CHAPTER 24

Establishing the Pool Sizes and Fluxes in CO_2 Emissions from Soil Organic Matter Turnover

E.A. Paul, W.R. Horwath, D.Harris, R. Follett, S.W. Leavitt, B.A. Kimball, and K. Pregitzer

I. Introduction

Soil organic matter (SOM) is known to act both as a source and a sink in global CO_2 cycles. Its role in the biological productivity of managed and unmanaged systems is equally well recognized. Central to any rigorous examination of the above roles are questions relating to what is there, and what happens to it? This means that we must ask what are the pools and the fluxes involved? The degradation of the plant components and plant respiration are an integral component of these fluxes. These are difficult to separate from the soil components. Most soils contain at least 5-20% of their organic carbon (C) as partially decomposed particulate residues (Cambardella and Elliott 1992) with native grassland having >30% of its SOM in this fraction. The method of determination and the C:N ratio of this fraction indicates that it contains partly decomposed plant residues, fine roots and extensive numbers of associated microorganisms. This fraction is closely related to the short term mineralization of C and N (Janzen et al., 1992).

Most present models consider plant residues as a separate group of pools. Difficulties arise in the characterization of partially degraded residues and their associated microflora. The extent, turnover and incorporation of soil exudates into humic constituents also is poorly defined. They vary widely depending on plant age, type, and abiotic conditions. The average result from many ^{14}C studies shows that 1 to 7% of net primary plant productivity is exuded into soil. The turnover of roots is not included in this calculation. This produces much more carbon than the exudates. Studies of wheat root distribution by washing have indicated that root weights comprise approximately 12% of the stover (Wilhelm et al., 1982). Other studies have shown larger amounts, e.g. 50% of total residues (Buyanovsky and Wagner, 1986).

The most useful techniques for testing and summarizing the various soil pool sizes and their dynamics involve mathematical models. They incorporate moisture and temperature as controlling factors in both plant productivity and SOM decomposition. They also reflect the effects of the amount of clay on stabilization of SOM and soil microbial biomass (SMB). Soil depth is taken into account and the majority of the models include information on plant nutrient and water uptake. There is some but not enough integration with landscape, and plant root development, nor is there an adequate understanding of the role of soil aggregates in the stabilization of plant residues, microbial biomass, and soil humic constituents. These models have greatly helped in the general management of soils and in initial approximations of the role of SOM in global C cycles. We have to refine our models if they are going to answer the important societal questions concerning global change, manure applications, fertilizer N additions, and atmospheric pollution effects on vegetation and water quality. This refinement requires better estimates of SOM fractions and their dynamics.

ISBN 1-56670-118-X/95/$0.00+.50

II. Pool Sizes and Fluxes in Soil Organic Matter Turnover

The first research requirement in determining pool sizes and fluxes is a well-characterized site with measured abiotic characteristics, SOM levels and above and beneath ground residue inputs. Since man's activities play such a profound role, whether it be in slash/burn agriculture, clear-cut timber harvesting, or tillage, these sites should be representative of both the major climatic areas and management procedures involved. The *Global Directory of Long-Term Agronomic Experiments* (Steiner and Herdt, 1993) helps identify some of these sites. Other sites include the NSF-LTER programs and the EPA study of agronomic management in the U.S. (Elliott et al., 1994).

Identification of the various pools of SOM is the second requirement. Soil organic matter is a complex series of related but different molecules with varying physical structure, molecular weight, chemical structures and functional groups. It is associated with itself, clays, and microorganisms and is within and between aggregates as well as residing on the soil surface and having been leached to greater depth. Therefore, it is not surprising that no one physical-chemical fractionation technique will separate the components into meaningful biological fractions. Of the many that have been tried, a few can be conducted with fairly straight forward equipment in a reasonable time period. These include the measurement of microbial biomass, long-term mineralization of C and N, the use of acid hydrolysis, the separation of aggregates on a size and stability basis, the measurement of particulate SOM, the determination of the light fraction, and the measurement of organic constituents associated with soil separates. Newer techniques, such as the use of XRD-4 resin, hold promise for separating polysacharides and possibly proteinaceous materials (Wilson, 1991).

