Review

The knowns, known unknowns and unknowns of sequestration of soil organic carbon

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ABSTRACT

Soil contains approximately 2344 Gt (1 gigaton = 1 billion tonnes) of organic carbon globally and is the largest terrestrial pool of organic carbon. Small changes in the soil organic carbon stock could result in significant impacts on the atmospheric carbon concentration. The fluxes of soil organic carbon vary in response to a host of potential environmental and anthropogenic driving factors. Scientists world-wide are contemplating questions such as: 'What is the average net change in soil organic carbon due to environmental conditions or management practices?', 'How can soil organic carbon sequestration be enhanced to achieve some mitigation of atmospheric carbon dioxide?' and 'Will this secure soil quality?'. These questions are far reaching, because maintaining and improving the world's soil resource is imperative to providing sufficient food and fibre to a growing population. Additional challenges are expected through climate change and its potential to increase food shortages. This review highlights knowledge of the amount of carbon stored in soils globally, and the potential for carbon sequestration in soil. It also discusses successful methods and models used to determine and estimate carbon pools and fluxes. This knowledge and technology underpins decisions to protect the soil resource.

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Contents

1. Introduction

Approximately 8.7 Gt (1 gigaton = 1 billion tonnes) of carbon (C) are emitted to the atmosphere each year on a global scale by anthropogenic sources (Denman et al., 2007; Lal, 2008a,b). However, the atmospheric increase has been in the order of 3.8 Gt C yr^{-1} (rate of increase in the year 2005, Denman et al., 2007), highlighting the important regulatory capacity of biospheric C pools (Le Quéré et al., 2009). In this context, soil organic carbon (SOC) and its potential to become a 'managed' sink for atmospheric carbon dioxide (CO_2) has been widely discussed in the scientific literature (e.g. Kirschbaum, 2000; Post and Kwon, 2000; Guo and Gifford, 2002; Lal, 2004a,b, 2008a,b; Post et al., 2004, 2009; Smith, 2008; Chabbi and Rumpel, 2009; Luo et al., 2010b). Here, we use the term SOC to define C in soil derived from organic origins. The term soil organic matter (SOM) is also used frequently in the literature and is generally agreed to contain about 58% SOC (i.e. elemental C). SOM is a mixture of materials including particulate organics, humus and charcoal along with living microbial biomass and fine plant roots.

To reward 'good' management of the soil C pool leading to enhanced soil carbon sequestration (SCS), there are a number of overarching questions that need to be considered in relation to the potential of the soil–plant system to 'sequester' organic C, where sequestering soil carbon requires a stipulated duration timeframe (usually 100 years) in order to be considered a 'permanent' increase under managed agricultural systems. SCS implies an increase in soil C for a defined period against a baseline condition where the increased C is sourced from atmospheric CO_2. This implication helps to frame the following questions:

1.1. What is the overall purpose of soil carbon sequestration?

What is the value in increasing the inputs to soil organic matter aside from its role in potential sequestration? For how long must this increase be maintained to be considered as SCS? Consequently, what is more important, long-term SCS or the functioning of the soil? Are these roles essentially inseparable?

1.2. How do we support SCS as a science community?

How can the benefits of SCS be promoted among policy makers/farmers/landholders (i.e. the potential of SCS to mitigate climate change, the use of SCS as a platform for sustainable agriculture) and how can suitable answers to questions such as measurement, modelling, monitoring and permanence for SCS and/or management advice be followed through? Is there a need to improve models of SOM dynamics in order (i) to demonstrate better understanding of the functioning of the soil ecosystem and (ii) to better assist landholders/farmers with management decisions?

Can both of these questions be addressed within the same model?

This article synthesizes current soil C research and highlights a number of key research areas.

2. Soil—A terrestrial pool of organic carbon

Globally, the quantity of C stored in the soil is second only to that in the ocean (38,400 Gt). While the terrestrial biotic C pool is ~560 Gt of organic C (Fig. 1), the soil C pool is more than four times this figure. The organic C pool capacity of world soils has been variously estimated for principal biomes (refer to Table 1). For instance, approximately 2344 Gt of organic C is stored in the top three meters of soil, with 54% or 1500 Gt of organic C stored in the first meter of soil and about 615 Gt stored in the top 20 cm (Jobbágy and Jackson, 2000; Guo and Gifford, 2002). This contrasts with the ~9 Gt addition of anthropogenically liberated 'new' C that is added to the atmosphere annually from fossil C sources (coal, oil and gas) and through ecosystem degradation. Several points quickly follow. For example, a change of just 10% in the SOC pool would be equivalent to 30 years of anthropogenic emissions and could dramatically affect concentrations of atmospheric CO_2 (Kirschbaum, 2000). Alternatively, small increases in rates of oxidation of soil C as a result of increasing temperatures could result in further increases in atmospheric CO_2 (Davidson and Janssens, 2006).

In general, plant production and patterns of biomass allocation strongly influence relative distributions of C with soil depth (Jobbágy and Jackson, 2000). The deeper in the soil profile, the older stored SOC is likely to be. For example, Trumbore (2009) postulated that low-density C and microbial phospholipid acids would increase in age with soil depth. Fontaine et al. (2007) proposed an increase of mean residence times of SOC of up to 2000–10,000 years for depths beyond 20 cm. Increased mean residence times reflect reduced microbial activity and SOC turnover at depth. This conceptual model is supported by patterns of root biomass and relative root density that also decline with soil depth (Jobbágy and Jackson, 2000) and by increasing concentrations of organo-mineral complexes with depth (Fontaine et al., 2007).

2.1. Effects of climatic conditions and ecosystem conditions on SOC pools

2.1.1. Climatic conditions

Worldwide, SOC stocks generally increase as mean annual temperature decreases (Post et al., 1982). Cool/cold, humid climate regions are characterized by their C-rich soils (Hobbie et al., 2000); for example, approximately 1672 Gt of C is stored in the arctic and boreal ecosystems of the northern hemisphere—a large proportion of the world's soil C (Tarnocai et al., 2009).

Increased greenhouse gas concentrations in the atmosphere are set to accelerate the rate of warming that may in turn change net primary productivity (NPP), the type of organic matter inputs to soil and soil microbial activity are all important drivers of SOC fluxes. This may release additional CO_2 from some soils. Kirschbaum (2000) concluded that global warming is likely to reduce SOC by stimulating rates of decomposition whilst simultaneously increasing SOC through enhanced NPP resulting from increased CO_2 levels with the net change in SOC stocks expected to be small over the coming centuries. On the other hand, Sitch et al. (2008) propose that in some instances, soil might be a comparatively stronger source of CO_2 in the future as temperature rises.

2.1.2. Ecosystem conditions and land use change

Typically, arable soils contain around 1–3% of SOC, whilst grassland and forest soils usually contain more (Jenkins, 1988). Guo and Gifford (2002) highlighted the influence of land use changes on soil C stocks. Their meta-analysis of 74 publications suggested that land use changes from pasture to plantation (−10%), native forest to plantation (−13%), native forest to crop (−42%) and pasture to crop (−59%) reduced total C stocks whereas changes from native forest to pasture (+8%), crop to pasture (+19%), crop to plantation (+18%) and crop to secondary forest (+53%) increased total C stocks. A reasonable summary is that changing land use from cropland to pasture or cropland to permanent forest result in the greatest gains of SOC. Post and Kwon (2000), for example, estimated average rates of SOC accumulation for pasture and forest establishment to be of 0.33 t C ha^{-1} yr^{-1} and 0.34 t C ha^{-1} yr^{-1}, respectively. Conversion from nearly all other land uses to cropping or monocultures result in losses of SOC. The effects of tillage and diversity are clearly important research foci.

The quantity, quality and timing of organic matter inputs to soil vary with species composition within community types (i.e. relative abundance of N-fixing species) as well as with wholesale changes in community structure (i.e. cropland, grassland, shrubland, woodland, forest). Changes may also result from either management or natural variations in edaphic conditions at a local scale (e.g. Binkley and Menyailo, 2005; Hart et al., 2005). These interactions can be complex and vary through time. A recent study investigating long-term (1930–2010) dynamics of SOC after land use change in Java, Indonesia (Minasny et al., 2011), showed that after nearly 40 years of decline as a result of conversion of primary forest to plantation and cultivated land, total SOC increased, beginning around 1970. This switch, from decreasing to increasing most likely resulted from human interventions to increase plant production through fertilizer application and other management (i.e. the 'Green-Revolution') The total SOC in Java showed a net accumulation rate of 0.2–0.3 t C ha^{-1} yr^{-1} in the first 10 cm during the period of 1990–2000.

2.2. Environmental conditions and decomposition

SOM is derived from the microbial decomposition of plant inputs, either directly, as plant residues, or indirectly, as animal residues. During decomposition, the intact plant and animal residues are initially broken down into small particles of largely intact material. Eventually, following repeated recycling through the soil micro-organisms, as transient products of decomposition, original plant or animal residues give rise to a highly stable, black-brown substance referred to as humus. Globally, humus is calculated to have a mean turnover time of 27 years (Jenkins, 1988).

It is now generally considered that SOM decomposition is controlled more by biological and environmental conditions than by molecular structures of the carbon-based inputs (Schmidt et al., 2011). SOM is thus described as a continuum of materials in varying states of decomposition, with the chemical composition at any given site dependent upon the interplay of site conditions and biological limitations.

Litter provides the C that supports heterotrophic microbial activity, and the more readily decomposable the C, the more rapidly the microbial community can grow (Agren and Bosatta, 1996). Decomposition returns to the atmosphere most of the C added in litter to the soil surface—only a very small fraction becomes humus. Through their effects on microbial activity, moisture and temperature exert strong control over the rate of litter decomposition, followed by litter quality (decomposability) and soil microbial community composition (Meentemeyer, 1978; Melillo et al., 1982; Parton et al., 2007).

