

11. Soil Carbon Turnover

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The net carbon balance of forests is the small difference between very large fluxes of carbon uptake (via photosynthesis) and carbon loss (via respiration). If carbon uptake by forest vegetation exceeds C loss by vegetation and soils, the forest is a net sink of atmospheric carbon. Currently, regrowing temperate forest ecosystems of the eastern United States are storing C (Goulden et al. 1996; Greco and Baldocchi 1996; Hollinger et al. 1999; Curtis et al. 2002).

Estimates of net C balance can be made by eddy-covariance and biometric methods (direct measurements of carbon fluxes and/or stocks). Eddy covariance directly measures the net fluxes of C between atmosphere and forest at roughly half-hourly intervals. Those measurements are integrated to estimate an annual net C balance. The equipment can be automated and left in place to measure variations in annual C balance from year to year. However, eddy covariance gives little information on where the C is being stored in the ecosystem—in particular, whether storage occurs in vegetation or soils. Biometric methods calculate changes in C stocks from repeated surveys of vegetation and soils and, thus, can be used to determine where C is being stored. However, large spatial heterogeneity and errors associated with spatial scaling of field measurements prevent this method from detecting small changes in C stocks from year to year. Large discrepancies between eddy flux and biometric approaches have been found at two of five AmeriFlux sites in eastern deciduous forests of the United States (Curtis et al. 2002) and point to the need to resolve potential uncertainties in both approaches. One of the largest discrepancies is found at the Oak Ridge Reservation,

the site of the Throughfall Displacement Experiment (TDE) (Chapter 22, this volume). Understanding where and how C is stored is important not only for resolving the present carbon balance of ecosystems, but also for predicting future ecosystem response to climate change.

A major uncertainty in biometric approaches for estimating ecosystem carbon budgets is quantifying current and future rates of soil C accumulation (or loss). Quantifying soil C stocks to within 20% is challenging, and in rocky soils, it can require hundreds of careful, labor-intensive measurements (Huntington et al. 1988; Fernandez et al. 1993). Quantifying C cycling rates through these stocks is even more difficult because soil C is made up of a variety of compounds that cycle on timescales ranging from days to millennia. We report C stocks and dynamics in soil organic matter at the Walker Branch Watershed in the Oak Ridge Reservation, Oak Ridge National Laboratory, Tennessee. Our approach to determining carbon dynamics in soils is to (1) separate soil organic matter (SOM) by depth, density, and size to isolate components that have different characteristic turnover times and (2) use radiocarbon to estimate the turnover time of each isolated fraction of soil organic matter. The ^{14}C approach takes advantage of radiocarbon produced by atmospheric thermonuclear weapons testing in the early 1960s (bomb ^{14}C ; Fig. 11.1), which can be used as a tracer for the timing of photosynthetic uptake and C cycling rates in an ecosystem during the past several decades. Once we have estimates of C stocks and turnover times for isolated SOM fractions, we use a C-stock model to simulate changes in C inputs to the soil to assess the potential for future carbon storage.

Use of the bomb- ^{14}C approach requires knowledge of the time record of ^{14}C in atmospheric CO_2 and in C inputs to soil organic matter since 1950. At Walker Branch, the local time history for ^{14}C as recorded in tree-ring cellulose agrees with the records for CO_2 in Northern Hemisphere air prior to 1990 (Fig. 11.1). After 1990, the Walker Branch record shows no decline in ^{14}C , although the record for Northern Hemisphere CO_2 shows a 4–8‰ per year decrease. This is largely because of local releases of ^{14}C that occurred (from an unknown source) on or around the study site after 1995 (Trumbore et al. 2002). Thus, the isotopic signatures for SOM fixed between 1990 and 1998 are not distinguishable from one another at this site, and we can only quantify SOM dynamics for C pools that cycle at rates greater than ~10 years. However, this limitation does not affect the goals of this work, namely to quantify current and potential rates of soil C storage at Walker Branch, discuss these results in the context of eddy-covariance and biometric approaches also performed on Walker Branch, and put our results in the larger context of soil C accumulation potential for forests of the eastern United States.

Sites

The work discussed here was carried out at two sites on the Walker Branch Watershed. The first site is located on the upper slope of the ambient plot of the

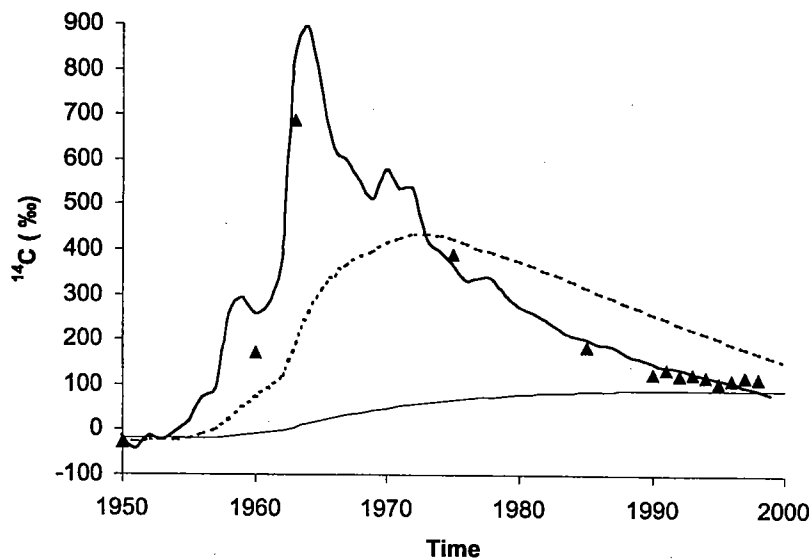


Figure 11.1. The global and local record of atmospheric ^{14}C of CO_2 since 1950. The thick solid line represents the time record of ^{14}C in the atmosphere (Northern Hemisphere) based on grapes grown in Russia (Burchuladze et al. 1989) for 1950 to 1976, direct atmospheric measurements for 1977 to 1996 (Levin and Kromer, 1997), and an assumed 4‰ per year decrease since 1996 (Levin and Hessheimer 2000). Filled triangles represent the local Walker Branch time record reconstructed with values for tree-ring cellulose from a *Quercus alba* tree taken on the Walker Branch Site (Trumbore et al. 2002). ^{14}C in cellulose from annual tree rings has been shown to be a good indicator of atmospheric ^{14}C of CO_2 (Hua et al., 1999). The accuracy of the ^{14}C measurement by accelerator mass spectrometry (AMS) is $\pm 6\%$; thus, the error bars are smaller than the symbols shown. Note that none of the tree-ring cellulose values from 1990 to 1998 are significantly different from each other. Thin dashed and solid lines represent the amount of radiocarbon in a carbon pool with turnover times of 10 and 50 years, respectively (as modeled using our steady-state homogenous-carbon-pool model; see text for equations).