The last requirement in determining SOM dynamics relative to CO_2 fluxes is that of measuring the rates of addition of substrate and turnover of the various pools. Tracers allow one to identify specific fractions, measure their turnover rate, mineralization-immobilization processes, plant uptake, and microbial growth. The use of tracers, therefore, is vital in identifying pool sizes and determining at what speed materials flow in and out of them, e.g. the fluxes. The most useful tracers include ^{13}C in both enriched ^{13}C experiments and those that use the isotopic signal provided by the discrimination that occurs in the C_3 (Calvin cycle) and C_4 (Hatch-Slack cycle) plants grown on the same site. Alternatively, ^{14}C can be used in the enriched form or as that occurring in the atmosphere. Thermonuclear bomb testing produced a ^{14}C signal during the period from 1960 to 1985 that is useful in following medium term SOM dynamics, especially if stored reference samples are available.

The availability of accurate, automated mass spectrometers for stable isotopes, and tandem accelerator mass spectrometers for naturally occurring ^{14}C now make it possible to process numerous samples containing microgram quantities of C. The use of enriched CO_2 in experiments to measure the effects of global climate change either in enclosures or unenclosed (FACE) can provide both a ^{14}C and a ^{13}C signal.

III. The Use of Carbon Dating

Carbon dates have looked at soil development, chrono sequences, landscape positions, the effects of cultivation, and of soil forming processes (Anderson and Paul, 1984; Campbell, et al., 1967; Coleman and Fry, 1992). Soils differ greatly depending on slope, aspect, vegetation type, and position in the landscape. Related differences in parent materials and moisture availability directly affect ecosystem productivity and agronomic yield. Techniques now available for the farming of fields on a soil type-landscape basis make it possible to manage individual soil types and landscape positions. We should not expect SOM to be similar across such a range of sites. The example in Table 1 shows that the SOM content of the upper position of a catena in Saskatchewan is lower than that of the more moist toe slopes and that there are buried horizons in the bottom slope position. This shows the effects of deposition from erosion. The soils at the top of the slope have a C age of 500 yr, but the surface of the moist, lowest position in the landscape dates modern. The reason for this is not well understood. If soil is eroded from the upper to the lower slope positions, C dating should reflect this. One must conclude that the more recent C is being transported to an area that has a generally higher turnover rate.

Agriculture, and to some extent forestry research, has been dominated in the last 60 yr by the measurement of the decrease in SOM contents following disturbance. It has been easier to follow and model the drop with disturbance than it will be for us to accurately predict the buildup of SOM as we change our management techniques in an increased CO_2 environment. The example given in Table 2 shows the

Table 1. The effect of landscape on soil organic matter contents and radiocarbon age; Oxbow Catena Saskatchewan

	Top slope	Mid slope	Toe slope	Depression
Ap Horizon cm	0-10 cm	0-15 cm	0-15	0-17.5
C%	3.3	3.6	4.6	3.9
Horizon	Radiocarbon age BP			
Ap	500	270	216	Modern
Ae 17-22.5 cm			685	-
B_T 22-37 cm			930	700
B_{Tg} 60-85 cm				4870

(Adapted from Martel and Paul, 1974.)

Table 2. The effect of cultivation on the radiocarbon age; 0-10 cm Oxbow Saskatchewan

	Virgin	15 yr cultivation	60 yr cultivation
C%	5.4	3.5	2.2
N%	0.45	0.34	0.23
Radiocarbon age BP	250 ± 65	295 ± 75	170 ± 60

(Adapted from Martel and Paul, 1974.)

Table 3. The effect of bomb ^{14}C on the radiocarbon age (Indian Head, Saskatchewan).

