Chapter 14
Cropland Soil Carbon Dynamics

Klaus Lorenz and Rattan Lal

Abstract Humans began thousands of years ago to cultivate land for growing crops after clearing the previous vegetation cover and plowing the soil. The soil disturbance altered soil carbon (C) dynamics which has been recently exacerbated by the increase in crop intensification (i.e., fertilization, irrigation, mechanization). For example, conversion to croplands may release up to 36% of soil organic carbon (SOC) to 27-cm depth in temperate regions, and up to 30% of SOC to 48-cm depth in tropical regions. In 2000, about 12% of Earth's ice-free land surface or 15 million km^2 were covered by croplands. Climate, geology and land and crop management practices control the size of the cropland soil C pool. A major fraction (25–70%) of the carbon dioxide (CO_2) fixed during plant photosynthesis in croplands by gross primary production (GPP) is respired autotrophically (R_a) back to the atmosphere. Globally, cropland GPP is about 14.8 Pg C year^{-1} (1 Pg = 10^{15} g). The remaining net primary production (NPP = GPP−R_a) is the main natural C input into cropland soils aside addition of manure and organic residues. Cropland NPP includes the production of biomass in foliage, shoots and roots, weed and seed production, root exudation, the C transfer to microorganisms that are symbiotically associated with roots, and the volatile organic carbon (VOC) emissions that are lost from leaves to the atmosphere. NPP enters soil by rhizodeposition and decomposition of plant litter but the major fraction is heterotrophically converted back to CO_2 by soil respiration and some lost as methane (CH_4). Aside decomposition, C losses from croplands

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occur also by fire, erosion, leaching, and most importantly harvest removing about 2.2 Pg C year^{-1} in the 1990s. Thus, a small amount of fixed C remains in cropland soils and accumulates in the SOC pool due to a combination of short- and long-term stabilization processes. Stabilization processes include physical protection of organic matter (OM) against decomposers and their enzymes, stabilization by organomineral complexes and organo-metal interactions, and some as biochemically recalcitrant black carbon (BC). Soil aggregation, in particular, may be the most important stabilization process in cropland topsoils. Site-specific factors including climate, physicochemical characteristics, soil and vegetation management determine the balance between C input and losses. Cropland soils can be recarbonized to some extent through adoption of recommended management practices (RMPs) such as conservation tillage, residue mulching and use of cover crops, practices which all contribute to soil C accumulation and sequestration by an additional transfer of C from the atmosphere to the soil. Whether cultivation of SOC-accreting crops can also contribute to the recarbonization of cropland soils needs additional research.

Keywords Soil organic carbon • Gross primary production • Net primary production • Recommended management practices

Abbreviations

AM	Arbuscular mycorrhiza
AUR	Acid-unhydrolyzable residue
BC	Black carbon
BIO	Microbial biomass
CQT	Carbon quality-temperature
DIC	Dissolved inorganic carbon
DOC	Dissolved organic carbon
DPM	Decomposable plant material
ECM	Ecto-mycorrhiza
ERM	Ericoid mycorrhiza
EU	European Union
FAO	Food and Agriculture Organization of the United Nations
GPP	Gross primary production
HI	Harvest index
HUM	Humified organic matter
NBP	Net biome production
NPP	Net primary production
NT	No-tillage
OM	Organic matter
PT	Plow tillage
PTF	Plant functional type
R_a	Autotrophic respiration

RMP	Recommended management practices
RothC	Rothamsted carbon model
RPM	Resistant plant material
SIC	Soil inorganic carbon
SOC	Soil organic carbon
SOM	Soil organic matter
UK	United Kingdom
USA	United States of America
VOC	Volatile organic carbon

14.1 Introduction

The global cropland area has increased dramatically as crops have been cultivated by humans for thousands of years on land reclaimed by converting forests, savannas and grasslands causing the largest emissions of carbon (C) from land-use change (Houghton 2010; Pielke et al. 2011). Predominantly in Asia, forests have been cleared for croplands, and a substantial amount of savanna and grasslands have been converted to croplands in North America, Africa, and the Former Soviet Union (Ramankutty et al. 2008). About 80% of new croplands are currently replacing forests in the tropics (Gibbs et al. 2010). Between 1985 and 2005, global cropland area increased by only about 2.4% but harvested area by about 7% (Foley et al. 2011). This cropland intensification is associated with large increases in energy, water and fertilizer consumption, and considerable losses in biodiversity (Foley et al. 2005). In 2000, about 12% of Earth's ice-free land surface or 15 million km^2 were covered by croplands (Ramankutty et al. 2008). The greatest cropland areas were located in South Asia, the Former Soviet Union and Tropical Africa (2.22, 2.07 and 1.94 million km^2, respectively). Cropland area can also be distinguished by biomes such as residential irrigated (i.e., irrigated cropland with substantial human populations), residential rainfed mosaic (i.e., mix of trees and rainfed cropland with substantial human populations), populated irrigated, populated rainfed and remote croplands (Ellis and Ramankutty 2008). In 2008, residential, rainfed mosaic was by far the most extensive among the cropland biomes with 4 million km^2 covered by crops, most abundant in Africa and Asia. Areas covered by crops in the other cropland biomes were 2.2, 0.97, 0.51, and 0.25 million km^2 in populated rainfed, residential irrigated, remote and populated irrigated cropland biomes, respectively (Ellis and Ramankutty 2008).

In 2000, the globally harvested areas for cereals, oil crops and forage were about 6.6, 1.8, and 1.4 million km^2, respectively (Monfreda et al. 2008) Annual grains were planted on 75% of the global cropland area. The largest cereal-harvested areas were located in Asia (2.7 million km^2), and Europe and the Former Soviet Union (in total 1.3 million km^2). On about 0.7 million km^2 oil crops were harvested in Asia, and on similar-sized area forage in combined Europe and the Former Soviet Union. Perennial crops can be distinguished from annual crops based on longevity.

Specifically, crops that live for more than one year under typical cultivation practices can be classified as perennials. Distinguishing between annual and perennial crop area is important with regard to soil organic carbon (SOC) sequestration. Specifically, perennial crops allocate a higher proportion of photosynthetically-fixed C belowground compared to annual crops by maintaining a permanent vegetation cover and a high root turnover with a high SOC input (Don et al. 2011). Further, compared with annual counterparts, perennial crops tend to have longer growing seasons and deeper rooting depths, and intercept, retain, and utilize more precipitation (Glover et al. 2010). About 1.8 million km^2 of harvested crop area in 2000 was under perennial orchards, grasses and pastures (Monfreda et al. 2008). High proportions of perennial crops were located in the tropics, particularly in Indonesia, Thailand, Malaysia, and Papua New Guinea. Often perennial crops produce high-value products like coffee (*Coffea arabica* L.), cocoa (*Theobroma cacao* L.), fruit, oil palm (*Elaeis guineensis* or *E. oleifera*), and nuts for export.

For the assessment of SOC sequestration potential it is useful to classify the cropland also based on other plant functional types (PFTs) aside annual/perennial PFTs (Smith et al. 1993). For example, the crop life form or physiognomy classification herbaceous (forbs and graminoids) dominated 91% of all harvested crop area in 2000 (12.3 million km^2; Monfreda et al. 2008). Annual crops, in particular, are exclusively herbaceous. The remaining 9% nonherbaceous crop area were under the life forms wood perennial shrubs and trees (0.4 and 0.8 million km^2, respectively). Shrubs were predominant in the coffee and cocoa growing regions of western and eastern Africa, the Central American countries south of Mexico, the eastern coast of Brazil, northern South America in Ecuador, Colombia, and Venezuela, and pockets of southeast Asia. Grapes (*Vitis* spp.) and berry bearing shrubs covered small areas in temperate countries. Further, herbaceous crops occupied much more area than either shrub or tree crops throughout the temperate zones. Although fruit bearing orchards were present in the humid tropics, coconuts (*Cocos nucifera* L.) and oil palm dominated. Specifically, Indonesia cultivated one quarter of the global coconuts area, and Indonesia and Malaysia together grew one half of all oil palm trees in 2000 (Monfreda et al. 2008).

Classification of croplands with respect to SOC sequestration is also possible by distinguishing crop functional types C_3 and C_4 based on photosynthesis pathways. Specifically, the proportion of C_3 and C_4 vegetation in a region affects the flux of carbon dioxide (CO_2) between plants and the atmosphere (Still et al. 2003). In 2000, the C_4 crops corn (*Zea mays* L.), sorghum [*Sorghum bicolor* (L.) Moench], millet, sugar cane (*Saccharum* spp.) and some grasses comprised 3.2 million km^2 or a disproportionate 24% of all harvested area (Monfreda et al. 2008). The greatest proportion of C_4 crops was located in central Mexico through the Yucatan Peninsula, the eastern portion of southern Africa extending from South Africa to Mozambique and Zambia, and a huge swath of the Sahel running from Mauritania in the west to Ethiopia in the east (Monfreda et al. 2008). Further, at least 50% of all crops planted in the U.S. Corn Belt and of the crops grown in northern Argentina and southern Brazil were C_4 crops. In these regions, corn is grown in rotation with soybeans [*Glycine max* (L.) Merr.] and fodder grass to produce animal feedstock.

The C_4 crops are, in particular, critical in arid, agriculturally marginal regions (Monfreda et al. 2008).

Legumes are the second most important source for nitrogen (N) in the terrestrial biosphere contributing about 40 Tg N year^{-1} (1 Tg = 10^{12} g), or about half of the amount released through synthetic fertilizer (Monfreda et al. 2008). In 2000, leguminous crops occupied 2.4 million km^2 or 18% of the global harvested crop area which included all pulses plus groundnuts, soybeans, alfalfa (*Medicago sativa* L.), and other leguminous forage crops, and half of the area of mixed grasses and legumes. Specifically, soybeans and alfalfa covered much of the central U.S., the region near the border of Argentina and Brazil, and southern Australia. Soybeans and alfalfa were primary protein source for livestock in these regions and were grown in rotation with corn, which supply livestock with the bulk of their energy requirements (Monfreda et al. 2008).

14.2 Cropland Soil Organic Carbon Pool

The fate of C recently fixed during plant photosynthesis (i.e., gross primary production or GPP) is the major determinant of the cropland SOC pool (Sanderman et al. 2010). A major fraction of GPP (25–70%) is expended during plant autotrophic respiration (R_a; Lambers et al. 2005). Only the fraction remaining after accounting for C losses by R_a (i.e., the net primary production or NPP = GPP–R_a) is stored in new plant biomass. Before reaching the soil, crop NPP is lost by biomass removal during harvest (i.e., grains, pulses, and other harvestable products). For example, 30–50% of the aboveground dry mass is harvested from cereal croplands (Johnson et al. 2006). In the 1990s, global crop harvest was about 2.2 Pg C year^{-1} (1 Pg = 10^{15} g) (Bondeau et al. 2007). Herbivory by insects and mammals, and emissions of volatile organic compounds (VOCs) also contribute to biomass loss from croplands (Ciais et al. 2010). And some cropland C is lost by fire, erosion and leaching (Torn et al. 2009). For example, about 1.6 kg CO_2 is emitted per kg of crop residues burned (Andreae 2004). Globally, about 1,200 Tg of crop residues was burnt in the late 1990s. Especially in the tropics, the ignition of fires is a common human activity to clear land for shifting cultivation by converting forests into croplands (Crutzen and Andreae 1990). Also, dry vegetation is removed by human-induced fires to promote crop productivity. However, burning of crop residues in developed countries has been greatly reduced during the industrial era due to the large-scale expansion of intensive cropping and active fire management (Marlon et al. 2008). Still being burned are crop residues in developing countries but comprehensive contemporary global data on crop burning are lacking (Crutzen and Andreae 1990; Pechony and Shindell 2010).

