecosystems (Sabine *et al.* 2003). The latter 2300 PgC can be further divided into 1500 PgC in the top soils to the depth of 1 meter and 800 PgC in the deeper soil layers to the depth of 3 meters according to distribution profiles of soil carbon along depths (Jobbágy and Jackson 2000). Plants contain 650 PgC, slightly smaller than the carbon pool size in the atmosphere (750 PgC). The sum of soil and plant carbon contents is 3800 PgC, five times the size of the atmospheric pool.

The burning of fossil fuels by humans presently adds about 6 PgCyr^{-1} to the atmosphere. Land clearing, deforestation, and fire release an additional 1.2 PgCyr^{-1} to the atmosphere. The amount of CO_2 added to the atmosphere by human activities may seem very small in comparison with the rates of fluxes through natural processes such as photosynthesis and respiration. But it takes only a small change to upset the balance of the global carbon cycle. Of the total anthropogenic emission, a little over half remains in the

atmosphere, while the rest is sequestered in land biosphere and the oceans. Modeling and experimental studies suggest that land ecosystems sequester approximately one-third of the anthropogenic emission in plant and soil pools (Schimel *et al.* 2001). As human activities continue to release CO_2, atmospheric CO_2 concentration is expected to keep increasing. Whether the terrestrial carbon sinks are sustainable, however, is highly uncertain.

To understand how the global carbon cycle responds to human perturbation and climate change, we have to understand different aspects of carbon processes, including soil respiration. Soil respiration accounts for a large portion of the total biosphere respiration and is the second largest flux from terrestrial ecosystems. A number of studies have compiled data from field measurements and scaled them up to estimate the global respiratory flux of CO_2 from soils. Schlesinger (1977) estimated global flux at a rate of approximately $75 \, Pg \, C \, yr^{-1}$, roughly 2.5 times larger than the input of fresh litter to the soil surface. Raich and Schlesinger (1992) compiled available data from the literature and estimated global flux to be $68 \, Pg \, C \, yr^{-1}$ from soils. Global soil respiration consists of $50 \, Pg \, C \, yr^{-1}$ from decomposition of litter and SOM, and $18 \, Pg \, C \, yr^{-1}$ from live roots and mycorrhizae. Using a global model, Raich and Potter (1995) updated the estimate of global soil respiration to $77 \, Pg \, C \, yr^{-1}$.

At the global scale, soil respiration releases carbon at a rate that is more than one order of magnitude larger than the anthropogenic emission. The soil pool from which soil respiration releases carbon is about four times the atmospheric pool. Thus, a small change in soil respiration can seriously alter the balance of atmosphere CO_2 concentration. To predict changes in the carbon cycle in response to global change, soil respiration has to be carefully studied.

Soil respiration is very sensitive to environmental changes. The sensitivity of soil respiration to changes in temperature, for example, is a critical parameter in the regulation of the global carbon balance. Results from seasonal measurements usually yield a relationship that the rate of soil respiration increases with temperature (Raich and Schlesinger 1992). In light of this relationship, global warming is expected to stimulate soil respiration and diminish the sink strength of terrestrial ecosystems.

Because of its crucial role in regulating the global carbon cycle, the temperature sensitivity of soil respiration has been extensively studied, using both experimental and modeling approaches. Giardina and Ryan (2000) and Liski *et al.* (1999) found that decomposition of old, recalcitrant SOM or organic carbon in mineral soils is less sensitive to temperature than labile carbon. Luo *et al.* (2001a) conducted a warming experiment in a natural grassland and revealed a phenomenon of acclimation whereby the sensitivity of soil respiration to warming decreases after the ecosystem is exposed to

experimental warming for a certain time. Thus, short-term data may not capture long-term characteristics of respiratory responses to rising temperature. Such results from those and many other studies challenge a common assumption in global models that respiratory carbon release from decomposing organic matter increases with global warming. However, recent soil incubation studies showed that the temperature sensitivity of the decomposition of SOM does not change with soil depth, sampling method, and incubation time (Fang *et al.* 2005). Using a three-pool model, Knörr *et al.* (2005) analyzed soil incubation data and claimed that the temperature sensitivity of slow carbon pools is even higher than that of the faster pools. We need data from well-designed, long-term experiments to resolve the issue of how soil respiration varies with long-term changes in temperature.

