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RESEARCH ARTICLE

Soil carbon stocks under present and future climate with specific reference to European ecoregions

Rattan Lal

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Abstract World soils and terrestrial ecosystems have been a source of atmospheric abundance of CO_2 ever since settled agriculture began about 10–13 millennia ago. The amount of CO_2-C emitted into the atmosphere is estimated at 136 ± 55 Pg from terrestrial ecosystems, of which emission from world soils is estimated at 78 ± 12 Pg. Conversion of natural to agricultural ecosystems decreases soil organic carbon (SOC) pool by 30–50% over 50–100 years in temperate regions, and 50–75% over 20–50 years in tropical climates. The projected global warming, with estimated increase in mean annual temperature of 4–6°C by 2100, may have a profound impact on the total soil C pool and its dynamics. The SOC pool may increase due to increase in biomass production and accretion into the soil due to the so-called "CO_2 fertilization effect", which may also enhance production of the root biomass. Increase in weathering of silicates due to increase in temperature, and that of the formation of secondary carbonates due to increase in partial pressure of CO_2 in soil air may also increase the total C pool. In contrast, however, SOC pool may decrease because of: (i) increase in rate of respiration and mineralization, (ii) increase in losses by soil erosion, and (iii) decrease in protective effects of stable aggregates which encapsulate organic matter. Furthermore, the relative increase in temperature projected to be more in arctic and boreal regions, will render Cryosols under permafrost from a net sink to a net source of CO_2 if and when permafrost thaws. Thus, SOC pool of world soils may decrease with increase in mean global temperature. In contrast, the biotic pool may increase primarily because of the CO_2 fertilization effect. The magnitude of CO_2 fertilization effect may be constrained by lack of essential nutrients (e.g., N, P) and water. The potential of SOC sequestration in agricultural soils of Europe is 70–190 Tg C yr^{-1}. This potential is realizable through adoption of recommended land use and management, and restoration of degraded soils and ecosystems including wetlands.

Keywords Global warming · Greenhouse effect · Soil carbon dynamics · Soil carbon sequestration · CO_2 fertilization effect

R. Lal (✉)
Carbon Management and Sequestration Center,
The Ohio State University, Columbus, OH 43210, USA
e-mail: LAL.1@osu.edu

Introduction

There is an urgent need to identify strategies of stabilizing atmospheric concentration of CO_2, responsible for 62% of the radiative forcing of Earth by long-lived greenhouse gases (GHGs), at less than doubling of the pre-industrial concentration of 280 ppm. For about 10,000 years before 1750, CO_2 concentration was about 280 ppm. Since the late 1700s, the CO_2 abundance has increased progressively and reached 377 ppm in 2004 with an overall increase of about 35% (WMO 2006). The CO_2 concentration is currently increasing at the rate of 1.9 ppm yr^{-1} or 0.47% yr^{-1} (WMO 2006). This increase is attributed to two principal sources: (i) land use conversion and deforestation, and (ii) fossil fuel combustion. The impact of land use conversion and agricultural activities on CO_2 abundance began with the onset of settled agriculture about 10,000 years ago (Ruddiman 2003).

CO_2 emissions due to land use conversion and deforestation intensified with the clearance of Northern Hemisphere forests in the 19th century. Emissions were exacerbated by deforestation of topical rainforests (TRF) during the 20th century. It is estimated that 350 Mha of TRF were deforested and another 500 Mha of secondary and primary tropical forests were degraded (Lamb et al. 2005) with substantial CO_2 emission to the atmosphere. Rapid expansion of agriculture during the 20th century, to meet the food demands of increase in world's population, also accentuated the release of CH_4 from rice paddies and livestock, and N_2O from fertilized croplands. Consequently, abundance of CH_4 increased from a pre-industrial level of 700 ppb to 1783 ppb in 2004, and is currently increasing at the rate of about 5 ppb yr^{-1} or 0.28% yr^{-1}. N_2O abundance increased from a pre-industrial level of 270 ppb to 319 ppb in 2004, and is currently increasing at the rate of 0.8 ppb yr^{-1} or 0.22% yr^{-1} (WMO 2006). Deforestation and land use conversion presently contribute 0.6 to 2.5 Pg C yr^{-1}. In contrast, fossil fuel combustion emits about 7 Pg C yr^{-1} (WMO 2006).

Some researchers and planners argue (e.g., IPCC 2000; Lal 2004) that land use and soil management technologies are feasible options of reducing the net rate of increase of CO_2 abundance. For example, Cox et al. (2000) observed that the biosphere will act as an overall C sink until about 2050, and will become a source thereafter when ocean will become a bigger sink at about 5 Pg C yr^{-1}. Pacala and Socolow (2004) proposed 16 technological interventions to stabilize CO_2 abundance over the 50-year period between 2004 and 2054, each with a CO_2-C sink capacity of 1 Pg C yr^{-1}. Three of the 16 options include: (i) producing biomass feedstock for fossil fuel based on establishing biofuel plantations on 250 Mha to produce ethanol from lingo-cellulosic feedstock, (ii) reducing tropical deforestation to zero, and establishing 300 Mha of new tree plantations, and (iii) converting 1500 Mha of cropland soils from plow tillage to no-till farming. Despite the promise of stabilizing CO_2 concentration at less than doubling of the pre-industrial loads, there remain numerous uncertainties in biotic strategies of C sequestration. Uncertainties are due to the complexity of the climate system (Lump 2002) and numerous feedback mechanisms (Cox et al. 2000).

The objective of this manuscript is to discuss processes, causes, and factors affecting the SOC dynamics in natural and managed ecosystems at present and with projected increase in mean global temperature. A comprehensive review of literature is out of the scope of this manuscript. Therefore, the focus is on citing specific examples from the Scandinavian and northern European regions with reference to specific processes, causes or factors moderating the SOC dynamics. The rational for the choice of these regions is the need for strengthening the data base and a comparatively larger impact of climate change on soils and ecosystems of the northern latitudes.

