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Bomb ^{14}C enrichment indicates decadal C pool in deep soil?

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Abstract Studies of changes in soil organic carbon (SOC) stocks normally limit their focus to the upper 20–30 cm of soil, yet 0–20 cm SOC stocks are only ~40% of 0–1 m SOC. Accounting for only the upper 20–30 cm of SOC has been justifiable assuming that deeper SOC is unreactive since it displays ^{14}C-derived mean residence times of hundreds or thousands of years. The dramatic increase in the ^{14}C content of the atmosphere resulting from thermonuclear testing circa 1963 allows the unreactivity of deep SOC to be tested by examining whether deep soils show evidence of 'bomb-^{14}C' incorporation. At depths of 40–100 cm, a well-studied New Zealand soil under stable pastoral management displays progressive enrichment of over 200‰ across samplings in 1959, 1974 and 2002, indicating substantial incorporation of bomb ^{14}C. This pattern of deep ^{14}C enrichment—previously observed in 2 well-drained California grassland soils—leads to the hypothesis that roots and/or dissolved organic C transport contribute to a decadally-reactive SOC pool comprising ~10–40% of SOC below 50 cm. Deep reactive SOC may be important in the global C cycle because it can react to land-use or vegetation change and may respond to different processes than the reactive SOC in the upper 20–30 cm of soil.

Keywords Soil organic matter · Carbon cycle · Radiocarbon · Model · Dissolved organic matter · Roots · Soil profiles · Soil depth

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Introduction

Globally, soil organic matter (SOM) contains 1500 Pg C to 1 m soil depth (Post et al. 1982) and 2300 Pg C to 3 m depth (Jobbagy and Jackson 2000)—more than biomass and atmospheric CO_2 combined. The soil organic carbon (SOC) pool is believed to represent a historic source of C to the atmosphere and a potential sink in the future. National greenhouse gas inventories reported under the United Nations Framework Convention on Climate Change and accounted for under the Kyoto Protocol contain estimates of soil organic carbon (SOC) changes associated with land use, land-use change and forestry. It has been assumed that SOC changes related to land-use change occur almost entirely in the upper 30 cm of soil, because deep SOC is inert on timescales <100 years and is transported slowly downward relative to plant sources (Baisden et al. 2002a; O'Brien and Stout 1978). As a result of these assumptions, most ecosystem biogeochemistry models that estimate the response of vegetation and soil to climate change, land management, and other factors, consider C dynamics in only the upper 20–30 cm of soil. Yet, estimates of SOC stocks generally recognize that SOC below 30 cm is a significant stock, and that SOC is distributed somewhat more deeply than plant roots (Jobbagy and Jackson 2000).

Since little mechanistic understanding exists to explain deep SOC dynamics, we resampled a New Zealand grassland soil for which now classic models of SOC turnover and transport were developed (O'Brien and Stout 1978). We build on the modeling methodology developed by Baisden et al. (2002a), and also compare the New Zealand grassland soil to two California grassland soils under substantially different climate. We therefore add recent soil radiocarbon measurements, performed to ~1 m depth, to measurements performed during 1949–1959 and 1974–1978. These measurements occur over a time period ideal for calculating the turnover rates of decadal C pools using the ^{14}C-enrichment of the atmosphere from thermonuclear weapons testing—a near doubling of atmospheric $^{14}C/^{12}C$ ratio in 1963 (Trumbore 1993). To minimize the influence of land management, we examined soil profiles believed to be under stable grazing management since at least ca. 1900. Our approach is designed to test for and quantify the amount of reactive SOC below 30 cm depth (henceforth referred to as deep RSOC) that cannot be considered to have residence times of centuries or millennia, and therefore can be expected to react to global changes.

