

The Characteristics of Soil Organic Matter Relative to Nutrient Cycling

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I. Introduction

Soil Organic Matter (SOM) is the primary sink and source of plant nutrients in natural and managed terrestrial ecosystems. It increases the ion exchange, water holding and infiltration capacity, promotes the formation of soil aggregates and is the major energy substrate for the soil microbiota (Sanchez et al., 1982). Soils that support high value crops where the required nutrients are replaced by fertilizers, those that do not have toxicity, aggregation or erosion problems or those supplied with ample irrigation and other management inputs can support crops without much SOM. These situations, which are basically field nutrient cultures, are rare and in the majority of soils, SOM is a necessary prerequisite for ecosystem health and productivity.

II. Soil Organic Matter as a Nutrient Reservoir

Generally, 95% or more of the N and S and between 20 to 25% of the P in surface soils are found in SOM. The C:N:S ratio of agricultural soils is approximately 130:10:1.3 whereas that of grassland and forests is approximately 200:10:1

(Stevenson 1994). The C:P and N:P ratios vary as a function of the parent material, degree of weathering, vegetation and management. The N:S ratio is less variable because these constituents are utilized in the formation of SOM by polymerization reactions (Paul and Clark, 1996). The weathered soils of the tropics (oxisols, spodosols and ultisols) usually contain less total P and a higher proportion of organic P than do the less weathered soils of temperate regions (Duxbury et al., 1989). Sanchez et al. (1982) compared the organic C and N in 61 soil profiles in the tropics with 45 from temperate regions. There were no significant differences in the percent of C between soils from tropical and temperate regions at any of the three depths measured (0 to 15, 0 to 50, and 0 to 100 cm depth intervals). Total N contents were significantly higher in the tropics at the 0 to 15 and 0 to 100 cm depths. Tropical forests had higher SOM contents and lower C:N ratios than savannahs. This is in contrast to temperate areas where grasslands are known to have higher SOM contents and lower C:N ratios.

Nitrogen is held primarily in the organic form. In moist soils where inorganic S does not accumulate, SOM is also the primary source of S. In soils with an accumulation of sulfate, organic matter still plays a major role in constituting the active forms of this nutrient. Phosphorus is present in all terrestrial systems in both organic and inorganic forms. Organic P forms the major labile source of P. In addition, SOM keeps inorganic P from being transformed into insoluble phases in many soils. Phosphorus is considered to be the ultimate control on SOM accumulation in both aquatic and terrestrial systems. Over geological time, N can be fixed from the atmosphere but the rates of weathering control P availability to plants. This in turn controls the input levels of plant residues. The availability of micro-nutrients (Fe, Mn, Cu, B, Mo, Zn) often is associated with SOM turnover. Soil organic constituents also complex and lower the toxicity of Al and Mn species found in many weathered soils (Brown et al., 1994).

The primary plant nutrient contained in SOM is C. It comprises slightly less than half the weight of plant residues but somewhat more than 50% of the weight of SOM. The increase in atmospheric concentrations of CO_2 from 260 uLL^{-1} in the mid 19th century to the present 360 uLL^{-1} has been in part responsible for the ability to increase agronomic yields through better management. The resulting increased residue inputs in both native and natural systems have increased the levels of SOM and thus the nutrient supplying power of soils throughout the world. A tropical rain forest is shown in Figure 1 to have a total system C storage of 34.6 kg C m^{-2} . Secondary forests and savannahs all have less than 10 kg m^{-2} . The primary rain forest is the only site where above ground C is greater than beneath ground storage. In most native systems, shoot production depends on the nutrients recycled within the plant as the need for leaf growth occurs. There also is a tight relationship between litter, roots, rhizosphere organisms and the mycorrhizal fungi. This ensures that degradation through the loss of nutrients does not occur to a significant extent in native systems. In addition the roots can withdraw nutrients from depth. It therefore is not surprising that forest regrowth is a major form of rehabilitation of degraded soils in the tropics. The effect of management and herbivory on residue additions also must be considered. Above ground herbivory,

