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# USING BULK SOIL RADIOCARBON MEASUREMENTS TO ESTIMATE SOIL ORGANIC MATTER TURNOVER TIMES: IMPLICATIONS FOR ATMOSPHERIC $CO_2$ LEVELS

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**ABSTRACT.** Although soil contains about three times the amount of carbon present in the preindustrial atmosphere, determining how perturbations (*e.g.*, changing land use, $CO_2$ fertilization, changing climate and anthropogenic nitrogen deposition) alter soil carbon storage and influence atmospheric $CO_2$ levels has proved elusive. Not knowing the soil carbon turnover times causes part of this uncertainty. I outline a strategy for using radiocarbon measurements to estimate soil organic matter turnover times and inventories in native soil. The resulting estimates of carbon exchange produce reasonable agreement with measurements of $CO_2$ fluxes from soil. Furthermore, derivatives of the model are used to explore soil carbon dynamics of cultivated and recovering soil. Because the models can reproduce observed soil $^{14}C$ measurements in native, cultivated, and recovering ecosystems (*i.e.*, the underlying assumptions appear reasonable), the native model was modified to estimate the potential rate of additional carbon storage because of $CO_2$ fertilization. This process may account for 45–65% of the "missing $CO_2$ sink."

## INTRODUCTION

Many researchers have estimated the global inventory of soil carbon using a variety of techniques: Schlesinger (1977) used vegetation types to estimate an inventory of 1456 Gt C; Post *et al.* (1982) estimated soil humus to hold 1395 Gt C using climatic life zones; and Eswaran, den Berg and Reich (1993) used soil orders to estimate an inventory of 1576 Gt C. Although there are differences among the techniques, the results generally agree. How this large pool of carbon influences atmospheric $CO_2$ is uncertain, because the discrete turnover times of the multitude of soil compounds are not well known. These turnover times may range from days to millennia. The purpose of this paper is to introduce a strategy for estimating the turnover time of soil organic matter (SOM) using bulk soil radiocarbon measurements.

The technique builds on the approaches of others: the Century and Rothamsted models, mass balance approaches and soil humus fractionation approaches. The Century and Rothamsted models use measurements of soil carbon decomposition as the foundation for sophisticated ecosystem models. The model structures are similar, having SOM consisting of fast, active and passive fractions. The Century model (Parton *et al.* 1987, 1989, 1993; Schimel *et al.* 1994) has an active carbon turnover time ranging from 20 to 50 yr, and, it assigns an 800- to 1200-yr turnover time for passive carbon. The Rothamsted model (Jenkinson 1990) uses a 20-yr turnover time for active carbon and a near infinite turnover time for passive carbon.

O'Brien and Stout (1978) use a sophisticated model to interpret their New Zealand soil $^{14}C$ measurements. Their model is constrained by the depth distribution of total carbon and $^{14}C$ and includes carbon input, decomposition rates and soil diffusiveness. They assign a 50-yr turnover time for active carbon and a near-infinite time for passive carbon.

Researchers have tried to separate active and passive components using physical and chemical fractionation techniques (Paul *et al.* 1964; Campbell *et al.* 1967; Martel and Paul, 1974b; Goh *et al.* 1976; Goh, Stout and Rafter 1977; Goh, Stout and Brown 1984; Scharpenseel, Ronzani and Pietig 1968; Scharpenseel, Tamers and Pietig 1968; Trumbore, Vogel and Southon 1989; Trumbore, Bonani and Wölfli 1990). Trumbore (1993) summarizes the results of various fractionation techniques. One way to test the effectiveness of these fractionation schemes is to see if the amount of bomb $^{14}C$ increases in the soil as predicted by estimates of the residence time for SOM. To date, the available fractionation schemes cannot do this (Trumbore 1993).