Methods

Sites and sampling

The Judgeford, Riverbank and Turlock Lake sites (Table 1) have been described elsewhere (Baisden et al. 2002a; Harden 1987; Lassey et al. 1996; O'Brien and Stout 1978). Briefly, the Judgeford soils formed on quartz/feldspathic loess under indigenous forest cover until European colonization of New Zealand, and have been converted to *Agrostis capillaris* dominated grassland since ca. 1900 (Parfitt et al. 1984). The Riverbank and Turlock Lake sites have soils derived from granitic alluvium (Harden 1987) and are dominated by annual grasses (mainly European species), but had a greater cover of native perennial grasses before ~150 years ago. Mean annual temperatures and precipitation are 12.6°C and 1290 mm at Judgeford and 16°C and 300 mm at Riverbank and Turlock Lake. Differences between the Riverbank and Turlock Lake sites can be associated with soil age (~200 versus ~600 ky, respectively) and texture (sandy loam versus loamy sand, respectively) (Harden 1987). None of the three sites has received significant inputs of N or P fertilizer.

The 1974 and 2002 samplings of the Judgeford soils occurred at locations ~1 km and ~200 m distant from the original site, at sites with closely matched landform, aspect, and hillslope position, as well as nearly identical soil horizons. For the 2002 sampling in particular, a matching site was carefully chosen based on stable land-use history, since the

Table 1 Site descriptions

Site	Soil Age	Soil Parent Material	Mean Annual Temp.	Mean Annual Precip.	Typical Soil Texture[d]
Judgeford[a], New Zealand (41.11° S, 174.95° E)	~15 ky[b]	Quartz/ Feldspathic Loess[b]	12.6°C	1290 mm	Silt Loam
Riverbank[c], California, USA (37.52° N, 120.59° W)	~200 ky	Granitic Alluvium	16°C	300 mm	Sandy Loam
Turlock Lake[c] California, USA (37.63° N 120.59° W)	~600 ky	Granitic Alluvium	16°C	300 mm	Loamy Sand

[a] O'Brien and Stout (1978)

[b] Parfitt et al. (1984)

[c] Baisden et al. (2002a)

[d] Textures do not vary widely in these profiles except Turlock Lake, where an argillic horizon below 62 cm has sandy clay and sandy clay loam textures

1959 site has undergone disturbance. Samples were taken from a soil pit, and sieved to remove roots >2 mm. Bulk density was measured by coring. No stones were encountered. Radiocarbon data are expressed in $\Delta^{14}C$ notation, corrected for isotopic fractionation using $\delta^{13}C$ data. All data represent bulk soils, combusted without any pretreatments and noting that no carbonates were present. Samples from 2002 were measured using AMS at the Rafter Radiocarbon Laboratory, Lower Hutt, New Zealand, while other results are drawn from the literature cited. We use the notation $\Delta^{14}C_{sample} = 1000 \times ((^{14}C/^{12}C_{sample})/(^{14}C/^{12}C_{standard}) - 1)$, where the standard is 95% of the activity of NBS Oxalic Acid—approximately representing the composition of pre-industrial atmosphere. The average analytical uncertainty reported by the AMS laboratories for the Judgeford, Riverbank and Turlock Lake profiles was 7‰, 5‰ and 4‰, respectively.

Modeling

Our mass-balance model for the turnover and transport of the 3 pools of SOC throughout soil profiles has already been described elsewhere (Baisden et al. 2002a). The model represents 3 pools of SOC uniformly as a continuous function of soil depth, assuming that each has a first-order decomposition rate and downward transport rate. Inputs to the SOC pool occur both through aboveground litter at the soil surface and through roots. The model calculates SOC storage and isotopic composition as the sum of all pools as a function of cumulative soil mass, which is calculated by normalizing soil depth to a bulk density of 1 g cm^{-3}. To this model, we added an empirical representation of a deep RSOC pool described by 4 parameters: the total amount of deep RSOC; an e-folding depth (representing the depth at which the concentration of deep RSOC is 1/e of the concentration at the soil surface); a downward transport rate; and a decomposition rate constant for the pool. The representation for deep RSOC does not represent a mass-balance model connected to the other 3 pools. Instead, it provides the minimum number of adjustable parameters to fit the changes in $\Delta^{14}C$ observed below 30 cm soil depth, and is consistent with interpreting deep RSOC as resulting from dissolved SOC transport or root inputs. Atmospheric $\Delta^{14}C$ data for the southern hemisphere was obtained from a spline fit to the Baring Head, New Zealand data (Manning et al. 1994).

Thus, RSOC is defined as SOC that is reactive on timescales <100 years: RSOC is the combination of the active ($\sim$annual turnover), stabilized (decadal) and deep RSOC pools. The passive or inert (millennial) SOC pool is therefore not part of RSOC, but part of total SOC.