Some of the fixed C in non-erosional croplands remaining after accounting for the losses by fire, harvest, herbivory and VOCs is deposited as above- and belowground residues. Aboveground but more importantly belowground residues are SOC sources (Rasse et al. 2005). Other belowground input aside residues is GPP transferred into the soil as root and mycorrhizal exudates including turnover of fine roots (i.e., <2 mm in diameter) and fungal hyphae, and as other dead microbial cells. Transfer of crop C into soils occurs also by the activity of the mesofauna (e.g., earthworms), hydrological flows of soluble C, and root senescence and mortality. About 50–70% of C fixed in croplands remains aboveground, and 20–30% is transferred belowground (Johnson et al. 2006). Aside by natural processes, organic C is also entering cropland soils by plant and soil management (i.e., addition of manure, and of non-harvested and non-burned residues).

The soil C inputs are substrate for decomposition and subject to losses by heterotrophic, primarily microbial respiration. The remaining SOC pool consists of organic C compounds in plant, microbial and faunal residues at various stages of decomposition with residence times ranging from days to millennia (Sanderman et al. 2010). Natural processes, environmental factors (i.e., climate, soil parent material), and vegetation and soil management (i.e., crop type, fertilization, irrigation, harvest, residue management, tillage) determine amount and composition of the SOC pool. The persistence of SOC is largely due to complex interactions between organic matter (OM) and its environment, such as the interdependence of compound chemistry, reactive mineral surfaces, climate, water availability, soil acidity, soil redox state and the presence of potential degraders in the immediate microenvironment (Schmidt et al. 2011). Among natural processes affecting the SOC pool in croplands are (i) C allocation and partitioning among plant organs (i.e., leaves, stems, roots), (ii) C fluxes within plant organs (i.e., to respiration, storage compounds, defensive compounds, structural components) and (iii) C fluxes among soil pools (U.S. Department of Energy 2008). The fraction of organic C accumulating in croplands (i.e., net biome production or NBP) which enters the soil (NBP_{soil}) primarily determines the amount of SOC sequestered (Schulze et al. 2010).

14.2.1 Cropland Soil Carbon Input

The SOC pool is primarily derived from photosynthetically fixed C (Sanderman et al. 2010). Thus, cropland GPP or the annual photosynthetic C uptake of all plant tissues (i.e., crops, weeds) over a specified cropland area determines the potential C input into cropland soils (U.S. Department of Energy 2008). Globally, cropland GPP is about 14.8 Pg C $year^{-1}$ (Beer et al. 2010). From a crop production perspective, photosynthesis includes all the events from light interception to the export of photosynthate for biomass accumulation and grain production (Murchie et al. 2009). Thus, GPP is the basis for food, fiber and biofuel production, and one of the major processes providing the capacity of croplands to partly offset anthropogenic CO_2 emissions (Beer et al. 2010). The net C stored in croplands as new plant material before harvest and other losses is the NPP. The portion of crop NPP at non-erosional croplands remaining after accounting for losses by harvest, herbivory and VOCs is deposited as above- and belowground residues. Decomposition of aboveground residues but more important of belowground residues together with root exudates, and biotic and hydrologic belowground C transfer are the major natural cropland soil C input processes (Lorenz and Lal 2005). Addition of manure and residues during land use and soil management may result in additional direct C input into cropland soils.

14.2.1.1 Natural Input Processes

The crop residue input depends on the biomass that grows from a unit input of solar radiation (Amthor 2010). The crop functional types C_3 and C_4 differ in their efficiency in converting solar energy into biomass. Specifically, crops using the C_4 photosynthetic pathway are generally more efficient than C_3 crops but this difference varies with solar radiation, temperature and water supply (Jansson et al. 2010; Sanderman et al. 2010). The reduced photosynthetic efficiency of C_3 crops arises, in particular, from reduced performance (Murchie et al. 2009). However, in spite of the greater capacity and water use efficiency of the C_4 photosynthetic pathway C_4 crops may be equally or even more sensitive to water stress than C_3 crops (Ghannoum 2009). Many economically important agricultural crops such as wheat (*Triticum* L.), rice (*Oryza sativa*) and soybean are C_3 crops (Schulze et al. 2005). The C_4 photosynthetic pathway occurs in the economically important crops such as corn, sugar cane, millet and sorghum.

Gross Primary Production

The GPP is controlled by leaf area, nitrogen (N) supply, season, temperature, light and atmospheric CO_2 concentration (Chapin et al. 2002). Differences in annual GPP among croplands depend primarily on the quantity of leaf area and the length of time this leaf area is photosynthetically active which are both ultimately determined by the interacting effects of soil resources, climate, vegetation, and disturbance regime (Falge et al. 2002).

Previously, plant- and stand-level GPP was calculated as two times biomass production with considerable variation among biomes and sites (Beer et al. 2010). However, the assumption that the NPP/GPP ratio is consistent regardless of ecosystem type has been challenged (Zhang et al. 2009). For example, under the assumption that NPP/GPP equals 2, global cropland GPP is estimated to be 8.2 Pg C year^{-1}, but to be 14.8 Pg C year^{-1} based on eddy covariance flux data and diagnostic models (Saugier et al. 2001; Beer et al. 2010). GPP can be measured but is generally estimated from simulation models as it is impossible to measure the net C exchange of all leaves of a cropland in isolation from other components (e.g., soil respiration; Chapin et al. 2002). Soil respiration is the flux of CO_2 from the soil surface to the atmosphere (Bond-Lamberty and Thomson 2010). Thus, estimations of GPP are based on (i) observations, (ii) diagnostic modeling approaches, (iii) process-oriented models and (iv) satellite-data based models (Yang et al. 2007; Beer et al. 2010). Field studies, for example, use tower eddy covariance systems to calculate seasonal and inter-annual dynamics of cropland GPP (Peng and Gitelson 2011). Some examples of estimated cropland GPP for different time periods, regions and crop species are given in the following section.

Global mean annual cropland GPP for the period 2000–2005 was estimated to be 765 gC m^{-2} $year^{-1}$ based on the first continuous satellite-derived dataset monitoring global vegetation productivity (Zhao et al. 2010). For the period 1982–2004, global crop GPP simulated by a terrestrial biosphere model ranged from 340 to 788 gC m^{-2} $year^{-1}$ (Bonan et al. 2011). For EU-25 (i.e., Austria, Belgium, Cyprus, Czech Republic, Denmark, Estonia, Finland, France, Germany, Greece, Hungary, Ireland, Italy, Latvia, Lithuania, Luxembourg, Malta, Netherlands, Poland, Portugal, Slovakia, Slovenia, Spain, Sweden and UK), mean cropland GPP was estimated to be 1,120 gC m^{-2} $year^{-1}$ for the period 2000–2005 (Schulze et al. 2009). Schulze et al. (2009) used a combination of top-down estimates based on atmospheric observations and bottom-up estimates derived from ground-based measurements. Based on a process-oriented and a remote sensing model, cropland GPP for EU-25 was 1,360 and 879 gC m^{-2} $year^{-1}$ over the period 1990–1999, respectively (Ciais et al. 2010). By applying a NPP/GPP ratio of 0.55, cropland GPP of Europe excluding Russia, Belarus and Ukraine was estimated to be 1,591 gC m^{-2} $year^{-1}$ (calculated based on Beer et al. 2007).

Seasonal changes in crop GPP can be characterized by CO_2 flux measurements. For mixed croplands, CO_2 fluxes within a season were as low as 0.2 μmol CO_2 m^{-2} s^{-1} and as high as 33.5 μmol CO_2 m^{-2} s^{-1} (Falge et al. 2002). The maximum flux was higher for C_4 than that for C_3 crops (60.5 *vs*. 27.3 μmol CO_2 m^{-2} s^{-1}). Estimated annual GPP based on these measurements were 599 gC m^{-2} $year^{-1}$ for C_3 crops, 1,101–1,396 gC m^{-2} $year^{-1}$ for mixed croplands, and 1,471 gC m^{-2} $year^{-1}$ for C_4 crops. Often, a second maximum of CO_2 emissions developed after harvest due to photosynthetic activity of inter-crops or weeds (Falge et al. 2002).

Over 6 years, GPP of irrigated and rainfed corn-soybean cropping systems were estimated based on continuous CO_2 measurements in eastern Nebraska, USA (Suyker and Verma 2010). Peak daily GPP values for irrigated and rainfed corn, and for irrigated and rainfed soybean were 24.9 and 22.9 gC m^{-2} day^{-1}, and 15.4 and 14.4 gC m^{-2} day^{-1}, respectively. Mean growing season totals of GPP for irrigated corn and soybean were 1,738 and 996 gC m^{-2}, respectively. Corresponding values for rainfed corn and soybean were 1,553 and 895 gC m^{-2}, respectively (Suyker and Verma 2010). Carbon fluxes were also measured in the North China Plain in a typical irrigated wheat/corn rotation cropland during four seasons (Lei and Yang 2010). The seasonal GPP for wheat ranged between 782 and 1,114 gC m^{-2}, and for corn between 872 and 880 gC m^{-2}, respectively. The annual GPP values for this rotational cropland were 2,008 gC m^{-2} $year^{-1}$ for the period 2006–2007, and 1,668 gC m^{-2} $year^{-1}$ for the period 2007–2008 (Lei and Yang 2010). In a corn cropland in Northeast China, GPP was simulated for a 3 years period using (i) a photosynthesis model and (ii) estimated based on tower CO_2 flux data (Wang et al. 2010). Annual simulated GPP in 2004, 2005 and 2006 were 310, 464 and 360 gC m^{-2} $year^{-1}$, respectively. In contrast, annual estimated GPP were 392, 504 and 437 gC m^{-2} $year^{-1}$, respectively, for the same time period (Wang et al. 2010).

For a winter wheat/summer maize cropping system in China, GPP varied between 1,370 and 1,531 g C m^{-2} $year^{-1}$ during 2 years (Yu et al. 2006). Depending on crop development stage GPP for winter wheat at a site in Belgium varied between 0.92 and 18.72 g C m^{-2} day^{-1} during three growing seasons (Dufranne et al. 2011). Growing season GPP at this site was as low as 1,568 g C m^{-2} and as high as 1,716 g C m^{-2}. Satellite-based remote sensing technologies can also be used to characterize GPP of croplands during the growing season. For example, Peng et al. (2011) used chlorophyll-related vegetation indices to estimate corn GPP over a period of 8 years at eight rainfed and irrigated sites in Nebraska, USA. GPP followed seasonal changes in chlorophyll content. Daytime growing season GPP varied strongly between 0 and 35 g C m^{-2} day^{-1} depending on weather conditions and growth stage with mean values between 10.4 and 15.1 g C m^{-2} day^{-1} (Peng et al. 2011).

Net Primary Production

The bulk of NPP in croplands is allocated to the production of biomass in foliage, shoots and roots. Cropland NPP includes also weed and seed production, root exudation (i.e., the soluble organic compounds that diffuse or are secreted by roots into the soil), the C transfer to microorganisms that are symbiotically associated with roots (e.g., mycorrhizae and N-fixing bacteria), and the VOC emissions that are lost from leaves to the atmosphere (Chapin et al. 2002; Ciais et al. 2010). Further, NPP includes also the biomass removed by herbivory often accounting for 5–10% of NPP. The processes of NPP loss are among the reasons why direct measurements of cropland NPP are not possible as not all of the biomass produced remains in croplands (Ciais et al. 2010). In addition, other components are rarely measured such as weed production, seed production, emission of VOCs to the atmosphere, exudation from roots and C transfer to root symbionts. Specifically, all components of cropland NPP have not yet been measured in a single study. However, satellites provide a tool for estimating cropland NPP globally (Monfreda et al. 2008). Other important estimation methods are statistical and process-based models. However, cropland NPP estimates mainly rely on census and survey data but these data measure agricultural production and not NPP (Ciais et al. 2010). The estimates of cropland NPP from yield inventory statistics are biased due to uncertainties in NPP definition, allometry, cropland area and input yield data themselves (Ciais et al. 2010). Some examples of cropland NPP for different regions and species are given in the following section.

