Influence of soil texture on carbon dynamics and storage potential in tropical forest soils of Amazonia

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[1] Stable and radiocarbon isotopes were used to investigate the role of soil clay content in the storage and dynamics of soil carbon in tropical forest soils. Organic matter in clay-rich Oxisols and Ultisols contains at least two distinct components: (1) material with light $\delta^{13}C$ signatures and turnover times of decades or less; and (2) clay-associated, ^{13}C-enriched, carbon with turnover times of decades at the surface to millennia at depths >20 cm. Soil texture, in this case clay content, exerts a major control on the amount of slowly cycling carbon and therefore influences the storage and dynamics of carbon in tropical forest soils. Soils in primary tropical forest have been proposed as a potentially large sink for anthropogenic carbon. Comparison of carbon stocks in Oxisols sampled near Manaus, Brazil, shows no measurable change in organic carbon stocks over the past 20 years. Simple models estimating the response of soil carbon pools to a sustained 0.5% yr^{-1} increase in productivity result in C storage rates of 0.09 to 0.13 MgC ha^{-1} yr^{-1} in soil organic matter, with additional potential storage of 0.18 to 0.27 MgC ha^{-1} yr^{-1} in surface litter and roots. Most storage occurs in organic matter pools with turnover times less than a decade. Export of carbon in dissolved form from upland terra firme Oxisols likely accounts for <0.2 MgC ha^{-1} yr^{-1}, but more work is required to assess the export potential for periodically inundated Spodosols. *INDEX TERMS:* 0315 Atmospheric Composition and Structure: Biosphere/atmosphere interactions; 0330 Atmospheric Composition and Structure: Geochemical cycles; 0322 Atmospheric Composition and Structure: Constituent sources and sinks; 1040 Geochemistry: Isotopic composition/chemistry; *KEYWORDS:* soil carbon, ^{14}C, radiocarbon, tropical soil, soil texture, turnover, decomposition, carbon cycle

1. Introduction

[2] Tropical forests are a key biome controlling the global carbon balance [*Friedlingstein et al.*, 1995; *Melillo*, 1996; *Thompson et al.*, 1996; *McGuire et al.*, 1996]. These forests contain approximately 60% of the total carbon stored globally in forests [*Dixon et al.*, 1994], and 26% of the C stored globally in soil organic matter [*Batjes*, 1996]. Rates of gross primary production are similarly high, and the average residence time of carbon in tropical forests is predicted to be less than that of other biomes, like boreal forests [*Raich and Schlesinger*, 1992; *Thompson et al.*, 1996].

[3] Using eddy-covariance analysis, *Grace et al.* [1995] estimated that a tropical forest area in Rondônia State (southwest region of the Amazon) was accumulating carbon at a rate of approximately 1 Mg C ha^{-1} yr^{-1}. The same technique applied in a forest area near Manaus (central Amazon) found a net gain of carbon nearly 6 times larger than the Rondônia study [*Malhi et al.*, 1999]. The potential for large carbon gains by tropical forests is contested by others [e.g., *Keller et al.*, 1996]. For instance, *Chambers et al.* [2001], using a empirical forest dynamics model, demonstrated that aboveground biomass in Central Amazon forests is unlikely to have the capacity to sequester much more than about 0.5 Mg C ha^{-1} yr^{-1}. *Malhi and Grace* [2000] concluded that the extremely large carbon gains measured by eddy covariance methods are implausible, but still estimate a net sink of carbon in primary tropical forest soils.

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[4] If tropical forests are indeed acting as sinks of carbon from the atmosphere, the extra carbon stored has three possible fates: accumulation in biomass, accumulation in soil organic matter (SOM), and export in a form other than CO_2 (such as nonmethane hydrocarbon emissions, or in dissolved form in percolating soil solution). A summary of permanent forest inventory plots in the tropics suggests that some carbon may be accumulating in vegetation [*Phillips et al.*, 1998]. On the other hand, *Malhi et al.* [1999] and *Malhi and Grace* [2000] have hypothesized that the majority of the extra carbon may be accumulating in the SOM of tropical regions, with up to 3.5 Mg C ha^{-1} accumulating every year in tropical soils. This amount of carbon represents roughly 3% of the estimated soil carbon stock in the first meter for Oxisols like those in the Manaus eddy covariance site [*Chauvel*, 1982; *Cerri and Volkoff*, 1987], a rate of increase that should provide detectable differences in C inventory after about a decade. Hence tropical soils are hypothesized to play a large role in the global carbon balance.

[5] Despite its potential importance, detailed quantitative studies of C dynamics in tropical soils are limited, and mostly confined to investigation of changes in carbon stocks due to the replacement of forest by pasture. These studies have shown that carbon stocks in SOM are large but that only a portion of the total soil organic matter changes its ^{13}C signature in the first decades after a change from C3 to C4 vegetation.

[6] Two factors are key to estimating the potential for soil organic matter to behave as a source or sink of CO_2: (1) the size of the flux of C into and out to the soil pool (and the degree of which that has changed), and (2) the time C resides in soil organic matter. Regarding this second factor, it is critical to understand how soil texture and mineralogy influence the residence time of C in forest soils. By way of illustration, *Schimel et al.* [1994] showed that the average residence time for SOM in the upper 20 cm of soils predicted by the Century model for tropical regions ranged from <14 years up to several decades, depending on how they parameterized the dependence of soil C dynamics on soil texture. *Silver et al.* [2000] found that the Century model underestimated the amount of C in clay soils, either due to insufficient allocation of low-quality litter belowground in sands, or an overestimation of the degree to which C is stabilized on clays. Clearly, better understanding of how the dynamics of C is related to soil texture and mineralogy is important for improving ecosystem C cycle models.

[7] We use measurements of carbon stocks, C inputs and losses, and radiocarbon in soil organic matter and litter to quantify below-ground carbon dynamics in several soil types in Amazônia. Previous work on soil C dynamics has focused on Oxisols, which are the dominant soil type in terra firme forest. Here, we expand to other soil types, compare approaches for separating bulk soil carbon into labile and refractory components, and explore the importance of soil texture in soil carbon residence time. Using carbon stocks and residence times, we use a simple model to predict the capacity for each soil to act as a sink for carbon given an increase in plant litter inputs. Finally, using data from the literature, we evaluate the potential rate of export of carbon from terra firme soil in dissolved phases. Our overall conclusion is that the capacity for soils to act as sinks for atmospheric CO_2 on decadal timescales is far smaller than has been predicted from some eddy covariance studies, though it is large compared to temperate forest ecosystems and can potentially contribute to significant interannual variation in forest carbon balance.