	Proportion of total	1963		1978	
Fraction	%	pmC	age yr	pmC	age yr
Soil	100	79	1900	83	1500
Nonhydrolyzable org-C	50	70	2800	77	2100

(Adapted from Anderson and Paul, 1984)

traditional 50% drop in SOM content on a weight basis. Changes in bulk density and measurement of SOM in the total profile must be taken into account, and the drop on a soil area basis is not as great as that shown when expresed as a percentage. Water erosion transports much of the soil into depressions in the same field from which it was originally displaced. Wind erosion usually is more extensive but also fills in low areas in a rolling landscape and areas with adjacent vegetation. The extent to which C deposited in the low areas is stored or lost to the atmosphere by decomposition is estimated at 10-50% of that transported.

In the example given, the radio C age of the virgin grassland soil is 250 yr. This increased through 60 yr of cultivation to 710 y showing the effects of the removal, by decomposition, of the more recently added materials. Bomb C effects are shown in Table 3. The samples collected in 1963 preceded the maximum incorporation of the ^{14}C into SOM. The overall soil at this site dated 1900 yr. This was decreased to 1500 y in 1978 as the thermonuclear-produced ^{14}C was incorporated into the SOM. Nonhydrolyzable organic C, comprising 50% of the total was very much older in the 1963 sample, it also was influenced, to some extent, by the effects of bomb C, for in 1978 this fraction had dropped an equivalent number of years. This is not surprising; the hydrolysis of modern, fairly high ligniferous plant residues also yields approximately 50% of the material as a nonhydrolyzable residue. The acid hydrolysis residue does not drop appreciably with cultivation showing that chemical recalcitrance is not the only factor causing these great ages. This led Martel and Paul in 1974 to suggest that physical protection played a major role in SOM dynamics. The role

Table 4. The effect of soil separates on radiocarbon ages; Melfort, Saskatchewan

Fraction	Proportion of organic C%	pm C	Age
Coarse silt	25	91	800 ± 50
Fine silt	29	89	965 ± 50
Coarse clay	31	86	1255 ± 60
Fine clay	8	98	170 ± 50

(Adapted from Anderson and Paul, 1984.)

Table 5. Comparison of the distribution of C and ^{14}C using radiocarbon dating and ^{14}C incorporation into microbial biomass in Western Canadian Wheat soil

	Radio-carbon age	Soil C %	Radio-carbon C %	Microbial ^{14}C %
Total soil	350 yr BP	100	100	100
Light fraction	240 pmC	6	8	5
0.5 N HCl hydro-lysate	107 pmC	35	37	50
6.0 N HCl hydro-lysate	161 pmC	22	24	20
6.0 N residue	1765 yr BP	37	31	25
NaOH extract	1900 yr BP	27	22	16
H_2O dispersion	1800 yr BP	3	3	3
Residue	1330 yr BP	7	6	6

(Adapted from Martel and Paul, 1974).

of microaggregates as outlined by Waters and Oades (1991), and Cambardella and Elliott (1992, 1993) further supports the concept of aggregation in SOM stabilization.

Numerous studies have shown that the fractionation of soil separates into sand, silt, and clay give significantly different C dates. The fine silt and course clay not only contain the largest amounts of C, they also contain the oldest C (Table 4). Fine clay is much younger. The possibility of contamination of the fine clay by microbial constituents dispersed during the fractionation procedure must be considered. The interaction of the soil separates with the soil aggregates must be further investigated, e.g. it is said that the microaggregates are the most stable. The tracer content of silt, course clay and fine clay in microaggregates and macroaggregates needs to be determined.

