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RESEARCH REVIEW

Soil organic carbon across scales

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Abstract

Mechanistic understanding of scale effects is important for interpreting the processes that control the global carbon cycle. Greater attention should be given to scale in soil organic carbon (SOC) science so that we can devise better policy to protect/enhance existing SOC stocks and ensure sustainable use of soils. Global issues such as climate change require consideration of SOC stock changes at the global and biosphere scale, but human interaction occurs at the landscape scale, with consequences at the pedon, aggregate and particle scales. This review evaluates our understanding of SOC across all these scales in the context of the processes involved in SOC cycling at each scale and with emphasis on stabilizing SOC. Current synergy between science and policy is explored at each scale to determine how well each is represented in the management of SOC. An outline of how SOC might be integrated into a framework of soil security is examined. We conclude that SOC processes at the biosphere to biome scales are not well understood. Instead, SOC has come to be viewed as a large-scale pool subjects to carbon flux. Better understanding exists for SOC processes operating at the scales of the pedon, aggregate and particle. At the landscape scale, the influence of large- and small-scale processes has the greatest interaction and is exposed to the greatest modification through agricultural management. Policy implemented at regional or national scale tends to focus at the landscape scale without due consideration of the larger scale factors controlling SOC or the impacts of policy for SOC at the smaller SOC scales. What is required is a framework that can be integrated across a continuum of scales to optimize SOC management.

Keywords: aggregate, biome, landscape, management, profile, scale, soil organic carbon, soil particle, soil policy, soil security

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Introduction

Much recent research on SOC has been driven primarily by the belief that soils can be used to offset increased concentrations of atmospheric CO_2. Much effort has focused on scale extremes, to develop a mechanistic understanding of SOC at the particle and aggregate scale (Schmidt *et al*., 2011) and to estimate global and national SOC stock (Schrumpf *et al*., 2011). There is some concern that soil may contribute to elevated CO_2 and global climate change (Davidson & Janssens, 2006; Smith *et al*., 2008; Blagodatsky & Smith, 2012). The response of SOC to changing global temperatures is being studied at small scales (see Von Lützow & Kögel-Knabner, 2009) and could in theory be extrapolated from predicted net primary production using climate and land-cover data (Del Grosso *et al*., 2008) to larger scales.

We now probably know enough about SOC to manage soils for SOC sequestration potential and stock protection. Whether this is operationally feasible at the farm level is unclear. Many complex challenges related to scale remain because SOC is a complex material with variable interactions of chemical composition with the mineral fraction and decomposition kinetics (Baldock & Nelson, 2000). Many climate, soil, crop and management factors affect SOC so quantitative data are not always transferable and applicable between sites, particularly in the case of land-use and land-cover change. Global issues such as climate change require consideration of SOC stock changes at biosphere or biome scales, but human interaction occurs at the landscape scale, with consequences at the pedon, aggregate and particle scales. An integration of our understanding of SOC at all scales is required to formulate solutions that are relevant and consistent from both scientific and policy perspectives because weak understanding of scale effects has limited the translation of SOC science into policy.

Growing demands on soils to provide food, water and energy security, protect biodiversity and abate climate change has led to the concept of 'soil security', a

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framework integrating biophysical, social, economic and political sciences (McBratney *et al.*, 2014). SOC might be a universal indicator of soil security (Koch *et al.*, 2013), where SOC lost from soil reduces soil security. Our objective is to review the processes controlling stabilization and distribution of SOC across scales, to examine the current synergy between science and policy and outline how SOC can be integrated into the framework of soil security relevant at each scale.

Particle scale

Primary mineral particles (sand; <2 mm to 50 μm, silt; <50 to 2 μm, and clay; <2 μm, Soil Survey Staff, 1975) largely determine inherent soil properties. Soil texture is used to estimate how soils will respond to management (Skopp, 2000) and is hypothesized to be a major determinant of SOC sequestration potential (Stockmann *et al.*, 2013). SOC associated with mineral particles is considered a basic unit in the soil (Christensen, 2002), and it is suggested the amount of complexed organic carbon controls soil physical behaviour (Dexter *et al.*, 2008). Meta-analysis of SOC accumulation in agricultural soils (West & Six, 2007) and its recovery after afforestation (Laganière *et al.*, 2010) have shown that capacity for carbon storage increases with clay content, but depending on global location and management. Far less attention is given to clay mineralogy that causes different microstructures (Oades & Waters, 1991; Barré *et al.*, 2014). Localized patches of OM on mineral surfaces (Mayer, 1999) are known to exist at active charge sites (Schoonheydt & Johnston, 2006). However, the monolayer-equivalent amount of SOC cover on fine particles has become a quantified descriptor based on the sum total of OM and mineral-specific surface area (SSA) (Mayer, 1994), and a maximum loading of 1 mg C m^{-2} has been used for soils and sediments (Keil *et al.*, 1994). Feng *et al.* (2013) demonstrated that the monolayer assumption might underestimate carbon stabilization and suggested an alternative method to calculate the organic carbon protective capacity using data from soils assumed to be at SOC saturation. They found SOC stabilization was more than double previous estimates (78 $\pm$ 4 g C kg^{-1} soil compared to 33 $\pm$ 1 g C kg^{-1} (Hassink, 1997) when an upper boundary line rather than a least square regression line was used to infer the maximum organic carbon protective capacity from a standard linear regression). Accounting for the SSA of clay and silt fractions when calculating SOC loading [mass of fine soil particles (mg C m^{-2})/SSA (m^2 g^{-1})] for agricultural soils receiving long-term manure showed that 44% of silt and clay fractions have greater loading than the assumed maximum of 1 mg carbon m^{-2} (Feng *et al.*, 2014). This indicates that SOC stabilization may be underpredicted because of the focus on texture, rather than its interaction with mineralogy.