Moisture and temperature also exert strong control of humus decomposition. There is currently significant debate in the scientific community about the temperature sensitivity of different fractions (e.g. litter vs humus) and different pools (e.g. labile vs stable) of organic matter mineralization. At present a single relationship
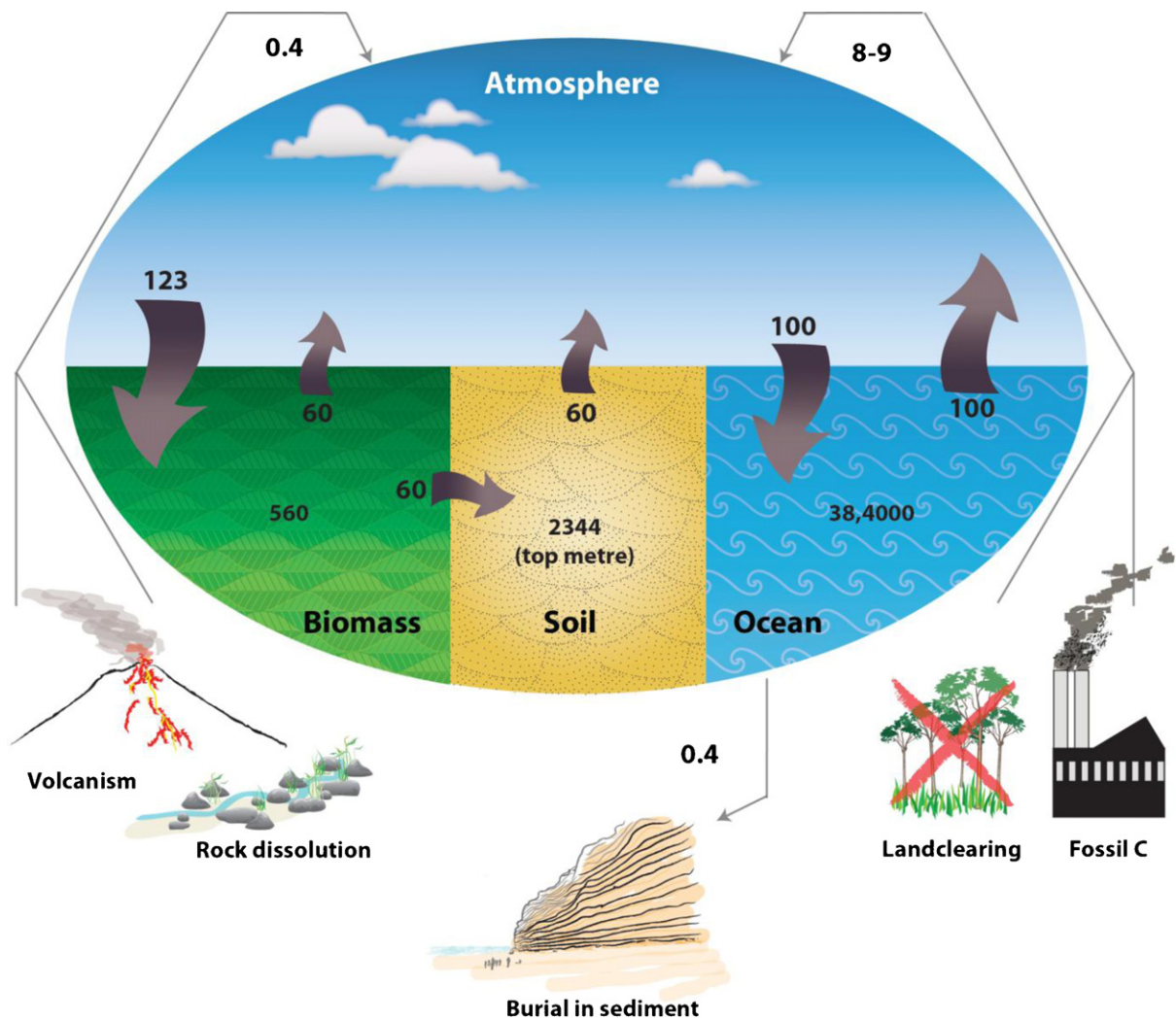
Fig. 1. Total annual flux of carbon in gigatons (billions of tonnes) through the most biologically active pools (as compared to the deep-ocean and lithosphere). Natural annual input of 'new' C to the biosphere through volcanism and rock dissolution is balanced by long term burial of C in ocean sediments. Annual anthropogenic input of 'new' C into the atmospheric pool (averaged over the last 40 years) comes from land clearing (1–2 Gt C annually) and fossil fuel sources (8.7 Gt C annually in 2008 (Le Quéré et al., 2009)). Mean uptake rates of terrestrial and ocean CO_2 sinks are 2.6 and 2.2 Gt C yr^{-1} for 1990–2000, respectively (Denman et al., 2007). If not cited otherwise, estimates adapted from Beer et al. (2010), Denman et al. (2007) and Volk (2008).

(i.e. a unique Q_{10}), is applied to all SOM pools in most decomposition models. On the other hand, chemical theory predicts that recalcitrant forms of organic matter should be more sensitive to temperature than labile forms. In the literature, results vary widely (Giardina and Ryan, 2000; Fang et al., 2005; Craine et al., 2010).

2.3. Biological processes affecting the decomposition of SOC

2.3.1. Priming effects

Priming effects were first recorded more than 50 years ago (Bingeman et al., 1953). In a review Kuzyakov et al. (2000) defined soil priming as "strong short-term changes in the turnover of SOM caused by comparatively moderate treatments of the soil". 'Priming effects' of added C (or N) on rates of mineralization of SOM are now well documented (e.g. Kuzyakov et al., 2000; Fontaine et al., 2003; Sayer et al., 2007), and may be either positive (i.e. increased C and N mineralization) or negative (i.e. immobilization). Addition of purified labile C substrates such as glucose (Dilly and Zyakun, 2008), fructose and oxalic acid (Hamer and Marschner, 2005), or cellulose (Fontaine et al., 2007) generally stimulate mineralization of SOM. Explanations of the priming effect include co-metabolism—additions of labile or fresh C stimulate growth of a suite of microorganisms that in turn leads to an increase

in microbial enzyme production (Kuzyakov et al., 2000; Hamer and Marschner, 2005). Fontaine et al. (2003) hypothesized that changes in land use and agricultural practices that increase the distribution of fresh C at depth could stimulate the mineralization of ancient buried C. For example, a change from shallow to deep-rooted grasses might have an overall negative effect on SOC due to the potential release of C buried at depth. Alternatively, might increased allocation of C below ground outweigh this potential effect? It remains to be seen if all soil microorganisms follow the same laws or if possible differences in inherent characteristics affect microbial diversity and SOC sequestration (Chabbi and Rumpel, 2009).

Quantifying priming effects under field conditions is challenging. Isotopic labelling is a preferred approach and well suited to small scale laboratory experiments, but difficult to apply in the field. In the laboratory, priming effects are generally of short duration and of small magnitude. Commonly, once the priming effect has dissipated, soil metabolism rapidly reverts to background rates. In systems where SOC accumulates most rapidly (i.e. some grasslands where a large proportion of NPP is belowground) priming effects may be of significance to total SOC. The true significance of priming effects to SOC on a global, or even ecosystem scale, awaits evaluation. It seems likely that priming effects will remain

Table 1
Estimates of SOC sink capacity of world soils, listed for principal biomes.

Biome	SOC storage (Gt C) by soil depth			Reference
	0–1 m	0–2 m	0–3 m	
Tropical regions	354–403	1078–1145		Batjes (1996)[a]
Other regions	616–640	1760–1816		
Global estimate	**1463–1548**	**2376–2456**		
Boreal forest	112	141	150	Jobbágy and Jackson (2000) [b]
Cropland	157	210	248	
Deserts	112	164	208	
Sclerophyllous scrubs	76	104	124	
Temperate deciduous forest	122	145	160	
Temperate evergreen forest	73	91	102	
Temperate grassland	105	143	172	
Tropical deciduous forest	119	175	218	
Tropical evergreen forests	316	408	474	
Tropical grassland/savannah	198	281	345	
Tundra	114	133	144	
Global estimate	**1502**	**1993**	**2344**	
Boreal forest	338			IGBP (Carter and Scholes, 2000)
Cropland	165			
Deserts and semi deserts	159			
Wetlands	N.A.			
Temperate forest	153			
Temperate grassland/shrubland	176			
Tropical forest	213			
Tropical grassland/savannah	247			
Tundra	115			
Global estimate	**1567**			
Boreal forest	471			WBGU (1988)
Cropland	128			
Deserts and semi deserts	191			
Wetlands	225			
Temperate forest	100			
Temperate grassland/shrubland	295			
Tropical forest	216			
Tropical grassland/savannah	264			
Tundra	121			
Global estimate	**2011**			

[a] Estimates based on 4353 soil profiles.
[b] Estimates based on 2700 soil profiles.

of greater significance to short-term effects – such as pulses of nutrient availability and of CO_2 and other greenhouse gases (GHG) to the atmosphere – than to long-term C sequestration.

2.3.2. Biodiversity

Increased plant diversity can increase SOC (e.g. Tilman et al., 2006) but this effect, while corroborated by some studies (see Section 2.1.2 above), is easier to demonstrate in artificial than 'natural' settings.

Many studies of litter decomposition contrast the influences of single plant species, and sometimes mixtures of leaf litter on decomposition (i.e. Melillo et al., 1982; O'Connell, 1986; Chapman et al., 1988; Briones and Ineson, 1996; Hobbie, 1996; McTiernan et al., 1997; Xiang and Bauhus, 2007).Much of the influence of individual plant species on soil microbial activity and nutrient cycling is through the quality and quantity of organic matter returned to the soil (Galicia and García-Oliva, 2004; Binkley and Menyailo, 2005). Plant species that host N-fixing bacteria (such as legumes) or root systems with mycorrhizal associations, often enhance nutrient uptake and can provide a pathway for the return of C substrate directly to microbes and soil (Hobbie, 1992). Increased microbial nitrogen (N) inputs, therefore, most likely increase soil C stocks by influencing decomposition processes. Studies by Kaye et al. (2000) and Resh et al. (2002) showed that compared with *Eucalyptus* species N-fixing trees increased soil N availability as well as the potential to sequester SOC. Moreover, N-fixing species tended to add more fresh C to the soil and that C was preferentially decomposed over older forms of C.

2.3.3. Roots and root exudates

Root-derived organic matter inputs are chemically diverse, ranging in complexity from readily decomposable substrates such as soluble sugars, amino and organic acids, to substrates from root turnover that require greater energy investment to decompose. It has been widely assumed that easily degradable plant exudates are almost exclusively degraded by bacteria (see reviews by Jones, 1998), with fungi playing important roles in the degradation of recalcitrant organic materials such as lignin, as well as cellulose and hemi-cellulose (reviewed by Boer et al., 2005). Perhaps owing to the difficulty (relative to inputs of leaf carbon to soil) of knowing exactly the amounts and chemical nature of root inputs, there is a wide range of views in the literature as to the significance of root inputs to soil carbon stocks and fluxes. Some research suggests root inputs to soil represent 5–33% of daily photoassimilate (Jones et al., 2009). Exudates fall into two categories: those that are a result of passive diffusion and over which the plant has little control, and those that have functional significance and at least some degree of regulation of their exudation by plants (see reviews by Jones et al., 2004; Paterson et al., 2007).

