



Review

Stabilization mechanisms of soil organic matter: Implications for C-saturation of soils

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Received 3 January 2001. Accepted in revised form 13 February 2002

Abstract

The relationship between soil structure and the ability of soil to stabilize soil organic matter (SOM) is a key element in soil C dynamics that has either been overlooked or treated in a cursory fashion when developing SOM models. The purpose of this paper is to review current knowledge of SOM dynamics within the framework of a newly proposed soil C saturation concept. Initially, we distinguish SOM that is protected against decomposition by various mechanisms from that which is not protected from decomposition. Methods of quantification and characteristics of three SOM pools defined as protected are discussed. Soil organic matter can be: (1) physically stabilized, or protected from decomposition, through microaggregation, or (2) intimate association with silt and clay particles, and (3) can be biochemically stabilized through the formation of recalcitrant SOM compounds. In addition to behavior of each SOM pool, we discuss implications of changes in land management on processes by which SOM compounds undergo protection and release. The characteristics and responses to changes in land use or land management are described for the light fraction (LF) and particulate organic matter (POM). We defined the LF and POM not occluded within microaggregates (53–250 μm sized aggregates as unprotected. Our conclusions are illustrated in a new conceptual SOM model that differs from most SOM models in that the model state variables are measurable SOM pools. We suggest that physicochemical characteristics inherent to soils define the maximum protective capacity of these pools, which limits increases in SOM (i.e. C sequestration) with increased organic residue inputs.

Introduction

Most current models of SOM dynamics assume first-order kinetics for the decomposition of various conceptual pools of organic matter (McGill, 1996; Paustian, 1994), which means that equilibrium C stocks are linearly proportional to C inputs (Paustian et al., 1997). These models predict that soil C stocks can, in theory, be increased without limit, provided that C inputs increase without limit, i.e. there are no assumptions of soil C saturation. While these models have been largely successful in representing SOM dynamics under current conditions and management practices (e.g. Parton et al., 1987, 1994; Paustian et

al., 1992; Powlson et al., 1996), usually for soils with low to moderate C levels (e.g. <5%), there is some question of their validity for projecting longer term SOM dynamics under scenarios of ever increasing C inputs (e.g. Donigian et al., 1997). Such scenarios are particularly relevant with the development of new technology designed to promote soil C sequestration through increasing plant C inputs.

Native soil C levels reflect the balance of C inputs and C losses under native conditions (i.e. productivity, moisture and temperature regimes), but do not necessarily represent an upper limit in soil C stocks. Empirical evidence demonstrates that C levels in intensively managed agricultural and pastoral ecosystems can exceed those under native conditions. Phosphorous fertilization of Australian pasture soils can increase soil C by 150% or more relative to the native condition

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(Barrow, 1969; Ridley et al., 1990; Russell 1960). Soil C levels under long-term grassland ('near native') vegetation have also been exceeded in high productivity mid-western no-tillage (NT) systems (Ismail et al., 1994) as well as in sod plots with altered vegetation (Follett et al., 1997). Hence, native soil C levels may not be an appropriate measure of the ultimate C sink capacity of soils.

There are several lines of evidence that suggest the existence of a C saturation level based on physiochemical processes that stabilize or protect organic compounds in soils. While many long-term field experiments exhibit a proportional relationship between C inputs and soil C content across treatments (Larson et al., 1972; Paustian et al., 1997), some experiments in high C soils show little or no increase in soil C content with two to three fold increases in C inputs (Campbell et al., 1991; Paustian et al., 1997; Solberg et al., 1997). Various physical properties (e.g. silt plus clay content and microaggregation) of soil are thought to be involved in the protection of organic materials from decomposer organism. However, these properties and their exerted protection seem to be limited by their characteristics (e.g. surface area), which is consistent with a saturation phenomenon (Hassink, 1997; Kemper and Koch, 1966).

A number of soil organic matter models have been developed in the last 30 years. Most of these models represent the heterogeneity of SOM by defining several pools, typically three to five, which vary in their intrinsic decay rates and in the factors which control decomposition rates (see reviews by McGill, 1996; Parton, et al., 1994; Paustian, 1994). Alternative formulations, whereby specific decomposition rate varies as a function of a continuous SOM quality spectrum (i.e. instead of discrete pools), have also been developed (e.g. Bosatta and Agren, 1996). However, in either case, the representation of the model pools (or quality spectrum) is primarily conceptual in nature. While such models can be successfully validated using measurements of total organic carbon and isotopic ratios of total C (e.g. Jenkinson and Rayner, 1977), the individual pools are generally only loosely associated with measurable quantities obtained with existing analytical methods. Consequently, it is not straightforward to falsify or test the internal dynamics of C transfers between pools and changes in pool sizes of the current SOM models with conceptual pool definitions because a direct comparison to measured pool changes is not possible.

A closer linkage between theoretical and measurable pools of SOM can be made by explicitly defining model pools to coincide with measurable quantities or by devising more functional laboratory fractionation procedures or both. The phrases 'modeling the measurable' and 'measuring the modelable' have been coined as representing the two approaches towards a closer reconciliation between theoretical and experimental work on SOM (Christensen, 1996; Elliott et al., 1996).

Various attempts have been made to correlate analytical laboratory fractions with conceptual model pools, with limited success. Motavalli et al. (1994) compared laboratory measurements of C mineralization with simulations by the Century model (Parton et al., 1994) for several tropical soils. When the active and slow pools in the model were initialized using laboratory determinations of microbial + soluble C for the active pool and light fraction for the slow pool, C mineralization was consistently underestimated, although all fractions were highly significantly correlated to C mineralization in a regression analysis. Magid et al. (1996) unsuccessfully attempted to trace ^{14}C labeled plant materials using three size-density fractionation methods to define an 'active' pool. Metherell (1992) found that the slow pool in Century was much larger than the particulate organic matter (POM) fraction isolated from a Haplustoll by Cambardella and Elliott (1992). However, Balesdent (1996) found that POM isolated after mild disruption corresponds to the plant structural compartment (RPM) of the Rothamsted carbon model (Jenkinson and Rayner, 1997). Acid hydrolysis has been used to estimate Century's passive C pool (Paul et al., 1997a; Trumbore, 1993), but it seems to slightly overestimate the size (Paul et al., 1997a; Trumbore, 1993), though not the C turnover rate, of the passive pool (Trumbore, 1993). Nevertheless, Paul et al. (1999) used extended laboratory incubations in combination with acid hydrolysis to define an active, slow and passive pool of C and were successful in modeling the evolution of CO_2 in the field based on these pools. These studies suggest that attempting to measure the modelable has had minimal success to date.

There have been a few recent attempts to more closely integrate models and measurements of physiochemically defined pools by 'modeling the measurable', although Elliott et al. (1996) and Christensen (1996) have presented conceptual models for this approach. Arah (2000) proposed an approach based on analytically defined pools and measurements of ^{13}C

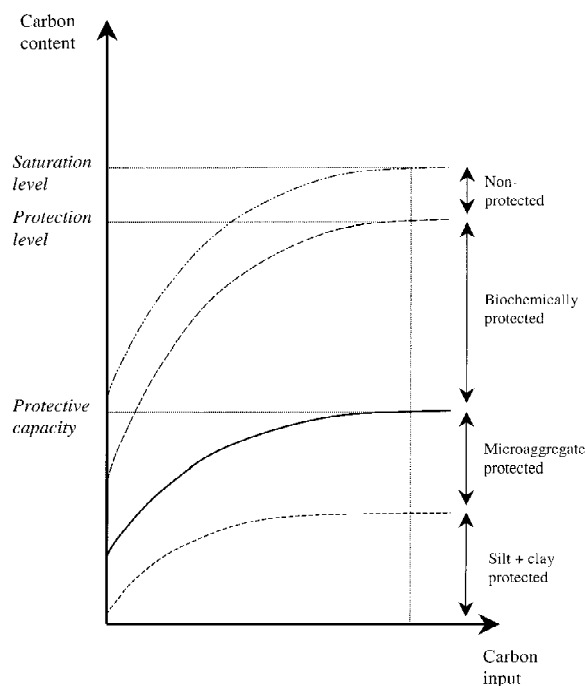


Figure 1. The protective capacity of soil (which governs the silt- and clay protected C and microaggregate protected C pools), the biochemically stabilized C pool and the unprotected C pool define a maximum C content for soils. The pool size of each fraction is determined by their unique stabilizing mechanisms.

and ^{15}N stable isotope tracers to derive parameters for a model with measurable pools. The approach considers all possible transformations between measured C and N pools and devises a system of equations using observed changes in total C and N and ^{13}C and ^{15}N for each fraction to solve all model unknowns. Necessary requirements of such an approach are that the analytical fractions are distinct and together account for the total carbon inventory.

