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Physical control of soil organic matter dynamics in the tropics

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Abstract

Management of soil organic matter (SOM) is esential to sustaining the quality and productivity of soils around the globe. This appears to be particularly true in the tropics where there is a greater proportion of nutrient poor, highly weathered soils that are more susceptible to losses of SOM. Developing management practices that promote the maintenance and storage of SOM in the tropics depends on understanding the factors that control SOM dynamics. This paper describes the role that soil physical properties (mineralogy, texture, and structure) play in regulating the accumulation and loss of SOM in tropical soils. Two different approaches are presented here. The first approach explores relationships between total SOM and soil physical properties in the tropics. These include effects of climate and mineralogy on latitudinal gradients in SOM, interactions between texture and mineralogy as determinants of SOM storage and relationships between SOM and the structural stability of soils. The second approach describes characteristics of SOM associated with different physical constituents of the soil, with particular attention to particle-size fractions and aggregated particles of different sizes. In each case we summarise findings on the distribution of SOM among fractions and characterise its biochemical composition, bioavailability and turnover. Evidence for and against the physical protection of organic matter from microbial attack in tropical soils is also given. Wherever possible, we compare and contrast the findings for tropical soils with those of temperate soils. The influence of landuse management on physical control of SOM dynamics is discussed as an overriding factor with each aproach. © 1997 Elsevier Science B.V.

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1. Introduction

Soil organic matter (SOM) plays an important role in determining the fertility and productivity of soils. This may be especially true in tropical areas where nutrient poor, highly weathered soils are often managed with few external inputs. Moreover, from a global change perspective, landuse management influences the ability of soils to serve as both a source and a sink of SOM and nutrients. Thus, there is a need to understand more clearly the factors which control SOM dynamics in the tropics and to determine how these factors are influenced by landuse management practices.

In the broadest sense, SOM dynamics are driven by climate, soil type and landuse management. These factors interact to determine the physical, chemical and biological controls on SOM. Together, they regulate the quality and quantity of SOM inputs, the composition and activity of decomposers communities, and rates of SOM loss by mineralization, leaching and erosion.

This paper focuses on physical control of SOM in the tropics, including the effects of soil physical properties (mineralogy, texture, structure) and land management factors, on the accumulation and loss of SOM. Many of our examples will come from well-drained tropical and sub-tropical soils, though reference to studies of other tropical and temperate soils are included for comparison.

Two general approaches will be distinguished: one describing the relationship between total SOM and soil physical properties and a second describing the different forms of SOM and their associations with various physical constituents of the soil, with particular emphasis on particle-size fractions and aggregated particles of different sizes. In each case, we discuss the implications of landuse management for altering these properties.

2. Total soil organic matter in relation to soil physical factors

2.1. Soil mineralogy

While there has been much discussion of, and indeed controversy over, differences in the SOM content of temperate and tropical soils based on gradients of climatological properties (Brown and Lugo, 1982; Post et al., 1982, 1985), increasingly the evidence suggests that when soils of a similar classification (e.g. Order) and treatment history are compared, the overlap in SOM content of soils from temperate and tropical regions is extensive (Sanchez, 1976; Anderson and Swift, 1983). As such, latitudinal gradients in SOM content are probably more closely related to differences in mineralogy. Three primary groups of tropical soils can be distinguished based on mineralogy: (1) the kaolinitic/halloysitic, or low activity clay (LAC) soils (e.g. Ultisols, Oxisols) which cover 60 to 70% of the tropical land areas, (2) the smectitic, or high activity clay (HAC) soils (e.g. Vertisols) and, (3) the more or less allophanic (ALL) soils (e.g. Andisols) with large amounts of amorphous or crypto-crystallized minerals.

The relationship between the soil organic carbon (SOC) content and clay + silt content (0–20 μm; % of total) in surface horizons (0–20 cm) of LAC, HAC and ALL soils is shown in Fig. 1 based on a collection of studies in West Africa, Antilles and Brazil (Feller, 1995a). In these studies, SOC concentrations did not differ widely between the LAC and HAC soils, however, ALL soils exhibited higher SOC levels under similar climatic and landuse management conditions (Albrecht et al., 1992a).

The large humus accumulation in Andisols is well known (Quantin, 1972; Wada, 1985) and mainly attributed to very stable humus–Al, Fe-complexes which may be protected from bacteria and enzymes in microaggregates rather than by a specific effect of allophane and associated minerals (Wada, 1985; Boudot et al., 1986; Oades et al., 1989). However, other studies indicate that SOM in ALL soils is highly decomposed (e.g. Calderoni and Schnitzer, 1984; Hatcher, 1989). In these soils, SOM is rich in carboxylic and aromatic groups, giving them a very high affinity to Al/Fe-oxides/hydroxides. As the clay + silt content of ALL soils is difficult to determine (Colmet-Daage et al., 1972) and often not significant in terms of andic properties, there is a need to establish correlations between SOC content and other physical parameters in these soils. Recent studies (Egashira et al., 1986) have investigated the use of water content at field capacity as a good indicator of the degree of andic character. On this basis, the SOC content of ALL soil was plotted against the water content at

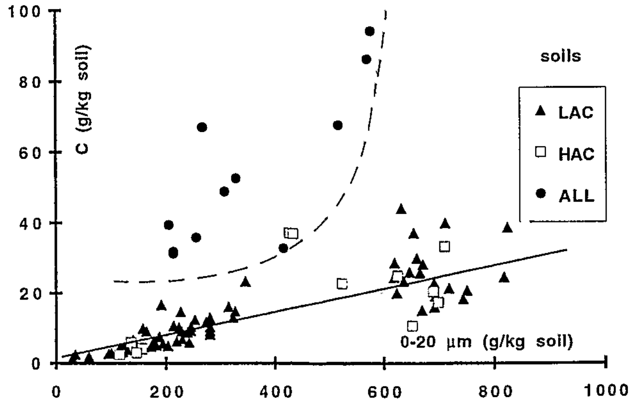

Fig. 1. Carbon content (g C/kg soil) of soils (0–20 cm) according to their mineralogy and texture (clay+silt content [$c+s<20$ μm], g/kg soil). LAC = low activity clay soils (kaolinitic), HAC = high activity clay soils (smectitic), ALL = allophanic soils. The regression equation for LAC soils is: $C = 0.037(c+s) + 0.69$ ($n = 65; r = 0.87; p < 0.001$).
The curved line delimits the domain of ALL soils (from Feller, 1995a).

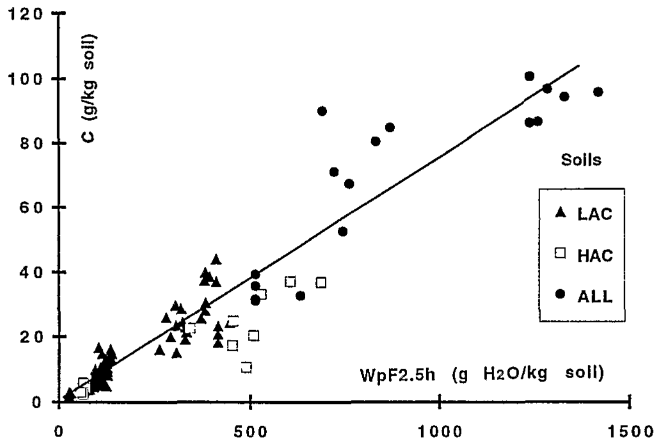

Fig. 2. Relationship between carbon content (g C/kg soil) and water content (g H_2O/kg soil) at pF2.5 of field moist soils can be described by the equation:
$C = 0.074(pF2.5) + 0.14\ (n = 95; r = 0.95; p < 0.001)$
LAC = low activity clay soils (kaolinitic), HAC = high activity clay soils (smectitic), ALL = allophanic soils. (Feller, unpubl.).

pF2.5 of the non-air dried sample (Fig. 2). These results indicate a strong relationship between the two parameters and a similar equation can be used for both LAC and ALL soils (Feller, unpubl.). The reason for this relationship is poorly understood, though it may be related to a distinct pore-size range which influences the accessibility of microorganisms to SOM. It would be useful to follow up this approach with Andisols of other tropical areas. Furthermore, it is clear that developing a fuller understanding of the clay–organic matter complex will contribute greatly to identifying those factors which most strongly influence organic matter storage (Goh, 1980; Tate and Theng, 1980).

2.2. Soil texture

Numerous authors have described the relationship between clay (or clay + silt) content and SOM in LAC soils (Martin, 1963; Perraud, 1971; Boissezon, 1973; Jones, 1973; Turenne, 1977; Talineau et al., 1980; Lepsch et al., 1982; Lepsch et al., 1994; Feller et al., 1991a; Ferry, 1992) from different sites in the tropics (West and Central Africa, Antilles, Brazil, South India). These studies have shown that clay (or clay + silt) content is a relatively important determinant of SOM levels in LAC soils. This relationship (Feller et al., 1991a) appears to hold equally well for cultivated soils as for those under native vegetation (Fig. 3) in situations covering a wide range of mean annual precipitation (600–3000 mm yr^{-1}). Therefore, the low organic matter content of sandy West African soils

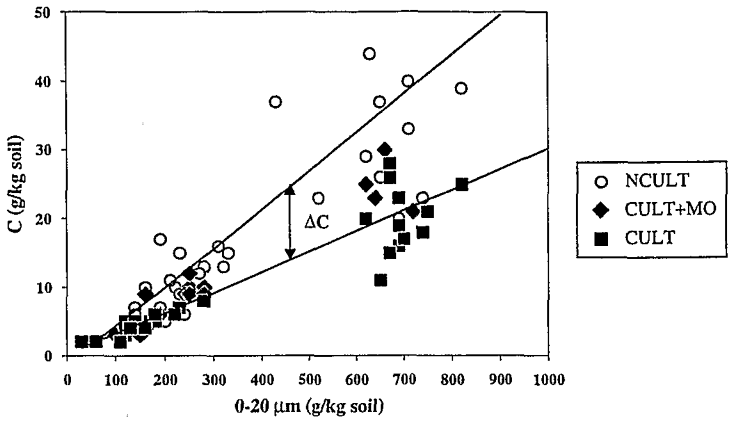

Fig. 3. Relationships between carbon content (g C/kg soil) and clay + silt content ($c + s \leq 20$ µm, g/kg soil) of LAC and HAC soils (0–20 cm) from different management histories; uncultivated (NCULT), cultivated (annual cropping) with organic inputs or temporary vegetated fallows (CULT + MO) and continuous cultivated (CULT) (Feller, 1995a). The regression equation for the LAC soils are:

NCULT situations $C = 0.049(c + s) - 0.23(n = 25; r = 0.91; p < 0.001)$

CULT + MO situations $C = 0.038(c + s) + 0.02(n = 16; r = 0.96; p < 0.001)$

CULT situations $C = 0.029(c + s) + 0.32(n = 25; r = 0.95; p < 0.001)$

ΔC represents the maximum variation between NCULT and CULT situations.

may be attributed to their low clay content (< 20%) in addition to management associated factors such as overgrazing and intensive farming.

Few data are yet available for high activity clay (HAC) soils (smectites) in the tropics. Of 69 Vertisols from the West African savanna (mostly in Nigeria), Jones (1973) found that SOM was negatively correlated to clay content in those (52) containing > 35% clay. In contrast, Yerima et al. (1989) reported a significant positive relationship between SOC and clay content for some Vertisols of Northern Cameroon. The negative relationship noted by Jones (1973) may be due to a strong influence of sheet erosion and/or the soils vertic properties resulting in contamination of surface horizons with soil from deeper horizons. Additional data will be required before any general conclusions can be drawn on the effect of soil texture on SOM storage in tropical smectitic soils.

2.3. Soil structure

The formation of stable soil aggregates is influenced by mineralogy, texture, landuse management and the quality and quantity of organic matter inputs. These factors interact to determine the now well established relationship between SOM content and water-stable aggregation. This relationship is probably best known for temperate agricultural soils (Douglas and Goss, 1982; Chaney

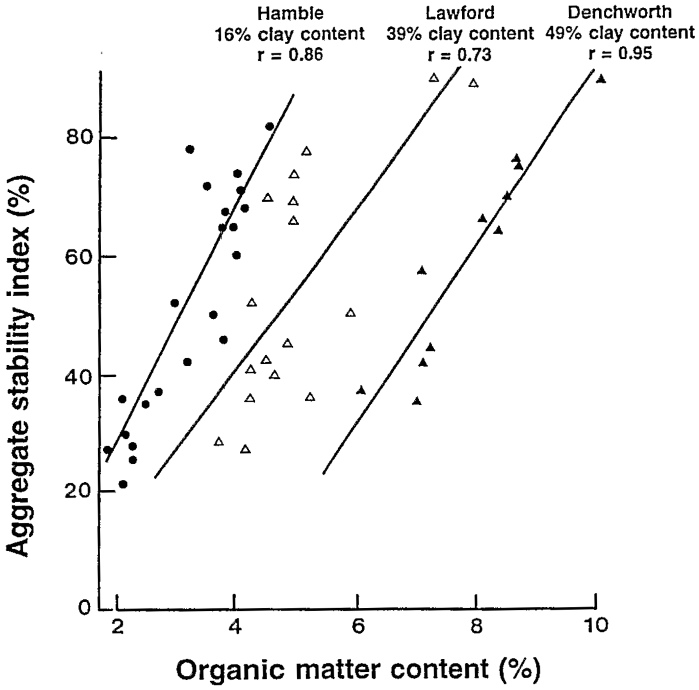

Fig. 4. Relationship between aggregate stability and organic matter content for samples of the Hamble, Lawford and Denchworth soil series. From Haynes and Beare (1995), redrawn from Douglas and Goss (1982).

and Swift, 1984; Bartoli et al., 1988) and is highly dependent on clay content. This fact is shown clearly in the studies of Douglas and Goss (1982) where increasingly higher quantities of C were required to achieve the same level of aggregate stability in soils of increasing clay content (16–49% clay) (Fig. 4).

