12 Soils

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KEY FINDINGS

1. Estimates for soil carbon stocks in the conterminous United States plus Alaska range from 142 to 154 petagrams of carbon (Pg C) to 1 m in depth. Estimates for Canada average about 262 Pg C, but sampling is less extensive. Soil carbon for Mexico is calculated as 18 Pg C (1 m in depth), but there is some uncertainty in this value (*medium confidence*).

2. Most Earth System Models (ESMs) are highly variable in projecting the direction and magnitude of soil carbon change under future scenarios. Predictions of global soil carbon change through this century range from a loss of 72 Pg C to a gain of 253 Pg C with a multimodel mean gain of 65 Pg C. ESMs projecting large gains do so largely by projecting increases in high-latitude soil organic carbon (SOC) that are inconsistent with empirical studies that indicate significant losses of soil carbon with predicted climate change (*high confidence*).

3. Soil carbon stocks are sensitive to agricultural and forestry practices and loss of carbon-rich soils such as wetlands. Soils in North America have lost, on average, 20% to 75% of their original top soil carbon (0 to 30 cm) with historical conversion to agriculture, with a mean estimate for Canada of 24% ± 6%. Current agricultural management practices can increase soil organic matter in many systems through reduced summer fallow, cover cropping, effective fertilization to increase plant production, and reduced tillage. Forest soil carbon loss with harvest is small under standard management practices and mostly reversible at the century scale. Afforestation of land in agriculture, industry, or wild grasslands in the United States and Canadian border provinces could increase SOC by 21% ± 9% (*high confidence*).

4. Large uncertainties remain regarding soil carbon budgets, particularly the impact of lateral movement and transport of carbon (via erosion and management) across the landscape and into waterways. By 2015, cumulative regeneration of soil carbon at eroded agricultural sites and the preservation of buried, eroded soil carbon may have represented an offset of 37 ± 10% of carbon returned to the atmosphere by human-caused land-use change (*medium confidence*).

5. Evidence is strong for direct effects of increased temperature on loss of soil carbon, but warming and atmospheric carbon dioxide increases also may enhance plant production in many ecosystems, resulting in greater carbon inputs to soil. Globally, projected warming could cause the release of 55 ± 50 Pg C over the next 35 years from a soil pool of 1,400 ± 150 Pg C. In particular, an estimated 5% to 15% of the peatland carbon pool could become a significant carbon flux to the atmosphere under future anthropogenic disturbances (e.g., harvest, development, and peatland drainage) and change in disturbance regimes (e.g., wildfires and permafrost thaw) (*medium confidence*).

Note: Confidence levels are provided as appropriate for quantitative, but not qualitative, Key Findings and statements.

12.1 Introduction

Globally, soils contain more than three times as much carbon as the atmosphere and four and a half times more carbon than the world's biota (Lal 2004); therefore, even small changes in soil carbon stocks could lead to large changes in the atmospheric concentration of carbon dioxide (CO_2). Despite their importance, however, stocks of soil organic carbon (SOC), which is the carbon component of soil organic matter (SOM), have been depleted through changes in land use and land cover and unsustainable land management practices associated with agriculture, grazing, and forest management. To better manage and sustain SOC stocks, a focused understanding of microbial and biogeochemical processes that interact in soils, regardless of land cover, to control soil carbon stabilization and destabilization is needed. Soil organic matter (the organic component of soil, consisting of organic residues at various stages of decomposition, soil organisms, and substances synthesized by soil organisms) also is considered a central indicator of

soil health because it regulates multiple ecosystem services that humanity derives from soils, including moderation of climate. SOM stores nutrients, increases water-holding capacity to promote plant growth, limits leaching of nutrients, and adds structure that improves drainage and reduces erosion (Oldfield et al., 2015).

The current best estimates for global SOC stocks are 1,400 ± 150 petagrams of carbon (Pg C) to 1 m in depth and 2,060 ± 220 Pg C to 2 m in depth (Batjes 2016). These values are derived from the Harmonized World Soil Database with corrections for underrepresented regions, including the Northern Circumpolar Region, using measured soil profiles and geospatial modeling. The resulting values are consistent with other global SOC pool estimates (Govers et al., 2013; Köchy et al., 2015). An estimated 90 to 100 Pg C is released by soils to the atmosphere as soil respiration each year, an efflux that represents both heterotrophic (approximately 51 Pg C) and autotrophic (approximately 40 Pg C) respiration (Bond-Lamberty and Thomson 2010; Hashimoto et al., 2015), roughly balanced by carbon incorporated into SOC from plant residues. This flux value can be compared to estimates from the most recent Intergovernmental Panel on Climate Change (IPCC) report that estimated the gross efflux from surface ocean water to the atmosphere as 78.4 Pg C per year (with a net sink of 2.3 ± 0.7 Pg C per year), carbon emissions from fossil fuel combustion and cement production as 7.8 ± 0.6 Pg C per year, and outgassing from freshwater as 1.0 Pg C per year (Ciais et al., 2013). Soil carbon storage and flux at a given location are controlled by variations in 1) soil-forming factors (Jenny 1941; McBratney et al., 2003; Mishra et al., 2010), 2) anthropogenic activities (Lal 2004), and 3) climatic forcings (Heimann and Reichstein 2008; Richter and Houghton 2011). Future change in the frequency of climatic extremes (Seneviratne et al., 2012) and land use and land management (Nave et al., 2013; Ogle et al., 2010; Wills et al., 2014) may alter SOC stocks and fluxes that affect land feedbacks to climate change, changing the magnitude of, or even reversing (i.e., change from sink to source), the land carbon sink (Friedlingstein et al., 2014).

Soils of North America store 366 to 509 Pg of organic carbon to 1 m in depth based on continental-scale analyses (Batjes 2016; Liu et al., 2013). Breakdown of SOC stocks by country are discussed in more detail later in this chapter. At the continental scale, nearly 75% of SOC stocks down to 1 m are found in the top 30 cm (Liu et al., 2013), which also is the portion of the soil profile most vulnerable to changes induced by land-use and land-cover changes, disturbance and extreme events, management practices, and climate change. Several knowledge gaps exist in the current ability to measure SOC stocks and fluxes across North America. Researchers employ diverse analytical methods to measure carbon concentration and take measurements at different depths; furthermore, many measurements lack bulk density estimates that are needed to calculate stock estimates. Most SOC stock estimates lack systematic uncertainty (i.e., error propagation) estimates. Consequently, this chapter shows many values of stocks and fluxes without companion uncertainty values. Therefore, significant risks exist for biased conclusions due to inadequate and uneven distributions of SOC profile observations, especially in permafrost regions (Mishra et al., 2013), for depths >1 m and in bulk density estimates for organic soils (Köchy et al., 2015). Recent updates to soil databases have improved coverage, but distributions of available samples across geographic regions are uneven and thus not sufficient to fully characterize SOC dependence on climate, edaphic factors, and land-cover types (Hengl et al., 2014; Mishra and Riley 2012). However, recent efforts, notably the U.S. Department of Agriculture's (USDA) Rapid Carbon Assessment (RaCA), will yield a much more consistent estimate of current soil carbon stocks (see Section 12.4.1, p. 479). Similarly, RaCA recently initiated a field-based soil carbon inventory for Mexico, and comprehensive stock estimates for different regions and land uses are forthcoming (see Section 12.4.2, p. 481).

Since cultivation of land began nearly 12,000 years ago, humans have been altering soil carbon stocks. Just since 1850, human degradation of soil worldwide may have resulted in a loss of 44 to 537 Pg SOC, largely through land-use change and conversion to agriculture (Lal 2001; Paustian et al., 1997). Globally, agricultural soils have lost 20% to 75%, or 30 to 40 megagrams of carbon (Mg C) per hectare (ha), of their antecedent SOC pool (Lal et al., 2015). In contrast, afforestation (the establishment of forest cover on land that previously did not have tree cover) and land restoration have the potential to recover depleted SOC stocks from the atmosphere (Lal 2004). For example, newly afforested lands cover 4 billion ha globally and have a carbon sequestration potential of 1.2 to 1.4 Mg C per year (Lal et al., 2015). Meta-analysis of afforestation effects on soil carbon storage in the United States and Canadian border provinces found that land conversion to forest from agriculture, industry, or wild grassland increased SOC by 21% ± 9% (Nave et al., 2013). The researchers found that the largest increase was in lands previously used for industrial purposes such as mining (173%), for areas with woody encroachment into unmanaged grassland (31%; see Ch. 10: Grasslands, p. 399), and for agricultural areas in the Northern Plains (32%; see Ch. 5: Agriculture, p. 229). Such SOC increases via afforestation and reforestation contribute to the net carbon sequestration by U.S. forests, currently estimated at 313 ± 40 teragrams of carbon (Tg C) per year (Lu et al., 2015).

12.2 Carbon Cycling Processes in Soils

Progress has been made over the last 10 years in understanding specific processes that determine the magnitude and direction of SOC stabilization and destabilization (see Figure 12.1, p. 473). This new information will not only help explain spatial patterns of SOC in North America, but also will help improve modeling of the large soil carbon pool in Earth System Models (ESMs). Outlined here are the processes that govern overall carbon stocks and fluxes through soils, from inputs through microbial transformations in the bulk soil and rhizosphere, and the protection mechanisms that govern the overall longevity of carbon in soils.

12.2.1 Precipitation

Overriding many soil carbon processes is the complicated role of precipitation and moisture on soil carbon stocks. Precipitation effects on SOC are complicated by the various and often opposing effects of precipitation on the various processes that control carbon stabilization and destabilization. On one hand, where moisture is limiting, increased soil moisture stimulates soil microbial activity, thus increasing soil respiration and destabilization of soil carbon. On the other hand, precipitation has strong effects on both vegetation type and plant production, and thus increases in precipitation in moisture-limited systems generally lead to increases in soil carbon through indirect effects on enhanced plant production, particularly increased root production (Jobbágy and Jackson 2000). In a global analysis (Jobbágy and Jackson 2000) total soil carbon content increased with precipitation and clay content and decreased with temperature. These results match numerous regional studies showing that precipitation in temperate ecosystems has a strong and positive relationship with SOC, likely through effects on total plant biomass, especially belowground biomass (Burke et al., 1989; Liu et al., 2012). Taken together these results suggest a greater response of plant production compared to decomposition from increased precipitation.