Throughfall Displacement Experiment (TDE) (Chapter 2, this volume). The second site (P5) is located about 1 km east in a very similar forest stand.

Methods

Soil Organic Matter Sampling

Samples for carbon and ^{14}C inventories were taken in 1998 from two soil pits ($0.5\text{ m} \times 0.5\text{ m}$) dug with the quantitative pit methodology of Huntington et al. (1989) and Hamburg (1984) and modified as described in Gaudinski et al. (2000). This method involves sampling a large volume of soil to allow calculation of horizon-specific bulk densities. Samples which integrated over each soil horizon

were collected from one of the pit faces for radiocarbon analysis and total C and N analyses. Carbon inventories are reported to a depth of about 80 cm.

Laboratory Analyses

Soil samples were separated into different fractions by a combination of hand sorting and density separations. Density separations are performed with a sodium polytungstate solution at $\rho = 2.1 \text{ g cm}^{-3}$. Material that floats is defined as low density, whereas material that sinks is defined as high density and assumed to be associated with minerals. Organic horizons were considered to be 100% low-density material composed of litter that can be identified as coming from leaves (L_L) and roots (L_R) and humified (H) material that is comprised of pieces too small to be identified as plant tissues. Oi horizons were composed entirely of large pieces of leaf litter (L_L). Oe + Oa horizons were hand picked to separate L_L , L_R , and H fractions. Mineral horizons were comprised of low-density fractions (L_R and H) and high-density mineral-associated carbon (M). Size separations (sieving with an 80- μm sieve) and hand picking were performed on the low-density fractions from the A horizons to separate L_R and H fractions. These methods are described in further detail in Gaudinski et al. (2000).

The rate of incorporation of bomb- ^{14}C in SOM fractions provides a measure of how fast C in a given fraction is exchanged with the atmosphere (on decadal timescales). This approach is particularly powerful when ^{14}C analysis can be performed on the same soil sampled at two different points in time (Trumbore 1993). The Plot 5 site was added to our study because we were able to obtain archived soils that were sampled in 1972. The archived P5 samples were dried, ground, and stored in a laboratory on Walker Branch. A pit description for the 1972 soil was not available; however, we discussed the 1972 sampling protocol with the original soil sampler (Todd, personal communication) and compared the archived and modern soil horizon colors to ensure that horizons sampled in 1998 were as similar as possible to those sampled in 1972. The 1972 pit remained open, and in 1998, we extended one of its walls for resampling. Density separations were performed on archived samples and the low-density fraction of the A horizon was separated into greater than and less than 80 μm fractions. Because the archived soil had been previously ground, hand picking to remove roots was not performed.

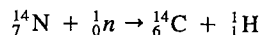
Radiocarbon Analyses

We converted all solid and gas samples to graphite by sealed-tube zinc reduction (Vogel 1992) and measured the ^{14}C content on an accelerator mass spectrometer (AMS) at the Center for Accelerator Mass Spectrometry at Lawrence Livermore National Laboratory. We express radiocarbon data as $\Delta^{14}\text{C}$, the difference in parts per thousand (per mil or ‰) between the $^{14}\text{C}/^{12}\text{C}$ ratio in the sample compared to that of a universal standard (0.95 times the activity of oxalic acid I, decay corrected to 1950), which is defined as "Modern." All samples are corrected to a common $\delta^{13}\text{C}$ value of -25‰ to correct for the effects of mass-dependent iso-

topic fractionation on measured ^{14}C values. This correction accounts for the discrimination of atmospheric ^{14}C during fixation of CO_2 by plant photosynthesis (^{14}C is assumed to fractionate twice as much as ^{13}C). The reported ^{14}C values, therefore, reflect the atmospheric ^{14}C of CO_2 from which the C was originally fixed.

Interpretation of Radiocarbon Data

Radiocarbon is produced naturally in the upper atmosphere when high-energy cosmic rays collide with atomic nuclei to produce neutrons. The neutrons then collide with nitrogen to form ^{14}C via the following reaction:



^{14}C is a radioactive isotope that decays back to ^{14}N via beta decay (electron emission with a maximum energy of 0.155 Mev) with a half-life of 5730 years. Once produced, the ^{14}C quickly oxidizes to form $^{14}\text{CO}_2$ and follows the same physical and chemical pathways as $^{13}\text{CO}_2$ and $^{12}\text{CO}_2$. The estimated natural production rate of ^{14}C over the Earth's surface is 2.50 ± 0.50 atoms of $^{14}\text{C} \text{ cm}^{-2} \text{ s}^{-1}$, and the steady-state natural $^{14}\text{C}/^{12}\text{C}$ ratio is $(1.2 \pm 0.2) \times 10^{-12}$ (Linick 1975). This $^{14}\text{C}/^{12}\text{C}$ ratio is referred to as "Modern" and, by definition, refers to the atmosphere of 1950.

Radiocarbon is also produced by aboveground thermonuclear explosions (bomb ^{14}C) when neutrons are released into the atmosphere. Atmospheric weapon testing in the late 1950s and early 1960s increased atmospheric ^{14}C by 10^{29} atoms (160 kg ^{14}C) or twice modern values (Linick 1975; Hesshaimer et al. 1994).

Globally, the $^{14}\text{CO}_2$ concentration has been decreasing since its peak values in the early to mid-1960s (Fig. 11.1). The decrease was rapid at first because of the dilution of the ^{14}C spike through exchange with carbon in the oceans and terrestrial biosphere. The rate of decline has slowed with time (Fig. 11.1), with the annual change of $\Delta^{14}\text{C}$ averaging $\sim -8 \text{‰ y}^{-1}$ in the 1980s and early 1990s. Between 1996 and 1999, the period of this study, rates of decrease slowed to $4\text{--}8 \text{‰ y}^{-1}$ (Levin and Hesshaimer 2000). The accuracy of the ^{14}C measurement AMS with graphite and sealed-tube zinc reduction is $\pm 6 \text{‰}$ based on repeated analysis of intercalibration standards. In general, this level of accuracy allows us to resolve the time elapsed since organic matter was fixed from the atmosphere to within 1–2 years over the past five decades.