Globally, cropland NPP for 1991 was 344 g C m^{-2} $year^{-1}$ derived from FAO agricultural production through application of crop-specific coefficients such as dry matter content, harvest index (HI), root production, and C and N content (Goudriaan et al. 2001). Productivity ranged from as low as 181 g C m^{-2} $year^{-1}$ for pulses to as much as 801 g C m^{-2} $year^{-1}$ for sugar crops. In 2000, average global cropland NPP was 397 g C m^{-2} $year^{-1}$ extrapolated by using harvest indices and harvest statistics (Haberl et al. 2007). The average EU-25 cropland NPP for the decade 1990–1999 ranged from 586 to 646 gC m^{-2} $year^{-1}$ based on yield data, from 482 to 585 gC m^{-2} $year^{-1}$ based on process-oriented and from 419 to 510 gC m^{-2} $year^{-1}$ based on remote sensing models, respectively (Ciais et al. 2010). Agricultural statistics (yields and cropped area) were also used to estimate cropland NPP for the conterminous U.S. (Hicke et al. 2004). The cropland NPP in the U.S. increased from about 350 gC m^{-2} $year^{-1}$ in 1972 to about 490 gC m^{-2} $year^{-1}$ in 2001. This gain was explained by more effective fertilization and pest management, higher yielding cultivars, more favorable climate, shifts to productive crop types (e.g., wheat to corn), and economic factors. Further, a substantial interannual variation in cropland NPP in the conterminous U.S. was likely driven by changes in climate, but economic and management decisions (e.g., increases in irrigation) may have also contributed (Hicke et al. 2004). A 41% increase in cropland NPP in the southern U.S. was reported by Tian et al. (2010) for the period 1895–2007. Specifically, the annual mean cropland NPP based on a process-based model increased from 369 gC m^{-2} $year^{-1}$ for the period 1895–1950 to 520 gC m^{-2} $year^{-1}$ for the period 1951–2007. Drought events were identified as major threats to cropland NPP in this study (Tian et al. 2010). Using plant C allocation coefficients obtained from studies published after 1970, Bolinder et al. (2007) estimated a NPP of 537 gC m^{-2} $year^{-1}$ for small-grain cereals in Canada. However, estimating belowground NPP in this study was associated with considerable uncertainty.

Among species, paddy rice had a higher NPP per unit land globally in 1991 than wheat (376 *vs*. 271 gC m^{-2} $year^{-1}$) as rice is produced under more intensified growing conditions than wheat (Goudriaan et al. 2001). Further, small and coarse grains together accounted for 60% of global crop NPP whereas oil crop and sugar crops each accounted for 9%. Globally, the highest cultivated crop NPP was estimated for western Europe, eastern Asia including Japan, South Korea, and China, and the central United States (Monfreda et al. 2008). High crop NPP was also estimated for southern Brazil and northern Argentina. These are the only large cultivated areas with NPP greater than 1,000 gC m^{-2}. The highest rates of all exceed 2,000 gC m^{-2} in the countries of northern Europe bordering the North Sea. Especially intensive agriculture, usually associated with irrigation, attains comparable rates in a smaller areas scattered throughout the world, including New Zealand, Israel, Egypt, Saudi Arabia, California, Oregon, and Washington in the western United States, Java, and pockets of Columbia, and India, and Mexico (Monfreda et al. 2008).

Using a method based on production and crop-specific factors, Prince et al. (2001) estimated that in 1992 county-level NPP in the U.S. ranged from less than 200 gC m^{-2} $year^{-1}$ in North Dakota, Wisconsin and Minnesota to over 850 gC m^{-2} $year^{-1}$ in central Iowa, Illinois and Ohio (assuming 50% of dry matter is C). Further, corn grain NPP reached very high values (>850 gC m^{-2} $year^{-1}$) over extensive areas whereas NPP for corn silage was lower, and the lowest for wheat and barley (*Hordeum vulgare* L.). Large interannual variability in NPP (<350 to >700 gC m^{-2} $year^{-1}$) was observed in Iowa over a 15-year period from 1982 to 1996, with the lowest NPP in 1983 (which had an unusually wet spring), in 1988 (which was a drought year), and in 1993 (which experienced floods). In 1992, NPP in Iowa ranged from 232 gC m^{-2} $year^{-1}$ for soybean to 547 gC m^{-2} $year^{-1}$ for barley (Prince et al. 2001). The increases in total NPP for croplands in the Great Plains from 1991 to 1996 were largely related to high NPP for corn (Bradford et al. 2005).

Belowground Carbon Input

Surface plant residues are incorporated into the mineral soil by physical mixing and solubilisation, transport and subsequent adsorption (Lorenz and Lal 2005). The relative importance of these processes *vs*. root litter and rhizodeposition for profile SOC distribution and dynamics depend on climate, soil and vegetation types (Rumpel and Kögel-Knabner 2011). However, most studies on SOC dynamics in sub-soil and, in particular, on dissolved organic carbon (DOC), have been done in temperate forest soils, and there are only a few studies on DOC in temperate arable soils (Chantigny 2003). DOC represents only a small proportion of SOC but the most mobile and actively cycling SOC fraction (Bolan et al. 2011).

Tillage mixes surface crop residues into the mineral soil which may result in subsurface SOC formation (Wright et al. 2007). For example, stubble incorporated into the soil makes a significant direct contribution to SOC sequestration (Sanderman et al. 2010). Over and above the effects of crop management and translocation of DOC, crop rooting and belowground biomass also impact SOC formation in the subsoil. Further, in extensively managed croplands and, in particular, those with low or no tillage activity, there is a myriad of soil biota that affect directly and indirectly SOC inputs into subsoil (Wilkinson et al. 2009). For example, earthworms, termites, ants, arthropods but also roots efficiently bury SOC while forming voids in the form of burrows, nests, chambers, galleries and root channels. Direct inputs include litter sequestration into nests, termitaria, borrows, and bioturbator disposal of dead tissues. The earthworms and litter arthropods are effective in fragmenting and mixing surface residues into the soil. Indirect inputs of SOC into subsoils may occur by infilling of biogenic voids with litter, redistribution of SOC and subsurface mixing and burial (Rumpel and Kögel-Knabner 2011).

Plant Roots

Annual crops transfer about 20% of the C from photosynthates to the roots during the growing season (Whalen and Sampedro 2009). However, not all fluxes and the entire root biomass can be directly measured. In particular, inventory data on root biomass are uncertain due to spatial and temporal heterogeneity, uneven sampling and methodological differences among studies (Robinson 2007). For example, only 64% of wheat roots were recovered by direct sampling after ^{13}C-labelling (Subedi et al. 2006). Yet, direct measurements of complete root profiles are rarely reported in the primary literature (Schenk and Jackson 2005).

About 50% of the C translocated belowground in croplands is used for root growth (Johnson et al. 2006). The root biomass ranges between 6% of the shoot biomass for sunflower (*Helianthus annuus* L.) and 50% for barley (Whalen and Sampedro 2009). About 30% of the C translocated belowground moves into the rhizosphere or is released during root respiration (Johnson et al. 2006). Further, as much as 30% of NPP may be transferred to mycorrhizal fungi. In soybean, about 20–30% of photosynthates are allocated to support rhizobia in nodules (Whalen and Sampedro 2009).

The root C is preferentially stabilized in SOC relative to surface residue C (Rasse et al. 2005). For example, root-derived C from hairy vetch (*Vicia dasycarpa* Ten.) is stabilized three times more readily in soils under corn cropping systems than is surface residue-derived C (Kong and Six 2010). Similarly, Bolinder et al. (1999) and Rasse et al. (2006) estimated that the humification ratio (i.e., SOC/C input ratio) of corn residues is about 0.12 for shoots and 0.19 for roots. Further, the humification ratio for barley (*Hordeum vulgare* L.) residues is 0.12 for shoots and 0.16 for roots (Broadbent and Nakashima 1974). For different crops, the contribution factor of roots *vs.* shoots to total SOC (i.e., [root-derived C/total root C input]/[shoot derived C/total shoot C input]) is 2.4 (Rasse et al. 2005). Thus, roots are a more important SOC source, in particular, in subsoils than shoots. However, the root C flux into soil is poorly understood because accurate measurements of root exudation and root cell sloughing are difficult (Rumpel and Kögel-Knabner 2011). Further, it is not known what root properties influence SOC stability or rhizodeposition (Mendez-Millan et al. 2010). While soil C inputs in annual crops can be modeled based on root-to-shoot ratios the above- and belowground biomass in perennial species may not be very representative of C inputs to the soil (Gill et al. 2002).

To overcome difficulties in measurements of root-derived C, belowground C inputs are often estimated from root-to-shoot ratios (Sanderman et al. 2010). However, much current knowledge about root-shoot biomass relations is based on unreliable data. Specifically, by omitting unreliable data, higher estimates for median root-to-shoot ratios are calculated than previously obtained. Accordingly, vegetation-specific root-to-shoot ratios must be used to predict root biomass (Mokany et al. 2006). Also, with regard to belowground C inputs, the often used static measurement of live root biomass may result in inaccurate estimates because a significant fraction of total root biomass may be short-lived fine roots. For example, turnover times of tree fine roots is 1 year or less based mainly on minirhizotron experiments and between 1.3 and 32 years based on changes in C isotopic ratios with time (Pritchard and Strand 2008). Minirhizotron measurement generally underestimate the turnover times whereas estimated fine root turnover times based on isotopic studies are always too high (Guo et al. 2008). The non-woody roots of many plants in croplands which lack secondary growth and cell wall thickening are short-lived (Hodge et al. 2009).

Not all of the about 40% net fixed C allocated belowground is used for root growth. Specifically, about 50% of the C allocated belowground (19% of net fixed C) is retained in root biomass and roughly 27% of C allocated to roots (11% of the net fixed C) is rhizodeposition (Jones et al. 2009). For example, when the total belowground C allocation is considered, the estimated root-to-shoot ratios of 0.50, 0.30 and 0.33 for wheat, corn and soybean increase to 0.82, 0.55 and 0.62, respectively

(Johnson et al. 2006). However, most isotopic labelling studies used to quantify the amount of photosynthate partitioned belowground have focused on young plants at a vegetative stage but partitioning is strongly affected by plant age. For example, the partitioning of C to the rhizosphere decrease by 43%, 28%, and 20% for roots, rhizosphere respiration and soil residues, respectively, for annual plants between plant ages ranging from 28 to 600 days (Nguyen 2003). The root-to-shoot ratios for wheat, corn, soybean and lupin (*Lupinus* L.) decrease from about 0.4–0.6 during vegetative growth phases to as low as 0.1 at flowering (Gregory et al. 1997). Further, almost half of the published data on rhizodeposition are for wheat and ryegrass, and 76% of the studies are related to only five crop/grassland species. Thus, the knowledge of C rhizodeposition, in particular, in mixed plant communities is scanty (Jones et al. 2009).