Many studies describing relationships between SOC content and aggregate stability in tropical soils are based on the use of Henin's test. Henin et al. (1960) described a structural instability index (IS) given by the formula:

$$\mathrm{IS} = \frac{(\mathrm{A} + \mathrm{LF})\,\mathrm{max}\%}{1/3\,\mathrm{Ag}\% - 0.9\,\mathrm{SG}\%} \qquad (1)$$

where (A + LF) max% represents the maximum amount of dispersed 0–20 μm fraction obtained after three treatments of the initial soil sample: without pretreatment (air) and with immersion in alcohol or in benzene. Ag% refers to the > 200 μm aggregates (air, alcohol, benzene) obtained after shaking, i.e. 30 manual turnings and sieving under water of the 3 pre-treated samples. SG% represents the content of coarse mineral sand (> 200 μm). The denominator of the formula (1/3 Ag%–0.9 SG%) is an estimation of the 'mean percent stable aggregates'. The instability index has been shown to be a good index (among 14 others) of detachability (Henin et al., 1960; De Vleeschauwer et al., 1979) in tropical soils.

Combeau (1960), Combeau et al. (1961), Martin (1963), Thomann (1963) and Feller et al. (1996) have applied the Henin index to soils of West and Central Africa, the Caribbean and Brazil and reported positive relationships between SOC content and aggregate stability. Similar conclusions based on water-stable aggregate tests were given by Goldberg et al. (1988) for 34 arid-zone soils from California, by Alegre and Cassel (1986) for a Peruvian Ultisol, by Dutartre et al. (1993) for sandy Alfisols, and by Arias and De Battista (1984) for Uruguayan Vertisols. The importance, however, of other variables (e.g. clay content) for determining this relationship in tropical soils is much less clear. For example, using multiple regression analyses, Goldberg et al. (1988) found that organic C, quartz and Fe content (%w/w) contributed most significantly to explaining the variability in aggregate stability (%) in the Californian soils. Quartz and montmorillonite content and particle surface area, but not clay content, were significant variables for maximum R^2 improvement. In contrast, Dutartre et al. (1993) found a positive relationship between organic C (%) and water-stable aggregation ($R^2 = 0.66$, $p < 0.1$) for twenty sandy soils of Burkina-Faso and Mali, but little or no influence of clay content.

Diverse families of organic compounds have been shown to affect soil aggregation. These include humic (humin and humic and fulvic acids) compounds (Combeau, 1960; Martin, 1963; Thomann, 1963; Chaney and Swift, 1984; Piccolo and Mbagwu, 1990; Dutartre et al., 1993), compounds extractable by polar or non-polar organic solvents (Giovannini and Sequi, 1976a,b; Hamblin and Greenland, 1977; Capriel et al., 1990; Wierzchos et al., 1992) and, most commonly, polysaccharides (Cheshire, 1979). In relation to the chemical structure of SOM, Monreal et al. (1995) showed for a cultivated Chernozem that the proportion of water-stable aggregates was highly correlated to concentrations of lignin dimers, sterols, alkylaromatics and lipids. Despite these relationships, many studies have shown higher correlations to aggregate stability using total SOC rather than individual SOM constituents.

There is little doubt that polysaccharides act as glues inside soil aggregates (see reviews by Cheshire, 1979 and Tisdall and Oades, 1982) even if their relative importance in comparison with other compounds (see above) may be questioned (Hamblin and Greenland, 1977). As the correlations between total C content or total carbohydrate content and aggregate stability are not necessarily significant on a short time scale (Baldock et al., 1987) recent studies have focused on the nature and origin of polysaccharide fractions as they related to aggregate stability. For example, Haynes and Swift (1990) and Haynes et al. (1991) showed for a Cambisol and Inceptisol from New Zealand, and Gisman and Thomas (1995) for Oxisols in Columbia, that aggregate stability was more closely correlated with the content of hot water-extractable carbohydrates than with total organic carbon (TOC), hydrolysable (HCl) carbohydrates or NaOH-extractable carbohydrates. They concluded that hot-water-extractable carbohydrates represent an important fraction for the stabilization of aggregates in these temperate soils. In contrast, Degens et al. (1994) found no significant correlation between aggregate stability and hot-water-extractable carbohydrates in Western Australia. Concerning the origin of polysaccharides, Cheshire et al. (1983, 1984) found that polysaccharides of microbial origin were more important than plant-derived polysaccharides for determining aggregate stability. This agrees with the results of Sparling and Cheshire (1985) who observed that the influence of polysaccharides on aggregate stability was less important in rhizosphere soils where a larger proportion of the polysaccharides are in the form of plant remains as compared to non-rhizosphere soil. Other studies (Benzing-Purdie and Nikiforuk, 1989) have stressed the importance of the polysaccharides of plant root origin. Explanations for differences concerning the role that plant derived polysaccharides play in aggregate stability may be found in the fact that they may originate from plant detritus or from plant exudates, two sources of SOM not distinguished in these studies and that they are likely to have different influences on soil aggregation.

Improvements in soil aggregation can have implications for a number of soil physical and chemical properties including water infiltration rates, oxygen supply and organic matter mineralization rates. In the latter case, stable aggregates may enhance the physical protection of SOM against losses due either to mineralization or detachability and erosion. The role of soil aggregates in the storage and protection of SOM will be discussed below.

2.4. Soil organic matter losses by erosion

The storage of SOC in tropical areas may be enhanced by reducing C losses due to erosion (particularly sheet erosion). These may be quantified by comparing the C content of eroded particles (E_c) to that of surface soil (S_c) layers (i.e. E_c/S_c ratio). In the West African LAC soils studied by Roose (1977), E_c/S_c ratios ranged between 1.5 to 12.8 indicating a relatively high potential for losses of SOC by sheet erosion.

Soil erodibility generally decreases with increases in SOM content and aggregate stability. For instance, De Vleeschauwer et al. (1979) showed that soil erosion (measured by rainfall simulator) and particle detachment were both influenced by SOC content. Similarly, Albrecht et al. (1992b) have shown for a Vertisol in Martinique that the turbidity of runoff water and the SOC exported are highly dependent on the SOC content of the 0–5 cm layer. They proposed a threshold value of about 20 mg C/g soil (0–5 cm) as an index of erosion risks for this soil. A similar relationship, but lower threshold value, was obtained for a clayey LAC soil from Martinique. These results suggest a complex interaction between SOC storage and aggregate stability, i.e. SOC plays a major role in the stabilization of aggregates but aggregation may also reinforce SOC storage by lowering the SOC losses resulting from sheet erosion.

2.5. *Landuse management and soil organic matter (SOM) losses*

Landuse and cultural practices influence SOM storage and loss in a variety of ways. They may alter the organic matter (OM) input into soils, the soil biological activity, the bioavailability of organic substrates and soil erodibility. In general, it is difficult to distinguish the contributions of these different factors; therefore much of our focus here is on the overall net effect of soil management on SOM dynamics. For example, this relationship has been illustrated for a collection of tropical LAC soils (Feller et al., 1991a; Feller, 1993) selected for their low incidence of erosion and covering a wide range of soil texture. As was shown previously (Fig. 3), SOC content depends on landuse management and becomes more important with increases in clay + silt content. SOC content under annual crops (in non-eroded plots) was approximately 60% of that under natural vegetation. The new equilibrium is generally reached in 3 to 5 years for coarse-textured soils and in 5 to 10 years for fine-textured ones. Vegetated fallow soils and those receiving organic amendments (Fig. 3, CULT + MO) have SOC contents intermediate to those of natural vegetation and continuously cropped soils. After a long period of continuous cultivation, artificial pastures were efficient in restoring SOC levels of clayey soils within about 5 years (Albrecht et al., 1992a). From these results and those reported by Pieri (1989) on the fertility of soils of the semi-arid tropics, Feller (1995b) proposed the following equation:

$$\mathrm{C}‰ = 0.32\ (\%\mathrm{silt} + \mathrm{clay}\ [0 - 20\mu\mathrm{m}]) + 0.87 \tag{2}$$

might represent a critical threshold of SOC content (horizon 0–10 cm) and, therefore, would be useful for predicting the sustainability (sustained fertility and productivity) of annual cropping systems in that area.

In many temperate soils, minimum and no-tillage (NT) practices tend to support higher standing stocks of SOM and greater soil aggregation than conventionally ploughed soils (e.g. Weill et al., 1989; Havlin et al., 1990). The exceptions to these cases are well known (Hamblin, 1980; Carter and Rennie, 1982). However, the effects of long-term reduced-tillage management on the maintenance of SOM levels in tropical and sub-tropical soils is poorly known. Recent studies of a subtropical Ultisol in the southeastern United States (Beare et al., 1994a,b) indicate that NT management practices may provide an important mechanism for the physical protection and maintenance of SOM that may otherwise be lost under conventional tillage (CT) practices. In these studies, Beare et al. (1994a) showed that for similar levels of plant productivity and, hence, similar levels of organic inputs, the SOC content of NT surface soils (0–5 cm) was 18% higher than that of CT after 11 years of continuous

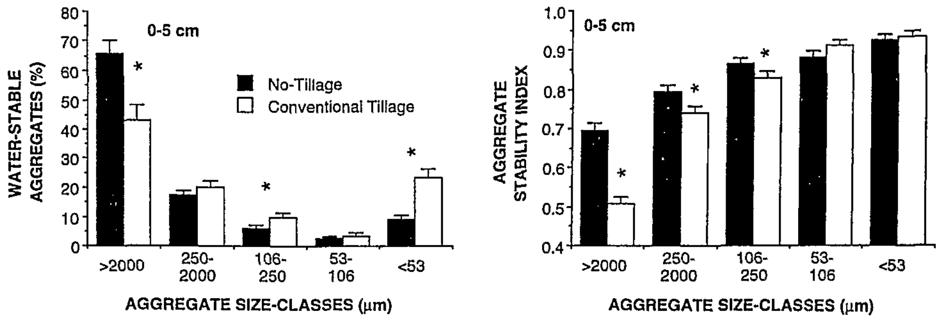

Fig. 5. Distributions and stabilities of intact water-stable aggregates from a subtropical Ultisol under conventional- (CT) and no-tillage (NT) management. Asterisks indicate significant differences ($P < 0.05$) between tillages within size class (adapted from Beare et al., 1994a).

treatment, while no difference was observed for the deeper horizons. These differences in SOM content were closely linked to the quantity and stability of macroaggregates (> 250 μm) which were much greater in NT than CT soils at this site (Fig. 5).

3. Fractionation of soil organic matter in relation to soil physical constituents

For over a century, soil organic matter (SOM) has been characterized by chemical fractionation (acid/alkaline extraction, acidic or alkaline hydrolysis), with some interest in pedogenesis research but with few significant applications in the field of agropedology. There are two possible reasons for this: (1) humic and fulvic acids generally have a low turnover rate (Anderson and Paul, 1984; Duxbury et al., 1989) and are, therefore, not necessarily implicated in the short-term processes (from days to decades) commonly studied in cultivated soils and (2) the importance of these chemical fractions to major soil processes such as aggregation and OM mineralization are not yet well established. This probably results from the fact that acid/alkaline extractions of organic matter are highly selective and linked to the solubility of the organic compounds, resulting in the extraction of similar compounds from completely different SOM pools. As a result, over the past two decades, much more of the research on SOM characterization in agricultural soils has involved physical fractionation techniques.

The association of organic matter with particular constituents of the mineral soil may be important to regulating the mineralization and storage of SOM. As indicated above, SOM dynamics are correlated with and presumably dependent upon soil texture and structure, the properties of which may strongly influence SOM decomposition processes. The following sections will review the relationships between SOM and soil physical constituents at two levels of the organization: (1) the nature of OM associated with primary mineral particles (sand, silt, clay) of the soil; i.e. the particle-size fractionation (PSF) approach allowing a characterization of SOM complexes by size and their relationship to mineral particles; (2) the nature of OM associated with macro- and microaggregates, defined here as the stable aggregate fractionation approach (SAF), allowing an investigation of how soil structural properties may influence the size, quality and distribution of SOM pools.