Due to the likelihood that multiple plausible solutions exist for a multi-pool model fitted to soil $\Delta^{14}C$ data (Trumbore 2000), a sophisticated approach is required to obtain automated solutions and uncertainty estimates, in contrast to the manual fitting procedures used previously (Baisden et al. 2002a). The bulk soil %C and $\Delta^{14}C$ estimated by the model for all 4 pools of SOC was optimized against the observed data for each profile in MATLAB (The MathWorks, Natick, MA, USA) using a Genetic Algorithm (GA) Toolbox (Sheffield 1994). Genetic algorithms have been used previously to fit complex C turnover models (Barrett 2002), and represent an alternative to the Markov chain Monte Carlo approach (Xu et al. 2006).

Briefly, the concept of a GA is that a range of solutions (where each solution comprises estimates for each parameter value specified in the model) can be seen as an analog to individuals in a population undergoing evolution. Each solution represents an individual with genetic material (chromosomes) represented by a set of parameter values. Initially, the population has individual parameter values spread randomly throughout the possible parameter space, and evolves by ‘mating’ individuals so that new parameter values are determined in a random space (a hypercube) in the vicinity of two parents, also allowing for mutations. Fitness of individuals in the population is determined by an objective function describing how well the solution fits observations, and the population is maintained by culling the least fit individuals at each generation.

The following details specify the approach used with the GA toolbox (Sheffield 1994). Solutions represent the fittest individuals from multiple runs of a single population of 50 real-valued chromosomes evolved over >300 generations. The objective function describing fitness was the sum of squares of differences between observed and modeled bulk soil $\Delta^{14}C$ and $\log_{10}$(%C), using a factor of 1000 for $\log_{10}$(%C) to ensure that both types of data were given similar weighting. For the model, the $\Delta^{14}C$ and $\log_{10}(\%C)$ data were averaged across horizon depth increments to match sampled horizons. In the options for the GA toolbox, we used the 'intermediate recombination' option allowing new parameter values to be between the parameter values of reproducing individuals (parents) as well as higher or lower by 50% of the difference between parents.

For the determination of uncertainties using a Monte Carlo approach, the genetic algorithm was rerun with random variation in $\Delta^{14}C$ (normal distribution; σ = 5‰) and %C (log normal distribution; σ = 5% coefficient of variance) to estimate uncertainty and sensitivity. For the Judgeford site, plant C inputs were allowed to vary within bounds (aboveground inputs between 0.5 and 1 times average aboveground NPP, belowground inputs between 0.5 and 3 times aboveground NPP) established based on field sampling (Saunders and Metson 1971). For the Turlock Lake and Riverbank sites, measured plant C inputs were used, but varied with random variation (normal distribution; σ = 15‰) to estimate uncertainty and sensitivity.

Table 2 reports the greater of two approaches for calculating uncertainty. These two approaches are, (1) the standard deviation of the 50 individual estimates from the genetic algorithm, and (2) the difference between the fittest individual obtained without random variation and the mean of the individuals obtained with random variation. These reported uncertainties represent our best estimate of 1 σ for each parameter or quantity estimated, but are not directly related to the difference between the model and measured data. Interpretation of the results therefore requires consideration of both the reported uncertainties (Tables 2 and 3) and model goodness of fit (Figs. 1 and 2).

Results and discussion

Within the upper 30 cm the measurements displayed expected values: $\Delta^{14}C$ values are just below 0‰ in 1959, strongly enriched by bomb ^{14}C in 1974, and returning toward the pre-bomb values in 2002 (Fig. 1). However, at depths of 40–100 cm the