Several analyses have noted a wide divergence in estimates of soil carbon stocks from terrestrial biosphere models (Tian et al., 2015; Todd-Brown et al., 2013). Todd-Brown et al. (2013) noted that the parameterization of soil heterotrophic respiration was a significant cause of the discrepancy in model predictions, while Tian et al. (2015) suggested that mechanisms such as changes in the proportion of labile to passive soil carbon pools, as well as sensitivities of respiration to climate, are significant sources of uncertainty in the modeling estimates of soil carbon. Thus, more accurate biome-specific analyses of the effects of precipitation on soil respiration,

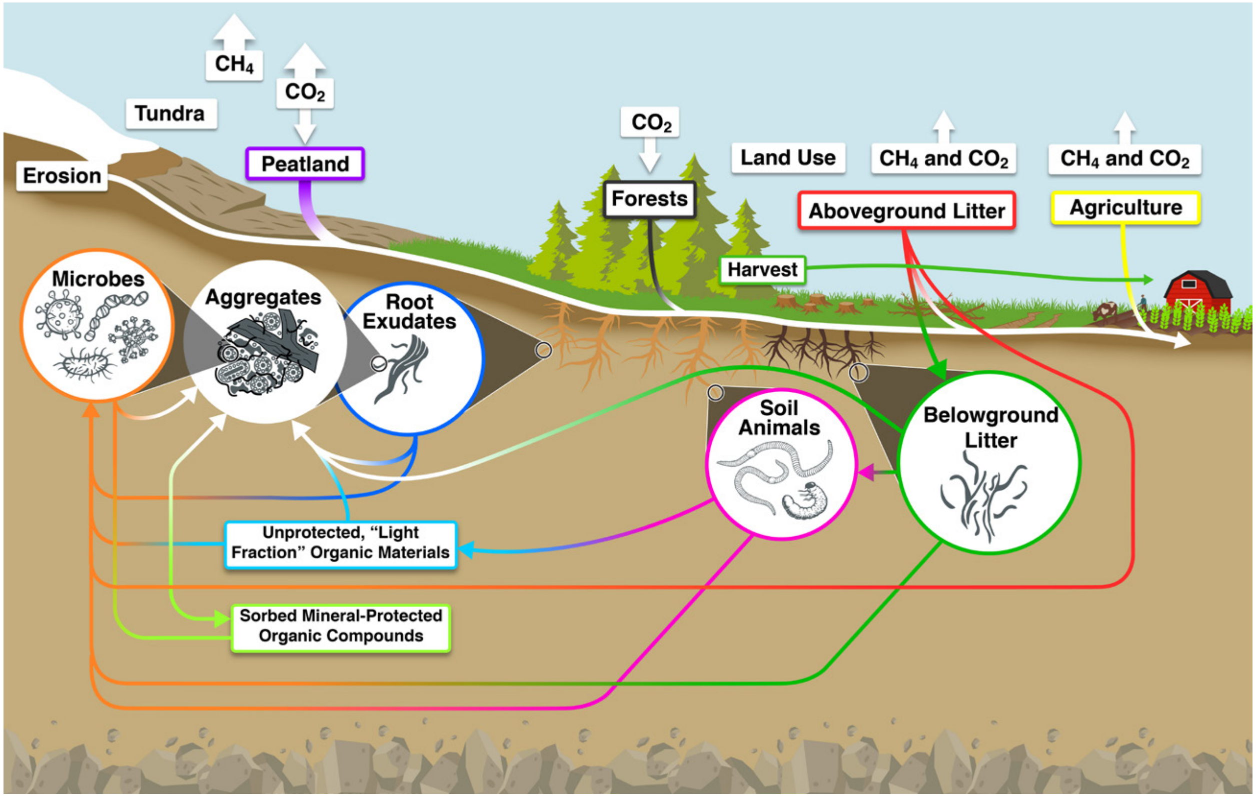

Figure 12.1. Processes Involved in Controlling Fluxes and Stabilization of Soil Carbon. A variety of soil animals and microbes can process plant litter that contributes to a pool of unprotected particulate organic matter (OM) with a relatively short turnover time. Alternatively, soil microbes also can process this litter into more stabilized forms such as aggregates or mineral-protected OM with relatively long turnover times. In this carbon pool, belowground litter appears to be preferentially stabilized, partly because of its proximity to both microbes and minerals. Root exudates may contribute to microbial carbon pools or to priming (i.e., the loss of mineral-protected soil carbon). Respiratory losses—occurring at all stages of biotic processing—can be affected by microbial carbon use efficiency and by conditions in the natural environment or those arising from land use. Not only can land use significantly affect both the quality and quantity of plant residues delivered to soils and their processing, it also can affect erosional losses and deposition. Climate change, especially in northern latitudes, may cause significant losses of soil carbon. (Key: CO_2, carbon dioxide; CH_4, methane.)

litter and root production, and vegetation type will be needed to improve soil carbon models.

12.2.2 Plant Litter Inputs

Many factors, including climate regime, atmospheric CO_2, land management, soil mineralogy and fertility, and nitrogen deposition strongly influence the structure of the plant community and thus the amount and quality of organic inputs (e.g., litter, wood, and root debris) to the surface of soils (Jandl et al., 2007; McLauchlan 2007; Smith et al., 2007). For example, elevated nitrogen deposition and high soil fertility generally increase plant shoot:root ratios and also decrease concentrations of plant protective compounds such as lignin (Haynes and Gower 1995; Luo and Polle 2009; Pitre et al., 2007). Chemical composition of litter, variably measured as carbon:nitrogen, lignin:nitrogen, or by the presence of complex aromatic compounds, has been shown to influence litter decomposition (Papa et al., 2013; Trofymow et al., 1995; Wardle et al., 2002), with high lignin or aromatic content observed to limit decomposition rates. However, the linkages among litter quantity, litter composition, and SOC stocks

are much less clear than would be expected due to other contributing factors. For example, several long-term litter manipulation experiments have shown that increased litter inputs do not always result in increased SOC storage (Lajtha et al., 2014a, 2014b; Mayzelle et al., 2014). Fresh carbon inputs can alter the decomposition of existing SOM because microbes, which play a major role as decomposers in soil ecosystems, will use the new inputs as fuel to decompose existing SOM (Bernal et al., 2016; Crow et al., 2009; Georgiou et al., 2015), resulting in a net decrease in SOC. Site-specific differences in soil mineralogy and microbial physiology also can influence the magnitude of response in SOC concentrations to changes in litter inputs (Geyer et al., 2016; see Section 12.2.3, this page). These kinds of interactions with soil minerals and microbes help to explain why chemical factors, such as lignin content, that are known to control litter decomposition do not always appear to be primary controls on SOC stabilization or destabilization (Rasse et al., 2006; Sulman et al., 2014). There also is evidence that root litter may be preferentially stabilized over shoot-derived litter (Iversen et al., 2008; Kong and Six 2010; Rasse et al., 2005; Russell et al., 2004). Thus, further research is needed to determine how changes in net primary production (NPP), vegetation, and litter quality due to rising atmospheric CO_2 concentrations will affect SOC stabilization in the future.

12.2.3 Soil Microbes

Soil microbes, including bacteria, fungi, and archaea, ultimately process all carbon inputs; consequently, microbes are referred to as "the eye of the needle through which all organic materials must pass" (Jenkinson 1977). The organic products and by-products of microbial decomposition, including microbial necromass, can accumulate in soils as SOM, and the chemistry of SOM is distinct from its source material including litter, roots, insect and animal necromass, and wood. The transformation from litter inputs through microbes and into SOM produces inorganic, carbon-containing gases such as CO_2 and methane (CH_4) through microbial respiration. Because of its important role in carbon transformation, the soil microbial community is key to understanding SOC stocks (Bernal et al., 2016; Guenet et al., 2012), even though the microbial biomass is typically only 1% to 2% of total SOM mass (Xu et al., 2013). Understanding microbial response to microclimate is key to understanding the carbon balance of soils under climate change, because soil balance under changing temperature and moisture is dependent on microbial community and physiological responses to changing temperature and moisture (e.g., Billings and Ballantyne 2013; Yan et al., 2016).

In addition to their direct role mineralizing SOM into inorganic gases, microbes contribute to physical mechanisms of SOC stabilization, indirectly affecting the rate and nature of SOC inputs from plants. A key mechanism of SOC stabilization is protection within soil aggregates (Six et al., 2002), and fungal mycelia and bacterial extracellular polysaccharides are important in forming and stabilizing these aggregates (Aspiras et al., 1971). SOC also is protected by chemical interactions with minerals, particularly silt and clay (Six et al., 2002), and microbes living on minerals may facilitate these interactions by depositing microbially derived carbon directly onto mineral surfaces (Uroz et al., 2015). Microbes can affect plant carbon inputs by regulating plant nutrient supply (Bever et al., 2010; van der Heijden et al., 2006), which affects plant community composition and the timing, mass, and properties of plant inputs of litter and exudates. Thus, although they compose a small fraction of SOC stocks, microbes play a central role in the SOC cycle, affecting inputs, storage, and outputs in diverse ways.

12.2.4 Macrofauna (Food Web)

Soil is home to millions of different organisms, from microorganisms to soil animals (fauna) such as microscopic roundworms (nematodes), tardigrades, rotifers, collembolans, mites, isopods, ants, spiders, and earthworms (Orgiazzi et al., 2015). These fauna exist in food webs containing multiple trophic levels—herbivores that feed directly on the roots of living plants, consumers that feed on living microorganisms associated with dead organic materials, predators that prey on other soil fauna, and plant or animal parasites and pathogens (Coleman and Wall 2015). Through soil bioturbation and feeding on plant roots, organic matter, and their associated microorganisms, soil animals are intimately involved in every step of SOM turnover and soil formation. Sometimes referred to as "ecosystem engineers," soil animals play a disproportionate role in the carbon cycle relative to their abundance and biomass. Carbon stocks of the soil fauna range from 0.3 to 50 kilograms of carbon per hectare, with desert soils containing the smallest faunal biomass and temperate grassland and tropical rainforest soils the greatest (Fierer et al., 2009). However, across biomes, the biomass of soil fauna typically represents less than 3% of the total biomass of living soil organisms, with soil microorganisms making up the majority. Despite their low biomass relative to soil microbes, soil fauna contribute significantly to carbon cycling through their regulation of microbial activity and through their physical mixing of organic materials and soil. The presence of soil fauna stimulates decomposition, respiration rates (i.e., CO_2 flux), and losses of dissolved organic carbon through leaching (de Vries et al., 2013). The positive impact of soil fauna on carbon cycling is attributed to organic matter fragmentation, which increases 1) the surface area available for microbial colonization; 2) the partial digestion of organic materials, enhancing their decomposability; 3) the direct contact of soil microbes with organic matter; and 4) the direct consumption of soil microbes—all impacts which stimulate microbial activity and the release of carbon and nutrients (Coleman and Wall 2015). However, one study found that the activity of earthworms increases carbon stabilization onto minerals to a greater degree than the increase in carbon mineralization, leading to net soil carbon increase (Zhang et al., 2013). Current ecosystem-scale models and ESMs typically overlook the significant effects of soil fauna on the carbon cycle, but guidelines for development of next-generation models call for explicitly incorporating soil food web properties and the responses of soil fauna to land use and climate change (de Vries et al., 2013).