It should be noted, however, that these two approaches are closely linked at the micrometric scale, i.e. 0–2 or 0–5 μm separates obtained by PSF are generally not dispersed particles but rather microaggregates, and will be referred to here as 'primary microaggregates' (sensu Christensen, 1992, 1996). Recent advances in SOM research using density fractionation techniques are described elsewhere (Elliott and Cambardella, 1991; Christensen, 1992; Gregorich and Ellert, 1993; Gregorich and Janzen, 1996).

3.1. The particle-size fractionation approach (PSF)

3.1.1. Objectives and methodological constraints

When particle-size fractionation is used for the characterization of OM associated with primary mineral particles, the major methodological objective is to achieve maximum dispersion of the soil (separation of sand, silt and clay fractions) with a minimum alteration (solubilization and/or transformation) of the associated OM. At present, in most studies (Elliott and Cambardella, 1991; Christensen, 1992) dispersion is obtained by ultrasonic treatment of the bulk soil in water at a constant energy level for a fixed time interval. Depending on the individual soil properties, there is usually a compromise between the energy input required to achieve soil dispersion and the risks of altering the organic or organo-mineral components. According to Christensen (1992) there appears to be a limited effect of ultrasonication on the mineral constituents of the soil or on the detachment of OM from organo-mineral complexes. He also notes, however, that "The intensity of ultrasonic treatments required to disperse soils completely most probably causes a substantial release of microbial SOM components, which may subsequently be redistributed among size separates and the water-soluable fraction". Another possible artifact of ultrasonication demonstrated by Balesdent et al. (1991) is the breakdown of soil plant debris (fraction > 50 μm) and the artificial transfer of C (about 50% under standard conditions) from the coarse (sand > 50 μm) to the fine (clay + silt < 20 μm) fraction. In view of these results, ultrasonication is not recommended for bulk soil samples (0–2 mm), rather it is more appropriately applied to the 0–50 μm soil–water suspension. Under these conditions, the complete dispersion of > 50 μm aggregates can be obtained by shaking the soil with water and glass beads or, if required, by the use of chemical dispersants such as sodium hexametaphosphate

(HMP) or sodic resin (Na-R) (Feller et al., 1991b; Gavinelli et al., 1995). These procedures generally solubilize less than 5% of total SOC. Na-R is recommended in studies of P fractionation where HMP is a significant source of P contamination.

Assuming limited transfer of C to smaller size-classes, the organic matter found in the silt- and clay-size fractions has a strong association with the mineral particles, i.e. the organic matter generally forms a complex with the mineral particles. This is not the case for the sand-size fraction, the association between sand and sand-size organic matter (particulate organic matter) being largely a function of size.

3.1.2. Physical and biochemical characterization of particle-size fractions

The majority of the information available on organic matter associations with particle-size fractions are from 2:1 clay-based warm to cool temperate soils. Much of this research has focused on the biochemical characterization of organic matter associated with particle-size fractions (Christensen, 1992). A general review of the findings for temperate soils is given here as a baseline against which to compare the somewhat more limited results from tropical soils.

Studies of mostly temperate soils have shown that the majority of soil organic matter is associated with silt- and clay-size fractions (Hinds and Lowe, 1980; Anderson et al., 1981a; Christensen, 1992). Values normally range from 10 to 30% of total soil C in the sand-size fraction (> 50 μm), 20 to 40% in the silt-size fraction (2–50 μm) and 35 to 70% in the clay-size (0–20 μm) fraction. In general, the fine-clay fraction is found to contain less stable organic matter associations than the somewhat coarser fine silt- and coarse clay-size fractions (Tiessen and Stewart, 1983). This fact appears to be related to the presence of more aliphatic, less aromatic, more easily hydrolysable and generally less stable humic acids in the fine clay fraction than in the coarser fractions (Anderson et al., 1981b; Catroux and Schnitzer, 1987; Oades et al., 1987). These observations are supported by natural abundance ^{15}N data which indicate the presence of more decomposed, stable organic matter in fine silt- and coarse clay-size fractions than that associated with fine-clays (Tiessen et al., 1984; Catroux and Schnitzer, 1987).

In contrast to the above observations, most studies show that the C/N ratio of SOM in size separates tends to narrow with decreasing particle-size (Chichester, 1969; Turchenek and Oades, 1979; Tiessen and Stewart, 1983; Zhang et al., 1988; and others reviewed by Christensen, 1992), suggesting that the SOM associated with finer fractions is more highly processed. The largest size fractions are dominated by coarse particulate organic matter, also called macro-organic matter (Gregorich and Ellert, 1993), which often has a very high C/N ratio. In most studies of temperate soils, amino acid-N and amino sugar-N show little variation across size separates, though some studies show that amino acid-N makes up a higher percentage of the total N in clay- and silt-size fractions as compared to values for sand (Christensen, 1992). The SOM associated with clay-size separates tends to be enriched in hydrolysable carbohydrates relative to silt- and sand-size fractions (Angers and Mehuys, 1990; Cheshire et al., 1990). Clay-size separates also tend to have higher galactose + mannose to arabinose + xylose ratios (GM/AX ratio), indicating a somewhat greater contribution of microbes than plants to the total carbohydrate pool in this fraction relative to larger fractions. Consistent with the above findings, clay-size separates have much higher concentrations of diaminopimelic acid, an organic constituent confined to the prokaryotic cell walls, than silt- and sand-size fractions (Christensen and Bech-Andersen, 1989).

Some examples for tropical soils come from studies in West Africa (Senegal, Togo, Ivory Coast), Lesser Antilles (Martinique, Guadeloupe, St. Lucie) and Brazil. In these studies, approximately 40 samples from surface horizons (0–10 cm) of LAC and HAC soils covering a wide range of soil texture and management practices (Fig. 6) were analysed for SOM fractions (Feller et al., 1991a). As recommended above, the particle-size fractionation method (Feller et al., 1991b) involved dispersion of the soil sample in water in the presence of a cationic resin (in sodic form) followed by ultrasonic treatment of the 0–50 μm soil–water suspension to improve the separation between clay and silt fractions.

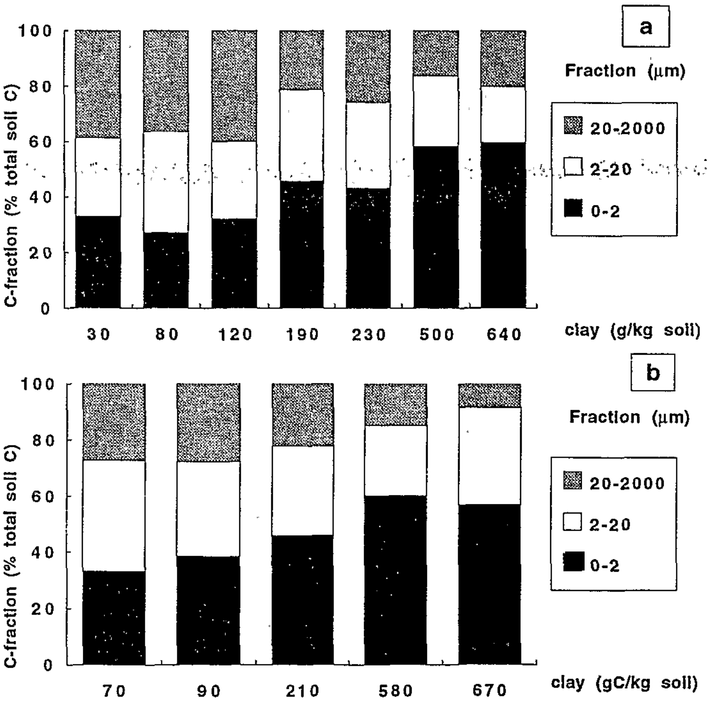

Fig. 6. Relative distribution of carbon (% total soil C) in particle-size fractions in relation to the clay content (g/kg soil) of different tropical LAC soils. (a) Uncultivated soils, (b) cultivated soils.

This method has the advantage (Balesdent et al., 1991) of reducing the risk of an artificial transfer of C from plant debris (coarse fractions > 50 μm) to fine fractions (< 20 μm) as may occur when the ultrasonic energy is applied to a whole soil sample using conventional methods.

According to Feller et al. (1991a,c) and Feller (1993) the results may be characterized as follows. The > 20 μm fraction was composed primarily of 'plant debris' at different stages of decomposition; having carbon-to-nitrogen (C/N) ratios ranging between 12 to 23 ($\bar{x} = \sim 20$). The 2–20 μm fraction, termed the 'organo-silt complex' was composed of highly humified plant and fungal debris associated with very stable organo-mineral microaggregates which are not easily disrupted during fractionation. Their C/N ratios vary from 10 to 21 ($\bar{x} = \sim 15$). The < 2 μm fraction, termed the 'organo-clay complex' had a predominance of amorphous OM acting to cement the clay matrix. Under forest or savanna, the presence of plant cell walls in the coarse clay fraction is sometimes detected but almost never in the fine clay fraction. Very often bacterial cells or colonies at different stages of decay can be observed in both fractions. Their C/N ratios vary from 7 to 12 ($\bar{x} = \sim 10$).

The carbohydrate composition of SOM in different size-fractions was determined (Feller et al., 1991c; Feller, 1995a) for two clayey LAC soils (Martinique and Brazil) and one Vertisol (Martinique). Consistent with expectation, the GM/AX ratio tends to increase from the coarsest to the finest fractions (Table 1) suggesting a greater contribution of plant-derived carbohydrates in the coarse sand-size fractions and of microbially-derived carbohydrates in the fine clay-size fractions. Similar observations were made by Guggenberger et al. (1995) for Oxisols under native savanna and pasture in Columbia. In contrast, the C/N ratio decreases markedly with decreases in particle size. The 'light fractions' ($d < 2.06$) also exhibit lower C/N ratios than the corresponding 'heavy fractions' ($d > 2.06$) (Whitehead et al., 1975; Murayama et al., 1979). Similar relationships between particle size and either the GM/AX or C/N ratio have been reported for temperate soils (Table 1).

In those soils studied by Feller et al. (1991a), the relative C and N distribution (% of whole soil C or N) in the different size fractions was largely dependent on soil texture regardless of cultivation history (Fig. 6). Although a similar pattern has been reported by Christensen (1992) for a wide range of Danish arable soils, the nature of the relationship appears to differ from that of the tropical soils studied by Feller (Fig. 7). In general, the clay-size fraction of the Danish arable soils (taken here as temperate soil examples) retain a much higher percentage of the whole soil C than the tropical soils and the differences tend to increase with increases in soil clay content. In contrast, the sand-size fraction of tropical soils retained a much higher percentage of the whole soil C than the temperate soils; carbon in the sand-size fraction decreased with increases in the clay content. There are several possible explanations for this result. Firstly, samples from the Danish and tropical soils do not correspond to the same layer, the Danish soils

Table 1
The galactose + mannose/arabinose + xylose (GM/AX) and C/N ratios of various soil particle-size (μm) or density (d) fractions reported in the literature

Reference	Site or treatment	Fraction (μm)	GM/AX	C/N
Whitehead et al. (1975)	1	$d < 2.06$	0.3	19.3
		$d < 2.06$	1.1	
	2	$d < 2.06$	0.5	18.5
		$d < 2.06$	1.0	
	3	$d < 2.06$	0.5	12.9
		$d < 2.06$	1.0	
Turchenek and Oades (1979)	Ur WFUr PPMelfort and Furner [b]	20–50	1.9	12.8
		5–20	1.7	14.1
		2–5	2.5	13.2
		0.4–2.0	3.5	11.0
		0.1–0.4	2.2	8.0
		0–0.1	2.4	9.8
Cheshire and Mundie (1981)	Countesswell	250–2000	0.36	
		20–250	0.66	
		2–20	1.14	
		0–2	1.71	
Murayama et al. (1979)	Countesswell	$d < 2.0$	0.7	
		$d < 2.0$ [a]	1.0 to 1.6	
Angers and Mehuys (1990)	Barley	50–2000	0.7	14.7
		2–50	1.4	18.3
		0–2	1.7	12.1
	Alfafa	50–2000	0.6	17.0
		2–50	1.4	16.6
		0–2	1.5	11.8
	Fallow	50–2000	0.7	16.6
		2–50	1.4	17.0
		0–2	1.7	11.3
Angers and N'Dayegamiye (1991)	0 t/ha (manure)	50–2000	0.47	17.6
		2–50	1.67	14.6
		0–2	1.96	11.5
	40 t/ha (manure)	50–2000	0.57	16.1
		2–50	1.76	13.7
		0–2	2.09	12.3
	80 t/ha (manure)	50–2000	0.56	17.8
		2–50	1.67	16.8
		0–2	2.04	12.3
Feller et al. (1991c); Feller (1995a)	Martinique (LAC) (sugarcane)	200–2000	0.4	25.8
		50–200	0.5	20.5
		20–50	0.8	16.4
		2–20	1.2	12.7
		0.2–2.0	1.7	9.3
		0–0.2	1.6	8.9
	Brazil (LAC) (sugarcane)	20–2000	0.4	23.2
		2–20	0.6	14.9
		0–2	0.6	11.5
	Martlinque (HAC) (pasture)	20–2000	0.2	25.9
		2–20	0.4	20.0
		0–2	1.4	8.7
Guggenberger et al. (1995)	Savanna	20–2000	0.4	26.7
		2–20	1.3	23.4
		0–2	2.1	14.5
	Pastures (avg. 2 sites)	20–2000	0.3	23.1
		2–20	0.9	20.4
		0–2	2.2	12.5
	Earthworm casts (pasture)	20–2000	0.5	22.2
		2–20	1.2	16.9
		0–2	1.9	12.4

[a] Various chemical fractions.
[b] Various particle size fractions of the four soil samples.