At the particle scale, the dominant SOC protection processes are chemical (SOC sorbed on to silt and clay particles as 'organo-mineral complexes') and biochemical (biologically stable or 'recalcitrant' SOC). Turnover rates for mineral-associated SOC likely relate to the time when organic molecules are sorbed, where molecules first sorbed are most stabilized and may represent ages of decades to centuries (Kögel-Knabner *et al.*, 2008). A conceptual model of organo-mineral association (Kleber *et al.*, 2007) describes a discrete zonal sequence of layers: (i) the contact zone, a stable inner sphere formed by absorption of amphiphilic N-rich molecules to mineral surfaces; (ii) the hydrophobic zone, a membrane-like bilayer of apolar amphiphilic molecules that protects the inner sphere; and (iii) the kinetic zone, an outer layer of organic molecules weakly retained by cation bridging and H bonds, with high potential for exchange with soil solution. Research on the attachment of new OM to pre-existing OM on clay particles has highlighted the role of Fe oxides in cohesion of organic carbon–iron chelates (Bachmann *et al.*, 2008; Lalonde *et al.*, 2012). Vogel *et al.* (2014) offered new insight into particle-scale spatial variability of SOC accumulation, showing SOC to be preferentially associated with rough surfaces (etch pits, micropores and cracks) that have a greater capacity for SOC storage than predicted by monolayer theory. They concluded that mineral particles with smooth surfaces might not be suited to OM attachment. However, the importance of molecular structure in the persistence of SOC remains unclear because data from 20 long-term experiments demonstrated the turnover of complex macromolecules is more rapid than bulk soil, or molecules sorbed in organo-mineral associations (Schmidt *et al.*, 2011).

Aggregate scale

Binding agents (clay, OM and polyvalent cations) cement primary mineral particles together to form micro-aggregates (<250 μm), which are bound by temporary binding agents (microbial- and plant-derived polysaccharides, fine roots and fungal hyphae) to form macro-aggregates (>250 μm) (Tisdall & Oades, 1982; Kay & Angers, 1999; Six *et al.*, 2004). Pore architecture reflects aggregation, with pores between and within micro-aggregates and macro-aggregates (Elliott & Coleman, 1988). As the decomposer microbial community has the ability to degrade most kinds of carbon, the key to ensuring long-term preservation of SOC may be the

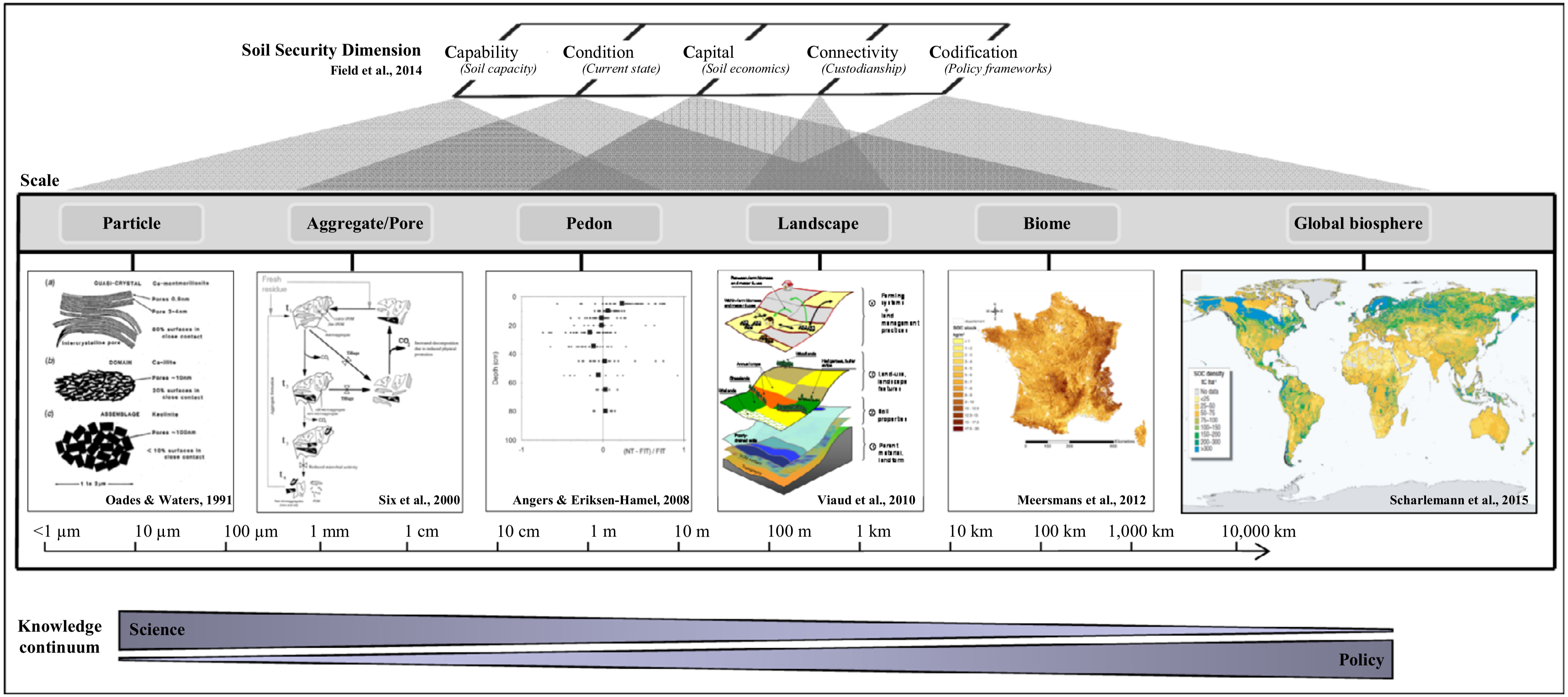

Fig. 1 A conceptual diagram depicting how we view soil organic carbon at each scale. The knowledge continuum is a representation of how informed the science and policy are at each scale, respectively. The relevance of soil security dimensions to each soil organic carbon scale is highlighted.

location of carbon substrates in biologically nonpreferred spaces or disruption of biological decomposition (Ekschmitt *et al.*, 2008). For example, a study of internal pore morphology and SOC distribution within micro-aggregates along a prairie chronosequence found micro-aggregates from cultivated prairie soil had the lowest concentration of SOC, but that 75% of the remaining SOC was in OM-filled pores. Following restoration, SOC in pore spaces increased to 90% (McCarthy *et al.*, 2008). Furthermore, Virto *et al.* (2013) showed that SOC in aggregates has longer turnover time than SOC associated with clay-sized particles in cultivated silty soils. SOC within the pore structure probably creates spatial and kinetic constraints on microbial activity because of microbial biogeography, where community structure differs by pore size (Ruamps *et al.*, 2011). More recent work emphasizes that microbial metabolism (*biotic*) is less important than habitat/location in the pore network (*abiotic*) in controlling mineralization of SOC (Ruamps *et al.*, 2013). Additionally, a positive feedback of increasing porosity with increasing carbon content arises from decomposition of fresh residue causing voids (De Gryze *et al.*, 2006) followed by the invasion of mycelium to access the decomposing OM (Emerson & McGarry, 2003). Due to temporary binding, macro-aggregates are more dynamic than micro-aggregates and experience fluctuation in size and abundance in response to weather and management (Six *et al.*, 2000a; Christensen, 2002) (Fig. 1).