The objective of this review paper is to summarize current knowledge on SOM dynamics and stabilization and to synthesize this information into a conceptual SOM model based on physicochemically defined SOM pools. This new model defines a soil C-saturation capacity, or a maximum soil C storage potential, determined by the physicochemical properties of the soil. We propose that the conceptual model developed from this knowledge may form the basis for a simulation model with physicochemically measurable SOM pools as state variables rather than with the biologically defined pools by Paul et al. (1999).

Protected SOM: Stabilization mechanisms, characteristics, and dynamics

Three main mechanisms of SOM stabilization have been proposed: (1) chemical stabilization, (2) physical protection and (3) biochemical stabilization (Christensen, 1996; Stevenson, 1994). Chemical stabilization of SOM is understood to be the result of the chemical or physicochemical binding between SOM and soil minerals (i.e. clay and silt particles). Indeed, many studies have reported a relationship between stabilization of organic C and N in soils and clay or silt plus clay content (Feller and Beare, 1997; Hassink, 1997; Ladd et al., 1985; Merckx et al., 1985; Sorensen, 1972). In addition to the clay content, clay type (i.e. 2:1 versus 1:1 versus allophanic clay minerals) influences the stabilization of organic C and N (Feller and Beare, 1997; Ladd et al. 1992; Sorensen, 1972; Torn et al., 1997). Physical protection by aggregates is indicated by the positive influence of aggregation on the accumulation of SOM (e.g. Edwards and Bremner, 1967; Elliott, 1986; Jastrow, 1996; Tisdall and Oades, 1982; Six et al., 2000a). Aggregates physically protect SOM by forming physical barriers between microbes and enzymes and their substrates and controlling food web interactions and consequently microbial turnover (Elliott and Coleman, 1988). Biochemical stabilization is understood as the stabilization of SOM due to its own chemical composition (e.g. recalcitrant compounds such as lignin and polyphenols) and through chemical complexing processes (e.g. condensation reactions) in soil. For our analyses, we divide the protected SOM pool into three pools according to the three stabilization mechanisms described (Figure 1). The three SOM pools are the silt- and clay-protected SOM (silt and clay defined as $<53\ \mu\text{m}$ organomineral complexes), microaggregate-protected SOM (microaggregates defined as $53\text{--}250\ \mu\text{m}$ aggregates), and biochemically protected SOM.

Chemical stabilization: Silt- and clay-protected SOM

The protection of SOM by silt and clay particles is well established (Feller and Beare, 1997; Hassink, 1997; Ladd et al., 1985; Sorensen, 1972). Hassink (1997) examined the relationship between SOM fractions and soil texture and found a relationship between the silt- and clay-associated C and soil texture, though he did not find any correlation between texture and amount of C in the sand-sized fraction (i.e. POM C). Based on these findings, he defined the capacity

Table 1. Regression equations relating silt plus clay proportion to silt and clay associated C

Size class ^a	Ecosystem	Intercept	Slope	r ²
0–20 μm	Cultivated	4.38 \pm 0.68 ^b	0.26 \pm 0.01	0.41
	Grassland	2.21 \pm 1.94	0.42 \pm 0.08	0.44
	Forest	–2.51 \pm 0.55	0.63 \pm 0.01	0.55
0–50 μm	Cultivated	7.18 \pm 3.04	0.2 \pm 0.04	0.54
	Grassland	16.33 \pm 4.69	0.32 \pm 0.07	0.35
	Forest	16.24 \pm 6.01	0.24 \pm 0.08	0.35
Size class	Clay type	Intercept	Slope	r ²
0–20 μm	1:1	1.22 \pm 0.37	0.30 \pm 0.01	0.74
	2:1	3.86 \pm 0.49	0.41 \pm 0.01	0.39
0–50 μm	1:1	5.5 \pm 5.93	0.26 \pm 0.13	0.38
	2:1	14.76 \pm 2.37	0.21 \pm 0.03	0.07

^aTwo size classes for silt and clay were reported in the literature.

^bValue \pm 95% confidence interval.

of soil to preserve C by its association with silt and clay particles. Studies investigating the retention of specific microbial products (i.e. amino sugars) corroborate the proposition of Hassink (1997) that C associated with primary organomineral complexes are chemically protected and the amount of protection increased with an increased silt plus clay proportion of the soil (Chantigny et al., 1997; Guggenberger et al., 1999; Puget et al., 1999; Sorensen, 1972). Puget et al. (1999) reported an enrichment of microbial derived carbohydrates in the silt plus clay fraction compared to the sand fraction of no-tilled and conventional tilled soils. However, the amount stabilized by silt and clay differs among microbial products. For example, Guggenberger et al. (1999) reported a higher increase of glucosamine than muramic acid under no-tillage at sites with a high silt plus clay content. A reexamination of the data presented by Chantigny et al. (1997) leads to the observation that the glucosamine/muramic acid ratio was only higher in perennial systems compared to annual systems in a silty clay loam soil and not in a clay loam soil. The silty clay loam soil had a higher silt plus clay content.

We expanded the analysis of Hassink (1997) of the physical protection capacity for C associated with primary organomineral complexes (Figure 2) across ecosystems (i.e. forest, grassland, and cultivated systems), clay types (i.e. 1:1 versus 2:1), and size ranges for clay and silt (0–20 μm and 0–50 μm ; see Ap-

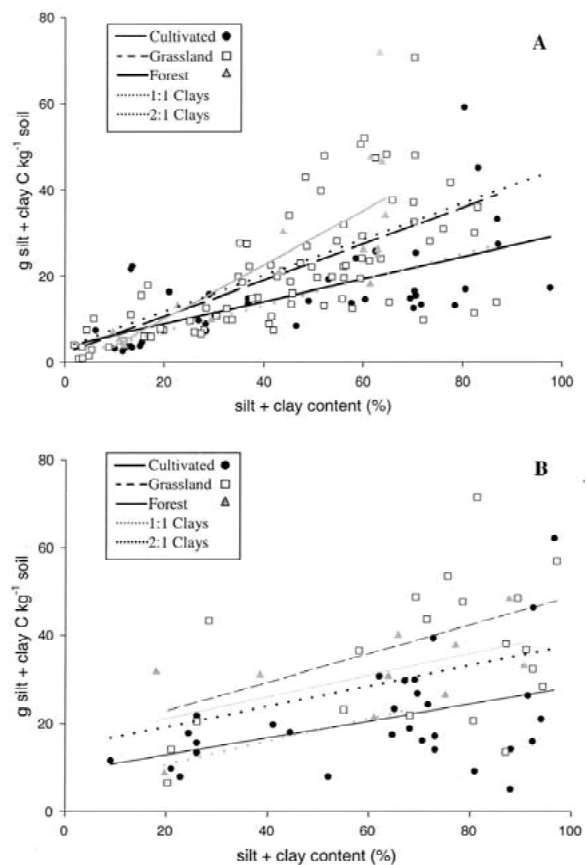


Figure 2. The relationship between silt+clay content (%) and silt+clay associated C (g silt+clay C kg^{–1} soil) for grassland, forest and cultivated ecosystems. A differentiation between 1:1 clay and 2:1 clay dominated soils is also made. The relationships indicate a maximum of C associated with silt and clay (i.e. C saturation level for the clay and silt particles), which differs between forest and grassland ecosystems and between clay types. Two size boundaries for silt+clay were used (A) 0–20 μm and (B) 0–50 μm .