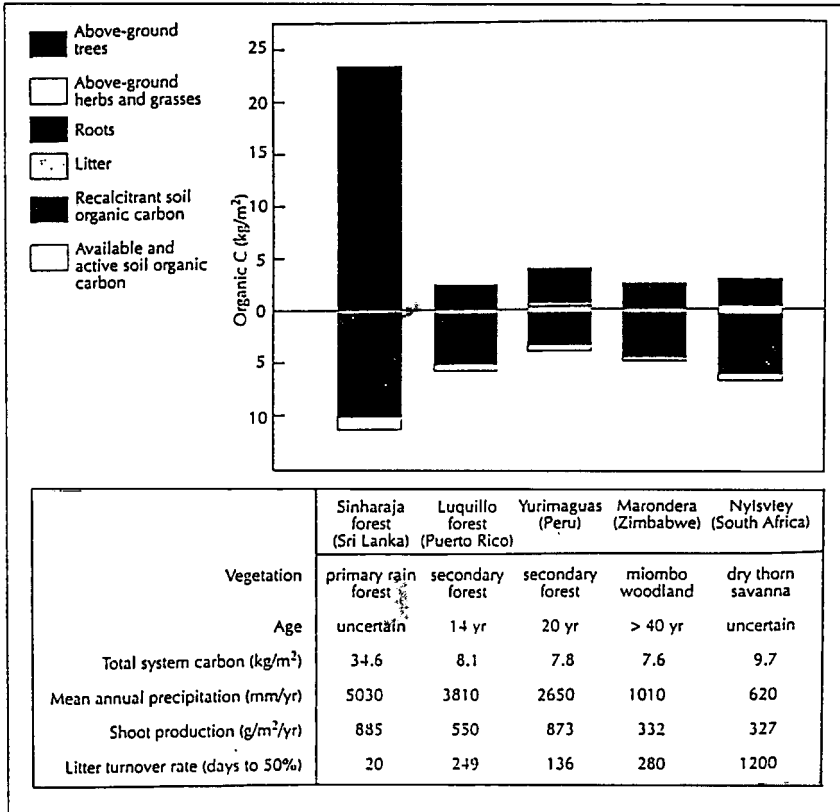


Figure 1. Carbon partitioning in primary and secondary forests. (From Woomer et al., 1994.)

the cutting of hay, and the pruning of trees result in a loss of beneath ground fine roots and associated mycorrhiza.

Concentrations of N, P, K, Ca and Mg present in several tropical lowland forests are given in Table 1. Carbon is concentrated in the above ground canopy. The N and P are concentrated in the litter and soil. The C:N:P ratio of the canopy is 250:1:0.1. That of the litter is 100:1:0.25 and soil is 5:1:0.1. Disturbance of this resource can lead to rapid degradation if N is lost to the atmosphere or groundwater or if P is chemically fixed. However, it should be possible to harvest the canopy and still leave the litter, soil, and root resources. Soil organic matter is a self-regenerating natural resource requiring continual inputs. It is not surprising that the removal of approximately 50% of the C from above ground by fires, through forest clearing, the increased decomposition due to initial cultivation or the lack of replacement of agricultural residues leads to rapid soil degradation. It should not

Table 1. Carbon and nutrient pools and fluxes of lowland tropical forests

	Org-C	N	P	K	Ca	Mg
Pool sizes g m ⁻²						
Above ground	34,000	133	11	96	180	29
Litter	3,700	40	1.2	2.6	18	2.8
Roots	4,100	44	1.2	9.6	56	4.9
Soil	2,300	458	41	25	360	43
Annual fluxes g m ⁻² y ⁻¹						
Rainfall	-	1.5	1.1	1.1	1.4	0.4
Litter fall	-	15	0.7	6.5	14	3.2
Stream water losses	-	3	0.07	1.2	6.3	3.2

(Adapted from Anderson and Spencer, 1991; Brown et al., 1994.)

be too radical to require that research and management policy be in place as to how the litter C from forest clearing can be retained on site or on adjacent sites before a site is clear cut. The most important single factor controlling the level of SOM is the incorporation of plant residues into that soil. To assume that the soils of the world can be utilized without appropriate input into this renewable resource is to guarantee degradation. If physical degradation such as compaction, erosion and lack of water infiltration does not occur, soils can be a very resilient and sustainable. Lugo and Brown (1993) show that increases in SOM contents have occurred due to improvements in plant productivity. Aggressively growing pasture grasses can compensate for the loss of SOM in recently deforested soils. The soil C of pastures was greater than that of 20 old forest plantations although the above ground biomass of the plantations was greater.