Physical protection (barriers between carbon substrate and microbes) is the dominant cause of SOC stabilization at the aggregate scale. Bacteria access is limited by micropores <0.2 μm diameter (Männik *et al.*, 2009); thus, physical protection is predominant in micro-aggregates compared to macro-aggregates (Six *et al.*, 2000a). Macro-aggregate formation is a rapid process driven by the amount of carbon input and fungal activity (De Gryze *et al.*, 2005). Macro-aggregates can develop within 7 days in a loess soil, containing many times more SOC than the parent material, but are relatively unstable so act as a carbon source for microbial decomposition (Andruschkewitsch *et al.*, 2014). The degree of aggregation depends on the size distribution of organo-mineral associations and their chemical characteristics (Christensen, 2002). In soil with similar clay mineralogy but different carbonate content subjects to the same agricultural management over decades, the carbonate-rich soil (20% clay) had about half the amount of >20 μm pores in macro-aggregates >2 mm compared to the decarbonated soil (10% clay) (Virto *et al.*, 2013). Soil disruption causes micro-aggregate pore space change that makes SOC diffusion into the centre of micropores more difficult. Soil aggregates (~5 mm) subject to conventional tillage have shorter path lengths (<200 μm) and nodal pore volumes (mode ~7.9×10^{-7} mm^3/63 μm^2) that tend to limit gas and water transport compared to aggregates in grassland (path lengths <600 μm; nodal pore volumes, mode ~5.0×10^{-6} mm^3/400 μm^2) and thus reduce the potential for OM accumulation in the aggregate interior (Peth *et al.*, 2008). Consequently, morphological characteristics of aggregate pores play a role in aggregate stability and prolonged protection of SOC.

Pedon scale

The amount of SOC in the soil pedon is the net balance between additions and losses through stabilization and destabilization processes (Sollins *et al.*, 1996). After addition of fresh carbon to the soil, decomposition results in a large proportion of the initial biomass, or labile SOC, being lost in 1–2 years (the 'fast' SOC pool), followed by slow decomposition over 10–100 years (the 'intermediate' SOC pool), and the remaining SOC enters a period of very slow turnover of 100 to >1000 years (the 'slow' SOC pool) (Jenkinson & Ladd, 1981; Falloon & Smith, 2000). It is the proportion of labile SOC to total SOC, rather than the total SOC pool that influences soil quality (Loveland & Webb, 2003) because the quantity and quality of the carbon input is the driving force for microbial mineralization (Lagomarsino *et al.*, 2006) and nutrient supply to growing plants. Although some soil-specific SOC thresholds have been determined for optimum crop production (Hassink, 1977; Zvomuya *et al.*, 2008), it is possible that universally acceptable SOC thresholds might be identified once the importance of labile SOC to specific soil functions can be quantitatively defined. Evidence for a constant stoichiometric ratio of C:N:P:S means that these nutrients will be necessary to stabilize SOC and will become locked-up in the soil (Kirkby *et al.*, 2011); therefore, a soil with a small SOC store turning over quickly might sustain more and better function than a soil with a large SOC store and little labile content (Janzen, 2006, 2015; Lagomarsino *et al.*, 2006). Climate change alters the soil stoichiometry (CO_2 enrichment increases P availability to plants and microbes relative to N, whereas warming reduces P availability relative to N (Dijkstra *et al.*, 2012) with feedbacks to plant productivity and SOC sequestration. Ultimately, the objective should be to determine the optimum balance of labile SOC and nutrient input for soil quality and SOC sequestration purposes.

Distribution of SOC in the pedon depends on the allocation of above- and below-ground plant litter. The root-to-shoot ratio is high for tundra, grasslands and cold deserts (4–7) and low for forests and croplands

(0.1–0.5) (Kögel-Knabner, 2002). Based on data for different crops, grasses and legumes, root-derived SOC has a 2.4-fold longer residence time than shoot C due to the chemical recalcitrance of root tissues (Rasse *et al.*, 2005), but fine root turnover undergoes more rapid decomposition and represents approximately 33% of annual net primary production, depending on species (Jackson *et al.*, 1997). Worldwide, about 75% of roots are in the top 40 cm (Jackson *et al.*, 1996) and management practices mainly influence the top 30 cm (Bationo *et al.*, 2007). In agricultural soils, full inversion tillage removes OM from the soil surface and buries it 15–30 cm in the soil, whereas no-till arable systems allow SOC to accumulate in soil surface layers (Angers & Eriksen-Hamel, 2008). Over the past few decades, no-till was promoted to increase SOC with benefits to soil physical properties important to maintain or boost crop yields where accumulated SOC began to be viewed as carbon sequestration and counted as SOC stock. There is current debate highlighting the risks attributed to the long-term nature of SOC accumulated under no-till, largely present in labile forms, which is at risk if ploughing were to occur again (Powlson *et al.*, 2014) and questioning if such an altered depth distribution, as a transfer of OM from surface to plough layer, represents SOC sequestration (Powlson *et al.*, 2014).