Table 2 Parameter values and uncertainty

Parameter	Explanation	Units	Judgeford	Riverbank	Turlock Lake
$1/k_1$	Active pool residence time	y	3.34 ± 0.09	0.50 ± 0.06	0.50 ± 0.05
$1/k_2$	Stabilized pool residence time	y	51 ± 4	21 ± 2	25 ± 3
$1/k_3$	Passive pool residence time	y	12000 ± 2000	9300 ± 500	>35,000[b]
k_{t1}	Partitioning coef.	–	0.32 ± 0.01	0.09 ± 0.01	0.08 ± 0.01
$k_{t2}/1000$	Partitioning coef.	–	0.34 ± 0.09	0.29 ± 0.04	0.43 ± 0.10
f_s	Surface litter C inputs	gC m^{-2} y^{-1}	265 ± 10	282[a]	178[a]
R	Root C inputs	gC m^{-2} y^{-1}	408 ± 91	279[a]	357[a]
v_1	Active pool downward velocity	mm y^{-1}	6.2 ± 1.3	0.6 ± 0.2	0.5 ± 0.4
v_2	Stabilized pool downward velocity	mm y^{-1}	0.9 ± 0.3	1.3 ± 0.1	0.5 ± 0.1
v_3	Passive pool downward velocity	mm y^{-1}	0.19 ± 0.01	0.25 ± 0.01	0.50 ± 0.05
L	e-Folding depth for root inputs	cm	29 ± 3	35 ± 1	9 ± 1
$1/P$	Rate of relocation of passive SOC to surface	y	13400 ± 400	123000 ± 9000	49000 ± 7000
D_p	$\Delta^{14}C$ of passive SOC relocation	‰	−600 ± 31	−520 ± 54	−323 ± 48
C_{DRSOC}	Normalized concentration of deep RSOC	–	0.39 ± 0.11	0.05 ± 0.01	0.10 ± 0.04
V_{DRSOC}	Downward velocity of deep RSOC	cm y^{-1}	3 ± 97[c]	14 ± 9	9 ± 25[c]
$1/k_{DRSOC}$	Residence time of deep RSOC	y	1.0 ± 1.0[c]	1.0 ± 0.2	11 ± 22[c]
L_{DRSOC}	e-Folding depth of deep RSOC	cm	623 ± 797	565 ± 67	668 ± 327

All parameters, excluding the 4 describing deep RSOC, are defined identically to those in Baisden et al. (2002a)

[a] Above and belowground vegetation C inputs were measured directly at these sites, rather than modeled

[b] Model result indicated an age greater than ages detectable by radiocarbon measurements

[c] High uncertainties for these parameters indicate two model solutions were possible ''on either side'' of the bomb−^{14}C spike

Table 3 Soil Organic Carbon (SOC) and Reactive SOC (RSOC) Pools Estimated by the Model

Site	Max. Depth Included in Analysis (m)	Total SOC (MgC/ha)	Total RSOC (MgC/ha)	Deep RSOC (MgC/ha)	Ratio of deep RSOC to Total SOC	Ratio of deep RSOC to total RSOC
Judgeford	1.0	192 ± 10	131 ± 7	10.3 ± 0.7[a]	0.054 ± 0.005[a]	0.078 ± 0.008[a]
Riverbank	3.6	39 ± 3	12 ± 2	9.0 ± 1.3	0.23 ± 0.05	0.74 ± 0.05
Turlock Lake	1.1	33 ± 2	13 ± 1	4.7 ± 3.3	0.14 ± 0.10	0.37 ± 0.27

[a] The model significantly underestimates the apparent size of the deep SOC pool. The gap between the model fit and $\Delta^{14}C$ data suggests deep SOC may be twice the size calculated by the model