The differences in temperature sensitivity of soil respiration nonetheless have global and regional implications. Grace and Rayment (2000) used simple models to illustrate that forest carbon sink diminishes if respiration rises with long-term increases in temperature. When respiration is insensitive to longer-term temperature changes, the forest ecosystems become increasingly effective at sequestering carbon as atmospheric CO_2 continues to increase. Thus, the assumption made about the temperature sensitivity of soil respiration has a profound effect on long-term projections of the global and regional carbon cycles and climate change.

The temperature sensitivity of soil respiration may also be a key factor in determining regional carbon balance. Results from a network of CO_2 flux sites across forests in Europe show that respiration increases, but photosynthesis does not vary along the latitudinal band from Iceland to Italy (Valentini *et al.* 2000). Tropical regions have large pools of SOM with relatively rapid turnover times. Carbon fluxes in the tropical regions are also larger than those in temperate and northern forests. Global warming potentially stimulates great losses of soil carbon in the tropics (Trumbore *et al.* 1996). Boreal forests and tundra have the largest store of labile organic matter and the greatest predicted rise in temperature. Organic carbon accumulated in the soil over previous, colder periods is now decomposing and being released through soil respiration as the soil warms in response to climate change. Thus, understanding soil respiration in different regions is critical in predicting regional and global carbon cycles.

2.4. SOIL RESPIRATION AND CLIMATE CHANGE

Soil respiration becomes relevant to climate change because the CO_2 released from soil respiration is one of the greenhouse gases. The greenhouse gases permit incoming solar radiation to reach the surface of the earth but restrict

the outward flux of infrared radiation. They absorb and reradiate the outgoing infrared radiation, effectively storing some of the heat in the atmosphere. In this way, greenhouse gases trap heat within the atmosphere, resulting in climate warming near the earth's surface.

The increased concentration of greenhouse gases in the atmosphere enhances the absorption and emission of infrared radiation. The atmosphere's opacity increases so that the altitude from which the earth's radiation is effectively emitted into space becomes higher. Because the temperature at higher altitudes is lower, less energy is emitted, causing a positive radiative forcing (IPCC 2001). If the amount of CO_2 is doubled instantaneously, with everything else remaining the same, the outgoing infrared radiation would decrease by about $4\,W\,m^{-2}$. That is, the radiative forcing corresponding to a doubling of the CO_2 concentration is $4\,W\,m^{-2}$. To counteract this imbalance, the temperature of the surface-troposphere system would have to increase by 1.2°C (with an accuracy of ±10%), in the absence of other changes. In reality, complex feedbacks in the climate system (e.g., via clouds and their interactions with radiation) are predicted to amplify the temperature increase to 1.5 to 4.5°C (IPCC 2001).

In addition to feedback loops within the climate system, the atmosphere interacts with the biosphere through climate-carbon cycle loops. The terrestrial ecosystems presently absorb approximately $2\,Pg\,C\,yr^{-1}$, primarily resulting from fertilization effects of rising atmospheric CO_2 concentration and N deposition on plants. As atmospheric CO_2 concentration continues to increase at the "business-as-usual" emission scenario (IS92a) (IPCC 1992), the land biosphere will take up an average of $7.5\,Pg\,C\,yr^{-1}$ by the end of the 21st century without the coupled climate-carbon cycle feedbacks (IPCC 2001).

Rising CO_2 concentration in the atmosphere enhances greenhouse effects, likely resulting in global warming. The global warming could substantially stimulate respiration, resulting in more release of CO_2 to the atmosphere to trap heat. Thus, the climate system and the global carbon cycle form a positive feedback loop to reinforce each other (Friedlingstein *et al.* 2003). Based on temperature sensitivity with a fixed Q_{10} value (e.g., 2.0) across the globe, global warming by 2°C would increase additional carbon release from soil respiration by more than $10\,Pg\,C\,yr^{-1}$, which is larger than the current anthropogenic carbon emission. The additional carbon release aggravates anthropogenic warming.