2. Study Sites

[8] Data are presented from sites that are part of a larger study of above-and below-ground carbon stocks and dynamics in primary forests of Amazônia. Soils were analyzed along transects spanning gradients in elevation and soil texture at two locations: (1) the ZF2 INPA reserve approximately 80 km north of Manaus, Amazon state (hereinafter referred to as the Manaus site), and (2) at km 83 on the Satarém-Cuiabá highway, in the Floresta Nacional do Tapajós (FLONA), Pará state (hereafter Santarém site). Studies of soil properties, as well as aboveground carbon cycling are ongoing at both sites as part of the LBA-Ecology project [*Chambers et al.*, 2003; *Silver et al.*, 2000]. We also use results from a previous study at the Reserva Ducke, located ∼20 km south of the ZF2 site near Manaus [*Trumbore*, 1993] and new results from a site studied previously for radiocarbon and carbon stocks near Paragominas, in eastern Pará state [*Camargo et al.*, 1999; *Trumbore et al.*, 1995] to show how ^{14}C changes with time in tropical forest Oxisols.

[9] The study sites have similar vegetation and climate (Table 1), though there is less marked seasonality of rainfall in Manaus than in Santarém and Paragominas. Other carbon cycle related characteristics, such as litterfall and soil respiration flux, are similar across the sites (Table 1).

[10] For analysis of spatial variability in carbon and nutrient stocks, 16 soils pits were dug at locations representing local soil textural end-members along each gradient (Table 1). The Manaus ZF2 site is characterized by plateaus dissected by small streams with local relief of ∼40 m. Soils were collected on the plateau, where Oxisols, rich in kaolinite clay are the dominant soil type (Manaus Oxisol, see Table 1). In addition, soil samples were also taken in lowland sites (locally called baixios), where soils are intermittently inundated and are made up of very clean sands mixed with organic matter (referred to as Manaus Spodosol). The FLONA Tapajós site similarly has plateaus with more clay rich soils (Santarém Oxisol), and areas on slopes or in topographic lows that contain more sand (Santarém Ultisol), though these are well-drained and lack the periodic inundation of the Manaus lowland soils. At the Paragominas site (Paragominas Oxisol), only clay-rich soils on plateaus were analyzed, and the results presented are for a single site only.

[11] Previously published radiocarbon data are available for an Oxisol sampled in 1986 at the Reserva Ducke, located ∼20 km south of the ZF2 transects [*Trumbore*, 1993]. The soils at Reserva Ducke are very similar to those of the Manaus Oxisol site reported here and the change in

Table 1. Site Characteristics

Soil	MAP, mm	MAT, °C	Clay,[a] %	Litterfall	Aboveground Biomass, Mg C ha^{-1} yr^{-1}	Annual CO_2 Flux
Paragominas Oxisol	1803 (555)[b,c]	-	70−90[c]	4.5[d]	132[d]	20[d]
Manaus Oxisol	2096[e]	26.7	60−80[f]	4.4[g]	217[g]	12[h]
Manaus Spodosol	2096[e]	26.7	<10[f]	3.3	-	8
Santarém Oxisol	2101[e]	25[i]	60−60[j]	3.10 (0.47)[b,j]	176[k]	15
Santarém Ultisol	2101[e]	25[i]	14−22[j]	4.39 (0.47)[b,j]	-	-

[a]Superficial and 1-m soil depth.
[b]One standard deviation.
[c]*Jipp et al.* [1998].
[d]*Camargo et al.* [1999].
[e]*Viera and Santos* [1987].
[f]*Ferraz et al.* [1998].
[g]*Higuchi et al.* [1997].
[h]*Chambers et al.* [2003].
[i]*Parrota et al.* [1995].
[j]*Silver et al.* [2000].
[k]*Sorrensen* [2000].

the amount of ^{14}C between 1986 and 2000 is an extremely useful indicator of C dynamics in these Oxisols.

3. Methods

[12] Sixteen soil profiles in each of the four soil types were sampled to 50 cm depth for analysis of bulk density and C stocks. Three pits for each soil were sampled to 2 m depth. Following analysis of C stocks and determining that the degree of spatial variation of root-free soil within each site was small [*Telles*, 2002] (Table 2), samples from one representative 2-m deep pit were analyzed for radiocarbon and stable isotopes. As mentioned previously, the data for both C stocks and radiocarbon from the Paragominas site are from a single pit.

3.1. Sample Preparation

[13] Soil samples from Manaus and Santarém sites were sieved to <2 mm, then homogenized and split into a smaller subsample that was handpicked to remove fine roots and charcoal. Soils were then ground for elemental and isotopic analyses.

[14] Samples from Paragominas (and those cited in the literature used here) were similarly sieved to <2 mm and homogenized. However, separation of roots and charcoal in those studies was accomplished by density separation, with roots and plant fragments removed by flotation in a liquid of density between 1 and 2 g cm^{-3}. Past comparisons have shown the Δ^{14}C of bulk soil treated in this manner to be the same as that from which low-density material has been removed by careful hand-picking. [*Trumbore and Zheng*, 1996].

[15] At the Paragominas and Manaus Reserva Ducke sites, dense organic matter was further fractionated by extraction with acids and bases [*Trumbore et al.*, 1995]. The Δ^{14}C of solubilized carbon (hydrolyzable fraction) is calculated as the difference between unextracted and residue (unhydrolyzable material). Dense fraction organic matter from Manaus Reserva Ducke soils was hydrolyzed only with 6N HCl, which has been shown to remove less ^{14}C than acid + base hydrolysis [*Trumbore and Zheng*, 1996].

[16] Gravimetric carbon content and carbon stable isotopic composition, expressed as δ^{13}C, were measured using an elemental analyzer coupled to a continuous flow stable isotope ratio mass spectrometer at the Centro de Energia Nuclear na Agricultura (CENA), Piracicaba, Brazil. Samples for radiocarbon analysis were sealed in evacuated

Table 2. Carbon Stocks (Mg/ha) of Roots, Mineral-Organic Matter and Total Soil Organic Matter Among Sites[a]

Depth, cm	Mineral-Organic Matter		Root		Total Carbon	
Oxisol Paragominas						
0−10	22		6		28	
10−40	32		2		34	
40−100	32		?			
Total	86					
Oxisol Manaus						
0−10	19.2	(2.9)	8.8	(3.6)	28.0	(4.9)
10−40	30.4	(2.7)	3.0	(1.7)	33.4	(3.2)
40−100	29.5	(2.0)	2.0	(0.8)	31.6	(2.2)
100−150	16.2	(0.8)	0.9	(0.6)	17.1	(1.1)
150−210	14.7	(0.6)	0.3	(0.2)	15.0	(0.6)
Total	110.0	(4.6)	15.0	(4.1)	125.0	(6.3)
Spodosol Manaus						
0−10	12.5	(3.9)	8.1	(3.4)	26.9	(5.5)
10−40	18.0	(4.7)	1.3	(0.2)	19.3	(4.7)
Total	30.5	(6.1)	9.3	(3.4)	46.2	(7.2)
Oxisol Santarém						
0−10	24.6	(4.5)	7.0	(3.2)	31.7	(5.7)
10−40	39.1	(4.0)	0.9	(0.4)	40.0	(4.1)
40−100	38.6	(3.1)	1.3	(0.7)	39.9	(3.2)
100−150	20.7	(1.6)	0.2	(0.1)	20.9	(1.6)
150−210	19.0	(1.3)	0.0	(0.0)	19.0	(1.3)
Total	142.1	(7.1)	9.4	(3.3)	151.5	(8.0)
Ultisol Santarém						
0−10	8.7	(2.1)	3.9	(1.7)	12.5	(2.7)
10−40	27.7	(1.8)	1.5	(0.9)	29.3	(3.7)
40−100	34.8	(2.2)	2.1	(0.9)	36.9	(4.5)
100−150	13.9	(2.0)	0.3	(0.1)	14.2	(4.0)
150−210	16.1	(1.5)	0.1	(0.0)	16.2	(3.0)
Total	101.2	(4.3)	7.9	(2.1)	109.1	(8.1)