Carbon sequestration in terrestrial ecosystems

Carbon sequestration implies transfer of atmospheric CO_2 into long-lived C pool (e.g., geologic, oceanic, biotic, pedologic). Technological options of C sequestration are based on engineering principles and natural processes (Fig. 1). Engineering principles involve capture, purification, transport and injection of CO_2 into geological formations or deep ocean. Engineering techniques of geologic sequestration have a large potential of thousands of Pg of C (Dooley et al. 2006). However, presently these techniques are a work in progress. Furthermore, these are costly measures affected by numerous uncertainties due to leakage and the necessities of measurement and monitoring. The oceanic uptake is increasing naturally due to the increase in CO_2 abundance (Feely et al. 2004). Sabine et al. (2004) estimated the natural oceanic sink of 118 ± 19 Pg C between 1880 and 1994, and the sink capacity is likely to increase to 5 Pg C yr^{-1}. However, injection

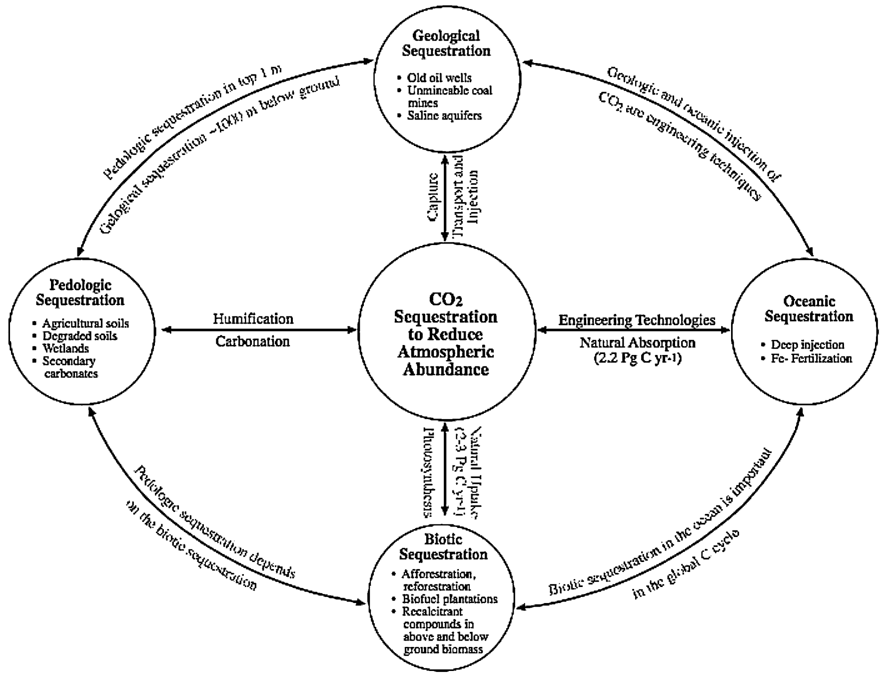

Fig. 1 Technological options of sequestering atmospheric CO_2 into long-lived pools. In addition, chemical transformation of CO_2 into stable minerals (e.g., $CaCO_3$) is another feasible option

of CO_2 deep under ocean surface has similar economic and ecologic concerns as does the geologic sequestration.

In contrast to geologic and oceanic techniques, the terrestrial C sequestration in biotic and pedologic pools is based on natural processes of photosynthesis, humification, illuviation, formation of char, and creation of secondary carbonates (Fig. 2). Four principal natural processes of terrestrial C sequestration are briefly outlined below:

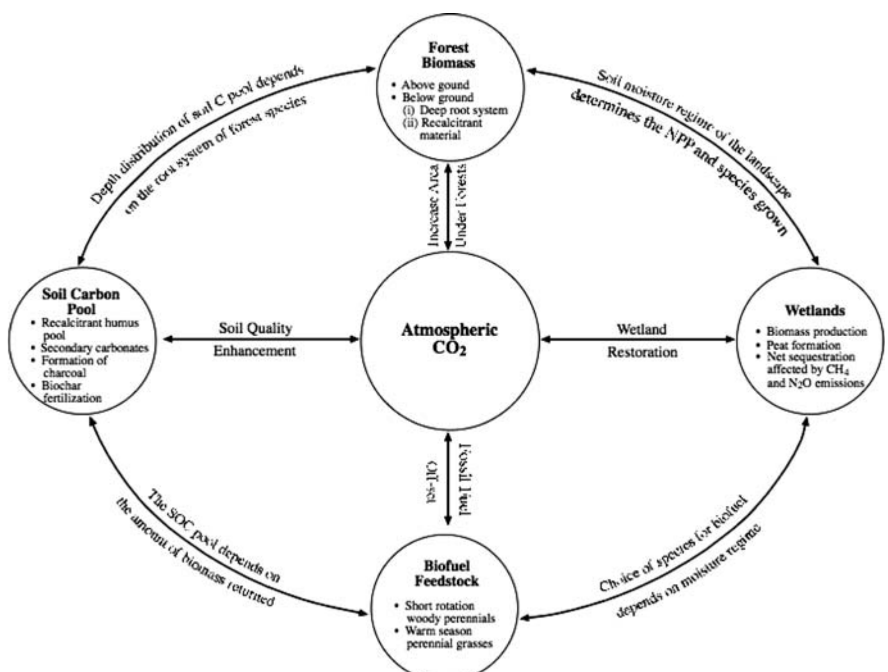

Fig. 2 Strategies of terrestrial carbon sequestration in biota and soil. Biofuel plantations generate renewable energy and sequester C in soil

Photosynthesis

The principal reaction that converts 120 Pg C yr^{-1} from atmosphere into biomass is given in Eq. (1) (Gislason 2005):

$$106CO_2 + 16NO_3^- + HPO_4^{2-} + 122H_2O + 18H^+ = C_{106}H_{263}O_{110}N_{16}P + 138O_2 \quad (1)$$

Of the 120 Pg CO_2-C absorbed by this reaction, 60 Pg is returned back to the atmosphere through plant respiration and decomposition of soil organic matter (SOM) or soil respiration. On a geologic time scale, this process has been extremely important in converting atmospheric CO_2into fossil fuel (coal, oil, gas) and peat soils. With innovative intervention, involving biotechnology and genetic manipulations that may enhance lignin: N ratio or below ground: above ground biomass allocation with placement of C deep into the sub-soil through a tap root system, it may be possible to increase the residence time of C within the terrestrial ecosystems. If 6% of the total photosynthate can be retained in the terrestrial ecosystem, it is enough to offset the fossil fuel emissions of 7 Pg C yr^{-1} in 2005.