Table 5 shows the results of another method of fractionation. The ^{14}C distribution was measured by two techniques, by radio C dating of naturally occurring ^{14}C and by scintillation counting of the ^{14}C that had been added as glucose and was then incorporated into microbial biomass. The light fraction floated in a dense liquid represents decomposing residues and associated microorganisms. This fraction is equivalent to the particulate organic matter of other studies. In this case, the light fraction had a ^{14}C content slightly greater than twice that of the pre-bomb C standard. It represented 8% of the naturally occurring ^{14}C as determined by C dating; however, it represented 5% of the material where the ^{14}C was derived from microbial biomass, showing that significant quantities of the biomass are associated with the plant residues. Multistep acid hydrolysis has been shown to be an effective way of separating labile constituents. The high occurrence of labile C in the radio C distribution and when incorporated as microbial C shows the effectiveness of this technique; 71% of the ^{14}C introduced as microbial biomass was removed by hydrolysis. The nonhydrolyzable residue in the C dated fraction shows great age. While representing 37% of the total C, it only represents 31% of the ^{14}C. That hydrolysis is not completely effective as a separation technique

Table 6. Distribution of ^{14}C (%) in a poplar-soil system two weeks and one year after labelling in September. Standard errors of the mean shown in parenthesis

	Day 14	Day 350	From litter[a]	Total
Leaves/litter	20 (1.5)	0.12	5.5 (0.25)	
Stems/branches	28 (4.6)	13 (1.6)		
Roots < 0.5 mm	7 (1.8)	5 (0.45)		
Roots > 0.5 mm	31 (3.2)	11 (1.0)		
Root-soil respiration	12 (1.50)			50.6
Microbial biomass	1.3 (0.2)	0.5 (0.8)	0.18	
Soil	0.7 (0.5)	3.7 (1.8)	3.4	
Total	100	34.3	9.1	43.4

[a] Derived from separately placed litter after 325 days.
(From Horwath, 1993).

is indicated by the fact that 25% of the recently incorporated microbial ^{14}C is included in this fraction. Further fractionation of the soil by sodium hydroxide extraction, followed by water dispersion leaves a residue of only 7%, which upon microscopic examination was shown to be bits of charcoal, insect skeletons, etc. It is not as old as the peptizable fractions.

IV. The Use of Enriched ^{14}C in Determining Plant Inputs

Enriched ^{14}C is particularly valuable in measuring both direct and indirect plant inputs such as exudates and root turnover. The example in Table 6 shows the results of labeling 3m high hybrid poplars with ^{14}C in September. This was followed by a 14 d equilibration period before analysis. Forty-eight percent of the labelled C remained above ground in leaves, stems, and branches. Fifty-two percent was translocated underground. The majority of this was found in large roots. Respiration during the 14 d period, while the ^{14}C was being transformed to longer-life compounds within the plant accounted for 12% of the net ^{14}C fixation. The soil microbial biomass plus the associated SOM represented 2% of the net fixation. If one assumes a 35% efficiency in the incorporation of root derived substrates into microbial biomass and microbial products, 5.7% of the plant fixed ^{14}C is accounted for. Fine roots did not lose ^{14}C, but stems and branches did. We concluded that fine root growth was derived from previously stored C, rather than directly from photosynthate. In the first year after labelling, the tree components plus the leaf litter lost 56.6% of the ^{14}C. During this period the microbial biomass showed a 6-fold dilution of its ^{14}C contents, indicating the turnover rate of this component.

The leaf litter from the labelled trees was placed onto soil of unlabelled trees; the labelled trees received the unlabelled litter. This replacement experiment made it possible to differentiate the leaf derived from the root derived C. The litter placed on the surface lost 73% of its label; 17% was translocated to the soil and associated microbial biomass. At this site the leaf litter plays a significantly larger role in forming soil microbial biomass but the proportion contributed to the soil itself is approximately equal to that of roots.

The September labelling (Table 6) moved much more C beneath ground than did a July labelling (data not shown). In addition, the material from the July labelling did not decompose as rapidly as that from the September labelling. This indicates that both the proportion of above and beneath ground C and the quality of the C changes with season. Significant misinterpretations can occur if one assumes that a uniform substrate is added to the soil over the complete growing season.