[a]Numbers in parentheses are standard deviations.

quartz tubes with cupric oxide wire and combusted at 900°C for 2 hours. Evolved CO_2 was purified cryogenically, and then reduced to graphite by the zinc reduction method [*Vogel*, 1992] for analysis by accelerator mass spectrometry at the Center for AMS at the Lawrence Livermore National Laboratory, Livermore, California. The analytical uncertainty associated with radiocarbon analyses on our graphite analyzed at LLNL is ±5‰, based on repeated determinations of secondary standards.

3.2. Derivation of Turnover Time From Radiocarbon

[17] Radiocarbon is produced at a relatively constant rate in the stratosphere by cosmic ray bombardment, and is subsequently oxidized to $^{14}CO_2$, which mixes through the atmosphere and biosphere. The decrease in ^{14}C activity in C derived from dead plant tissues is a measure of the residence time of carbon, for residence times longer than a few hundred years [*Trumbore*, 2000]. Radiocarbon was also produced by nuclear weapons testing in the atmosphere, which approximately doubled the background amount of ^{14}C in the early 1960s. Incorporation of this ''bomb'' ^{14}C into soil organic matter is a valuable tracer of C exchange with the atmosphere over the past few decades [*Trumbore*, 2000].

[18] Measurement and modeling of ^{14}C in soil organic matter pools are discussed in detail elsewhere [*Trumbore*, 2000; *Gaudinski et al.*, 2000]. We report radiocarbon data in $\Delta^{14}C$ (‰) notation, where

$$\Delta^{14}C = 1000 * \left(R_{sample}/(0.95 * R_{standard}) - 1\right), \qquad (1)$$

and R_{sample} is the $^{14}C/^{12}C$ ratio of the sample (corrected to $\delta^{13}C$ of −25‰), $R_{standard}$ is the $^{14}C/^{12}C$ ratio of the oxalic acid standard, decay corrected to 1950, with $\delta^{13}C$ of −19‰. The factor $0.95 * R_{standard}$ equals the $^{14}C/^{12}C$ ratio of CO_2 in air in 1950 (also known as Modern). The reporting of ^{14}C data corrected to a common ^{13}C value removes any mass-dependent isotope fractionation effects. Using $\Delta^{14}C$ notation, positive values mean greater ^{14}C activity in the sample than in carbon fixed during 1950, indicating the influence of ^{14}C derived from atmospheric weapons testing. Negative $\Delta^{14}C$ values indicate that the average residence time of C is long enough for significant radioactive decay to have occurred.

[19] The rate of incorporation of bomb ^{14}C into the soil C reservoir provides a useful tool for deciphering C turnover in soils on timescales of decades to hundreds of years. For a given C pool assumed to be at steady state, with annual rate of C inputs (I), equal to C losses assuming first order decomposition (k * C(t), where k is the decay constant for decomposition and C(t) the amount of carbon), $dC(t)/dt = I - k * C(t)$. Assuming the radiocarbon signature of C inputs to the soil organic matter pool or fraction reflect that of the current atmosphere (R_{atm}, where $R_{atm} = \Delta^{14}C_{atm}/1000 + 1$), the amount of C in at time t (C(t)) and its $^{14}C/^{12}C$ ratio relative to the standard ($R(t) = \Delta^{14}C(t)/1000 + 1$)) can be estimated from that of the previous time step (C(t − 1); R(t − 1)) using a model with 1-year time step,

$$C(t)R(t) = [C(t-1)R(t-1)(1 - k - \lambda) + (I)R_{atm}(t)], \qquad (2)$$

where λ is the decay constant for radioactive decay of ^{14}C (0.00012584 yr^{-1}). Dividing by C(t) and, given the steady state assumptions, C(t) = C(t − 1), and I = kC, this reduces to

$$R(t) = [R(t-1)(1 - k - \lambda) + (k)R_{atm}(t)]. \qquad (3)$$

Using a yearly time step and values of R_{atm} for the years of interest, and given the constraints of measured $\Delta^{14}C$ values at two points in time since 1964, one can determine values of k that correspond to measured ^{14}C values (for each isolated C pool).

[20] Data for $\Delta^{14}C$ of CO_2 in the tropical atmosphere are available from *Levin et al.* [2000] from 1900 to 1996. For years since 1996, we have averaged $\Delta^{14}C$ values measured in our laboratory at a variety of different field sites. We have removed air samples with $\delta^{13}C < -9.5‰$ from this data set, assuming that the samples (taken roughly 0.5 m above the soil surface) are affected by re-respired carbon. We added 3‰ to these values because they were predominantly made in the Northern Hemisphere, and *Levin et al.* [2000] show an increase of ∼3‰ between northern hemisphere and tropical sites for the past decade. The values we use in the model for tropical soil C dynamics are: 1996: 100.1 ± 7‰; 1997: 90.9 ± 7‰; 1998: 91.9 ± 9‰; 1999: 84.0 ± 9‰; 2000: 83.3 ± 9‰.

[21] The model as described above assumes that C entering the soil pool has the $\Delta^{14}C$ of C fixed from the atmosphere in that year. Hence values of k derived from the model reflect the overall residence time of C in the living plant and soil [*Trumbore*, 2000]. Significant lag times are found in plant tissues such as roots [*Gaudinski et al.*, 2001], which means the k values derived using bomb ^{14}C models do not directly correspond to decomposition rate constants for soil organic matter. We have modified the model to account for the time lag between photosynthesis and addition of dead tissues to soil organic matter pools and using the $\Delta^{14}C$ of leaf and root litter to estimate the length of time C resides in live vegetation pools (see results, below).

4. Results

4.1. Soil Carbon Stocks

[22] For the three soils sampled 100 cm depth, C stocks ranged from 102 to 151 MgC ha^{-1}, with 2–12% of the total C in fine roots (Table 2). To 40 cm depth, the largest C stocks were found in the two Oxisols (61–71 MgC ha^{-1}), with significantly lower C stocks in the soils with higher sand content (46 MgC ha^{-1} Spodosol and 42 MgC ha^{-1} Ultisol). Total soil carbon stocks in the upper 10 cm of soil were similar across sites, with 26–32 MgC ha^{-1}, except for the Santarém Ultisol, which had only 12 MgC ha^{-1} (Table 2). Our results for the Santarém sites are in accord with those of *Silver et al.* [2000], whose sites were nearly colocated with ours.