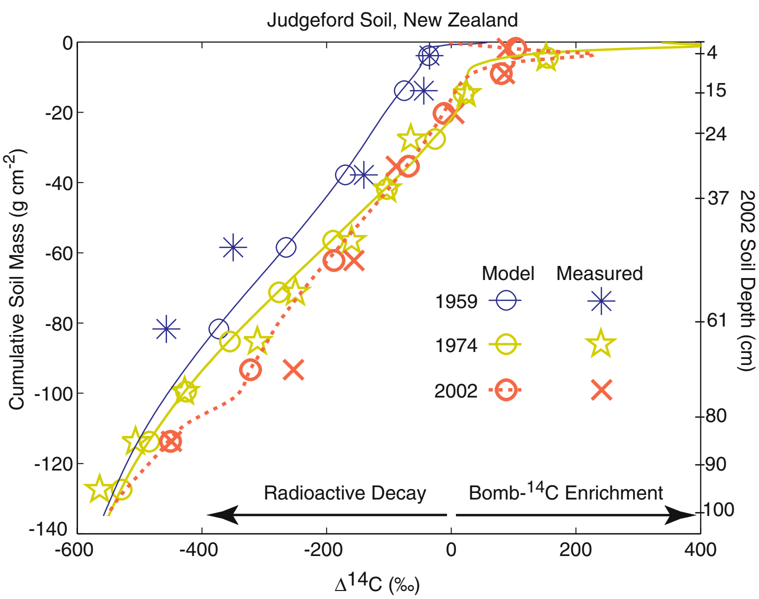

Fig. 1 Observed and modeled $\Delta^{14}C$ values for the 1959, 1974 and 2002 samplings of the Judgeford soil profile, New Zealand. Cumulative soil mass is equivalent to soil depth in centimeters with bulk density normalized to 1 g cm^{-3}

time-series profiles displayed a surprising and progressive enrichment of up to 200‰ over the 43 years, indicating incorporation of bomb ^{14}C. This incorporation of ^{14}C fixed since the 1960s cannot be explained by accumulation of C in the soil profile since the total profile C stocks to 1 m depth are 22.4 and 19.6 kg m^{-2} in 1959 and 2002, respectively. Similarly, for the 0.3–0.1 m depth increment, C stocks appear to have decreased from 9.3 kg m^{-2} to 7.2 kg m^{-2} between 1959 and 2002, respectively. While it is possible that this enrichment could have resulted from site selection, Fig. 2 shows that a similar but smaller enrichment was observed in both of the well-drained California grassland soil profiles (Baisden et al. 2002a).

We model the dynamics of C and ^{14}C in the soil profile using a 3-pool mass balance model that recognizes C inputs from the surface litter and roots, downward transport, and turnover (oxidation) of each C pool (Baisden et al. 2002a) (see methods). To account for the apparent downward transport and retention of bomb$-^{14}C$ below 30 cm depth, which was also observed in two well-drained California grassland soils (Baisden et al. 2002a) (Fig. 2), we added a fourth pool of SOC to the model—representing deep RSOC. Deep RSOC was empirically modeled based on four parameters: the total amount of deep RSOC; an e-folding depth; a downward transport rate; and a decomposition rate constant for the pool.

In contrast to previous studies providing estimates of SOC turnover rates for the three pool model (Baisden et al. 2002a; Baisden et al. 2002b), we optimized the model presented here to determine the mass and turnover rates of the SOC pool associated with bomb-^{14}C enrichment observed below 30-cm