Rhizodeposition

The C flow at the soil-root interface is bidirectional with C being lost from roots and taken up from the soil simultaneously (Jones et al. 2009). Rhizodeposition describes the release of organic C compounds by roots which results in dramatic changes in the physical, biological and chemical nature of the soil. Specifically, rhizodeposition fuels the soil microflora leading to an abundant microbial population in the rhizosphere which is involved in the great majority of soil biological activity (Bais et al. 2006; Nguyen 2003). Rhizodeposition processes include (i) loss of root cap and border cells, (ii) death and lysis of root cells, (iii) flow of C to root-associated symbionts living in the soils (e.g., mycorrhiza), (iv) gaseous losses, (v) leakage of solutes from living cells (root exudates), and (vi) insoluble polymer secretion from living cells (mucilage). Most C lost during root growth is in the form of complex polymers. For example, root exudates contain a diverse array of C-containing primary and secondary metabolites aside ions, free oxygen and water, enzymes and mucilage (Bais et al. 2006). However, the relative importance of each rhizodeposition process is unknown as it is extremely difficult to discriminate between them in both space and time. The C flow in the rhizosphere is extremely complex, being highly plant and environment dependent and also varying both spatially and temporally along the root. Thus, amount and type of rhizodeposits is highly context specific (Jones et al. 2009).

Rhizdeposition has direct and indirect effects on C sequestration in croplands as root exudation, in particular, may represent a significant C loss for crop plants. The magnitude of photosynthates secreted varies with soil type, age, and physiological state of the plant, and nutrient availability (Bais et al. 2006). The root exudates may affect plant-plant, plant-microbe and plant-nematode interactions. In particular, plant-microbe interactions in the rhizosphere may contribute to C sequestration in plant and soil as micronutrient acquisition is enhanced by the stimulation of microbial communities through exudates. Rhizodeposition may range between less than 10% of the net C assimilation by a plant to 44% of a nutrient-stressed plant's total C (Bais et al. 2006). However, many of the published data on C flow in both soil and roots are severely biased (Rees et al. 2005; Kuzyakov 2006). Nevertheless, from the about 1.5–2.2 Mg $\mathrm{C\,ha^{-1}}$ allocated belowground by cereals during the vegetation period roughly 0.4–0.6 Mg $\mathrm{C\,ha^{-1}}$ enters the soil in the form of rhizodeposition (Jones et al. 2009; Kuzyakov and Domanski 2000). However, these estimates are highly uncertain as the partitioning of rhizosphere respiration from mycorrhizal respiration is almost impossible. Also, whether root exudates directly contribute to the SOC pool is under discussion (Sanderman et al. 2010). Due to the rhizosphere priming effect the decomposition rate of soil organic matter (SOM) in the rhizosphere may increase three- to five-fold in response to root exudation but the response of the subsoil SOC pool is less well known (Kuzyakov 2002, 2010). For example, laboratory studies by Fontaine et al. (2007) indicated that adding an additional energy source similar those in rhizodeposits to the subsoil prime microbial decomposition of subsoil SOC. However, the stimulation of stable subsoil C decomposition in the field by addition of labile material may be small (Sanaullah et al. 2011). Further, the subsoil priming effect is not always observed but an increase in the subsoil SOC pool from inputs of root exudates is likely (Salomé et al. 2010). Also, complex compounds derived from root turnover may contribute indirectly to the SOC pool by enhancing aggregation (Rees et al. 2005).

Mycorrhizal Fungi

Mycorrhizal fungi live in symbiosis with the roots of plants. Almost all agricultural crops are mycorrhizal plants with the exception of the Brassicaceae family (cauliflower [*Brassica oleracea* L. Botrytis group], cabbage [*Brassica oleracea* L. Capitata group], canola [Brassica campestris L.], and others; Whalen and Sampedro 2009). Soil yeasts may affect the colonization of crop plant roots by arbuscular mycorrhizal (AM) fungi (Botha 2011).

AM fungi are important for sustainable crop production (Sen 2003). For example, AM fungi may facilitate uptake of up to 80% of the P, 25% of the N, 10% of the K, 25% of the Zn and 60% of the Cu requirements of the host plant (Marschner and Dell 1994). Other functional benefits provided by AM fungi are pathogen protection and facilitation of water uptake by the host plant (Verbruggen and Kiers 2010). AM fungi receive about 3–20% of photosynthate from their host plant, and the hyphae outside of the plant root play a major role in C translocation into the soil (Treseder and Cross 2006). This may enhance SOC sequestration as C is translocated away from the high respiratory activity around the root. The large network of hyphae outside of the root may represent up to 15% of the SOC pool (Leake et al. 2004).

AM fungi produce the glycoprotein glomalin within their hyphal walls (Wright and Upadhyaya 1996). Glomalin is deposited in the soil as the hyphae senesce, and may comprise 0.7–2.4% of the SOC pool in agroecosystems and enhance soil aggregation (Purin and Rillig 2007; Treseder and Turner 2007). Thus, glomalin may represent a reasonably large C influx to SOC in the order of tens to hundreds $\mathrm{g\,C\,m^{-2}\,year^{-1}}$ with relatively slow turnover rates (Rillig et al. 2003; Treseder and Turner 2007). However, glomalin cannot be directly isolated and methods for measuring soil glomalin have limitations (Schindler et al. 2007). Thus, glomalin is present in measurable but modest quantities in most cropland soils. Important may be indirect effects of mycorrhizal fungi on SOC storage as hyphal growth improves soil structure and aggregate stability (Sanderman et al. 2010). Thus, AM fungal hyphal abundance, soil structure and C storage in croplands may be closely correlated (Wilson et al. 2009). In addition to contributing to SOC gain, mycorrhizal fungi also contribute to direct SOC loss through decomposition (Talbot et al. 2008). However, compared to AM fungi ericoid mycorrhizal (ERM) and ecto-mycorrhizal (ECM) fungi have more extensive enzymatic capabilities and are, thus, more efficient decomposers (Smith and Read 2008). In summary, mycorrhizal fungi may moderate SOC dynamics in croplands by enhancing SOC sequestration and by the formation and maintenance of soil aggregates (Talbot et al. 2008; Wilson et al. 2009). Further, AM fungi facilitate shifts in the structure and function of host plant communities by symbiotic interactions (Cameron 2010).

14.2.2 Cropland Soil Carbon Sequestration

The net balance between soil C inputs and losses determines the capacity of a crop soil to sequester C. The activity of soil microorganisms and fauna leads to decomposition of OM and mineralization to CO_2 (Whalen and Sampedro 2009). Abiotic processes such as photodegradation of surface OM may also contribute to decomposition. Photodegradation is the decomposition of OM compounds directly by solar irradiance which increases CO_2 fluxes through either microbial facilitation or more important photochemical mineralization (Rutledge et al. 2010). Photodegradation may be an important contributor to CO_2 loss from croplands where SOC, litter and/or standing dead material are exposed to solar irradiance. Thus, photodegradation occurs in croplands of arid and semi-arid areas, bare burnt areas, sparsely vegetated croplands like shrublands, croplands after cultivation or harvest especially when crop residues are left on the surface, during prolonged drought and croplands with large amount of exposed standing dead material like croplands under no-tillage (Rutledge et al. 2010). Photodegradation of SOM may also be an important process after crop residues have been removed or where erosion exposes SOM to solar radiation (Feng et al. 2011). For example, photo-oxidation increased the solubility of SOM but did not substantially affect the organochemical composition of corn litter. Thus, photo-oxidation may contribute to soil C loss through leaching and oxidation (Feng et al. 2011).

Net C losses from croplands may also occur by soil erosion. This involves (i) detachment and breakdown of soil aggregate, (ii) transport of soil particles by runoff water or wind, (iii) redistribution of eroded material over the landscape, and (iv) deposition of eroded material in depressional sites and protected areas (Lal 2004). Historically, crop soils have lost significant amounts of SOC by erosion (Lal 2003). However, eroded SOC can also be a net C sink relative to the atmospheric CO_2 pool (Van Oost et al. 2007). Specifically, erosion can induce changes in soil-atmosphere C exchange via altered patterns of oxidation of SOC remaining at the eroding site and of the SOC originally present at the depositional site as well as via SOC production at both locations (Billings et al. 2010). Three key mechanisms can alter the flux of C between the soil and the atmosphere. First, dynamic replacement describes the partial replacement of the depleted SOC pool at the eroding site by newly assimilated C. The continued C input and decrease in SOC available to decomposition can lead to a net C gain (Van Hemelryck et al. 2009). Secondly, topsoil SOC is buried and decomposition reduced. Specifically, the decomposition rate of SOC in depositional settings can be reduced due to a combination of physical and chemical processes, such as increased soil wetness, limited aeration, compaction and physical protection of the deposited soil material within newly formed aggregates, leading to a long-term preservation of buried C (Quinton et al. 2010). The third mechanism affecting the C flux is transport and increased decomposition. Specifically, aggregates may breakdown by the disruptive energy of forces applied to the soil by water erosion (raindrop impact, the shearing force of flowing water and collision with other aggregates). This process of disaggregation exposes previously protected SOC to microbial decomposition and combined with a relatively greater proportion of labile SOC within larger soil aggregates may lead to rapid mineralization of this easily decomposable C following water erosion (Van Hemelryck et al. 2009). Thus, erosion can induce a net C sink or source (maximum net source and sink of 1.1/3.1 Pg C year^{-1}, respectively), depending on management practices, the extent to which SOC oxidation and production characteristics change with erosion, and the fate of eroded SOC (Billings et al. 2010).

Some cropland C may be lost by emission of CH_4 (Denman et al. 2007). Among the biogenic CH_4 sources are the anaerobic decomposition of OM in wetland soils emitting 100–231 Tg CH_4 year^{-1}, flooded soils under rice cultivation emitting 31–112 Tg CH_4 year^{-1}, and crop residues under very wet field conditions (Greenhouse Gas Working Group 2010). Microbial methanogenesis is the dominant soil process generating CH_4 carried out by a group of anaerobic Archaea converting substrates produced by fermentation of organic macromolecules (Singh et al. 2010). However, a large proportion of CH_4 produced in drier soil is often consumed by methanotrophic bacteria before escaping to the atmosphere (Reay 2003).

Cropland soils may also lose small amounts of C by emitting VOCs, compounds which may also interact with C cycling in soils (Gray et al. 2010; Insam and Seewald 2010). Plants, microorganisms and animals are the main producers of VOCs in soils (Laothawornkitkul et al. 2009; Leff and Fierer 2008; Loreto et al. 2008). Decomposing litter including the microbial metabolism has been identified as a major VOC source in laboratory incubation studies (Gray et al. 2010). However, the relative contributions of abiotic and biotic sources to soil VOC emissions in croplands are not entirely known (Leff and Fierer 2008).

Carbon losses through leaching are important for the C balance of croplands (Kindler et al. 2011). For example, leaching losses of biogenic C (DOC plus biogenic dissolved inorganic carbon or DIC) increase the net C loss from European cropland soils by 25%. Leaching of DOC from European croplands can be up to 4 g C m^{-2} year^{-1}. However, DOC concentrations are much smaller than DIC concentrations, particularly in subsoils (Kindler et al. 2011). Potentially leachable C compounds are derived mainly from plant litter and SOM but also from root exudates and the microbial biomass (Kalbitz et al. 2000). Plant residues are the major DOC source in cultivated soils (Bolan et al. 2011). The addition of manure and sewage sludge increases the amount of DOC in soils. DOC is produced largely by the activity of the soil biota as litter is physically and chemically altered during decomposition, and substances are leached from litter and soluble compounds are formed. Abiotic processes (i.e., desorption and dissolution from the potentially leachable C) control the DOC concentration of the soil solution. DOC compounds can be grouped into labile DOC compounds such as simple carbohydrates, low molecular weight organic acids and proteins, and amino sugars (Marschner and Kalbitz 2003). Recalcitrant DOC compounds consist of polysaccharides, other plant compounds and/or microbially derived degradation products. Only the mobile DOC fraction in macro- and mesopores is subjected to convective transport by seepage. DOC in micropores is immobile and interacts with the mobile fraction by diffusion. DOC in the soil solution may be decomposed or removed from solution by various processes (Kalbitz et al. 2000). Thus, DOC concentrations strongly decrease with increase in soil depth and small amounts of leached C is advectively transported to aquatic ecosystems (Sanderman and Amundson 2008). Leaching of DOC from subsoils is controlled, in particular, by sorption to poorly crystalline Fe and Al (hydr) oxides with a high specific surface area (Kindler et al. 2011). Further, DOC leached from soils may partly be retained in the vadose zone before reaching aquifers. In general, recently deposited crop residues and application of organic amendments such as biosolids and manures are the most important sources of DOC in arable soils (Bolan et al. 2011). Minor sources are root decay, exudates and microbial metabolites. However, very few studies have been published on boreal and tropical agroecosystems (Chantigny 2003).