There are also suggestions, however, that root exudates contribute to depletion of SOC stocks through a 'rhizosphere priming effect' based on evidence that overall rates of SOC decomposition

may increase dramatically (up to 5-fold) in response to root exudates (see Sanderman et al., 2010 for examples).

Root-C seems to be preferentially stabilized compared to shoot-C (see review by Rasse et al., 2005). Greater chemical recalcitrance of root tissues as compared to that of shoots seems to be responsible for only a small portion of this preferential stabilization. Other SOM protection mechanisms might be enhanced by root activities, such as physico-chemical protection of root exudates, especially in deep soil horizons, and micrometer-scale physical protection through mycorrhiza and root-hair activities that place root C in very small pores and aggregates. This may have implications for C storage in as much as it might suggest; favouring rotations, cropping systems and plant species that allocate C below ground. Much more research is needed to determine and predict the balance between gain versus losses as precipitated by root growth and turnover.

2.4. Chemical and physical processes affecting the decomposition of SOC

SOC can be protected in the soil matrix through physical-(chemical) stabilization processes, as well as by inherent 'chemical recalcitrance'.

Decomposition has been described as a key or even a 'bottleneck' process in C and nutrient cycling (Attiwill and Adams, 1993). The chemical composition of decomposing material, especially characteristics such as C:N ratios and lignin content, are crucial for determining how quickly decomposition proceeds (Meentemeyer, 1978; Melillo et al., 1982). In general, litter decomposition rate is considered negatively related to C:N ratios, lignin content and lignin:N ratios, and positively related to N concentrations (Melillo et al., 1982, 1989). Even so, litter dynamics appear to differ from SOM dynamics. The mineralization of SOM proceeds at a much slower rate than the decomposition of the plant and animal residues from which it is formed. The latter, containing large polymeric molecules of biological origin such as proteins, carbohydrates, cellulose, etc. are highly favoured for enzymic attack, due to their relatively simple and regularly repeated chemical structures. In contrast, SOM lacks such a simple structure, and is a highly unfavourable substrate for enzymic mineralization (Kemmitt et al., 2008).

Recent stable isotope-based research, as synthesized by Amelung et al. (2008), has shown that the residence time of SOM is not correlated to its chemical composition. As a consequence, while relative chemical complexity helps explain short-term decomposition of litter and added organic matter decomposition (e.g. at seasonal, annual scales), it does not explain SOM decomposition in the long term (decades, centuries).

Mineralization is often directly linked to SOM via soil microbial communities and their molecular size, specific activity or composition (Marschner and Kalbitz, 2003; Fontaine and Bardot, 2005). A recent, somewhat 'controversial' hypothesis challenges this conventional view and proposes that the mineralization rate of humified SOM is independent of the size, structure or activity of the soil microbial community. Based on observations of fumigation experiments, Kemmitt et al. (2008) argued that the rate-limiting step in SOM mineralization is governed by abiotic rather than microbial processes termed the 'Regulatory Gate'. The 'Regulatory Gate' hypothesis processes include, among others: diffusion, desorption from soil surfaces, oxidation or stabilized extracellular enzymes. Further research is required to understand better the significance of abiotic versus microbial processes in SOM mineralization at the global scale. The regulatory gate hypothesis might suggest controls over C stocks depend on abiotic mechanisms of protection of C and our (human) capacity to influence those. Adsorption/desorption of SOM appears difficult to control in soils, but soil structure is strongly affected by soil management and land

use. Certainly, the 'Regulatory Gate' theory is not yet universally accepted (Kuzyakov et al., 2009).

Two main physico-chemical) stabilization processes are: (i) protection within aggregates which translates to spatial inaccessibility of soil microbes to organic compounds and a limitation on O_2 availability and, (ii) interactions with mineral surfaces and metal ions (e.g. see Six et al., 2004; Von Lützow et al., 2006 for details). These aspects are of importance because both can provide *a priori* limits to the soil C sequestration potential of some soils (via the surface properties of minerals such as functional groups and charge) and capacity for soil aggregation and stability (as influenced by soil particle size distribution). It must be noted that increasing SOM has positive effects on both, the aggregation of a soil and the amount of surface charge present. Ongoing discussions regarding the chemical and physical processes determining the composition of SOM were reviewed in detail by Kleber and Johnson (2010) and are summarized below:

2.4.1. Does the solubility of SOM, as determined by chemical (alkaline/acid) extractions, sufficiently characterize recalcitrant SOM?

Traditional chemical extraction procedures leave unextractable residues that are assumed to be resistant or recalcitrant because of their complex polymeric macromolecular structure. This observation led to the development of the 'humus concept' which postulates that decomposition processes create humic substances having different turnover times, i.e. fulvic acids (decades/centuries), humins and humic acids (millennia) (Schlesinger, 1977). However, such inherent chemical stability can be questioned since at least some extracted humic substances are a product of the extraction procedure rather than a real component of SOM (Piccolo, 2002). Quickly developing technologies such as nuclear magnetic resonance (NMR) and synchrotron-based, near-edge X-ray fine structure spectroscopy do not show clear evidence of discrete humic molecules in undisturbed soil (Lehmann et al., 2008b). Applying Curie point pyrolysis-gas chromatography coupled on-line to mass spectrometry (Py-GC/MS) and isotope ratio mass spectrometry (Py-GC IRMS), Gleixner et al. (1999) showed that instead of macromolecules, substances with lower molecular weight (proteins or peptides) are more likely to be preserved in soil during decomposition and humification processes. In addition, the assumed existence of a high proportion of aromatic C in humic materials has been questioned (Gleixner et al., 2002).

These new findings support microbial activity as the primary active agent for SOM stabilization, and that C integrated into new, microbe-derived molecules is what remains in the soil and not the precursor substances (see Chabbi and Rumpel, 2009).

2.4.2. Is distinction of SOM in terms of soil function and as conceptualized in process models of SOM dynamics, driven by knowledge of rates of decomposition and humification pathways?

Non-living SOM is now conventionally divided into at least three C pools: (1) an active pool with turnover rates of years (root exudates, rapidly decomposed components of fresh plant litter) (2) an intermediate or slow pool with turnover rates of decades and (3) a passive pool with turnover rates of centuries to millennia (stabilized organic matter due to chemical or physical mechanisms) (Trumbore, 1997, 2009), including charcoal formed through pyrolysis. Excluding charcoal, evidence for the nature of the passive pool relies mostly on radiocarbon dating and detection of C_3-plant residues in SOM of soils that have been cultivated with C_4-plants for centuries or longer. The magnitude of this passive or inert pool may be grossly underestimated, if charcoal is not quantified separately (Lehmann et al., 2008a).

Counter-evidence for the existence of a passive pool is that major organic materials (e.g. lignin, cellulose and hemicellulose,

lipids and proteins) can decompose fully under 'optimal' conditions (e.g. sufficient oxygen supply), even though some organic materials might take longer to decompose than others (i.e. wood > leaves). While environmental constraints help explain incomplete decomposition, there is evidence that interactions with mineral surfaces help stabilize decomposed plant residues and/or remnants of microbes and fungi.

2.4.3. Can the physical nature of SOM be better represented by a 'molecular aggregate' model as proposed by Wershaw (2004) in place of a 'humic polymer' model?

The principal basis of the molecular aggregate model is that SOM is composed of partially degraded products of the soil biota (i.e. plant polymers). Natural organic matter, therefore, consists of molecular aggregates (molecular fragments are amphiphilic in nature) of different degradation products that are held together by entropic interactions and/or noncovalent bonds. The model does not provide any structural reason for inherent stability against decomposition.

2.4.4. Taken together, these questions suggest a new view of decomposition processes and recalcitrance

Recently, the oft used concept of 'recalcitrance' in SOM research (which has been used as either a direct or proxy mechanism within SOM models) has been questioned as little more than iteration of a belief in inherent molecular 'resistance' to decomposition and as an operationally defined characteristic that reflects a long residence time for some SOC (Kleber, 2010). Thus, decomposition is perhaps more appropriately viewed as a logistical problem (Kleber, 2010) where the key factors are: (i) the microbial ecology; (ii) enzyme kinetics (iii) environmental drivers and (iv) matrix protection. Constraining SOM decomposition on a logistical basis has compatibility with systems approaches – that is: structure as defined by i–iv (above) is at least partially determinant of function and likely to include (potentially) comprehensible feedback mechanisms. These functions could include characteristics such as level of soil aggregation, nutrient cycling and, disease suppression.

It is currently unclear if a change of view as to the composition of SOM affects process-oriented models of SOM that were built using older concepts of humification and decomposition. It may be necessary to explicitly include concepts of stabilization as discussed in 2.4.2, 2.4.3 and 2.4.4 (e.g. as attempted in the Struc-C model by Malamoud et al. (2009)). The question remains as to whether this adjustment would improve the modelling of SOM dynamics.

3. Soil carbon measurement—Pools and fluxes

3.1. SOC pools

As described above, SOM consists of a complex mixture of (partially) decomposed substances (i.e. organic molecules such as polysaccharides, lignin, aliphatic biopolymers, tannins, lipids, proteins and aminosugars) derived from plant litter as well as faunal and microbial biomass (Totsche et al., 2010). It also represents a variety of pools that are related to microbial function (Krull et al., 2003; Trumbore, 2009). These pools are divided according to biological stability (labile, stabile, refractory and inert), decomposition rate (fast-active, slow-intermediate and very slow/passive/inert) and turnover time (short, long, very long). This 'division' both originated from and led to methods of SOC fractionation (Table 2).

3.1.1. Measurement of SOC

A considerable effort has been made by the scientific community to measure the amount of SOC using a variety of techniques including both *ex situ* as well as *in situ* methods (Table 3) (see Chatterjee et al., 2009 for an overview of these techniques).

Ex situ methods such as dry combustion have long been regarded as 'standard'. However, C sequestration programs using terrestrial systems will require precise and cost-effective measurement of SOC in order to verify and potentially monitor SOC stocks. Along with the development of efficient sampling methods at the farm- or landscape unit scale that allow definition of 'minimum' replication that in turn lower the costs involved with sample processing and measurement in the laboratory, *in situ* analytical methods have also been developed such as near-infrared spectroscopy (NIR) and mid-infrared spectroscopy (MIR). Combined, these sampling and analytical methods offer considerable cost-effectiveness compared to traditional methods (Janik et al., 2007).

Even so, modelling may require well-defined 'measureable' C 'fractions' in order to represent SOC dynamics. The following sections summarize the state of knowledge of these fractions.

3.1.2. SOC fractionation

The methods used to fractionate SOM including physical (size, density, aggregation) and chemical (solubility, mineralogy) separation procedures are shown in Table 4.