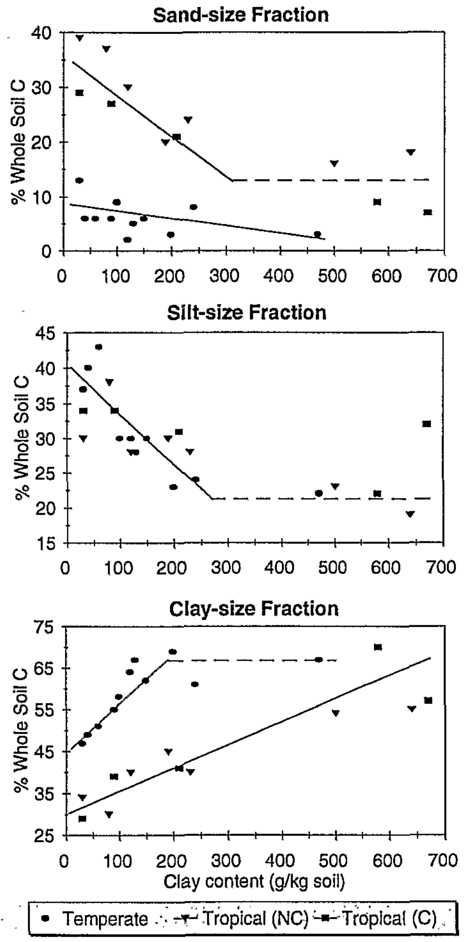

Fig. 7. Relationships between the distribution of C in particle-size fractions and the clay content of cultivated temperate soils (plotted from Christensen, 1992) and cultivated (C) and non-cultivated (NC) tropical soils (adapted from Feller, 1995a).

represent the Ap horizon (about 0–30 cm layer) while the tropical soils were taken to a depth of 10 cm. It is likely that soil from the 0–10 cm layer is relatively more enriched in sand-size C and less enriched in clay-size C as compared to the 0–30 cm layer. Secondly, difference in the distribution and growth pattern of roots in temperate and tropical soils may explain difference in the enrichment of the sand-size fraction. Alternatively, the differences may be explained by methodology; that is: (1) the ultrasonic dispersion method used by Christensen (1992) may involve greater physical fragmentation of particulate OM than that used by Feller et al. (1991a) resulting in a transfer of C from the sand- to clay-size fraction (Balesdent et al., 1991) or (2) the sand-size separates studied by Feller et al. (1991b) were contaminated by silt and clay-size fractions following the incomplete dispersion of sand-size aggregates. However, the efficiency of sodic resin dispersion was confirmed by microscopy and weight differentials determined before and after SOM oxidation with H_2O_2. Finally, the differences may be attributed to weaker aggregate structure and clay–OM bonding in tropical soils, leaving a greater proportion of clay-size organic matter susceptible to oxidation and, therefore, a greater proportion of the total SOM in the sand-size fraction. These hypotheses deserve greater investigation.

Notably, the relatively high proportion of C associated with the sand-size fraction in the studies by Feller et al. (1991a) is consistent with a number of other studies from tropical soils. For example, in his study of a weakly leached ferruginous soil (3.6–4.5% clay) from Senegal, Bernhard-Reversat (1981) found that the sand-size fraction (obtained by differential sieving of soil under distilled water) contained from 29–64% ($\bar{x} = \sim 49\%$) of the total soil C. Similarly, Cameron and Posner (1979) reported that 32 to 40% of the total soil C in a Merredin sandy loam from Australia was associated with the sand-size fraction (obtained by vigorous hand-stirring in water and sedimentation). Further, Desjardins et al. (1994) showed for a medium textured Typic Kandiudult from Brazil that the sand-size fraction contained 35 and 52% of the total soil C in a native forest soil and an adjoining 10 year old pasture, respectively. The reasons for these differences from the temperate soil examples provided by Christensen (1992) remain unclear and deserve further study.

3.1.3. Dynamics of soil organic matter (SOM) associated with particle-size fractions

Soil organic matter (SOM) dynamics can be studied at different temporal scales (seasonal or pluriannual variations), using different approaches (laboratory or field experiments, with or without utilization of isotope techniques, etc.). In this section we will distinguish three types of studies: (1) those of potentially mineralizable C (PMC) and N (PMN) on daily to monthly scale based on laboratory incubation experiments; (2) those of soil C dynamics at the pluriannual scale using investigations of ^{13}C natural abundance of SOC: the 'in situ ^{13}C approach' and (3) those describing the effects of the soil management practices on pluriannual variation in SOM associated with particle-size fractions.

3.1.3.1. Potentially mineralizable organic matter in particle-size fractions. Although much of the information on the mineralization of SOM associated with particle-size (or density) fractions comes from temperate soils (i.e. Chichester, 1969; Cameron and Posner, 1979; Lowe and Hinds, 1983; Sollins et al., 1984; Catroux and Schnitzer, 1987; Christensen, 1987; Gregorich et al., 1989; Hassink, 1995) some data are also available for tropical situations (Bernhard-Reversat, 1981, 1987, 1988; Sollins et al., 1984; Feller, 1993).

The bioavailability of SOM associated with different particle-size fractions can be estimated by quantifying the PMC and PMN in each fraction (often

Table 2
Potentially mineralizable carbon (PMC) and nitrogen (PMN) in different particle-size fractions of temperate and tropical soils, adapted from the literature

Site no. and source	PMC (%)			PMN (%)		
	Sand (50–2000 μm)	Silt (2–50 μm)	Clay (0–2 μm)	Sand (50–2000 μm)	Silt (2–50 μm)	Clay (0–2 μm)
Temperate situations						
1				7.4	5.4–5.0	13.9–32.9
2					4.5–17.4	5.4–25.4
3				0.4–2.6	0.6–3.6	5.3–7.1
4				9.0–8.0	2.4–4.5	7.8–11.0
5	0.9–6.5 [a]	0.5–1.4 [a]	0.9–2.0			
6	4.9	1.5	0.7			
7 [c]	5.0–9.5	2.5	2.0			
Tropical situations						
8a				3.0–7.0	9.0–13.1 [b]	
					11.0–13.9 [b]	
8b				6.8–18.2		
					5.3–5.8 [b]	
9a				0.2–3.3		
9b	8.0–12.0	12.0–13.0 [b]		0–10.2	11.5–12.7	
10a	2.6–4.2	0.2–0.8	1.5–1.6	0.4–0.7	1.8–2.8	3.9–4.1
10b	9.8 [a]	1.3 [a]	0.9	5.6	4.0	3.2

Results expressed as a percentage of total C and N in each fractions.

[a] For sand (20–2000 μm) and silt (2–20 μm) fractions, respectively.

[b] Values for silt + clay (0–50 μm).

[c] Values for 150–2000, 20–150 and < 20 μm fractions, respectively.

Data source and experimental conditions (horizon, incubation duration in days, temperature, vegetation).

1: Chichester, 1969 (topsoil, 14 days, 35°C).
2: Cameron and Posner, 1979 (0–3 cm, 28 days, 30°C, pasture).
3: Lowe and Hinds, 1983 (7 cm, 56 days, 37.5°C).
4: Catroux and Schnitzer, 1987 (0–15 cm, 28 days, 28.5°C).
5: Christensen, 1987 (0–25 cm, 50 days, 20°C, barley).
6: Gregorich et al., 1989 (0–15 cm, 20 days, 24°C, corn).
7: Hassink, 1995 (0–10 cm, 67 days, 20°C, mean values of sandy, loamy and clay grassland soils).
8a, 8b: Bernhard-Reversat, 1981 (0–5 cm, 20 days, 32°C, savannah: a = under trees, b = in the open).
9a: Bernhard-Reversat, 1988 (0–10 cm, 20 days, 32°C, savannah and tree plantation).
9b: Bernhard-Reversat, 1987, 1988 (0–10 cm, 20 days, 32°C savannah and tree plantation).
10a, 10b: Feller and Nicolardot (unpubl.) (0–10 cm, 28 days, 28°C, a = sugarcane, b = millet) Samples that were previously used for greenhouse experiments (Feller et al., 1983; François et al., 1991).

expressed as the percentage of total C or N). Some estimates of PMC and PMN in particle-size fractions taken from the literature are reported in Table 2. The following points will be emphasized from these data. In temperate soils the absolute and relative amounts of potentially mineralizable (biologically active) C and N tend to increase and decrease, respectively with decreases in particle size (Lowe and Hinds, 1983; Catroux and Schnitzer, 1987). In tropical soils, PMC in the sand-size fraction (plant debris) is usually higher than that of silt- and clay-size fractions. There is generally no trend with regard to silt- and clay-size fractions, PMC in the silt-size fraction being either higher (Gregorich et al., 1989; Hassink, 1995; Feller and Nicolardot, unpubl.) for a sandy soil or lower (Christensen, 1987; Feller and Nicolardot, unpubl.) for a clayey soil than that of the clay-size fraction. Moreover, Hassink (1995) showed for temperate grassland soils that PMC did not differ significantly among size fractions in soils of different textures (sandy, loam, clayey). He concluded that the lower PMC "Commonly observed in fine textured soils compared with coarse textured soils is due to a different distribution of SOM among fractions and not to differences in decomposition rates of the fractions". On the other hand, PMN generally increases from sand- to clay-size fractions, the low values of the sand fraction being attributed to nitrogen immobilization in the presence of plant debris. For sandy soils of Senegal, Bernhard-Reversat (1981) found a significant correlation between the PMN and the C/N ratio of the different size fractions. For a collection of temperate soils and tropical Andisols, Sollins et al. (1984) showed

Table 3
Relative contributions of particle-size fractions to the total C and N mineralized, adapted from the literature

Soil type (site)	C mineralized (%)			N mineralized (%)		
	Sand (50–2000 μm)	Silt (2–50 μm)	Clay (0–2 μm)	Sand (50–2000 μm)	Silt (2–50 μm)	Clay (0–2 μm)
Temperate situations						
Loamy sand (5a) [c]	10	25	65			
Sandy loam (5b) [c]	6	21	73			
Tropical situations						
Sandy (8a,b)				15–41	50–84 [b]	
Sandy (10a)	86 [a]	3 [a]	11	43 [a]	15 [a]	42
Clayey (10b)	31	11	58	1	13	86

Results expressed as a percentage of the total C and N mineralized.

[a] For sand and silt, fractions 20–2000 and 2–20 μm respectively.

[b] Values for silt+clay (0–50 μm).

[c] 5a and 5b have 6 and 15% clay, respectively. Values for the 'burned' treatment.

For primary references and site description information see footnote of Table 2.

that a negative relationship also existed between the 'light fractions' ($d < 1.6$) and PMN of these soils.

The contribution of each fraction to the total C (or N) mineralized is dependent on its mineralization potential (e.g. bioavailability), its relative mass and its C concentration. In Table 3 we reported some data for tropical sandy or clayey LAC soils. The relative contributions of the different fractions to the total mineralized C and N depends greatly on soil texture. In sandy soils the sand-size fraction which contains a proportion of plant debris represents a major source of potentially mineralizable C and N. In clayey soils, however, most of the mineralized N (86%), and a large portion of the mineralized C (58%), originates from the organo-clay fraction. Though covering a narrower range of textures, a similar trend to that of clayey tropical soils was observed by Christensen (1987) for temperate arable soils in Denmark (Table 3).

3.1.3.2. Natural abundance ^{13}C enrichment of particle-size fractions. Where vegetative histories are well known, changes in the δ^{13}C signature of organic matter originating from different vegetation types (C3 versus C4 plants) can be used to describe the accumulation of C derived from 'new' vegetation (Y) and the loss of native C derived from 'old' vegetation (X). This technique has been applied to both whole soils and particle-size fractions.