## ESTIMATING SOIL CARBON TURNOVER TIMES USING BULK RADIOCARBON MEASUREMENTS

In my research, I use a time-step one-box model and bulk soil $^{14}C$ measurements to estimate turnover times and inventories of active and passive carbon. The model has atmospheric $^{14}C$ values and $CO_2$ concentrations for every year from AD 1800 until the present. The user selects the carbon inventory and the turnover time. The turnover time equals the carbon inventory divided by the exchange flux. The exchange flux equals the amount of carbon that is added to the box (from photosynthesis) or lost from the box (respiration). Losses through erosion and dissolution are thought to be small (Schlesinger 1986) and are not considered. The model can be run in either a steady-state mode (where the flux in equals the flux out) or in a non-steady-state mode (in which carbon is either accumulating or decreasing). I use this model and soil $^{14}C$ data to show that soil carbon has more than one component and to estimate the turnover time of the passive fraction, the proportions of active and passive carbon in surface soil, and the active soil carbon residence time.

Many researchers have concluded that soil consists of a complex mix of organic molecules whose turnover times range from a few years to thousands of years. This pool cannot be characterized by a single turnover time (O'Brien 1984; Balesdent, Wagner and Mariotti 1988; Parton *et al.* 1987, 1989, 1993; Jenkinson and Raynor 1977). Figure 1 shows how a single residence time of 650 yr fails to characterize soil humus. Six prebomb values of surface soil had an average $^{14}C$ content of 92% modern (pMC) (Table 1). This 92 pMC value translates into a 650-yr turnover time, which shows very little increase in bomb $^{14}C$ with time. The soil $^{14}C$ values increase in the 1960s and then level off

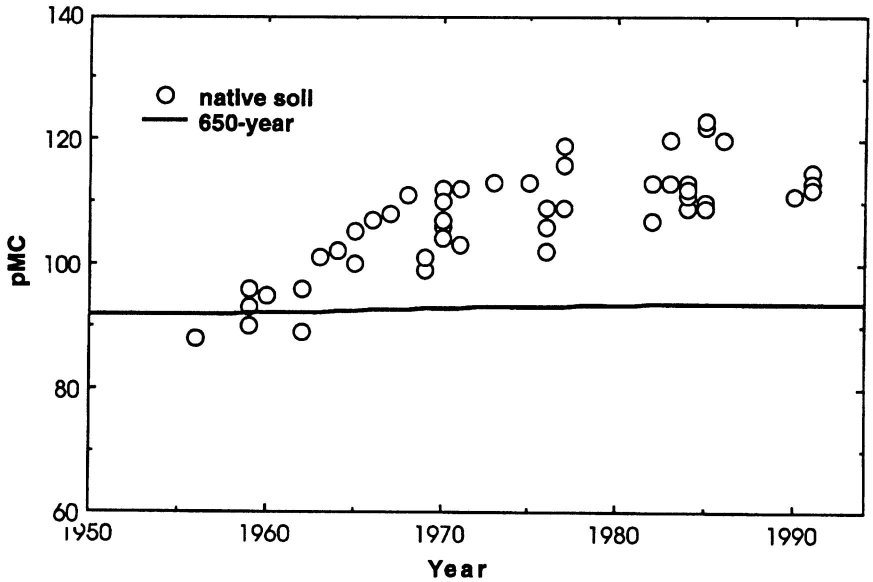

Fig. 1. Soil $^{14}C$ values *vs.* time. Measured soil $^{14}C$ values for non-cultivated soils are plotted against time from 1950 to 1991. The values tend to increase during the 1960s and then level off. Model results for a theoretical carbon pool having an 650-yr residence time are shown by the line. If soil carbon consisted of components having this single residence time, one would expect the observations to fall around this line. Data for these figures from Tables 1 and 2 and Harrison (1994).

(Fig. 1). This increase suggests that soil organic material contains an active component with a turnover time significantly <650 yr. This active component must be diluted with a passive component having a turnover time of >650 yr.