Humification

It is the process of conversion of labile biomass C returned to the soil into a relatively recalcitrant SOM pool. The decomposing products by microbial processes include soluble organic compounds which constitute dissolved organic carbon (DOC), and relatively amorphous organic compounds such as humic acids, fats, waxes, oils, lignin and polyuronoides. The recalcitrant humus faction has a long residence time of thousands of years.

Aggregation

Formation of organo-mineral complexes through polyvalent cations is another process of stabilization of SOM (Edwards and Bremner 1967). Biomass C encapsulated within stable micro-aggregates is protected against microbial attack. A micro-aggregate is schematically represented by Eq. 2.

$$\underset{\text{(Domain)}}{(\text{Clay–P–OM})} \underset{\text{dispersion}}{\overset{\text{Aggregation}}{\rightleftharpoons}} \underset{\text{(Micro-aggregate)}}{y\,(\text{Clay–P–OM})_x} \underset{\text{dispersion}}{\overset{\text{Aggregation}}{\rightleftharpoons}} \underset{\text{(Macro-aggregate)}}{[(\text{Clay–P–OM})_x]_y}. \quad (2)$$

In the schematic presented in Eq. 2, clay particles are attached to long chain organic molecules (OM) by polyvalent cations (P), most important among which are Ca^{+2}, Mg^{+2}, Fe^{+3}, and Al^{+3}. Sesquioxides (Fe_2O_3, Al_2O_3, Mn_2O_3) are extremely important to formation of stable micro-aggregates in soils of the tropics.

Formation of secondary carbonates

There are two types of carbonates in soil: (i) lithogenic carbonates which are formed from the weathering of parent materials, and, thus, are primary carbonates, and (ii) pedogenic carbonates which are formed from dissolution of CO_2 in soil air to form carbonic acid and its reaction with Ca^{+2} or Mg^{+2} to form carbonates which precipitate on gravels or pebbles. Pedogenic carbonates are also called secondary carbonates , and their formation in soil leads to CO_2 sequestration (Eq. 3).

$$CO_2 + H_2O \rightleftharpoons H_2CO_3 \rightleftharpoons H_2CO_3 + Ca\ CO_3 \rightleftharpoons Ca^{+2} + 2HCO_3^- \quad (3)$$

An increase in CO_2 concentration of soil air or decrease in pH increases formation of secondary carbonates. Dissolution of CO_2 is also favored by increase in soil moisture content. The rate of formation of secondary carbonates is low and ranges from 5 to 10 Kg C ha^{-1} yr^{-1} (Lal 2004).

Weathering of silicates

Weathering of Ca–Mg silicates (e.g., plagialese, olivine, pyroxene, volcanic glass) to form clays (e.g., kaolinite, halloysite, imogolite) also causes net removal of CO_2 from the atmosphere (Gislason 2005) as shown in Eq. 4:

$$CaAl_2Si_2O_8 + 2CO_2 + 3H_2O \rightarrow Al_2Si_2O_5(OH)_4 + Ca^{++} + 2HCO_3^- \rightarrow CaCO_3 + CO_2 + H_2O \quad (4)$$

Cations (Ca^{+2} and Mg^{+2}) released are transported from the weathering site by runoff into oceans where they eventually precipitate as carbonates. One mole of CO_2 is sequestered for each mole of Ca^{+2} or Mg^{+2} released. However, the rate of CO_2 sequestration by this process is slow and significant only on a geological time scale.

Deep placement of biomass carbon

Transfer of biomass C into the sub-soil horizons can increase its residence time in the soil, because sub-soil C is farther away from the zone disturbed by agricultural activities, impacted by climatic elements, and transported by erosional processes. Deep placement of SOC, below 1 m depth, can increase the residence time (Lorenz and Lal 2005). It is estimated that 5% of the previously "unaccounted C" in the global C budget may be stored in mineral sub-soil under boreal mires.

Leaching of biocarbonates

Movement of HCO_3^- into ground water or closed systems isolated from the ambient environment is another mechanism of C sequestration within soil. Restoration of saline soils (through application of gypsum, green manure and other biosolids) increases leaching of $Ca(HCO_3)_2$ with irrigation water, especially if the latter is not saturated with biocarbonates.

Biochar

Production of black carbon or biochar is another technique of enhancing C storage in soil. High fertility of the "terra preta" or "black land" in the Amazon is attributed to enrichment of soil by biochar (Marris 2006). Increase in C pool in soils managed by slash and burn agriculture is also attributed to high proportion of charcoal (Rumpel et al. 2006a, b). Biochar fertilizers are also being considered an option to improve soil fertility while enhance C pool in soil (Woods et al. 2006; Hayes 2006).

Technological options of carbon sequestration in terrestrial ecosystems

There is a close link between the atmospheric and the terrestrial C pools, both of which are linked through the process of photosynthesis and respiration/oxidation. Photosynthesis (Eq. 1) annually transfers 120 Pg of C into the terrestrial pool. Prior to settled agriculture about ten millennia ago, terrestrial ecosystems comprising biota and soil were sink for atmospheric CO_2. However, soil and vegetation have been sources of CO_2 since the dawn of settled agriculture (Ruddiman 2003, 2005). Thus, the objective of ecosystem management is to restore the depleted C pool of soil and biota through adoption of recommended management practices (RMPs).