Soil organic carbon stock is usually examined to a maximum depth of 1 m depth. On a global scale, SOC ranges from 3 to 38 kg C m^{-2} in the top 1 m of mineral soil and up to 125 kg C m^{-2} in organic soil (Batjes, 1996). Recently, emphasis has been placed on quantifying deep stocks of SOC as many soils greatly exceed 1 m. Considering maximum rooting depths can reach up to 2.5 m in grassland, 5 m in forests and 7 m in shrubland, the subsoil offers substantial potential to stabilize SOC, estimated at between 760 and 1520 Pg C below 1 m (Lorenz & Lal, 2005). SOC determined for deeply weathered soils (5–38 m to bedrock; mean 21 m) across 5 locations in south-western Australia reported 51–75% of SOC occurred in the top 5 m (mean SOC concentration, $0.12 \pm 0.01\%$), below which there was little change to bedrock ($0.04 \pm 0.002\%$). However, the contribution of the deep SOC below 5 m, even at low concentrations, was on average 41% of the total SOC in the pedon (Harper & Tibbett, 2013) and suggests normal reporting might be a gross underestimation of SOC stocks. High concentrations of SOC exist at depth in wetlands, peatlands and permafrost soils (Davidson & Janssens, 2006). Other exceptions of high SOC concentrations occur as soil horizons (>1 m) buried by volcanic, aeolian, alluvial, colluvial, glacial and anthropogenic processes (Chaopricha & Marín-Spiotta, 2014). One option to increasing SOC in agricultural soils may be placing of SOC lower in the pedon where there are greater SOC binding sites (Hassink, 1996; Poirier *et al.*, 2014), and microbial and environmental conditions favour slower turnover due to low oxygen diffusivity, soil moisture and nutrient availability (Chaopricha & Marín-Spiotta, 2014). In ploughed agricultural soils, OM is buried to between 15 and 30 cm in the soil, but a meta-analysis of a large number of full inversion tillage and no-till comparative sites found the increase in SOC content around the depth of the plough was not enough to offset the effects of soil disruption (Angers & Eriksen-Hamel, 2008).

An alternative approach to placing SOC at depth is to increase carbon inputs at depth by selection of plants with long root systems (Glover *et al.*, 2010) such as perennial wheat, which can achieve 10-fold root biomass compared to annual wheat (Larkin *et al.*, 2014). Such crops mimic natural prairie ecology (Brummer *et al.*, 2011) with benefits for enhancing SOC being (i) continuous year-round supply of root litter, (ii) minimal disruption of the soil structure and (iii) increased total biomass when grown in bi-/polyculture with grasses and legumes. For example, a switch to tall fescue (*Lolium arundinaceum* Schub.) following 15 years of grain production increased SOC by 17% by the equivalent soil mass approach over 0–60 cm with a significant increase in the 40–60 cm soil depth after 7 years (Carter & Gregorich, 2009).

Probably, the greatest barrier to increasing SOC stock at depth is the stimulation of decomposers by adding fresh inputs of OM that result in microbial mineralization of old, stable SOC, a process known as the priming effect. For example, metabolically active microbes present at 60–80 cm with very little fresh SOC (500 times lower SOC compared to 0–20 cm) were simulated after the addition of cellulose, causing decomposition of new as well as >2500-year-old SOC (Fontaine *et al.*, 2007). This means that microbial biomass is not only a pool but also a driver of SOC turnover (Blagodatsky *et al.*, 2010; Kuzyakov, 2010). Various manipulations have been proposed to enhance the stabilization of SOC. One suggestion is to alter the microbial community to achieve SOC utilization efficiency through a microbial community shift from bacteria to fungi, although little is known about how to achieve this (Jastrow *et al.*, 2007). Another option is to manipulate the amount of C substrate to between 2 and 5 times the microbial biomass carbon so micro-organisms will preferentially utilize fresh substrate (Blagodatskaya & Kuzyakov, 2008). Such manipulations need to be tested in the field to verify the mechanisms of SOC persistence. This is possible using long-term, controlled experiments on pedons, such as large-scale lysimeters (Schmidt *et al.*, 2011).

Landscape scale

At the landscape scale, SOC dynamics are affected by natural and anthropogenic processes occurring in lateral and vertical dimensions (Rosenbloom *et al.*, 2006; Viaud *et al.*, 2010). Lateral movement of SOC by air or wind erosion is recognized as one of the greatest sources of uncertainty in the global carbon balance because there is debate whether anthropogenic soil erosion represents a sink or source of atmospheric carbon. If a posterosion watershed C balance is increased, relative to a pre-erosional or a noneroding scenario, by replacement of SOC by new photosynthate in the eroded slopes and stabilization of eroded SOC in depositional settings, then a watershed is defined as a net sink for atmospheric CO_2 (Berhe *et al.*, 2007, 2008). Erosional processes on hillslopes cause SOC mineralization to occur during transport, with aggregate breakdown followed by restabilization by burial that results in an increase in the mean residence time of SOC due to preservation of labile SOC at depth. The rate of CO_2 mineralization during soil erosion is regulated by the interaction of soil moisture and soil temperature on soil microbial activity (Van Hemelryck *et al.*, 2011), and the effects vary with slope position and weather (Wei *et al.*, 2014). This makes it difficult to predict CO_2 efflux during erosion and to determine the carbon source–sink function in contemporary agricultural landscapes. Such a lack of quantitative information on rates of erosion leads to uncertainty in SOC sequestration of between 0.3 and 1.0 t CO_2 ha^{-1} yr^{-1} (Sanderman & Chappell, 2013), which is similar to the expected sequestration rates of many agricultural options. It is estimated that globally 35 ± 10 Pg yr^{-1} of sediment is mobilized by approximately 80% water, 15% tillage and 5% wind erosion and corresponds to a SOC flux of 0.5 ± 0.02 Pg yr^{-1}, and a proportion of that is removed from the terrestrial system 0.08 ± 0.02 Pg, and delivered to rivers (Quinton *et al.*, 2010).