TABLE 1. Prebomb Soil $^{14}C$ Values for Uncultivated Soil

| pMC | Depth (cm) | Year | Reference | Soil type | Location |
|---|---|---|---|---|---|
| 96 | 1–8 | 1959 | O'Brien (1986) | Mollisol | New Zealand |
| 96 | 0–23 | 1959 | Trumbore (1993) | Temp. forest | USA |
| 82 | 0–12 | 1927 | Trumbore *et al.* (1990) | Spodosol | FSU |
| 96 | A-hor. | 1962 | Campbell *et al.* 1967 | Chernzemic | Canada |
| 90 | A-hor. | 1962 | Campbell *et al.* 1967 | Mollisol | Canada |
| 94 | 0–2 | 1959 | Vogel 1970 | Forest | Germany |

The passive soil turnover time can be estimated from soil $^{14}C$ measurements made at depths where little or no active soil carbon is present. $^{14}C$ concentrations decrease with depth, reflecting a decrease in the proportion of active to passive carbon. Both carbon content and $^{14}C$ soil values tend to decrease with increasing depth (Harrison, Broecker and Bonani 1993a). At some depth, the soil $^{14}C$ measurements approach an inflection value below which they tend to decrease very slowly. These values and depths vary for different locations (see Table 2). The average value for the sites listed in Table 2 was 55 pMC, which corresponds to a 4700-yr turnover time for passive soil carbon. However, some tropical soils have active soil carbon several meters below the surface (Nepstad *et al.* 1994; Fisher *et al.* 1994)

TABLE 2. Deep Soil $^{14}C$ Values for Cultivated and Uncultivated Soil

| pMC | Depth (cm) | Reference | Soil type | Location |
|---|---|---|---|---|
| 49 | 65–105 | Becker-Heidmann *et al.* (1988) | Mollisol | China |
| 43 | 74–94 | O'Brien and Stout (1978) | Mollisol | New Zealand |
| 60 | 60–140 | Scharpenseel and Becker-Heidmann (1989) | Vertisol | Israel |
| 60 | 60–140 | Becker-Heidmann (1989) | Udic | India |
| 62 | 85–110 | Tsutsuki *et al.* (1988) | Mollisol | Germany |

For the prebomb condition, one can estimate the proportions of active and passive components in surface soil using the 55% modern passive soil $^{14}C$ value and the 92% modern values measured for the bulk soil (Table 1). If the active component turns over quickly (<100 yr), enough so that we can assume that its $^{14}C$ value is almost 100 pMC ($^{14}C$ has a half-life of 5700 yr), a mixture of 17% passive and 83% active leads to the observed average $^{14}C$ value of 92 pMC (Fig. 2).

The postbomb increase in soil $^{14}C$ concentrations can be used to estimate the active soil carbon turnover time (Harrison 1993a). A 25-yr turnover time produces the best fit to the available data. Most of the points are for temperate ecosystems, so warmer tropical ecosystems may have faster turnover times, whereas cooler boreal turnover times may be slower. The proportions of active and passive soil may differ for tropical and boreal climates.

This approach can be validated by looking at a specific site where deep, prebomb, and a time series of surface soil data can be used to compare model predictions with soil $^{14}C$ measurements. O'Brien