Many of the issues concerning the use of ^{13}C natural abundance as tracer of SOM dynamics were recently reviewed by Mariotti (1991). Other methodological considerations (precision and possible biases) were offered by Balesdent (1991). This approach has been used extensively in the temperate zone (Balesdent et al., 1987, 1988, 1990; Skjemstad et al., 1990; Balesdent and Balabane, 1992; Angers et al., 1995; Gregorich et al., 1995; Puget et al., 1995) as well as in tropical areas (Cerri et al., 1985; Martin et al., 1990; Feller et al., 1991d; Trouvé et al., 1991; Bonde et al., 1992; Trouvé, 1992; Desjardins et al., 1994; Garcia-Oliva et al., 1994; Golchin et al., 1995). The accumulation and loss of bulk soil C at these sites is given in Table 4. In several of these studies the δ^{13}C approach was used to investigate the dynamics of SOM associated with particle-size fractions from surface soil horizons (Tables 5 and 6).

The accumulation or loss of C under changing vegetation is most often described from first order kinetics and expressed as the mean residence time (MRT) or the half-life of C in years. Some examples of the MRT of organic matter associated with different particle-size fractions are given in Table 5. Direct comparisons between sites are made difficult by many factors (e.g. climatic, edaphic and agronomic factors, depth of horizons, fractionation methods and sizes of the fractions) that influence the MRT of SOC. However, within any one site, there is a tendency for MRT to increase as fraction size decreases, a pattern that appears to hold across sites. In France, for example, Balesdent et al. (1987) and Balesdent (1991) have shown that MRT of SOC in the top 30 cm of an Eutrochrept soil (%clay = 27) undergoing C3–C4 succession differed

Table 4

Studies reporting the accumulation and loss of C under changing vegetation using the natural abundance ^{13}C approach

Site [a]	Country (location)	Vegetation change (x to y)	Duration (years)	Depth (cm)	g C/kg soil x	g C/kg soil y	Clay (%)	Reference
Temperate								
A	France (Doazit)	Pine–corn	23	0–30	24	14.5	11	Balesdent et al., 1987
B	France (Auzeville)	Wheat–corn	11	0–30	9.5	8.5	27	Balesdent et al., 1987; Balesdent, 1991
C	USA (Sanborn)	Meadow–wheat	27	0–10	44	12.2	15	Balesdent et al., 1988
				0–10	44	9.1	16	
				0–10	44	12.7	15	
C bis	Canada (Newbois)	Meadow–corn	11	0–8	3.34 [b]	1.93 [b]	23	Angers et al., 1995
C ter	Canada (Winchester)	Forest–corn	25	0–5	138.5	20.4	nd	Gregorich et al., 1995
Tropical								
D	Ivory coast	Grass savanna–tree savanna	16	0–10	7.9	7.1	5	Martin et al., 1990
E	Brazil (Piracicaba)	Forest–sugarcane	12	0–10	32	16.5	63	Cerri et al., 1985, Feller et al., 1991a
			50	0–10	32	15.7	58	
F	Brazil (Capital Poco)	Forest–pasture	9	0–10	13.7	1.8	13	Desjardins et al., 1994
G	Congo (Pointe-Noire)	Grass savanna–eucalyptus or pine	6, 11, 21 0–10	0–10	6.5	6, 6, 1	13	Trouvé et al., 1991, Trouvé, 1992
J	Mexico (Chamela)	Forest–pasture	7	0–6	1.23 [b]	1.07 [b]	25	Garcia-Oliva et al., 1994

[a] Letters refer to abbreviations used by Feller (1995a). Site abbreviations also apply to Tables 5 and 6.

[b] Carbon content in kg/m^2 over the sample profile.

Table 5
Mean residence time (MRT in years) of organic matter associated with different particle-size fractions in some temperate and tropical soils, adapted from the literature

Site*	Particle-size fractions (μm) 2000	200	50	20	10	2	0
Temperate							
B	←5.5→		←15→			←70→	
Tropical							
Ea	←<1→		←40→		←40→		
Eb			←4→		←6→		←59→
G		←10→			←16→		←25→

The horizontal lines distinguish the limits of the fractions studied.
[a] Site descriptions given in Table 4. Ea and Eb according to Cerri et al. (1985) and Bonde et al. (1992), respectively.

substantially by particle-size fraction; averaging 5.5, 15 and 70 years for the 200–2000, 50–200 and 0–10 μm size separates, respectively. In contrast, Cerri et al. (1985) showed for a clayey Oxisol from Brazil that the MRT of organic matter in the 50–200 and 0–50 μm fractions was relatively long (ca. 40 years) while the coarse plant debris fraction (200–2000 μm) exhibited a very rapid turnover rate (ca. annual time scales). The relatively stable C in the 500–2000 μm fraction of this Oxisol (Cerri et al., 1985; Feller et al., 1991d) may result from charcoal debris left after forest burning. Similar observations were reported by Skjemstad et al. (1990) following a change from forest to pasture vegetation in Australia. Using a more complete fractionation method, Bonde et al. (1992) calculated MRTs of 4, 6 and 59 years for the 20–2000, 2–20 and 0–2 μm fractions, respectively. In a recent thesis, Trouvé (1992) applied the ^{13}C approach to changes from savanna to forest (pine or eucalyptus) vegetation. From her results we calculated MRTs of approximately 10, 16 and 25 years for the > 50 μm, 2–50 μm and 0–2 μm fractions, respectively. Although based on a very limited data set, the MRTs of the clay (or clay + fine silt < 10 μm) fraction are lower in tropical than temperate soils suggesting a higher turnover rate for this fraction in the tropics.

The accumulation of C derived from new vegetation with a different $\delta^{13}C$ signature has been used to describe the dynamics of SOC in particle-size fractions. For comparative purposes, the results presented in Table 6 were recalculated from raw data to conform to the > 50, 2–50 and 0–2 μm fractions described elsewhere. The percentage of total C in each size fraction which was derived from 'new' vegetation [$y(\%)$] is presented in relation to the time under

Table 6
The accumulation of C derived from new vegetation (y) and the loss of native C derived from old vegetation (x) in sand-, silt- and clay-sized fractions of temperate and tropical soils

Site	Years under y	y (% of fraction C)			x (% of initial fraction C)		
		Size fractions (μm)			Size fractions (μm)		
		> 50	2–50	0–2	> 50	2–50	0–2
Temperate							
A	23	43	11	19	nd [a]	nd	nd
B	13	64	14	11	nd	nd	nd
C	27	58	42	30	94	74	79 [c]
C	50	95	46	32	98	81	84 [c]
C	98	80	67	56	95	88	78 [c]
C bis	11	27		9 [b]	46		49 [b]
C ter	25	45	21	25	nd	nd	nd
Tropical							
D	16	92	55	30	93	50	51
E	12	36	23	19	54	75	60 [c]
E	50	53	48	34	80	77	72 [c]
F	10	54	40	33	42	48	52 [c]
G	6	55	18	17	54	37	0
G	11	47	25	22	57	47	13
G	21	75	59	49	73	55	46
J	7	43	22	24	nd	nd	nd

[a] nd, no data available.

[b] Values are for 0–50 μm fraction.

[c] Values were calculated from data of the 0.2–2.0 and 0–0.2 μm fractions. Source data are as follows: C27: 81 and 74, C50: 87 and 75, C98: 82 and 64, E12: 62 and 58, E50: 70 and 75, F10: 51 and 58 for the 0.2–2.0 and 0–0.2 μm fractions, respectively.

new vegetation (Fig. 8a1, b1, c1). Similarly, the effect of time under new vegetation on the loss of C derived from 'old' vegetation [x(%)] was estimated (Fig. 8a2, b2, c2). In each case, differences between temperate and tropical situations are distinguished. Several interesting observations can be drawn from these findings.

In general, the accumulation of new C tends to decrease particle size (Table 6), an observation which appears to hold for both temperate and tropical soils. With only two exceptions (B-13 year and E-50 year), the results also suggest that the rate of SOC accumulation in sand-, silt- and clay-size fractions is higher in tropical than temperate soils (Fig. 8). This finding is consistent with the assumption that SOC turnover rates are higher in tropical soils as compared to temperate situations (Ayanaba and Jenkinson, 1990). Of the two exceptions, the B-13 year site represents a shift from wheat to 13 years of corn production in a coarse-textured soil. In this case, wheat-derived C in the sand-size fraction (> 50 μm) appears to be more susceptible to decay than that derived from other sources, thereby enhancing the enrichment of corn-derived C. Soils from the E

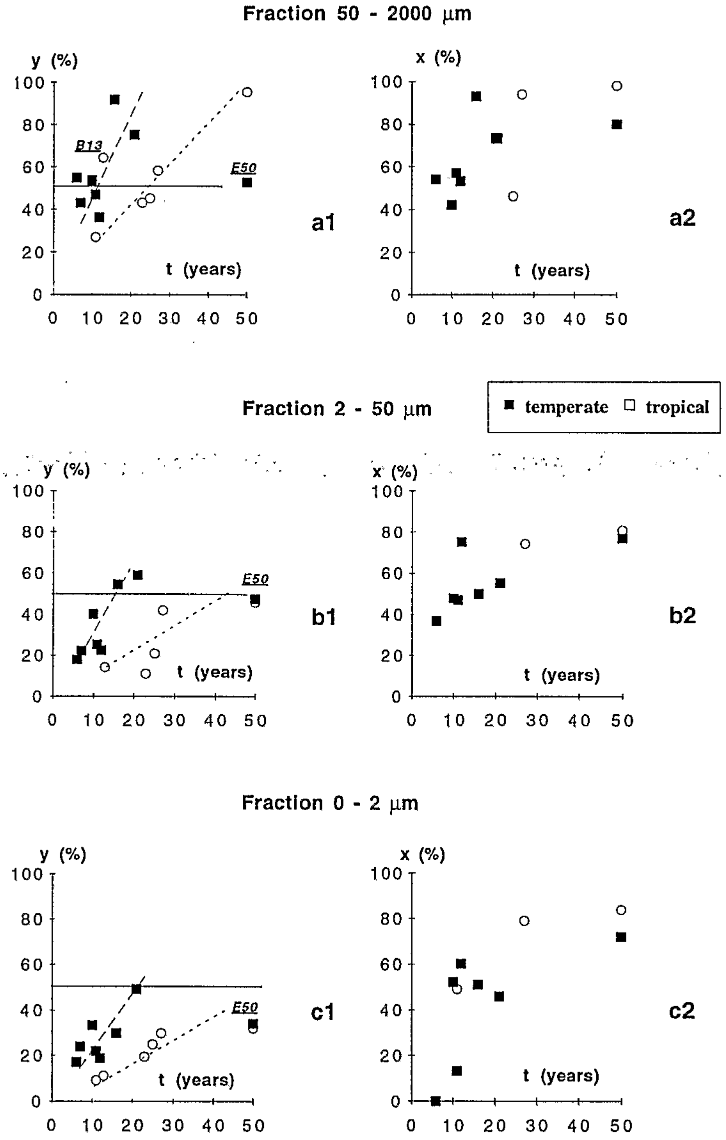

Fig. 8. Relative storage [$y(\%)$] of the carbon derived from new vegetation (y; as % total carbon content) in sand- (a1), silt- (b1) and clay-size (c1) fractions of temperate (○) and tropical (■) soils in relation to time under new vegetation. Relative loss [$x(\%)$] of the carbon derived from the native vegetation (x; as % of initial total carbon content) in sand- (a2), silt- (b2) and clay-size (c2) fractions of temperate (○) and tropical (■) soils in relation to time under new vegetation.

sites were described as clayey Oxisols while those of other tropical sites represented coarse to medium textured soils. For all size fractions the $y(\%)$ values of the E-50 year site are curiously low in comparison to other tropical sites. The low value of the > 50 μm fraction may be explained by the presence of very stable SOC in the form of charcoal (mentioned above), while the low values of the < 50 μm fractions may be attributed to a protective effect of clay and/or oxides in microaggregates (< 50 μm). These explanations, however, require experimental confirmation.

These results can also be used to estimate the time required under the new vegetation to obtain equivalent quantities of SOC derived from 'old' and 'new' vegetation [i.e. y (50%)]. For tropical sites with coarse-textured soils (site E excluded) y (50%) is achieved after about 12, 16 and 22 years for sand-, silt- and clay-size fractions and after 25, 45 and 60 years respectively, in the same fractions of temperate sites. These data provide further evidence in support of a higher turnover rate of organic matter in tropical than temperate soils.

Where available, data on losses of native C (derived from 'old' vegetation) paint a somewhat more confusing picture. Although results from some sites (C, D, G) suggest a trend toward a decrease in $x(\%)$ from sand- to clay-size fractions, data from other sites (Cbis, E12, F) do not support this pattern (Table 6). Furthermore, there was no clear difference in the rate of 'old' C loss between temperate and tropical sites (Fig. 8) though additional data are required to properly evaluate this question. The half-life of C derived from the 'old' vegetation [$x(\%) = 50$] can be estimated from Fig. 8 at about 6 to 12 years for the > 50 μm and 2–50 μm fractions and about 10 to 20 years for the 0–2 μm fraction.