Turnover times for SOM fractions were estimated from their respective ^{14}C signatures and a steady-state, time-dependent model that tracks C and ^{14}C in each fractionated SOM pool. We assume that the C pools remain at steady state (inputs = losses) and that the C inputs added to each pool for each year have ^{14}C contents determined from the time record of atmospheric ^{14}C of CO_2 [see Trumbore et al. (1995) and Gaudinski et al. (2000) for further details]. The balance of ^{14}C in year t is given by

$$C_{(t)} R_{\text{som}(t)} = IR_{\text{atm}(t)} + C_{(t-1)} \times R_{\text{som}(t-1)} - kC_{(t-1)} R_{\text{som}(t-1)} - \lambda C_{(t-1)} R_{\text{som}(t-1)}; \quad (1)$$

collecting terms, we have

$$R_{\text{som}(t)} = \frac{IR_{\text{atm}(t)} + (C_{(t-1)} R_{\text{som}(t-1)} (1 - k - \lambda))}{C_{(t)}}, \quad (2)$$

where

C = stock of carbon for the given C pool (in g C m^{-2})

I = inputs of C above and below ground (in $\text{g C m}^{-2} \text{y}^{-1}$)

k = decomposition rate of SOM (in y^{-1})

$$R = \left(\frac{\Delta^{14}\text{C}}{1000} \right) - 1$$

R_{atm} = the ratio of ^{14}C in the atmosphere normalized to a standard

R_{som} = the ratio of ^{14}C in the given SOM pool: L, H, or M, normalized to a standard

λ = radioactive decay constant for $^{14}\text{C} = 1/8267$ years

t = time (year) for which the calculation is being performed.

Note that the R_{som} at any time t depends not only on the $R_{\text{atm}(t)}$ but also on both C inventory and R_{atm} of previous years.

We varied turnover times (i.e., k^{-1}) until we matched the measured $\Delta^{14}\text{C}$ signature for the given SOM pool in the year of sampling (1998). This approach assumes that (1) the C pool being modeled is homogeneous and (2) there is no time lag between fixation of C from the atmosphere and the time when C is added to the SOM pool [i.e., C inputs each year have $^{14}\text{C}/^{12}\text{C}$ ratios equal to $R_{\text{atm}(t)}$]. Once a turnover time was estimated from ^{14}C data, the flux of C into and out of the pool were calculated as the stock of C in that pool divided by the turnover time.

Results

Soil C and ^{14}C Inventory

Soil organic matter inventories (not including roots) were 4610 g C m^{-2} for the soil pit dug on the ambient portion of the TDE (to a depth of 70 cm) and 5695 g C m^{-2} for the P5 site (to a depth of 80 cm; Table 11.1). Carbon concentrations decreased rapidly with depth with 46% C in the O horizons and 0.2 % C in the B horizons. Most of the SOM C stocks, however, resided in mineral soils, with an average of 615 g C m^{-2} in the O horizons (12% of the total) and 4540 g C m^{-2} in the mineral A, E and B horizons combined (88% of the total). Low-density ($L_L + H$) carbon made up 64% of the total soil carbon stocks. Humified organic material (H) makes up 90% of the total low-density C. Fine roots (< 2 mm in diameter) to a depth of 90 cm are assumed to contain 375 g C m^{-2} (live + dead roots with 47% carbon; Joslin and Wolfe 1999).

Radiocarbon contents of low-density fractions were consistently higher than those of dense fractions, though ^{14}C in both components decreased with soil depth (Fig. 11.2). Comparison of fractions separated from the bulk low-density material shows that it is made up of material that is higher in ^{14}C (fine roots) and lower in ^{14}C (humified fraction; H) relative to the bulk ^{14}C value (Fig. 11.2 and Table 11.2). Where low-density material was separated into components by sieving, the

Table 11.1. Soil organic matter inventory (not including roots)

Pit	Horizon	Bulk density ^a (g cm ⁻³)	Soil carbon (g C kg ⁻¹ soil)	Bottom depth (cm)	Low-density SOM			High-density SOM	
					Total C stock ^b (g C m ⁻²)	Leaf litter ^{c,d} (g C m ⁻²)	Humified ^d (g C m ⁻²)	Mineral associated ^d (g C m ⁻²)	
TDE	Oi	— ^e	456	2	360	360	NA ^f	NA	
	Oe + Oa	—	378	3	270	NA	270	NA	
	A	0.44	50	8	615	NA	565	50	
	E	1.01	7	50	2475	NA	885	1590	
Total	Bt	1.18	3	70	890	NA	240	650	
					4610				
P5	Oi	—	456	3	190	190	NA	NA	
	Oe + Oa	—	378	5	410	NA	410	NA	
	A	0.52	203	7	3095	NA	3025	70	
	E	1.01	7	24	1245	NA	450	795	
	B1/EB	1.13	2	41	430	NA	90	340	
	Bt	0.92	2	63	325	NA	65	260	
Total					5695				

^a Gravel-free bulk density (i.e., less than 2 mm).

^b Oi and Oe horizons calculated on a per area basis.

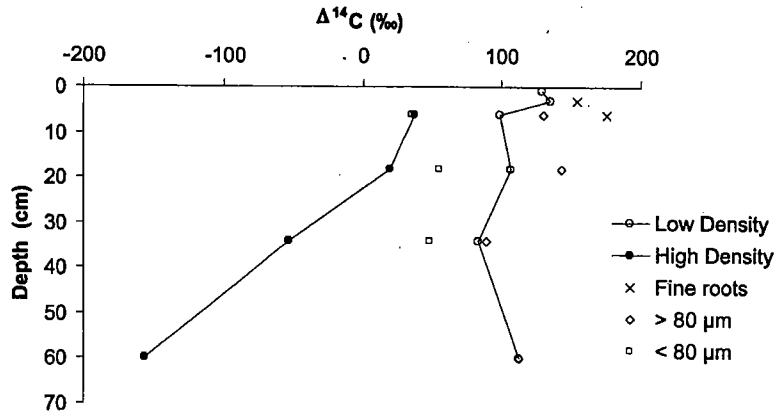
^c On a dry weight basis.