III. Forms of Soil Organic Matter

It is difficult to separate the plant residues undergoing decomposition from the soil biota carrying out the processes and the soil humic constituents resulting from the process. Most soil analyses are conducted after sieving through a 2 mm sieve. This removes the larger particles but not the partially decomposed residues and their associated organisms. Organic matter by definition consists of the partially decayed plant residues that are no longer easily recognizable as plant materials, the microorganisms and microfauna involved in decomposition and the by-products of microbial growth and decomposition. These by-products undergo humification to form the materials known as humus (Figure 2). Humic materials consist of dark colored organic condensates that have a higher C and lower O content than most plant and animal residues. They are approximately 50 to 55% C, 4.5% N, and 1% S with varying amounts of P and metals. These materials are closely associated with the soil's inorganic constituents and often occur within aggregates. They thus

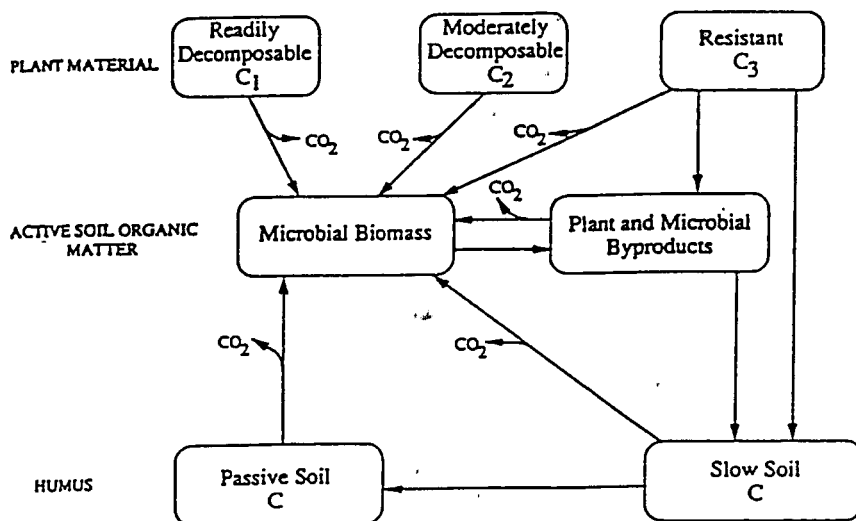


Figure 2. The role of plant residues and microbial by-products in the formation of soil organic matter. (From Paul and Clark, 1996)

decompose very slowly, accumulating in nature as SOM as well as in peats, coals, oils and organic sediments (Paul and Clark, 1996).

Because the humification process is primarily a chemical one, it is not enzymatically controlled but is primarily a free radical type of reaction. Although no two of the resultant molecules can be expected to be identical, the general form of humates in many parts of the world is similar. There is enough original plant C remaining unaltered that the imprint of the original plant structures can be measured with modern instrumentation. Measurement of the ^{18}O content has shown that the majority of O in soil's humic substances comes from cellulose and other plant carbohydrates rather than from the plant lignin (Dunbar and Wilson, 1983).

A. Chemical Characterization

Since SOM plays many roles, no one technique can characterize it relative to its many functions. It has been difficult to adequately establish appropriate analytical techniques to define the pools and fluxes of SOM relative to nutrient availability. Chemical analyses can be conducted on an elemental basis and on the basis of the organic components that can be identified. Less than 50% of the organic N in soil can be identified as belonging to any particular chemical class recognizable in plant, animal or microbial constituents. Amino acids comprise the largest majority of N. They do not vary greatly in various soils of the world or under different cultivation

practices (Sowden et al., 1977). Amino sugars which occur initially as microbial cell wall constituents make up a smaller percentage. They vary with soil type and management (Stevenson, 1994). The unidentified N is assumed to be heterocyclic in nature and includes pyridine and phenoxazone derivatives. Much of this N participates in the condensation reactions that stabilize the humics and therefore by definition is different than the enzymatically formed products occurring in plants, animals and microorganism (Stevenson, 1994).

A majority of S in soil is of the ester form, C-O-S or C-N-S. Such materials called HI reducible-S make up to 30 to 70% of the organic S in soils. The remaining organic S occurs directly bonded to C and thus is not reducible by HI. The ester bonded form is considered to have the highest turnover rates. Phosphorus is present as organic esters especially as materials such as Ca phytates. Most fractionation techniques that determine polysaccharides, proteins, etc. of plant, animal and microbial residues leave a large percentage of unidentifiable residue in soil. The polycondensed humic materials have a large range of molecular weights and chemical structures and cannot be readily be broken down into their constituent groups. Six-normal acid hydrolysis that breaks the constituent proteins to amino acids and ammonia and polysaccharides to individual sugars has been often been utilized. It does not completely solubilize constituents such lignocellulose, but has been found useful both in characterizing the biochemically identifiable components and as an overall measure of the resistant components of temperate soils. When utilized prior to ^{14}C dating, it leaves an insoluble residue that on average is 1,200 years older than the total soil both in the surface and subsurface horizons (Paul et al., 1996). Acid hydrolysis leaves behind much of the polyaromatic C. It selectively removes the N, P, and organic S and cannot be directly used in the measurement of the effect of disturbance on these elements. These, however, are closely associated with C; conclusions as to their fate can be drawn from the associated C and a knowledge of the C:N:P:S ratios. Acid hydrolysis while removing 30 to 80% of the soil C is not as sensitive to the effects of disturbance as it is to site characteristics (Collins et al., 1996).