The results of these analyses tend to confirm the observation that C-turnover rates are higher for the particulate organic matter associated with the sand-size fraction (> 50 μm fraction) than clay-associated organic matter. Furthermore, these results provide evidence that C-turnover is higher in tropical than in temperate soils and that the differences may be linked to a faster rate of C-turnover in the sand-size fraction (i.e. particulate organic matter) and a slower accumulation of SOC in the clay-size fraction. There is, however, a need for additional data from tropical soils to understand the effects of soil physical properties on SOM dynamics and to evaluate the role that agricultural management plays in moderating these influences. Nevertheless, it is clear that the $\delta^{13}C$ approach has contributed significantly to our understanding of SOM dynamics in both temperate and tropical soils.

3.1.3.3. Effects of landuse on soil organic matter (SOM) associated with particle-size fractions. Feller et al. (1991a) and Feller (1995a) investigated the effect of landuse management on changes in the SOC content of the particle-size fractions in several of the studies described above. Fig. 9A,B illustrate the effects of changing landuse management on the SOC content of tropical LAC soils (schemes a1, b1, c1) and the contributions of specific particle-size fractions to these changes (schemes a2, b2, c2). Comparisons are made between sandy, sandy-clay and clayey soils.

The effects of landuse management on losses of SOC are illustrated by the

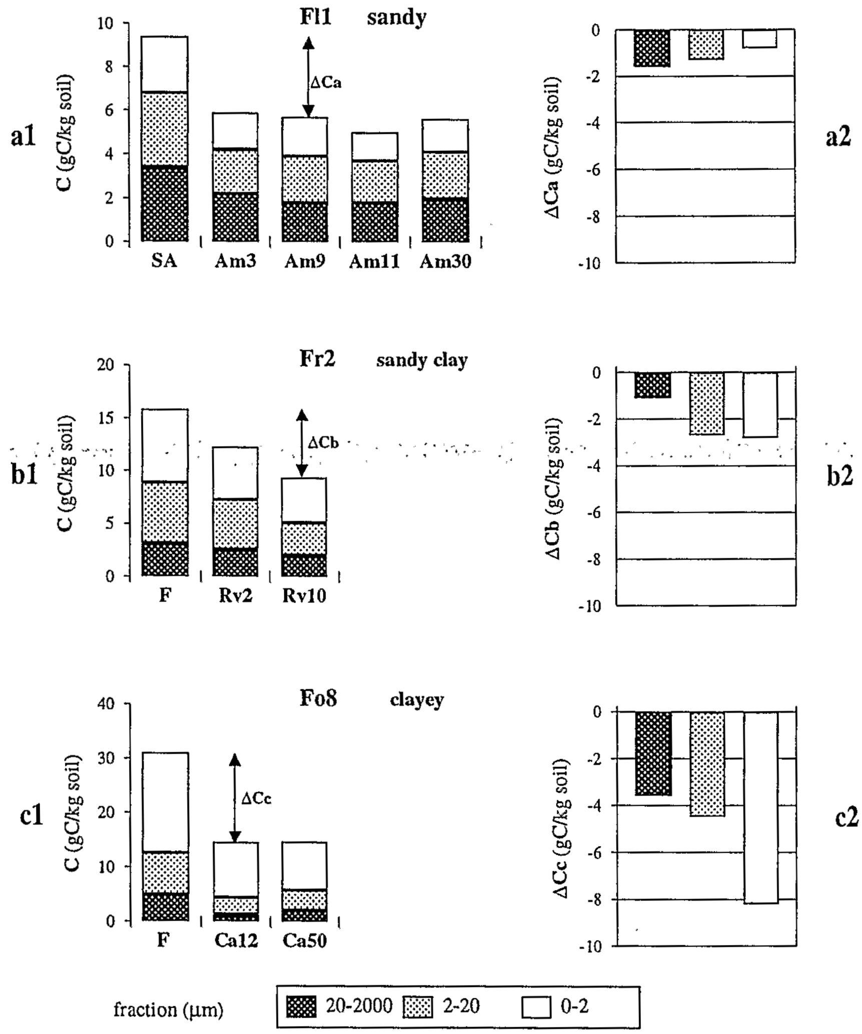

Fig. 9. (A) Decreases in the carbon content (g C/kg) of LAC soils of different textures (sandy [a1], sandy clay [b1] and clayey [c1]) and their fractions after clearing of native vegetation (i.e. savanna [SA] followed by annual cultivated cropping [Am; 3–30 years]; forest [Fa] followed by annual cultivated cropping [Rv; 2 or 10 years] or forest [Fa] followed by sugarcane [Ca; 12 or 50 years]). Fl1 = Senegal; Fr2 = Côte d' Ivoire, Fo8 = Brazil (replotted from Feller, 1995a). The contribution of each size fraction to the total C decline is also shown for the sandy [a2], sandy clay [b2] and clayey [c2] soils. (B) Increases in the carbon content (g C/kg) of LAC soils of different textures (sandy [a1], sandy clay [b1] and clayey [c1] and their fractions in vegetated fallows (herbaceous [Jh], bush [Ja] after continuous cultivation and cropping to maize [Ms, 14 years] or annual rotations [Rv, 10 years or Rc, 10 years]. Fr3 = Togo; Fr2 = Côte d'Ivorie, Fr7 = West Indies (Ste Lucie) (reported from Feller, 1995a). The contribution of each size fraction to the total C increase is also shown for the sandy [a2], sandy clay [b2] and clayey [c2] soils.

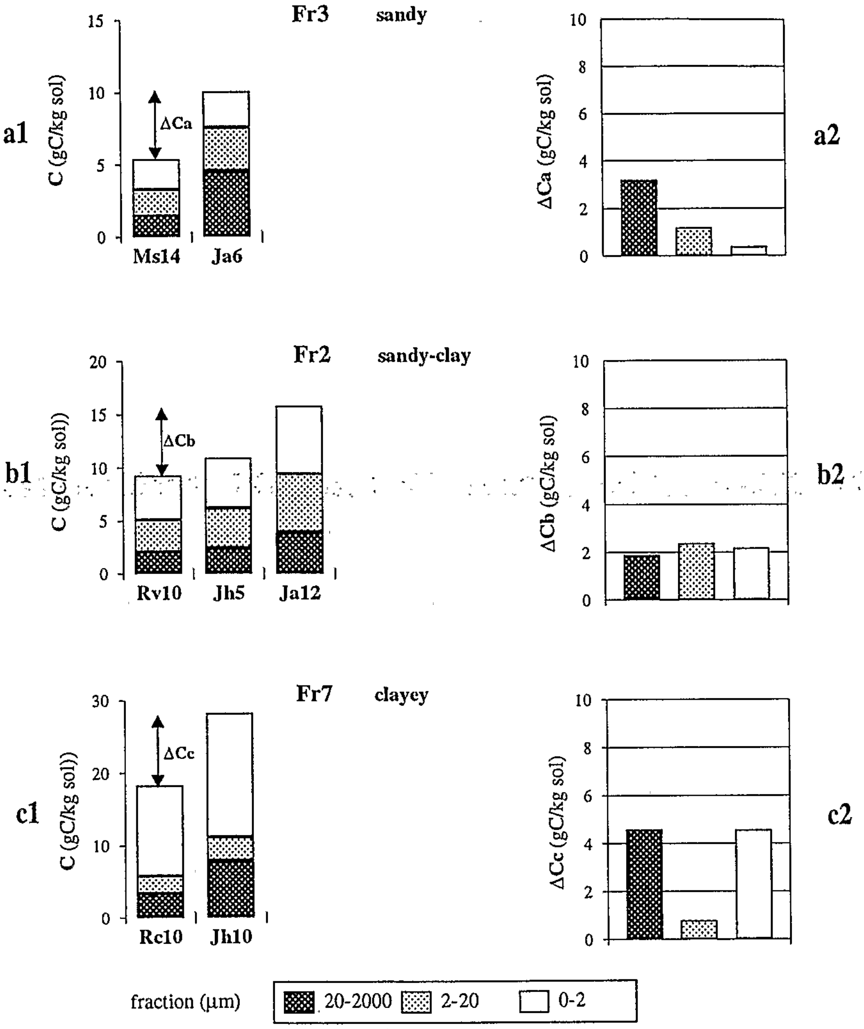

Fig. 9 (continued).

effect of continuous cultivation (cropping annual rotations Am, Rv or sugarcane Ca) after clearing of native vegetation (savanna SA or forests F) in Fig. 9A. The effect of converting continuously cultivated soils (maize Ms or cropping annual rotations Rv and Rc) to vegetated fallows (herbaceous Jh or bush Ja) on the accumulation of SOC is shown in Fig. 9B. Under continuous cultivation, the decrease in SOC contents (Fig. 9A,a1–c1) was primarily due to a loss of particulate organic matter in the sand-size fraction in sandy soils (Fig. 9A,a2) and of clay-associated C in clayey soils (Fig. 9A,c2). The 2–20 μm fraction plays a relatively greater role in the sandy-clay soil (Fig. 9A,b1,b2). Under vegetated fallow, the increase in SOC contents (Fig. 9B,a1–c1) was mainly due to an accumulation of particulate organic matter in the sand-size fraction in sandy soils (Fig. 9B,a2) and to an accumulation of C in the sand- and clay-size fractions of the clayey soil (Fig. 9B,c2). Increases in SOC in the 2–20 μm fraction were limited.

A similar approach (not shown here) was used by Feller et al. (1991a) and Feller (1995a) to describe the effect of conversion to meadows on SOC storage in clayey LAC or HAC soils. Their findings were similar to those reported for the vegetated fallows above and there were no clear differences between LAC or HAC soils. Feller et al. (1991a) also investigated the effect of composted straw on the SOC dynamics of sandy LAC soils and demonstrated that the observed increases in SOC contents were mainly attributable to an accumulation of particulate organic matter in the sand-size fraction (> 50 μm).

Overall, these results suggest that over medium to long time scales (> 5 years), the effects of soil management on SOC accumulation and loss differ quantitatively and qualitatively in relation to soil texture. In coarse-textured soils a significant proportion of the SOC dynamics are due to changes associated with the sand-size fraction (i.e. particulate organic matter). Similar results were reported by Djegui et al. (1992) and Bacye (1993) for sandy to sandy-clay LAC soils of Benin and Burkina-Faso, respectively. In contrast, much of the variation in SOC in fine-textured soils, especially those with a high clay content was associated with the 0–2 μm fraction. These results imply that a significant proportion of the SOC associated with clay fraction is relatively labile, a conclusion that is supported by results using the $\delta^{13}C$ approach described above (i.e. 50% loss of clay associated C in 10–20 years). The observation provides support for the structure of the 'Century' model (Parton et al., 1987) which considers the clay-size fraction as an 'active' to 'slow' pool of SOC in tropical soils. Further research is needed to properly characterize and describe the nature of this pool.

3.2. The stable aggregate fractionation approach (SAF)

Unlike the particle-size fractionation approach which separates primary mineral particles based on size alone, the aggregate fractionation approach separates particles based on both their size and the stability of their binding agents. For many decades, measures of stable soil aggregates have been used to quantify the structural stability of soils and these have most often been related to the SOM content. It was not until relatively recently that research into the formation and stabilization of soil aggregates has been used to describe mechanisms for the physical protection and storage of soil organic matter.

3.2.1. Objectives and methodologies

Efforts to describe the quality and quantity of aggregate-associated organic matter stems from two particular interests: (1) understanding the importance of organic matter constituents for determining the structural stability of aggregates, and (2) identifying the mechanisms by which the aggregation of particles contributes to the physical protection and storage of SOM. Essential to both of these interests is the need to define the size and stability of aggregates using methods that are quantitative and reproducible. The measurement of stable soil aggregates depends on both the forces that bind particles together and the magnitude of the disruptive forces applied. Many different approaches have been used (Jastrow and Miller, 1991; Beare and Bruce, 1993).

An important consideration in comparing the organo-chemical composition of aggregates from different soils and size-classes is their sand content. Sand is excluded from relatively small size-classes (< 0.05 mm) but in larger size-classes it may be found as a contaminant not contained within the aggregated particles. Though sand may be coated with organic matter the organic matter content is relatively low and not contained within a lattice structure as with silt and clay. For these reasons, it is often useful to normalize the organic matter concentrations of different aggregate size-classes for the sand content (Elliott and Cambardella, 1991); that is to express the organic matter concentration on a silt plus clay basis.

The effect of aggregation on the accessibility of organic matter to biological or biochemical mineralization can be studied by comparing results obtained before and after disaggregation of a soil. Various approaches have been used to physically disrupt soil aggregates and release aggregate-bound organic matter. These include sonication, crushing or grinding aggregates to a defined size and slaking in water. Each has very different effects on the organic matter released for mineralization. In particular, it is important to distinguish between those approaches that expose the inter- and intra-microaggregate organic matter; as the former is the primary source of organic matter that is mineralized following cultivation (Elliott, 1986), whereas, the latter is composed of much more persistent, physically protected organic matter that is less susceptible to mineralization (Edwards and Bremner, 1967).