[23] About a third of the total carbon to 10 cm depth in each soil was in fine (<2 mm) roots. However, the non-root portion of soil carbon also included organic matter too fine to pick by hand in addition to mineral-associated organic matter. In the Manaus Spodosol, this fraction accounted for

virtually all of the non-root carbon in the soil, as shown in subsequent density separations [Telles, 2002].

[24] Gravimetric C contents (Figure 1) show the same general patterns as C stocks. The %C in sieved (<2 mm) soil from which roots have been removed decreases rapidly from 2 to 3%C at the surface to values <0.8%C below 50 cm depth. Analytical uncertainties increase in importance at very low values of %C (<0.4%C), which dominate the calculation of uncertainty for the entire soil profile (Table 2).

4.2. Radiocarbon

[25] Radiocarbon measured in bulk soil organic matter (<2 mm, roots removed) for the Oxisols and Ultisol decrease rapidly from positive values (indicating the predominance of carbon fixed since the 1960s) near the surface to large negative values (indicating that radiocarbon has resided in the soil long enough for significant radioactive decay to occur) below 1 m depth (Figure 2). Steepest rates of decline occur within the upper 20 cm of soil; ^{14}C values continue to decline but at a slower rate below 1 m depth. Values of D^{14}C below 1 m depth vary between sites, with most negative values occurring in the soils with highest clay content and less negative values in soils with highest sand content (Santarém Ultisol).

[26] The Manaus Spodosol differs markedly from the other soils in Figure 2. D^{14}C values in this soil are positive and demonstrate no trend to the maximum depth sampled (50 cm), indicating relatively recent origin of the organic matter sampled in this soil.

[27] D^{14}C values for surface litter (points at 0 cm depth on Figure 2) show the presence of bomb ^{14}C in all soils, and are 20–30% more enriched in ^{14}C than the atmospheric CO_2 in the year of collection (2000; +81%). Values are slightly higher in Santarém than in Manaus. Near-surface live fine roots from these sites also have ^{14}C values significantly higher (by +20 to +90%, depending on size and depth) than contemporary atmospheric values (unpublished data).

4.3. Stable Carbon Isotopes

[28] Stable carbon isotope data give information about the sources of organic matter found in soils, as well as the

Figure 2. Radiocarbon in organic matter, reported as D^{14}C (%) from bulk soil after it has been sieved to <2 mm and hand-picked to remove root detritus and charcoal.

degree of alteration by microbial degradation. For the Oxisols at all three sites and the Santarém Ultisol, ^{13}C values are most negative (most depleted in ^{13}C) in the near-surface (0–10 cm), and show enrichment in ^{13}C with depth to values ranging from 25.5 to 26.5% (Figure 3). Again, the Manaus Spodosol differs markedly from the other soils, with ^{13}C values (between 29.5 and 30%), close to those of fine roots (29 to 30%) and surface plant litter (30%). Changes in ^{13}C and ^{14}C are related, with heavier ^{13}C values associated with more ^{14}C-depleted organic matter, as found in other tropical forest soils with no evidence of past vegetation change [Saniotti et al., 2002].

4.4. Turnover Times Estimated From Radiocarbon Model

4.4.1. Surface Litter and Litterfall

[29] Comparing the D^{14}C values of surface litter with the history of ^{14}C in the atmosphere, and using equation (3), we would estimate a turnover time for this pool of 3–4 years (see Figure 4). This is similar to the 3-year turnover time reported by Trumbore et al. [1995] for primary forest surface litter in Paragominas, Para (sampled 1992) and also matches the value observed by Trumbore [1993] for surface litter in Reserva Ducke. Leaf litter in tropical forests decomposes rapidly [Gholtz et al., 2000], with the majority
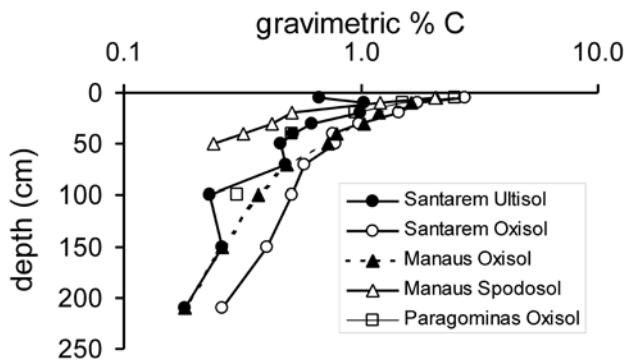

Figure 1. Gravimetric percent carbon in <2 mm soil that has been picked for roots. Roots can make up to approximately one third of the total carbon in the upper 10 cm of soil.
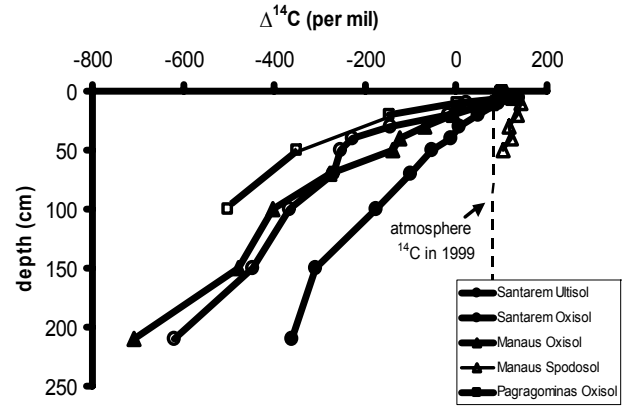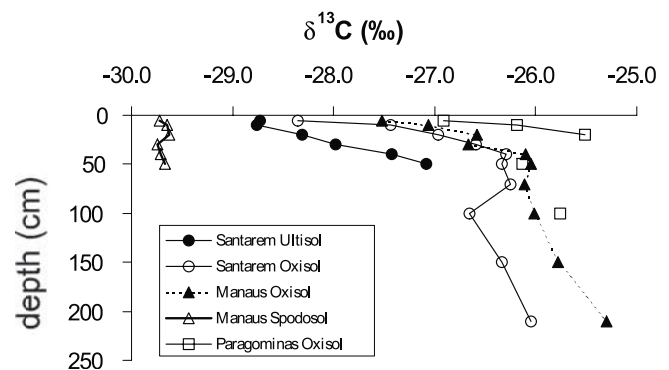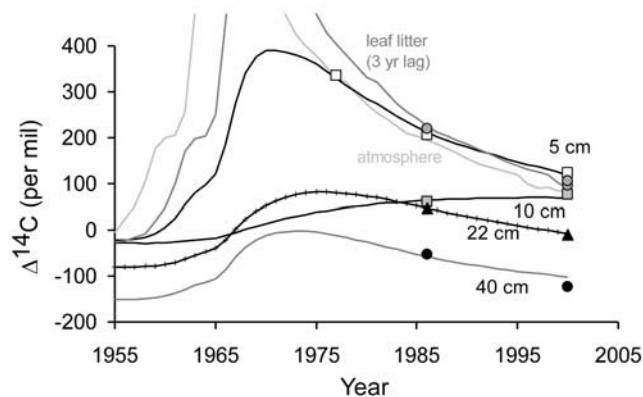

Figure 3. Stable carbon isotopes (reported as % deviation from the PDB standard) for soil organic matter.