14.2.2.1 Stable Soil Organic Matter and Humic Substances

Historically, SOM has been thought to consist of parts with a molecular structure that decomposes easily and other dark, amorphous parts that decompose slowly if at all (Kleber and Johnson 2010). The molecular properties of these dark ‘humus compounds’ have been thought to render them refractory (Alexander 1965). During the humification process, humus compounds supposedly evolve into polymeric macromolecular humic substances, i.e., the part of OM in soils most resistant to microbial attack because of its complexity. However, it is questionable whether operationally defined humic substances artificially prepared by an alkaline extraction procedure can serve as models of SOM fractions (Baldock and Nelson 2000). Materials extracted from soil with alkali do not occur as such in natural soils and there is no molecular-level evidence for the existence of distinct humic molecules in soils (Kelleher and Simpson 2006; Lehmann et al. 2008). Thus, SOM can be rather defined conceptually as a mixture of organic compounds in various stages of decomposition (Kleber and Johnson 2010).

The humification process does not create stable organic macromolecules out of labile plant litter precursor materials. SOM is not a stable (recalcitrant, refractory) product of secondary syntheses in soil as previously suggested (Stevenson 1994), and not the true end product of decomposition which is CO_2 (Kleber and Johnson 2010). All plant residues decay rather rapidly in aerobic soils with adequate moisture and temperature, and even lignin is more or less transformed with only parts of lignin accumulating and potentially being stabilized in soils (Hofmann et al. 2009; Thevenot et al. 2010; Vancampenhout et al. 2009). Some fresh plant C is reworked by the biotic community, and the remnants of soil biota attach to mineral surfaces and may be protected against decomposition for centennial to millennial timescales (Kleber and Johnson 2010). Seemingly labile OM persists or is stable in soil because of sorptive protection, aggregation, occlusion and facultative nonutilization (Ekschmitt et al. 2008; Salomé et al. 2010). Thus, whether recalcitrant OM exists in soil aside black carbon (BC) due to inherent molecular characteristics is a matter of debate (Kleber 2010a,b; von Lützow and Kögel-Knabner 2010). However, important for SOC sequestration is the fact that simple and relatively fresh OM inputs can directly contribute to the stable SOM pool as a slow aging process to produce stable SOM is apparently not required (Sanderman et al. 2010).

14.2.2.2 Decomposition

Decomposition is second only to photosynthesis in driving the quantity of C cycled through ecosystems (McGuire and Treseder 2010). The unharvested remnants of crops such as leaf, stem and root tissue not removed from the field and the die-back of legumes such as leaves and stems are subject to decomposition (Whalen and Sampedro 2009). The residues may be decomposed rapidly by a bacterial food chain or slowly by a fungal food chain depending on tillage practices as litter placement can strongly influence the decomposer community composition and decomposition rates (Beare et al. 1992). In particular in no-till croplands, macrofauna such as earthworms fragment litter and redistribute it in the soil profile.

Decomposition is the physical and chemical breakdown of dead OM of animal, microbial and plant origin that emit C into the atmosphere and release nutrients in forms that can be used for microbial, plant and animal production (Chapin et al. 2002). Litter decomposition is the result of the three interlinked processes leaching, fragmentation and catabolism (Cotrufo et al. 2009; Sanderman et al. 2010). Decomposition is driven primarily by bacteria and fungi accounting for up to 90% of the total decomposer biomass, and about 85–90% of OM decomposition is mediated microbially. Fungi mainly decompose fresh plant material whereas gram-positive bacteria consume both fresh and older SOM (Amelung et al. 2008). About 10–15% of decomposition is performed by soil animals (Wolters 2000). Decomposition is a key ecological process for maintaining supply of most plant-essential nutrients. For example, nutrient recycling via decomposition may account for >90% of plant-available N and P, and for >70% of plant-available K and Ca in natural ecosystems (Chapin et al. 2002).

Leaching is most important during tissue senescence and when plant litter is deposited at the soil surface. Mainly labile compounds such as low-molecular weight soluble sugars, polyphenols and amino acids are leached from plant litter (Berg and McClaugherty 2008). Some of the leached and solubilised material may be absorbed by soil organisms. However, depending on adequate water supply a major portion of leached OM may be transported deeper into the mineral soil and adsorbed to SOM and soil minerals, or transported into adjacent aquatic ecosystems. Soil fauna also partially solubilises fresh plant and fragments residues, and facilitates the establishment of microbial decomposers, for example, by mixing residues into the soil for more intimate contact with decomposer microorganisms (Wolters 2000). However, soil fauna has only a limited ability to enzymatically decompose organic residues compared to soil microorganisms (Whalen and Sampedro 2009).

The main initial decomposers fungi are mostly concentrated closer to the soil surface (Chapin et al. 2002; Fierer et al. 2003). Fungal C is about 75% of total microbial residue C in arable soils, and fungal respiration is about 61% of microbial respiration in cropland soils (Joergensen and Wichern 2008). Fungal enzymes can break down virtually all classes of plant compounds, and hyphal fungi can adjust mycelium growth across considerable distances towards OM substrates (Ritz 1995). Fungi have higher C assimilation efficiencies (i.e., store more C than they metabolize) than bacteria (Singh et al. 2010). Thus, SOM in croplands favoring a fungal-dominated community is improved quantitatively (Six et al. 2006). In contrast to fungi, bacteria primarily decompose more labile substrates (Moorhead and Sinsabaugh 2006). However, bacterial biomass extends to deeper depths. Bacteria rely on solute transport and soil perturbation for their translocation towards organic residues (Jiang et al. 2005). Thus, the composition of the initial microbial community decomposing a residue may gradually shift through time concurrently with changes in residue composition (Hättenschwiler et al. 2005).

In addition to the primary resource plant OM, biomass of microorganisms and microfauna (Ø<0.2 mm) are themselves secondary resources for decomposition. The C polymers in fungal cell walls (chitin, melanin) are more resistant to decomposition than those in bacterial cell membranes and walls (phospholipids, peptidoglycan) (Singh et al. 2010). Thus, respiration rates in soils dominated by fungi are typically low which increases the potential for SOC sequestration (Six et al. 2006). Ultimately, however, most natural compounds are fully mineralized to inorganic forms under favorable environmental conditions (Kleber 2010a; Marschner et al. 2008). Specifically, about one-half to two-thirds of plant residues entering the soil may be decomposed in one year (Whalen and Sampedro 2009).

Less well known is the long-term fate of leaf and root residues in cropland soil as decomposition is often studied only for 1–2 years (Silver and Miya 2001). In contrast, some litter decomposition experiments in forests run for 10 years and longer (Harmon et al. 2009; Moore et al. 2011). Thus, studies monitoring crop litter decomposition for several years are required as some litter may decompose very slowly depending on environmental conditions and biochemical characteristics. For example, up to 65% of the initial *Triticum aestivum* L. leaf mass and up to

55% of the initial *Andropogon gerardii* Vitman root mass may remain after 10 years decomposing in a soil (Harmon et al. 2009). However, very slow mass loss in late-stage decomposition may occur only in cold northern latitudes or other suboptimal conditions as it does, for example, not appear to occur in warmer climates (Prescott 2010).

Most studies on above-ground litter decomposition use the litter bag method to follow the progressive loss of mass from fresh litter through time (Bocock and Gilbert 1957). This method is still widely accepted for quantifying and comparing litter decomposition rates (Torn et al. 2009). However, a compromise on mesh size based on the experimental question is needed and the method has its limitations as soil faunal activities may be excluded depending on mesh size (Prescott 2005). In particular, the process of litter fragmentation may be prevented by studying decomposition in mesh bags (Cotrufo et al. 2009). Another limitation is that decomposition is often studied using single-species litterbags but mixed-species litterbag decomposition studies indicate that ecosystem litter decomposition is not predictable from decay rates of component species (Gartner and Cardon 2004). Thus, available litter decomposition data for croplands should be used with caution and referred to the specific decomposition component process (i.e., leaching, fragmentation, catabolism) measured (Cotrufo et al. 2009). For example, the use of litter bags is suitable for testing the effects of plant protection products on OM decomposition in agroecosystems (Knacker et al. 2003).

14.2.2.3 Organic Matter Stabilization

Inputs of OM to soils can increase the SOC pool when they are stabilized and not completely mineralized to CO_2 (Kleber and Johnson 2010). Beyond the decadal timeframe selective preservation of relatively unaltered plant-derived compounds due to biochemical recalcitrance (i.e., aliphatic compounds such as lipids and waxes) may not be as an important long-term OM stabilization mechanism as previously thought (Krull et al. 2003; Lorenz et al. 2007). However, charring of OM may result in the formation of biochemically recalcitrant BC. The molecular structure of plant inputs and OM plays only a secondary role in determining C residence times over decades to millennia. The soil C stability is instead mainly an ecosystem property as it depends on the biotic and abiotic environment (Schmidt et al. 2011). The ^{14}C age of SOM fractions is not necessarily related to molecular structure or thermodynamic stability (Kleber et al. 2011). Thus, old C cannot be viewed as being composed of complex or recalcitrant compounds. Further, old and stable SOM is not necessarily biochemically recalcitrant (Kleber et al. 2011). Similarly, microbial derived OM may not be recalcitrant or composed of complex compounds. Thus, mean residence times of microbial biomarkers in soils do not exceed several hundred years (Amelung et al. 2008). Biomarkers are structurally unique biomolecules that retain their C skeleton information and can be used to determine their biological origins and/or environmental settings. However, microbial derived materials may play a special role in SOM stabilization as the SOM turnover appears to be a function of microbial ecology and the resource availability within a given physical soil environment (Kleber et al. 2011). Thus, processes which slow down mineralization are major centennial-scale stabilization mechanisms for SOM (Sanderman et al. 2010).

The two important groups of processes for long-term stabilization of OM are (i) processes which lead to physical protection, rendering OM spatially inaccessible to decomposers or their water-soluble degradative enzymes and (ii) organomineral complexes and organo-metal interactions, i.e., interactions of OM with minerals, metal ions, and other organic substances (Von Lützow et al. 2006). Physical protection may retard decomposition for decades to centuries whereas organomineral complexes or organo-metal interactions may be responsible for most of the highly stable (centuries to millennia) non-charred SOM. However, biochemical recalcitrance and physical protection may allow OM to remain in the soil longer, giving time for organomineral complexes to form (Six et al. 2000). Thus, SOM stabilization is a combination of short- and long-term processes (Nair et al. 2010). Any disruption of the stabilization process may result in decomposition of SOM even if it's thousands of years old (Ewing et al. 2006). In summary, the persistence of SOM is largely due to complex interactions between OM and its environment, such as the interdependence of compound chemistry, reactive mineral surfaces, climate, water availability, soil acidity, soil redox state and the presence of potential degraders in the immediate microenvironment (Schmidt et al. 2011).