The classical chemical fractionation technique for fractionating SOM is based upon differences in solubilities in alkaline and acid solutions (Aiken et al., 1985; Stevenson, 1994). The technique involves extraction with an alkaline reagent, usually NaOH or $\text{Na}_4\text{P}_2\text{O}_7$. The extract is further subdivided into humic and fulvic acids by adjustment with acid to a pH of 2. The fulvic acids are soluble in both alkali and acid. Humic acids are soluble in alkali but precipitate at pH 2. The material insoluble in base constitutes the humin. Humin is considered to be comprised of materials closely associated with clay and bound by multivalent cations as well as constituting truly insoluble products such as waxes and charcoal. The fulvic and humic acids differ in molecular weight, elemental composition and acidity reflecting differences in soil forming processes. The humic to fulvic acid ratio can vary by a factor of 5. Grassland soils such as the mollisols have the highest concentration of humic acids. Forest soils such as ultisols and spodosols with mobile soil organic matter have the lowest ratios. The humic fraction is comprised of highly condensed and aromatic and paraffin structures (Aiken et al.,

1985). The humic acids are the most stable fraction (Martel and Paul, 1974; Anderson and Paul, 1984) and should increase in percentage as the soil organic matter drops upon soil degradation.

Chemical characterization of total SOM or its fractions can also be conducted by nondestructive techniques such as solid state nuclear magnetic resonance with cross polarization and magic angle spinning spectroscopy. These expand on the earlier measurements with ultra violet-visible and infrared spectroscopy. Skjemstad et al., (1986) utilized multiple techniques to investigate the nature of the organic matter in a vertisol under native acacia scrub and after up to 45 years of cultivation. Continuous cultivation for 20 resulted in 35% loss of organic C; 45 resulted in loss of 66% of the SOM. Solid state ^{13}C NMR spectroscopy performed directly on samples from both native and degraded sites showed only minor differences in the in situ organic matter. Long chain alkyl groups represented greater than 50% of the organic C. The light fraction ($< 1.6 \text{ g cm}^{-3}$) was greatly reduced on cultivation while the heaviest fraction, which contained the greatest concentration of alkyls, remained unchanged. In these soils, cultivation decreased the relative amount of humic acids but increased the amount of clay organic matter materials.

Solid state ^{13}C NMR has been utilized to quantify the chemical changes associated with decomposition of added substrates (Baldock and Preston, 1995). The expected decrease in the content of carbohydrate C and increases in the amount of lignin were observed. Increases in the contents of alkyl and carbonyl C also were noted. Temperature increases during decomposition increased the rate of transformations but not the chemical outcome. Nutrients can affect the outcome of decomposition. In wood, the absence of N stimulates lignin degradation by white root fungi; under high N conditions polysaccharide degradation is stimulated. Whether this occurs only in woody litter is not known. The white rot fungi such as *Phanerochaete* are not good soil competitors and the effect of available N on lignin degradation in soils has not been established.

B. Physical Fractionation

The clay and silt content of many but not all soils plays a major role in stabilization of SOM. The type of clay, as with allophanic soils, also plays a role. These factors together with the effects of aggregates are now considered under the general heading of physical protection. Factors involved include adsorption of the organic substrates, the enzymes and the microorganisms themselves. Formation into aggregates also affects water availability, aeration, the movement of biota and substrate accessibility. Both the size and the stability of the organic matter are known to be related to particle size distribution in soil. The sand fraction contains the majority of relatively undecomposed plant residues. These vary with vegetation type and management with degraded soils having a much lower proportion of undecomposed plant residues than native sites. The silt size fraction contains not only the greatest concentration but also generally the most resistant organic carbon

(Martel and Paul, 1974; Woormer and Swift, 1994; Dalal and Mayer, 1986 a,b). Soil fractionation based on the density of the organo mineral particles involved is now widely conducted. It considers both the association of organics with minerals and the role of the aggregate structure. (Biederbeck et al., 1994; Buyanovsky et al., 1994).

Plant residues comprise the most labile components and can account for significant proportions of the soil C. They have most often been measured as the light fraction by using dense liquids that are organic or inorganic in nature (Beare et al., 1994 a,b; Janzen et al., 1992). The light fraction which has a density of 1.5 to 2.0 g cm⁻³ has a wide C:N ratio and is biologically very active. The heavy fraction has a narrower C:N ratio and decomposes more slowly. The light fraction comprises 2 to 17 % of the SOM and is highest under perennial forages or where soils are continuously cropped. The respiration rate and microbial biomass were highly correlated to the light fraction (Janzen et al., 1992). The light fraction obtained with these techniques is similar but not identical to the materials separated on a sieve with a 53 μm mesh and designated as particulate organic matter (POM) by Cambardella and Elliott (1994).