pendices for details). Following the methodology of Hassink (1997) we performed regressions (Figure 2 and Table 1) between the C content associated with silt and clay particles (g C associated with silt and clay particles kg^{–1} soil; Y axis) and the proportion of silt and clay particles (g silt plus clay g^{–1} soil; X axis). All regressions were significant ($P < 0.05$) and comparison of regression lines revealed that the influence of soil texture on mineral-associated C content differed depending on the size range used for clay and silt particles. Consequently, we did regressions for two different size classes of silt and clay particles (i.e. 0–20 μm and 0–50 μm ; Figure 2 and Table 1). The intercept for the 0–50 μm silt and clay particles was significantly higher than for the 0–20 μm silt and clay particles (Table 1). This difference in intercept was

probably a result of the presence of larger sized (20–50 μm) silt-sized aggregates in the 0–50 μm than in the 0–20 μm silt and clay particles. These larger silt-sized aggregates have more C per unit material because additional C binds the primary organomineral complexes into silt-sized aggregates (Tisdall and Oades, 1982). However the difference in intercept might also be the result of POM particles of the size 20–50 μm associated with the 0–50 μm fraction (Turchenek and Oades, 1979). Intercepts for cultivated and forest ecosystems were significantly different for the 0–50 μm particles, but were only marginally significantly different ($P < 0.06$) for the 0–20 μm particles. Slopes for grassland soils (0–20 μm particles) were significantly different than those for forest and cultivated soils. The differences between grasslands and cultivated lands are likely due to differences in input and disturbance, which causes a release of SOM and consequently increased C availability for decomposition. An explanation for the significantly different slopes for grassland and forest soils (Table 1) is not immediately apparent. Especially that the slope is higher for forest than grassland slopes. This is in contrast to the suggestion that grassland-derived soils have a higher potential of C stabilization than forest-derived because of their higher base saturation (Collins et al., 2000; Kononova, 1966). Consequently, this difference in C stabilization by silt and clay particles between forest and grassland systems should be investigated further.

In contrast to Hassink (1997), we found significantly different relationships for 1:1 clays versus 2:1 clays regressions and for the cultivated versus grassland regressions (Figure 2 and Table 1) for the 0–20 μm particles. The effect of clay type was also significant for the 0–50 μm particles. This lower stabilization of C in 1:1 clay dominated soils is probably mostly related to the differences between the clay types (see below). However, the effect of climate can not be ignored in this comparison because most 1:1 clay dominated soils were located in (sub)tropical regions. The higher temperature and moisture regimes in (sub)tropical regions probably also induce a faster decomposition rate and therefore contributes to the lower stabilization of C by the 1:1 clays. Nevertheless we believe that the type of clay plays an important role because different types of clay (i.e. 1:1 and 2:1 clays) have substantial differences in CEC and specific surface (Greenland, 1965) and should, consequently, have different capacities to adsorb organic materials. In addition, Fe- and Al-oxides are most often found in soils dominated by 1:1 minerals and are strong flocculants. By

being strong flocculants, Fe- and Al-oxides can reduce even further the available surface for adsorption of SOM. We are not certain why soils examined by Hassink (1997) did not follow this reduced capacity to adsorb organic materials; few soils dominated by 1:1 clays, however, were included in the data set used by Hassink (1997) and most of them had a low carbon content. Nevertheless, the difference between the two studies might also be a result of the contrasting effect the associated Fe- and Al-oxides can have. The strong flocculating oxides can reduce available surface (see above) but they might also co-flocculate SOM and consequently stabilize it. Therefore, it appears that mechanisms with contrasting effects on SOM stabilization exist and the net effect still needs to be investigated. The different regression lines for grassland and cultivated systems are in accordance with Feller et al. (1997). They also found a significant lower slope for the regression line between the amount of 0–2 μm particles and the C contained in the 0–2 μm fraction of cultivated soils compared to non-cultivated soils. The lack of influence of cultivation on the silt and clay associated C observed by Hassink (1997) was probably a result of the low proportion of silt and clay and high SOM contents of the soils used. The silt- and clay-associated C formed a small fraction of the total C in his soils. Consequently, sand-associated C accounted for the majority of total soil C. Given this dominance of sand-associated C and its greater sensitivity to cultivation than silt- and clay-associated C (Cambardella and Elliott, 1992), in which C is transferred from the sand associated fraction to the silt- and clay-associated fractions during decomposition (Guggenberger et al., 1994), a loss of silt- and clay-associated C upon cultivation is likely to be minimal.

In summary, we found, as Hassink (1997) did, a direct relationship between silt plus clay content of soil and the amount of silt- and clay-protected soil C, indicating a saturation level for silt and clay associated C. This relationship was different between different types of land use, different clay types, and for different determinations of silt plus clay size class. Also, the silt- and clay-associated soil organic matter was reduced by cultivation.

Physical protection: Microaggregate-protected SOM

The physical protection exerted by macro- and/or microaggregates on POM C is attributed to: (1) the compartmentalization of substrate and microbial bio-

mass (Killham et al., 1993; van Veen and Kuikman, 1990), (2) the reduced diffusion of oxygen into macro- and especially microaggregates (Sexstone et al., 1985) which leads to a reduced activity within the aggregates (Sollins et al., 1996), and (3) the compartmentalization of microbial biomass and microbial grazers (Elliott et al., 1980). The compartmentalization between substrate and microbes by macro- and microaggregates is indicated by the highest abundance of microbes on the outer part of the aggregates (Hattori, 1988) and a substantial part of SOM being at the center of the aggregates (Elliott and Coleman, 1988; Golchin et al., 1994). In addition, Bartlett and Doner (1988) reported a higher loss of amino acids by respiration from the aggregate surfaces than from within aggregates. Priesack and Kisser-Priesack (1993) showed that the rate of glucose utilization decreased with distance into the aggregate. The inaccessibility of substrate for microbes within aggregates is due to pore size exclusion and related to the water-filled porosity (Killham et al., 1993).

Many studies have documented a positive influence of aggregation on the accumulation of SOM (Angers et al., 1997; Besnard et al., 1996; Cambardella and Elliott, 1993; Franzluebbers and Arshad, 1997; Gale et al., 2000; Golchin et al., 1994, 1995; Jastrow, 1996; Monreal and Kodama, 1997; Paustian et al., 2000; Puget et al., 1995, 1996; Six et al., 1998, 1999, 2000a). Cultivation causes a release of C by breaking up the aggregate structures, thereby increasing availability of C. More specifically, cultivation leads to a loss of C-rich macroaggregates and an increase of C-depleted microaggregates (Elliott, 1986; Six et al., 2000a). The inclusion of SOM in aggregates also leads to a qualitative change of SOM. For example, Golchin et al. (1994) reported significant differences in chemical structure between the free and occluded (i.e. within aggregates) light fraction. The occluded light fraction had higher C and N concentrations than the free light fraction and contained more alkyl C (i.e. long chains of C compounds such as fatty acids, lipids, cutin acids, proteins and peptides) and less O-alkyl C (e.g. carbohydrates and polysaccharides). These data suggest that during the transformation of free into intraaggregate light fraction there is a selective decomposition of easily decomposable carbohydrates (i.e. O-alkyl C) and preservation of recalcitrant long-chained C (i.e. alkyl C) (Golchin et al., 1994). Golchin et al. (1995) also found that cultivation decreased the O-alkyl content of the occluded SOM. They suggested that this difference is a result of the continuous

disruption of aggregates, which leads to a faster mineralization of SOM and a preferential loss of readily available O-alkyl C. Hence, the enhanced protection of SOM by aggregates in less disturbed soil results in an accumulation of more labile C than would be maintained in a disturbed soil.