V. The Use of ^{13}C

The stable isotope of C representing 1.1% of the atmosphere will receive increasing use because of safety concerns with enriched ^{14}C and the availability of accurate, sensitive, reasonably priced automatic mass spectrometers. The signal provided by the differential fractionation of $^{13}CO_2$ by C_3 or C_4 plants will be increasingly utilized. This technique is very useful for a number of reasons: 1) many fields can be sampled

and multiple replicates can be utilized because of the reasonable cost of the analysis: 2) the replacement of C_4 grassland with C_3 wheat or soybeans; the growth of C_4 corn on C_3 grasslands in cooler climates and the substitution of C_4 plants onto C_3 forest vegetation in the tropics provides an extensive range of sites. 3) the length of time since cultivation of these sites ranging from 10 to 110 y is in the period of time where enriched ^{14}C experiments and naturally occurring C dates are most difficult to use.

There is an excellent range of ^{13}C:^{12}C ratios extending across the plains of North America where C_3 wheat is being grown on largely C_4 grasslands. On some of these sites wheat is now being replaced with maize. We thus have an identifiable ^{13}C signal moving back toward the signal provided by the original C_4 vegetation. In the eastern U.S., C_4 plants such as maize and sorghum grown on former forest vegetation provide equally good sites for analysis. The ^{13}C signal was used by Balesdent et al. in 1988 to show that in the Sanborn long-term plots at least 50% of the original tall grass prairie C remained after 100 yr of cultivation.

The ^{13}C signal has been effectively utilized in the tropics to differentiate soils from largely C_4 savannahs and the adjoining C_3 forests. Bonde et al., 1992, found that 12 yr of cropping of a forest soil to C_4 grasses resulted in a 90% turnover of the forest SOM in the sand fraction. In the clay fraction, only 40% of the C was from the C_4 grasses with the remainder being from the original forest. This represented turnover times of 4, 6, and 59 yr for the sand, silt, and clay fractions of soil, respectively. Other studies in the tropics have shown the ^{13}C:^{12}C ratios to change with depth. The ^{13}C signal of a podzol B horizon was from an ancient forest rather than the present savannah (Schwartz et al., 1986).

VI. Estimation of the Kinetics of Decomposition During Long-Term Incubation

The utilized chemical and physical fractionation techniques have limitations in adequately separating SOM fractions for analysis of their dynamics. An alternative, best used in conjunction with fractionation techniques, is to allow the microorganisms and their enzymes to identify the biologically active soil C. Extended incubations (100-250 d) have been found to be sensitive indicators of management effects. Table 7 illustrates that a grassland developed on a previously forested site with no cultivation during recorded time periods contained more C than cultivated sites and released significant levels of CO_2. At the time of sampling the reversion field had been allowed to revert to an old field status for 4 yr after a significant length of time in a corn-soybean rotation. Although total C contents were not yet different, the reversion treatment evolved 930 μg C g^{-1} 200 d^{-1} vs. 560 μg C g^{-1} 200 d^{-1} for the associated corn-soybean rotation. The C mineralized over the 200 d represented 7.3% of the total C in the grassland, 9.8% in the reversion and 5.9% in the corn-soybean plots, showing the rapid buildup of soil C within the reversion treatment.

Table 7 also shows that microbial biomass can be a sensitive indicator of changes, although it represents a smaller percentage of the total C than does the C mineralized in 200 d. The reversion treatment had the largest percentage of its C present as microbial biomass. The mineralization curves were fitted to the sum of two first-order equations. Carbon dating has shown that 50% of the C in most soils is very old, i.e. greater than 500 yr, therefore, it should not contribute significant amounts of C to a mineralization such as this. The definition of the second pool as $C_2 = (C_T/2)-C_1$ made it possible to obtain a good fit with the restricted number of points.

CO_2 analysis does not give an exact figure of the available C since microbial growth utilizes some of the substrate attacked during decomposition. The pool C_1 was corrected for microbial growth by utilizing growth efficiency calculations (Paul and Clark, 1989). In this study, as in many others, the decomposition rate constant is fairly consistent, but the pool size, C_1, was affected by treatment. Even though the incubation was only carried out for 200 d, kinetic analysis showed that the mean residence time of the C_2 pool ranged from 5.3 to 9.5 yr with the reversion treatment being the most active and the corn-soybean being the least active.