Debate is centred on the timescales of eroded SOC that are not well understood because hillslopes and floodplains developed during the Holocene might have buried sufficient material to offset the effects of carbon emissions caused by anthropogenic land-cover change (Hoffmann *et al.*, 2013a). Study of anthropogenically eroded sediment in Central Europe during the Holocene showed that sediment storage increases with drainage-basin size, with greater amounts of eroded sediment in floodplains of ~10^3 km^2 and on hillslopes in floodplains of ~10^5 km^2 (Hoffman *et al.*, 2013b). Despite this, sedimentary archives showed net SOC accumulation in floodplains (0.7 ± 0.2 g C m^{-2} a^{-1}) always exceeds that on hillslopes (0.4 ± 0.1 g C m^{-2} a^{-1}) (Hoffman *et al.*, 2013b), which likely reflects a greater net primary production in the floodplain. Attempts to model the evolution of soil slope sequences estimated erosional processes could represent a significant proportion of SOC inventory perhaps up to 40% (Rosenbloom *et al.*, 2006).Vertical and lateral redistribution of SOC in the landscape occurs as dissolved organic carbon (DOC) (Lambert *et al.*, 2011). Podzolization is a combination of leaching and translocation of SOC and cations to deeper locations in the pedon, or into the groundwater. Although DOC represents a minor amount of SOC (Cole *et al.*, 2007), introduction of clean air legislation in North American and northern Europe, which successfully reduced anthropogenic sulphur deposition [reported as a 90% reduction since 1980 by the European Monitoring and Evaluation Programme (Torseth *et al.*, 2012)], has decreased soil water acidity and ionic strength (Evans *et al.*, 2006) and exacerbated mobilization of DOC in soils (Roulet & Moore, 2006). DOC provides a nutrient and energy source for heterotrophic bacteria, and the subsequent mineralization of DOC turns aquatic systems into net sources of CO_2 (Duarte & Prairie, 2005). In small homogenous catchments, hydrological functions control the temporal variability of stream water DOC, whereas in large heterogeneous catchments, it is a combination of hydrological mechanisms and major landscape elements (i.e. wetland or forested areas) that control movement of DOC (Laudon *et al.*, 2011). To understand the fate of SOC movement in the landscape scale, greater understanding of soil surface water hydrology, hydrological pathways and catchment hydrology is required to quantify the lateral and horizontal movement of eroded sediment and this hydrological information needs to be linked to accurate and spatially relevant information on SOC with which to effectively model SOC dynamics (Viaud *et al.*, 2010). Advancement of digital soil mapping, combined with soil depth functions such as equal area spline (Malone *et al.*, 2009) or exponential depth functions (Minasny *et al.*, 2006), makes it possible to estimate vertical and lateral distribution of SOC across a landscape with limited soil data.

Biome scale

The biome describes a major and distinct regional element of the biosphere, characterized by typical communities of plants and animals, consisting of several ecosystems within a region of similar climate (IPCC, 2014). At the biome scale, climate and its interaction with vegetation control the SOC storage capacity of soils. From a global perspective, total SOC content is positively correlated with mean annual precipitation and clay content and negatively correlated with mean annual temperature (Jobbágy & Jackson, 2000). There is little process-level understanding of SOC at this scale, so meta-analysis is the most common approach to understanding biome-scale control. Whole-ecosystem carbon stock (defined as carbon in live biomass and SOC to 1 m depth) and carbon turnover time (Carvalhais *et al.*, 2014) are estimated for major biomes as follows: tropical forests (702 Gt; 14.2 years), boreal forest (505 Gt; 53.3 years), temperate forest (292 Gt; 23.5 years), tropical savannas and grasslands (285.3 Gt; 16.0 years), temperate grasslands and shrublands (182 Gt; 41.3 years), deserts (250 Gt; 36.3 years), tundra (156 Gt; 65.2 years), croplands (362 Gt; 22.1 years), and wetlands (20 Gt; 19.7 years). There is a dependence of turnover time on temperature, but a strong association between SOC turnover time and precipitation suggests that hydrological control may be as relevant as temperature in future climate/carbon cycle feedbacks, which will require a better understanding of changes to the hydrological cycle (Carvalhais *et al.*, 2014). The most uncertain part of the global C cycle is terrestrial carbon or the net carbon flux arising from land-use and land-cover change (LULCC) (Houghton *et al.*, 2012), estimated to have contributed approximately 11% of global C emissions from 2000 to 2009 (7.8 ± 0.4 Gt C yr^{-1} fossil fuel; 1.0 ± 0.5 Gt C yr^{-1} LULCC) (Le Quéré *et al.*, 2013). Insufficient data exist to determine the effects of land-use change on SOC stock within any particular biome. Instead, data on land-use change are more indicative of the landscape scale; for example, meta-analysis has determined that decreases in SOC occur due to conversion from pasture to plantation (−10%), native forest to plantation (−13%), native forest to crop (−42%) and pasture to crop (−59%), and increases in SOC occur due to conversion from forest to pasture (+8%), crop to pasture (+19%), crop to plantation (+18%) and crop to secondary forest (+53%) (Guo & Gifford, 2002). Bulking of land-use data across biomes, and considering vegetation in the absence of climate, results in a loss of information about the climatic controls on SOC and should be used with caution in carbon accounting and national policy. The same caution should be applied to regional or national investigation of the impact of climate change on SOC because several types of biome which respond to climate in different ways are bulked in the analysis to produce a mean rate of SOC change over time. For example, SOC rates of change are −0.6% yr^{-1} from 1979 to 2003 in England and Wales (Bellamy *et al.*, 2005; biomes = temperate forest, grassland, shrubland, cropland, wetlands), −0.9% yr^{-1} between 1990 and 2004 for a mountainous French region (Franche-Comté) (Saby *et al.*, 2008; biomes = temperate grassland, croplands) and −2% yr^{-1} from 1930 to 1940, −0.7% yr^{-1} from 1960 to 1970 and +1.1% yr^{-1} by 2000 in Java, Indonesia (Minasny *et al.*, 2010, biomes = croplands). It is reasonable to summarize that the biome represents a scale boundary that relates to a transition from understanding SOC in terms of processes at the smaller scales to thinking about SOC as large pools subject to carbon flux. A lack of process-level understanding at the biome scale along with the difficulty of relating biomes to the scales of human management (i.e. farm to national boundaries) means that there is great difficulty in approaching SOC management from the biome scale.

Biosphere scale

Soil organic carbon is an important component in the global carbon cycle, the biogeochemical cycle by which carbon is exchanged among the biosphere, pedosphere, geosphere, hydrosphere and atmosphere. In the global carbon cycle, SOC represents the largest component of the terrestrial biosphere carbon pool (Scharlemann *et al.*, 2015). Global SOC stocks down to a depth of 1 m are estimated to be about 1500 Pg C. Scharlemann *et al.* (2015) calculated a median estimate of 1437 Pg C with a range of 504 to 2469 Pg C from 7 studies using spatially explicit data. One issue with uncertainty of SOC stocks is that they are estimated from soil maps developed in the 1970s, which map the components of a landscape into soil classes. The development of a new digital soil map of the world using emerging technologies for soil mapping (see Arrouays *et al.*, 2014) is currently underway. This involves techniques such as spatial disaggregation using modern numerical techniques to map the soil classes of polygons in legacy soil maps individually (Odgers *et al.*, 2014). This will allow for prediction of SOC at pedon to landscape scale to provide the best estimate of SOC along with the confidence of prediction to a depth of 2 m at a spatial resolution of 100 m at the global scale in the future. The most detailed map to date represents SOC distribution to a depth of 1 m at a spatial resolution of 1 km (Hiederer & Köchy, 2011), although SOC stocks may be underestimated where SOC is stored at depth in mineral soils and in deep organic soils beyond 1 m (Lorenz & Lal, 2005; Davidson & Janssens, 2006; Chaopricha & Marín-Spiotta, 2014). An examination of the terrestrial carbon pool by IPCC climate region, a bigger scale than the biome (tropical wet, tropical moist, tropical dry, tropical montane, warm temperate moist, warm temperate dry, cool temperate moist, cool temperate dry, boreal moist, boreal dry, polar moist and polar dry), highlighted that the largest SOC and biomass carbon stock are subject to the greatest risks (Scharlemann *et al.*, 2015).