Recent studies indicate that the macroaggregate ($>250\ \mu\text{m}$) structure exerts a minimal amount of physical protection (Beare et al., 1994; Elliott, 1986; Pulleman and Marinissen, 2001), whereas SOM is protected from decomposition in free (i.e. not within macroaggregates) microaggregates ($<250\ \mu\text{m}$) (Balesdent et al., 2000; Besnard et al., 1996; Skjemstad et al., 1996) and in microaggregates within macroaggregates (Denef et al., 2001; Six et al., 2000b). Beare et al. (1994) and Elliott (1986) found an increase in C mineralization when they crushed macroaggregates, but the increase in mineralization only accounted for 1–2% of the C content of the macroaggregates. In addition, no difference in C mineralization between crushed and uncrushed macroaggregates has been observed (Pulleman and Marinissen, 2001). In contrast, C mineralization of crushed free microaggregates was three to four times higher than crushed macroaggregates (Bossuyt et al., 2002). Gregorich et al. (1989) observed a substantial higher C mineralization when microaggregates within the soil were disrupted than when lower disruptive energies were used that did not break up microaggregates. Jastrow et al. (1996), using ^{13}C natural abundance technique, calculated that the average turnover time of C in free microaggregates was 412 yr, whereas the average turnover time for macroaggregate associated C was only 140 yr in the surface 10 cm. These studies clearly indicate that C stabilization is greater within free microaggregates than within macroaggregates. Further corroborating evidence for the crucial role microaggregates play in C sequestration were reported by Angers et al. (1997), Besnard et al. (1996), Gale et al. (2000) and Six et al. (2000b). Angers et al. (1997) found in a field incubation experiment with ^{13}C and ^{15}N labeled wheat straw that wheat-derived C was predominantly stored and stabilized in free microaggregates. Gale et al. (2000) reported similar C stabilization within free microaggregates in an incubation study with ^{14}C -labeled root material. Upon conversion of forest to maize cultivation, Besnard et al. (1996) found a preferential accumulation of maize- and forest-derived POM-C in microaggregates compared to other soil fractions. Six et al. (1999) observed a decrease in fine intra-macroaggregate-POM (i.e. 53–250 μm sized POM

(fine iPOM) predominantly stabilized in microaggregates within macroaggregates (Six et al., 2000b)) under plough tillage compared to no-till. However, there was no difference in coarse intra-macroaggregate POM (i.e. 250–2000 μm POM not stabilized by the microaggregates within macroaggregates) between tillage systems at three of the four sites studied. They concluded that the incorporation and stabilization of fine POM-C into microaggregates within macroaggregates and free microaggregates under no-tillage is a dominant factor for protection of the fine-sized fraction of POM. Nevertheless, the dynamics of macroaggregates are crucial for the sequestration of C because it influences the formation of microaggregates and the sequestration of C within these microaggregates (Six et al., 2000b). That is, rapid turnover of macroaggregates reduces the formation of microaggregates within macroaggregates and the resulting stabilization of C within these microaggregates (Six et al., 1998, 1999, 2000b).

Though the incorporation of POM into microaggregates (versus bonding to clay surfaces; i.e. chemical mechanism) seems to be the main process for protection of POM, the clay content and type of soil exert an indirect influence on the protection of POM-C by affecting aggregate dynamics. Franzluebbers and Arshad (1997) suggested that physical protection of POM within aggregates increases with clay content since mineralization of POM-C relative to whole-SOM-C after dispersion and aggregation both increased with increasing clay content (Franzluebbers and Arshad, 1996). Different clay types lead to different mechanisms involved in aggregation (Oades and Waters, 1991) and will therefore influence differently the protection of POM through microaggregation. Within the 2:1 clay minerals, clay minerals with a high CEC and larger specific surface, such as montmorillonite and vermiculite, have a higher binding potential than clay minerals with a lower CEC and smaller specific surface, such as illite (Greenland, 1965). In contrast to the 2:1 minerals, kaolinite and especially Fe- and Al-oxides have a high flocculation capacity due to electrostatic interactions through their positive charges (Dixon, 1989; Schofield and Samson, 1954). Even though, different mechanisms prevail in soils with different clay types, soils seem to have a maximum level of aggregate stability. Kemper and Koch (1966) observed that aggregate stability increased to a maximum level with clay content and free Fe-oxides content. Since the physical protection of POM seems to be mostly determined by microaggregation, we hypothes-

ize that the maximum physical protection capacity for SOM is determined by the maximum microaggregation, which is in turn determined by clay content, clay type.

Biochemical stabilization: Biochemically-protected SOM

In this review, a detailed description of the influence of biochemical stabilization on SOM dynamics will not be given, we refer to an excellent review on this subject by Cadisch and Giller (1997). Nevertheless, biochemical stabilization of SOM needs to be considered to define the soil C-saturation level within a certain ecosystem (Figure 1). Biochemical stabilization or protection of SOM occurs due to the complex chemical composition of the organic materials. This complex chemical composition can be an inherent property of the plant material (referred to as residue quality) or be attained during decomposition through the condensation and complexation of decomposition residues, rendering them more resistant to subsequent decomposition. Therefore the third pool in our model (Figure 1) is a SOM pool that is stabilized by its inherent or acquired biochemical resistance to decomposition. This pool is akin to that referred to as the 'passive' SOM pool (Parton et al., 1987) and its size has been equated to the non-hydrolyzable fraction (Leavitt et al., 1996; Paul et al., 1995; Trumbore 1993). Using ^{14}C dating, it has been found that, in the surface soil layer, the non-hydrolyzable C is approximately 1300 years older than total soil C (Paul et al., 1997a, 2001). Several studies have found that the non-hydrolyzable fraction in temperate soils includes very old C (Anderson and Paul, 1984; Paul et al., 1999; Trumbore, 1993; Trumbore et al., 1996) and acid hydrolysis removes proteins, nucleic acids, and polysaccharides (Schnitzer and Khan, 1972) which are believed to be more chemically labile than other C compounds, such as aromatic humified components and wax-derived long chain aliphatics (Paul et al., 1997a). The stabilization of this pool and consequent old age is probably predominantly the result of its biochemical composition. However, Balesdent (1996) did not find any great differences in dynamics between the non-hydrolyzable and hydrolyzable C fraction and therefore questioned the relationship between biodegradability and hydrolyzability. Nevertheless, we chose the hydrolysis technique to differentiate an older and passive C pool, because we think it is the simplest and best available technique to define

such pool and others (Paul et al., 1995, 1999, 2001; Trumbore, 1993) have shown it to be an appropriate technique for such applications. The size of this pool is also likely to be limited since the mean age of C in the passive pool is generally much younger than the pedogenic age of the soil.