The use of C mineralization makes it possible to anlayse treatments from many long term experiments without the use of tracers (Elliott et al., 1993). Conducting a C mineralization study on a previously labelled sample, provides even more information. This can be either from C_3-C_4 vegetation or from enriched ^{14}C or ^{13}C studies. This type of analysis also makes it possible to further calculate pool sizes by isotope dilution calculations (Voroney et al., 1991).

Analysis by Horwath of the poplar plots (Table 6) shows that 60 d after the addition of ^{14}C the most active pool (C_1) comprised 1.4 to 3% of the total C and had a mean residence time of 25 to 50 d. Extended

Table 7. Kinetic analysis of CO_2 mineralization curves for three treatments of the KBS-LTER (Michigan State University) $C_m = C_1 (1 - e^{-k1t}) + C_2 (1 - e^{-k2t})$; $C_2 = (C_T)/2 - C_1$: standard error of the mean shown in parenthesis

	Grassland	Reversion	Corn soybeans
Total C (C_T) $\mu g\ g^{-1}$	15,000	9,500	9,500
C mineralized (C_m) $\mu g\ C\ g^{-1}\ 200\ d^{-1}$	1,100	930	560
C_m/C_T %	7.3	9.8	5.9
Microbial C $\mu g\ g^{-1}$	345 (51)	251 (38)	141 (22)
Microbial C/C_T %	2.3	2.6	1.5
Pool size (C_1) $\mu g\ g^{-1}$	623 (103)	404 (42)	170 (24)
Mineralization kinetics			
C_1^a/C_T%	4.1	4.2	1.8
$K_1\ d^{-1}$	0.022	0.029	0.024
$K_2\ d^{-1}$	0.00035	0.00052	0.00029
K_1 MRT d	45	34	41
K_2 MRT y	7.8	5.3	9.5
R_2	0.997	0.999	0.999

[a]C_1 corrected for microbial growth

mineralization studies of the SOM one year after labelling showed that the ^{14}C was now in much closer equilibration with the total soil C. The mean residence time of the ^{14}C derived CO_2 in the second pool now approximated 10 yr verses 3 yr in the sampling on 60 d after labelling.

VII. Enriched CO_2 Studies.

The use of enhanced CO_2 in experiments to measure the effect of future climate change represents a unique opportunity for determination of pool sizes and fluxes of SOM. Most of the CO_2 that is utilized in these experiments either in enclosed chambers, in enclosures open to the atmosphere at the top or completely free as in the Free Atmosphere Carbon Enrichment Experiments (FACE) comes from petroleum derived materials. The tank CO_2 has both a ^{14}C (0 PMC) and a ^{13}C signal of about -30‰. When mixed with air this becomes -10‰. The results from the FACE experiments in Arizona in which CO_2 was applied to cotton plants grown in the field for a 3 y period is shown in Figure 1. Cotton plants grown in a field with a normal atmosphere were compared to those with FACE air at 550 μmol mol^{-1} of CO_2. The FACE air had a δ ^{13}C content of -17‰ and a ^{14}C of 75 pmC compared to the control of -7.5‰δ ^{13}C and 115 pmC for ^{14}C. This resulted in control plants of -26.7‰δ ^{13}C and 111.8 pmC. The FACE plants had a δ ^{13}C content of -38.4‰ and a ^{14}C of 72.1 pmC. The soil, after 3 y of cotton growth, was found to be labelled at -23.8‰ for the FACE compared to 22.4‰ for the control.

The associated ^{14}C signal was used to obtain further information on the dynamics of plant residues (Table 8). SOM left after flotation of the soil in the control plot at a specific gravity of 1.2 had an average age of 250 yr BP. Careful microscopic picking of the plant residues raised the total soil to a 1000 yr BP. On hydrolysis, this gave a residue of 2400 y BP and a supernatant of 250 y BP. The equivalent FACE experiments show that the incorporation of FACE residue depleted of ^{14}C changed the age to 950 yr BP. This was essentially unaltered by picking for it then approximated the picked sample from the control soil at 1000 yr BP. The HCl residue of the FACE plots was again very old, not being different than the control. This shows that after 3 yr of cotton growth most of the label was still in identifiable plant residues that could be removed by flotation and careful picking. The carbonates, at 2500 yr BP were unaffected by the increased CO_2 levels and their tracer signals. Further equilibration in the field will make it possible to follow the movement of both signals into the various SOM fractions of this site.