'Best practice' approaches have been developed that include both physical and chemical fractionation methods (e.g. see method proposed by Sohi et al., 2001; Six et al., 2002; Zimmermann et al., 2007a (Fig. 2)). These newer approaches are very promising even so they do not fully incorporate concepts such as stabilization on mineral surfaces or protection by aggregates. Additionally, cost-effective *in situ* methods are also being developed. For instance, MIR calibrated via partial-least-squares (PLS) has proven to be quite promising for the prediction of soil C fractions (total organic C, particulate organic C, resistant-(charcoal) C) (Janik et al., 2007; Zimmermann et al., 2007a; Calderón et al., 2011) and spectral libraries for predicting C fractions with MIR techniques are currently being developed.

3.1.3. Imaging SOC in situ with structure

Physical protection has been long cited as a key mechanism in stabilizing C (Adu and Oades, 1978) and much of what we know about the relation between soil structure and C dynamics is obtained using techniques that physically disrupt the soil (Young et al., 2008).

Transmission electron microscopy preparation techniques can be adapted to preserve the fabric of clay particles and the SOM in nano to micrometer aggregates (Chenu and Plate, 2006). Techniques exist for creating two-dimensional thin sections of soil that preserve microbes *in situ* and provide some information on the relative location of microbes and pores (Nunan et al., 2003). However, because of anisotropy and non-stationarity, two-dimensional images on their own provide an incomplete picture. This is especially true in relation to understanding transport processes relevant to resource flows where three-dimensional information on pore topology is required (Werth et al., 2010). Until recently, standard methods were not available for directly imaging the three-dimensional soil structure, C and microorganisms, or the true complexity of the soil habitat (O'Donnell et al., 2007).

X-ray microtomography has been exploited as a tool for imaging soil for several decades and is increasingly accessible to soil science (Peth et al., 2008). Microtomography provides non-destructive, high-resolution, three-dimensional images. The indirect effects of SOM on the three-dimensional features of the soil microbial habitat have been measured using tomography in a number of studies (Feeney et al., 2006; Deurer et al., 2009; Papadopoulos et al., 2009). In each case it has been observed that SOM has a profound impact on the physical structure of soil resulting in greater structural order and therefore increased soil porosity. Further significant advances

Table 2
Forms of (soil) organic C found in the scientific literature (table adapted from Chapter 2 of the Victorian Government report 2010, based on Baldock (2007)).

Form	Composition	Pool category
Surface plant residue	Plant material residing on the surface of the soil, including leaf litter and crop/pasture material	Fast (or labile) pool Decomposition occurs at a timescale of days to years
Buried plant residue	Plant material greater than 2 mm in size residing within the soil	Fast (or labile) pool Decomposition occurs at a timescale of days to years
Particulate organic matter (POC)	Semi-decomposed organic material smaller than 2 mm and greater than 50 μm in size	Fast (or labile) pool Decomposition occurs at a timescale of days to years
'Humus'	Well decomposed organic material smaller than 50 μm in size that is associated with soil particles	Slow (or stable) pool Decomposition occurs at a timescale of years to decades
Resistant organic carbon (ROC)	Charcoal or charred materials that results from the burning of organic matter (resistant to biological decomposition)	Passive (or recalcitrant) pool Decomposition occurs at a timescale of decades to thousands of years

Table 3
Techniques to determine the SOC content (after Chatterjee et al., 2009).

Ex situ methods	In situ methods
Wet combustion i.e. oxidization of SOC by an acid solution such as the Walkley-Black method	Infrared techniques such as MIR and NIR Laser-Induced Breakdown Spectroscopy (LIBS)
Dry combustion i.e. weight-loss-on-ignition such as the muffle furnace and automated CN(S) analysers such as vario EL or vario max	Inelastic Neutron Scattering Remote sensing

are needed to study the consequences of observed structure for soil function, including more sophisticated modelling approaches.

3.2. C fluxes and their measurement

Fluxes of C entering soils are directly linked to both extant soil C pools and soil C effluxes (SCE). Root exudates fuelled by photosynthesis and decomposition of aboveground and belowground litter provide C to soils. Along with decomposition of native SOM and root respiration these factors also exert strong control of soil respiration or SCE (Luyssaert et al., 2007).

Soil respiration is usually defined as CO_2 released from soil to the atmosphere via the combined activity of (1) roots (root or autotrophic respiration), and (2) micro- and macroorganisms decomposing litter and organic matter in soil (heterotrophic respiration) (Högberg et al., 2005). Root exudates are, in the main, rather quickly decomposed while leaf and root litter are less readily decomposed. SOM is then considered the least easily decomposed in part due to it representing materials that have already been

microbially processed (referred to as humus). Each component returns C to the atmosphere on different average time scales: rapid C cycling is often associated with root/rhizosphere respiration (days to months), while litter C decomposition is slower (months to years), while inherent (native) SOM decomposition (decades to centuries) seems the slowest of all (Kuzyakov and Larionova, 2005; Cisneros-Dozal et al., 2006). The relatively slow turnover rate of SOM via heterotrophic respiration has profound implications for long-term storage of SOC.

The stable isotope $\delta^{13}C$ is used in different ways to study the dynamics of soil C: Pulse-labelling is used to trace the flux of recently-fixed C through the plant–microbe–soil continuum (Högberg et al., 2008), while isotopic analysis of SOM from disturbed ecosystems that have experienced a shift in dominance from C_3 to C_4 plants (e.g. tropical forests to crops or grasslands) gives information on soil C cycling over the years or decades following land use changes (Balesdent et al., 1987; Ehleringer et al., 2000; Trumbore, 2000). The partitioning of heterotrophic and autotrophic sources of respiration using natural abundance $\delta^{13}C$ isotopes is very

Table 4
Methods of SOC fractionation (e.g. see review by Von Lützow et al., 2007).

| Physical fractionation				
Aggregate size fractionation	Particle size fractionation	Density fractionation		
Macroaggregates (>250 μm)/wet sieving/slaking/dispersion (ultrasonic)	Clay-sized, silt-sized and sand-sized particles	Heavy fraction (organo-mineral complexes)	Light fraction	POM (particulate organic matter)
		Separated using liquids with a certain density (between 1.6 and 2.0 g cm^{-3})		
DOC (dissolved organic matter)/less than 0.45 μm in solution				
Chemical fractionation				
Chemical extractions	Hydrolysis	Oxidation		
Used to separate 'humic substances' into humic acids, fulvic acids and humin/based on solubility in alkali and acid solutions, the most common are NaOH and $Na_4P_2O_7$ Microbial biomass C/i.e. chloroform used as fumigant	Used to separate hydrolytic bonding of carbohydrates/protein molecules, etc./for instance HF to separate mineral-OM associations/or HCl to quantify proportion of SOC association with proteins, amino acids, amino sugars	Used to remove labile/active fraction of SOM (i.e. plant residues)/most common agents are H_2O_2, NaOCl and $KMnO_4$		
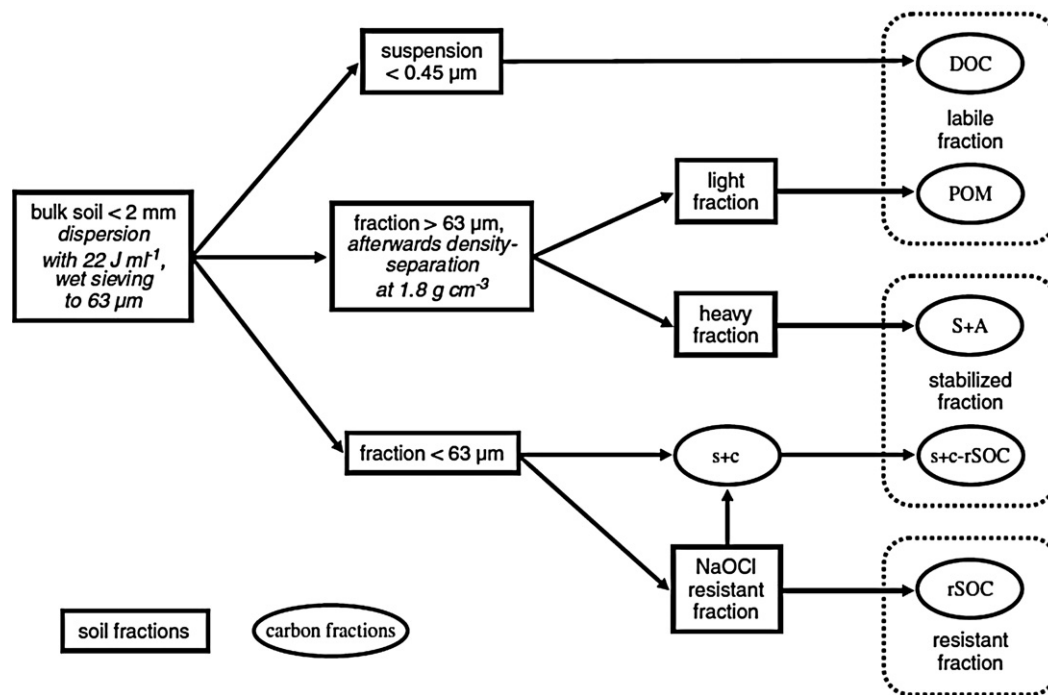
Fig. 2. Combined density fractionation procedure (taken from Zimmermann et al., 2007a, used with permission). Abbreviations: dissolved organic matter (DOC), particulate organic matter (POM), sand and stable aggregates (S + A), silt and clay particles (s + c) and oxidation-resistant carbon (rSOC).

promising (Millard et al., 2010). Radiocarbon ^{14}C is also used to estimate the origin of the C in both soil pools and in soil effluxes (Trumbore, 2000, 2009).

C balance approaches can also be used to estimate one or more pools or fluxes (e.g. Giardina and Ryan, 2002). Of these, soil respiration is usually monitored *in situ* using dynamic or static chamber methodologies coupled with an Infra Red Gas Analyser (IRGA) to measure CO_2 efflux at the soil surface (Luo and Zhou, 2006). Components of soil respiration can be further separated (Hanson et al., 2000; Högberg et al., 2005; Kuzyakov, 2006) using methods grouped into four broad categories: (1) integration of physically-disintegrated respiratory components (including root, leaf litter and SOM); (2) exclusion of live roots from soil monoliths via trenching; (3) the use of non-invasive stable or radioactive isotopes that rely on different 'isotopic signatures' of CO_2 derived from living roots and SOM; and, (4) stem girdling in forest ecosystems involving instantaneous termination of photosynthetic C flow to roots and associated microorganisms without affecting micro-climate, at least initially (Högberg et al., 2005).