Unprotected SOM: Characteristics and dynamics

Recently derived, partially decomposed plant residues that are not closely associated with soil minerals constitute the unprotected SOM pool. We suggest that this unprotected SOM pool is measurable as either the light fraction (LF) or POM fraction and therefore consider these two fractions as conceptually similar pools even though differences in characteristics do exist between them (see below). Many studies have found that the LF and POM, especially coarse POM ($>250\ \mu\text{m}$), are relatively easily decomposable and are greatly depleted upon cultivation (e.g. Cambardella and Elliott, 1992; Six et al., 1999; Solomon et al., 2000), indicating their relatively unprotected (biochemical and physical) status. The POM and the LF are isolated by size and/or density separation. Many different size classes, densities, and degrees of soil dispersion before isolation are used, confounding comparisons among studies. The LF is isolated by flotation in a high-density liquid after a certain degree of disruption or even complete dispersion of the soil (Janzen et al., 1992; Meijboom et al., 1995; Sollins et al., 1984). The yield of LF depends strongly on the density used and the level of soil dispersion before the density flotation. In order to isolate unprotected LF, microaggregates should not be broken up before flotation (see above). Consequently, unprotected LF is only isolated after a minimal level of dispersion of the soil, i.e. the disruption level used, should break up only macroaggregates and not microaggregates. We define unprotected POM as the 53–2000 μm sized POM not contained within microaggregates and protected POM as 53–250 μm sized POM contained within microaggregates. Since the macroaggregate versus microaggregate boundary is size determined, a size separation methodology is more standardizable than a density separation methodology for the isolation of protected versus unprotected SOM as defined in this review. A specific method is described in the last section of this review. For this paper we review the available information for both the LF and POM independent of methodology used because: (1) characteristics of the two fractions are fairly

similar (see Table 2), and (2) most studies did not differentiate between protected and unprotected LF and POM.

Plant Origin of unprotected SOM

Both the LF and POM are mainly comprised of plant residues, but also contain seeds and microbial debris, such as fungal hyphae and spores (Besnard et al., 1996; Dalal and Mayer, 1986a,b; Molloy and Speir, 1977; Oades et al., 1987). Several authors have also reported the presence of charcoal in LF and POM (Cambardella and Elliott, 1992; Greenland and Ford, 1964; Molloy and Speir, 1977; Skjemstad et al., 1990). The predominantly plant origin of LF and POM is easily verifiable under a microscope and is also confirmed by many biochemical characterization studies on these fractions. First, several studies have indicated a lower carbohydrate concentration in POM and LF compared to the smaller-sized or heavier fractions, respectively (Angers and Mehuys, 1990; Catroux and Schnitzer, 1987; Cheshire and Mundie, 1981; Greenland and Ford, 1964; Guggenberger et al., 1994; Murayama et al., 1979; Oades et al., 1987; Skjemstad et al., 1986; Solomon et al., 2000). However, a higher carbohydrate-C enrichment ratio in the POM fraction than in the clay-sized fraction has been reported (Dalal and Henry, 1988; Guggenberger et al., 1994). Despite these contradictory results on carbohydrate content among the above-mentioned studies, all studies observed a lower $((\text{M})\text{annose} + (\text{G})\text{alactose}) / ((\text{A})\text{rabinose} + (\text{X})\text{ylose})$ ratio for POM and LF compared to smaller-sized and heavier fractions, respectively. Second, the thermogram of POM obtained by pyrolysis-field ionization mass spectrometry had a sharp maximum at 370 °C, symmetrical peak shapes and a simultaneous evolution of phenols, lignin monomers, lignin dimers and alkylaromatics (Schulten et al., 1993). Third, POM was enriched in phenolic CuO oxidation products (vanillyl (V), syringyl (S), and cinnamyl (C), (VSC)) with a low acid-to-aldehyde ratio's of the vanillyl units $[(\text{ac}/\text{al})_{\text{v}}]$ and had high syringyl-to-vanillyl ratio (S/V), indicating a high lignin content which is only little altered by microbes (Amelung et al., 1999; Guggenberger et al., 1994; Six et al., 2001; Solomon et al., 2000).

Even though the above described characteristics indicate the predominantly plant origin of the LF and POM fraction, they also indicate a partial decomposition stage and microbial contribution to these fractions. The intermediate stage between plant mater-

Table 2. General characteristics of the light fraction (LF) and particulate organic matter (POM)

Characteristics
1. Consists of plant residues in various stages of decomposition.
2. Presence of charcoal
3. Mannose+Galactose/Arabinose+Xylose ratio is low
4. High O-alkyl content
5. High C/N ratio
6. Low net N mineralization potential
7. Labile SOM pool
8. High lignin content (Vanillyl, Syringyl, Cinnamyl content high; Phenylpropenoic acid/benzoic acid ratio high)
9. Microbial biomass and microbial debris are associated with LF

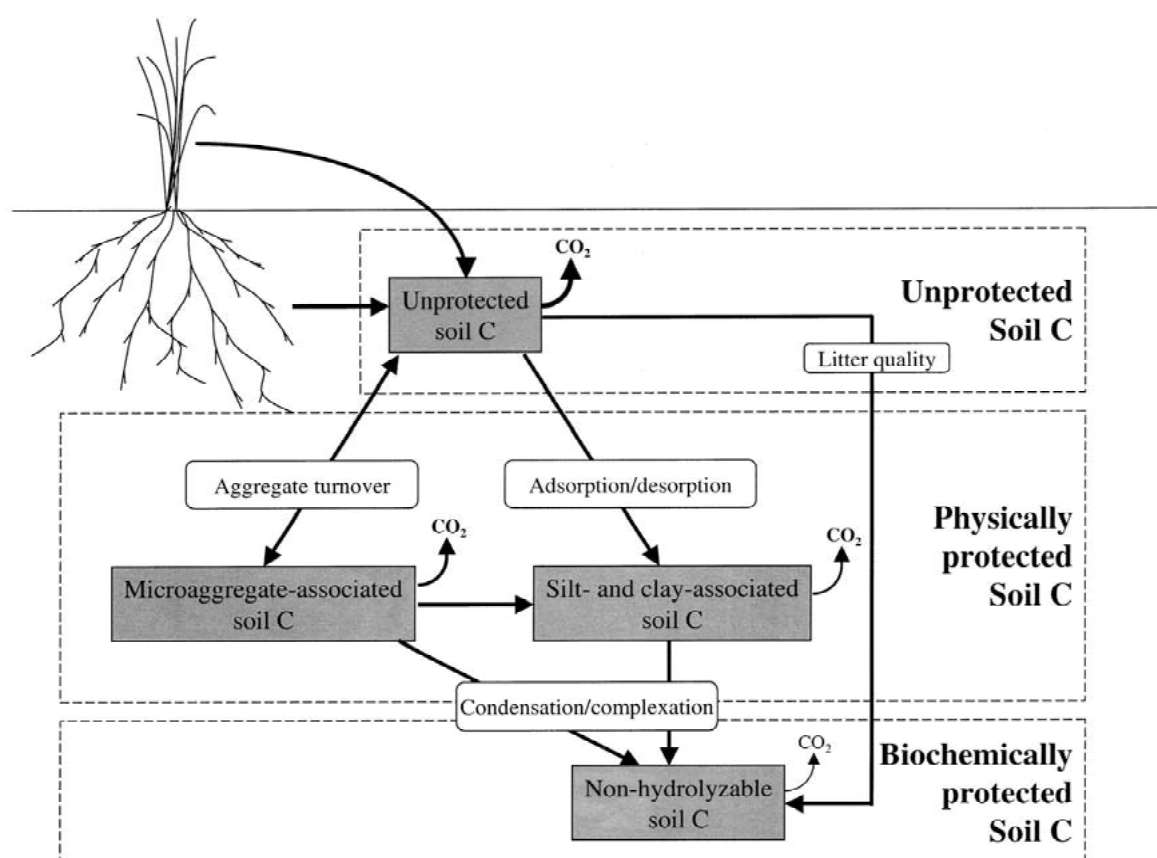


Figure 3. Conceptual model of soil organic matter (SOM) dynamics with measurable pools. The soil processes of aggregate formation/degradation, SOM adsorption/desorption and SOM condensation/complexation and the litter quality of the SOM determine the SOM pool dynamics.

ials and humified organic matter of the LF is confirmed by its amino acid composition (Greenland and Ford, 1964; Turchenek and Oades, 1979) and it has been observed that microorganisms are associated with the LF (Chotte et al., 1998; Jocteur Monrozier et al., 1991). In addition, amino sugar, glucosamine, and muramic acid

analyses indicate a significant microbial contribution to POM, which is mostly fungal (Six et al., 2001; Solomon et al., 2001). Consequently, POM and LF or the unprotected SOM fraction is a mixture of compounds caused by a regenerating plant residue pool and partial microbial decomposition.