Figure 4. Radiocarbon data from three sets of samples of similar Oxisols sampled near Manaus. The upper contours are for litter and 0–5 cm of mineral soil; the lower contours show deeper mineral soil depths. Circles show surface litter, which consistently lags the atmospheric ^{14}C curve by $\sim$3 years. Diamonds show data from *Trumbore* [1993] for mineral-associated organic matter. The 1977 point is humic material extracted from the soil at Reserva Ducke and reported by *Trumbore* [1993]. Best fit lines for the radiocarbon time history for each depth interval are shown; each have inputs lagged by 3 years to account for the residence time of carbon in living plant tissues. These curves represent the following combinations of ''active'' and ''passive'' organic matter pools: 0–5 cm, 65% 3 yr + 35% 60 yr; 10 cm, 70% 70 yr + 30% 500 yr; 22 cm, 40% 8 yr + 65% 1000 yr; 40 cm, 28% 5 yr + 72% 2000 yr.

of mass disappearing within <1.5 years. Hence, high ^{14}C values in surface litter must reflect the residence time of litter on trees prior to addition to the soil surface. We therefore use a 3-year lag for litter inputs to the soil for modeling non-litter soil organic matter components.

4.4.2. Root Inputs

[30] The Δ^{14}C of fine (0.25 to 1 mm) roots picked from the soil correspond to residence times of <1 to 8 years. These values are similar to those reported for temperate forest soils [*Gaudinski et al.*, 2001], and, as for leaf litter ^{14}C values, are a measure of the time elapsed since the carbon in root structural tissues was fixed from the atmosphere. We are currently investigating the source of high root ^{14}C values using root screen methods similar to those discussed by *Gaudinski et al.* [2001] and will discuss implications of fine root ^{14}C data elsewhere.

4.4.3. Soil Organic Matter

[31] The best constrained turnover times from radiocarbon measurements come from comparison of the Manaus Oxisol profiles from Reserva Ducke (collected 1986 [*Trumbore*, 1993]) and ZF2 sites (Figure 4). In all but one case (the 5–10 cm depth interval), the amount of radiocarbon in the soil organic matter for a given depth interval decreased between 1986 and 2000. Using equation (3), components with relatively rapid (<8 years) turnover times show a decrease in Δ^{14}C values over the past 15 years, while those with longer turnover times show an increase in Δ^{14}C. A portion of the organic matter in these soils therefore must have turnover

times of less than a decade. The Δ^{14}C values of the total organic matter at depths >10 cm fall well below the record of Δ^{14}C in the atmosphere, indicating the presence of a second pool of organic matter with much slower turnover times that is diluting the more rapidly cycling component.

[32] Figure 4 shows best fit curves for 1986 and 2000 Δ^{14}C data from the Manaus Oxisols. A minimum of two components is required to produce a curve that fits observations made in 1986 and 2000. For the 0–5 cm depth interval, the best fit is achieved with 65% of the SOM having a turnover time of 3 years (assuming a 3-year lag to soil C inputs), and 35% with a turnover time of $\sim$60 years. For 22 and 40 cm depths, $\sim$30% of the organic matter is in a pool with turnover times of 5 to 8 years, and $\sim$70% in a pool with turnover times of 1000–2000 years. The 10-cm depth interval differs from the others in that Δ^{14}C values increased between 1986 and 2000. For this interval, the Δ^{14}C data are consistent with most ($\sim$70%) of the organic matter having turnover times of 70 years, and $\sim$30% of the organic matter with a turnover time of 500 years.

[33] For the Manaus baixio soil, we have no past soil sample for comparison. All Δ^{14}C values are positive, so the majority of this carbon has been fixed from the atmosphere since the onset of atmospheric weapons testing. Further, the organic matter is of one type, as evidenced by consistently negative δ^{13}C values, and the fact that nearly all carbon not hand-picked from the soil floated in a liquid of density close to 2.0 g cm^{-3}. We therefore treat the organic matter in the Spodosol as a single, homogeneous pool. Assuming 3-year time lags before C is added to the soil as litter, the Δ^{14}C values of +102‰ to +143‰ are reproduced by pools with turnover times ranging from either 3–6 years or 50–60 years. We may determine which of the two is more likely to be correct by comparing the fluxes we calculate by dividing the inventory of soil organic matter by the ^{14}C-derived turnover time with CO_2 fluxes measured in the sites [*Chambers et al.*, 2003]. For example, if all of the 30 MgC ha^{-1} of organic carbon in the Manaus Spodosol to 40 cm depth (excluding roots) has turnover times of 3–6 years, the annual rate of C addition to and loss from this soil must be 5 to 10 MgC ha^{-1} yr^{-1}. The higher range of this estimate exceeds the estimated NPP for this ecosystem of $\sim$8 MgC ha^{-1} yr^{-1}; *Chambers et al.* [2003]. For turnover time of 50–60 years, the annual flux would be only 0.5–0.6 MgC ha^{-1} yr^{-1}. Roots at this site have a stock of $\sim$9 MgC ha^{-1}; if they are decomposing rapidly, they may be responsible for much of the heterotrophic CO_2 flux from these soils [*Silver et al.*, 2000]. Hence it is likely that the longer turnover times are appropriate. The calculations above assume steady state conditions, which may not apply in this occasionally inundated system. For estimating potential response, we will use the faster turnover times for the upper 10 cm of soil, and the longer turnover times for 10–40 cm. This likely overestimates the response of these soils to changes in C inputs with time.

[34] Table 3 summarizes the estimated turnover times for carbon in Manaus soils. The important point here is that the majority of soil carbon to 1 m depth must be in ''passive'' pools with turnover times of many decades to several thousands of years in the Oxisol, while most of the carbon

Table 3. Turnover Times of Organic Matter Used in Estimating Soil C Storage Potential[a]

	Active	Slow	Passive	Total
Manaus Oxisol 1 (From Physical and Chemical Fractionation)				
C stock, Mg ha^{-1}	19.2	10.2	20.9	50.3
Turnover time, yr	8–10	70.0	500–1500	405
Estimated annual flux, Mg C ha^{-1} yr^{-1}	2.9	0.1	0.0	3.0
Storage rate after 500 years of NPP Increase, Mg C ha^{-1} yr^{-1}	0.3	0.2	0.1*	0.6
Manaus Oxisol 2 (From ^{13}C)				
C stock, Mg ha^{-1}	16.6	5.4	28.3	50.3
Turnover time, yr	5.0	82.0	180–1500	710.0
Estimated annual flux, Mg C ha^{-1} yr^{-1}	3.3	0.1	0.0	3.4
Storage rate after 500 years of NPP Increase, Mg C ha^{-1} yr^{-1}	0.3	0.1	0.2*	0.6
Manaus Spodosol				
C stock, Mg ha^{-1}	18.1	20.6	0.0	38.7
Turnover time, yr	5.0	60–70	N/A	31.0
Estimated annual flux, Mg C ha^{-1} yr^{-1}	3.6	0.3	N/A	3.9
Storage rate after 500 years of NPP Increase, Mg C ha^{-1} yr^{-1}	0.4	0.4	N/A	0.8
Single Pool Soil [Malhi et al., 1999]				
C stock, Mg ha^{-1}				50.0
Turnover time, yr				16.0
Estimated annual flux, Mg C ha^{-1} yr^{-1}				3.1
Storage rate after 500 years of NPP Increase, Mg C ha^{-1} yr^{-1}				1.0

[a]Active pools are those with turnover times <10 years in all soil layers (including low-density organic matter and organic matter with δ^{13}C of −29.7 per mil); slow pools are those with turnover times of decades (and δ^{13}C of −25.3 per mil); and passive pools are those with turnover times of centuries to millennia (with δ^{13}C of −25.3 per mil). Values are less than those expected if the time is extended beyond 500 years due to the slow response time of these pools.

must have turnover times of decades or shorter in the Spodosol.