4. Soil carbon modelling

4.1. Process-oriented versus organism-oriented models

4.1.1. Model characteristics

There is a need to assess SOM dynamics and the sequestration potential of soil C at the landscape scale (Post et al., 2007) as well as simulate the response of soils to environmental pedoturbation (Smith et al., 1998). As well as measuring SOM, models of SOM dynamics are used to address the needs listed above (Post et al., 2007; Batlle-Aguilar et al., 2010). Ultimately, they should provide reliable predictions to the size of soil C stocks for different soil types, with differing management practices (e.g. tillage practices, crop rotations, SOM additions or N fertilizer applications) and climate regimes. The main features of process-oriented and organism-oriented SOM models are briefly outlined in Table 5 (see

Brussaard, 1998; Smith et al., 1998; Post et al., 2007; Manzoni and Porporato, 2009; Batlle-Aguilar et al., 2010).

Organism-oriented models provide understanding of C and N flow through food webs and explore the role of soil biota in C and N mobilization (Fig. 3). They also allow analysis of environmental risks and provide a guide to above and below-ground linkage of food webs (Brussaard, 1998; Smith et al., 1998; Susilo et al., 2004).

However, process-oriented SOM models have been dominant in efforts to simulate changes in SOM resulting from management practices. The main characteristics of the most popular process-oriented SOC and N turnover models are presented in summary in Table 6. Process-oriented models predict SOM dynamics based on different conceptual C pools that alter in size via decomposition rates and stabilization mechanisms (Fig. 4). The majority of these incorporate the biomass of the soil in at least one model compartment. Noticeably, only a few, such as SOMM, include explicit descriptions of meso- and macrofauna functions and distinguish different forms of organic matter based on the abundance of the soil fauna (Smith et al., 1998). Within process-oriented models, CENTURY and RothC are the most frequently used to simulate SOM dynamics at a farm-scale (Viaud et al., 2010).

Process-based models have been developed and tested using long-run data sets. For instance, Smith et al. (1997) tested the performance of nine SOM models for predicting long-term changes in SOM using data from seven long-term (>20 years) experiments. In summary, most showed a good ability for predicting SOM dynamics over decades across a range of land use, soil types and climatic regions (Smith et al., 1998). Clearly, model calibrations play a major role in influencing their predictive ability.

That overall comparison also revealed that soil texture is used in only some models to modify decomposition processes (e.g. texture assigned physical protection). Recently, Malamoud et al. (2009) developed the Struc-C model (based on RothC) that specifically incorporates soil structure (aggregate) hierarchies and physical protection of SOC via aggregates. In the Struc-C model, clay content plays an important role in complexing C and only this complexed
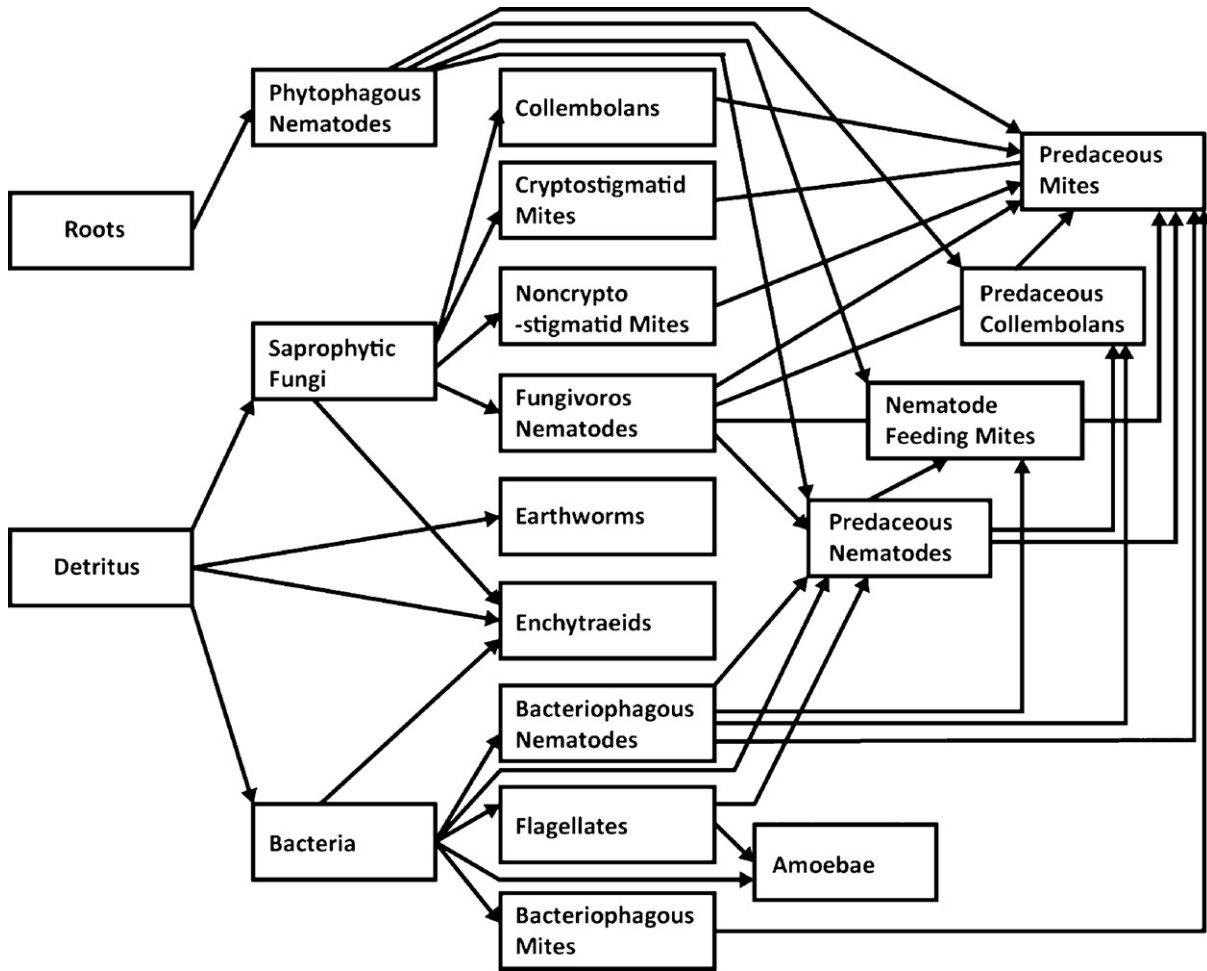
Fig. 3. Structure of an organism-oriented model (detrital food-web model, the model exemplifies the food-web at Lovinkhoeve experimental farm, The Netherlands based on De Ruiter et al. (1993), cited in Smith et al. (1998), used with permission).
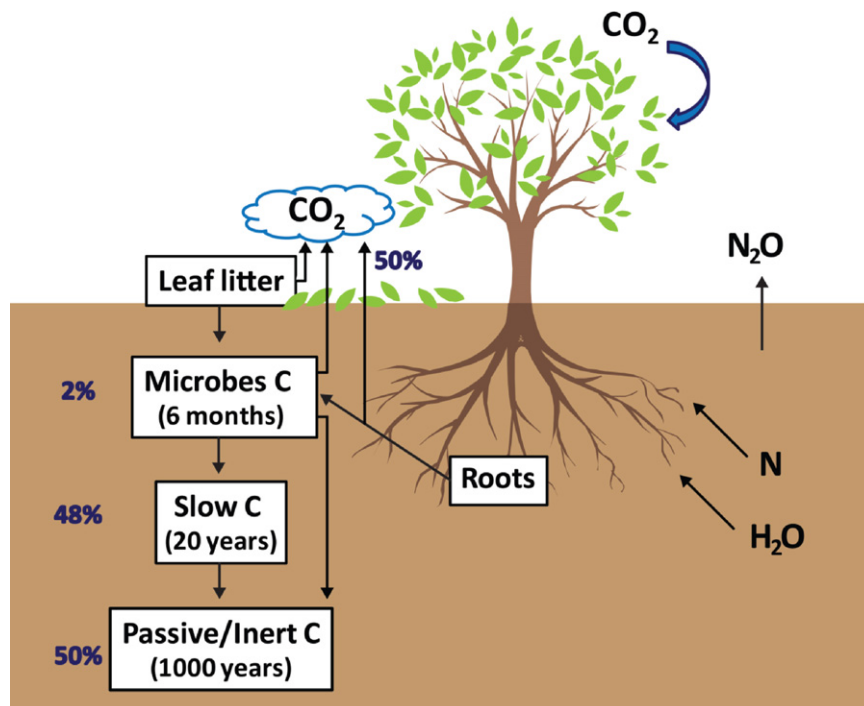

Fig. 4. Conceptual pools and fluxes often used in process-oriented SOM models.

Table 5
Main characteristics of processes-oriented versus organism-oriented models (based on Batle-Aguilar et al., 2010; Brussaard, 1998; Krull et al., 2003; Manzoni and Porporato, 2009; Post et al., 2007; Smith et al., 1998).

	Process-oriented models	Organism-oriented models
Model type	Mechanistic, predictive value	Mechanistic, explanatory value
Aim	Simulate processes involved in SOM migration and transformation	Simulate SOM using functional/taxonomic groups of the soil
Examples	CANDY, CENTURY, DAYCENT, DAISY, DNDC, ITE, NCSOIL, RothC, Socrates, SOILN, SOMM, Struc-C and the Verbene model	Fungal-growth models/Models of decomposition of OM that incorporate functional groups of microbial biomass/Food web models based on taxonomic groups (mostly detrital models)
Representation of SOM	Different conceptual C pools with similar chemical or physical characteristics/differ by decomposition rates, stabilization mechanisms/generally soil biota only included in form of microbial biomass (exception: SOMM) Generally, more than one compartment of SOM degradation: (a) Active pool (fresh plant material, root exudates, microbial biomass) with MRT of 1 year (b) Slow pool (SOC that decomposes at intermediate rate) with MRT of 100 years (c) Passive or inert pool (SOC with physical or chemical stability) with MRT of 1000 years	SOC dynamics represented through different pools of soil biota (classified according to their taxonomy or metabolism) i.e. representation of soil biota by functional groups (food web models): Microorganisms (bacteria, mycorrhizal and saprotrophic fungi) SOM and litter (represented in form of roots, detritus)
Mechanism	SOC decomposition based on first-order kinetic rates	C and N fluxes simulated through functional groups based on their specific death rates and consumption rates, applying energy conversion efficiencies and C:N ratios of the organisms
Time-step	Daily, weekly or monthly	Daily
Scale	Include top 30 cm of the soil Small-plot to regional-scale	Small-plot
Application	Have been applied to a range of ecosystems (grassland, arable land, grass-arable rotations, forest)	Have been applied to arable land and grassland
Others	Successfully coupled with GIS software (e.g. CANDY, CENTURY, RothC)	Include changes of soil biota communities in the modelling of SOM dynamics (i.e. simulating feedback mechanisms due to changes in biota activity or characteristics)

C enters the aggregate pool. Turnover of aggregates has predefined rate constants associated with disruption and aggregation that are modified by the time since the last input of fresh C. Disruption of aggregates is associated with a loss of a fraction of C as CO_2 that depends on clay content. Struc-C was tested using data sets from England and Australia. The model yielded promising results compared with RothC. Struc-C, however, is not spatially explicit and does not include soil biota and thus is limited in its ability to probe the link between C sequestration and the biological and physical functioning of soil.