With these differences in mind, we shall report some results concerning the mineralization of carbon and nitrogen from intact and physically disrupted aggregates as a means of quantifying pools of aggregate-protected SOM. The implications of these methodological differences are considered in the following sections, however, because examples from tropical soils are rare, subtropical and temperate situations will be considered as well.

3.2.2. The composition and distribution of aggregate-associated soil organic matter (SOM)

In contrast to the relatively well described relationship between bulk soil organic matter and aggregate stability, the literature is rich in conflicting results regarding the relationship between aggregate size-classes and SOM constituents. Most of the studies investigating the organo-chemical composition of aggregate size-classes have focused on their total organic C (TOC) and carbohydrate contents, and less frequently on their neutral sugar composition. In many cases, TOC and carbohydrate concentrations do not follow a consistent pattern with respect to aggregate size-classes (Dormaar, 1987; Piccolo and Mbagwu, 1990). In others, the relationship between aggregate size and SOM content is more predictable. Tisdall and Oades (1980), for example, reported the following trend for a red brown earth: high SOC contents (1 to 2%) for > 250 μm, low SOC contents (0.3 to 0.7%) for 20–250 μm and medium SOC contents (0.7 to 1.7%) for < 20 μm aggregates. Decreases in TOC and carbohydrate content with decreasing aggregate size were reported by Haynes and Swift (1990) for Inceptisols (comparison of > 2.0 and < 0.25 mm), by Oades and Waters (1991) for a Mollisol and an Alfisol, and by Dormaar (1984) for a Mollisol (0.1 to 2.0 mm aggregates). In contrast, Baldock et al. (1987) showed for a silt loam Brunisol that carbohydrate concentrations were similar for aggregates 1.0 to 8.0 mm in diameter but increased in smaller aggregates (0.1–1.0 mm). The same problems appear in comparing the relative sugar composition (individual sugars as a % of total sugar concentration) of different aggregate size-classes. Although, Dormaar (1984) found no differences in the relative sugar composition of variously sized aggregates, Baldock et al. (1987) observed that in soil under continuous bromegrass the contribution of plant carbohydrates increased as aggregate size decreased, while the reverse was true of soil under continuous grain-corn.

As discussed previously, few studies have considered differences in the mineral and organic components of aggregate size-classes and yet these factors may be critical to interpreting the results obtained. Elliott (1986) compared the nutrient concentrations, normalized to a sand-free basis, of seven aggregate size-classes that were obtained from slaked soils of cultivated and native sod treatments. In general, his results showed that microaggregates contained less organic matter associated with silt plus clay than macroaggregates. Elliott et al. (1991) applied a similar approach to a chronosequence of tropical Peruvian Ultisols under cultivation; in this case correcting for both the sand and light-fraction material in each aggregate size-class. Using this approach the authors demonstrated that the OC concentration of the 'heavy' fraction was, with few exceptions, not different among size classes or treatments. Similarly, Albrecht et al. (1992b) reported fairly uniform concentrations of C (20 mg C/g aggregate) in the 'heavy' fraction (light-fraction free) of a wide range of aggregates (5 to 2000 μm in diameter) collected from a Vertisol and a clayey LAC soil in Martinique.

The results of these studies contrast sharply with those of Beare et al. (1994a) for a sub-tropical Ultisol from Georgia, USA. In this study, the concentrations of total C differ substantially between aggregate size-classes when normalized to a sand-free basis. In general, the highest concentrations were found in the largest microaggregate (106–250 μm) of a no-tillage (NT) soil, decreasing in both larger and smaller aggregate size-classes (Fig. 10). When the total C concentrations were corrected for particulate organic matter (POM), the remaining mineral-associated organic matter was distributed similarly to that of total C, though the magnitude of the differences between fractions was greater for mineral-associated organic matter. These findings provide support for an alternative view of aggregate organization in which microaggregates are formed at the center of macroaggregates (Oades, 1984; Elliott and Coleman, 1988; Beare et al., 1994a). In this view, fragmented organic debris (e.g. roots, fungal hyphae, fecal matter) may become incorporated into macroaggregates by, for example,

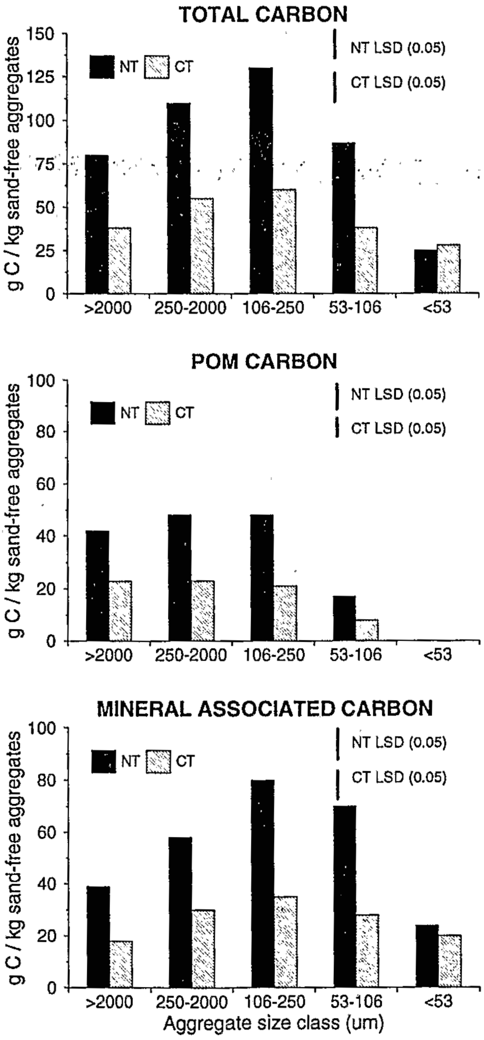

Fig. 10. Total, particulate (POM) and mineral-associated C (normalized to a sand-free basis) in water-stable aggregates from the surface (0–5 cm) of a subtropical Ultisol under conventional- (CT) and no-tillage (NT) management. LSD ($P < 0.05$) ranges are given separately for CT and NT aggregates (from Beare et al., 1994a).

the feeding and casting activities of soil fauna or the shrinking and swelling of soil with drying and rewetting. As these fragments decompose, they may become encrusted with microbial mucilages and clay particles, leaving the organic matter increasingly occluded and inaccessible to microbial attack. The microaggregates formed in this way, and subsequently released from macroaggregates by either physical or chemical stresses, would be relatively stable and enriched in SOM.

Total carbohydrate concentrations also varied significantly between aggregate fractions at this site. The results of Hu et al. (1995) showed that the total carbohydrate concentrations of macroaggregates from NT soils were significantly higher than that of microaggregates (Table 7). Carbohydrates made up 10% of the total C content of macroaggregates but less than 5% of the C in microaggregates. The mannose-to-xylose ratios in these fractions suggest that microbially derived carbohydrates make a much higher contribution to the total carbohydrates found in microaggregate as compared to macroaggregates. This result may be explained by the fact that macroaggregates, in comparison to microaggregates, are enriched in macro-organic matter (plant debris > 50 μm) whose mannose-to-xylose ratio is lower than the SOM in silt- and clay-size fractions (see Table 1).

Support for this alternative view of aggregate organization is found in other studies of subtropical Ultisols from Georgia. In a study comparing no-tilled and intensively tilled soils receiving ^{15}N-labelled grain sorghum residues, Beare et al. (unpublished results) found quite strong effects of management on the distribution of ^{15}N-enriched organic matter among aggregate fractions (Fig. 11). While there was fairly uniform enrichment of all aggregate size-classes in intensively tilled soils, under no-tillage the highest enrichment of ^{15}N was found in the largest microaggregates (106–250 μm) and the smallest macroaggregates (250–2000 μm). Buyanovsky and Wagner (1994) investigated the MRT of ^{14}C

Table 7
The distribution of carbon and carbohydrates in water-stable aggregates from a subtropical Ultisol under no-tillage management (Hu et al., 1995)

Aggregate size class (μm)	Carbon	Total carbohydrates	Sugars/total C	Mannose/xylose ratio
	(g/kg sand-free) [a]		(%)	M/X
> 2000	72.9 [c]	7.45 [c]	10.2 [b]	1.70 [b]
250–2000	95.9 [b]	9.95 [b]	10.4 [b]	1.48 [b]
106–250	101.0 [b]	5.00 [d]	4.9 [c]	2.98 [c]
53–106	87.1 [b,c]	4.28 [d]	4.9 [c]	3.71 [c]

Soil depth 0-5 cm.

[a] All values are expressed on a sand-free basis.

[b,c,d] Means within a column followed by the same letter are not significantly different (ANOVA/LSD; $P < 0.05$).

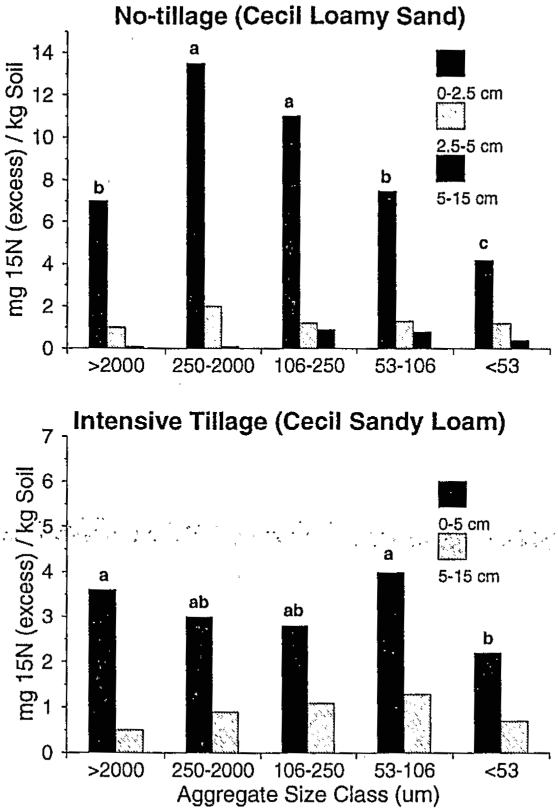

Fig. 11. Incorporation of organic matter from ^{15}N-labelled grain sorghum residues into aggregate fractions of subtropical Ultisols under no-tillage and intensive-tillage management. Results are given for samples collected one year after addition of labelled residues. Bars with letters in common are not significantly different at $P < 0.05$ level. Results of mean separation tests are given only for surface soil samples (Beare et al. unpublished).

released from incorporated residues in various aggregate fractions of a Mollisol under long-term cultivation. In general, their findings showed that for a period of 18 months after incorporation the residence time of carbon increased with decreases in aggregate size: from 1–3 years for macroaggregates (> 250 μm) to 6 years for microaggregates (100–250 μm).

3.2.3. *Physical protection of aggregate-associated organic matter*

As outlined by Ladd et al. (1993), ''Electron microscopy studies (SEM, TEM) have provided the visual evidence to reinforce conclusions drawn from other studies that physical protection mechanisms are important determinants of the stability of organic matter in soil''. Ultramicroscopic observations (TEM) of a tropical Vertisol under pasture (Feller et al., 1996) agree with this 'visual evidence' that plant cell wall debris, bacteria colonies and amorphous OM can be protected from decomposer organisms in microaggregates as they become encrusted in a dense clay fabric. Ladd et al. (1993) also reviewed the various experimental approaches for investigating the physical protection of SOM in non-tropical situations. In general, the authors conclude that with increases in the clay content and/or the structural stability of soils, there are ''Limitations in the accessibility of substrates to decomposer microflora and of microorganisms to microfaunal predators, by virtue of differences in their pore size location within aggregates''.

As discussed previously, applying varying degrees of physical disruption to soils can help to describe the quantity, quality and location of physically protected, biologically active pools of organic matter. This approach has been particularly useful in distinguishing macroaggregate- from microaggregate-protected organic matter. Experimental evidence for a physically protected pool of organic matter depends greatly on the methods employed. For example, using a coarse sieving approach (6 versus 2 mm sieved soil), Ross et al. (1985) found no consistent influence of sieve size on a wide range of microbial and biochemical indices (microbial biomass, mineral-N flush, P-biomass and ATP content), although there was a trend toward higher net N mineralization in more finely sieved samples. In contrast, Scheu et al. (1996) and Stenger et al. (1995) measured significant increases in N and C mineralization when undisturbed aggregates were crushed to pass 1 and 2 mm sieves, respectively.