At the biosphere scale, it is possible to estimate SOC pools and their fluxes, but there is little process understanding of SOC. Processes operating at this scale are influenced by climate and latitude and require a top-down approach to isolate the effects of climate, climate change and climate–plant community interaction. For example, attempts to characterize the greatest risks to SOC stocks require assessment over several scales. Approximately 25% of the total global SOC stocks at northern latitudes in permafrost regions included in 'boreal moist' climate are at threat from climate change (*biosphere scale*) (Schuur *et al.*, 2008), whereas deforestation remains the largest source of net carbon flux from land-use change (*landscape scale*) (Friedlingstein *et al.*, 2010) which continues to pose a threat to 'wet' and 'moist' tropical forests (*biome scale*) which store almost 60% of biomass carbon. In contrast, processes operating at the smaller scales require a bottom-up approach. These approaches meet in the middle at the landscape scale, where the influence of large- and small-scale processes has the greatest interaction. Current SOC management reflects the sum of policy at regional and national boundaries aimed at the landscape scale. For effective global management of SOC, it is important that the correct scale is identified for SOC policy and management.

Integration of SOC science into policy

Primary soil particles have a definitive effect on SOC stabilization. Translating what is known about SOC at the particle scale into meaningful policy can be achieved through the concept of SOC saturation. The saturation of SOC can in theory be used to estimate the unique and finite capacity of a soil to stabilize OM (Hassink, 1997; Baldock & Skjemstad, 2000; Six *et al.*, 2002; Chung *et al.*, 2008; Angers *et al.*, 2011; Beare *et al.*, 2014; Wiesmeier *et al.*, 2014), although this cannot be estimated with absolute confidence yet. In simple terms and without consideration of soil mineralogy, an upper limit of potential SOC stabilization can be predicted from the mass proportion of the fine fraction (clay + silt) (Six *et al.*, 2002; Angers *et al.*, 2011). Soil is likely to reach its SOC saturation level asymptotically (Stewart *et al.*, 2008), whereby the final phase of SOC sequestration is slowest and not fully understood in terms of SOC kinetics (Stewart *et al.*, 2007). A reasonable policy setting might be to set an upper level of steady-state SOC concentration (i.e. such as 90% of SOC saturation). This approach still requires a site- or region-specific component to account for climate and interacting factors which influence the theoretical ability of soil to reach SOC saturation (Post *et al.*, 1982; Stockman *et al.*, 2013). Estimation of SOC saturation without consideration of SSA will lead to an underestimation of potential SOC storage (Feng *et al.*, 2013). Eventually, our understanding of SSA and SOC saturation must match up to avoid an underestimation of SOC sequestration potential. At the particle scale, policy can progress with targets based on SOC saturation estimates. Lower limits for SOC, below which a serious decline in soil quality and crop yields can occur (Stockmann *et al.*, 2013), also need to be identified in policy. This will depend on the same kind of mineralogical, texture and climatic conditions.

At the aggregate scale, encapsulation of OM in micro-aggregates is an important physical protection mechanism (Six *et al.*, 2000a; Ekschmitt *et al.*, 2008). Soil disturbances such as tillage increase the rate of macro-aggregate turnover and diminish the formation of new micro-aggregates along with the opportunity for sequestration of SOC within these micro-aggregates (Six *et al.*, 2000a,b). While the extent of SOC change in response to no-till management varies between soils (Six *et al.*, 2000b; Angers & Eriksen-Hamel, 2008; Powlson *et al.*, 2014), conservation tillage is universally promoted as an optimal management practice to maintain soil structure and enhance SOC. Adoption of conservation agriculture practices is currently a market-based mechanism driven by crop productivity. Uptake of reduced/no-till practices is high when benefits such as water efficiency are maintained through good soil structure and increased crop yields in countries such as Australia (Derpsch *et al.*, 2010); however, these practices are unfavourable when crop yields are reduced in cooler and/or wetter climates in no-till systems when wet soils delay planting and machinery operations (Ogle *et al.*, 2012; Soane *et al.*, 2012) and ploughing improves soil drainage. To promote a more widespread adoption of conservation agriculture, a soil suitability index for reduced/no-till agriculture should address the spatial and temporal soil moisture response to climate (Soane *et al.*, 2012) and maintenance of crop productivity levels at the local landscape scale. Some policy frameworks recognize that a loss in SOC translates as deterioration in soil physical properties and impairment of soil nutrient cycling mechanisms with implications for crop productivity (Loveland & Webb, 2003). The European Single Farm Payment includes conditions of Cross-Compliance (Council Regulation (EC) No. 73/2009; a mechanism that ensures agricultural land is kept in 'good agricultural and environmental condition') and sets a 3.4% OM content (or 2.0% SOC) as baseline condition for arable soils. While this threshold SOC value needs validation (Verheijen *et al.*, 2005) beyond the range of soils from which it was derived (Zvomuya *et al.*, 2008), it is not yet known if bulk OM content equates to stable and secure SOC. The EU Common Agricultural Policy (Freibauer *et al.*, 2004; Smith *et al.*, 2005) and US Conservation Reserve Program (Follett, 2009; Gelfand *et al.*, 2011) are both targeted at the aggregate scale and incorporate measures to set-aside agricultural land for short (1 year) to medium (10–15 years) terms, with the intended benefit of improved soil structure and SOC sequestration, but the value of these policy instruments is not clear.