To examine the positive feedback loop between the climatic system and global carbon cycle, Cox *et al.* (2000) carried out three simulations. The first simulation set the atmospheric CO_2 concentration in the model as in the IS92a scenario without climate warming. The model projects that soils in the land ecosystems absorb a net of nearly $400\,Pg\,C$ from 2000 to 2100 (Fig. 2.4). The second simulation examines climate warming and its effects on the global
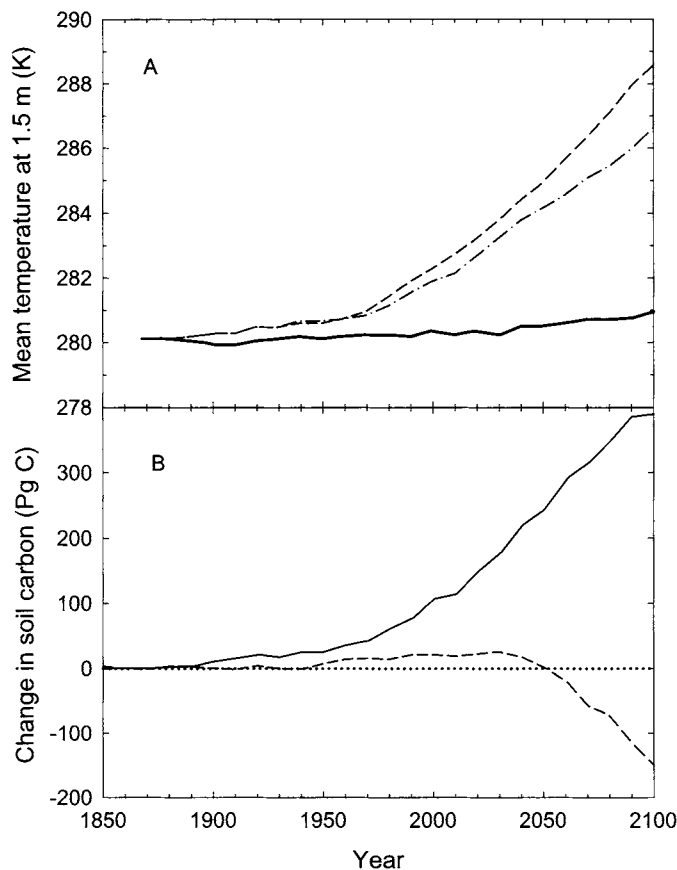
FIGURE 2.4 Simulated mean temperature over land (panel A) and carbon storage in soil (panel B) as affected by rising atmospheric CO_2 concentration, climate warming, or both. The dashed line indicates simulated land surface temperature (panel A) and soil carbon storage (panel B) by the fully coupled carbon cycle climate model, the dot-dashed line in panel A indicates the simulated temperature by a standard global circulation model of climate change with prescribed CO_2 concentration (IS92a) and fixed vegetation, and the solid line indicates simulated temperature (panel A) and carbon storage (panel B) by a model that neglects direct CO_2-induced climate change. The slight warming in the latter is due to CO_2-induced changes in stomatal conductance and vegetation distribution (Redrawn with permission from Nature: Cox et al. 2000).

carbon cycle without the climate-carbon cycle feedback. Rising atmospheric CO_2 concentration, as predefined by the IS92a scenario, induces a 5.5°C warming over land. The climate warming stimulates plant and microbial respiration. The land ecosystems become a source of 60 PgC to the atmosphere over the 21st century. In the third simulation, the climate model is

coupled with the carbon cycle model. The simulation by the coupled model projected an atmospheric CO_2 concentration of 980 ppm in 2100, rather than the 700 ppm as in IS92a. The land ecosystems release 170 Pg C in the simulation, with the coupled model due to stimulated respiration. The global temperature was projected to increase by 8.0°C over land, 2.5°C greater than the simulation of the climate model not coupled to the carbon cycle model. The dramatic increase in global temperature is largely due to stimulated respiration and oxidation of organic matter in warmer soils. Similar positive feedbacks between climate warming and global carbon cycling are demonstrated in simulations by Dufresne et al. (2002). Thus, soil respiration is a critical process that is involved in the positive feedback between climate change and the global carbon cycle. An understanding of responses of soil respiration to global warming is now urgently needed in order to evaluate uncertainty in global climate change projections.

2.5. SOIL RESPIRATION AND CARBON STORAGE AND TRADING

Climate change is not merely a scientific issue but also one of the main challenges facing humanity. To address this challenge, business opportunities have been created for carbon trading in a global market. The market provides incentives for reducing atmospheric CO_2 by those countries seeking to meet their obligations under the framework of the Kyoto Protocol as well as by voluntary national or regional jurisdictions outside the Kyoto Protocol. The Kyoto Protocol, formally known as the United Nations Framework Convention on Climate Change (UNFCCC), was forged in Kyoto, Japan, in December 1997. It has been ratified by most of the world's developed countries and took effect as an international treaty in February 2005. Under the treaty, the participating countries (i.e., the developed and/or market-oriented ones) are legally bound to reduce their greenhouse gas emissions by 2008–2012 to 5% below their levels in 1990 (Sanz et al. 2004).