Most models do not differentiate SOM content with depth and do not account for the influence of even simple factors like soil pH or management factors such as tillage which influence crop residue and SOC distribution with depth (Angers and Eriksen-Hamel, 2008). Most only include the top 30 cm of soil, the zone that inhabits the majority of roots, plant inputs and microbial activity. It may well be worthwhile to simulate the degree to which soil C below the 'plough layer' may be subjected to disturbances (Trumbore, 2009). Jenkinson and Coleman (2008) modified RothC to RothPC-1 to predict the turnover of organic C in subsoils to up to 1 m of depth using multiple layers and introduced two additional parameters, one that transports organic C down the soil profile by an advective process, and one which reduces decomposition processes of SOM with depth.

There have also been important changes to the CENTURY series models of SOM dynamics (DayCent, ForCent and PhotoCent). One major improvement has been the ability to simulate full greenhouse gas dynamics for agricultural, forest and grassland systems. For example, DayCent can simulate soil CH_4, N_2, N_2O and NO_x gas fluxes, plant production dynamics, soil N dynamics, soil NO_3 leaching and SOM dynamics using daily time steps. DayCent is

also considered to be sufficiently well calibrated to project future changes in GHG dynamics (Del Grosso et al., 2009; De Gryze et al., 2010).

Clearly, calibration with long-term data-sets is critical. Gollany et al. (2011) used the data of five long-term experiments (>50 yr) and the processes-based model CQESTR to assess the potential of maintaining SOC stocks by applying no-till practices whilst removing crop residues. Their results showed that sustainable cropping depended on initial SOC, crop rotation intensity, tillage practices, crop yield and environmental conditions. Their simulations also showed that manure application, cover crops or intensified crop rotation under no-tillage (NT) practices could overcome loss of crop residue C. In summary, process-based models such as CQUESTR have been used successfully to simulate long-term effects of management changes from tillage to NT practices under a range of environmental and soil properties, fertility management, cropping systems and crop residue removal practices.

4.1.2. Model limitations

The current preference for process-oriented models over organism-oriented models as predictive tools for policy makers and other stakeholders (Smith et al., 1998) stems from (1) the easier estimation of internal parameters and calibration for specific purposes, (2) their testing with a variety of ecosystems, (3) their relatively simple structure and often larger time-steps (i.e. months versus days for organism-oriented models), (4) their ease of coupling to GIS software, (5) their suitability for larger scales (landscape or regional), and (6) their typically lower computational intensity (Brussaard, 1998; Smith et al., 1998; Post et al., 2007; Batle-Aguilar et al., 2010).

Table 6
Main characteristics of the most frequent referred to models in the scientific literature (based on Smith et al., 1997; Grace et al., 2006; Krull et al., 2003).

Model	Main characteristics	Reference
CANDY	Modular system combined with data base system for model parameters, measurement values, initial values, weather data, soil management data Simulates soil N, temperature and water to predict N uptake, leaching, water quality Uses proportion of soil particles to separate IOM (<6 µm)	Franko (1996)
CENTURY	Designed for long-term (up to centuries) SOM dynamics, plant growth and N, P and S cycling Developed for grassland, but extended to cropping systems, forests and savanna systems Monthly time step Implements two forms of litter: metabolic and structural Implements three SOM compartments: active (MRT 1–5 yr), slow (MRT 25 yr, 30–60% of SOC) and passive (MRT 1000 yr, 30–50% of SOC) Soil texture (clay content) determines separation of C from active OM pool into CO_2 or slow pool Basic ideas similar to RothC Biomass included	Parton (1996)
DAISY	Simulates crop production and soil water and N dynamics Developed as field management tool for agricultural systems Portioned into hydrological model, soil N model with a SOM submodel and a crop model with a N uptake model Clay content influences rate constants Semi-cohort accounting system used for litter decay Biomass included	Mueller et al. (1996)
DNDC	Couples denitrification and decomposition processes 4 submodels: soil climate, decomposition, denitrification, plant growth Clay absorption of humads Biomass included	Li et al. (1992)
ITE	Developed for grassland environments Aims to simulate N cycling 3 submodels: grazing-animal intake model, vegetative grass-growth model, SOM model Decomposition rates are function of quantity of microbial biomass	Thornley and Verbene (1989)
NCSOIL	Simulates N and C through soil microbes and organic components 4 organic compartments: plant residues, microbial biomass, humads, stable organic matter (stability of SOM results from metabolism) Decomposition independent of microbial biomass Biomass included	Molina (1996)
RothC	Developed for arable land, but also applied to temperate grasslands and forest soils Monthly time step 5 compartments: decomposable plants, resistant plant material, microbial biomass, humified organic matter (MRT 50 yr 80–90% of SOC), inert organic matter (MRT up to 10,000 yr, 5–15% of SOC) Decomposition rate, and ratio of humus, microbial biomass and CO_2 dependent on soil clay content Basic ideas similar to CENTURY Biomass included	Jenkinson and Coleman (1994)
Socrates	Weekly time step 5 compartments: decomposable plant material, resistant plant material, unprotected microbial biomass, protected microbial biomass, humus (stabilised pool) Decomposition rate (into humus, microbial materials and CO_2) dependent on soil CEC Biomass included	Grace et al. (2006)
SOMM	Developed for forest systems Process rates regulated by N and ash content of litter fall 3 soil litter layers: L, F, H Soil animals influence C fluxes (i.e. distinction into forms of humus such as mull and mor based on role of soil fauna—microarthropods and earthworms) Models C accumulation in soil organic horizons	Chertov and Komarov (1996)
Struc-C	Updated, modified version of the Roth-C model Monthly time step Incorporates soil structure (aggregate) hierarchies within physical protection of SOC Simulates formation of organo-mineral associations and aggregates (physically protected SOC)	Malamoud et al. (2009)
Verbene	Developed for grasslands Implements a plant growth submodel 3 Submodels: Soil Water, Som (Plant Residues: Decomposable, Structural, Resistant, Om: Stabilized, Protected, Unprotected), Soil N Physical protection caused by soil clay Decomposition rate modified by temperature and soil moisture, not influenced by microbial activity Biomass included	Verbene et al. (1990)

4.1.2.1. Process-oriented models.

(a) *Measuring the modelable versus modelling the measureable.*Assignment of model compartments to measureable SOM fractions is often difficult and may require isolation of numerous operational fractions which in turn restricts the validation of these models with real world data (Krull et al., 2003; Post et al., 2007). There can also be discordance between conceptual residence times/stabilization mechanisms and methodologies that can selectively fractionate these particular C pools (Von Lützow et al., 2007; Chabbi and Rumpel, 2009). For example, Zimmermann et al. (2007b) compared measureable SOM fractions to the conceptual C pools used in the RothC model and found good agreement for a range of environmental conditions (i.e. shift from management practices such as till to NT farming). This is not always the case and it may be worthwhile to modify SOC pools used in models so that they are based on measureable C fractions, and globally consistent data sets on measureable soil C fractions are needed to calibrate and validate SOC models.

(b) *The representation of soil biota.*Post et al. (2007) noted that there was no scientific evidence that abundance of soil biota limited decomposition. Wutzler and Reichstein (2008) have since proposed a simple model with a more explicit description of soil biota functioning to investigate the representation of priming effects, SOC accumulation limits and inputs of fresh organic matter. Their results indicated that priming effects did not have a major influence on overall SOC dynamics and supported thinking that C was likely to be stabilized in deeper layers of the soil profile in the absence of fresh, energy-rich C supply. Furthermore, Lawrence et al. (2009) investigated whether addition of microbial mechanisms of decomposition would improve models of SOM dynamics. They showed that the inclusion of exoenzyme and microbial controls in kinetic representation of decomposition rates improved the ability of some models to simulate changes in soil C stocks under different moisture conditions.

4.1.2.2. Organism-oriented models. Limitations of organism-oriented (food web) models for predictive purposes have been listed by Brussaard (1998) as:

(a) The quality of organic matter consumed at each trophic interaction is not well known;
(b) A number of functional groups are not included in the existing models;
(c) The possible spatial habitat restriction of certain functional groups is not incorporated and;
(d) Many biological interactions in the soil are actually non-trophic in nature.

In summary, at the present time validation of food web models with real world data is both cost intensive and unlikely due to highly challenging estimations of intensive parameters such as feeding preferences, N content, life cycles, assimilation efficiencies, production:assimilation ratios, decomposabilities and population sizes (Smith et al., 1998). Furthermore, model uncertainties are likely to be very large, reducing their overall effectiveness.

In summary, for the foreseeable future, process-oriented models are most likely to be relied upon for predicting long-term SOM dynamics at the farm scale including management practices. Organism-oriented models will be useful for providing advice on specific management practices such as 'how does change in soil biota affect the overall functioning of the agroecosystem?' or 'what options do farmers have for managing populations of organisms?' (see Susilo et al., 2004).

4.2. Other models used and currently being developed to potentially predict SOM stocks or fluxes

4.2.1. Empirical regression models
Annual changes in C stocks have been estimated from empirical regression 'models' based on long-term field experiments or derived from the literature (Viaud et al., 2010). Estimation of soil C stocks for whole regions can be cost-effectively based on stratified, random sampling provided there is sufficient data for a defined set of environmental covariates (McKenzie et al., 2000). This approach can be used to monitor changes in C stocks (after the adoption of management practices for at least 5 years) under a C trading scheme.

4.2.2. Landscape models
Moving towards a perception of farms as whole ecosystems (Lemke and Janzen, 2007), and therefore the importance of landscape-scale modelling of SOM dynamics is widely recognized (e.g. Voinov et al., 2004; Gaucherel et al., 2006; Viaud et al., 2010). Three gaps in knowledge have been identified: (1) the development of optimal, but still simple, 3-dimensional representations of landscapes (vertically and horizontally), (2) the implementation of functional interactions and SOM transfers (such as erosion, transport of dissolved organic matter and effects related to different soil depths) and (3) the availability of adequate datasets for model validations (especially the representation of fluxes between different landscape elements) (Viaud et al., 2010).