Several processes may be the reason why OM is physically inaccessible to decomposers and extracellular enzymes. First, decomposers and OM are sparsely and heterogeneously distributed in soils but the mobility of most decomposer organisms is limited (Ekschmitt et al. 2008; Young et al. 2008). This process of spatial separation may control SOC dynamics, in particular, in the subsoil (Salomé et al. 2010). Thus, a proportion of SOC of any chemical composition located in biologically non-preferred soil spaces is not subject to decomposition (Ekschmitt et al. 2008). Secondly, pores <0.2 μm in diameter are too small for bacteria (typical dimensions 0.5–1 μm), fungi (3–10 μm), microfauna (diameter <0.2 mm), mesofauna (0.2–2 mm) and macrofauna (>2 mm) to enter (Wolters 2000; Young et al. 2008). As pore size decreases to <50 nm, enzymes may also be inhibited from entering (Zimmerman et al. 2004). Third, the hydrophobicity of OM, in particular, those of partially-oxidized OM may greatly reduce accessibility as microbial decomposition is governed by distribution of and accessibility to water (Bachmann et al. 2008). Fourth, soil aggregation at multiple spatial scales may cause inaccessibility of OM by creating strong diffusional limitations to enzymes, and also oxygen which is another important controlling factor of decomposition (von Lützow et al. 2006).

In cropland topsoils, soil aggregation may be most important for spatial separation between decomposers or their extracellular enzymes and OM (Sanderman et al. 2010). Thus, OM losses from crop soils increase when soil aggregates are disrupted (Rovira and Greacen 1957). Aggregates are formed when mineral particles combine with organic and inorganic substances (Bronick and Lal 2005). Microaggregates (20–250 μm in diameter) are formed from small microaggregates (<20 μm) primarily within macroaggregates (>250 μm; Oades 1984). The smallest microaggregate fraction <53 µm is usually referred to as silt + clay fraction. The size classes occur in an aggregate hierarchy. Macroaggregates are less stable with a life span of years, break up most easily by slaking, have the highest concentration of C and the youngest C on average. In contrast, the smallest microaggregates are most stable with a life span of decades, have the lowest concentration of C and the oldest C. The OM in microaggregates is relatively stable as their organomineral complexes are hydrophobic and clogging of micropores creates physical barriers for bacteria and enzymes which retards decomposition (Bachmann et al. 2008).

In cropland soils, fungal extracellular polysaccharides and hyphae are primarily responsible for the formation of macroaggregates which protect plant- and microbial-derived OM from decomposition (Six et al. 2006). Further, fine roots contribute to the physical enmeshment of soil particles in aggregates (Miller and Jastrow 1990). Creating and maintaining the stability of aggregates critically depends on soil fauna, especially earthworms (Edwards and Bohlen 1996). Stabilization of OM within macroaggregates may be restricted to surface horizons in croplands but stabilization within silt-size aggregates is also important in subsurface horizons (Moni et al. 2010). Protection of OM by aggregation against microbial decomposition is effective primarily for soils with a large pool of labile OM (Goebel et al. 2009).

Other processes contributing to OM stabilization in soils involve sorptive reactions with mineral surfaces (Fe-, Al-, Mn-oxides, phyllosilicates), metal ions, and other organic substances (von Lützow et al. 2006). The binding of OM on minerals differs in stability depending on the suite of soil minerals and solution chemistry (Sanderman et al. 2010). For example, the quantity and characteristics of clay minerals, and the presence of polyvalent cations such as Ca^{2+} or Mg^{2+} facilitate sorptive OM stabilization (Whalen and Sampedro 2009). Multiple layers of OM may range outward from the mineral surface with decreasing strength of association (Kleber et al. 2007). Thus, the outer layers may be most actively cycling mineral-stabilized OM (Torn et al. 2009). The destabilization of SOM at the mineral surface may occur by mineral dissolution or soil evolution rather than by direct degradation. For example, the evolution of metastable non-crystalline minerals into crystalline minerals with lower surface area and charge density during soil development reduces the ability to stabilize SOM (Torn et al. 1997). Most of the SOM in soils of croplands with poor structure after cultivation for many years is often associated with mineral surfaces and stabilized in organomineral complexes (Basile-Doelsch et al. 2009; Flessa et al. 2008). However, sandy soils do contain less organomineral complexes than clayey soils and their aggregates are typically weaker (Sarkhot et al. 2007).

14.2.2.4 Controls on Decomposition and Stabilization

Organic substrates for decomposition in cropland soils are plant, microbial and animal residues, rhizodeposits, animal manure and SOM (Whalen and Sampedro 2009). Litter decomposition is influenced by the physical-chemical environment, decomposer organisms and substrate quality (Swift et al. 1979).

Globally, rates of plant litter decomposition are mainly controlled by variations in litter quality (i.e., N, C:N, lignin (%), lignin:N, and P, K, Ca, Mg), microclimate, soil properties and microbial community composition (Zhang et al. 2008). Early rates of litter mass loss can be predicted from climate and litter chemistry (Prescott 2010). Litter chemistry has the most direct influence on decay rates with the acid-unhydrolyzable residue (AUR):N ratio as the most consistent predictor. AUR was formerly known as "lignin". Further, leaf dry matter content and specific leaf area are useful predictors of mass loss rates. Apparently, thresholds exist at which specific factors have an overriding influence on decomposition (Prescott 2010). In contrast to leaf litter decomposition, root chemistry is the primary controller of root decomposition whereas climate and environmental factors play only secondary roles (Silver and Miya 2001). Most of the plant litter compounds decompose rapidly whereas the remaining recalcitrant compounds decompose slowly. However, plant tissues are not decay-resistant as, for example, lignin, cutin and suberin have residence times in years to decades (Prescott 2010). The C dynamics in aboveground litter *vs.* SOM in mineral soil bear little resemblance as both litter quality and physical-chemical mechanisms control SOM protection and degradability (Giardina and Ryan 2000). Thus, plant litter decomposition does not produce SOM with slow turnover time as the link between litter quality and SOM is not well established (Torn et al. 2009).

Under the same environmental conditions, litter quality controls the decomposition rate. Litter of high quality decomposes faster than that of lower quality. For example, alfalfa (*Medicago sativa* L.) residues decompose faster than residues of grain sorghum (*Sorghum bicolor* [L.] Moench) and winter wheat (*Triticum aestivum* L. emend. Thell.) under the same environmental conditions (Schomberg et al. 1994). Thus, alfalfa litter supposedly has a higher quality than sorghum and wheat litter. However, there is neither a common definition nor a quantitative index of 'quality' (Cotrufo et al. 2009). As high energy and nutrient supply for microbial use is required for fast decomposition, a high quality litter has high nutrient (i.e., N and P) concentrations, a high proportion of easily degradable C-compounds (e.g., sugars), and low concentrations of substances inhibiting microbial activity (Swift et al. 1979). Thus, indices of litter quality under discussion are N, C:N, AUR:N, holocellulose:lignocellulose, leaf width and specific leaf area (Cotrufo et al. 2009; Prescott 2010). In later decomposition stages, factors controlling microbial degradation of lignin such as concentrations of N and Mn become key quality parameters (Berg and McClaugherty 2008). However, a general and quantitative index of litter quality/decomposability is lacking (Cotrufo et al. 2009).

Litter decomposition is a biological process (Prescott 2010). Thus, primary controls of decomposition are also microbial activity and ultimately enzymatic activity, except for the abiotic process of photodegradation, and initial litter communication and mixing by soil fauna (Sanderman et al. 2010). Decomposition occurs as a result of secretion of extracelluar enzymes by microorganisms (Prescott 2010). Thus, decomposition depends on the successful diffusion of enzymes to appropriate OM and successful diffusion of decay products back into a microorganism.

Adequate temperature, a water film, aeration, suitable pH, enzymes, diffusion conditions, and an accessible and susceptible substrate in the same place at the same time are requirements that decomposition occurs (Prescott 2010). The size, diversity and activity of the microbial community as well as interactions with large soil biota affect the decomposition rate (Whalen and Sampedro 2009). Soil fauna comprises of 10–20% of total soil biomass and control the abundance and diversity of the microbial community by predation (Cotrufo et al. 2009). However, the effects of fauna such as microarthropods on litter decomposition are poorly understood (Kampichler and Bruckner 2009).

Fungi use C more efficiently than bacteria, and fungi are composed of more recalcitrant structural C. Thus, fungal-dominated food webs in croplands may be associated with higher amounts of SOC than those dominated by bacteria (van der Heijden et al. 2008). The optimal soil pH for bacteria-mediated decomposition is 6.5–8.0 whereas optimal soil pH ranges between 5.5 and 6.5 for fungi-mediated decomposition. Due to a variety of interacting factors, decomposition is more rapid in neutral than in acidic soils (Chapin et al. 2002). Soil bacteria are more tolerant to anaerobic conditions than fungi and are, thus, probably responsible for most decomposition in anaerobic soils. Soil texture determines the habitat available for microorganisms and the amount of predation by larger soil organisms. Further, the activities of microbial decomposer may be limited by insufficient quantities of essential nutrients such as N, P, K and others required to sustain microbial growth (Whalen and Sampedro 2009).

Soil fauna mediate decomposition and typically increase rates of mass loss from litter and SOM (Ayres et al. 2009). Faunal species in soil include collembolans, mites, enchytraeids, isopods and earthworms. Soil fauna enhances decomposition by (i) partial digestion of OM and returning it to the soil, (ii) fragmentation of plant litter thereby increasing the surface area available for microbial colonization, (iii) bringing microbes and OM into direct contact (during gut passage and by dragging litter from the surface into the soil), and (iv) grazing on soil microbes (Bardgett 2005). After gut passage, OM may be either readily available or less available for decomposition (Fox et al. 2006; Osler and Sommerkorn 2007). However, only the effects of grazing on decomposition are better characterized (Ayres et al. 2009). OM is partially degraded in the gut of saprotrophic soil fauna and some OM is returned to the soil as faecal pellets. The OM returned with collembolan, isopod and termite faeces, and with earthworm casts is typically easier to decompose than the original OM which leads to accelerated decomposition. However, decomposition rates of earthworm casts decline rapidly and casts may enhance below-ground C sequestration over longer time scales (Martin 1991). Whether other faunal groups also contribute to fragmentation resulting in increased surface area available for microbial colonization is less well known.

Earthworms may increase the amount of OM in soil aggregates and, thus, stabilize soil C (Bossuyt et al. 2005). Earthworms play a key role in decomposition by their partial digestion and fragmentation of OM, by dragging litter into the soil and bringing microbial decomposers into direct contact with OM (Wolters 2000). Further, grazing by microarthropods and nematodes may alter decomposition rates by influencing activity and growth of soil bacteria and fungi, and through selective feeding (Ayres et al. 2009). For example, the oribatid mite *Scheloribates moestus* (Acari: Oribatida) which is abundant in many U.S. ecosystems, stimulate extracellular enzyme activity, enhance microbial respiration rates and increase water-extractable organic C during decomposition of corn litter (Wickings and Grandy 2011). The mites decrease the relative abundance of polysaccharides in decomposing litter. Further, the feces have a higher relative abundance of polysaccharides and phenols and a lower relative abundance of lignin compared to unprocessed corn litter. Thus, *S. moestus* may play a key role in soil C cycling dynamics (Wickings and Grandy 2011).