Global change markets have existed for carbon trading since 2002. The markets traded approximately US$10 million worth of emission allowances in European Union countries in 2002 and will trade as much as US$1 billion per year in allowances by 2010 (Johnson and Heinen 2004). This emerging carbon market is potentially quite substantial (estimated at US$10 billions per year) and introduces a clear financial value for the capture and mitigation of CO_2 emissions in land ecosystems.

Under the Kyoto Protocol, management of natural terrestrial carbon sinks can earn a direct cash award in the carbon mitigation market. The natural sinks reside primarily in expanded forest stocks and increased soil sinks,

which can be managed to increase sink strength and reduce atmospheric CO_2. The emission-trading market provides the opportunity for farmers and foresters to profit by selling emission credits to those parties looking to partially offset their CO_2 reduction obligation. The buying parties may find it less costly to outsource part of their emission mitigation commitment in the natural sinks than to take other measures to reduce emissions. This market-trading practice provides the selling parties with new financial incentives for environmentally friendly land management and forest rehabilitation.

Forests cover about $42 \times 10^{12} m^2$ globally. Forest carbon storage can be achieved in three principal ways: (1) improving the management of currently forested areas, (2) expanding the currently unforested area via afforestation and agroforestry, and (3) reducing the rate of deforestation. All the management options alter the balance between carbon fluxes into the forest ecosystems (i.e., photosynthesis) and fluxes out of the forests via plant and microbial respiration and biomass harvests, resulting in increased carbon stocks in tree biomass, litter mass, soil SOM, and wood products. Potential forest sequestration could approach $1 \, Pg \, C \, yr^{-1}$. But more realistic estimates of achievable sequestration are approximately $0.17 \, Pg \, C \, yr^{-1}$ from improved management of existing forests and $0.2 \, Pg \, C \, yr^{-1}$ from afforestation on formerly wooded and degraded lands (Watson et al. 2000). Financial costs are modest to high (US$3 to $120 per ton of carbon) in so-called Annex 1 countries (i.e., industrialized countries or those that are undergoing the process of transition to a market economy) but often low (US$0.2 to $29 per ton of carbon) elsewhere. Management measures to improve carbon storage in forestry include prolonging rotations, changing tree species, continuous-cover forestry, fire control, combined water storage with peat swamp afforestation, fertilization, thinning regimes, and mixed species rotation. Once management improvements saturate forest carbon sinks, forest ecosystems achieve a steady state, so that any further net carbon storage is unlikely to occur.

Cessation of deforestation is another major method of promoting carbon storage in forest ecosystems. Currently, land use changes result in a net release of $1.2 \, Pg \, C \, yr^{-1}$ (Fig. 2.3). Deforestation, mainly in the tropics, accounts for a large portion of the net release. While complete cessation of deforestation is unrealistic for a variety of social and economic reasons, it offers the single most effective potential solution to mitigate climate change by forest ecosystems. Agroforestry has been widely practiced in the Punjab and India, where crops grow under a canopy of trees. The combinations of trees, crops, and forages in agroforestry may promote carbon sequestration and the sustainable use of other resources.

The other major natural sink in terrestrial ecosystems can be realized mainly through the recapturing of some portion of carbon released from cultivation in world soils. Natural soils retain carbon in stable microaggregates