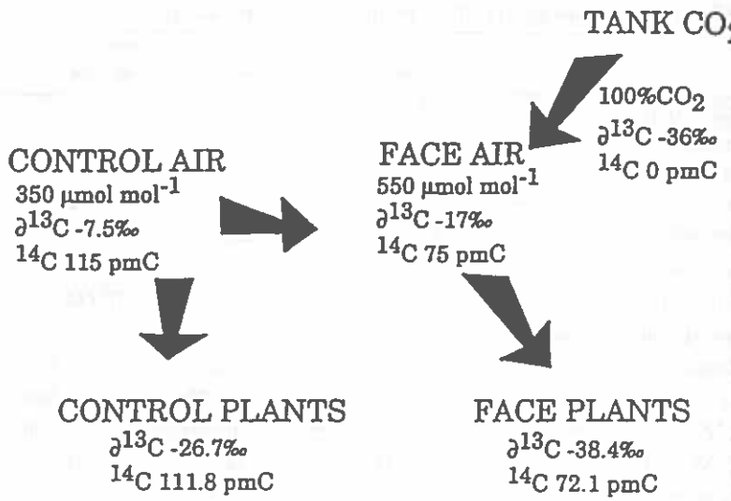

Figure 1. Summary of ^{14}C and $\partial^{13}C$ of FACE experiment air and cotton plants. (Adapted from Leavitt et al. 1992.)

Table 8. Effect of incorporation of geologically dead carbon into soil of FACE study in Arizona

	Control		FACE	
	pmC	yBP	pmC	yBP
Floated soil	0.96	250	0.89	1000
Floated &picked	0.89	1000	0.88	950
HCl-residue	0.72	2400	0.75	2300
HCl-supernatant	0.96	250	1.00	Modern
Carbonates	0.73	2500	0.73	2500

(Adapted from Leavitt et al. 1993.)

VIII. Summary and Conclusions

We have attempted to show, using examples from a number of sites, that the prerequisites for adequately determining the dynamics of SOM relative to global CO_2 fluxes are now available. Field sites with long-term management and abiotic and biological productivity measurements can provide the necessary soil samples. The use of added ^{14}C and ^{13}C either as added plant residues in enclosures or in FACE experiments gives very useful information. Modern techniques now make it possible to obtain SOM pool turnover rates without having to resort to expensive and lengthy field tracer experiments. Carbon dating using naturally occuring ^{14}C and use of the ^{13}C signal supplied by photosynthetic discrimination can be effectively used for determination of root versus soil respiration, plant residue decomposition rates, the source of mineralized CO_2, and the turnover rate of SOM fractions. Both the Saskatchewan and the Arizona data show a significant resistant fraction that is thousands of years old. More data must be obtained on intermediate sites, especially those that are affected by variations in landscapes and specific management effects.

Acknowledgements

Material present in this report was obtained from research projects supported by NSFFSA 890565 and 87-802331 as well as by USDA.

References

Anderson, D. and E.A. Paul. 1984. Organo-mineral complexes and their study by radiocarbon dating. *Soil Sci. Am. J.* 48:298-301.

Balesdent, J., G.H. Wagner, and A. Mariotti. 1988. Soil organic matter turnover in long-term field experiments as revealed by carbon-13 natural abundance. *Soil Sci. Soc. Am. J.* 52:118-124.

Bonde, T.A., B.T. Christensen, and C.C. Cerri. 1992. Dynamics of soil organic matter as reflected by natural carbon-13 abundance in particle size fractions of forested and cultivated oxisols. *Soil Biol. Biochem.* 24:275-277.