^d Low density means $\rho < 2.1 \text{ g cm}^{-3}$, high density means $\rho > 2.1 \text{ g cm}^{-3}$.

^e Not available.

^f NA = not applicable.

A. TDE Site



B. P5 Site

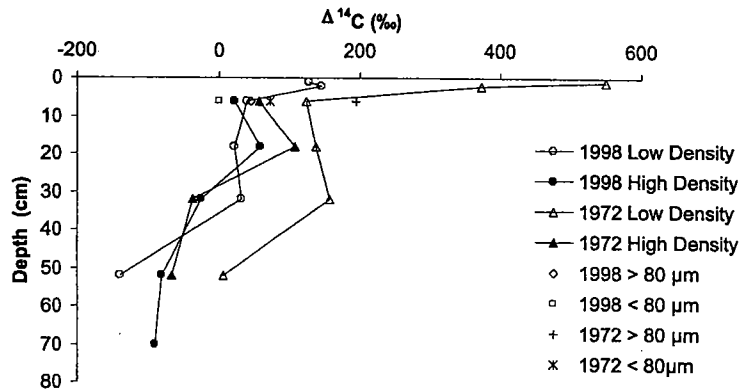


Figure 11.2. Radiocarbon profiles of low-density (open symbols) and high-density (closed symbols) soil organic matter and fine roots. The uppermost value in the profile represents the Oi horizon, which is bulk leaf litter that is still recognizable to the species level. The value directly below represents the humified component of the Oe + Oa horizon after roots and recognizable leaf parts have been removed. The remaining profile values are in the mineral soil and represent bulk low-and high-density soil organic matter samples or low-density samples sieved with an 80- μm sieve. When the bulk and sieved samples are shown together at the same depth, the bulk represents a mass-weighted composite of the greater than and less than 80 μm samples. At the P5 site, values are shown for archive (1972) and modern (1998) soils.

Table 11.2. ^{14}C values and associated turnover times (TTs) and corresponding fluxes

Horizon	Stock (g C m^{-2})	$\Delta^{14}\text{C}$ of SOM (%)	TT short (year)	TT long (year)	CO_2 flux ($\text{g C m}^{-2} \text{y}^{-1}$)	CO_2 flux horizon total ($\text{g C m}^{-2} \text{y}^{-1}$)	$\Delta^{14}\text{C}$ of SOM horizon total (%)
TDE							
Oi (L _L)	358	130	5	68	NA	NA	NA
Oe + Oa (L _R)	— ^a	154	NA ^b		NA	NA	
Oe + Oa (H)	269	136 ^c	6	60	45	6	97
A (LR)	—	175	NA		NA		
A (H)	563	99 ^d	91		6.2		
A (M)	46	37	189		0.2		
E (H)	883	107	83		10.6	14	84
E (M)	1582	19	435		3.6		
Bt (H)	240	113	80		3.0	3	81
Bt (M)	647	-156	1580		0.4		
P5							
Oi (L _L)	191	127	5	67	NA	NA	NA
Oe + Oa (L _R)	—	—	NA		NA		
Oe + Oa (H)	406	144 ^c	7-13	55	58	16	39
A (LR)	210		NA		NA		
A (H)	3023	40	195		15.5		
A (M)	70	22	230		0.3		
E (H)	452	21	235		1.9	8	49
E (M)	796	59	142		5.6		
Bt (H)	91	31	210		0.4	1	-1
Bt (M)	342	-25	625		0.5		
Bt1 (H)	63	-139	1400		0.0	0.4	-87
Bt1 (M)	256	-80	830		0.3		

Note: Fluxes are derived by dividing inventory by the steady-state ^{14}C -derived TT. Where there is more than one possible TT shown, the bold values indicate the one used to calculate the flux. See text for further details.

^a Not available.

^b NA = Not applicable, see text for details.

^c Represents the bulk $\Delta^{14}\text{C}$ value; however, picking of different components resulted in values from 130‰ to 136‰.

^d Represents the $\Delta^{14}\text{C}$ value of a composite sample after fine roots have been picked out.

finer size fraction ($< 80 \mu\text{m}$) was lower in ^{14}C than the coarser size fraction ($> 80 \mu\text{m}$; Fig. 11.2).

Comparison of low-density organic matter sampled at the P5 site in 1972 and 1998 showed decreases in radiocarbon between the two sampling dates (Fig. 11.2B) ranging from 423‰ to 229‰ in the Oi and Oe + Oa horizons, respectively, to 296–84‰ in the mineral horizons. In contrast, the $\Delta^{14}\text{C}$ signature of high-density mineral-associated SOM samples are similar to samples from 1972 having $\Delta^{14}\text{C}$ values only 23–48‰ greater than those in 1998. The less than 80 μm fraction of the low-density material in the A horizon contains less radiocarbon than the greater than 80 μm fraction for both 1972 and 1998.

Radiocarbon values for the L_L, L_R, H, and M fractions and their associated ^{14}C -derived turnover times, and CO₂ fluxes (calculated by inventory divided by turnover time) are shown for each horizon below the Oi horizon in Table 11.2. Turnover times increase with depth from <10 to 1400 years for low-density material and from 200 to 1500 years in the high-density material. Calculated CO₂ fluxes associated with the decomposition of SOM decrease rapidly with depth from 45 to 60 g C m⁻² y⁻¹ in the Oe + Oa horizons to < 3 g C m⁻² y⁻¹ in the lower B horizons. Decomposition fluxes are not calculated for the Oi horizons because lagged radiocarbon inputs (from either stored photosynthate or components that reside on the plant for more than 1 year) make flux estimations too uncertain for material that cycles on less than 5-year timescales. As discussed earlier, the unusual ^{14}C history of this site further complicates the use of the atmospheric record of $^{14}\text{CO}_2$ for distinguishing among turnover times faster than 10 years at the Walker Branch site.