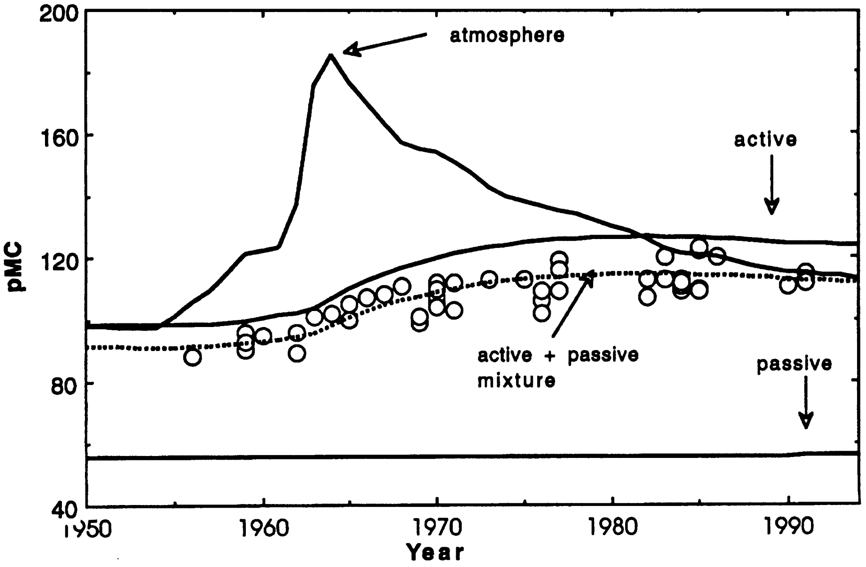

Fig. 2. Soil $^{14}C$ model predictions and soil $^{14}C$ observations *vs.* time. A mixture of 83% active and 17% passive soil carbon produces the best visual fit to the native soil $^{14}C$ measurements (○). The concentration of atmospheric $^{14}C$ almost doubled because of nuclear bomb testing.

and Stout (1978) reported measurements for a New Zealand grassland soil that included a deep soil value and a time series of surface soil values that extended from prebomb times into the mid-1970s. The model that best fits the data has a 12% passive component plus a 88% active portion that turns over every 25 yr. Figure 3 shows the agreement between the model and the data. Although this model reproduces the prebomb soil $^{14}C$ values and the postbomb increase in $^{14}C$ values in native soil, the model's robustness can be evaluated by seeing if derivatives can explain $^{14}C$ measurements in cultivated and recovering soil.

## SOIL CARBON DYNAMICS IN CULTIVATED ECOSYSTEMS

Another way to test the model's assumption is to see if it can be modified to reflect carbon dynamics in cultivated ecosystems. Soil loses *ca.* 25% of its carbon when cultivated (Schlesinger 1986; Post and Mann 1990; Davidson and Ackermen 1993). This loss stems from reduced inputs of organic matter and increased rates of organic matter mineralization. Cultivated soil generally has lower $^{14}C$ values than native soil that has been sampled at the same time (Fig. 4; Martel and Paul 1974a). Hsieh (1992, 1993) has developed a two-component model that reproduces temporal changes in $^{14}C$ values in cultivated soil. Using a similar approach, Harrison, Broecker and Bonani (1993b) assumed that carbon lost from soil due to cultivation would be lost from the active carbon pool. However, the oxidation and loss of a fraction of the active-soil carbon could only explain about half the observed $^{14}C$ depletion. Mixing subsurface soil with the shallow surface soil through cultivation (*i.e.*, the plow mixes up the soil) can account for the remaining depletion in $^{14}C$ values. The model, which included mixing and oxidation, produced good agreement with the available data for changes in soil $^{14}C$ upon cultivation (Fig. 4).

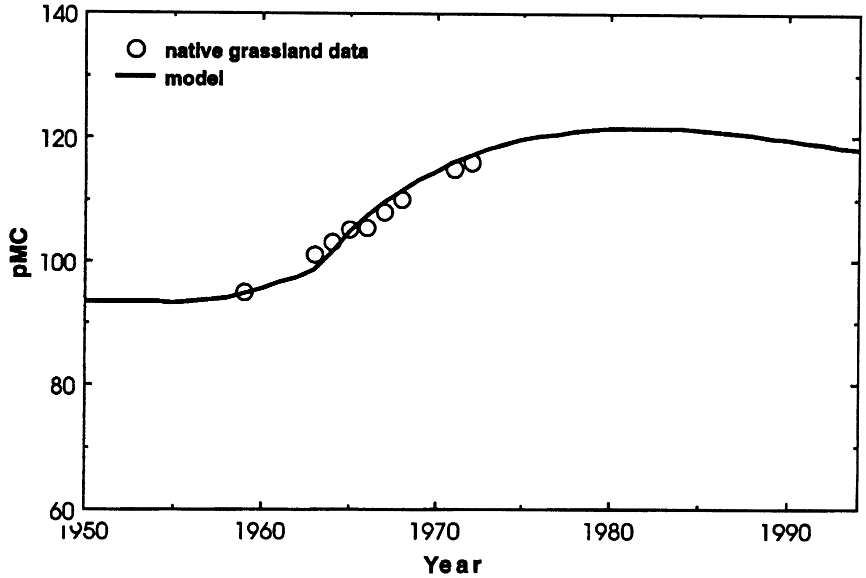

Fig. 3. New Zealand test case. O'Brien and Stout (1978) published $^{14}C$ data for a New Zealand grassland site comprising a time series of surface soil, including one early-bomb, and one deep soil $^{14}C$ value. I used this information to attempt to validate the model for a specific site. This model that best fit the data consisted of 88% active and 12% passive carbon. The active carbon had a 25-yr turnover time.