12.2.5 Rhizosphere Interactions

The rhizosphere is defined as an area of soil where microbial activity is stimulated by the presence of roots. A substantial portion of plant biomass is located below ground in the form of roots. Estimates of belowground NPP based on root:shoot ratios assign 30% to 60% of total plant biomass to roots, depending on the biome (Bolinder et al., 2007; Rytter 2001). Regularly shedding sloughed cells and mucilage, roots exude a variety of simple carbon compounds into the soil immediately surrounding them (Hirsch et al., 2013). These root "exudates" comprise primarily organic acids, sugars, and amino acids (Hirsch et al., 2013; Jones 1998). These exudates can interact with minerals by sorption or can liberate organic compounds and nutrients for plant or microbial uptake (Dessureault-Rompre et al., 2007; Keiluweit et al., 2015). In general, the mass of soil in the rhizosphere makes up a smaller fraction (<40%) of total soil than does root-free soil, but it disproportionately affects carbon cycling. For example, microbial biomass, extracellular enzyme activity, decomposition, and mineralization rates are consistently higher in rhizosphere soil compared with those in bulk soil. Fungal hyphae can extend >40 cm away from roots (Finlay and Read 1986), extending the influence of root carbon past the rhizosphere (Zak et al., 1993). Dead root biomass is a substrate source for saprotrophic microbes and detritivores, while living roots are a source of carbon to mycorrhizal fungi.

Mycorrhizal material, shown to be a dominant pathway through which carbon enters the SOM pool, exceeds the input via leaf litter and fine-root turnover (Godbold et al., 2006). Mycorrhizae also may stimulate the decomposition of soil carbon to mine nutrients, paradoxically causing destabilization of soil carbon pools. The effects of mycorrhizae on soil carbon balance are thus complicated by the balance between carbon stabilization effects and soil carbon priming effects (Brzostek et al., 2015). However, recent research (Averill and Hawkes 2016;

Averill et al., 2014) demonstrated that ecosystems dominated by plants with symbiotic ectomycorrhizal fungi store more carbon in soils than ecosystems dominated by arbuscular mycorrhizae–associated plants.

12.2.6 Nitrogen Effects on SOM Dynamics

There are substantial interactions between biogeochemical cycles of carbon and nitrogen. Human activities (e.g., fertilizer production, fossil fuel combustion, and industry) have substantially increased nitrogen supply to ecosystems (Vitousek et al., 1997). Global annual nitrogen deposition has increased tenfold over the past 150 years (Lamarque et al., 2005; Yue et al., 2016), although nitrogen deposition has decreased significantly across North America over the last decade due to pollution control. Historic nitrogen loading increased NPP (Elser et al., 2007; LeBauer and Treseder 2008; Xia and Wan 2008), which in turn increased carbon inputs to the forest floor and overall production of plant biomass (Hyvonen et al., 2007; Vitousek et al., 1997). Across biomes, total soil carbon tends to increase with experimental nitrogen addition (Yue et al., 2016), yet this may result less from increases in inputs and more from altering the extent or rates of decomposition (Frey et al., 2014; Liu and Greaver 2010). Microbial decomposition of soil carbon is generally retarded by nitrogen deposition (Hagedorn et al., 2003), but carbon allocation to roots also decreases with nitrogen deposition, limiting new carbon inputs to soil. However, a recent meta-analysis suggested that the reduction in soil carbon respiration, and thus increase in soil carbon stocks resulting from nitrogen deposition, might be equal in magnitude to the amount of additional carbon sequestered by aboveground vegetation (Janssens et al., 2010). Literature surveys suggest that the soil carbon response to anthropogenic nitrogen will fall in the range of 0 to 23 grams of carbon per gram of nitrogen added (Reay et al., 2008), but the uncertainty around this value is very high.

12.2.7 Protection Mechanisms

The extent of carbon protection (i.e., resistance to microbial decomposition) in soil historically has been attributed to litter chemistry, and this remains an element of carbon persistence (Clemente et al., 2011) in organic soils or organic soil horizons that accumulate on the surface of the mineral soil in forests. In recent decades, studies have shown that the controls on carbon stability in mineral soils are more likely dominated by physical and biological factors in the soil environment (Jastrow et al., 2006; Lehmann and Kleber 2015; Lin and Simpson 2016). Physical protection by spatial isolation (i.e., aggregate formation; McCarthy et al., 2008) and chemical associations with soil minerals (i.e., sorption) are both key drivers of carbon persistence in soils. Protection of carbon within soil aggregates (i.e., physical associations between soil minerals and organic compounds) can lead to long-term carbon storage in soils (Jastrow et al., 1996; Six et al., 2004). Compromising the physical structure of aggregates such as by tillage can result in substantial carbon losses because SOC becomes more available physically to decomposition (Navarro-Garcia et al., 2012). Alternatively, carbon may be protected via sorption to soil minerals in which reactive surfaces, including phyllosilicates, oxides, and other minerals, bind carbon molecules via chemical bridges and bonds. The types of compounds sorbed range from discrete chemical compounds (Solomon et al., 2012) to fragments of partially decayed microbial biomass (Courtier-Murias et al., 2013). Mineral-associated carbon stocks can have half-lives ranging from 30 to 4,500 years (Hall et al., 2015a, 2015b; Heckman et al., 2014), yet they can be rendered vulnerable as local environmental conditions change in ways that alter the chemical binding strength, such as changes in precipitation, infiltration, or temperature. In addition, larger-scale processes can serve to protect soil carbon, such as freezing, waterlogging, cryoturbation, or erosion deposition (Kaiser et al., 2007; Grosse et al., 2011; Berhe et al., 2007; Kroetsch et al., 2011).

12.2.8 Losses

Gas Fluxes

Gases including CO_2 and CH_4 are released from soils as a result of SOM and litter decomposition by soil microbes. Respiration of live roots and their

associated mycorrhizal symbionts also release CO_2 into the subsurface (Bond-Lamberty et al., 2004; Hanson et al., 2000; Subke et al., 2006; Tang et al., 2005). Globally, approximately 90 to 100 Pg C per year was released to the atmosphere from microbial soil respiration, and the projected rate increase is about 0.1 Pg C per year under a warming climate (Bond-Lamberty and Thomson 2010; Hashimoto et al., 2015). Soil respiration is affected by soil temperature, soil moisture, and organic carbon availability (Davidson and Janssens 2006). Typically, warming increases microbial respiration, while increases in moisture variably affect microbial respiration with maximum CO_2 emissions observed under partially saturated conditions. As soils saturate, methanogenesis is likely to emerge as the dominant carbon emission. Other global change factors such as elevated atmospheric CO_2 and naturally and anthropogenically altered soil nitrogen status also interactively affect soil respiration in direct and indirect ways (Billings and Ziegler 2008; Zhou et al., 2016). Also observed are vast differences in the amount of gas evolution as a function of landscape heterogeneity, underlying geology and soil type, and vegetative cover, as well as daily and seasonal temporal changes. Consequently, ESMs have not fully used soil respiration data for validation and calibration (Phillips et al., 2016).

Compared with CO_2, CH_4 has 28 times higher global warming potential over a 100-year time horizon (Saunois et al., 2016). Worldwide biogenic (i.e., associated with plants, animals, and microbes) sources of CH_4 emissions, including those from natural ecosystems, agriculture, biomass burning, and landfill waste, are estimated to be 0.33 Pg C per year or 12.4 Pg CO_2 equivalent[1] (CO_2e) per year, including anthropogenic biogenic sources of 7.4 Pg CO_2e per year (Tian et al., 2016). The U.S. inventory of greenhouse gases (GHGs) estimated anthropogenic total CH_4 emissions of 0.87 Pg CO_2e per year in 2015 if the 100-year global warming potential of 28 is used to calculate the CO_2 equivalent for CH_4, including anthropogenic biogenic sources of 0.42 Pg CO_2e per year, mostly from agriculture, landfill, and waste management (U.S. EPA 2017). Methane in North American soils is produced primarily under anaerobic conditions by methanogenic microbes, mostly in freshwater wetlands and rice paddies. However, CH_4 emissions are the net balance of both CH_4 production and oxidation (i.e., CH_4 destruction) by methanotrophic microbes (Tate 2015). The oxidation (i.e., consumption) of CH_4 in wetlands is important and may reduce potential CH_4 emissions by over 50% (Segarra et al., 2015).

Erosion

Soil erosion mobilizes about 75 Pg of soil each year by water and wind, with most erosion stemming from agricultural lands (Berhe et al., 2007). This accelerated movement of soil has major effects on the carbon cycle, most obviously because erosion physically removes SOC from soil profiles, exposing some fraction to oxidation during transit or upon deposition (Lal 2003). However, the degree to which soil erosion contributes to atmospheric CO_2 depends on several additional factors. Erosion can alter SOC mineralization and stabilization at both eroding and depositional sites, for example by burying and partially preserving SOC at the depositional site (Billings et al., 2010; Dialynas et al., 2016). Oxidation of eroded SOC is, therefore, only one component of net SOC change (Van Oost et al., 2012). Stallard (1998) first introduced the concept of new SOC production at an eroding site, a process which can balance the oxidation of eroded SOC (Berhe et al., 2007; Billings et al., 2010; Dialynas et al., 2016; Fang et al., 2006; Harden et al., 1999; Jenerette and Lal 2007; Liu et al., 2003; Quine and Van Oost 2007; Rosenbloom et al., 2006; Smith et al., 2001; Van Oost et al., 2007). Global estimates of the carbon sink strength of erosion and deposition vary widely. Several studies suggest that soil net erosion and deposition may result in a small net carbon sink, perhaps up to about 0.1 Pg C per year (Van Oost et al., 2007), although Berhe et al. (2007) suggest a modern erosion-induced carbon sink strength of about 0.7 to 1 Pg C per year. Wang et al. (2017) estimate a cumulative offset of atmospheric carbon of 78 ± 22 Pg C

[1] Carbon dioxide equivalent (CO_2e): Amount of CO_2 that would produce the same effect on the radiative balance of Earth's climate system as another greenhouse gas, such as methane (CH_4) or nitrous oxide (N_2O), on a 100-year timescale. For comparison to units of carbon, each kg CO_2e is equivalent to 0.273 kg C (0.273 = 1/3.67). See Preface, p. 5, for details.

due to agriculturally enhanced erosion during the period 6000 BC to AD 2015, which represents approximately 37 ± 10% of carbon emissions linked to contemporary anthropogenic land-cover change. Carbon burial rates have increased by a factor of 4.6 since AD 1850, consistent with erosion-induced carbon fluxes occurring disproportionately in recent centuries. Extrapolating globally, Billings et al. (2010) suggest an upper limit of a maximum net global sink of 3.1 Pg C per year (if all eroded carbon were protected from oxidation) and a net source of 1.1 Pg C per year if all eroded carbon were oxidized.