[35] For the Santarém soils, we have no archived soils to determine how Δ^{14}C values have changed with time. We will argue below that the dynamics of these soils may be estimated from the others and from relationships between soil texture and the components of soil organic matter.

4.4.4. Physical and Chemical Fractionation of Soil Organic Matter

[36] Results of physical and chemical fractionation of organic matter from the Paragominas soil are very similar to those obtained previously ([*Trumbore et al.*, 1995; *Camargo et al.*, 2000] from soils collected in 1992). Low-density organic matter and surface litter have high carbon contents, more negative δ^{13}C values, and high Δ^{14}C values compared to organic matter associated with dense mineral surfaces (Table 4). The majority of soil carbon is in the dense fraction, though in surface layers, nearly 40% is in low-density forms. Removal of organic matter associated with components that are hydrolyzed in acid and base (including sesquioxide coatings) leaves a residue that represents ~20% of the original soil carbon (or about a third of the carbon associated with minerals) that is depleted in both ^{13}C and ^{14}C. Radiocarbon values in this residue show

evidence of some bomb ^{14}C in the uppermost mineral soil layer, and Δ^{14}C values decline steeply with depth.

[37] Comparison of Δ^{14}C values for the dense (mineral-associated) fraction of C for soils sampled in 1992 [*Trumbore et al.*, 1995; *Camargo et al.*, 2000] and 1996 (this work), show a decline in the Δ^{14}C of dense fraction organic matter in the 0–10 cm layer, little to no change in the 20–25 cm and 50 cm samples, and a larger difference at 100 cm. The decline at 0–10 cm is consistent with ~2/3 of the carbon having turnover times of 10–20 years, and ~1/3 having turnover times of ~100–200 years. In 1992, we sampled only 0–10 cm, which may be too coarse a resolution in this zone where Δ^{14}C and C content change rapidly with depth.

5. Discussion
5.1. Separating Soil Organic Matter Into Refractory and Labile Components

[38] Often ecosystem models treat soil organic matter as a single, homogeneous pool, for which the turnover time is determined by dividing the standing stock of carbon by the

Table 4. Physical and Chemical Fractionation of Forest Soils From Paragominas, 1996

	Litter	0–5 cm	10 cm	20 cm	40 cm	100 cm
Distribution of Carbon in Fractions (as Percent of Total Carbon in the Fraction)						
Light	100	37	40	32	22	15
Dense	0	63	60	68	78	85
Residue	0	20	15	19	22	32
Hydrolyzable (calculated)		43	45	49	56	53
Carbon Content of Fractions, Weight %						
Bulk	29.8	2.46	1.45	0.95	0.51	0.3
Dense		2.11	1.19	0.89	0.54	0.34
Residue		0.91	0.42	0.35	0.21	0.17
δ^{13}C, ‰ PDB						
Light	−28.2					
Dense		−26.91	−26.12	−25.51	−26.13	−25.76
Residue		−28.78	−27.99	−27.69	−25.98	−25.72
Hydrolyzable (calculated)		−26	−25.6	−24.6	−26.2	−25.8
Δ^{14}C, ‰						
Light	160					
Dense		113.2	1	−146.2	−351.8	−501.1
Residue		15.2	−190.5	−369.3	−634.3	−749.1
Hydrolyzable (calculated)		159	67	−57	−238	−350
Turnover Time (Assuming Homogeneous Within Each Pool)						
Light	6					
Dense		71	300	1400	4500	8300
Residue		240	2000	4840	14300	24700
Hydrolyzable		6[a]	120	650	2580	4450
From ^{13}C calculation "passive"		80		1500	7000	11000
From change in Δ^{14}C between 1992 and 1996			200	2060[b]	4250[b]	17500

[a]Two turnover times (6 years and 45 years) can give the value of +159 in 1996; we choose the shorter one because it fits better with the observed decline in δ^{14}C between 1992 and 1996 (see text).
[b]Based on Δ^{14}C data for dense fraction soils in 1992, depths are 25 cm and 50 cm.

rate of loss by heterotrophic decomposition [e.g. *Raich and Schlesinger*, 1992; *Malhi et al.*, 1999]. *Malhi et al.* [1999] use 16 years as the average turnover time for 162 MgC ha^{-1} to 1 m depth for soil at a site in the ZF2 reserve near our Manaus Oxisol. Physical and chemical fractionation of soil organic matter (Table 4) and comparison of $\Delta^{14}C$ values in soil organic matter collected at different times in the past (Figure 4) clearly demonstrate that this single pool model is not applicable for predictions of turnover on timescales shorter than centuries. Not only our radiocarbon data for these soils, but many studies of carbon dynamics in forests that are converted to pasture reported in the literature [e.g., *Cerri et al.*, 1991; *Davidson and Ackerman*, 1993], show that the use of a single turnover time for all soil organic matter to 1 m depth is unrealistic and will over predict the capacity of soils to store carbon over the next century [*Schimel et al.*, 1994; *Townsend et al.*, 1995; *Gaudinski et al.*, 2000].

[39] To predict carbon storage changes on decadal time-scales, a minimum of three components of organic matter are required: (1) leaf and root detritus; (2) "active" components of SOM with turnover times of decades or less, and (3) "passive" components of organic matter with turnover times of many decades to millennia. The "active" component in our soils has low $\delta^{13}C$ values and high $\Delta^{14}C$ values. It includes low-density organic matter too small to be removed by hand-picking (but removable by density separation), and material weakly associated with soil minerals and soluble in the density separation solution. "Passive" components with longer turnover times include those associated with soil minerals (density >2.0 g/cm^3), characterized by lower $\Delta^{14}C$ content and higher $\delta^{13}C$ values. In the upper part of the soil, this material has turnover times of decades. Below 5 cm depth the turnover time of this material is longer than a century, and it makes up the majority of the organic carbon in the soil. Although it can itself be separated into components with longer and shorter turnover times by acid and base hydrolysis (Table 4) the differences are not significant for predicting changes in C stocks over the next century except in the 0−5 cm depth interval.