For example, the fields of hydrology and geomorphology have developed soil C models that operate at a landscape scale. Porporato et al. (2003) incorporated soil C and N cycles in a hydrological model to study the influence of soil moisture dynamics on soil C and N dynamics whereas Yoo et al. (2006) found that soil thickness controlled SOC storage within grass-covered hillslopes, with 70% of SOC stored in depositional areas.

4.2.3. 'Whole systems' modelling
Modelling the functioning of whole ecosystems, incorporating all interactions of soil processes and soil-biota has many logistical challenges. Primary among these is the whole-system comprehension required to build such a complex network, closely followed by the level of spatial and temporal knowledge needed to populate local interactions within the network. Another challenge is the lack of knowledge of the behaviour of whole networks, including direction change, 'metastable' states and non-linear behaviour (chaos, i.e. see paper by Phillips, 1998). In complex systems causation can become iterative; that is its effects in each 'instant' are somewhat fluid and 'effects' in and of themselves are also considered causative agents. All of the above makes modelling of SOM from a systems point of view (at practical scales) well beyond current computing power even if sufficient comprehension and modelling constructs existed. At this point, such applications remain theoretical rather than predictive.

4.2.4. Prospects for potential 'Next Generation' soil carbon models
There remain important gaps in our knowledge about the processes of C stabilization in soil (Lal, 2008a,b). Furthermore, the understanding of the link between how C is sequestered in soil and its consequences for soil function is also appearing more limited in light of a growing list of 'known unknowns'. However, understanding soil function is crucial to sustaining and enhancing productivity to meet future food demands. Current global trends in degradation of agricultural soils have to be reversed. It is important to understand the extent to which the processes of C stabilization are related

Table 7
Potential rates of C sequestration by adoption of best management practices for principal biomes (adopted from Lal, 2008b).

Best management practices	Temperate climate		Tropical climate	
	Humid (t C ha^{-1} yr^{-1})	Dry (t C ha^{-1} yr^{-1})	Humid (t C ha^{-1} yr^{-1})	Dry (t C ha^{-1} yr^{-1})
Conservation tillage	0.5–1.0	0.25–0.5	0.25–0.5	0.1–0.25
Cover cropping systems	0.2–0.5	0.1–0.2	0.1–0.2	0.05–0.1
Integrated nutrient management/manuring	0.3–0.6	0.2–0.4	0.2–0.4	0.1–0.2
Improved grazing	0.02–0.4	0.1–0.2	0.4–0.8	0.1–0.2

to biotic and abiotic components of soil over time. A theory of soil C dynamics that links physical and biological processes through time must be developed.

Some tentative steps towards such an approach have been taken by Malamoud et al. (2009). Before that, Tisdall and Oades (1982) had proposed a conceptual model coupling aggregation and biotic process that expresses the importance of microbial activity and roots in soil aggregation. Whilst the mechanisms are generally accepted, this conceptual model has never been operationalized. Thus, such a model could be useful in addressing the link between C and important soil functions. Accordingly, Dupraz et al. (2006) modelled the interaction between sedimenting particles and photosynthesizing cyanobacteria in a study of the development of growth pattern in stromatolites.

Crawford et al. (1997) also presented a model for particle aggregation in soil based loosely on the Tisdall and Oades conceptual model. While Crawford et al. (1997) proposed only a two-dimensional model with no feedback between structure and microbial activity, a more sophisticated three-dimensional model is currently being developed. Models such as this can identify new kinds of dynamical behaviour in the soil–microbe system and *stimulate* thinking. Whether or not the level of complexity currently precludes *simulation* of real soil, such a model clearly could not be used to predict C dynamics at the landscape scale.

4.3. Where to from here in modelling of SOM dynamics?

Validation of existing models with real world data has shown convincingly that recent advances in SOC models (i.e. RothC and CENTURY) facilitate prediction of SOC stocks for a variety of natural and agricultural systems. Models are also becoming increasingly more powerful in predicting soil biological activity. Multiple model

Table 8
Measured rates of C sequestration by adoption of best management practices for principal biomes.

Country	Agricultural system/Management practice or change		SOC sequestration rate	Reference
Australia	Long fallow wheat-sorghum rotation	No-till	No change	Young et al. (2009)[a]
	Continuous winter cereal	No-till	No change	
	Intensive cropping (>1 crop/year)	No-till	Some accumulation of SOC observed	
	Perennial pasture		~0.35 t C ha^{-1} yr^{-1}	
Australia	Crop rotation		0.20 ± 0.04 t C ha^{-1} yr^{-1}	Sanderman et al. (2010)[b]
	Stubble retention		0.19 ± 0.08 t C ha^{-1} yr^{-1}	
	Reduced tillage		0.34 ± 0.06 t C ha^{-1} yr^{-1}	
	Cropping to pasture		0.30–0.60 t C ha^{-1} yr^{-1}	
Brazil	Crop rotations with legumes	No-till	0.04–0.88 t C ha^{-1} yr^{-1} in the top 30 cm	Boddey et al. (2010)[c]
			0.48–1.53 t C ha^{-1} yr^{-1} in the top 100 cm	
Canada	Intensive-till to no-till		0–0.16 t C ha^{-1} yr^{-1}	VandenBygaart et al. (2008)[d]
	Wheat-fallow to continuous cropping		0.20–0.30 t C ha^{-1} yr^{-1}	
	Annual to perennials		0.45–0.77 t C ha^{-1} yr^{-1}	
Denmark	Silage maize		0.25–0.49 t C ha^{-1} yr^{-1}	Kristiansen et al. (2005)
	Silage maize and manure at 8 t ha^{-1}		0.71–0.98 t C ha^{-1} yr^{-1}	
South Africa	No-till		No change	Loke et al. (2012)[e]
Sweden	Management effects on topsoil SOC		Highly variable	Karlsson et al. (2003)
Global	No-till compared to conventional tillage		~0.57 ± 0.14 t C ha^{-1} yr^{-1}	West and Post (2002)[f]
	Enhancing crop rotation complexity		~0.20 ± 0.12 t C ha^{-1} yr^{-1}	
Global	No-till compared to conventional tillage		4.9 t ha^{-1} more SOC under NT in the top 30 cm	Angers and Eriksen-Hamel (2008)[g]
Global	No-till compared to conventional tillage		No significant change in total SOC Change of distribution of SOC in soil profile e.g. SOC increase by 3.15 ± 2.42 t ha^{-1} in top 10 cm	Luo et al. (2010a)[h]
Global	No-till compared to conventional tillage		3.4 t ha^{-1} more SOC under NT in the top 30 cm	Virto et al. (2012)[i]

[a] Replicated field experiments from 1994 to 2000.
[b] Review of Australian agroecosystems.
[c] Three long-term experiments on Ferrosols (subtropical climate).
[d] Review of Canadian agroecosystems.
[e] Semi-arid climate.
[f] Meta-analysis of 67 long-term tillage/crop rotation studies.
[g] Meta-analysis of long-term tillage studies, 24 comparisons.
[h] Meta-analysis of 69 paired-experiments.
[i] Meta-analysis of long-term tillage studies, 35 comparisons, some variability (up to 30%) in response to NT attributed to differences in yield and C inputs.

development, refinement and assessment are needed to maintain the scientific momentum behind the modelling of SOM dynamics.

Given current research on SOM dynamics within different disciplines, the following seems a reasonable list of empirical outcomes that we should be able to better predict:

- Soil structure and SOC interactions;
- The composition of SOC which is still under debate;
- Potential changes in the composition of SOM due to different environmental (e.g. climatic) and management regimes;
- The distribution of SOM in the soil matrix and the consequences for biological, chemical and physical processes over time;
- The amount and dynamics of dissolved SOM;
- The behaviour of SOM dynamics with depth which will lead to the modelling of potential changes in SOC dynamics based on the whole soil profile (e.g. predicting SOC dynamics beyond the surface layer);
- The role of surface versus belowground biomass residues in SOM dynamics;
- Enhanced or suppressed biological processes;
- The amount of respired C versus C transferred to soil during decomposition;
- Potential changes in SOC dynamics on the landscape scale including:
    SOC stocks for a set of different management practices;
    The redistribution of SOC to different parts in the landscape due to erosion and deposition (buried soil C).

5. Practical measures for enhancing soil carbon

SOC has received increasing attention due to its potential capacity to play an important role in mitigating (human) GHG emissions (Wander and Nissen, 2004). At a global scale, this is due to:

- The large size of the soil C pool compared to other biologically 'active' pools (Paustian et al., 2000);
- The estimated loss of organic C from the soil pool due to anthropogenic influences over the last century. For instance, globally around 42–78 Gt of C have been lost due to soil management practices and soil erosion (Paustian et al., 2000; Lal, 2004b);
- The considerable annual flux of C through the soil pool relative to annual anthropogenic emissions; and
- The total area of soil that is under some form of human management.

Rhetorically, efforts to increase the SOC pool have been dubbed *soil carbon sequestration* (or SCS, Paustian et al., 2000). According to Bernoux et al. (2006) "SCS for a specific agroecosystem, in comparison with a reference, should be considered as the result for a given period of time and portion of space of the net balance of all GHG expressed in C-CO$_2$ equivalent or CO$_2$ equivalent computing all emissions sources at the soil–plant–atmosphere interface, but also all the indirect fluxes (gasoline, enteric emissions and so on)." Thus, SCS is described as long-term or permanent (i.e. 100 years) removal of CO$_2$ from the atmosphere into the soil. SCS may have the potential to offset fossil-fuel emissions by 0.4–1.2 Gt C yr^{-1}, equal to 5–15% of the global emissions (Lal, 2004a; Powlson et al., 2010). The term SCS, therefore, implies a 'lock up' of C from atmospheric CO$_2$ in the soil either through accumulation of very long lived C (e.g. charcoal) or by alteration of the relative magnitude of soil C pools with different residence times. Not all increases in SOC are a true net transfer of C from the atmosphere to land. Some are simply a movement from one terrestrial C pool to another, and have no influence, either positive or negative, on climate change (e.g. see discussions

on manure application in Janzen et al., 1998; Schlesinger, 1999; Viaud et al., 2010, or organic farming in Leifeld and Fuhrer, 2010).