Unprotected SOM as a nutrient source

The LF and POM have a high C/N ratio and a low net N mineralization (N_{\min}) potential (Catroux and Schnitzer, 1987; Chichester, 1969; Greenland and Ford, 1964; Lowe and Hinds, 1983; Sollins et al., 1984; Tiessen and Stewart, 1983; Turchenek and Oades, 1979). For example, Boone (1994) measured the N_{\min} potential of LF for a cornfield, pine stand and maple stand and found that LF only represented 11% (maize), 13% (pine) and 2% (maple) of the N_{\min} potential of the whole mineral soil. The net N_{\min} of the LF and POM has been directly related to its N concentration and inversely related to its C:N ratio (Barrios et al., 1996; Hassink, 1995; Sollins et al., 1984), indicating the importance of the net N immobilization process, resulting from the N demand associated with initial microbial growth, for the decomposition of these fractions (Recous et al., 1999).

It has been suggested that the LF and POM are good indicators for labile organic matter (Janzen et al., 1992). Research has shown a strong correlation between LF C and/or N amount and soil respiration and/or mineralizable N (Alvarez et al., 1998; Curtin and Wen, 1999; Janzen et al., 1992; Sierra, 1996). For example, LF content explained 40–50% of the field variation in N_{\min} of a maize field in the Argentine Pampa (Sierra, 1996). As the LF, POM is a significant contributor to N mineralization and turnover. A highly significant relationship between POM N and N uptake by maize was observed in a bioassay study (Vanlauwe et al., 1998) and in field trials (Vanlauwe et al., 1999).

Influence of management on unprotected SOM

Since LF and POM are labile organic matter pools, they are sensitive to management practices (Balesdent, 1996; Biederbeck et al., 1994; Bremer et al., 1994; Janzen, 1987; Janzen et al., 1992; Tiessen and Stewart, 1983; Solomon et al., 2000) and consequently highly influenced by the cultivation history of the soil (Greenland and Ford, 1964). Examples are: (1) the N pool size of the LF was significantly different between no-tillage, chisel tillage and plough tillage, whereas there were no differences in the heavy fraction among these treatments (Alvarez et al., 1998); (2) the frequency of fallow in wheat rotations in southwestern Saskatchewan was found to be the dominant factor affecting the LF-C and LF-N content (Biederbeck et al., 1994); and (3) the amounts of LF-N were significantly different between tree fallow systems with different tree species after 2 and 3 years of establishment (Barrios et

al., 1997). Bremer et al. (1994) concluded that the LF is the most robust indicator of management-induced changes in SOM. They calculated the sensitivity of total C, light fraction and mineralized C to different spring wheat crop rotations by subtracting the value of the treatment with the lowest level of the respective parameters from the value of the treatment with the highest level and dividing it by the lowest value. The results were 0.2, 2.5 and 1.5 for total C, light fraction and mineralizable C respectively.

Using ^{13}C natural abundance techniques, Balesdent (1996) concluded that POM has a short mean residence time relative to C associated with clay- and silt-sized organomineral complexes, indicating the relative high lability of POM. Tiessen and Stewart (1983) observed a strong decrease in POM after 60 years of cultivation; the biggest reduction of SOM was observed in the floatable POM. Others have reported that coarse organic matter ($>100\ \mu\text{m}$ or $>250\ \mu\text{m}$) was the SOM pool mostly influenced by type of management (Quiroga et al., 1996; Solomon et al., 2000). These results indicate that POM, especially floatable and coarse POM, provides an earlier indication of the consequences of different soil managements than did total organic matter (Dalal and Mayer, 1986a,b).

New conceptual SOM model with measurable pools

The information reviewed here suggests that SOM behaves in ways that can be explained by generalizing basic physicochemical soil processes and that a model structure relating these processes is capable of describing the behavior of measurable SOM pools (Figure 3). The model (currently at a conceptual stage) integrates traditional first order decomposition dynamics of SOM with the main SOM stabilization mechanisms that have been identified. An important ramification of this model is the concept that storage capacity of soils may become saturated with respect to C. Indications for a carbon saturation level in soils were given by Campbell et al. (1991) and Solberg et al. (1997). Both studies found no increase in soil C content with a two to three fold increase in C inputs. In addition, by relating C inputs with C content for 48 agricultural systems across 11 sites (Paul et al. 1997b), we found that an asymptotic relationship explained slightly more of the observed variability than a linear relationship (Figure 4).

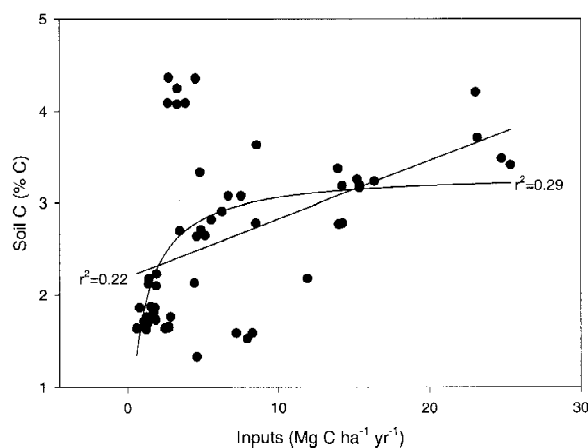


Figure 4. Relationship between carbon inputs ($\text{Mg C ha}^{-1} \text{ yr}^{-1}$) and total soil organic carbon (Mg C ha^{-1}) across 11 agricultural field experiment sites with an array of experimental units (Paul et al., 1997b).

Basic, mechanistic soil processes are incorporated in the model but the conceptual pools (active, slow and passive) developed elsewhere (Jenkinson and Rayner, 1977; Parton et al., 1987) are still reflected within the structure of the model (Figure 3). We hypothesize that the microaggregate-protected C plus silt- and clay-protected C represents part of the slow pool whereas the unprotected pool represents the active fraction and part of the slow pool. The non-hydrolyzable fraction of the silt- and clay-associated C represents the biochemically protected pool (Figure 3) and is hypothesized to be comparable to the passive pool. The unprotected C pool is POM or LF not occluded within microaggregates whereas the microaggregate-protected C pools is fine POM occluded within microaggregates.

The processes of microaggregate formation/degradation (i.e. microaggregate turnover), adsorption/desorption, and condensation/complexation are the main mechanisms of protection and release of SOM (Figure 3). The direct transformation of unprotected soil C to the biochemically stabilized non-hydrolyzable soil C is mainly determined by litter quality. Incorporation of the processes of desorption and adsorption was used by Hassink and Whitmore (1997) to model silt- and clay-protection of SOM. Condensation and complexation dynamics have been studied and information is available in the literature (Stevenson, 1994). Microaggregate formation and degradation are beginning to be understood, but more detailed information about microaggregate behavior must be obtained to accurately model these processes.