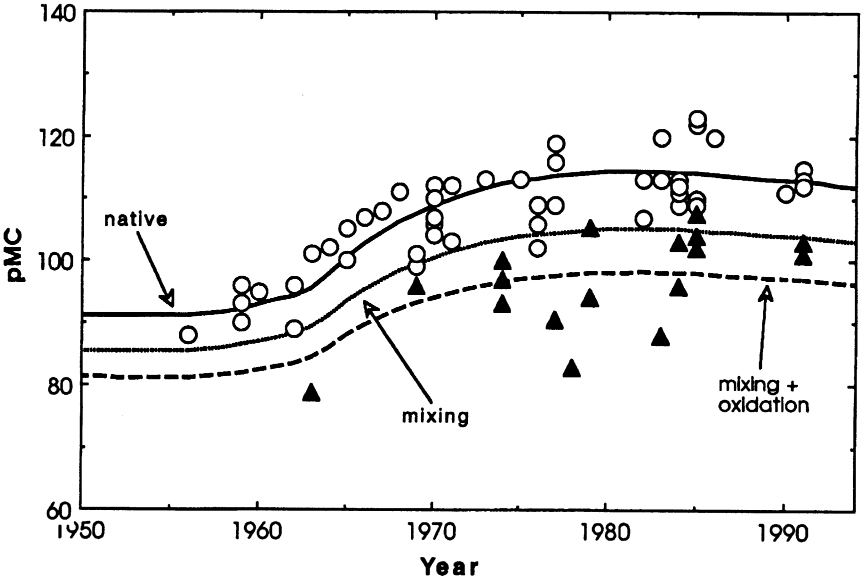

Fig. 4. Native *vs.* cultivated soil $^{14}C$ values. Cultivated soil (▲) has lower $^{14}C$ values than native soil (○). This difference is caused by mixing and oxidation. The farmer's plow mixes deeper $^{14}C$ depleted soil with $^{14}C$ rich surface soil, diluting the amount of active soil carbon in the surface. Increased SOM oxidation further reduced the inventory of active soil carbon.

## SOIL CARBON TURNOVER IN RECOVERING ECOSYSTEMS

To test the model further, the baseline model was modified to explore the carbon dynamics of a recovering soil that was increasing its carbon stores. One example includes a recovering temperate forest located in the Calhoun National Forest, described by Richter *et al.* (1994; 1995), Harrison (1994), and Harrison *et al.*(1995). This site contains a Loblolly pine (*Pinus taeda*) forest that was planted in 1959 on land that had been cultivated for the previous 150 yr. The soil has been accumulating carbon since the pines were planted. From 1962 to 1968, the surface carbon concentration increased from 5.9 to 8.0 Mg ha$^{-1}$. The native soil carbon model was modified to take into account this carbon accumulation by increasing the flux of carbon into the active-soil carbon pool. The turnover time that best reproduced the observed $^{14}$C measurements was 12 yr, which is about twice as fast as carbon turnover in native ecosystems. Figure 5 shows the agreement between the model and the data for this accumulating ecosystem.