Total sink capacity and the rate of C accumulation in soil and biota are determined by climate, soil properties, landscape position and the natural vegetation cover determined by the interaction among these factors. The maximum C pool, in soil and biota as determined by the ecosystem characteristics, occurs under natural vegetation cover. The soil organic carbon (SOC) pool is prone to depletion upon conversion of natural to agricultural ecosystems. The rate of depletion is exacerbated by onset of degradative processes such as soil erosion. Therefore, the maximum sink capacity of a soil equals the amount of SOC pool depleted by the historic land use and severity of degradation (Fig. 3). The magnitude of SOC depletion may be 30% to 50% over 50–100 years after conversion to an agricultural land use in a temperate climate, and 50–75% over 10–20 years in a tropical climate (Fig. 3). Both rate and magnitude of SOC depletion are exacerbated by soil erosion. The process can be reversed through adoption of a restorative land use and recommended soil and vegetation management practices which can restore the depleted SOC pool. The rate of SOC sequestration follows a sigmoid curve (Fig. 3), and is represented by the slope of the curve ($\Delta y/\Delta x$). The attainable SOC pool, the new equilibrium level achieved under the RMPs, is generally lower than the antecedent pool under natural conditions. In general, the attainable SOC pool is about 2/3rd of the historic loss of SOC Pool. Reactions 1, 2 and 3 outlined above can be managed through soil and vegetation management to enhance C sequestration in terrestrial ecosystems. Technological options to enhance C sequestration are outlined in Fig. 4.

Potential rates of SOC sequestration with adoption of RMPs are outlined in Table 1. In general, rates of SOC sequestration are more in cool than warm climates, soils with severely than slightly depleted SOC pool, poorly than well-drained soils, and in light than heavy-textured

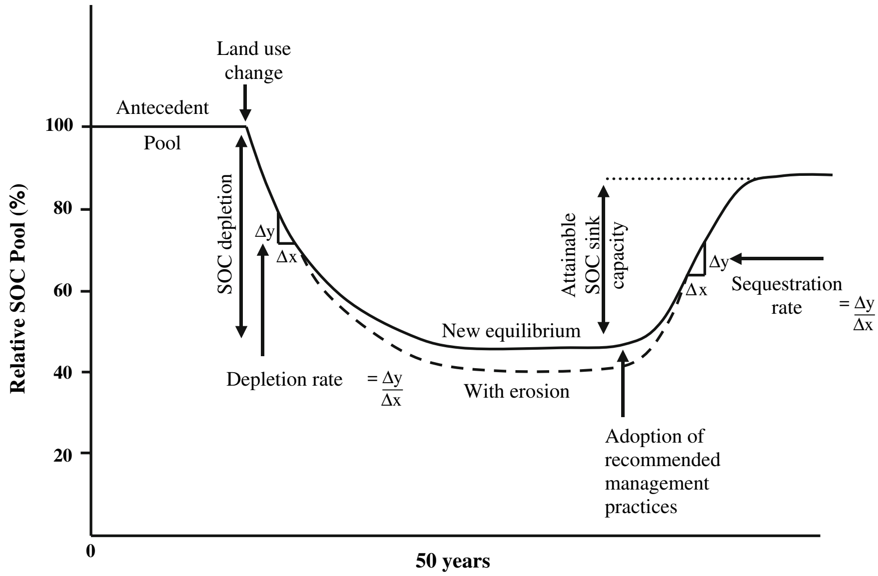

Fig. 3 SOC depletion by land use conversion from natural to agricultural ecosystem. The attainable pool is generally lower than the antecedent pool. The rate of SOC sequestration depends on soil type, landuse, climate and management

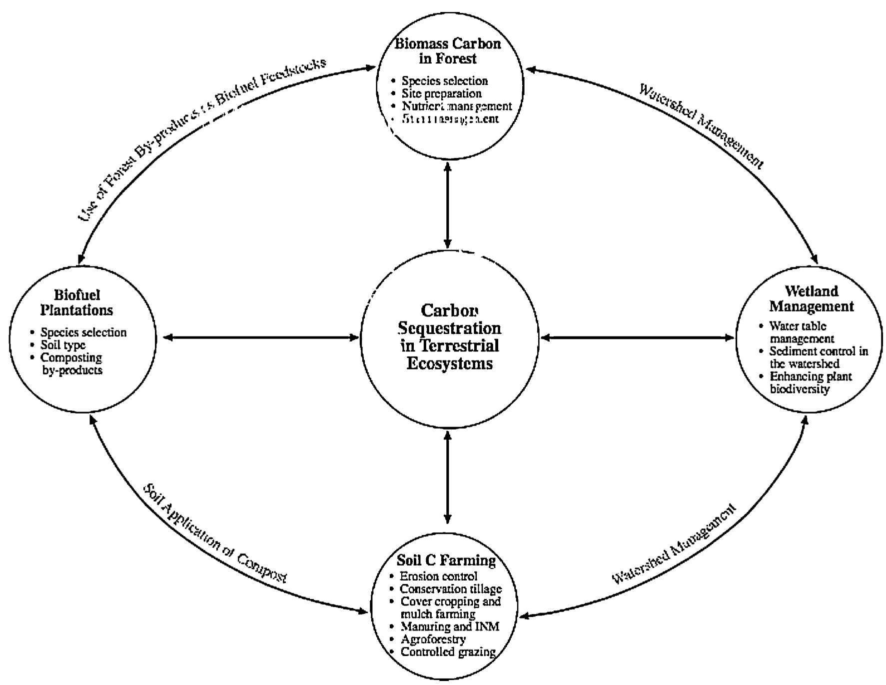

Fig. 4 Technological options for carbon sequestration in terrestrial ecosystem. Soil C farming can be promoted through trading of C credits and generating another income stream for farmers

soils. Clay content has a strong impact on SOC pool and its dynamics (Vejre et al. 2003).