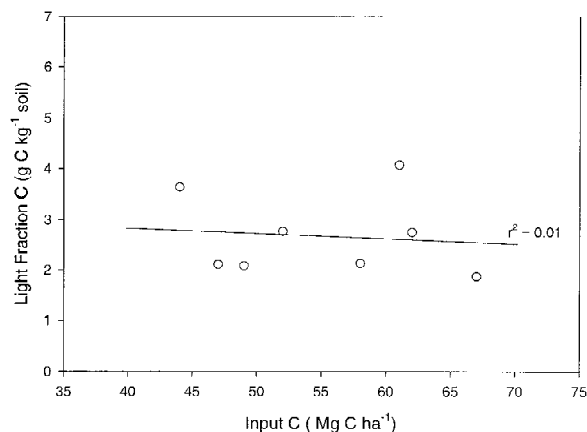


Figure 5. Relationship between carbon inputs (Mg C ha^{-1}) and light fraction C ($\text{g C kg}^{-1} \text{ soil}$) in an agricultural field experiment at Melfort, Saskatchewan, Canada (Campbell et al., 1991; Janzen et al., 1992).

Finally, the size of the unprotected SOM pool is a function of C inputs and the specific decomposition rate of the various C components. The decomposition rate of the unprotected C pool is, by definition, independent of the level of chemical and physical protection, but it is related to soil moisture, temperature, intrinsic biodegradability and N availability as principal controls on microbial activity. There are indications that the unprotected C pool also becomes saturated. Solberg et al. (1997) found a higher amount of LF-C when 25 kg N ha^{-1} was added to a barley cropping system than when no fertilizer was added, but there was no additional increase in LF-C with 50 and 75 kg N ha^{-1} additions even though yield responses were observed. Based on data presented by Janzen et al. (1992) and Campbell et al. (1991) for the seemingly C saturated soil at Melfort, we related LF-C to C inputs and found no relationship (Figure 5).

In Figure 6, we present a fractionation scheme that could isolate the four SOM pools forming the basis of our new conceptual model, making it a model with measurable pools (Figure 3). In a first step, coarse non-protected POM, microaggregates, and silt+clay associated C are isolated from 2 mm air-dried sieved soil. These fractions can be isolated with a recently developed method by Six et al. (2000b). The method accomplishes a complete break up of macroaggregates without breaking up microaggregates, which are then separated by sieving. In a second step, fine non-protected POM that was collected together with the microaggregates on the sieve is isolated by density flotation (Six et al., 2000b). Subsequently, microag-

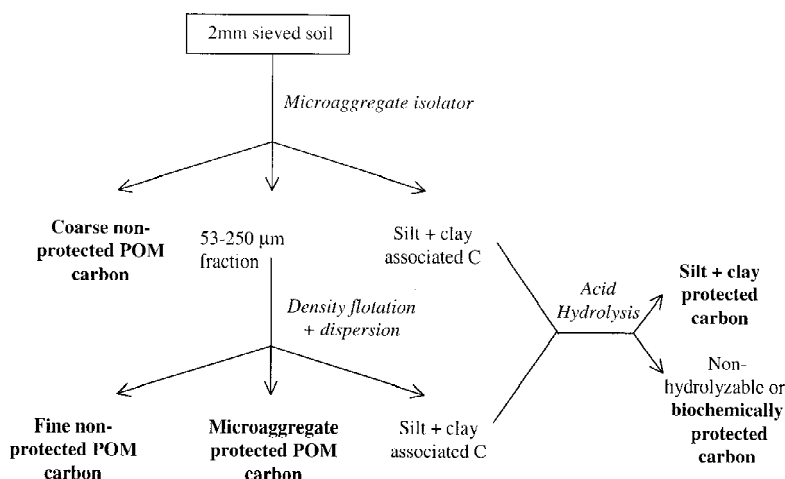


Figure 6. Fractionation scheme to isolate SOM fractions used in conceptual SOM model.

gregates are dispersed to isolate microaggregate protected POM versus silt and clay associated C (Six et al., 2000b). The silt and clay associated C fractions from step 1 and 2 are then hydrolyzed to differentiate the silt+clay protected C versus biochemically protected carbon.

Summary and conclusion

From a synthesis of knowledge on stabilization mechanisms for different soil organic matter pools we have conceptualized a model of SOM dynamics based on four measurable pools: (1) a biochemically-protected C pool, (2) a silt- and clay-protected C pool, (3) a microaggregate-protected C pool, and (4) an unprotected C pool. Each pool has its own dynamics and stabilizing mechanisms, which we hypothesize to determine a level at which soil C becomes saturated.

The silt and clay protected C pool is the C that is protected by association with the mineral particles and is by our definition hydrolyzable. The size of the pool depends on the silt and clay proportion in soil and the relationship between the silt and clay proportion differs between forest and grassland ecosystems. With the same proportion of silt and clay particles, 2:1 clay mineral dominated soils have a greater silt- and clay-protected C pool than 1:1 clay mineral dominated soils. Upon cultivation, silt- and clay-protected C is lost.

The physical protection of SOM exerted by aggregates is indicated by a positive influence of aggregation on the accumulation of SOM. In addition, cultivation has been found to result in a loss of SOM

by breaking up the aggregates. Along with the accumulation of SOM by inclusion in aggregates, a qualitative change in the chemical structure of SOM has been found. The physical protection of SOM exerted by aggregates is predominantly at the microaggregate level. Therefore our second protected SOM pool is the microaggregate-protected C pool.

The biochemically-protected C pool turns over very slowly and is often equated to the non-hydrolyzable fraction. It is the inherent and attained complex chemical composition of the non-hydrolyzable fraction that inhibits decomposition.

The unprotected C pool is isolated as LF or POM not occluded within microaggregates. The N_{min} in soil has been positively related to the N content of the unprotected C and negatively to the C/N ratio of the unprotected C because of immobilization during initial stages of incubation. The unprotected C pool is labile, is an important nutrient source and is very sensitive to management practices.

In order to develop a quantitative model based on our conceptual model the controls on the dynamics and sizes of our proposed pools need to be better quantified. The formation and degradation processes of microaggregates are less understood than that of macroaggregates and minimal quantitative data is available for these processes. Even though we delineated the four pools as conceptually different pools with different stabilization mechanisms because we believe that they are the main stabilization mechanisms for them, there might be some overlap in stabilization mechanisms between the different pools. For example, it is conceivable that the non-hydrolyzable fraction is not only biochemically stabilized but is also partially sta-

bilized by association with clay and silt particles or the silt and clay protected pool is also partially stabilized by incorporation in microaggregates. The relative importance of the different mechanisms should be investigated. One of the major gaps in knowledge and consequently a research priority is the mechanistic explanation of a saturation level, if it exists, for the unprotected and biochemically protected pools.

Acknowledgements

We would like to thank Christian Feller, Sylvie Recous, Roel Merckx, Denis Angers, Alan Franzuebbers and Jan Skjemstad for comments on an earlier version of this manuscript. This research was funded by a U.S. Department of Energy grant: Soil Carbon Saturation: Determining Limits on Carbon Sequestration Capacity (DE-FG03-00ER62997).