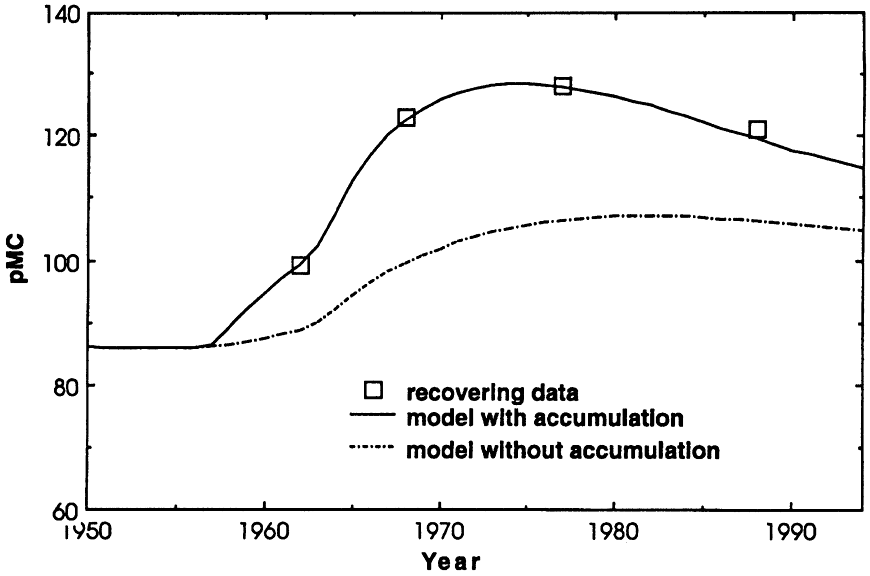

Fig. 5. $^{14}$C measurements and model results for South Carolina. The model results that best fit the surface soil $^{14}$C measurements were those from an accumulation model that took into account the increase in carbon inventory. The carbon increased from 5.9 Mg ha$^{-1}$ to 8.0 Mg ha$^{-1}$. The active reservoir turnover time is about twice as fast as the global average for native soils.

## DETERMINING THE GLOBAL INVENTORY OF ACTIVE SOIL ORGANIC MATTER

Table 3 lists the current best estimates of the global inventory of non-wetland SOM. These values can be translated to give a global inventory of active soil carbon. The active-to-passive proportions found in surface soil cannot be applied to all carbon present in non-wetland soil because the proportion of active to passive carbon decreases with increasing depth. The integrated inventories suggest that the global pool is *ca.* 50% passive and 50% active (Harrison 1993a). Although extrapolating these distributions involves uncertainty (*i.e.*, the proportions are likely to differ for other climates and types of vegetation), the global inventory of active carbon in soils ranges between 500 and 625 Gt C.

TABLE 3. Soil Carbon Sequestration Due to $CO_2$ Fertilization*

| | Reference | | |
|---|---|---|---|
| Category | Schlesinger (1977, 1991) | Post *et al.* (1982) | Eswaran *et al.* (1993) |
| Total soil carbon (Gt C) | 1500 | 1400 | 1600 |
| Non-wetland soil carbon (Gt C) | 1000 | 1200 | 1250 |
| Active soil carbon (Gt C) | 500 | 600 | 625 |
| Exchange flux (Gt C $yr^{-1}$) | 20 | 24 | 25 |
| Potential C sequestration due to $CO_2$ fertilization (Gt C $yr^{-1}$) | 0.5 | 0.6 | 0.7 |
| % "missing $CO_2$ sink" | 45 | 55 | 65 |

*This table uses estimates of the inventory of active soil carbon and a simple $CO_2$ fertilization model to estimate the potential carbon sequestration in soil due to $CO_2$ fertilization. Only non-wetland soil was included in the calculation. The inventory of active soil was estimated assuming that a soil profile contains *ca.* 50% active carbon; a $CO_2$ fertilization factor of 0.35 was used in the $CO_2$ fertilization model to calculate the amount of carbon sequestered during an average year in the 1980s (Harrison 1993a). Dixon *et al.*'s (1994) missing sink estimate for the 1980s of 1.1 Gt C $yr^{-1}$ was used.