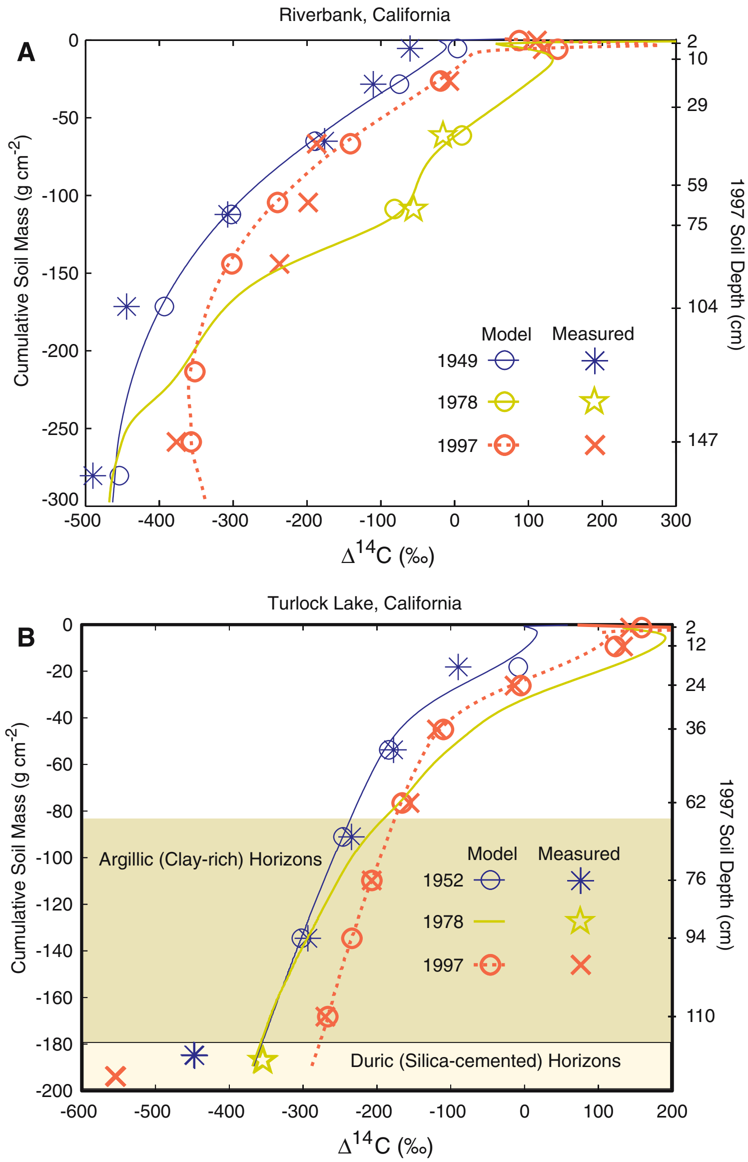

Fig. 2 Observed and modeled $\Delta^{14}C$ values for the Turlock Lake and Riverbank sites, California, USA, with data previously reported by Baisden et al (2002a). Cumulative soil mass is equivalent to soil depth in centimeters with bulk density normalized to 1 g cm^{-3}

depth. Parameter values (Table 2) for the Judgeford soil differ approximately as expected from those determined for the California soils given the differences in climate and soil parent material. The parameter values are within the range expected based on existing models such as CENTURY and RothC (Parshotam 1996; Parton et al. 1996; Smith et al. 1997).

Figures 1 and 2 demonstrate that the model captures the general pattern of variation seen in soil $\Delta^{14}C$ as a function of soil depth. The fit in the upper 30 cm is similar to that in Baisden et al (2002a) but is improved below this depth showing that the model now includes a mechanism for fitting bomb-^{14}C enrichment at deeper depths. In the California soils (Fig. 2), the model including the deep SOC pool fully fits the deep soil bomb-^{14}C enrichment, supporting the use of the pool sizes and uncertainties reported in Table 3. However, the model does not fully account for the ~200‰ bomb-^{14}C enrichment observed in the Judgeford soil profile (Fig. 1). The inability of the model to capture the full bomb-^{14}C enrichment most likely results from the model's assumption of uniformity of processes as a function of soil depth. For example, hydrologically driven transport rates are known to decrease with soil depth as plants remove

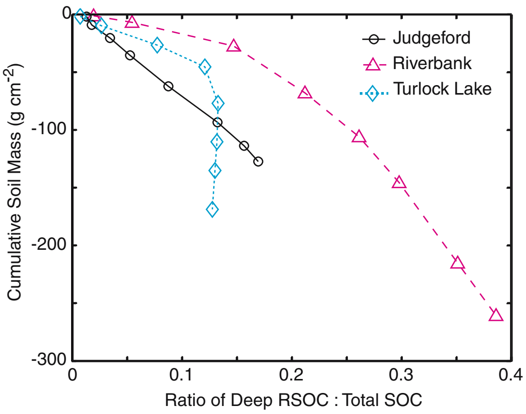

Fig. 3 Model derived estimates of the proportional contribution of deep RSOC to total SOC as a function of soil depth. The symbols show values calculated for each soil horizon. Cumulative soil mass is equivalent to soil depth in centimeters with bulk density normalized to 1 g cm^{-3}

water from the rooting zone and return it to the atmosphere via transpiration. Soil biological processes can also be expected to vary with soil depth as a result of non-linear effects of soil temperature and moisture (Metherell et al. 1993). It is therefore likely that a model of accounting for soil water transport, plant root dynamics and/or DOC dynamics (Neff and Asner 2001) would provide a better fit of the Judgeford soil's $\Delta^{14}C$ profile over time.