Estimating the rates of the erosion-induced redistribution of soil carbon has many uncertainties (Berhe et al., 2007; Regnier et al., 2013). These uncertainties derive from 1) the dynamics of eroded and deposited SOM (Hu and Kuhn 2014); 2) the texture and mineralogy of the soil being eroded; 3) the geomorphological nature and potential for decomposition in depositional environments; 4) the history and future of land uses, especially in intensively managed landscapes such as harvested forests and agriculture (Papanicolaou et al., 2015); and 5) changes to climate and hydrological cycles, including the timing and frequency of extreme events. Additional watershed-based studies, experimental studies, and modeling can address these uncertainties.

12.3 Modeling SOC Dynamics

At the global scale, the response of SOC to the influences of land use, disturbances, and climate change is projected using ESMs, which include simplified versions of soil carbon cycling models (Harmon et al., 2011; Tian et al., 2015). These early soil carbon models (e.g., CENTURY, Bolker et al., 1998; RothC, Gottschalk et al., 2012) largely assume exchanges of carbon between soil carbon pools are first-order exchanges defined by pool turnover times (Todd-Brown et al., 2013), and such assumptions (and model frameworks) continue into contemporary large-scale ESMs such as the Community Land Model (Huang et al., 2018) or the E3SM Land Model (Tang and Riley 2016). However, different models use different strategies to simplify and represent the complex cycling processes that were discussed in Section 12.2, p. 472; thus, model simulation results tend to diverge. For example, model outputs can vary widely in their projections of global carbon stocks and microbial respiration (Tian et al., 2015) based on nonmodeled outputs such as deep carbon storage and wetland carbon storage. The addition of land use to some models has indicated that soils previously projected to be sinks for CO_2 may actually be sources (Eglin et al., 2010). Because SOC stocks are so large compared to other global compartments (e.g., vegetation and atmosphere), the wide variations in projections of SOC stocks contribute a great deal of uncertainty to future carbon cycle projections (Todd-Brown et al., 2013). Wider adoption of global data products including the Harmonized World Soil Database and SoilsGrid (FAO/IIASA/ISRIC/ISSCAS/JRC 2012; Hengl et al., 2014) may facilitate the development of new tools to better integrate both local SOC observations (Dietze et al., 2014; Xia et al., 2013; Xu et al., 2006) and global data products into future models (Hararuk et al., 2014).

At a finer scale, the recognition that small-scale processes, including microbial respiration, nutrient limitation, and soil microclimate (Luo et al., 2016; Tian et al., 2015), affect overall soil carbon fluxes has prompted the emergence of microbially explicit and process-rich models for soil carbon cycling (Manzoni and Porporato 2009; Sulman et al., 2014; Tang and Riley 2014; Wieder et al., 2013). Models that include the size of the microbial biomass, microbial dormancy, and enzyme functions (Wang et al., 2014) are beginning to represent previously ignored processes such as priming (accelerated decomposition of stable carbon), mineral association, and temperature sensitivities, as well as their feedbacks to the Earth's physical system in the form of altered GHG emissions. The most recent soil-specific models, such as the Millennial Model (Abramoff et al., 2018), further classify SOC into measurable physicochemical categories (e.g., mineral-associated carbon, carbon physically entrapped in aggregates, dissolved carbon, and fragments of plant detritus) and include explicit processes regulating the transfers of carbon between pools, in contrast to the earlier models based on empirical turnover times (Abramoff et al., 2018).

These modeling types reflect very different scales, with ESMs simulating kilometer-scale landscapes and the more process-rich models simulating regional processes at finer scales such as centimeters to meters. Bridging these scales requires further empirical understanding and new mathematical frameworks (e.g., Wang et al., 2017). As models continue to advance, other challenges include determining which new models and approaches can be parameterized with empirical data and used for larger-scale decision making.

12.4 North American and Regional Context

12.4.1 United States

Scientists have used several approaches to estimate U.S. SOC stocks. These stocks may be aggregated in specific land areas such as geopolitical boundaries (i.e., states) or Land Resource Regions, or they may be grouped by soil-order or land-cover classes (Guo et al., 2006; Wills et al., 2014). Most efforts have developed estimates for the conterminous United States (CONUS), but results vary based on methods and assumptions. Guo et al. (2006) estimated SOC stocks for CONUS as between 30 and 150 Pg (0 to 2 m in depth) by soil order using the State Soil Geographic database (STATSGO; USDA Soil Conservation Service 1993) and another 23 to 94 Pg C stock as inorganic carbon within the top 2 m of surface. Compared with CONUS, fewer studies have estimated soil carbon stocks for Alaska. Mishra and Riley (2012) estimated stocks in Alaska as 77 Pg C, an update from the value of 48 Pg estimated by Bliss and Maursetter (2010). The U.S. Geological

Table 12.1. Estimates of Soil Carbon Storage in the Conterminous United States in Different Land-Use Classes[a–d]

Land Cover	Soil Organic Carbon (from RaCA[e])	Soil Organic Carbon (Bliss et al., 2014)	Soil Organic Carbon (Sundquist et al., 2009)	Soil Organic Carbon (Other Estimates)
Forests and Woodlands	20	13.1	25.1	28[f]
Agriculture	13	13.4	27.4[d]	
Shrublands		5.6	9.7	
Urban		3.3		1.9[g]
Wetlands	14	8.9		13.5[h] – 11.5[i]
Rangelands (+ Pasture)	19	12.3	11.2[d]	
Totals	65	57.2[j]	73.4	

Notes

a) Storage measured in soil down to 1 m in depth.

b) All values are in petagrams of carbon (Pg C).

c) No total is given for "Other Estimates" values because the values do not represent all land-use classes and some land-use classes likely overlap (e.g., urban is partially accounted for in agriculture [see d] and developed; range estimates likely include some agricultural land).

d) "Agriculture" is listed in Sundquist et al. (2009) as "agriculture and developed"; "rangelands and pasture" is listed as "other" and includes all grasslands.

e) RaCA, U.S. Department of Agriculture's Rapid Carbon Assessment.

f) Domke et al. (2017).

g) Pouyat et al. (2006).

h) From the *Second State of the Carbon Cycle Report* (SOCCR2), Ch. 13: Terrestrial Wetlands, p. 507.

i) Nahlik and Fennessy (2016).

j) Total soil profile of carbon is 73 Pg.

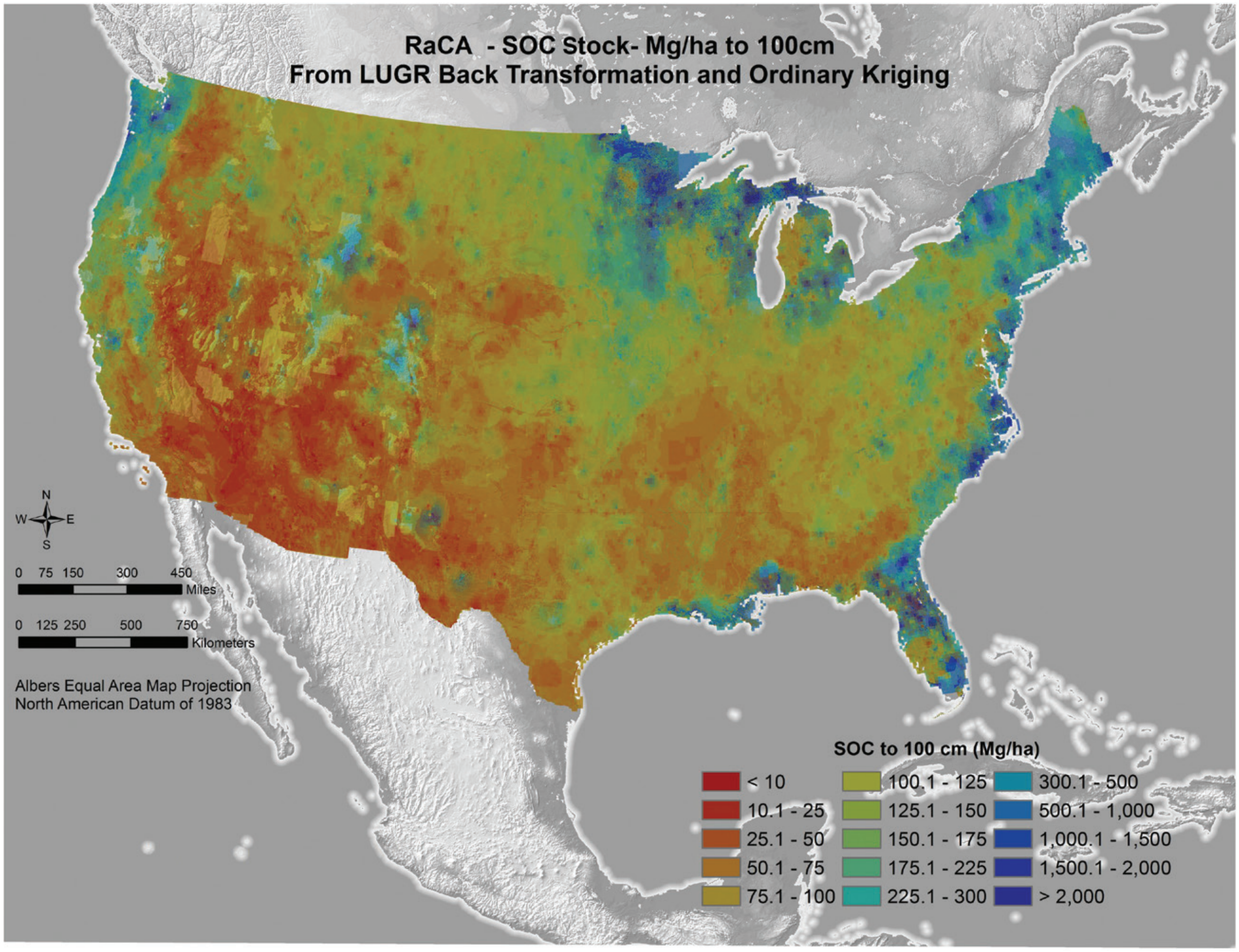

Figure 12.2. Rapid Carbon Assessment (RaCA) of Soil Organic Carbon (SOC) Stock Values. Data are in megagrams (Mg) of carbon per hectare (ha) to 100 cm. Soil group strata and land use and land cover (LULC) strata were linked together into a LULC-Soil Group Combination, designated as "LUGR." Prepared using the geometric mean of pedon stocks according to RaCA methodology. [Figure source: Reprinted from U.S. Department of Agriculture Natural Resources Conservation Service, Soil Survey Staff, RaCA project. Prepared by Skye Wills, 2016]

Survey (USGS) calculated CONUS SOC storage as 77.4 Pg C from the Soil Survey Geographic (SSURGO) database, developed by the USDA Natural Resources Conservation Service (NRCS). This information is supplemented with data from the Digital General Soil Map of the United States (STATSGO2; catalog.data.gov/dataset/u-s-general-soil-map-statsgo2-for-the-united-states-of-america; Sundquist et al., 2009; see Table 12.1, p. 479).