Before analyses, aggregates can be disrupted by shaking, the addition of dispersing agents or sonication (Christenson, 1992). In some studies, since aggregates play a major role in controlling turnover in the field, little disruption is carried out. A combination of techniques that measure aggregate size and density fractions was used by Hassink, (1995) to measure the complex interactions between SOM and soil degradation. Silica gel, which is nontoxic and does not penetrate the SOM, prevents the artificially high densities obtained when utilizing organic or other inorganic high density fluids. Hassink disrupted the largest aggregates by forcing soil through a 250 μm sieve. The materials passing the 250 μm sieve but retained on a 150 μm sieve were separated into three fractions by flotation in silica gel of > 1.37, 1.37 to 1.13, and < 1.13 g cm⁻³ (Meijboom et al., 1995). The material between 20 to 150 μm in size formed the fourth fraction, the < 20 μm size the fifth fraction. Subjection of the fractions to long-term incubation showed the < 1.13 g cm⁻³ material of the 150 to 250 μm sized fraction decomposed the most rapidly. The 1.13 to 1.37 density material in turn was more decomposable than the > 1.37 g cm⁻³ fractions. The 20 to 150 μm sized fraction decomposed more slowly and the < 20 μm sized materials decomposed the most slowly. This fractionation was said to give meaningful fractions that could be utilized directly for modeling purposes. However, the slowest rates obtained for any fraction were higher or similar to the decomposition rates found for the totally undisturbed soil. They were much higher than rates that would be calculated when utilizing ¹⁴C dating for a soil such as this. We suggest that these fractionations still include a mixture of old and young C. Physical fractionation by itself does not give the necessary pool sizes and flux rates required for determining the effects of the SOM relative to degradation. It does supply very meaningful estimates of the effect of physical protection on the decomposition rates of the more active fractions.

IV. The Use of Tracers in Soil Organic Matter Studies

The use of tracer-C and N in conjunction with long-term field plots and physical and chemical fractionation have made it possible to determine the pools and fluxes of SOM. Soil organic matter is a complex series of related molecules with varying physical structures, molecular weight, chemical components and functional groups. It is associated with itself, with clays and biota in various parts of an aggregate as well as often residing on the soil surface (Aiken et al., 1985; Stevenson, 1994). Tracers allow the investigator to identify specific components and measure their turnover rates as well as identifying nutrient uptake by plants and soil organisms. The tracers include ^{14}C naturally occurring in the atmosphere due to cosmic irradiation of atmospheric N. This ^{14}C , utilized in carbon dating, has been augmented during the last 40 years by the ^{14}C produced during thermonuclear bomb testing (Harkness et al., 1991). Also available are ^{14}C -enriched compounds produced either as pure chemicals or by growth of plants in a $^{14}\text{CO}_2$ atmosphere. Other radioactive nutrient elements include ^{32}P and ^{33}P and ^{35}S . Radioactive isotopes can have limited half-lives such as is the case for ^{11}C and ^{13}N . More often the limitation to their use is the fact that they constitute health hazards and require strict licensing. Their advantage, however, is the great sensitivity with which they can be measured such that very low concentrations can be utilized.

Carbon dating is based on the principle that the ^{14}C with a half-life of 5568 years, once incorporated into plants, suffers little further discrimination as SOM is formed. The use of modern carbon dating techniques such as tandem accelerator mass spectrometers makes possible the characterization of milligram levels of soil (Coleman and Fry, 1991). However, carbon dating is expensive and prone to contamination from enriched C sources. It therefore is used most often to obtain background information. Carbon-13 measurements are most often used to measure turnover rates under a wide variety of conditions.

The work with tracers at Rothamsted, UK is an example of the power of the tracer approach. The Rothamsted plots have a 150 year record of crop inputs and management practice. Carbon dating was utilized to characterize the turnover time of the resistant fractions. The addition of ^{14}C -enriched plant residues determined the more rapid decomposition rates. Bomb C inputs were utilized to help verify C addition rates to the soils. The data were used to develop a mathematical model that described the various decomposition rates and inputs (Jenkinson and Rayner, 1977). The flux of a C through five SOM pools included: 1) decomposing plant materials, 2) resistant plant materials, 3) soil microbial biomass 4) physically protected, and 5) chemically stabilized organic matter. Ages of these pools ranged from several months for plant materials to several thousand years for physically and chemically protected SOM fractions. Paul and van Veen (1978) in their review of tracer use in SOM studies, applied the physical protection concept to both resistant and more labile SOM components.