At the pedon scale, existing policy is targeted at maintaining or enhancing SOC (i.e. the recent move by the EU to retain permanent grassland). Here, soil processes interact with vegetation and require both factors to be considered in policy to ensure both soil quality and SOC sequestration are appropriately balanced for crop production and building SOC stock. Policy at this scale will affect what the landscape looks like and will drive inputs to aggregate processes, but SOC accrual will occur at the pedon scale. Spatial variation in the soil landscape will mean that policy will have a heterogeneous effect because of differences in the mediating influence of the soil pedon due to major pedological and biophysical processes interacting with climate. At the pedon scale, management of deep SOC is not a policy focus because of uncertainty due to lack of information and the priming effect (Fontaine *et al.*, 2007). If policy were to evolve to regulate SOC at depth, then the inorganic carbon pool and loss of geologic carbon [i.e. in calcite and dolomite minerals in dry lands through irrigation (Amundson & Smith, 1988)] must also be considered. The permanence of SOC stock, or how long it remains in the soil over a specific time horizon (Smith *et al.*, 2005), is perhaps the most important factor for policymakers in SOC sequestration because financial contracts will need to cover periods up to 100 years or more. A greater understanding of the magnitude and frequency of risks of SOC reversal is needed in soil policy (Schmidt *et al.*, 2011; Murray & Kasibhatia, 2013).

At the landscape scale, crop production is well managed using integrated farming practices. Tapping into this management using decision support systems that integrate nutrient management with carbon sequestration in target zones based on both saturation deficit and yield potential might be a useful direction of research to align both SOC sequestration and food security. Such an approach would highlight the impact of adopting conservative farming practices for SOC sequestration and agricultural productivity on farm income (Kragt *et al.*, 2012). It would also highlight zones to avoid, such as agricultural lowlands prone to pollution swopping or loss of eroded sediment. A decision support system would be required to integrate measures of catchment management (i.e. policies such as the EU Water Framework Directive and US Watershed Protection and Flood Prevention) that aim to improve hydromorphology and reduce nutrient pollution from agriculture, which in turn reduces DOC loss to surface/groundwater. A key policy issue at the landscape scale is that information about landscape processes needs to be integrated to avoid unwanted side effects and to achieve win-win outcomes to increase farm productivity and SOC stock.

Soil policy implemented at national jurisdiction scale is predominantly concerned with soil contamination and is focused on the landscape (examples of countries with national soil policy are Germany (Federal Soil Protection Act) and the Netherlands (Dutch Soil Protection Act)). Agreement has not yet been reached on a European Union Soil Framework Directive. Similarly, there is no US-wide soil protection instrument. Often the argument presented for such multilateral policy is that soil can be managed better at landscape scale because of its relatively stationary nature and does not have the same transboundary concerns as air or water. This is at odds with international efforts to monitor SOC where biome-scale estimates of SOC stocks and changes are relevant as environmental indicators, but for the UNFCCC, the United Nations Convention on Biodiversity (UNCBD) and the Organization for Economic Co-operation and Development (OECD), these are approximated by national estimates of varying quality. SOC does indeed respond to transboundary drivers, particularly climate, although at much slower rates than air and water. Furthermore, SOC sequestration is the net removal of atmospheric CO_2 and thus has a direct transboundary effect. There is a case to be made that soil should be subject to policy at the biome scale to ensure consistency across national boundaries and for noncontradictory policy within individual biomes. Part of the problem is that biomes cross national boundaries, do not have sharply defined boundaries because of transition zones between biomes and are not uniform but contain a mosaic of patches. For example, China contains a large number of biomes: alpine tundra, montane forest, xeric scrubland, arid desert, temperate forest, temperate steppe, dry steppe, subtropical rainforest and tropical rainforest. A number of these biomes also overlap with neighbouring jurisdictions; in China, scrubland overlaps with Mongolia and tropical rainforest with South-East Asia. The influence of the biome is poorly translated in national policy, particularly when the focus of soil policy is not centred on SOC or its climatic controls. To properly monitor SOC at the biome scale, there is a need for sampling design within biomes that captures landscape and pedon variation.

At the biosphere scale, CO_2 levels in the atmosphere are rising at a faster rate than ever recorded (average increase to 4.1 ± 0.1 Gt C yr^{-1} in the period 2000–2005; Keeling & Whorf, 2005), leading to irreversible global climate change (Solomon *et al.*, 2008). Climate change policy can only be advanced to better manage terrestrial carbon stocks once the ambiguity surrounding the impacts of LULCC at the landscape and biome scales can be translated into large-scale processes at the biosphere. Thus, science at the biosphere scale is thus lagging behind policy. The reasons are threefold: (i) there is a well-recognized need for greater availability of data on rates of land-use change, SOC stocks of lands undergoing change and the effects of land management on SOC and biomass carbon; (ii) a consensus of LULCC terminology is required to ensure consistent accounting of terrestrial carbon flux (Pongratz *et al.*, 2014); and (iii) uncertainty arises from incomplete understanding of all the processes affecting net C flux from LULCC, such as the effects of nitrogen (Jain *et al.*, 2013). From a scientific point of view, it is the lack of process-level understanding at the biosphere scale that is the greatest deficit. A reasonable aim at the biosphere scale would be to quantify the total global residual SOC sink in terms of historic SOC lost via LULCC emissions. Having an accurate estimate of the global SOC sink potential would act as a driver for better linkage between climate change policy and SOC sequestration schemes.

Soil organic carbon management within a framework of soil security

It is clear that SOC processes at different scales can be used to understand how we should manage SOC in the context of national inventory, the farm and the field. The soil security framework suggests five dimensions that allow us to codify how we think of SOC management for long-term environmental management and agricultural productivity. The dimensions are as follows: (i) *capability*, the intrinsic capacity of a soil to produce products and ecosystems services; (ii) *condition*, the current state of the soil as modified by human activities; (iii) *capital*, the economics of soil services to health, environment and food production; (iv) *connectivity*, the stewardship of soil managers of soil products and services of the soil; and (v) *codification*, the framing of soil within policy and regulatory frameworks to secure soil (McBratney *et al.*, 2014). Each dimension can be related to scale for SOC management.