Using a slightly more destructive technique (e.g. crushing), Elliott (1986) and Gupta and Germida (1988) showed that from 18 to 43% of the potentially mineralizable N in macroaggregates from native sods was protected from microbial attack by its association within the intact structure of macroaggregates. This organic matter presumably represents the inter-microaggregate organic matter that is protected from microbial attack by its association within macroaggregates; that which is most susceptible to mineralization following cultivation. With a similar objective, Golchin et al. (1994) described methods for the separation of 'free' and 'occluded' particulate organic matter in soils. The 'free' fraction was obtained by densimetric fractionation of the whole soil without disruption of macroaggregates and therefore included organic matter external to the aggregate. The 'occluded' fraction was separated after disruption of aggregates by ultrasonication. The two fractions differed in their morphology, chemical composition and SOM turnover (Golchin et al., 1995). The 'occluded' fraction exhibited a slower turnover than the 'free' fraction. The macroaggregate-protected pool of OM appears to be much greater in fine-textured than in the coarse-textured soils (Borchers and Perry, 1992; Hassink, 1992, 1995) and is relatively undisturbed as compared to cultivated soils (Gupta and Germida, 1988; Beare et al., 1994b).

Other studies have used even more energetic techniques, such as ultrasonication and fine grinding, which expose both micro- and macro-aggregate-protected organic matter (Adu and Oades, 1978; Gregorich et al., 1989; Borchers and

Table 8

Standing stocks of aggregate protected and unprotected C and their C/N ratios in aggregates from a subtropical Ultisol under CT and NT management (from Beare et al., 1994b)

Sample depth	Size class (µm)	Standing stocks g/m^2						Carbon-to-nitrogen ratios					
		Aggregate, unprotected C			Macroaggregate, protected C			Aggregate, unprotected C			Macroaggregate, protected C		
		NT	CT	$P < 0.05$	NT	CT	$P < 0.05$	NT	CT	$P < 0.05$	NT	CT	$P < 0.05$
0–5 cm													
	> 2000	24.9	13.2	* [a]	7.5	0.07	*	10.0	9.1	*	6.8	8.2	*
	250–2000	9.6	10.2		2.8	1.2	*	9.2	9.4		7.5	13.6	*
	106–250	3.4	3.3					8.7	8.5				
	53–106	0.6	0.7					3.7	4.7				
	Total	38.4	27.4	*	10.3	1.9	*	9.4	8.8		6.9	10.9	*
0–15 cm		74.2	74.7		17.2	8.5	*	9.1	8.7		8.9	17.3	*

[a] Asterisks indicate significant differences ($P < 0.05$, two-tailed t-test) between tillage treatments within aggregate size-classes.

Perry, 1992). In general, these studies have shown that a large proportion of the physically protected biologically active organic matter is contained within micro-pores or is associated with microaggregates. As an extension of this work, Powlson (1980) used chloroform fumigation to investigate the contributions of microbial organic matter to the C and N released by fine grinding (mortar and pestle). His results showed that on the order of 2.7 to 3.9 times more organic C was made available for mineralization by fine grinding (releasing microaggregate-protected OM) and that 23–48% could be attributed to killed biomass. Still, other studies using highly disruptive techniques have reported little or no evidence of aggregate-protected organic matter (Catroux and Schnitzer, 1987). The extent to which the differences in these findings reflect specific properties of the aggregate-associated OM or methodological differences remains unclear.

Although the above studies provide reasonably strong evidence of aggregate-protected SOM in temperate soils, the importance of aggregation, particularly macro-aggregation, to SOM storage in tropical soil is much less clear. For example, in a recent study of LAC (kaolinitic) soils, Feller (1993) found no differences in the quantities of potentially mineralizable C and N from intact 'aggregated' soils and the sum of that measured from dispersed particle-size fractions. Similarly, Robinson (1967) found little or no effect of grinding samples into increasingly smaller particle-size fractions on the quantity of nitrogen mineralized from several tropical red earths (Latisols and Ferrisols) of East Africa. Bernhard-Reversat (1981) also found no effect of particle-size fractionation on the nitrogen mineralized in short-term incubations (< 1 month) of a sandy LAC soil from northern Senegal.

In contrast to the above studies, Beare et al. (1994b) found that the quantity of 'unprotected' potentially mineralizable C in 250–2000 μm aggregates of a sub-tropical Ultisol from Georgia, USA was higher than that of either larger (> 2000 μm) or smaller (< 106 μm) aggregate size-classes (Table 8). Concentrations of macroaggregate-protected C were also somewhat higher in smaller macroaggregates than larger ones. The estimated rates of C mineralization from different aggregate fractions suggested that the unprotected pool of C in the smallest microaggregates was somewhat more labile than that of larger aggregate size-classes. Although quantities of macroaggregate-protected C (that released by aggregate disruption) are much lower than unprotected C, their rates of mineralization indicate that protected pools of C are more labile than unprotected pools (Table 9). Presumably, these macroaggregate-protected pools of organic matter are rapidly mineralized when aggregates are physically disrupted by tillage or slaking.

It is important to note that physical controls on the storage and loss of organic matter cannot be viewed in isolation from biological influences. Soil biota are clearly important in mediating physical changes in soil structure that may alter the storage and transformations of SOM. Biological constituents ranging from roots and fungi to microarthropods and earthworms can influence the formation

Table 9
Mineralization potentials and rate constants for protected and unprotected pools of C in aggregates from a subtropical Ultisol (0–5 cm depth) under CT and NT management (from Beare et al., 1994b)

Size class (μm)	Carbon mineralization potential, C_o [a], g C/kg s.f. [b]						Rate constants (k) [a], (day^{-1})					
	Aggregate, unprotected C			Aggregate, protected C			Aggregate, unprotected C			Aggregate, protected C		
	NT	CT	$P < 0.05$	NT	CT	$P < 0.05$	NT	CT	$P < 0.05$	NT	CT	$P < 0.05$
> 2000	2.35	1.44	* [d]	0.68	0.08	*	0.127	0.137		0.196	0.300	*
250–2000	3.17	2.26	*	0.92	0.28	*	0.124	0.146	*	0.183	0.203	
106–250	2.85	1.62	*				0.137	0.129				
53–106	1.28	1.07					0.217	0.236				
LSD ($P = 0.05$) [c]	0.44	0.38		0.32	0.10		0.019	0.031		0.040	0.082	

[a] Asymptotic parameter estimates based on non-linear regression fitting of the first-order model: $C_{\mathrm{min}} = C_o(1 - e^{-kt})$.
[b] Units are g CO_2-C respired per kg sand-free (s.f.) aggregates.
[c] Fisher's least significant difference (LSD) values for comparison of means of size-classes within tillage and soil depth (ANOVA, $P < 0.05$).
[d] Asterisks indicate significant differences ($P < 0.05$, two-tailed t-test) between tillage treatments within aggregate size-classes.

and stabilization of soil aggregates. For example, Martin (1992) and Lavelle and Martin (1992) observed an important effect of aggregation on C mineralized from casts of endogeic earthworms (*Millsonia anomala*) from an Alfisol of the Ivory Coast. Their studies showed that after 400 days of incubation, the organic C content of the earthworm casts was 20% higher than that of the control (non-ingested soil). This was attributed to the higher C content and stability of the earthworm casts than the mineral soil (Blanchart, 1992; Blanchart et al., 1993). Other organisms may also contribute to the physical protection of organic matter through their influence on soil aggregation. For example, in a recent study of a Georgia Ultisol, Beare et al. (1997) used field applications of a fungicide to investigate the influence of mycelial fungi on the stability of soil aggregates and the retention of soil organic carbon. Their results showed that where densities of fungal hyphae were reduced in NT soils there was a 40% reduction in the largest macroaggregates (> 2000 μm) and a redistribution of particles into smaller size-classes. However, in CT soils, where hyphal densities are somewhat lower, there were no significant effects of the fungicide applications on aggregate size-distributions.

3.2.4. *Effects of landuse management on aggregate-associated soil organic matter*

The initial stability of aggregates used in studies of aggregate-protected OM is often poorly defined. It is likely that the protective effect of aggregates is lower in soils with low aggregate stability (sandy and/or cultivated soils) than in those with higher aggregate stability and greater C storage (e.g. fine textured, carbon-rich soils under native vegetation or reduced tillage). Evidence for this is provided in several studies (Borchers and Perry, 1992; Hassink, 1992; Beare et al., 1994a,b). In a recent study of a subtropical Ultisol, Beare et al. (1994a) showed that macroaggregates of CT soils were fewer and less stable than those of NT soil. Furthermore, macroaggregates accounted for much more of the total SOM in NT (43%) as compared to CT (26%) soils. Measurements of biologically active pools of aggregate-associated SOM provide further evidence of the relationship between agricultural management and the physical protection of organic matter. Beare et al. (1994b) found that macroaggregate-protected SOM (i.e. the additional organic matter released by aggregate disruption) accounted for 18.8 and, 19.1% of the total mineralizable (protected plus unprotected) C and N in NT soils as compared to only 10.2 and 5.4%, respectively, for CT treatments (Table 8).

As such, continuously cultivated soils often lack a measurable pool of macroaggregate-protected SOM: a factor that may help to explain the negative findings reported by Robinson (1967), Bernhard-Reversat (1981) and Feller (1993), all of which were obtained from cultivated or poorly aggregated sandy soils. For this reason, management history is likely to be an important factor in determining the size and composition of aggregate-protected SOM pools. These observations point out the clear need to systematically investigate factors which regulate the physical protection of organic matter in soil aggregates and their contribution to the storage of SOM in both temperate and tropical soils.

4. Summary and conclusions

In this paper we evaluated the effects of soil physical properties (mineralogy, texture, structure) and landuse management practices on SOM storage and dynamics in the tropics. Despite early descriptions to the contrary, the overlap in SOM content of soils from temperate and tropical regions is now known to be extensive. Although the importance of climatic factors (e.g. mean annual temperature) in regulating SOM dynamics cannot be denied, long held assumptions with regard to latitudinal gradients in SOM content now appear to be more closely related to differences in mineralogy and landuse history.

The importance of mineralogy for SOM storage is well-known for soils with vitric or andic characteristics such as allophanic (ALL) soils. However, little is known of the extent to which SOM storage in smectitic (HAC) soils differs from that of kaolinitic-oxidic (LAC) soils. Furthermore, there is a particular gap in information on the contributions of soil texture (e.g. clay content) and erosion processes to SOM storage and loss in tropical HAC soils.

In contrast, the importance of texture to SOC storage in LAC soils is well documented and appears to hold true for both cultivated and uncultivated situations. In native and uncultivated soils, the positive relationship between SOC and clay content concerns the storage of OM in the form of particulate organic matter (POM; plant debris > 50 or 20 μm) as well as OM associated with the organo-silt and organo-clay fractions (dominated by amorphous OM). In cultivated soils, however, differences in SOC are mainly associated with the organo-clay fraction, though in most cases the POM pool makes up a relatively small proportion of the total SOM. These findings suggest that the POM fraction is highly susceptible to loss during cultivation and yet responsive to differences in clay content where soils are not disturbed by tillage. This may be particularly important in tropical soils where 20 to 40% of the SOC is associated with the sand-size (POM) fraction as compared to 2 to 14% for many temperate soil examples. However, results comparing cultivated and uncultivated soils indicate that from 30–40% of the SOC associated with the clay fraction belongs to a labile-to-slow pool of organic matter. This pool is much larger than that normally associated with microbial metabolites (i.e. active pool) and further research is needed to clearly define its nature and composition.

The effect of cultivation on the net dynamics of SOM (studies without tracers) appear to be very different for coarse- and fine-textured soils. The differences are due mainly to the plant debris (POM) fraction of sandy soils whereas the organo-clay fraction plays a greater role in fine-textured soils.

Studies of the in situ turnover of SOM (the $\delta^{13}C$ approach) in tropical C3–C4 successions have helped to confirm that OM turnover rates decline with decreases in particle-size fraction, allowing a more precise evaluation of findings obtained by non-isotopic techniques. Results of these studies confirm that greater than 50% of SOC associated with the clay fraction and derived from native vegetation may be lost after clearing and cultivation over a period of 10–20 years. Furthermore, the storage of SOM originating from newly established vegetation is strongly influenced by soil texture. The sand-size fraction (i.e. particulate organic matter) appears to be particularly important in regulating the SOC dynamics in coarse-textured soils. In fine-textured soils, both the sand- and clay- size fractions appear to play an important role in organic matter dynamics.

The influence of soil structure on SOM storage may result from two different factors: (1) the physical protection of SOM against mineralization due to its inaccessibility to microbial attack and (2) a reduction in the detachment of fine particles that contribute to losses of SOM by erosion, particularly in semi-arid and subhumid tropical soils. Although much more information is needed to draw any general conclusions with regard to the physical protection of OM in tropical soils, the existing data suggest that the size and composition of aggregate-protected SOM pools in LAC soils may be particularly sensitive to management as compared to temperate soils. While physical protection of SOM can account for up to 20% of the mineralizable N in well aggregated (e.g. direct drilled) cropping soils, less than 6% of the potentially mineralizable N is protected in intensively cultivated soil. The importance of SOM content to determine the detachability and erodibility of SOC appears to be particularly important in clayey soils of low structural stability. This effect should be taken into consideration when estimating the loss of SOC at the field or landscape level.

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