The NRCS's recent RaCA project captures information on the carbon content of soils across CONUS at a relatively uniform point in time (Soil Survey and Loecke 2016). A secondary goal was to capture SOC stocks in different kinds of soils and land uses. For this assessment, RaCA collected 144,833 samples from the upper 1 m of 32,084 soil profiles at 6,017 randomly selected locations across the United States. Independently developed soil groups for each RaCA region were combined with land-use, land-cover information, yielding an estimate of the total carbon stock across CONUS of 65 Pg C (see Figure 12.2, this page). Different estimates of soil carbon pools are expected to differ; individual soil and land-cover classes have different levels of uncertainties surrounding their carbon pool estimates, and errors can include land-classification differences and different ways of aggregating sparse data. For example, Domke et al. (2017) used the USDA Forest Service's Forest Inventory and Analysis (or FIA) data to project SOC density in CONUS forest types and parts of Alaska and compared regional projections to those from RaCA. These modeled SOC density projections were substantially smaller than those of RaCA for most NRCS Land Resource Regions, at times by more than a factor of three.

Carbon storage in interior CONUS wetlands are assessed (see Ch. 13: Terrestrial Wetlands, p. 507) using a combination of NRCS SSURGO data and the U.S. Fish and Wildlife Service's (USFWS) National Wetland Inventory. These estimates of the upper 1 m indicate that terrestrial wetlands store about 13.6 Pg C, a value very similar to that of Nahlik and Fennessy (2016), who reported a value of 11.5 Pg. Storage of carbon in CONUS saline wetlands is significantly lower. Estimates of tidal wetland soil stocks along the freshwater-to-saline transition area plus the seagrass soil stocks are 0.8 Pg C for "blue carbon" ecosystems (see Ch. 15: Tidal Wetlands and Estuaries, p. 596). Given that more than half the historical U.S. wetland area has been lost due to anthropogenic activities, further loss of wetland soils represents a key vulnerability that could result in a net transfer of carbon from the soil to the atmosphere.

12.4.2 Mexico

The most recent estimate of soil carbon stocks in Mexico is reported to a depth of only 30 cm. According to Jobbágy and Jackson (2000), the top 20 cm of soil typically represents 40% of total soil carbon stocks averaged across vegetation communities in Mexico. At 9.13 Pg C in the top 30 cm, this reported SOC stock is 73% of the country's total terrestrial stock (CONAFOR 2010), but a conservative estimate of SOC stocks to 1 m in depth might be 18 Pg C, assuming that the top 30 cm represents about half the total soil carbon stocks. However, this estimate remains highly uncertain as acquisition of field data to fill data gaps (e.g., bulk density measurements) and spatial extrapolation methods continue to evolve (de Jong et al., 2010). For example, simply using different versions of land-cover maps for spatially extrapolating mean SOC values results in significant differences for semitropical low forests and mangroves (Paz Pellat et al., 2016). Despite these issues, almost half (48%) of Mexico's SOC appears to be contained in forests, especially the dry deciduous, semi-evergreen, and oak forests (see Tables 12.2, this page, and 12.3, p. 482). Furthermore, grazing lands accounted for 23% of the total SOC stock, mostly due to their extensive area. Finally, despite the relatively low soil carbon density of shrublands, they were extensive enough to account for 7% of the total SOC stock (Paz Pellat et al., 2016).

Table 12.2. Soil Organic Carbon Distribution in Mexico by FAO FRA[a] Classes[b]

FAO FRA Classes[a]	Area in Millions of Hectares	Petagrams of Carbon
Forestlands	65	4.3
Other Forestlands	20	0.6
Other Lands	108	4.1
Planted Forest	0.33	< 0.01
Totals	**194**	**9.1**

Notes

a) Global Forest Resources Assessment (FRA) of the United Nations Food and Agriculture Organization (FAO).

b) From Paz Pellat et al. (2016).

Table 12.3. Soil Organic Carbon Distribution in Mexico for Vegetation Types with Top Five Highest Total Soil Carbon Estimates[a]

Vegetation Types (Top Five)	Area in Millions of Hectares	Teragrams of Carbon	Percent of Total
Grazing Lands	50	2,115	23
Deciduous Dry Forest	14	690	8
Desert Microphyll Shrub	22	600	7
Medium Semi-Evergreen Forest	5	570	6
Oak Forest	11	564	6

Notes

a) From the National Institute for Statistics and Geography of Mexico for 2007 (from Paz Pellat et al., 2016).

At the national scale, CO_2 fluxes from mineral soils to the atmosphere were estimated as 30.2 Tg CO_2 per year, mostly from deforestation of secondary oak, pine-oak, and tropical dry forests (de Jong et al., 2010). About 10% of Mexico's land is strongly affected by soil erosion, with about 36% remaining stable (Bolaños-González et al., 2016).

Temperate forests in Mexico are potential areas of carbon sequestration because about 10% of total GHG emissions in Mexico are attributed to land-use change from opening new areas to cultivation and logging. Tropical forests in Mexico also experience much of the same pressures of land-use change, but they occur over stronger gradients of precipitation. Land-use change from forest to pasture appears to interact strongly with precipitation. For example, dry tropical forest conversion to pasture may increase SOC (3.7% at 788 mm per year), yet this same land-use change appears to decrease SOC as precipitation increases (–0.2% at 2,508 mm per year; –2.2% at 4,725 mm per year; Campo et al., 2016). Mangroves in Mexico have the highest density of soil carbon (364 Mg C per hectare), located throughout Mexico's extensive coastline and riverine systems. A variety of disturbances affect mangroves and, as in many parts of the world, include erosion, increasing sea level change, and salt intrusion (Gilman et al., 2008). Due to the difficultly in sampling these soils, few estimates are available, especially if attempting to quantify this stock to the bottom of the organic layer. Nevertheless, the Gulf of Mexico region generally has the highest carbon stocks (1,300 Mg C per hectare) of SOC compared with those of the other regions in Mexico (100 to 1,100 Mg per hectare; Herrera Silveira et al., 2016).

12.4.3 Canada

Canada has a total land area of 998.5 megahectares (Mha) that contains 72.2 gigatons of carbon (Gt C) to a depth of 30 cm (Tarnocai 1997). The total of 55.2 Mha of land currently used for agriculture contain about 4.14 Gt C to a depth of 30 cm and 5.5 Gt to 1 m. As about 80% of agricultural land is located in the Canadian Prairies, most (approximately 88%) SOC is also found in Prairie soils, which are mostly carbon-rich Chernozemic soils developed under grassland. Tarnocai (1997) estimated a total of 262.3 Pg C in soils within the tundra, forest, and agricultural regions of Canada. Over half the carbon (147.1 Pg C; Tarnocai 2006) is in organic (peat) soils, some of which are affected by permafrost. Total soil carbon estimates for Canada likely will increase as knowledge of deep carbon stocks in permafrost soils increases (Hugelius et al., 2014). For example, Kurz et al. (2013) estimated that soils in Canada's boreal forest region alone contain 208 Pg C, which is about 80% of the Tarnocai (1997) estimate of the total carbon stocks in Canada. Of this 208 Pg, the majority (137 Pg) of the boreal soil carbon stocks are in the deep organic soils of the country's extensive peatlands, and the remainder (71 Pg) are in upland

Table 12.4. Estimates of Soil Carbon Storage in Canada[a–b]

Land Cover	Soil Organic Carbon
Organic (Peat) Soils	147.1[c], 137[e]
Agriculture	5.5[d]
Boreal Forest Region	208[e, f]
Upland Forest Soils	71[e]
Total	262.3[c, g]

Notes

a) Storage measured in soil down to 1 m in depth.
b) Values in petagrams.
c) Tarnocai (2006).
d) Tarnocai (1997).
e) Kurz et al. (2013).
f) Note that this overlaps with estimates of organic peat soil carbon.
g) Columns do not add up due to overlap in categories.

forest soils that often have thick organic soil horizons (42 to 55 Mg C per hectare; estimated from Letang and de Groot 2012) that overlay the mineral soil (Kurz et al., 2013; see Table 12.4, this page).

Canadian forest soil carbon research over the last decade has focused on understanding the dynamics of SOC as influenced by 1) mosses (Bona et al., 2013, 2016); 2) forest composition and soil taxonomy (Laganiere et al., 2015; Shaw et al., 2008, 2015); 3) invasive earthworms (Cameron et al., 2015); 4) response to temperature changes (Laganiere et al., 2015; Smyth et al., 2011); 5) response to wildfire, specifically in peatlands (Granath et al., 2016; Kettridge et al., 2015); and 6) recovery patterns (Ward et al., 2014). Under development is a national peatland carbon modeling system (Webster et al., 2016) that will fill information gaps previously identified, including a peatland-type map; landscape-scale modeling of forested, treed, and nontreed peatland types; water table fluctuation in response to climate change; and CH_4 fluxes (Shaw et al., 2016). Eventually, responses to permafrost thaw, wildfire, and anthropogenic disturbances will be included (Shaw et al., 2016; Webster et al., 2016).

Several new spatial products and databases have improved the understanding of relationships among vegetation types (Beaudoin et al., 2014; Thompson et al., 2016) and changes in disturbance-type patterns (Hermosilla et al., 2016), improving accuracy and enhancing the ability to scale up and integrate results from fine-scale to landscape-scale studies reporting national GHG emissions.