Carbon dating has recently been applied to a broad range of North American soils by Paul et al. (1996). This verified earlier work that showed the carbon dating age of SOM to be dependent on landscape position, cultivation and soil forming

processes (Cambell et al., 1967; Martel and Paul, 1974; Anderson and Paul, 1984). Upper, often eroded portions of landscapes tended to be old; the lower, moist depressions tended to have young SOM. Paul et al. (1996) also verified earlier work by Scharpenseel (1993) that the ^{14}C age increased dramatically with soil depth. Acid hydrolysis in the work of Martel and Paul (1974) was shown to separate old from younger soil organic fractions. Figure 3 shows the effect of acid hydrolysis, the ^{14}C age and the calculated decomposition rate constants of a grassland soil in Colorado after 84 years of cultivation. The native soil had 13.6 g C kg^{-1} soil but degraded to 8.4 g C kg^{-1} after extended cultivation. The radiocarbon ages of the surface soil were modern for the native soil and 1300 years for the cultivated. The nonhydrolyzable C had an age of 2900 years in the native and 3300 in the cultivated. Thus the total soil SOM had much more rapid decomposition rates at the surface than did the nonhydrolyzable residue; the two rates approached each other at depth (Figure 3c).

Carbon dating best characterizes the resistant fractions of SOM. Acid hydrolysis together with density gradient measurements and carbon dating were also found by Trumbore et al. (1993) to separate SOM fractions into approximately equal amounts of C with residence times of ten, hundred, and thousand years respectively. Acid hydrolysis removed the enriched fractions of temperate but not of tropical soils where most of the organic matter may be in the more labile form. The stable isotopes of nutrient elements ^{13}C , ^{15}N and ^{34}S can now be easily be measured with automated mass spectrometry. This together with the occurrence of low levels of these isotopes produced by natural biological discrimination has made the stable isotopes the method of choice especially where field work in isolated regions needs to be conducted. The stable isotope ^{13}C occurs naturally in atmospheric CO_2 at a concentration of 1.1%. During photosynthesis, C_3 plants incorporate less $^{13}\text{CO}_2$ than do C_4 plants (Vogel et al., 1980; O'Leary, 1981; Boutton, 1991). The CO_2 of the atmosphere has a ^{13}C content of -7 ‰ expressed relative to a limestone standard. Plants with a C_4 photosynthetic pathway such as sorghum, warm season grasses and maize have a $\delta^{13}\text{C}$ of -11 to -14 ‰ . Plants with a C_3 pathway such as most trees, cool season grasses, wheat and rice have a $\delta^{13}\text{C}$ of -26 to -28 ‰ . The interchange of C_3 and C_4 plants in the field under a long-term basis, i.e., 30 to 50 years or even for 1 to 2 years, provides a usable signal that can be measured with today's mass spectrometers (Martin et al., 1990).

The ^{13}C discrimination in plants provides a signature that with very little change shows up in the SOM (Balesdent et al., 1987; Boutton, 1991). Balesdent et al. (1988) utilized ^{13}C on the long-term plots in Sandborn, Missouri to estimate the age of the stable SOM as 600 years after 100 years cultivation. The age of >600 years was later verified by carbon dating (Hsieh, 1992). The data in Figure 4 for ^{13}C analysis complement the earlier ^{14}C dating of the Akron, CO site (Figure 3). Cultivation from 1909 to 1992 resulted in a 50% drop in the SOM of the top 15 cm when expressed on an area basis after taking bulk density into account. The drop at the 15 to 30 cm level was less pronounced. The use of ^{13}C made it possible to measure the contribution of wheat residues to the SOM during the 84 years of wheat-fallow cropping (Follett et al., 1996). Only 5% of the plant C added as litter, roots and weeds over this time period was accounted for as SOM.