A way to test the model is to determine if fluxes measured from the soil agree with those values predicted by the model. A 500 Gt C pool turning over every 25 yr emits 20 Gt C from the soil annually, *i.e.*, a flux of 150 g C $m^{-2}$ $yr^{-1}$. The 600 and 625 Gt C values result in 24 Gt C $yr^{-1}$ (180 g C $m^{-2}$ $yr^{-1}$) and 25 Gt C $yr^{-1}$ (190 g C $m^{-2}$ $yr^{-1}$), respectively. These estimates are significantly lower than the observed flux from a temperate forest soil, which ranged from 400 to 500 g C $m^{-2}$ $yr^{-1}$ (Raich and Schlesinger 1992). It must be realized that the measured values record sources of $CO_2$ besides microbial respiration of SOM, such as root respiration and the oxidation of litter and fine roots. While it is impossible to resolve the contribution from these additional $CO_2$ sources, it is unlikely that they account for >50% of the total flux. Furthermore, soils such as desert soil that have low organic carbon contents make it difficult to compare global and regional values. The ecological complexities make it virtually impossible to get better agreement between the $CO_2$ fluxes predicted by the model and those measured for a temperate forest ecosystem. However, even if one considers the measured flux to represent an upper limit, the predicted values fall well below it.

## CARBON DIOXIDE FERTILIZATION

Having estimated the turnover time and inventory of fast-cycling carbon, it is possible to assess the amount of carbon that might be stored in soil as a consequence of $CO_2$ fertilization. $CO_2$ fertilization occurs when plants increase their growth when exposed to elevated $CO_2$ levels (Strain and Cure 1985; Bazzaz and Fajer 1992). A convenient way of expressing the effect is *via* a $CO_2$ fertilization factor (*i.e.*, the percentage increase in growth that results from a doubling of $CO_2$ concentration).

Although many indoor $CO_2$ fertilization experiments have shown increased growth at elevated $CO_2$ levels (Strain and Cure 1985), the extrapolation of these results to natural vegetation is highly controversial (Bazzaz and Fajer 1992). Further, the application of these $CO_2$ fertilization factors to the carbon input to soil involves speculation. Zak *et al.* (1993) have shown that the soil carbon under plants grown in doubled $CO_2$ concentrations in open-top chambers was greater than their nonelevated paired counterparts, although their results were not statistically significant. Norby *et al.* (1992) found evidence of increased fine root density for trees grown in elevated $CO_2$ concentrations, lending credibility to the belief that if plant growth is stimulated, the inventory of soil carbon will be as well.

For this study, a $CO_2$ fertilization model has been developed to estimate the additional amount of carbon stored in soil because of $CO_2$ fertilization. The flux of carbon into soil organic matter can be increased by adding the term

$$\beta \, \delta pCO_2 \, EF \,, \tag{1}$$

where $\beta$ = $CO_2$ fertilization factor (0.35 after Harrison *et al.* 1993a)
$\delta pCO_2$ = fractional change in $CO_2$
EF = exchange flux.

As the flux of carbon into the box increases in the box, the decay flux (*i.e.*, the decay constant times the amount of carbon in the box) also increases. If the level of atmospheric $CO_2$ stops increasing, the soil will attain a higher steady-state carbon content with a turnover time of 25 yr.

Table 3 lists the model results. The amount of carbon sequestered in soil because of $CO_2$ fertilization ranges from 0.5 to 0.7 Gt C $yr^{-1}$ for the 1980s. Dixon *et al.* (1994) estimate that the "missing sink" is 1.1 Gt C $yr^{-1}$ for this time period. Thus, $CO_2$ fertilization might well provide the mechanism to explain much of the much of the "missing $CO_2$ sink" as carbon held in soil.

## CONCLUSION

I have presented a strategy for estimating the global inventory and turnover time for carbon stored in non-wetland soil. The available $^{14}C$ data suggest that active soil carbon has a 25-yr turnover time and a 500–625 Gt C inventory. Therefore, active soil carbon is well placed to respond significantly to perturbations such as $CO_2$ fertilization, changing climate and anthropogenic nitrogen deposition. The effects of $CO_2$ fertilization on the flux of detrital organic matter input to soil can potentially store 0.5–0.7 Gt C $yr^{-1}$, thus explaining a major portion of the "missing $CO_2$ sink."

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