Given the simple empirical formulation of the deep RSOC in the model, examining uncertainty in parameter estimates provides a useful examination of the value of the model and assumptions in constraining the existence and size of the RSOC pool based on the data. The differences resulting from the recognition of deep RSOC in the model are best illustrated by comparing the parameter values obtained for the California soils to those in Baisden et al. (2002a). Most importantly, the rate of turnover for the passive SOM pool, k_3, decreased by a factor of two or more. This change is the logical result of recognizing the separation of a pool containing more recently fixed C (and ^{14}C from the atmosphere) below the rooting zone. However, other parameters indicated a greater tendency for SOM to enter deeper soil depths, through a doubling of downward transport rate of stabilized SOM, v_2, and a tripling of the characteristic depth of root inputs, L. Most other parameters showed little change resulting from the addition of the deep RSOC pool.

At present, the parameter values and pool sizes given in Tables 2 and 3 must be interpreted as potentially conservative estimates of the importance of the deep RSOC in the Judgeford soil. Despite this, the parameter values in Table 2 suggest that recognizing deep RSOC may shift estimates of conventional turnover parameters as defined in previous work (Baisden et al. 2002a). These changes can be seen as an advancement in understanding and reducing model selection uncertainty. This advancement is similar to the recognition of lag times in the entry of C derived from litter pools and other SOM pools into long lived or deep soil SOM pools (Baisden et al. 2002a; Gaudinski et al. 2000).

The model calculations suggest the deep RSOC represented 5–23% of total SOC within the sampled soil profiles, and 8–74% of RSOC with turnover rates of years to decades (Table 3). Below 50 cm depth, deep RSOC represented an even larger fraction of SOC—in the range of 10–40% (Fig. 3). Moreover, the lower end of these estimates probably underestimates the true size of the deep SOC because the model does not fully account for the deep bomb-^{14}C enrichment observed in the Judgeford soil (Fig. 1), and because the deep SOC extends below the soil depths sampled in this study. The divergence in the shape of the contributions of deep RSOC to total SOC in the Turlock Lake and Riverbank soils suggests considerable uncertainty in whether deep RSOC accounts for an increasing proportion of total SOC below the 100 cm soil depth typically sampled in studies of C cycling.

In support of our finding, 10 years after ^{14}C labeling of Colorado short-grass prairie, 17.5% of recovered ^{14}C tracer was found in SOC below 50 cm soil depth (Gill et al. 1999). Similarly, C and N mineralization consistent with deep decadal SOC has been observed to >7 m depth in prairie and agricultural soils (Ajwa et al. 1998) and may be consistent with both the downward translocation of soil C and N following disturbance (Ajwa et al. 1998) and the presence of biologically available N in sediments and weathered rock (Holloway and Dahlgren 1999). Identifying deep RSOC with turnover times of decades or less is most important for understanding the potential effects of land-use change on the C cycle, particularly where new land uses may accelerate loss or fail to replace deep RSOC. The recent use of $\delta^{13}C$ to track forest- and pasture-derived SOC indicates that tropical deep SOC responds to land-use change on decadal timescales (Veldkamp et al. 2003). Similarly, after accounting for soil compaction, SOC stocks to 1 m depth show significant changes between forest, pasture and agriculture, the combined use of δ^{13}C and SOC inventories identifies the failure of C4 crops to replace SOC turnover throughout soil profiles, including decreased stabilization of SOC below 30 cm depth (Osher et al. 2003).

Given the limited evidence discussed above, why have previous studies utilizing archived and contemporary samples spanning the bomb-^{14}C period (Paul et al. 1997; Trumbore 1993; Trumbore 1997) not clarified the significant enrichment of bomb-^{14}C below 30 cm depth? First, work has focused primarily on surface soils due to expectation that most human-, vegetation- or climate-induced changes in SOC will occur in the upper 30 cm of soil. Where Δ^{14}C measurements have been made to >50 cm soil depth, measurements are rarely made on both pre- and post-bomb profiles. In some cases involving grasslands, bulk Δ^{14}C values are reported only after pretreatment with acid to remove carbonates (Torn et al. 2002), and therefore also causing likely solubilization of deep RSOC. In forests, light density fractions have been reported to have modern Δ^{14}C values but in association with forest disturbance during the period between archived and contemporary sampling (e.g., Trumbore 1993). In the rare cases in which bulk Δ^{14}C values of both pre- and post-bomb profiles are reported, horizons are often >30 cm thick and depth intervals sampled are difficult to match between pre- and post-bomb profiles (Trumbore 1997; Wang et al. 1999). This difficulty largely results from greater variability in soil properties in colluvial upland soils sampled on elevational climate transects (Trumbore 1997; Wang et al. 1999) and on glacial till (Gaudinski et al. 2000), compared with alluvial and loess grassland soils. Similarly, due to differences in the size of individual plants and root architecture, forest soils are more variable than grassland soils at the scale of soil pits used to collect historic samples.