Land use, soil degradation and historic carbon loss from soils of Europe

The total land area of Europe is 2.26 billion hectares (Bha), of which 291 million hectares (Mha) is arable, 17 Mha is under permanent crops, 183 Mha is under pasture and 158 Mha is under forest and woodland (Table 2). Because of a mild climate and relatively resilient soils, soil degradation is less severe in Europe than in Asia, Africa or Australia (Oldeman 1994). Land area prone to soil degradation in Europe is estimated at 218 Mha of which 194 Mha is at moderate plus level of severity (Table 3). Among principal soil degradative processes, soil erosion by

Table 1 Potential rates of soil organic carbon sequestration

Land use and management	Temperate climate		Tropical climate	
	Humid, Mg C ha^{-1} yr^{-1}	Dry, Mg C ha^{-1} yr^{-1}	Humid, Mg C ha^{-1} yr^{-1}	Dry, Mg C ha^{-1} yr^{-1}
Conservation tillage	0.5–1.0	0.25–0.5	0.25–0.5	0.1–0.25
Cover cropping/cropping systems	0.2–0.5	0.1–0.2	0.1–0.2	0.05–0.1
Integrated nutrient management/manuring	0.3–0.6	0.2–0.4	0.2–0.4	0.1–0.2
Water management/irrigation	0.1–0.2	0.25–0.5	0.05–0.1	0.2–0.4
Erosion control/waste conservation	0.4–0.8	0.1–0.2	0.2–0.4	0.05–0.1
Agroforestry	0.3–0.6	0.2–0.4	0.4–0.8	0.2–0.4
Improved grazing	0.2–0.4	0.1–0.2	0.4–0.8	0.1–0.2
Soil restoration	0.5–1.0	0.4–0.6	0.8–1.2	0.2–0.4

A negative sequestration in humid climates is caused by drainage of poorly drained wetland soils

Table 2 Land use and soil degradation in Europe and Spain (FAO 2000)

Land use	Area	
	Europe, Mha	Spain, Mha
Total area	2298.7	50.6
Land area	2261.0	49.9
Arable land	291.1	13.7
Permanent crops	17.0	4.9
Permanent pasture	183.1	11.5
Forests and woodland	157.6	15.8

Table 3 Estimates of soil degradation in Europe (adapted from Oldeman 1994)

Process	Land area affected by soil degradation	
	Total, Mha	Moderate+, Mha
Water erosion	114	93
Wind erosion	42	39
Chemical degradation	26	26
Physical degradation	36	36
Total	218	194

water and wind comprise 72% of the total area and 68% of the moderate plus area affected by soil degradation (Table 3). Strongly interacting with erosion is the process of physical degradation that affects 36 Mha in Europe. Dregne and Chou (1992) estimated the extent of desertification in Europe. Of the total dryland area of 146 Mha, 94 Mha (or 65%) is prone to desertification (Table 4).

Table 4 The extent of desertification in Europe (adapted from Dregne and Chou 1992)

Land use	Total dryland area, Mha	Land area prone to moderate + desertification, Mha
Irrigated cropland	11.9	1.9
Rainfed cropland	22.1	11.9
Rangeland	111.6	80.5
Total	145.6	94.3

Conversion of natural to managed ecosystems, and susceptibility to soil degradative processes can lead to depletion of SOC pool with attendant release of CO_2 and other GHGs into the atmosphere. Assuming SOC depletion of 10–20 Mg C ha^{-1} for arable land, 5–10 Mg C ha^{-1} for pasture, 2–5 Mg C ha^{-1} for permanent crops, and 5–10 Mg C ha^{-1} for forest and woodland, the historic depletion of SOC pool in soils of Europe may be 4–8 Pg of C. The magnitude of SOC depletion may be relatively more in soils prone to degradation and desertification. Of the total agricultural land area of 491, 218 Mha prone to degradation and desertification may lose an additional 5–10 Mg C ha^{-1}. Thus, total historic loss of SOC from soils of Europe may be 5–11 Pg C. These estimates are tentative and preliminary at best, and need to be improved through systematic evaluation of the current and historic land uses and their impact on SOC pool through appropriate statistical/biotic techniques (Peng et al. 1995). Tentative as these estimates are, they provide a reference point with regard to the potential of soils of Europe to sequester

C through land use change. Further, assuming that two-thirds of the SOC lost can be sequestered, total potential of SOC sequestration in soils of Europe may be 3–7 Pg over 25–50 years.

Impact of climate change on soil organic carbon pool, soil quality and biomass productivity

Assessment of the SOC dynamics due to past climate changes requires a clear picture of the dependence of soil quality on climate change as mediated by changes in vegetation and relative distribution of terrestrial biomes. Biomes or ecosystems, and the terrestrial C pool in them, respond to climate-induced changes in temperature and moisture regimes. At the ecosystem level, the soil affects vegetation through its influence on water availability and the C/N cycle (Cheddadi et al. 2001). Change in soil moisture regime can affect species composition in the ecosystem. It is predicted that doubling of CO_2 would lead to an average increase of about 10% in precipitation at the global level (Hennessy et al. 1997). Prentice and Fung (1990) simulated climate-driven spatial arrangement of vegetation at global scale. They estimated a mass transfer of C from the terrestrial biosphere to the atmosphere ranging from 30 Pg (or 15 ppm of CO_2) to 50 Pg (25 ppm of CO_2). White et al. (1999) predicted the impact of climate change on net primary productivity (NPP). By 2020, they predicted a partial loss of the Amazonian rainforest, C_4 grasslands and temperate forest in areas of southern Europe and eastern USA, but expansion in the boreal forest area. Projected increase in temperature and decrease in effective rainfall may decrease NPP in many tropical areas, but increase that in the Boreal forests. Global net ecosystem productivity (NEP) may increase from about 1.3 Pg C yr^{-1} in the 1990s to 3.6 Pg C yr^{-1} in 2030 and then decline to zero by 2100. Peng et al. (1995) reported a decrease in terrestrial C pool in Europe, between –31% and –47%, during the late glacial period due to the persistence of ice sheets.

Kleemola et al. (1995) predicted a longer growing season in Finland leading to introduction of new/more productive cultivars and even new species in the region. In addition to the favorable effect of a longer growing season, NPP may also increase due to the CO_2 fertilization effect. A similar favorable response may occur in other cold regions (e.g., Scandinavia, Siberia). There will be favorable effects on animal and plant communities, especially in polar areas (Sohlenius and Boström 1999). For conditions in central and southern Sweden, Eckersten et al. (2001) reported that winter wheat production may increase by 10–20% (depending on soil type) with attendant increase in litter biomass returned to the soil.