[40] For soils where we did not perform physical and chemical fractionation of organic matter, we used $\delta^{13}C$ data to separate organic matter into two components, one with $\delta^{13}C$ values consistent with fresh low-density organic matter ($\delta^{13}C = -29.7‰$; the "active," or "A," fraction), and one with $\delta^{13}C$ equal to the heaviest values observed at 1 m depth in Oxisols ($\delta^{13}C = -25.3‰$; the "passive," or "P," fraction). For each depth interval, we calculate how much of the organic matter is in each fraction from the mass balance of ^{13}C,

$$P * \delta P + (1 - P) * \delta A = \delta T; \text{ or } P = (\delta T - \delta A)/(\delta P - \delta A), \quad (4)$$

where P is the fraction of total organic matter in the passive fraction, A is the active fraction, δT is the observed $\delta^{13}C$ of bulk mineral soil organic matter (Figure 3), and δA and δP are the two end-member ^{13}C signatures (−29.7 and −25.3‰, respectively). Values of P, the amount of "passive" soil carbon (Figure 5a) generally correlate with the clay contents for soils (see Table 1; Figure 6).

[41] The $\Delta^{14}C$ value for end-member A may be estimated from the $\Delta^{14}C$ of fresh organic matter (which ranges from +110‰ for surface litter up to +200‰ for roots), or from the organic matter in the Manaus Spodosol (+140‰). The $\Delta^{14}C$ of end-member P (Figure 5b) is then calculated from the ^{14}C mass balance.

$$R_P = (R_T - (1 - P) * R_A)/P, \quad (5)$$

where R_A equals ($\Delta^{14}C$ of fraction A)/1000 + 1, and R_P and R_T are the analogous ratios for fractions P and T. The $\Delta^{14}C$ of fraction P is then ($R_P - 1$) * 1000. A turnover time (1/k) derived from the $\Delta^{14}C$ value of fraction P is calculated using either equation (3) if it contains bomb carbon (i.e., $\Delta^{14}C$ is positive), or by solving the relation $R_P = k/(k + \lambda)$ for k, where λ is the radioactive decay constant for ^{14}C (see *Trumbore* [2000] for details).

[42] Turnover times calculated for the passive fraction P (Figure 5c), using ^{13}C values to partition soil organic matter into labile and refractory components, compare well with values obtained by physical and chemical fractionation of the soil obtained for the Paragominas Oxisol in Table 4, and for the Manaus Oxisol (Table 3). Given the expense of measuring ^{14}C in a large number of fractionated samples, using $\delta^{13}C$ to separate organic matter into active and passive components appears to be a useful alternative for tropical forest Oxisols.

5.2. Relationship to Clay Content

[43] Stabilization on clay surfaces has long been recognized as a mechanism for storing organic matter in soils on long timescales. However, clay content alone is not a good predictor of the total amount of organic matter; other factors, such as clay mineralogy [*Torn et al.*, 1997], or the abundance of phyrophosphate extractable iron [*Percival et al.*, 2000] may dominate when comparing very different soil types. In our study, all soils are made up of mixtures of kaolinite clay, sesquioxides, and quartz, so that changes in organic matter properties that correlate with soil texture will also correlate with other factors like pyrophosphate extractable iron.

[44] Carbon inventories are higher in the Oxisols, which have higher clay content and higher fractions of clay-associated organic matter (fraction P). The amount of "passive" carbon calculated using $\delta^{13}C$ to fractionate C correlates well with overall clay content in these soils (Figure 6).

[45] The residence time of fraction P, the more recalcitrant and ^{13}C-enriched C, increases rapidly with depth in all soils (Figure 5c). The controls of C on century to millennial timescales in these soils are linked to processes like formation and dissolution of secondary mineral phases, and processes that transport C and minerals vertically in soils. Clearly, fraction P itself consists of different components with shorter and longer residence times (Table 4). The oldest C stored in these soils (residue after acid and base hydrolysis of mineral-associated organic matter) has $\delta^{13}C$ values close to those of fresh organic matter, which may be an indication that it is derived from components such as plant waxes that can persist for a long time in soils.

[46] The mean turnover time of soil organic matter in the upper 40 cm of soil, calculated as the weighted mean of active and passive (A and P) components, are inversely
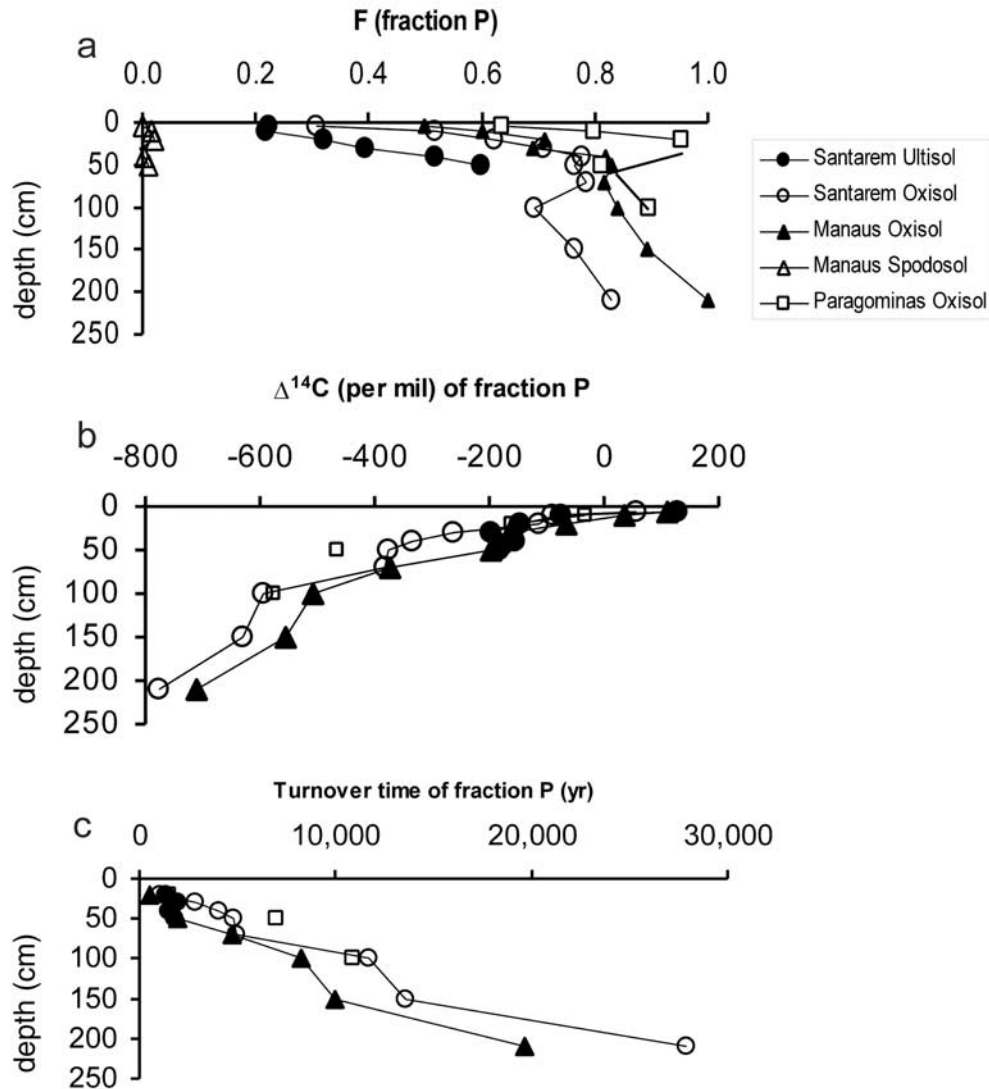
Figure 5. (a) Fraction of total carbon (F), (b) $\Delta^{14}C$, and (c) estimated turnover time of passive fraction C (fraction P) calculated using ^{13}C mass balance (equation (4) in text).