Discussion

Radiocarbon-Derived SOM Dynamics

Knowledge of C stock sizes and their average turnover times allows for evaluation of the potential for SOM to accumulate or lose C during the coming centuries under different scenarios of climate change or ecosystem productivity. The residence time of C in SOM pools is reflected in how much bomb ^{14}C they have incorporated since 1950 and in whether or not ^{14}C values have increased or decreased between 1972 and 1990 (Fig. 11.1). At Walker Branch, the greatest amount of bomb ^{14}C is in low-density fractions in the uppermost portion of the soil profile. The amount of bomb ^{14}C decreases with depth (Fig. 11.2). Most C with less than 10-year turnover times is in the uppermost soil horizons (to about a 5-cm depth) and represents a very small percentage (9%) of the total soil C stocks (excluding roots; Table 11.2). Carbon with turnover times of 10–100 years is found mostly between 5 and 40 cm in depth in both low- and high-density fractions and makes up 77% of the total SOM. Carbon cycling on timescales longer than 100 years predominates below a depth of 40 cm and constitutes ~14% of SOM stocks.

The model we use to estimate turnover times assumes that each fraction of SOM is homogeneous (with respect to turnover). Clearly, the low-density fraction of organic matter can be separated into different components with different cycling rates and, thus, $\Delta^{14}\text{C}$ values. This is especially true in surface horizons (Fig. 11.2A). For example, in the A horizon at the TDE site, roots make up 16% of the total sample mass and have ^{14}C signatures 130‰ higher than the $> 80 \mu\text{m}$ fraction from the same horizon. In this case, removing roots increases the composite turnover time for the nonroot fraction from 65 to 90 years. The organic Oi and Oe + Oa horizons at Walker Branch, unlike the mineral horizons, have 1998 $\Delta^{14}\text{C}$ values above the $\Delta^{14}\text{C}$ of CO_2 for the current atmosphere. In such cases, two different k values predict the same $\Delta^{14}\text{C}$ values for the SOM fraction (Fig. 11.3). The $\Delta^{14}\text{C}$ values of organic matter for archived soils sampled in 1972 at the P5 site allow us to distinguish which of the two possible k values is the more reasonable. $\Delta^{14}\text{C}$ values in O horizons decreased between 1972 and 1998, indicating that faster turnover times are the best fit to the data (Fig. 11.3). Estimates of turnover time close to 10 years are limited in precision because of the lack of change in ^{14}C values for atmospheric CO_2 between 1990 and 1998 (Fig. 11.1). However, the radiocarbon data clearly show the Oi and Oe + Oa horizons have turnover times of < 10 years and 10 ± 5 years, respectively. For comparison, the Oe + Oa horizons at Howland, Maine and Harvard Forest, Massachusetts have turnover times of $\sim 40 \pm 10$ years, with the Howland site also constrained

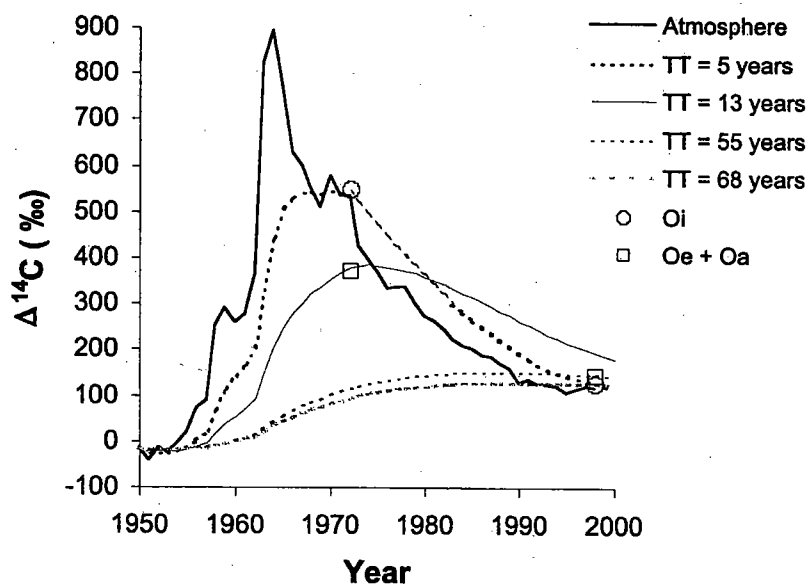


Figure 11.3. $\Delta^{14}\text{C}$ values for atmospheric CO_2 and organic horizon components for archived and modern soils and modeled (steady-state) turnover time curves. The Walker Branch input data are derived from tree-ring cellulose starting in 1990 (to account for the unusual atmospheric ^{14}C history at this site; see text for details).

by comparison of archived and contemporary soil samples (Gaudinski 2001). Our turnover-time estimates may, in fact, be too long by 1–5 years if there are any lagged ^{14}C inputs caused by carbohydrate storage prior to plant tissue (root or leaf) formation (Gaudinski et al. 2000; Trumbore 2000).

We estimate the flux of CO_2 derived from heterotrophic decomposition of each SOM fraction by dividing its stock by its ^{14}C -derived turnover time. We limited this analysis to the humified low-density and mineral-associated SOM fractions of the Oe + Oa and mineral horizons with turnover times of 10 years or more. Combined fluxes for low- and high-density fractions range from 45 to 58 $\text{g C m}^{-2} \text{y}^{-1}$ in the Oe + Oa horizons, from 6 to 16 $\text{g C m}^{-2} \text{y}^{-1}$ in the A horizons, and from 8 to 14 $\text{g C m}^{-2} \text{y}^{-1}$ in the E horizons and are less than 3 $\text{g C m}^{-2} \text{y}^{-1}$ in the lowermost B horizons (Table 11.2). The majority is from the Oe + Oa horizon (53%) and the A + E horizons (45%), whereas only 2% comes from the B horizons. Therefore, the total calculated decomposition flux from SOM decomposing on 10-year time scales or greater is 77 $\text{g C m}^{-2} \text{yr}^{-1}$ (9% of the total soil respiration flux) with 6% from the Oe + Oa horizon and 3% from the mineral horizons. These fluxes represent an average over several years to decades and are not necessarily indicative of decomposition dynamics in any given year. We do not include the Oi horizon in this calculation because of our inability to determine turnover time accurately from radiocarbon for the reasons described.