Buyanovsky, G.A. and G.H. Wagner. 1986. Post-harvest residues in cropland. *Plant and Soil* 93:57-65.

Cambardella, C.A. and E.T. Elliott. 1993. Methods of physical separation and characterization of soil organic matter fractions. *Geoderma*. In press.

Cambardella, C.A. and E.T. Elliott. 1992. Particulate soil organic-matter changes across a grassland cultivation sequence. *Soil Sci. Soc. Am. J.* 56:577-583.

Campbell, C.A., E.A. Paul, D.A. Rennie, and K.J. McCallum. 1967. Factors affecting the accuracy of the carbon dating method of soil humus studies. *Soil Sci.* 104:81-85.

Coleman, D.C. and B. Fry. 1991. *Carbon Isotope Techniques*. Academic Press, San Diego.

Elliott E.T., K. Paustian, H.P. Collins, E.A. Paul, C.V. Cole, I.C. Burke, R.L. Blevins, D.J. Monz, and S.D. Frey. Terrestrial carbon pools and dynamics: Preliminary data from the corn belt and Great Plains. ASA Special Publication, Soil Quality Symposium (in press).

Horwath, W.R. 1993. The dynamics of carbon, nitrogen and soil organic matter in populus plantations. Ph.D. dissertation, Michigan State University.

Janzen, H.H., C.A. Campbell, S.A. Brandt, G.P. Lafond, and L.Townley-Smith. 1992. Light-fraction organic matter in soils from long-term crop rotations. *Soil Sci. Soc. Am. J.* 56:1799-1806.

Leavitt, S.W., E.A. Paul, B.A. Kimball, and G.R. Hendrey. 1992. Isotopic estimation of inputs of carbon to cotton soils under FACE CO_2 enrichment. DOE Research Summary, U.S. Department of Energy Global Change Program, Carbon Dioxide Information Analysis Center, Oak Ridge National Laboratory, Oak Ridge, TN.

Leavitt, S.W., E.A. Paul, B.A. Kimball, G.R. Hendrey, J.R. Mauney, R. Rauschkolb, H. Rogers, Jr., K.F. Lewin, J. Nagy, P.J. Pinter, Jr., and H.B. Johnson. 1993. Carbon isotope systematics of FACE cotton and soils. *Agric. For. Meteorol.* Submitted.

Martel, Y.A. and E.A. Paul. 1974. The use of radiocarbon dating of organic matter in the study of soil genesis. *Soil Sci. Soc. Am. Proc.* 38:501-506.

Paul, E.A. and F.E. Clark. 1989. Soil Microbiology and Biochemistry. Academic Press, San Diego. 273 pp.

Schwartz, D., A. Mariotti, R. Lanfranchi, and B. Guillet. 1986. $^{13}C/^{12}C$ ratios of soil organic matter as indicators of vegetation changes in the Congo. *Geoderma* 39:97-103.

Steiner, R.A. and R.W. Herdt (eds.). 1993. A Global Directory of Long Term Agronomic Experiments. Vol. 1: Non-European Experiments. The Rockefeller Foundation, New York, N.Y.

Voroney, R.P., J.P. Winter, and E.G. Gregorich. 1991. Microbe/plant/soil interactions. p. 77-99. In: D.C. Coleman and B. Fry (eds.). *Carbon Isotope Techniques*, Academic Press, San Diego

Waters, A.G. and J.M. Oades. 1991. Organic matter in water stable aggregates. In: W.S. Wilson (ed.).*Advances in Soil Organic Matter Research: The Impact on Agriculture and the Environment*. The Royal Soc. Chemistry, Cambridge, United Kingdom.

Wilhelm, W.W., L.N. Mielke, and C.R. Fenster. 1982. Root development of winter wheat as related to tillage practice in western Nebraska. *Agronomy J.* 74:85-88.

Wilson, W.S. 1991. *Advances in Soil Organic Matter Research: The Impact on Agriculture and the Environment*. The Royal Society of Chemistry. Cambridge, United Kingdom.