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Section editor: R. Merckx

Appendix A. Silt plus clay (0–20 μm) content and silt- and clay-associated C concentration for various soils arranged by land use

Author	Year	Clay type	Silt+clay (%)	Silt+clay C (g C kg soil ⁻¹)
<i>Cultivated</i>				
Balesdent et al.	1998	2:1	57.7	13.8
			62.5	25.8
Bonde et al.	1992	1:1	70.3	16.5
			70.5	15.5
Christensen	1985	2:1	28.3	8.9
			29.0	15.8
			60.4	14.6
			78.4	13.2
			97.6	17.2
Christensen	1986	2:1	36.8	14.9
Christensen	1987	2:1	36.8	13.8
			13.3	21.7
			13.6	22.4
			30.1	10.7
			30.2	11.1
Feller and Beare	1997	1:1	12.5	3.7
			15.2	3.9
			10.1	3.2
			12.3	3.5
			13.3	3.8
			13.6	3.5
			15.6	4.7
			26.9	9.7
			28.4	7.2
			69.4	14.8
			70.1	12.6
			71.6	13.3
			80.5	16.9
			86.9	33.3
			87.1	27.4

Author	Year	Clay type	Silt+clay (%)	Silt+clay C (g C kg soil ⁻¹)
Feller and Beare	1997	2:1	11.7	2.7
			70.5	25.4
Guggenberger et al.	1994	2:1	58.5	24.2
Hassink	1997	2:1	3.6	3.6
			6.2	7.4
Schulten et al.	1993	2:1	21.0	16.3
Turchenek and Oades	1979	2:1	83.0	45.2
Turchenek and Oades	1979	Mixed	46.5	8.4
			80.2	59.1
Zhang	1988	2:1	49.0	14.3
			53.0	19.2
<i>Forest</i>				
Balesdent et al.	1998	2:1	61.4	47.4
Bates	1960	1:1	9.7	7.1
Bonde et al.	1992	1:1	62.9	24.8
Feller and Beare	1997	1:1	11	5.5
			22.7	13.1
			29.4	9.8
			61.4	18.0
			62.9	26.1
			61.4	34.2
Feller and Beare	1997	2:1	12.0	3.9
Guggenberger et al.	1994	2:1	63.2	71.7
			63.7	46.4
McKeague	1971	2:1	43.0	20.8
			43.8	30.3
			60.0	26.1
<i>Grassland</i>				
Amelung et al.	1998	2:1	19.9	7.4
			26.1	7.0
			28.0	7.5
			32.5	12.6
			32.6	9.9
			33.6	9.9
			35.1	27.6
			38.7	15.1
			41.3	22.6
			45.3	18.7
			45.7	16.1
			47.3	23.1
			48.6	27.0
			49.6	22.2
			51.4	39.9
			53.7	19.8
			57.8	12.6
			59.7	24.2
			59.8	29.4

Appendix A. Continued

Author	Year	Clay type	Silt+clay (%)	Silt+clay C (g C kg soil ⁻¹)
Dalal and Mayer	1986a	2:1	64.6	48.2
			77.4	41.8
			52.1	13.2
			55.8	14.8
Dalal and Mayer	1986a	Mixed	65.2	14.0
			27.4	6.7
			82.3	11.5
			86.8	14.0
Elustondo et al.	1991	2:1	15.4	15.7
			16.7	17.9
			34.8	19.9
			43.7	21.3
			61.2	23.7
			63.5	24.1
			65.8	37.6
Feller and Beare	1997	1:1	4.9	1.5
			11.9	5.0
			17.4	6.0
			19.3	7.5
			19.3	7.8
			25.2	9.7
			28.4	12.7
			30.5	10.7
			72	9.8
			82.3	30.1
Feller and Beare	1997	2:1	82.9	36.1
			70.1	32.7
			59.4	19.4
Guggenberger et al.	1995	1:1	56.0	22.3
			56.1	19.6
			56.5	22.6
Guggenberger et al.	1994	2:1	60.1	52.0
Hassink	1997	2:1	1.9	4.1
			2.8	0.9
			3.5	3.6
			3.6	1.0
			4.4	7.5
			5.4	3.0
			5.8	10.2
			8.9	3.6
			13.3	11.1
			35.6	18.6
			36.5	27.5
			42.6	19.9
			45.5	13.5
			76.0	31.0
Leinweber and Reuter	1992	2:1	37.6	14.8
			41.0	8.9
			41.5	10.6
			41.9	7.4

Appendix A. Continued

Author	Year	Clay type	Silt+clay (%)	Silt+clay C (g C kg soil ^{−1})
McKeague	1971	2:1	37.0	22.4
			45.0	34.3
			48.3	43.1
			52.1	47.9
			59.4	50.6
			62.4	47.4
			70.0	37.2
			70.2	70.7
			70.3	48.0
			73.3	28.1
Nacro et al.	1996	1:1	16.0	6.0
Turchenek and Oades	1979	Mixed	50.7	19.7
Zhang et al.	1988	2:1	54.7	28.2
			56.3	32.0
<i>Unknown</i>				
Feller and Beare	1997	1:1	5.2	1.2
			6.5	1.4
			20.8	7.9
			24.2	8.4
			27.4	11.8
			67.8	20.3
			68.5	16.3

Appendix B. Silt plus clay (0–50 μm) content and silt- and clay-associated C concentration for various soils arranged by land use

Author	Year	Clay type	Silt+clay (%)	Silt+clay C (g C kg soil ⁻¹)
<i>Cultivated</i>				
Anderson et al.	1981	2:1	68.0	16.0
			70.5	18.9
Angers et al.	1993	2:1	65.0	23.3
			67.0	29.7
			69.0	29.9
Angers and Ndayegamiye	1991	2:1	73.0	17.1
			73.0	17.1
			73.0	14.0
Balesdent et al.	1998	2:1	88.1	14.1
			91.4	26.3
^a Catroux and Schnitzer	1987	2:1	69.5	26.7

Appendix B. Continued

Author	Year	Clay type	Silt+clay (%)	Silt+clay C (g C kg soil ⁻¹)
Elustondo et al.	1990	2:1	9.0	11.7
			24.4	17.8
			41.0	19.8
			44.4	18.0
			64.6	71.4
			71.6	24.2
			72.3	39.3
Shang and Tiessen	1997	1:1	22.8	7.9
^c Stemmer et al.	1999	2:1	52.2	14.1
			82.1	11.2
Tiessen and Stewart	1983	1:1	68.0	18.9
Tiessen and Stewart	1983	2:1	21.0	9.8
			26.0	21.8
			26.0	15.6
			26.0	13.7
^b Turchenek and Oades	1979	2:1	92.4	46.2
^b Turchenek and Oades	1979	Mixed	80.6	20.7
			96.5	61.9
Young and Spycher	1979	2:1	52.0	7.9
			88.0	8.0
Zhang et al.	1988	2:1	92.3	15.8
			94.0	21.1
<i>Forest</i>				
Balesdent et al.	1998	2:1	87.6	48.2
Bates	1960	1:1	38.4	30.9
McKeague	1971	2:1	61.0	21.3
			63.8	30.6
			65.7	39.9
			75.0	26.3
			90.6	32.9
Shang and Tiessen	1997	1:1	19.7	8.7
Young and Spycher	1979	2:1	18.0	31.6
			77.0	37.6
<i>Grassland</i>				
McKeague	1971	2:1	28.3	43.2
			55.0	23.1
			58.0	36.5
			69.1	48.7
			71.3	43.5
			75.4	53.3
			78.4	47.6
			81.2	71.4
			87.0	38.0
			89.3	48.4
			94.3	28.3
			97.0	56.9
Nacro et al.	1996	1:1	20.3	6.5

Appendix B. Continued

Author	Year	Clay type	Silt+clay (%)	Silt+clay C (g C kg soil ⁻¹)
^c Stemmer et al.	1999	2:1	59.5	11.6
Tiessen and Stewart	1983	1:1	68.0	21.8
Tiessen and Stewart	1983	2:1	21.0	14.1
			26.0	20.5
^b Turchenek and Oades	1979	Mixed	80.6	20.7
Young and Spycher	1979	1:1	73.0	10.6
			87.0	13.5
Zhang et al.	1988	1:1	91.0	36.7
			92.3	32.3
<i>Unknown</i>				
Young and Spycher	1979	Allophanic	71.0	67.5

^aParticles less than 45 μm .^bParticles less than 53 μm .^cParticles less than 63 μm .