Soil properties are secondary controls but climate has an overriding control on OM decomposition and stabilization (Sanderman et al. 2010). Specifically, temperature and rainfall (soil moisture) exert important direct controls on soil microbial activity (Whalen and Sampedro 2009). Water addition through irrigation, for example, may increase decomposition rates of crop residues (Schomberg et al. 1994). In general, faster litter decomposition rates are measured under warmer and wetter conditions when soil temperature and moisture conditions are ideal for microbial activity. However, which climatic index is the best predictor of decomposition rate is under discussion (Cotrufo et al. 2009). In soils with high moisture content, C accumulation is the greatest as decomposition is more restricted than is NPP (Chapin et al. 2002). Further, soil C respiration rate indicative of SOM decomposition roughly doubles for a 10°C warming when microbial activity is not limited by substrate availability or soil moisture (Davidson and Janssens 2006). The carbon quality-temperature (CQT) theory of the temperature sensitivity of OM decomposition links the temperature sensitivity of old SOM to an increase in molecular complexity. However, old SOM fragments are not large, complex, polymeric humic macromolecules. Thus, old SOM may not be particularly sensitive to temperature increases as predicted by the CQT theory (Kleber et al. 2011). Indirectly, temperature may affect decomposition by altering soil moisture, and quantity and quality of OM inputs (Chapin et al. 2002). In summary, over large geographical areas climate (i.e., temperature and soil moisture) exerts the strongest controls on C losses from soil (Sanderman et al. 2010).

Any control or processes that constrain microbial and enzymatic activity such as aggregation and association with mineral surfaces affect the decomposition rate (Sanderman et al. 2010). The soil-aggregate formation and SOC stabilization are affected, in particular, by soil type, climate, landscape position, ecology, and anthropogenic factors (Christensen 1996). Soil texture, for example, plays a large role in the number and kind of primary organomineral complexes formed. Specifically, soils high in clay content exhibit strong aggregate formation and stability. Thus, clay is positively correlated with SOC and the importance of clay for SOC stabilization increases with increase in soil depth (Jobbágy and Jackson 2000). Texture exerts strong controls on soil C losses by reducing apparent decomposition rates through a range of stabilization mechanisms (von Lützow et al. 2006). Clay minerals reduce the decomposition rate of SOM as the higher the mineral surface area the higher may be the amount of OM that can be adsorbed and protected from microbial enzymes (Chapin et al. 2002). However, clay content is not the most important determinant of SOC storage in contrast to the common assumption (Kutsch et al. 2009). Rather than texture *per se* the reactivity of the mineral surface determines the protection of OM (Kleber et al. 2007). For example, SOC in tropical soils such as Oxisols is poorly correlated with soil aggregates as oxides are the key binding factors (Six et al. 2002). Thus, SOM storage and turnover are often closely related to the mineral properties, in particular, the poorly crystalline phases (Torn et al. 2009). Subsoils have a larger potential for long-term stabilization due to abundance of reactive surfaces, smaller and less diverse biotic communities, and limitations on oxygen diffusion compared to topsoils (Rumpel and Kögel-Knabner 2011). In contrast, environmental conditions in topsoils are more often favorable for decomposition because of large and diverse biotic community, adequate moisture, available nutrients, good aeration (Sanderman et al. 2010). However, the topsoil environment is less favorable for C stabilization due to scarcity of reactive mineral surfaces.

Temperature and soil moisture affect OM breakdown in aggregates through their effect on microbial activity. For example, aggregate turnover is slower in cold or dry climates but faster in moist and/or warm climates. However, very wet climates may reduce OM breakdown under anaerobic soil conditions. In clayey soils, wetting and drying cycles can increase aggregation (Horn and Smucker 2005). In temperate climates, freeze thaw cycles may also promote aggregate formation and SOC stabilization (Chen et al. 1997). Plant species in croplands can also affect aggregation. For example, N fixing species and mycorrhizal associations may boost microbial populations and result in higher levels of aggregate stability. Further, molecular-level properties of SOM under different plant species may have an impact on resistance to degradation in microaggregates (Bachmann et al. 2008). Higher levels of SOC and aggregation are observed under plants with great root density (Nair et al. 2010).

In summary, decomposition of a substrate and destabilization of SOM are accelerated when: (i) conditions are suitable for microbial activity, enzyme production and diffusion, (ii) OM is vulnerable to enzymatic degradation, (iii) microbial transformations that lead to the creation of recalcitrant products are constrained, (iv) chemical reactions that lead to formation of stabilized SOM are constrained, (v) physical protection of OM in soil aggregates is weak and, (vi) chemical protection of OM in soil is also weak (Prescott 2010).

14.2.3 The Net Balance of Cropland Soil Organic Carbon

At steady state and under similar soil and vegetation management, the C inputs to cropland soil and losses from it are approximately balanced depending on site-specific factors (Sanderman et al. 2010). Quantitative descriptions of SOM dynamics were previously based on the assumption that SOM consists of a single homogenous pool decomposing at varying relative rates (Shibu et al. 2006). Others consider SOM as comprising of heterogeneous components and decomposition of the components occurs at different relative rates. The components are represented by arbitrary SOM pools. For example, the ‘active pool’ decomposes on time scale of hours to months to years, and may consist of root exudates, microbial cell contents and some fresh litter compounds (Torn et al. 2009). The ‘intermediate’ or ‘slow’ SOM pool has turnover times in the range of decades to centuries, and may consist of structural components of plants more resistant to decay or OM stabilized by association with soil minerals or aggregate structures. The ‘passive’ or ‘millennial cycling’ SOM pool persists in soils for thousands of years and consists of highly stabilized OM, typically associated with soil minerals or stable aggregates (Torn et al. 2009). However, the molecular model for stable SOM is under discussion but the specific structure of OM may be responsible for long-term preservation of OM in soil mineral fractions (Clemente et al. 2011; Kleber et al. 2011).

In comprehensive SOM models essential elements are thoroughly understood based on the scientific knowledge (Shibu et al. 2006). In contrast, essential aspects are formulated in less detail in summary SOM models and these models are more suitable for applicative and predictive purposes. For example, the Rothamsted carbon model (RothC) and the CENTURY model are two of the most widely used summary SOM models (Coleman and Jenkinson 1995; Parton et al. 1987). RothC is purely concerned with soil processes and not linked to a plant production model (Falloon and Smith 2009). In contrast, CENTURY is part of a larger ecosystem model than simulates crop, grass and tree growth, and the effects of management practices on both plant production and SOM.

In RothC, quality of residue/litter entering the soil is defined using the ratio of decomposable plant material (DPM) to resistant plant material (RPM) (Falloon and Smith 2009). In contrast, CENTURY uses the lignin:N ratio to define litter quality. In RothC, the SOC is split into four active compartments that decompose by a first-order process and have their own characteristic rate constants. In the order of decreasing maximum decomposition rates, the active compartments are DPM, RPM, microbial biomass (BIO) and humified organic matter (HUM). RothC assumes that a small amount of SOC is resistant to decomposition (inert organic matter [IOM]). In contrast, the CENTURY SOM sub-model includes three SOM pools, two surface and two sub-surface litter pools, and a surface microbial pool. All pools decompose by first-order kinetics and have characteristic decomposition rate constants. The pools can be ordered in decreasing maximum decomposition rates as the soil metabolic litter pool, the surface metabolic litter pool, the active SOM pool, the surface microbial pool, the soil structural litter pool, the surface structural litter pool, the slow SOM pool and the passive SOM pool. CENTURY also simulates leaching of OM whereas RothC is not capable of modeling DOC loss from soils (Falloon and Smith 2009).

Both RothC and CENTURY are applicable to cropland ecosystems (Falloon and Smith 2009). Both simulate the effects of farmyard manure application on SOM but CENTURY can also simulate the impact of tillage, harvesting, organic amendments, irrigation, erosion and fire on SOM. Both models have rarely been applied to assess short-term aspects of the C cycle and soil C cycling on a small scale. The majority of applications have focused on evaluating RothC and CENTURY against datasets of changes in SOC over decadal to century time scales. Both models can simulate the effects of management on SOC in the long term. Large scale applications of both models have also been performed (Falloon and Smith 2009).

Both RothC and CENTURY models were originally developed to predict turnover of C in topsoils. However, RothC has been modified to a multi-layer model to describe also the turnover of C in the top meter of soil (Jenkinson and Coleman 2008). It is recognized that the soil profile cannot be treated as a homogenous unit to predict the acceleration of SOC decomposition by global warming. Both RothC and CENTURY fail to account for pH effects on SOM turnover (Falloon and Smith 2009). Most SOM models are unable to simulate SOC changes in permanently waterlogged, very dry, highly organic and recent volcanic soils. The SOM models may be further limited in their applicability to tropical croplands. For evaluating SOM models, there are relatively few long-term experiments related to land use change rather than land-management changes (Falloon and Smith 2009).

The confidence in SOM model predictions is particularly limited as SOC measurements from available long-term experiments are rarely replicated (Falloon and Smith 2003). Challenges in measuring SOC include obtaining representative undisturbed soil cores for different layer depths, accurately measure bulk density for conversion to SOC pool, and addressing the high spatial variation in SOC (Falloon and Smith 2009). Accurately measuring litter lignin concentration which is among the input variables for CENTURY is also challenging (Preston et al. 1997). Most importantly, the SOM pools in models are theoretical without measurable counterparts. Whether a totally inert IOM pool (RothC) receiving no C inputs and being resistant to decomposition exists is unclear as was discussed previously regarding the lability of BC and SOC stabilization (Hammes et al. 2008; Von Lützow et al. 2006). Further, uncertainties are associated with the use of radiocarbon (^{14}C) age as input data in CENTURY and output data produced by RothC (Falloon and Smith 2009). Specifically, old (^{14}C age) and stable SOM is not necessarily chemically recalcitrant (Kleber et al. 2011). Also, the simplifying assumption in many models that SOC pools are near equilibrium has been challenged (Wutzler and Reichstein 2007). For example, agricultural soils may never reach a theoretical equilibrium SOC level because of changing conditions (i.e., climate change, land-management change) and partial resets by disturbances such as erosion (Polyakov and Lal 2004; Bell et al. 2011).

It has been hypothesized that the SOC pool in a particular soil eventually saturates at a maximum C-saturation level depending on inherent physicochemical characteristics (Stewart et al. 2007). In soils close to their saturation value, any additional increases in C inputs may then remain as unprotected POM and rapidly cycled back to the atmosphere (Stewart et al. 2008). However, while data from some long-term agricultural field experiments indicate that SOC saturation does occur others did not observe saturation behavior even after many years of additions of OM by manure application (Stewart et al. 2007; Blair et al. 2006a, b). Even though the whole soil may not be saturated, the chemically and biochemically protected pools may be influenced by C-saturation behavior (Stewart et al. 2009). Once the chemically protected SOC pool is filled, added C may accumulate in the physically and in the non-protected fractions. Mineral-associated SOC pools, in particular, eventually saturate. The reactive mineral surface area is a finite resource in topsoils and C-saturation may occur (Séquaris et al. 2010). Thus, arbitrarily defined soil fractions may have different C-saturation dynamics (Stewart et al. 2009).

In summary, recent advances in mechanistic understanding of soils have not yet been incorporated into widely used models of SOM cycling (Schmidt et al. 2011). Current models assume a pool of organic material that will have an intrinsic decay rate. These models rely on simple proxies such as soil texture as a surrogate for sorption and other organo-mineral interactions. Further, litter quality (e.g., lignin:N ratios or structural C groupings) is used a proxy of partitioning plant inputs into pools of different turnover times. However, these parameters are not consistent with the observations that emerge. Also, global models largely ignore deep mineral soils (Schmidt et al. 2011).