The soil security dimension relevant to SOC at the particle scale is soil capability. Soil capability is largely defined by a set of long timescale/very slowly varying soil characteristics (such as soil texture) which define the potential functionality of a soil (McBratney *et al.*, 2014). The mass and SSA of the fine fraction dictate the upper limit of SOC stabilization. Although the timescale required to reach an upper limit of SOC stabilization in the bulk soil is relatively long, between 20 and 60 years (West & Post, 2002), recent insight into SOC dynamics under incubation conditions has demonstrated that unaggregated silt and clay respond faster to C additions than mineral-associated micro-aggregate fractions, reaching SOC saturation in 2.5 years (Stewart *et al.*, 2009). This demonstrates that management can have a rapid effect on SOC accrual in the fine fraction or organo-mineral associations. Once the free mineral pool reached its C capacity, further inputs of C accumulate in other pools, the physically protected, biochemically protected and nonprotected pools. These findings fit the suggestion that soil C saturation follows a hierarchy from small to larger pools (Gulde *et al.*, 2008) in keeping with the hierarchical model of aggregate formation (Tisdall & Oades, 1982). It is important to note that in a reverse situation, where rapid loss of SOC follows disruption such as cultivation, it is the particulate/uncomplexed SOC that is rapidly lost (Cambardella & Elliot, 1992).

Soil security dimensions relevant at the aggregate/pore scale are soil capability and condition. The capability of any given soil refers to its potential functionality (McBratney *et al.*, 2014). This dimension recognizes that there are intrinsic differences between soils. Within soils, management can cause a shift between the natural/reference state (*genoform*) to an altered state (*phenoform*) (Droogers & Bouma, 1997) and thus cause degradation to, or enhancement of, soil capability. In the context of SOC storage, the soil capability depends on the type of soil particles present and the extent to which aggregate hierarchy is expressed, as SOC concentration does not always increase with aggregate size (Six *et al.*, 2000b). Soil condition refers to the current state of the soil, or in other words, how reflective the soil is of its genoform compared to some other phenoform. For a given capability (i.e. soil texture), the condition can vary quickly under management, and this will be observed at the aggregate scale first. For example, an increase in cultivation intensity results in a loss of C-rich macro-aggregates and an increase in C-depleted micro-aggregates (Six *et al.*, 2000a,b). Alternatively, a soil actively managed to sequester C can exceed its equivalent genoform SOC capacity, with exceeding SOC accumulating in unbound forms (Hassink, 1997). This is achieved by shifting the genoform C equilibrium in the direction of C saturation by enhancing soil C inputs, decreasing soil C losses or through a combination of both. Therefore, to achieve an enhanced phenoform, a larger pool of labile SOC pool must be established and maintained. This benefits overall soil quality and guarantees

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a higher proportion of SOC progresses to C pools with intermediate to slow turnover.

At the soil pedon-scale capability, condition and capital dimensions are relevant. Fertile soil is considered as replenishable natural capital (Aronson *et al.*, 2007). Placing a monetary value on soil assets, where SOC plays a pivotal role in maintaining plant nutrition, greater aggregate stability and water holding capacity, will serve to secure those assets (McBratney *et al.*, 2014). Together the soil capability and condition contribute to the value, or capital, associated with existing soil C stock. In current SOC sequestration programmes, payment is calculated on a flat rate of tonne of C sequestered per hectare (Graff-Zivin & Lipper, 2008). Based on the perceived SOC that can be sequestered, soil capacity and condition may influence entry into and payment under more sophisticated SOC sequestration programmes in the future.

Soil organic carbon is managed at the landscape scale, largely through farming, and thus most dimensions overlap here: condition, capital, connectivity and codification. Connectivity refers to the 'stewardship' of the soil. For landowners, this carries the recognition of soil as a non-renewable resource that is to be valued as an asset within the farm resource base (soil, water, habitat, climate and people). Life cycle assessment may pave the way for determining the actual value of soil fertility and SOC in agricultural products (Garrigues *et al.*, 2012; Petersen *et al.*, 2013) (as investigated for energy crops; Brandão *et al.*, 2011; Lindorfer *et al.*, 2014). Land management practices chosen by farmers influence the soil condition. Increasingly, society is becoming interested in commitment to good stewardship of land, and in turn, food industries are keen to demonstrate they are caring for the land. Codification, or public policy and regulation on soils, is implemented at the farm/landscape scale. Such policies serve to recognize and pay for the provision of ecosystem services such as SOC storage and to prevent agriculture from being the source of disservice such as greenhouse gas emissions (Power, 2010).

Codification, in recognition of national capital, is implemented by jurisdictions, where regional or national policy often has to deal with the complexity of managing multiple biomes across several climatic zones. To date, policy based on sustainable development has been largely responsible for shaping environmental programmes, in which soils are an undermanaged component of the agricultural ecosystem. Internationally, it has been recognized that a soil-centric policy framework is required for sustainable development (Koch *et al.*, 2013). Under such a policy framework, greater status would be given to SOC as an important regulator of soil functions and a universal indicator of soil security, placing greater emphasis on the protection and enhancement of SOC stock and restriction on large-scale LULCC. Managing SOC at the biosphere scale will allow for factors influencing SOC at the biome to be managed consistently. Such a policy framework would provide linkages to other soil-related global issues including food security, water security, energy sustainability, ecosystem service delivery, biodiversity protection and climate change abatement. A policy instrument designed to manage SOC stocks will in turn contribute to more effective policy to mitigate climate change.

Conclusion

A desirable outcome for SOC management is to be able to prescribe the most beneficial management to achieve desirable crop yield, quantify how much SOC is sequestered per cropping cycle, determine how stable the sequestered SOC is in soil, and estimate how long it will take to utilize the sink potential of that specific soil and arrive at SOC saturation. This calls for better integration of scientific knowledge across scales for effective SOC management.

Soil organic carbon processes at the biosphere to biome scales are not well understood. Instead, SOC has come to be viewed as a large-scale pool subject to carbon flux. A top-down approach is necessary to isolate the effects of climate, latitude and plant community type on SOC. In contrast, better understanding exists for SOC processes operating at the smaller scales of the particle, aggregate and pedon. A bottom-up approach is necessary to integrate SOC processes in the soil system. These two approaches meet in the middle at the landscape scale, where the influence of large- and small-scale processes has the greatest interaction and is exposed to the greatest modification through agricultural management. Policy at regional or national scales tends to focus at the landscape scale without due consideration of the factors controlling SOC at larger scales or the impacts of policy for SOC at the smaller SOC scales. What is required is a framework that can be integrated across a continuum of scales to optimize SOC management.

The concept of soil security with biophysical-, social-, economic- and political science-based dimensions incorporates scale-related considerations and lends itself to the scale-related aspects of SOC management. Management of SOC through such a multidimensional framework enforced at the appropriate scales will facilitate in more connected science, give rise to more transparency in upscaling/downscaling of SOC management decisions and in turn deliver soil C security.

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