Another key reason for difficulty in recognising deep RSOC has been the limitations of methodologies. In particular, density fractionation remains the tool of choice, but suffers from several concerns. First, density fractionation methods vary in the density used to make the separation (Trumbore 1993, Gaudinski et al. 2000, Baisden et al. 2002b), and it is known that the level of shaking or sonication employed causes significant variation in the fraction recovered (Golchin et al. 1994). Furthermore, density fractionation in sodium polytungstate can fail to recover in excess of 10% of the total SOC in some samples, presumably as a result of dissolution (Baisden et al. 2002b, S.E. Crow, pers. comm.).

As a result, density fractionation can, when employed with the common practice of calculating light fraction C content and isotopic composition by difference using bulk values and measurements of the dense fraction, be misinterpreted as suggesting that the 'light fraction' is entirely particulate or non-mineral-associated SOC. Baisden et al. (2002a) used the Δ^{14}C values of dense fractions and bulk SOC at depth in Riverbank and Turlock Lake soils to conclude that the deep RSOC pool might be as much 24% of total SOC at depth. The study could not however differentiate between particulate SOC or dissolution of mineral-associated SOC. If dissolution is an important contributor to the ability of density fractionation to quantify deep RSOC, then more careful methods should be developed because heavy liquids may dissolve an arbitrary amount of SOC that is related neither to biological mechanisms of depolymerization and destabilization (Schimel and Bennett 2004), nor the definable pool size that can be derived from isotherm studies (Neff and Asner 2001). Based on these considerations, we suggest that the use of bulk Δ^{14}C values from time-series samples spanning the bomb-^{14}C era combined with mathematical separation of pools represents an important alternative to soil fractionation methods for estimating the size and dynamics of deep RSOC, as well as associated uncertainties.

Recognition of deep RSOC challenges studies of the terrestrial C cycle in grasslands, and perhaps other terrestrial ecosystems, by potentially doubling the size of SOC pools understood to react to disturbance, land use, management and climate change on timescales <100 years. This recognition is highly consistent with broad recognition that biological activity extends meters below the soil surface (Richter and Markewitz 1995). Given the apparent size of the deep RSOC pool, what is its likely source? The primary candidates are deep plant roots and dissolved organic matter (DOM) transport. Previously, significant C fluxes have been associated with plant roots in tropical forests to depths of 8 m (Nepstad et al.

1994; Trumbore et al. 1995). Plant roots were not, however, commonly observed at the depths where the deep bomb-^{14}C enrichment was observed in the California soils (Baisden et al. 2002a). Moreover, the deep bomb-^{14}C enrichment in the Judgeford soil (Fig. 1) increased between 1974 and 2002, a result that cannot be attributed to plant roots when atmospheric Δ^{14}C was decreasing.

We therefore hypothesize that the deep RSOC represents a large ~decadal SOC pool that is fed by DOM inputs deep in the soil profile, with a possible small contribution from deep plant roots. Improving our knowledge of the deep RSOC will be critical for improving models of SOC cycling to include SOC turnover and transport below 30 cm depth—an essential aspect of C accounting (Jackson et al. 2002; Jobbagy and Jackson 2000). Further research will require radiocarbon measurements in additional soils as well as the development of independent methods of estimating deep RSOC in soils with greater variation in Δ^{14}C values, including colluvial upland soils and forest soils. Improved understanding will result primarily from investigations of processes controlling the transport and fate of dissolved organic matter in soil—research that will not only extend terrestrial ecosystem C models beyond ~30 cm of surface soil, but will shed light on the recently recognized roles of DOM in nutrient cycling (Neff et al. 2003) and loss (Perakis and Hedin 2002).

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