The 55.7 Mha of land that currently are used for agriculture in Canada are estimated to contain about 4.3 Pg C to a depth of 30 cm and 6.6 Pg C to 1 m using the Canadian Soil Information Service (CanSIS) National Soil Database. As of 2013, Canadian agricultural land removed 11 Tg CO_2 per year, an amount which represents about 2% of the total national GHG emissions (ECCC 2015). This is due largely to a reduction in the use of summer fallow lands and increased adoption of no-till practices in the Canadian Prairies. However, this value has declined from the reported 13 Tg in 2005 because changes in SOC stocks and fluxes tend to reach equilibrium at some point after a change in conditions.

12.4.4 Arctic and Boreal Ecosystems

Arctic and boreal ecosystems cover about 22% of the global land surface (Chapin et al., 2000) and contain 1,035 ± 150 Pg C in the upper 3 m of surface soil (Hugelius et al., 2014), amounts which equal about 33% of the total global surface SOC pool (Jobbágy and Jackson 2000; Schuur et al., 2015). The presence of permafrost and waterlogged soils in boreal and Arctic soils has allowed the accumulation of large quantities of carbon in this biome (McGuire et al., 2009; see Ch. 11: Arctic and Boreal Carbon, p. 428, for more details). Deep soils (>3 m in depth) contain significant stocks estimated between 210 ± 70 Pg C and 456 ± 45 Pg C, particularly in carbon-rich Pleistocene-age sediments called "yedoma" found in unglaciated parts of Alaska and Siberia, as well as in their alluvial deposits (Hugelius et al., 2014).

The changing disturbance regime can strongly affect soil carbon storage and flux. Permafrost thaw (Schuur et al., 2015) is tied to changes in the timing, frequency, and severity of wildfires

(Chapin et al., 2010; Kasischke et al., 2010), plant community composition (Mann et al., 2012), and alterations in the hydrological cycle (Jorgenson et al., 2001, 2010; Roach et al., 2013). Thaw will affect both storage and fluxes of carbon as the climate continues to warm. An estimated 5% to 15% of the terrestrial permafrost carbon pool is thought to be vulnerable to decomposition and release to the atmosphere, based on a synthesis of experimental studies, ecosystem models, and expert assessments (Schuur et al., 2015). Carbon loss from peatlands has shown large responses to water table fluctuations (Waddington et al., 2015), wildfire events (Turetsky et al., 2011), and permafrost thaw (Jones et al., 2017; Wisser et al., 2011). Key uncertainties as to the future of carbon storage in Arctic and boreal regions include the extent to which plant community productivity will respond to elevated CO_2 (McGuire et al., 2009), whether landscapes will become wetter or drier in the future (Schuur et al., 2015), the magnitude of winter fluxes (Commane et al., 2017), and the extent of the permafrost carbon feedback (Schaefer et al., 2011; Schuur et al., 2015).

12.5 Societal Drivers, Impacts, and Carbon Management

12.5.1 Agriculture

Because more than 50% of the Earth's vegetated surface is dedicated to agriculture (e.g., cropland and grazing land), understanding the role of agricultural management on SOC stocks is critical (see Ch. 2: The North American Carbon Budget, p. 71). Virtually all management choices (e.g., crop type, rotation, tillage, fertilization, irrigation, and residue management) will affect carbon inputs (e.g., crop residues and manure) and the decay rate or erosional loss of SOM (Paustian et al., 1997; Smith 2008). In most cases, SOC changes occur slowly and short-term (annual) changes are difficult to measure, but studies from long-term experiments, together with improved predictive models, provide a basis for guiding management and policies to improve SOC stocks (NAS 2010; Ogle et al., 2014; Paustian et al., 2016).

Causes of SOC loss include 1) reduced biomass carbon inputs; 2) enhanced erosion and leaching; and 3) increased decomposition rates due to tillage disturbance (Paustian et al., 2016). A meta-analysis for Canadian soils reported that, when native soil was converted to agricultural land, there was an average loss of 24% ± 6% of soil carbon (VandenBygaart et al., 2003). Globally, agricultural soils have lost, on average, 20% to 45% of their original top soil carbon (0 to 30 cm) but with much higher losses in cultivated organic soils and where extensive erosion has occurred (Don et al., 2011; Ogle et al., 2005). Following restoration of perennial forest and grassland vegetation on annual cropland (e.g., for soil restoration or retiring marginal lands from production), much of the lost soil carbon stocks eventually can be recovered. Conversion of annual cropland to perennial grassland in temperate environments increased soil carbon stocks, on average, by 13% to 16%, with greater relative increases occurring in more mesic climates (Ogle et al., 2005).

In recent decades, SOC stocks in agricultural soils in the United States and Canada have stabilized and in some cases begun to increase (Follett et al., 2011; U.S. EPA 2015) as new conversion of land to agricultural use has largely halted and adoption of soil conservation practices and crop yields have increased (Chambers et al., 2016; Johnson et al., 2006). Effects of agriculture on soil carbon stocks, along with effects of conservation measures, are reviewed and quantified in Angers and Eriksen-Hamel (2008), Hutchinson et al. (2007), Luo et al. (2010), Palm et al. (2014), Paustian et al. (2016), Powlson et al. (2014), and many others. Improved residue management, added forage in crop rotations or adoption of agroforestry, double-cropping, conservation reserve planting, increased use of perennials in rotation, and use of practices that increase plant growth such as effective fertilization are successful in increasing soil carbon (Hutchinson et al., 2007; Luo et al., 2010; Palm et al., 2014), especially if more than one practice is used. In Canada, the wide adoption of reduced tillage and summer fallow over many regions has resulted in soil carbon increases and reduced erosion

(Agriculture and Agri-Food Canada 2016; Soil Conservation Council of Canada 2016).

An analysis of no-till only versus conventional till by Palm et al. (2014) found that carbon gains occurred in only half the paired comparisons and that increased residue retention had a greater effect on soil carbon than reduced tillage. Powlson et al. (2014) argue that adoption of no-till agriculture can improve crop production and reduce erosion in many cases, but it may not have significant effects on carbon sequestration. However, a meta-analysis by Kopittke et al. (2017) saw an overall small positive (+9%) effect of conversion to no-till from conventional till methods. Most analyses of tillage effects do not account for SOC erosion. Montgomery (2007) calculated a mean erosion rate difference between conventional agriculture and no-till agriculture of about 1 mm per year. Although this eroded soil causes a net movement of carbon from the site with associated negative effects on soil fertility and health, this movement might not represent a net loss of soil carbon globally and could represent a net sink, because the eroded carbon can be buried and therefore protected. Meanwhile, carbon accumulation can continue in the site from which the erosion originally occurred via the usual processes of additions and transformations of plant residues (Wang et al., 2017).

Estimates of the current SOC balance for U.S. agricultural lands suggest a small net sink on long-term cropland (6.4 Tg C per year) and on land recently converted to grassland (2.4 Tg C per year), while small net losses of SOC were estimated for long-term grassland (3.3 Tg C per year) and land recently converted to cropland (4.4 Tg C per year; U.S. EPA 2015). A similar picture appears for Canadian agricultural soils with an estimated net sink of about 3 Tg C per year (ECCC 2015). A full soil carbon inventory for Mexican agricultural soils is still in progress; however, with ongoing forest conversion to agricultural uses (see Section 12.4.2, p. 481), there likely is a substantial loss of SOC due to agricultural activities.

Other chapters present more information on management of agricultural soils and its effects on carbon (see Ch. 5: Agriculture, p. 229; Ch. 7: Tribal Lands, p. 303; and Ch. 10: Grasslands, p. 399).

12.5.2 Forestry

A wide variety of forest management practices affect around 204 Mha of timberlands in CONUS (see Ch. 9: Forests, p. 365). Those practices typically involve a combination of harvesting, stand regeneration, and stand tending. The intensity of those practices and their resulting effects on soils depend on landowner management objectives.

To date, most research on forest harvest effects on soil carbon has suggested that mild to moderate intensity harvesting does not cause measurable changes in upland soils (Johnson and Curtis 2001), but that intensive harvesting and plantation management may cause reductions in mineral soil carbon (Buchholz et al., 2014; Johnson and Curtis 2001), especially if imposed on old-growth natural stands. A meta-analysis of studies measuring effects of forest harvest on soil carbon stocks by Nave et al. (2010) found that while forest floor carbon generally was reduced after harvest, mineral soil carbon was less affected, although certain soil orders were more susceptible to mineral soil carbon loss than others. Forest soil carbon stores have the ability to recover to preharvest stages, although recovery might take decades (Nave et al., 2010) to a century or more (Diochon et al., 2009); thus, rotation length plays a significant role in the degree of harvest impacts on soil carbon. Several chronosequence studies have observed reductions in mineral-bound carbon pools in successional stands decades after harvesting (Diochon et al., 2009; Lacroix et al., 2016; Petrenko and Friedland 2015). Because this timing of carbon loss corresponds to periods of high nutrient demands during biomass re-accumulation, the cause could be mining of SOM by plants and mycorrhizal fungi to alleviate nutrient limitation. Dean et al. (2017) argue from a modeling standpoint that there are more significant losses of soil carbon with forest harvest of primary forests when calculated over centuries, but this model result is not supported by empirical studies.

Afforestation and agroforestry (the practice of integrating woody vegetation with crop and/or animal production systems) have been cited as having potential for increasing soil carbon sequestration (IPCC 2000; Upson et al., 2016). Several meta-analyses conducted on afforestation effects on former croplands have produced a general consensus that soil carbon gains may take more than 30 years to be measurable (Barcena et al., 2014; Li et al., 2012; Nave et al., 2013) but can increase carbon stocks by 19% to 53% (Guo and Gifford 2002; Nave et al., 2013). However, while tree establishment in both grasslands and croplands showed greatly increased aboveground biomass carbon storage, meta-analysis of studies found that tree establishment on pastureland led to losses or no changes in soil carbon (Shi et al., 2013).

12.6 Synthesis and Outlook

Soil carbon is vulnerable to both pervasive warming and moisture disturbances, as well as to land-use decisions, all of which can strongly affect soil carbon contents. In northern latitudes, which are particularly vulnerable to soil carbon loss, some of the fastest warming trends (Cohen et al., 2014) and largest carbon stocks (Ping et al., 2008) occur. A significant portion of northern soil carbon is stored as organic peat horizons, which play a pivotal role in insulating permafrost from temperature changes but are particularly sensitive to changes in soil moisture (Johnson et al., 2013). Thus, the feedbacks among warming, moisture, and wildfire have important consequences to the carbon cycle at a global scale (Olefeldt et al., 2016). Meanwhile, localized "hotspots" for soil carbon storage, while also vulnerable to warming and soil moisture, can be sensitive to management practices as well and, therefore, can offer potential mitigation opportunities to avoid carbon emissions. For example, maintaining high water tables in carbon-rich peatlands potentially avoids carbon emissions that otherwise would accompany drainage.