for up to hundreds and thousands of years unless environmental conditions are changed and the stable soil structure is damaged. Cultivation practices such as plowing break soil aggregates, expose originally protected organic matter in soils to microbial attacks, and thus accelerate decomposition and respiratory carbon losses to the atmosphere. Soils degraded by cultivation are more susceptible to accelerated erosion, which carries carbon to rivers and oceans, where it is partially released into the atmosphere by outgassing (Richey et al. 2002). After conversion of natural to agricultural ecosystems, organic carbon in soils has been depleted by as much as 60% in temperate regions and 75% or more in the tropics. Some soils have lost as much as 2000 to $8000 \, g \, C \, m^{-2}$. Land clearance by humans for agricultural activities began 8000 years ago in Eurasia (McNeill and Winiwarter 2004) and became substantial enough to cause preindustrial CO_2 anomalies in the atmosphere 2000 years ago. Ruddiman (2003) estimated that land conversion during the preindustrial era may cause carbon loss at a rate of $0.04 \, Pg \, C \, yr^{-1}$ for 7800 years and that the total carbon emission from terrestrial ecosystems is $320 \, Pg \, C$, including carbon losses from plant and soil pools. The global cumulative loss of carbon from terrestrial ecosystems is estimated to be 136 to $160 \, Pg \, C$ over the past 200 years. Carbon loss from soils is approximately $78 \, Pg \, C$, including $52 \, Pg \, C$ by soil respiration and $26 \, Pg \, C$ by soil erosion (Lal 2004), with 2.0 (± 1.4) $Pg \, C \, yr^{-1}$ in the 1980s and 1990s alone (Houghton 2002). In comparison, carbon emission from fossil fuel combustion was $270 \, Pg \, C$ between 1850 and 1998 and approximately $5 \, Pg \, C$ in the 1990s. Land use change transformed land covers of temperate regions before about 1950 to the tropics in recent decades (Achard et al. 2002, DeFries et al. 2002, Houghton 2003), resulting in substantial CO_2 effluxes from soils in every continent except Antarctica (DeFries et al. 1999).

The potential carbon sink capacity in soils through ecosystem management approximately equals the cumulative historical carbon loss. The attainable soil sink is 50 to 66% of the potential capacity. The optimistic rate of soil carbon sequestration is at 0.6 to $1.2 \, Pg \, C \, yr^{-1}$ (Lal 2003) and a more likely rate at 0.3 to $0.5 \, Pg \, C \, yr^{-1}$ (Sauerbeck 2001). Carbon sequestration at the optimistic rate would restore most of the lost carbon within 50 to 100 years. Thus, carbon sequestration in soils potentially offsets fossil fuel emissions by 0.4 to $1.2 \, Pg \, C \, yr^{-1}$, or 5 to 15% of the global fossil fuel emissions.

Based on the principles of increasing plant carbon inputs, slowing soil carbon decomposition rates, or both, soil carbon can be built through a variety of agronomic management techniques (Fig. 2.5). Carbon inputs can be enhanced by growing higher biomass crops, by leaving more crop biomass to decompose in situ, by increasing belowground NPP, and by growing cover crops during portions of the year. Decomposition rates can be slowed by reducing tillage and by growing crops with low residue quality. No tillage
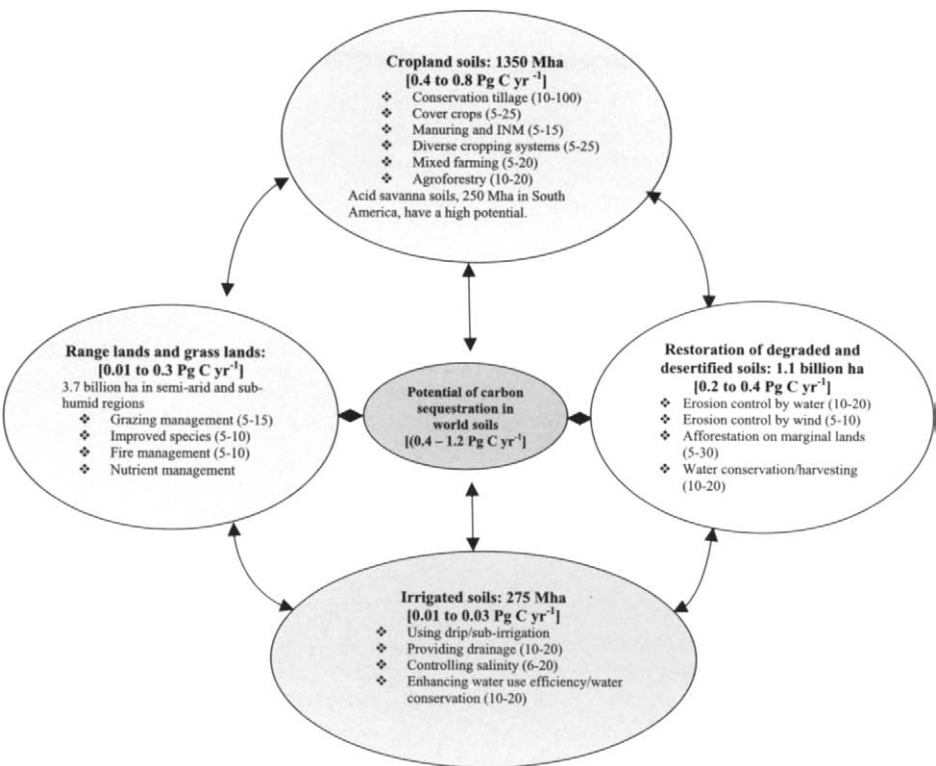
FIGURE 2.5 Soil C sequestration potential in cropland, grazing/range land, degraded/deserti-fied lands, and irrigated soils. Rates of C sequestration given in parentheses are in $gC m^{-2} yr^{-1}$. These are not additive and low under on-farm conditions (Redrawn with permission from Nature: Lal 2004 with references to original papers for the listed rates).