The benefits of SOM for soil health, plant growth and production are well known and are related to its fundamental role in the function and fertility of terrestrial ecosystems (Janzen, 2006). Sparling et al. (2006) recently proposed that management to improve SOC could have an environmental protection benefit of up to 40–70 times its benefit to productivity (yield). This is an example of the win-win of SCS (e.g. see Chapter 8.8 IPCC report, Smith et al., 2007). Similarly, Post and Kwon (2000) reported losses of up to 50% of SOC in the top 20 cm of arable soil, over 30–50 years of cultivation. Simple switching to NT farming practices thus can offer enormous benefits. SCS will, however, depend on a variety of factors such as existing soil C, soil type, climate and management practices (Smith et al., 2007). Furthermore, the feasibility of enhancing SCS will not only depend on these natural factors but also on the social and economic conditions in place such as labour cost and sufficient food production to name a few. There is some theoretical basis to suggest that the size of the potential SOC sink should be equal to the amount of C lost due to past managements. This is probably an overly optimistic estimate and Lal (2004a) suggested that the C sink capacity of agricultural and degraded soils might be only 50–66% of the historic C loss, due to the need to account for irreplaceable losses of mineral soil mass through processes such as soil erosion. Even so, some of the C lost via erosion may end up 'buried' via terrestrial sedimentation, up to 1 Gt C yr^{-1} (see Stallard, 1998; Quinton et al., 2010).

The latest report of the IPPC (Smith et al., 2007) implies an annual sequestration potential of 1.4–2.9 Gt of CO$_2$-equivalents through global agricultural soils, where soils would reach C saturation after 50–100 years. In contrast, a recent publication by Chatterjee and Lal (2009) suggests a sequestration potential of agricultural soils of up to 6 Gt of CO$_2$-equivalents per year by 2030. In this regard, Table 7 summarizes potential rates of SOC sequestration by adoption of best management practices for principal biomes whereas Table 8 compiles actual measured rates of SOC sequestration. For instance, most meta-data analysis (Table 8) suggest that if NT farming is adopted, there is a slight overall increase in SOC in the surface soil compared to full-inversion-tillage (FIT) and that this increase improves with time (Angers and Eriksen-Hamel, 2008; Luo et al., 2010a; Virto et al., 2012). However, when considering the whole soil profile, there seems to be a limited effect of NT on SOC stocks (Luo et al., 2010a). Virto et al. (2012) found that some of the variability (up to 30%) in response to NT can be attributed to differences in yield and C inputs. As seen in Table 8 there are some case studies where NT does not increase SOC (e.g. Loke et al., 2012) or where NT results in SOC increase at very great depth (Boddey et al., 2010).

SCS cannot continue indefinitely at rates like those listed above. Initial increases in SOC following a change in management are rapid, but then slow and will reach new quasi-equilibrium at some point in future.

SCS can also be reversed, as shown by history. Reducing C inputs and increasing tillage will remain threats to SOC and rates of loss can quickly increase under some scenarios. For example, Aune and Lal (1997) proposed a critical SOC concentration of 1.08% for corn yield production in tropical soils (low activity clays), based on crop yield being about 80% of total yield. A critical value of 2% SOC was proposed by Loveland and Webb (2003), below which a serious decline in soil quality would occur in agricultural soils in temperate regions. While there is a lack of quantitative evidence for such thresholds, the review of cropping data by Loveland and Webb (2003) suggests there is reasonable evidence of a threshold for N-supply from plant residues that corresponds to approximately 1% SOC. The review also drew attention to the possibility that the proportion of fresh C (biomass input) in the total SOC pool may be
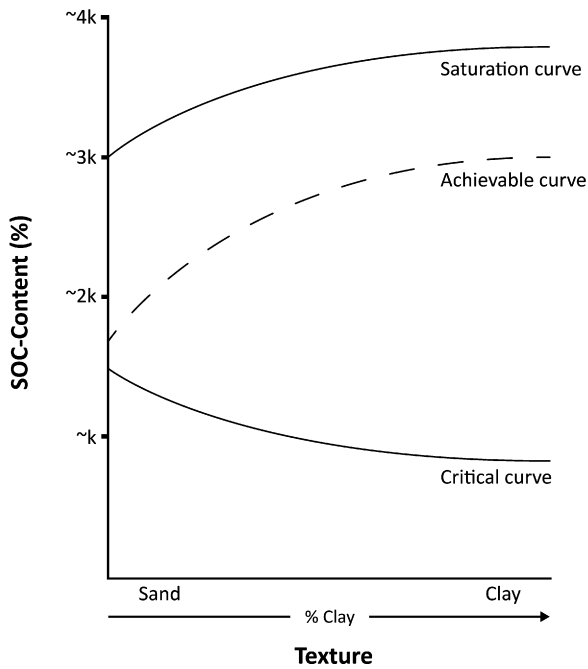
Fig. 5. Notional saturation, achievable and critical curves of SOC accumulation as a function of soil texture. Here, k is a site- or region-specific proportionality constant' that depends on a variety of interacting factors (e.g. climate, soil material and topographic position).

of greater importance to secure soil quality than the size of the C stock itself. On the basis of the literature (e.g. see Hassink, 1997), it is posited that critical concentrations of SOC (e.g. ~1.5 k for sandy soils, and ~0.8 k for clay rich soils) can be set as guide for farmers. Graphically (Fig. 5) it might be reasonable to propose that,

depending on clay types and content, there is a potential *saturation curve*, *achievable curve* and *critical curve* for the amount of organic C in agricultural soils. However, to establish such standardized curves across a range of soil types, more quantitative studies are needed.

6. Conclusions

The 'future of SOC research' requires collaboration and communication between the 'science community' (Fig. 6a) and the 'practice sector' (Fig. 6c), facilitated by individuals that are knowledge brokers (Fig. 6b) as defined by Bouma et al. (2011) with "hard knowledge and social intelligence". In order to be applicable to the practice sector, new findings in SOC dynamics need to be addressed in a conceptual framework for communicating the need for change as being 'compatible', 'observable', 'achievable' and foremost 'economical' (Robertson et al., 2012).

This review has drawn together various perspectives on some of the key issues that should be addressed in order to sequester more soil C and enhance soil productivity. In particular:

- In situ, time-efficient and cost-effective, SOC measurement methods, such as NIR techniques, should be developed further in order to measure or predict SOC stocks and the composition of SOC through the on-going generation of spectral libraries from different soil types and climatic conditions for the purpose of validation.
- The understanding of the functionality of the soil system, such as the influence of soil structure on the functioning of the soil-biota habitat, should also be improved.
- There is potential to improve SOC dynamic models through dialogue between 'empirical scientists' and 'modellers', for instance:
  - It might be worthwhile to consider a change from the current conceptual C pools of process-oriented models to measureable components of soil C and/or new views on SOM composition
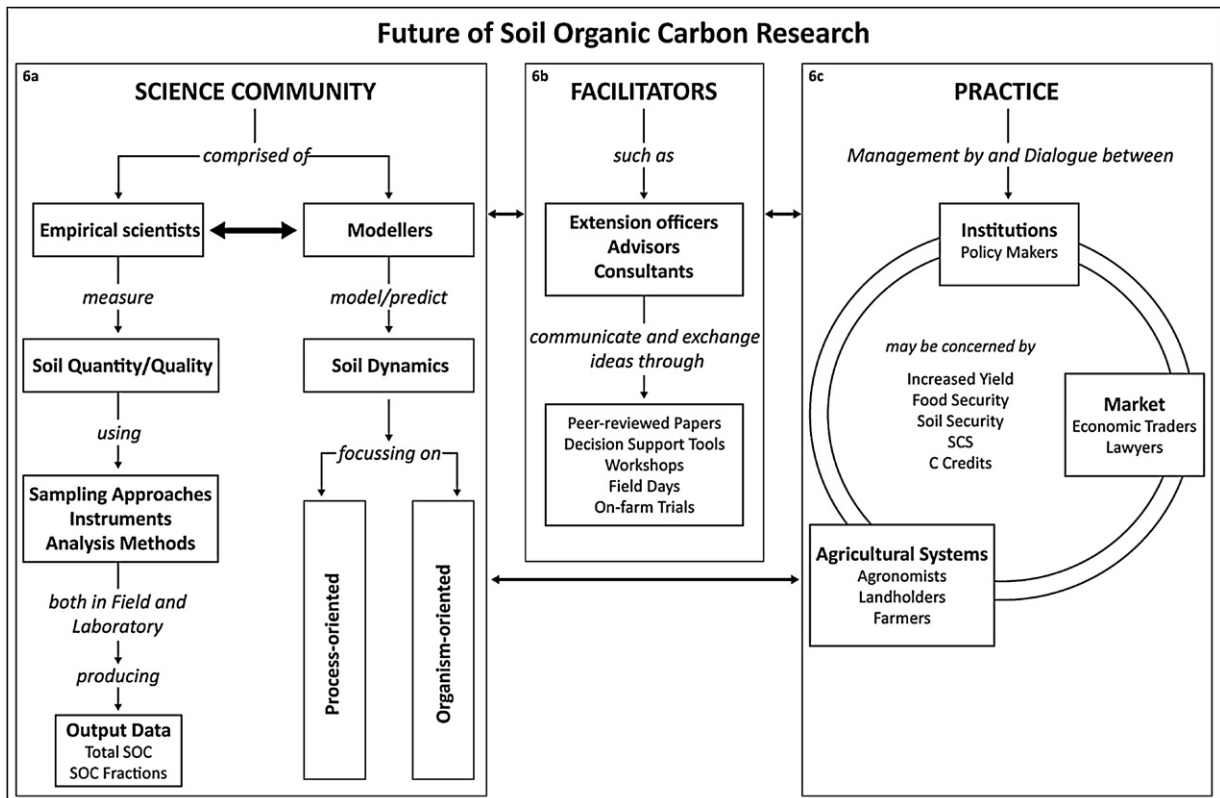

Fig. 6. Interdisciplinary research.

which in turn might improve the predictive capacity of models. This will also show how robust SOC models are and identify how they will evolve;

- To move towards landscape modelling of SOC dynamics to account for a range of different management practices and the redistribution of SOC to/from different parts in the landscape;
- To model changes in SOC for the whole soil profile. This also highlights the need for more studies that investigate potential effects of different management practices on the SOC distribution beyond the surface layer;
- Sampling strategies should be optimized on the regional scale such as stratified, random sampling, for potential use within a SOC accounting scheme.
- It should be investigated further if soil erosion enhances SOC loss or increases SOC by burying SOC at depth;
- The 'practice' of short-term versus long-term SCS should be more clearly defined in terms of 'which C 'pools' can equate to a permanent sequestration of C' and 'how can short-term C pool maintenance be rewarded to sustain natural and managed ecosystems'?.

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