Management options for actively sequestering carbon into soil are important opportunities for climate mitigation, but several issues arise before there is confidence in the outcome for a given soil under a given management setting. Topographical and mineralogical characteristics and disturbance histories (e.g., fire-return interval and land-use change history) likely influence the net balance between input and loss and yet are highly variable across North America. Strategic experimental designs with consistent oversight and methodologies could constrain the uncertainties and understanding of the processes that control carbon storage. Building spatially and temporally explicit databases could improve process-based models to provide better estimates for soil carbon trajectories and thereby empower land managers to chart the trajectory of soil carbon.

Increasingly, the development of policies to 1) promote improved soil health (Kibblewhite et al., 2008; Vrebos et al., 2017), 2) encourage soil carbon sequestration for GHG mitigation (Chambers et al., 2016; Follett et al., 2011), and 3) satisfy consumer demands for more sustainable products (Lavallee and Plouffe 2004) will demand strong scientific support for improved understanding of SOC dynamics, new technologies to increase SOC stocks, and decision-support tools to effectively assess options and monitor progress. Along with new research on more conventional practices to build soil carbon (e.g., improved rotations, reduced tillage, and cover crops), scientists are investigating newer practices and technologies to increase SOC stocks, including 1) applying biochar (Woolf et al., 2010) and compost (Ryals et al., 2015), 2) using deep tillage to increase the total depth and storage of SOC-rich soil (Alcantara et al., 2016), 3) deploying new crop varieties with increased allocation of carbon below ground and deeper into the soil profile (Paustian et al., 2016), and 4) planting perennial plants in place of annual crops (Cox et al., 2006). New research and best practices in forestry such as selective harvesting and residue management (Peckham and Gower 2011), tailored for particular soils (Hazlett et al., 2014), also have the potential to increase carbon retention in forest soils. As new knowledge is generated about the applicability of various practices in different environments, incorporating this new information into improved decision-support tools (see Ch. 18: Carbon Cycle Science in Support of Decision Making, p. 728) will guide land managers, industry, policymakers, and other stakeholders in building heathier soils that are rich in organic matter.

SUPPORTING EVIDENCE

KEY FINDING 1

Estimates for soil carbon stocks in the conterminous United States plus Alaska range from 142 to 154 petagrams of carbon (Pg C) to 1 m in depth. Estimates for Canada average about 262 Pg C, but sampling is less extensive. Soil carbon for Mexico is calculated as 18 Pg C (1 m in depth), but there is some uncertainty in this value (*medium confidence*).

Description of evidence base

The value range of soil carbon to a depth of 1 m for the United States is based on several compilations: Alaska is estimated in Mishra and Riley (2012) as 77 Pg C, an increase from the value reported by Bliss and Maursetter (2010) of 48 Pg. The sampling for the Mishra and Riley (2012) estimate is quite extensive, and land types for areal weighting are well known and documented. Modern estimates for the conterminous United States (CONUS) span the range from the U.S. Geological Survey (USGS) estimate of Sundquist et al. (2009) at 77 Pg C and the Rapid Carbon Assessment (RaCA, initiated by the Soil Science Division of the U.S. Department of Agriculture's National Resources Conservation Service in 2010) estimate (Soil Survey and Loecke 2016) at 65 Pg C (see Table 12.1, p. 479). The RaCA estimate is based on 144,833 soil samples and extrapolation using detailed soil maps. The soil carbon value of 9 Pg C for Mexico is based on Paz Pellat et al. (2016), but that estimate is based on sampling to a depth of only 30 cm. Based on conversion factors in Jobbágy and Jackson (2000), a conservative extrapolation to 1 m yields a value of 18 Pg C. The estimates for Canada are from Tarnocai (1997, 2006). This assessment recognizes that 1 m is a very arbitrary depth to consider; Batjes (1996) reported a 60% increase in the global soil organic carbon (SOC) budget when the second meter of soil was included.

Major uncertainties

There is medium high confidence in the estimates from CONUS due to new extensive and intensive sampling, although estimates for specific land-use classes still vary with different estimates. Confidence is relatively high for estimates in the agricultural areas of Canada but lower for forested areas. In Canada, uncertainty for the large peatlands areas in the boreal and Arctic regions is high due to low-sampling intensity and low-resolution mapping of peatland types. Uncertainty for estimates from Mexico are likely high due to low sampling coverage, and available data are only to a depth of 30 cm.

Assessment of confidence based on evidence and agreement, including short description of nature of evidence and level of agreement

Soil carbon was extensively sampled in three independent studies for CONUS, so the confidence for the range of values reported here is very high. Due to the complex nature of estimating soil carbon in boreal and peat regions, the uncertainty is greater surrounding values for Canada. There is low confidence in values reported for Mexico as sampling is not as extensive and the depth of sampling is not as great.

Summary sentence or paragraph that integrates the above information

The estimates of total soil carbon stores are reasonably accurate for CONUS and Canada but are less accurate for Mexico.

KEY FINDING 2

Most Earth System Models (ESMs) are highly variable in projecting the direction and magnitude of soil carbon change under future scenarios. Predictions of global soil carbon change through this century range from a loss of 72 Pg C to a gain of 253 Pg C with a multimodel mean gain of 65 Pg C. ESMs projecting large gains do so largely by projecting increases in high-latitude soil organic carbon (SOC) that are inconsistent with empirical studies that indicate significant losses of soil carbon with predicted climate change (*high confidence*).

Description of evidence base

A description of the scientific concerns with current ESMs is presented in He et al. (2016). They analyzed ^{14}C data from 157 globally distributed soil profiles sampled to a depth of 1 m to demonstrate that ESMs currently overestimate the soil carbon sink potential. Todd-Brown et al. (2014) also pointed out major sources of error in current ESMs and suggested that most ESMs poorly represented permafrost dynamics and omitted potential constraints on SOC storage, such as priming effects, nutrient availability, mineral surface stabilization, and aggregate formation. For example, many ESMs simulated large changes in high-latitude SOC that ranged from losses of 37 Pg C to gains of 146 Pg C. The poor performance of current ESMs can result from biases in model structure, parameterization, initial values of carbon pools, and other variables (Luo et al., 2016).

There is currently a great deal of controversy over how to improve the representation of soil carbon in models (Chen et al., 2015); several authors suggest that microbial dynamics, including the priming effect, need better representation (Georgiou et al., 2015; Sulman et al., 2014; Wieder et al., 2014), as does soil carbon response to nitrogen enrichment (Janssens and Luyssaert 2009; Riggs and Hobbie 2016). However, there is no evidence that suggests how much detail is needed to adequately represent future soil carbon dynamics and soil carbon pools.

Deep carbon (>1 m in depth) generally has been found to be more stable and resistant to management or climate change than carbon in surface soils (Rumpel and Kögel-Knabner 2010; Schrumpf et al., 2013), but, given that subsurface horizons contain more than half the soil carbon (Jobbágy and Jackson 2000), small changes could significantly affect carbon budgets. Although less well studied, deep carbon has been shown to be sensitive to management practices (Alcantara et al., 2016; Ward et al., 2016).

Microbial dynamics, including the priming effect, are key controls on soil carbon turnover (Bernal et al., 2016; Guenet et al., 2012). Carbon-use efficiency of different substrates by microbes might be a key factor in soil carbon stabilization (Cotrufo et al., 2013).

Major uncertainties

How much detailed information on microbial physiology, coupled carbon-nitrogen cycles, or other processes is needed to improve soil carbon models is not well known.

Assessment of confidence based on evidence and agreement, including short description of nature of evidence and level of agreement

Models can be tested against empirical data, and they do not perform very well; thus, determining the accuracy of future projections is difficult.

Summary sentence or paragraph that integrates the above information

The poor performance of current ESMs can result from biases in model structure, parameterization, initial values of carbon pools, and other variables. Most ESMs poorly represent permafrost dynamics and omit potential constraints on SOC storage, such as priming effects, nutrient availability, mineral surface stabilization, and aggregate formation.

KEY FINDING 3

Soil carbon stocks are sensitive to agricultural and forestry practices and loss of carbon-rich soils such as wetlands. Soils in North America have lost, on average, 20% to 75% of their original top soil carbon (0 to 30 cm) with historical conversion to agriculture, with a mean estimate for Canada of 24% ± 6%. Current agricultural management practices can increase soil organic matter in many systems through reduced summer fallow, cover cropping, effective fertilization to increase plant production, and reduced tillage. Forest soil carbon loss with harvest is small under standard management practices and mostly reversible at the century scale. Afforestation of land in agriculture, industry, or wild grasslands in the United States and Canadian border provinces could increase SOC by 21% ± 9% (h*igh confidence*).

Description of evidence base

Converting native forests or pastures to cropland can reduce soil carbon by 42% to 59%, respectively (Guo and Gifford 2002). A meta-analysis for Canadian soils reported that, when native soil was converted to agricultural land, there was an average 24% loss of soil carbon (VandenBygaart et al., 2003). Estimates for Mexico also suggest that loss of soil carbon due to management remains significant (Huber-Sannwald et al., 2006).

Agricultural effects on soil carbon stocks, including effects of conservation measures, are reviewed and quantified in Angers and Eriksen-Hamel (2008), Hutchinson et al. (2007), Luo et al. (2010), Palm et al. (2014), Paustian et al. (2016), Powlson et al. (2014), and many others. Specific conservation measures for improved soil carbon retention have been shown to be effective in both Canada and the United States. In Canada, conservation measures, including reduced summer fallow and reduced tillage, have been widely adopted over many regions and have resulted in soil carbon increases and reduced erosion (Soil Conservation Council of Canada 2016). Agriculture and Agri-Food Canada (2016; AAFC) has 30 years of data showing that, in the Canadian Prairies, reduced tillage combined with reduced summer fallow have led to significant SOC increases. Improved residue management, including adding forage in crop rotations or adopting agroforestry, and practices that increase plant growth such as effective fertilization are effective in increasing soil carbon (Hutchinson et al., 2007; Palm et al., 2014). A meta-analysis by Angers and Eriksen-Hamel (2008) suggested that, although significant increases in surface soil carbon with reduced tillage are commonly observed, the slight decreases in soil below the plow layer also are common, thus making overall increases in total soil carbon profiles averaged across studies small but significant. In a more recent meta-analysis by Luo et al. (2010), increased soil carbon with reduced tillage was seen only for double-cropping systems, a finding which agrees with the AAFC result that reduced summer fallow and reduced tillage together caused significant increases in soil carbon.