Changes in terrestrial C pool, NPP and NEP also influence soil C pool. Temperature is the primary rate determinant of microbial processes. Dalias et al. (2001) concluded that in addition to temperature controlling rates of C mineralization in soil, it also affects the processes of decomposition. Dalias and colleagues observed that decomposition of by-products produced at higher temperatures may be more recalcitrant than those at lower temperatures.

The effects of projected climate change may be different in the Boreal, Tundra and Polar regions than in mid latitudes. In cold region soils (Cryosols), peat and other organic soils are presently net C sink. Increase in temperatures may have dramatic influences on soil organisms and the mineralization process (Sohlenius and Boström 1999), rendering soils of these regions a net source of atmospheric CO_2. In the U.K., Buckland et al. (2001) predicted that climate change will give rise to new opportunities for grassland invasion. Some native species may establish beyond their current range of distinction.

The sensitivity of SOC pool to climate change, both in terms of its magnitude and quality, warrants an objective evaluation of the impact of projected climate change on rendering soils as a source or sink of atmospheric CO_2, and cycles of principal elements (e.g., N, P, S) and water, which are closely linked to the turnover rate of SOM. The impact of projected climate change of 3°C increase in the Mediterranean Basin on terrestrial/SOC pool was estimated by Bottner et al. (1995). They predicted that a 3°C increase would cause an average altitudinal upward shift of the vegetation belts by 500 m. The productivity of the cold ecoregions would increase as a result of longer growing season and reduced winter cold stress. In most of the arid zone, however, the temperature increase would affect the length of the growing season. Consequently, an increase in temperature would exacerbate the loss of SOC pool. The magnitude of SOC loss was estimated at 28% in the hyper-humid zone, 20% in the sub-humid zone, and

15% in the hyper-arid warm or cool zones of the Mediterranean region (Bottner et al. 1995). In addition, the depletion of SOC pool may be confined to the upper soil layers.

Cheddadi et al. (2001) also studied the effects of projected increase in CO_2 on the Mediterranean vegetation. They projected that an increase in atmospheric CO_2 to 500 ppmv, with an attendant increase in mean temperature of 2°C and a severe reduction (>30%) of the present annual precipitation, may significantly change the present vegetation surrounding the Mediterranean. However, an annual increase in temperature by 2°C would not increase risks of desertification on any part of the Mediterranean unless annual precipitation decreased drastically.

In Atlantic Europe, Duckworth et al. (2000) observed that a 2°C increase in temperature would shift towards vegetation associated with warmer conditions, a shift of about 100 km on the ground. Theoretically, an average rise in mean annual temperature of 1°C is the equivalent of an approximate poleward shift of vegetation zones by 200 km (Ozenda and Borel 1990). However, the projected climate change may be gradual, and the initial effects subtle (Hendry and Grime 1990). In fact, when the effect of climate change is considered on vegetation as a whole, particularly the land use, both interspecific interactions and interactions with environmental factors, reduce the potential of major change in the ecosystem or the biome (Duckworth et al. 2000).

Climate change may also alter the rate and magnitude of emission of GHGs from arable land. Modelling studies by Olesen et al. (2004) showed that total GHG emission increased with increase in temperature. The increase in total GHG emission was 66–234 kg CO_2 equivalent ha^{-1} for a temperature increase of 4°C. In addition, higher rainfall increased total GHG emission in winter-cereals dominated rotations. An increase in rainfall of 20% increased total GHG emissions by 11–53 kg CO_2 equivalent ha^{-1} yr^{-1}, and a 50% increase in atmospheric CO_2 concentration decreased emissions by 180–269 kg CO_2 equivalent ha^{-1} yr^{-1}. Further, GHG emissions increased with increase in use of nitrogenous fertilizer (Olesen et al. 2004). Fraction of SOC lost by respiration is also influenced by the ambient CO_2 concentration, it increases with increase in ambient concentration (Sindhoj et al. 2000). Possible warming and drought due to projected climate change may, especially in the Mediterranean climate, lead to shift in the species composition of seedlings, reduction in biodiversity, increase in the threat of wild fire, and increase in risks of soil erosion (Wessel et al. 2004). Drought may also decrease the flow of C from roots to the soil. A modeling study was done to assess changes in SOC pool in forest soils of Sweden due to the projected increase in temperature (Agren and Hyvonen 2003). Increase in temperature by 4°C will increase SOC losses by 0.9 Tg yr^{-1}. The projected increase in temperature in the northern latitudes can enhance NPP with a positive impact on SOC pool (Callesen et al. 2003). Therefore, climate change may affect ecosystem functioning through increase in temperatures or changes in precipitation patterns. Agren and Hyvonen (2003) estimated that increase in NPP in Swedish forests may be 0.7 Tg C yr^{-1}, leading to a net loss of SOC from Swedish forest soils with increase in temperatures.

Potential of soil carbon sequestration in Europe

Anticipated land use changes and adoption of improved soil/crop/vegetation management practices may cause increase in the overall SOC pool in Europe over the next 50–100 years.

Land use change

Bouma et al. (1998) reported that major changes in land use may be anticipated because of technological, socio-economic and political developments. Technical possibilities of modern agriculture may produce the same yields on only 30–50% of the current agricultural area. That being the case, there is a potential of conversion of agricultural land to other uses (e.g., pasture, forest, reserves or set-aside land). With the possibility of taking much of the land out of production, there is a large potential of sequestering most of the historic C lost. Katterer et al. (2004) reported that increase in crop yields due to adoption of RMPs enhanced SOC pool because of increase in C input into the soil.