implants seeds without turning the soil with a plow and reduces the loss of SOM. The low-quality residue contains organic carbon that is more difficult for microbes to decompose. Thus, soil restoration and woodland regeneration, no-till farming, cover crops, nutrient management, manuring and sludge application, improved grazing, water conservation and harvesting, efficient irrigation, agroforestry practices, and growing energy crops on spare lands are recommended management practices (RMPs) to increase the soil carbon sequestration (Silver *et al.* 2000, Nordt *et al.* 2001, West and Marland 2002, Lal 2004). Those management practices add high amounts of biomass to the soil, cause minimal soil disturbance, conserve soil and water, improve soil structure, enhance activity and species diversity of soil fauna, and strengthen mechanisms of elemental cycling (Fig. 2.5).
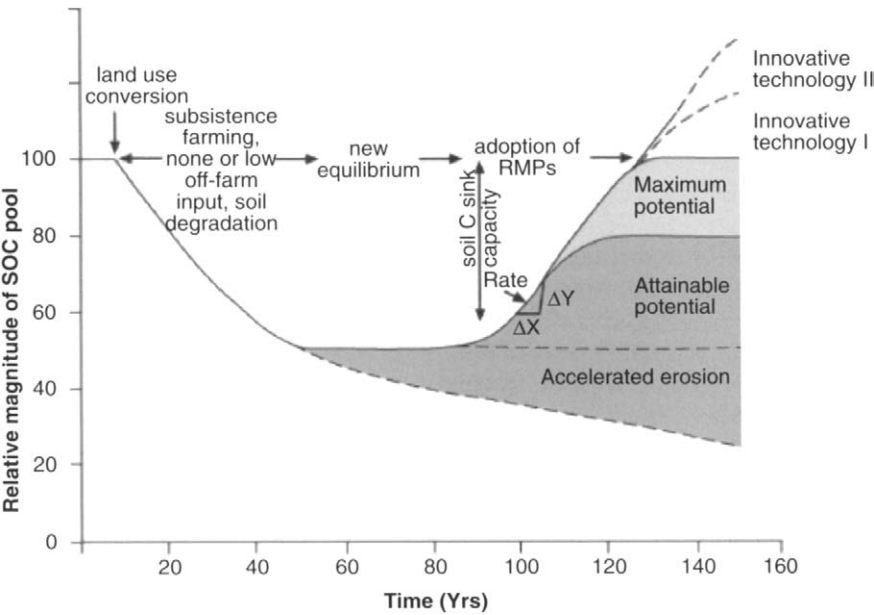
FIGURE 2.6 Schematic illustration of soil carbon dynamics after conversion from natural to agricultural ecosystems and subsequent recovery using recommended management practices (RMPs). The maximum potential equals the magnitude of historical carbon loss (adapted with permission from supplemental material of Lal 2004).

The capacity of soil carbon sequestration varies with time (Lal 2004). The rate of soil carbon sequestration through land managements usually follows a gradual decline. It reaches a maximum in the first 5 to 20 years after land conservation and continues until SOM attains a new equilibrium (Fig. 2.6). The rates of soil carbon sequestration in agricultural and restored ecosystems range from 0 to $15\,g\,C\,m^{-2}\,yr^{-1}$ in dry and warm regions (Armstrong *et al.* 2003) and 10 to $100\,g\,C\,m^{-2}\,yr^{-1}$ in humid and cool climates (West and Post 2002). These rates may continue for 20 to 50 years with the continuous uses of recommended management practices and then decline as the soil carbon content reaches a steady state. The global carbon-trading markets can be a major incentive in promoting the management practices that increase carbon storage in soils. To implement the carbon-trading markets, on the other hand, we have to develop the ability to measure photosynthesis, respiration, and short-term (three- to five-year) changes in SOM pool for verification of carbon credits.