Palm et al. (2014) point out serious methodological flaws with many tillage comparisons that include sampling by depth not equivalent soil mass, flaws which cause significant overestimates of soil carbon in no-till soils with higher bulk densities. In their 2014 meta-analysis, about half the paired comparisons showed small increases in soil carbon from reduced till but half did not, suggesting that increased residue retention is more significant than reducing tillage. A similar meta-analysis by Kopittke et al. (2017) that also corrected for changes in bulk density found an overall small positive (+9%) effect of conversion to no-till practices from conventional till. Powlson et al. (2014) point out that the gains in surface soil carbon with adoption of no-till methods can improve crop production and reduce erosion in many cases, but the reverse can be true in cool, wet climates or the wet tropics.

Several meta-analyses of afforestation effects on former croplands have been conducted, and there is general consensus that soil carbon gains may take more than 30 years to be seen (Barcena et al., 2014; Li et al., 2012; Nave et al., 2013) and can increase carbon stocks by 19% to 53% (Guo and Gifford 2002; Nave et al., 2013).

Data on forest harvest effects are from a comprehensive meta-analysis by Nave et al. (2010), who report variable and low changes in mineral soil carbon stocks with forest harvest but significant decreases in forest floor carbon. Several chronosequences support this meta-analysis. Dean et al. (2017) argue from a modeling standpoint that there are significant long-term losses of soil carbon with forest harvest of primary forests; however, much of this argument is based on assumptions about the relationship between plant inputs and soil carbon sequestration that are not necessarily supported by empirical studies.

Wetland estimates are based on information in this report's (SOCCR2) two wetland chapters. All chapters showed findings of strong evidence that loss of wetlands is a significant factor for total soil carbon loss, given the very high carbon density of wetland soils.

Wear and Coulston (2015), using data from the National Greenhouse Gas Inventory (NGHGI), report annual forest carbon accumulation, including both sequestration and land-use transfers in the United States as 223 teragrams of carbon (Tg C) per year, roughly 0.5% of the stored forest carbon. This likely translates into increased soil carbon storage, although this distinction was not made in the analysis. Similar estimates have not been made for Canada or Mexico.

Major uncertainties

The certainty for forest harvest effects on soil carbon appears to be very robust and based on many studies across North America, although a recent modeling study suggests that these other studies, carried out over decades, miss a multicentury-scale slow loss of soil carbon with forest harvest. However, there are no data to support that model result. Uncertainty arises because there are few empirical studies that compare soil carbon stocks in true primary forests to forests that have undergone centuries-long harvest cycles.

Uncertainties for agricultural effects have to do with site-specific variation in management implementation and lack of knowledge of deep soil carbon dynamics. However, convergence of the different meta-analyses on similar figures and research in this field is quite extensive (Li et al., 2012).

The wetland estimate also is quite robust given the high sampling density of the National Wetland Condition Assessment (NWCA) of the National Aquatic Resource Surveys. The NGHGI estimate of forest cover increase is quite robust given the quality of input data.

Assessment of confidence based on evidence and agreement, including short description of nature of evidence and level of agreement

The meta-analyses of Nave et al. (2010, 2013) suggest very good agreement over forestry effects on soil carbon, although Dean et al. (2017) suggest that, over centuries, logging has had more significant effects on soil carbon. Given that the Dean et al. (2017) study is based on modeling with assumptions that are not supported in this analysis, such as that SOC is strongly related to biomass inputs, SOCCR2 is placing greater confidence in the Nave analyses (Nave et al., 2010, 2013).

The analysis by Paustian et al. (2016) suggests that there is some disagreement over agricultural management effects on SOC and that these effects are specific to local site and climatic conditions. The Li et al. (2012) meta-analysis suggests that afforestation of former croplands globally results in net SOC increases but that local results are so variable that local projection is difficult and results depend on soil type, management, and the type of tree species.

The wetland estimate is quite robust given the high sampling density of the NWCA.

Estimated likelihood of impact or consequence, including short description of basis of estimate

Conversion to agriculture is a significant source of greenhouse gases to the atmosphere and loss of soil carbon. However, across North America, mitigation strategies such as conversion to no-till or reduced-till methods, adoption of crop rotations that provide greater carbon inputs, increased residue retention, and the use of cover crops during fallow periods are reducing the impact of agriculture (Paustian et al., 2016). Similar results are seen in Canada (Soil Conservation Council of Canada 2016). Erosion of soil carbon from agricultural lands is still a significant concern (Montgomery 2007). Afforestation has caused increases in soil carbon across CONUS.

Summary sentence or paragraph that integrates the above information

Studies have shown that conversion of native land to agriculture significantly reduced soil carbon, although improved management of agricultural land has the potential to have significant positive effects on soil carbon reserves. While modeling exercises suggest that logging and management of primary forest cause a significant SOC loss, robust meta-analyses suggest that this loss is quite minimal with effective forestry management.

KEY FINDING 4

Large uncertainties remain regarding soil carbon budgets, particularly the impact of lateral movement and transport of carbon (via erosion and management) across the landscape and into waterways. By 2015, cumulative regeneration of soil carbon at eroded agricultural sites and the preservation of buried, eroded soil carbon may have represented an offset of 37 ± 10% of carbon returned to the atmosphere by human-caused land-use change (*medium confidence*).

Description of evidence base

Best estimates of the effects of erosion are summarized in Billings et al. (2010), Van Oost et al. (2007), and Wang et al. (2017). Erosion can significantly affect productivity in agricultural regions, and some authors have argued that loss of eroded carbon represents a true loss to the atmosphere (Lal and Pimentel 2008). However, work based on multiple eroding profiles indicates that approximately 26% of eroded SOC can be replaced at the eroding site, representing a small but significant carbon sink (Van Oost et al., 2007). Harden et al. (1999) suggest that U.S. cropping patterns before 1950 likely resulted in about a 20% to 30% reduction of original SOC but that on-site recovery of soil organic matter (SOM) levels occurred after the 1950s. In Canada, VandenBygaart et al. (2012) also note a net carbon sink for eroded agricultural soils. Van Oost et al. (2007) suggest that replacement of eroded SOC, along with damped SOC mineralization upon burial, may combine to generate a small net carbon sink up to about 0.1 Pg C per year. Wang et al. (2017) calculate that cumulative, agriculturally accelerated erosion prompted SOC replacement and buried SOC preservation, representing an offset of 70 ± 16% of carbon emissions by anthropogenic land-cover change up to AD 1600; after this period, the cumulative value represented a smaller offset (37 ± 10% in 2015).

Major uncertainties

The fate of eroded agricultural soil can only be modeled, not directly measured, and the production of new soil carbon after exposure of new mineral surfaces also cannot be directly measured.

Assessment of confidence based on evidence and agreement, including short description of nature of evidence and level of agreement

Erosion of soil is known to occur, but the fate of the eroded SOC is less clear. Currently, findings conclude that the eroded SOM appears to represent a small sink of carbon but that not all material is accounted for, and the geographic extent of full carbon budget studies is quite limited. Although subsurface soil carbon appears to be relatively stable, the responses to future changes in management and climate are not well understood.

Estimated likelihood of impact or consequence, including short description of basis of estimate

In the United States, conservation measures introduced after the Dust Bowl of the 1930s suggest that the potential for massive erosional losses of soil carbon are unlikely, but similar measures are not used in Mexico. In Canada, conservation measures including zero-till have been widely adopted over many regions and have resulted in soil carbon increases and reduced erosion (Soil Conservation Council of Canada 2016). Estimates for Mexico suggest that loss of soil carbon due to management practices remains significant (Huber-Sannwald et al., 2006).

Summary sentence or paragraph that integrates the above information

Large uncertainties remain in specific key areas, including the impact of lateral movement and transport of carbon through erosion and management.

KEY FINDING 5

Evidence is strong for direct effects of increased temperature on loss of soil carbon, but warming and atmospheric carbon dioxide increases also may enhance plant production in many ecosystems, resulting in greater carbon inputs to soil. Globally, projected warming could cause the release of 55 ± 50 Pg C over the next 35 years from a soil pool of 1,400 ± 150 Pg C. In particular, an estimated 5% to 15% of the peatland carbon pool could become a significant carbon flux to the atmosphere under future anthropogenic disturbances (e.g., harvest, development, and peatland drainage) and change in disturbance regimes (e.g., wildfires and permafrost thaw) (*medium confidence*).

Description of evidence base

Although many laboratory experiments have shown that soils respond to increased temperature with increased respiration, there are many potential causes for this increase, including increased belowground inputs (Giardina et al., 2014) or increased plant production (Phillips et al., 2016). A global meta-analysis has shown that soil respiration increases with temperature (Bond-Lamberty and Thomson 2010), but how much of this is due to turnover of new, labile plant inputs is unclear (reviewed in Bradford et al., 2016). Empirical relationships developed by Crowther et al. (2016) suggest that global soil carbon stocks in the upper soil horizons will fall by 30 ± 30 Pg C under a temperature increase of 1°C, and 55 ± 50 Pg C with expected warming in the next 35 years, depending on the rate at which the effects of warming are realized.

Many studies have suggested that peatlands and boreal ecosystems are particularly vulnerable to warming (Bridgham et al., 2008; Dise 2009; Hicks Pries et al., 2015; Koven et al., 2015) because of factors such as permafrost thawing and drying effects on decomposition (Ise et al., 2008), increased fire from drying (Turetsky et al., 2014), and poleward expansion of low-carbon ecosystems (Koven 2013). Thawing of sporadic and discontinuous permafrost may release up to 24 Pg C currently stored in boreal peatlands over decades to centuries (Jones et al., 2017). Wildfire combustion of organic soils across permafrost-dominated landscapes can produce carbon losses ranging from 2.95 ± 0.12 to 6.15 ± 0.41 kilograms of carbon per m^2, depending on the season (Turetsky et al. 2011).

Major uncertainties

Most laboratory experiments demonstrate that warming causes the loss of soil carbon, but how soils in natural ecosystems will respond to global warming is less predictable, given the different possible trajectories of plant production responses in different ecosystems and the possibility of increased plant production matching elevated soil respiration (Xu et al., 2016). Acclimation of soil microbes to warming could modulate the response of soils (Luo et al., 2001), although a meta-analysis (Wang et al., 2014) suggests that heterotrophic activity will not significantly acclimate to warming.

Assessment of confidence based on evidence and agreement, including short description of nature of evidence and level of agreement

At current rates of carbon dioxide and temperature increase, peatlands are highly likely to release a significant amount of stored soil carbon. Less certain is whether soils in other ecosystems, especially those subject to drought, will respond similarly to elevated temperature.

Summary sentence or paragraph that integrates the above information

The release of carbon from peatland soils could represent a major positive feedback loop to continued disturbance regimes related to climate change and human activities.

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