Importance of forests

Forests play an important role in the C cycle both at regional and global scales. Distribution of the

Table 5 Distribution of forest in Europe and other regions in the northern temperate zone (modified from Nabuurs et al. 1997)

Region	Area, Mha
Europe (excluding USSR)	149.3
USSR	755.0
North America	456.7
Temperate Asia	179.8
Total	1540.8

forested area in Europe is shown in Table 5. Combined with the former USSR, total area under forest in Europe is 904.3 Mha or 59% of the total forest in the Northern Hemisphere temperate and boreal zones. Review of the C budget studies in European forests show that forest area of Europe of 149 Mha (excluding the FSU) yields a whole tree C sink of 101.3 Tg C yr^{-1} which is equivalent to 9.5% of the European emissions. The total C stock of these forests (whole tree) has been estimated at 7.9 Pg C, and a wood C sink of 29.2 Tg C yr^{-1} (Nabuurs et al. 1997). In comparison, the annual sink capacity of FSU forest has been estimated at 517 Tg C yr^{-1} by Kolchugina and Vinson (1993a, b) and 740 Tg C yr^{-1} by Dixon et al. (1994). Janssens et al. (2003) estimated the net biome productivity of the European forest sector, with a total area of 384 Mha, at 377 Tg C yr^{-1}. The sink capacity of all temperate forests has been estimated at 1.4 Pg C yr^{-1} by Sedjo (1992), and for northern boreal and temperate zone forest at 0.7–1.3 Pg C yr^{-1} by Sampson et al. (1993).

Despite its importance, a few studies have accounted for the C sink capacity of soils supporting forests. In this regard, a preliminary estimate provided by Nabuurs et al. (1997) is extremely relevant. They reported the total SOC pool of soils supporting European forests at 12.0 Pg, but did not provide an estimate of the rate of SOC sequestration.

Liski et al. (2002) estimated C budget of soils and trees in the forests of 14 EU countries plus Norway and Switzerland from 1950 to 2040. The SOC pool increased throughout the study period. The SOC sink was 26 Tg C yr^{-1} in 1990 and projected to be 43 Tg C yr^{-1} by 2040. The SOC sink in forest soil can be managed by choice of species, soil fertility, etc. Positive impact of CO2 fertilization on C sequestration in the forest biomass and soils may be limited by lack of nutrients or soil fertility constraints (Oren et al. 2001).

Soil carbon sequestration

Soil C sequestration is an important strategy of offsetting fossil fuel emissions from Europe or from any other geographical region. Several studies have reported the potential of C sequestration in European soils through adoption of RMPs (Smith et al. 1998, 2000b). Conservation tillage (no-till farming, direct drilling) has a potential to sequester about 23 Tg C yr^{-1} in the European union or about 43 Tg C yr^{-1} in the wider Europe including FSU (Smith et al. 1998). In addition to enhancing SOC pool, up to 3.2 Tg C yr^{-1} may also be saved in agricultural fossil fuel emissions. The potential of no-till farming to sequester C is greatly enhanced (twice as much to 46 Tg C yr^{-1}) whereby soils are amended with organic materials. The SOC content of soils was estimated at 15.3 Pg for the European Union compared with 34.6 Pg for the wider Europe, and 7.5 Pg for agricultural land for the European Union compared with 13.5 Pg for the wider Europe. The SOC pool of the arable land was 3.9 Pg for the European Union and 7.2 for the wider Europe. Using this baseline, Smith et al. (1998) concluded that 100% conversion to no-till agriculture in Europe could mitigate all fossil fuel C emissions from agriculture in Europe (Tables 6, 7).

Smith and Powlson (2000) also estimated the importance of manuring on SOC sequestration in Europe. They estimated that 820 million metric tons of manure are produced in Europe each year, and only 54% is applied to arable land and the remainder to non-arable agricultural land (e.g., grassland). With a high antecedent SOC level, application of manure to grassland does not enhance the SOC pool.

Table 6 Fertilizer and manure use in Europe (FAO 1998)

Nutrient source	Amount (10^6 Mg yr^{-1})
Manure	820
Fertilizer	
N	20.7
P_2O_5	6.3
K_2O	5.5

Table 7 Measured rates of soil C sequestration in some European countries

Country	Cropping system	SOC sequestration rate (Kg C ha^{-1} yr^{-1})	Reference
Denmark	(i) Silage maize	250–490	Kristiansen et al. (2005)
	(ii) Silage + manure @ 8 Mg ha^{-1}	710–980	
Denmark	Miscanthus plantations (0–100 cm)	2071	Hansen et al. (2004)
Sweden	Management effects on topsoil C	Highly variable	Karlsson et al. (2003)
Finland	C accumulation in mineral soil beneath peat over 150–3000 yrs	2–140	Turunen and Moore (2003)
Europe	Wetland restoration	0–800	Janssens et al. (2003)
Iceland	Restoration of eroded soils	600–1000	Arnalds et al. (2000); Gudmundsson et al. (2004)

However, applying manure on arable land can enhance its SOC pool. Smith and Powlson (2000) estimated that if all manure were incorporated into arable land in the European Union, there would be a net sequestration of 6.8 Tg C yr^{-1}, which is equivalent to 0.8% of the 1990 CO_2-C emissions for the region.

In addition to no-till and manuring, there are also other management practices with a potential to enhance the SOC pool. Important among these are crop rotations and cover crops (Singh et al. 1998). In Hungary, Berzseny and Gyrffy (1997) reported higher crop yields and SOC pool when cereals were rotated with legumes. Growing a cover crop in rotation with grain crops can accentuate the rate of SOC sequestration (Fullen and Auerswald 1998; Uhlen and Tveitnes 1995). In the U.K., Fullen and Auerswald (1998) reported the rate of SOC sequestration at 0.02% yr^{-1} for 12 years in a set-aside land. Increase in SOC concentration by including a cover crop in the rotation cycle was reported in Sweden (Nilsson 1986), Netherlands (Van Dijk 1982), U.K. (Johnston 1973) and Europe (Smith et al. 1997).

These discussions on SOC sequestration do not include the potential of restoring peat soils by inundation. For example, peat soils occupy about 0.3 Mha of land area in Sweden (10% of Swedish arable land). However, these soils emit 1 Tg C yr^{-1} (Andren et al. 2004). Furthermore, soil respiration in peat soils can be 2–3 times as high as in the mineral soils because of the high SOC concentration (Lohile et al. 2003). Restoration of these soils may convert these lands into a net sink of about 0.3 Tg C yr^{-1} at the mean rate of 1 Mg C ha^{-1} yr^{-1}.