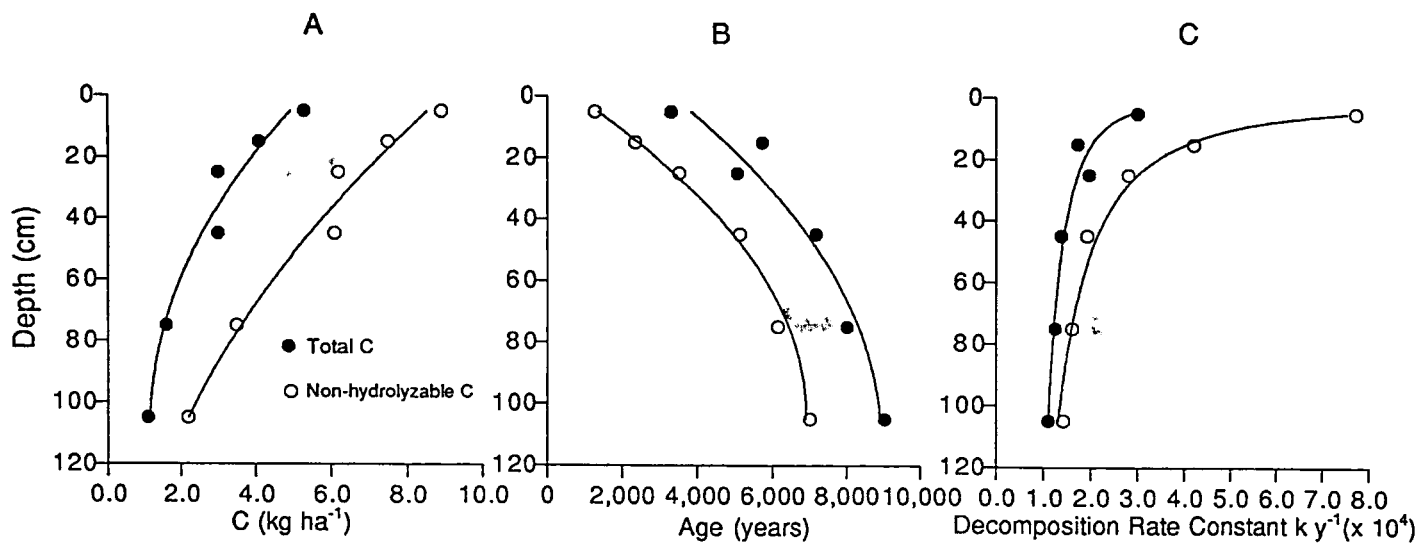


Figure 3. The relations between: A) soil organic matter content, B) ¹⁴C age and C) decomposition rate constants and depth in the Akron, Colorado cultivated soil. (From Paul et al., 1996)

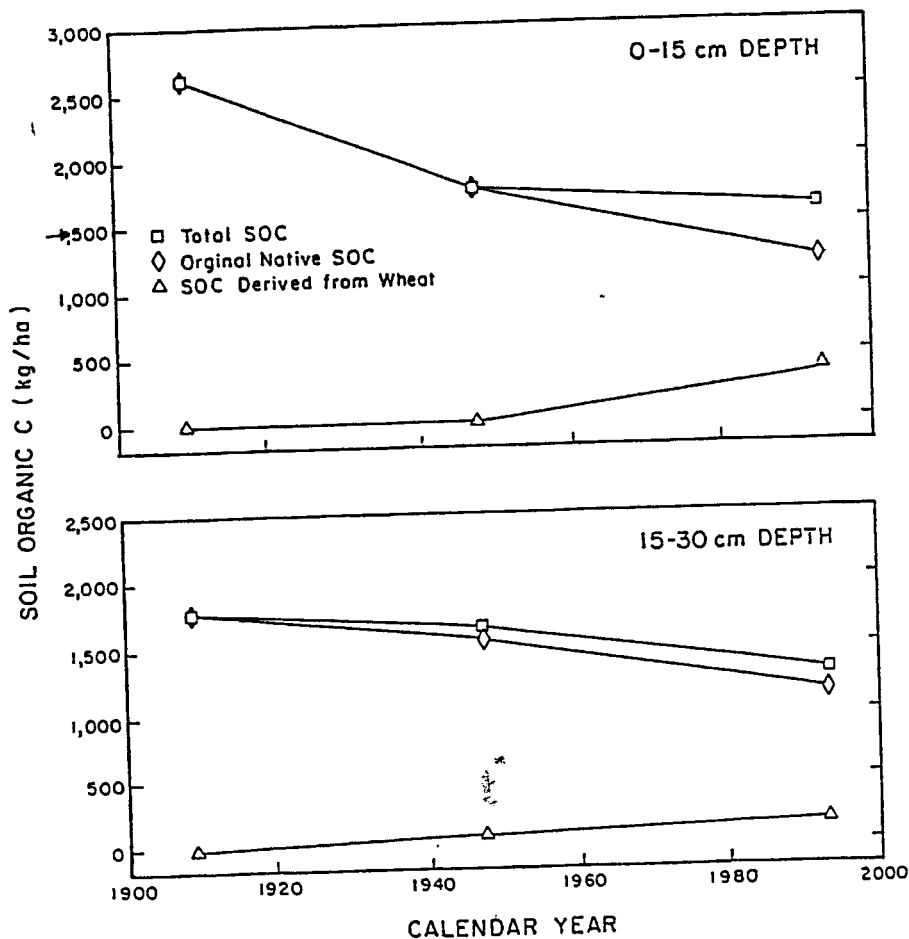


Figure 4. Total soil organic C, original native soil organic C, and soil organic C derived from wheat from 1909 to 1993 at Akron, Colorado. (From Paul et al., 1996.)