Our results compare well to component-integration studies also performed on the Oak Ridge Reservation. Edwards et al. (1989), working in a *Quercus prinus* stand on Walker Branch, measured respiration rates for the whole soil profile, litter, and roots (autotrophic + heterotrophic sources) and estimated (by difference) that 32 $\text{g C m}^{-2} \text{y}^{-1}$ came from decomposition of SOM in the mineral horizons (5% of the total soil respiration flux). Using a similar approach, Edwards and Harris (1977) found 23 $\text{g C m}^{-2} \text{y}^{-1}$ came from the decomposition of SOM in the mineral horizons (2.2% of the total soil respiration flux) and 103 $\text{g C m}^{-2} \text{y}^{-1}$ came from the O₂ horizon, which is comparable to the Oe + Oa horizon (10% of the total soil-respiration flux).

Potential for Soil C Accumulation

Using the soil C stock and turnover time information presented earlier, we can model the potential for the soils of Walker Branch to store carbon in response to climate change. To do this, we created a four-pool model that tracks C stocks:

$$\frac{dC}{dt} = \sum_{j=1}^{j=4} I_j - k_j C_j, \quad (3)$$

where C_j represents the carbon stock in pool j , k_j equals $(\text{turnover time}_j)^{-1}$, and I_j equals the inputs to pool j . The four pools represent (1) leaf litter (Oi horizon), (2) the Oe + Oa horizon, (3) low-density SOM in the mineral horizons, and (4) high-density SOM in the mineral horizons. Inputs to each pool were calculated on the basis of initial steady-state conditions and the ^{14}C -derived k (turnover time).

The commonly cited process driving C sequestration in the coming century is an increase in forest net primary production (NPP) (caused by CO₂ or N fertilization) that will be larger than any decreases in overall turnover times (i.e., increased decomposition rate because of increases in mean annual temperature or precipitation changes). Thus, we modeled the response of Walker Branch soils to a 25% increase in NPP, which is similar to that observed in a young loblolly pine plantation after 2 years of CO₂ enrichment (at 565 ppm) (DeLucia et al. 1999). We ran the four-pool carbon model for average, high, and low cases. The high and low cases use the range of measured C stock and turnover-time values that respectively maximize or minimize C storage potential (Table 11.2 and Table 11.3). Our results indicate total C accumulation after 100 years of 505, 1140 and 235 g C m⁻² for the average, high, and low cases, respectively (which are 39%, 63%, and 31% of the new steady-state value under the 25% NPP increase; Table 11.3 and Fig. 11.4A). The rate of C accumulation in the first year after the increase in NPP is between 30 and 120 g C m⁻² y⁻¹ (Fig. 11.4B). However, by the fifth year, the rates are between 7 and 22 g C m⁻² y⁻¹, and after 100 years, they drop to between 1 and 6 g C m⁻² y⁻¹. This rapidly decreasing rate of C storage is reflected in the *e*-folding times of 89, 38, and 151 years, respectively (Table 11.3). During the coming century, we estimate the average annual C accumulation rate in these soils to be between 2 and 11 g C m⁻² y⁻¹ (with 5 g C m⁻² y⁻¹ for the average case; calculated as the average for the initial 100-year period after a sustained 25% increase in NPP).

On timescales of human interest, C storage within the soil profile at Walker Branch will be predominantly within low-density humified organic components that have turnover times of a decade to several hundred years. Detrital C, such as leaf litter in the Oi horizon, will be a poor C sink because it turns over rapidly (< 10 years), has small C stocks, and reaches steady state within years to decades (Fig. 11.5). The Oe + Oa and A + E horizons, however, have low-density C stocks with decadal turnover times and therefore, the potential to store hundreds of grams of carbon on human timescales (Fig. 11.5, Table 11.2). High-density, mineral-associated carbon stocks, which dominate total C stocks below the A + E horizon, have low inputs and long turnover times (200–2000 years) and can store large amounts of C (> 100 g m⁻²) only over several centuries (Fig. 11.5; Table 11.2).

Validity of the Steady-State Assumption and Current Rates of C Accumulation

The time-dependent ¹⁴C model we used to determine turnover times and the flux estimates we have based on those turnover times both assume that Walker Branch soils are at steady state with respect to carbon. The ¹⁴C and C-stock data available are too few to determine the validity of this assumption. However, there are several reasons why it might be reasonable to assume that C may be accumulating in soils at this site. First, a nearby (2 km away) eddy flux tower shows large amounts of annual C uptake at this site and there are published estimates that a

Table 11.3. Parameters used in soil carbon model and the resulting carbon increase after a 25% increase in net primary production

Pool	Average			High			Low			Harvard Forest		
	Stock (g C m ⁻²)	TT ^a (year)	Input ^b (g C m ⁻² y ⁻¹)	Stock (g C m ⁻²)	TT (year)	Input ^b (g C m ⁻² y ⁻¹)	Stock (g C m ⁻²)	TT (year)	Input ^b (g C m ⁻² y ⁻¹)	Stock (g C m ⁻²)	TT (year)	Input ^b (g C m ⁻² y ⁻¹)
Leaf litter	275	1	220	360	1	360	190	2	95	150	1	150
Oe + Oa	340	10	36	410	6	68	270	13	21	1800	40	45
LD	2660	168	16	4150	85	49	1170	250	5	1800	100	18
HD	1900	850	2	2300	200	12	1450	1500	1	4620	500	9
SS total ^c	5175			7220			3080			8370		
C increase at new SS ^d	1295			1805			770			2095		
C increase after 100 years ^e	505			1140			235			950		
e-folding time ^f	89			38			151			67		

Note: First three sets of columns refer to Walker Branch. The fourth set of columns refers to Harvard Forest, MA. See text for further details.

^a TT = turnover time.

^b Inferred steady-state inputs (i.e., stock/TT).

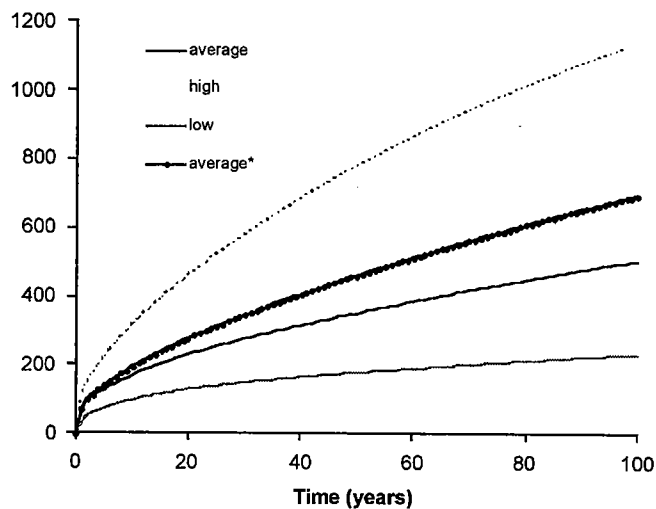
^c Steady-state carbon inventory.