Potential of European soils for carbon sequestration

The Kyoto Protocol recognizes soil and forest as C sink, and the European Union is committed to reduction in CO_2 emissions to 92% of the 1990 level during the first commitment period.

Janssens et al. (2003) estimated that Europe's terrestrial biosphere absorbs 7–12% of European anthropogenic CO_2 emissions. In consideration of the land use and soil characteristics, options for C mitigation on agricultural land on a national scale have been discussed for Norway by Singh and Lal (2001), U.K. by Smith et al. (2000a) and Europe by Smith et al. (2000b). In Norway, soil/crop management practices with potential for C sequestration include conservation tillage along with mulch farming techniques, integrated nutrient management (INM) combining a judicious use of fertilizers and manures, diverse crop rotations including ley farming and cover crops, erosion control measures and restoration of peatlands (Singh and Lal 2001). The overall potential of these practices for SOC sequestration ranges from 0.6 to 1.0 Tg C yr^{-1}. Of the total potential, 59% is due to adoption of erosion control measures, 21% to conversion of plow till to conservation tillage and mulch farming, 14% to improved cropping systems including rotations and manuring, and 6% to restoration of peatlands. In the U.K., Smith et al. (2000a) estimated C mitigation on agricultural land at 20.4 Tg C yr^{-1} comprising 3.7 Tg by using animal manure, 0.3 Tg by sewage sludge, 1.9 Tg by cereal straw incorporation, 3.9 Tg by no-till farming, 3.3 Tg by agricultural extensification, 3.2 Tg by natural woodland regeneration, and 4.1 Tg by bioenergy crop production. Smith and colleagues observed that a realistic potential for C mitigation on U.K. agricultural land is 10.4 Tg yr^{-1}, which is about 6.6% of 1990 U.K. CO_2-C emissions. Extending the methodology used in estimating U.K. potential to Europe, Smith et al. (2000b) assessed a range of options for C mitigation by agriculture. Important among all options examined, bioenergy crops showed the greatest potential for C mitigation. They also observed that most important resource for C mitigation through agriculture in Europe is the surplus arable land. The SOC sequestration potential for soils of Europe was estimated at 56 Tg C yr^{-1} (Smith et al. 2000b) compared with the global potential of 400 to 800 Tg C yr^{-1} (IPCC 1996). When aboveground C is included, the total C mitigation potential of Europe is 113 Tg C yr^{-1}. However, these estimates do not consider emission of other GHGs such as CH_4 and N_2O. These GHGs may be important in those agricultural practices, which involve the use of biosolids as amendments (e.g., manure, sewage sludge and crop residues) (Smith et al. 2001). Janssens et al. (2003) estimated the net C sink of 135 and 205 Tg C yr^{-1} in Europe's terrestrial biosphere, or the equivalent of 7–12% of the anthropogenic C emissions.

Carbon mitigation potential of European agriculture

Presently, European cropland soils are losing 300 Tg C yr^{-1} to the atmosphere (Janssens et al. 2003), which essentially offsets the sink in the forest sector. In contrast to croplands, grassland soils are a net sink of 101 Tg C yr^{-1} (Janssens et al. 2003). Restoration of eroded/degraded soils and adoption of RMPs on remaining agricultural and forestry lands are important strategies of soil C sequestration. Similar to the soils and agroecoregions of the U.S. (Lal et al. 1998), the potential of soil C sequestration is high for temperate humid climates. In this regard, a judicious management of manure and fertilizers is critical. Even if only 10% of the 820 million tons of manure produced in Europe can be converted into a recalcitrant humic fraction (Table 5), the potential of SOC sequestration is 32 Tg C yr^{-1} assuming a C concentration of 40%. However, the manure, along with other organic amendments, have to be applied on degraded and other agricultural soils. Similarly, other soil/site specific RMPs have to be used to intensify agricultural/forestry production.

The data in Table 8 shows a tentative estimate of SOC sequestration potential of European agriculture and forestry, without consideration of the biofuel and other options for fossil fuel offset or emission avoidance. The SOC sequestration potential of 67–191 Tg C yr^{-1} (129 ± 62) can be achieved through adoption of appropriate policies. The large uncertainty also needs to be addressed through development of site-specific options at regional and national scales. It has been argued that SOC sequestration potential of European soils has been overestimated (Smith et al. 2005). Thus, these results should be used with caution.

Conclusions

The land use, land use change, forestry activities, and agricultural soils can be used to meet national regional commitments under the Kyoto emission reduction targets. Potential for C mitigation by European agriculture can be an important factor to

Table 8 Potential for carbon mitigation by European agriculture

Strategy	Area (Mha)	C sequestration rate (Mg/ha/yr)	Total potential (Tg C yr^{-1})
Restoration of degraded soils as set-aside land	94.3	0.2–0.6	18.9–56.6
Adoption of improved RMPs on cropland (minus degraded soils)	196.8	0.1–0.4	19.7–78.7
Improved pastures	183.1	0.1–0.2	18.3–36.6
Improved permanent crops	17.0	0.1–0.2	1.7–3.4
Improved forest and woodlands	157.6	0.05–0.1	7.9–15.8
Total potential			66.5–191.1

meeting its commitment of reduction in CO_2 emissions to 92% of the 1990 baseline level during the first commitment period of 2008–2012. Important practices with a substantial potential of soil C sequestration include soil restoration and set-aside land, use of manure and other biosolids, adoption of no-till farming or direct seeding, and use of improved systems of management on agricultural and forestry lands. The total potential for C mitigation by agriculture and forestry is 66–119 Tg C yr^{-1}. This is a gross potential and does not account for the hidden C costs of input involved, which must be accounted for. However, realization of this potential depends on identification and implementation of relevant policy options that facilitate adoption of RMPs and conversion of marginal/degraded soils to restorative land uses. Soil C sequestration, being an important political and economic factor in meeting Kyoto emission reduction commitments, needs a serious consideration of scientists, policy makers and land managers.

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