Long-term mineralization of the soil C by microorganisms in extended laboratory incubations can be utilized to determine the in situ availability of C and N to the microbiota of the soil (Stanford and Smith, 1972). The percentage of total organic C evolved as CO_2 has been used to compare soil under different management conditions (Collins et al., 1992; Carter and Rennie, 1982; Campbell et al., 1989). The analysis of the CO_2 release rates makes it possible to derive estimates for functional pool sizes and their turnover rates especially if combined with acid

Table 2. Dynamics of corn-soybean field in Michigan as determined by ^{14}C dating and ^{13}C analysis and extended mineralization

	-----Soil depth (cm)-----		
	0-20	25-50	50-100
Soil C g kg ⁻¹ (C _T)	12.1	2.4	1.3
^{13}C ‰	23.1	22.0	22.7
% C from crop residues ^a	32.7	37.5	24.0
C ₁ /C _T (%)	3.2	3.9	5.6
MRT C ₁ (d) ^b	52.2	27.0	39.7
C ₂ /C _T (%)	67.3	63.4	69.3
MRT C ₂ (y)	16.8	11.6	33.4
C ₃ /C _T (‰) ^c	30.0	32.7	25.1
MRT C ₃ (y)	1435	--	--

^a % from crop = $(\delta^{13}\text{C}$ cropped soil - $\delta^{13}\text{C}$ native soil) / ($\delta^{13}\text{C}$ crop residue - $\delta^{13}\text{C}$ native soil).

^b C₁ and C₂ determined by curve fitting of CO₂ from extended laboratory incubation. MRT = (1/k) k multiplied by a factor of 2 to transform laboratory data to field data.

^c C₃ and MRT C₃ determined by carbon dating.

(Adapted from Collins et al., 1996.)⁴

hydrolysis and carbon dating to determine the turnover of the old pool. The old resistant pools contribute little to CO₂ evolution even in extended incubation but are very important from a soil C storage and stabilization standpoint. The release of N during extended incubations gives important information on the mineralization of this nutrient. Stanford and Smith (1972) used leaching experiments in which the nitrate was removed every few weeks. A number of other investigators have used static incubation where soil subsamples were removed from larger samples during the extended incubation. The possibilities of denitrification occurring under these conditions must be considered in the interpretation of the curves obtained.

A corn-soybean field (Table 2) cultivated for over 100 years in Michigan had approximately 60% as much SOM as the native deciduous forest. The ^{13}C analyses showed that only 33% of today's surface soil came from the agricultural crop. Carbon dating showed the resistant pool, accounting for 30% of the total C, to be 1435 years in age. This corroborated the ^{13}C measurement showing that most of soil C still came from the original forest soil even after 100 years of cultivation. Extended incubation and curve fitting (Collins et al., 1996) showed a turnover time of 52 days for the most active pool comprising 3% of the soil C. The intermediate pool, representing 67% of the C released CO₂ equivalent to a turnover time of 12 years.

V. Summary

A range of techniques are available to measure the role of organic matter in soil degradation and rejuvenation. The most straightforward, and probably the most meaningful, measurement is the determination of the total C, S, and P content of the soils. This must be conducted at various depths together with bulk density measurements. Fractionations and characterization of SOM have given the necessary background to show that the SOM in the coarse sand fractions is composed of partially decomposed material with high turnover rates and thus the greatest potential for contribution of soil nutrients. The materials associated with silt size fractions often tend to be the most stable. The clay fraction, because it also contains some of the active microbial metabolites, can be quite young and show high turnover rates. Clay and silt fractions stabilize microbial biomass and microbial products as well as participating in aggregation. It therefore is not surprising that clay content in many soils is correlated to SOM content and helps control turnover rates.

The determination of pool sizes and fluxes with tracers in conjunction with long-term incubation and SOM fractionation is now possible. Measurements of undecomposed-partially decomposed plant materials and associated biota (the light fraction or the POM) as well as the measurements of aggregation when combined with long-term field studies can provide the background information necessary for the development of concepts relative to proper management of SOM. This natural resource is renewable and does not have to be degraded. Being renewable it can continue to supply the food and fiber of a growing world population without degradation if removed nutrients are replaced by biological or chemical means and if erosion and salinity are controlled.

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