^d Carbon stock increase due to 25% increase in net primary production after reaching new steady state.

^e Carbon stock increase 100 years after net primary production increased by 25%.

^f Time (years) for the stock to change to 1/e times the new steady-state value.

A.



B.

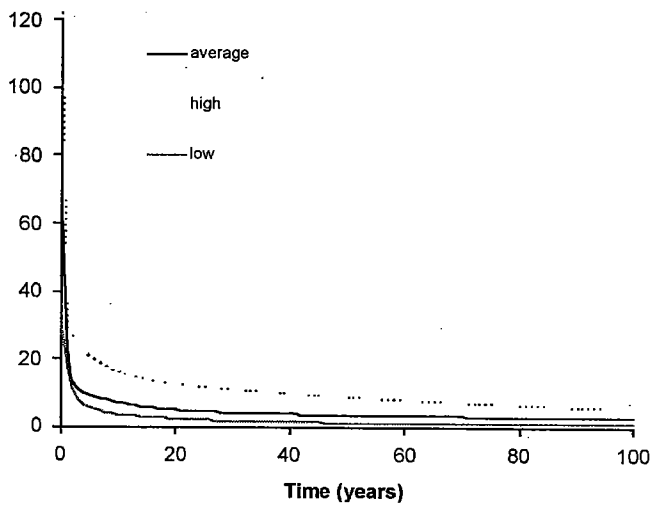


Figure 11.4. Soil carbon response after a 25% increase in NPP for the high, low, and average soil carbon stock and cycling rate scenarios (Table 11.3). (A) The three average, high, and low scenarios initially at steady state plus the average case initially accumulating $10 \text{ g C m}^{-2} \text{ y}^{-1}$ (average*). (B) Rate of change in soil C storage over time for the high, low and average case shown in (A).

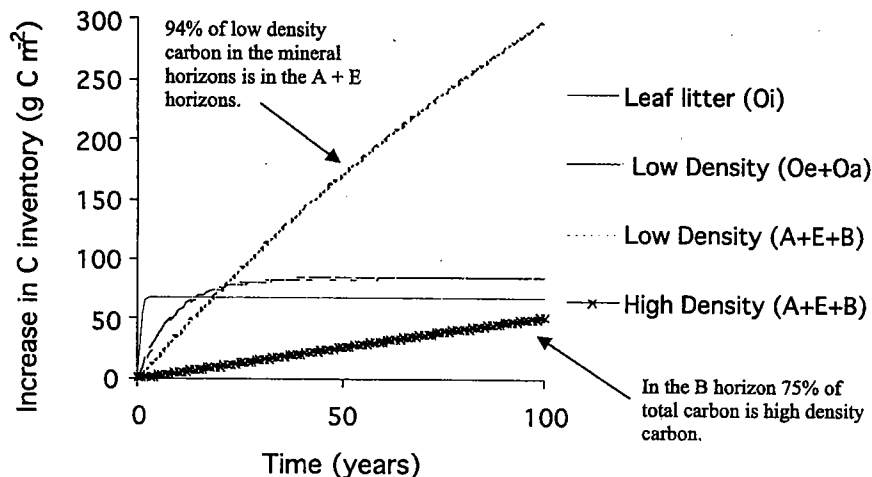


Figure 11.5. Carbon accumulation in response to a sustained 25% increase in NPP (average case) for different soil fractions.

large portion of that storage must be going into the soil (Malhi et al. 1999). Second, this site was selectively logged and grazed in the early 1900s, and the forest has regrown and been undisturbed since the 1940s. Finally, historic increases in atmospheric CO₂ concentrations and enhanced N deposition are thought to lead to increases in NPP that increase soil C storage.

To determine the magnitude of and any possible error associated with the steady-state assumption in an aggrading system, we used a non-steady-state accumulation model (Gaudinski et al. 2000) that matches the current C and ¹⁴C stocks in 1998 with different values for inputs (*I*) and turnover times (*1/k*; where both *I* and *1/k* are constants) and assumes that all C in the humified components of the Oe + Oa horizon has accumulated in the past 100 years. Inputs of 30–50 g C m⁻² y⁻¹ and turnover times of 6–7 years matched the C and ¹⁴C data, respectively, and our model then predicts accumulation rates of << 1 g C m⁻² y⁻¹ in 1998. Hence, C and ¹⁴C stocks are consistent with a steady-state assumption for the Oe + Oa horizons. Turnover times of 6–7 years are on the low end of the measured range (6–13 years) (Table 11.2). However, turnover times of 10–15 years in an accumulating scenario produce ¹⁴C values larger than any measured by 40–70‰ and thus provide further confirmation that this horizon is at steady state. Using this same approach for the low-density fractions of the A horizon (with inputs of 10–20 g C m⁻² y⁻¹ and turnover times between 250–350 years), a minimum of 200 years of accumulation is required to match the C and ¹⁴C data. The model then results in 5–10 g C m⁻² y⁻¹ of accumulation in 1998.

Current accumulation rates in the E and B horizons will be even less than those in the O and A horizons because of their relatively small stocks of low-density SOM (Fig. 11.5, Table 11.1). Thus, our turnover time and flux estimations likely have little error from the steady-state assumption in the O and mineral E and B

horizons. However, our steady-state assumption for the A horizon may not be entirely valid. If we initialize our average C stock model with accumulation in the low-density fraction of $10 \text{ g C m}^{-2} \text{ y}^{-1}$ (the A horizon contains 68% of the low-density material in the mineral horizons), the added C storage over 100 years (relative to a steady-state case) is less than 200 g C m^{-2} and still less than the range bracketed by our high scenario (Fig. 11.4A), thus having no effect on our estimate of potential C storage following a sustained 25% increase in NPP.