14.2.4 Anthropogenic Drivers of Cropland Soil Organic Carbon

The SOC balance of croplands is a function of past and present agricultural technology and farming practice (Ciais et al. 2011). Examples are manure and tillage management. Technology may impact the soil C input directly by affecting the management of harvest residues and indirectly via effects on yield and NPP. Anything that increases crop biomass production such as rotations, nutrients, improved cultivars and irrigation (i.e., agricultural intensification) impacts the soil C input (Robbins 2011). As croplands are often intensively managed they offer opportunities to deliberately alter the SOC dynamics (Smith et al. 2008). However, C budgeting studies over croplands are scarce in comparison to studies on crop yields and on processes controlling plant and soil fertility (Ciais et al. 2011).

Improving cropland management, restoring degraded crop lands and cultivated organic soils greatly affect SOC dynamics (Smith et al. 2008). For example, improved agronomic practices that increase yields and generate higher inputs of residue C can lead to increased SOC storage. Practices include (i) using improved crop varieties, (ii) extending crop rotations, notably those with perennial crops which allocate more C below-ground, and (iii) avoiding or reducing use of bare fallow (Lal 2004; Smith and Conen 2004). Adding more nutrients, when deficient, can also promote soil C gains (Alvarez 2005). By providing temporary vegetative cover between agricultural crops, ‘catch’ or ‘cover’ crops also add C to cropland soils (Freibauer et al. 2004). Since soil disturbance tends to stimulate soil C losses through enhanced decomposition and erosion, reduced- or no-till agriculture often results in soil C gain, though not always (Govaerts et al. 2009). Cropland systems that retain crop residues (e.g., avoiding the burning of residues) also tend to increase soil C because these residues are the precursors for SOM, the main store of C in the soil (Smith et al. 2008).

The management of croplands is an important factor in altering SOC dynamics (Franzluebbers 2010). Conservation practices, in particular, have a great potential to increase SOC. Guiding conservation principles that can be globally applied are to (i) minimize soil disturbance, (ii) maximize surface cover by managing crops and crop residues and, (iii) stimulate biological activity through crop rotations, cover crops and integrated nutrient and pest management (Franzluebbers 2010). However, the mechanisms that govern changes in SOC after reducing tillage operations are less clear (Govaerts et al. 2009). Among factors that may play a role are root development and rhizodeposits, baseline soil C content, bulk density and porosity, climate, landscape position, and erosion/deposition history. Further, altering crop rotation may influence SOC by changing quantity and quality of OM input (Govaerts et al. 2009).

Cropland irrigation may affect SOC dynamics by altering crop yields and residue returns (Lal 2004). About 25% of the global harvested crop area (i.e., area of all major food crops and cotton (*Gossypium* L.) as well as those of perennial, annual and fodder grasses) was irrigated in 2000 (Portmann et al. 2010). Further, the irrigated harvested area for rice was 1 million km^2 while it was 0.7 and 0.3 million km^2 for irrigated wheat and corn, respectively. The average crop yield of irrigated cereals was 442 Mg km^2 while average yield of rainfed cereals was 266 Mg km^2 (Siebert and Döll 2010). Crop yields and residue returns and, thus, soil C input may decrease when crop irrigation is discontinued. For example, the global production of dates (*Phoenix dactylifera* L.), rice, cotton, citrus (*Citrus*) and sugar cane would decrease by 60%, 39%, 38%, 32% and 31%, respectively, if currently irrigated crops were not irrigated. Further, cereal production on irrigated land would decrease by 47% without irrigation (Siebert and Döll 2010). In contrast, drainage of agricultural lands in humid regions can promote productivity and hence SOC accrual (Smith et al. 2008).

Agroforestry refers to the practice of purposeful growing of trees and crops, and/or animals, in interacting combinations, for a variety of benefits and services such as increasing crop yields, reducing food insecurity, enhancing environmental services and resilience of agroecosystems (Ajayi et al. 2011; Nair et al. 2008). In the Tropics, agroforestry systems includes alley cropping, homegardens, improved fallows, multipurpose trees on farms and rangelands, silvopastoral grazing systems, shaded perennial-crop systems, shelterbelts, windbreaks, and taungya (i.e., growing agricultural crops during early stages of establishment of forestry plantations). In temperate regions, agroforestry practices include alley cropping, forest farming, riparian buffer strips, silvopasture and windbreaks (Nair et al. 2009). In particular, planting trees may affect SOC dynamics and the SOC pool may be higher compared to that of croplands, pastures or natural grasslands replaced by the agroforestry system (Nair et al. 2010).

Organic soils contain high C densities because of delayed decomposition under flooded conditions. Thus, when they are used as croplands organic soils must be drained (Freibauer et al. 2004). SOC is lost, specifically, from organic soils by deep drainage and intensive mechanical disturbance such as deep plowing. Potential alternative uses for organic soils are maintaining a more shallow water table and avoiding deep plowing as well as avoiding cultivation with potatoes (*Solanum tuberosum* L.) and sugar beets (*Beta vulgaris* L.) and instead cropping to permanent cultures (Freibauer et al. 2004).

14.3 Recarbonization of Cropland Soils

The depletion of the SOC pool by cropland cultivation can be partially reversed by recarbonization. Soil C losses may occur when land under other uses is converted for the cultivation of crops. For examples, about 25–30% of the SOC stored in the top meter of soil is released by cultivation of native soils, whether under forest or prairie vegetation (Houghton 2010). Specifically, in temperate regions conversion from forest to cropland and from grassland to cropland may cause the loss of 31 ± 20% SOC to 28.5 ± 13.5 cm depth and of 36 ± 5% SOC to 27.1 ± 11.1 cm depth, respectively (Poeplau et al. 2011). New cropland SOC equilibrium may be reached if at all 23 years after conversion from forest and 17 years following conversion from grassland (Bell et al. 2011). In tropical regions, SOC losses of 25% to 36 ± 4 cm depth and of 30% to 48 ± 8 cm depth occurred by conversion of primary forest to cropland or perennial crops, respectively (Don et al. 2011). Losses of 21% SOC to 39 ± 5 cm depth occurred when secondary tropical forest was converted to cropland but no changes were observed to 51 ± 9 cm depth when converted to perennial crops. When tropical grassland was converted to cropland, 10.4 ± 6.1% of SOC were lost to 38 ± 11 cm depth (Don et al. 2011). However, cropping practices, irrigation, use of fertilizers and different types of tillage affect changes in SOC density. Thus, both conversions of native soils to crop soils and cultivation of croplands may decrease SOC pools. Croplands can partially recarbonized through adoption of recommended management practices (RMPs) such as conservation tillage, residue mulching and use of cover crops, practices which all contribute to soil C accumulation and sequestration by an additional transfer of C from the atmosphere to the soil (Lal 2007; Powlson et al. 2011).

Agricultural systems can be specifically managed to enhance C sequestration (Power 2010). For example, a net gain of the SOC pool can be achieved by conversion of plow tillage (PT) to no till (NT) and other conservation tillage practices, along with crop residue mulch or cover crops (Lal 2009). The transfer of C from the atmosphere into both the soil inorganic carbon (SIC) and SOC pool for enhancement of soil C sinks can be accelerated (Macías and Arbestain 2010). This may be achieved by (i) favouring growth of crop biomass which is the major source for SOC, (ii) promoting and facilitating carbonation processes to increase the SIC pool, (iii) reducing erosional C loss from croplands and favouring pedogenesis for build-up of the soil profile C pool, (iv) developing OM-rich horizons, and/or (v) recovering degraded or contaminated crop soils to restore the soil C sink. Some of the C lost in the past from cropland soils by changes in land use and cultivation can be recovered through improved management, thereby withdrawing atmospheric CO_2 (Smith et al. 2008). Most agricultural soils contain 30–75% less C than their potential capacity as determined by soil, climate, terrain, drainage, land use, and soil and crop management practices (Lal and Follett 2009b). The term 'soil C sequestration' implies that the total C pool in the soil profile is increasing through managerial interventions aimed at transferring atmospheric CO_2 to the soil C pool by moderating either organic and/or inorganic transformations (Lal and Follett 2009b). Thus, practices that retains or returns more of the C captured by growing plants increases the soil C pool (Sanderman and Baldock 2010). Carbon sequestration in crop soil implies an additional transfer of C from the atmosphere to the soil and, thus, a genuine contribution to climate change mitigation (Powlson et al. 2011).

The rate of C sequestration in croplands with adoption of RMPs depends on soil texture and structure, rainfall, temperature, farming system, and soil management (Lal 2004). The SOC pool in croplands can be enhanced by increasing use efficiency of input, decreasing losses by erosion and leaching, and improving soil structure (Lal and Follett 2009b). Further, the SIC pool can be enhanced by application of biosolids, liming/application of cations, and conserving water in the root zone. However, the potential of SIC sequestration by pedogenic carbonate formation is less well known. Soils of irrigated croplands may sequester both SIC and SOC. The rate of soil C sequestration ranges from about 100–1,000 kg ha^{-1} $year^{-1}$ for SOC and 5–15 kg ha^{-1} $year^{-1}$ for SIC (Lal and Follett 2009a). However, some agricultural field trials indicate that the relative increase in soil C pool with adoption of RMPs is not an actual increase but rather due to a reduction or cessation of soil C losses (Sanderman and Baldock 2010). The global C sink capacity of agricultural pools is estimated to be up to 78 Pg C and can be filled at the potential maximum rate of about 1 Pg C $year^{-1}$. However, the attainable and actual cumulative global rate of soil C sequestration may be lower because of managerial, economic, and policy constraints (Lal and Follett 2009a). The duration of soil C sequestration may be 25–50 years (Lal 2004). With increasing saturation of the soil C pool the sink activity diminishes but sequestration may continue due to climate change and land-management change (Bell et al. 2011).

Common RMPs for SOC sequestration are mulch farming, conservation tillage, agroforestry and diverse cropping systems, cover crops, and integrated nutrient management, including the use of manure, compost, and biosolids (Lal 2004). Irrigation can be used to buffer against soil moisture deficits and to sustain crop productivity. Where irrigation is constrained, the selection of drought-resistant crops promotes SOC sequestration. Further, crop production can also be adapted to temperature. For example, wheat in North America is now cultivated in environments once considered too arid, too variable, and too harsh to cultivate (Olmstead and Rhode 2011). Rising atmospheric CO_2 concentrations may directly alter crop yield and the SOC pool (Ainsworth and McGrath 2010). Free-air CO_2 enrichment experiments indicate that soybean and rice grain yield may increase by 13% at 550 ppm CO_2. In contrast, grain yield of sorghum and corn are not expected to increase at elevated CO_2 when water supply is adequate. However, in the long term responses of row crop agroecosystems to management practices may be more important than responses to elevated CO_2 (Moran and Jastrow 2010).

14.4 Conclusions

By converting land under other uses to cropland, by cropland management and its intensification (i.e., fertilization, irrigation, mechanization) humans have altered SOC dynamics on 12% of Earth's ice-free land area. Up to 36% of SOC may have been lost from cropland topsoils. Thus, most cropland soils contain less C than their potential capacity as determined by soil, climate, terrain, drainage, land use, and soil and crop management practices. Photosynthesis is the major natural C input into cropland soils while direct input occur by addition of manure and organic residues. Harvest removes a major proportion of cropland NPP before it enters the soil. Natural crop soil C losses occur by decomposition, erosion and leaching. Cropland SOC can be increased by mulch farming, conservation tillage, agroforestry and diverse cropping systems, cover crops, and integrated nutrient management, including the use of manure, compost, and biosolids. Recent advances have improved our understanding of SOC dynamics and SOC persistence. However, the deep mineral cropland soils have been largely ignored. Further, it remains to be studied whether SOC-accreting crops (e.g., crop plants with a bushy and deep root system, perennial crops) can be cultivated to recarbonize cropland soils.

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