Our results of relatively low C-storage potential in Walker Branch soils are in accord with the results reported by Trettin et al. (1999), who measured changes in C stocks for eight long-term plots located in similar forest stands on Walker Branch. Their results showed C gains and losses between 1972 and 1993 in Oi and Oa horizons ($\sim 100\text{--}400 \text{ g C m}^{-2}$ over the 11-year period), with most actually showing losses. The A horizons (0–15 cm) in their study tended to increase in % C (although not significantly) in five of the eight long-term plots. The B horizons (45–60 cm) actually show % C losses in five of their eight long-term plots between 1972 and 1993 (with significant losses in two plots). The net soil C balance shown by Trettin et al. (1999) for the whole soil profile (0–60 cm) between 1982 and 1993 is a trend toward C loss (significant C loss in only one plot). Johnson et al. (1988), however, show that soils (from 0–45 cm) on the Oak Ridge reservation adjacent to Walker Branch accumulated 1160 g C m^{-2} during a 15-year period (from 1980 to 1995; $\sim 80 \text{ g C m}^{-2} \text{ y}^{-1}$ if the rate was constant). Time-series measurements of C stocks have inherent problems relating to large spatial variability; however, the data of Trettin et al. (1999) urge caution in any implicit assumption of C accumulation in soil C stocks, even in sites where accumulation seems very plausible.

Implications for Ecosystem Carbon Budgets of Walker Branch

Net forest C balance is the sum of C gains or losses in aboveground versus belowground biomass. At Walker Branch, eddy covariance techniques for a similar site 2 km away show net ecosystem production (NEP) of $580 \text{ g C m}^{-2} \text{ y}^{-1}$ (Wilson and Baldocchi 2001). Current estimates of aboveground biomass accumulation in wood on Walker Branch are $220 \text{ g C m}^{-2} \text{ y}^{-1}$ and for coarse woody roots are $45 \text{ g C m}^{-2} \text{ y}^{-1}$ (Chapter 22, this volume). Solving for the remaining belowground storage by difference would yield $315 \text{ g C m}^{-2} \text{ y}^{-1}$ of soil C storage [an increase that should be measurable with the methods of Trettin et al. (1999)]. Our results, however, show that the soils of Walker Branch are currently storing $< 15 \text{ g C m}^{-2} \text{ y}^{-1}$, and under changing climatic conditions, they likely have the potential to store no more than an additional $11 \text{ g C m}^{-2} \text{ y}^{-1}$. Although some of the above-discussed assumptions introduce errors in our estimates, sustained storage rates $> 100 \text{ g C m}^{-2} \text{ y}^{-1}$ are not possible with reasonable scenarios of changing C inputs. This discrepancy will be discussed further in Chapter 22.

In forest ecosystems where eddy-covariance techniques show net uptake, an increase in soil C should not necessarily be assumed. As discussed earlier, Trettin et al. (1999) show a general trend (although significant in only one plot) of soil

C loss (for 0–60 cm) for eight long-term plots on Walker Branch between 1972 and 1993. An alternative biometric estimate of net ecosystem C storage at Walker Branch based on summing directly measured aboveground and belowground C fluxes (and assuming heterotrophic respiration is 50% of total soil respiration) yields an NEP of $250 \text{ g C m}^{-2} \text{ y}^{-1}$ (Curtis et al. 2002). Using the same estimates of C accumulation in wood and coarse roots discussed earlier in this chapter and in Chapter 22, soil C accumulation is then estimated to be close to zero and decreasing.

Beyond Walker Branch

Current soil C-storage rates at Walker Branch ($< 15 \text{ g C m}^{-2} \text{ y}^{-1}$) are similar to that of Harvard Forest, MA ($10\text{--}30 \text{ g C m}^{-2} \text{ y}^{-1}$) (Gaudinski et al. 2000), a more northerly mixed deciduous temperate forest in the eastern United States. We calculate that a 25% increase in NPP would cause C storage over the next century of 950 g C m^{-2} at Harvard Forest and 505 g C m^{-2} at Walker Branch using an average of measured C stocks and turnover times for both sites (Table 11.3) (Gaudinski et al. 2000). Total C-storage potential following this perturbation is larger at Harvard Forest primarily because of larger stocks and longer relative turnover times in Oe + Oa horizons (Table 11.3). Although such a gradient is not surprising (given the differences in climate), the increased annual C accumulation rate at both sites is still $< 10 \text{ g C m}^{-2} \text{ y}^{-1}$ (calculated as the average over the initial 100-year period after a sustained 25% increase in NPP).

Our rates of current and potential C accumulation are in accord with research done in other natural ecosystems. Schlesinger and Lichter (2001) predict the potential for accumulation of $< 60 \text{ g C m}^{-2} \text{ y}^{-1}$ in a young loblolly pine plantation as atmospheric CO_2 concentrations increase through time. In a literature review, Post and Kwon (2000) found that, after agricultural abandonment, average rates of soil C accumulation were $34 \text{ g C m}^{-2} \text{ y}^{-1}$ for reestablishing forests or grasslands. Agricultural soils in the United States are also estimated to be able to store $\sim 20 \text{ g C m}^{-2} \text{ y}^{-1}$ under certain management conditions that increase C inputs to soil (Lal et al. 1999). Rates of soil C storage in tropical forest soils are also on the order of tens of $\text{g C m}^{-2} \text{ y}^{-1}$ (Trumbore, unpublished data).

Conclusions

Using radiocarbon to constrain turnover rates of SOM, we have shown that bulk SOM consists of fractions that will respond at quite different rates to a change, such as an increase in NPP. Five percent of the carbon resides in leaf and root residues with turnover times of < 10 years. These pools will respond quickly to changes and have small inventories, so they are not expected to play a large role as C sinks or sources, although they may dominate year-to-year variation in soil heterotrophic respiration rates. Thirty-five percent of the C is stabilized by association with soil minerals and has turnover times > 100 years; although C stocks in this fraction are large, they are not expected to change dramatically

during the next century of climate or even land-use change. Sixty percent resides in low density and humified organic matter in mineral soils with turnover times of 10–100 years; it is these fractions that will determine C storage potential of soils and that require further investigation as to their sources and what determines their turnover time in soils.

Our estimates of current and future C-storage capacity at Walker Branch are an order of magnitude smaller than those needed to reconcile eddy flux estimates of annual net carbon uptake and estimates of aboveground carbon storage at this site. This result suggests that methodological issues (perhaps those associated with an underestimation of nocturnal respiration of C by ecosystems) seriously affect eddy-covariance results from Walker Branch and should be investigated further.

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