correlated with clay content. The mean turnover times range from 700 years (Paragominas Oxisol) to ~400 years (Manaus Oxisol; Table 4), to 300 years for the Santarem Ultisol, and <100 years for the Manaus Spodosol (Table 4). We do not find that turnover times of more active soil components vary with clay content, though the limitations of radiocarbon methods (especially including uncertainties in the residence time of C in plant tissues) do not allow us to measure this definitively. Mean ages of carbon for the upper ~20 cm of soil range from 200–380 years for clay-rich soils. These are longer than the mean age of organic matter estimated by the Century model [*Schimel et al.*, 1994], largely because of underestimation of the turnover time of "passive" pool carbon by the model.

5.3. Modeling Potential C Uptake in Manaus Soils

[47] The mean age of soil organic matter is not a good predictor of the response of the soil carbon pool to a change in carbon inputs, because it represents an average of components with much faster and much slower turnover times. We used the estimates of C storage and turnover summarized in Tables 1 and 3 and equation (3) to compare the potential for soils in the Manaus region to store carbon assuming a 25% increase in C inputs (Figure 7; Table 4). We assumed ecosystem production increased 0.5% per year for 50 years, then stayed constant at 1.25 times the initial C inputs for each soil organic matter component, similar to the scenario used by *Chambers et al.* [2001] for estimating C storage potential of wood in primary tropical forests in Manaus. Estimates for the Manaus Oxisol (using dynamics derived from $\delta^{13}C$-based fractions, and from Figure 4), and the Manaus Spodosol are compared with the single-pool model assuming 16-year turnover time for all soil carbon proposed by *Malhi et al.* [1999].

[48] Predicted C storage in all soils increases while inputs increase, though the rate of increase slows after 50 years
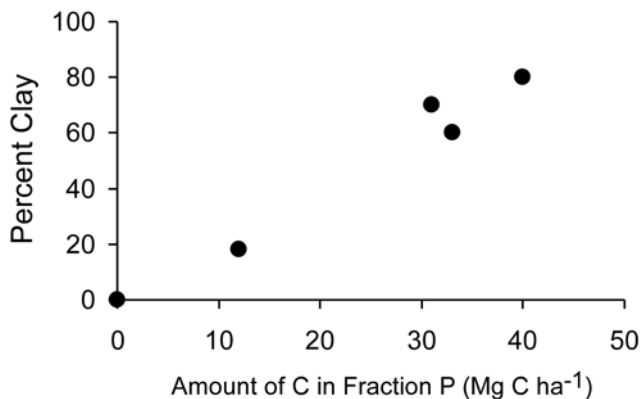
Figure 6. Comparison of the amount of passive fraction C (fraction "P"), calculated using δ^{13}C data (equation (4) in text), with actual soil clay content.

(Figure 7a). While ultimately, when a new steady state is reached, the total increase in C storage will be 25% (given that inputs increased with no change in decomposition rates), the increase in the first century is far less because only a fraction of the total soil organic matter will reach the new steady state on decadal or shorter timescales. Estimated rates of soil C storage in the initial 50 years when NPP is steadily increasing (Figure 7b) are 0.09 to 0.13 MgC ha^{-1} yr^{-1} in soil organic matter, with an additional 0.18 to 0.27 MgC ha^{-1} yr^{-1} potentially stored in litter and roots. Most of the added carbon is being stored in carbon pools with turnover times less than a decade; once the increase in NPP stops, carbon continues to accumulate at very slow rates (<0.05 MgC ha^{-1} yr^{-1}) in carbon pools with turnover times of decades to millennia. In contrast, the single-pool model of *Malhi et al.* [1999] stores up to 0.3 Mg C ha^{-1} yr^{-1} in the upper 40 cm. If applied to the entire 1-m soil depth, assuming initial C storage is ~136 MgC ha^{-1}, we estimate a maximum storage rate of 0.65 MgC ha^{-1} yr^{-1}, far less than the 3.6 MgC ha^{-1} yr^{-1} proposed in their hypothetical C budget.

[49] The rates we calculate for C storage potential in the initial years following an increase in NPP in primary tropical forests are 2–10 times larger than those estimated for temperate and boreal forests, given similar scenarios of a 25% increase in NPP over 50 years (estimated from data presented by *Gaudinski et al.* [2000] and *Trumbore* [2000]).

5.4. Validity of the Steady State Assumption: Constancy of C Inventory Over Time

[50] Our results suggest that C is not presently accumulating rapidly in tropical forest soils in response to a recent increase in NPP, even if productivity of tropical forests has increased in the past decades. A simple test of whether carbon has been accumulating at significant rates at the Manaus Oxisol site is to compare C stocks measured by this study with those reported in the literature. If C were accumulating in the upper meter at a rate of 1–3 Mg C ha^{-1} yr^{-1}, we would expect an increase in the past 20 years of 10–60 Mg C ha^{-1}. Figure 8 compares our data (based on 36 soil pits) with those of *Chauvel* [1982] and *Cerri and Volkoff* [1987]. C inventory and bulk density were meas-

ured using the same or very similar methods in both cases. No significant difference can be found in C inventory over the past 20 years for Oxisols sampled in Manaus ZF2 sites. Hence, our assumption of steady state in modeling radiocarbon data is justified; moreover, it appears impossible for soils at this site to be acting as major carbon sinks over the past 20 years.

5.5. Potential for Export of Soluble Carbon

[51] A question remains as to the potential for soils to export carbon in dissolved organic and inorganic forms with percolating water. Estimates of this export can be made using published data on dissolved organic and inorganic C, and the amount of precipitation that escapes transpiration in the soil profile. *Jipp et al.* [1998] estimated drainage in the Paragominas primary forest sites to be ~10% of the annual precipitation using a time series of soil water content. Assuming a value of 200 mm per year (the equivalent of 200 L m^{-2} year^{-1} of water drainage), and a dissolved organic carbon concentration of ~1 mg C per liter of water, the annual C export as dissolved organic carbon is only 0.002 MgC ha^{-1} yr^{-1}. Dissolved inorganic carbon export will be higher, because the percolating water equilibrates with high CO_2 partial pressures (up to ~10% CO_2 at 8 m depth) in soil pore space [*Davidson and Trumbore*, 1995]. The export of C dissolved in percolating waters equilibrated
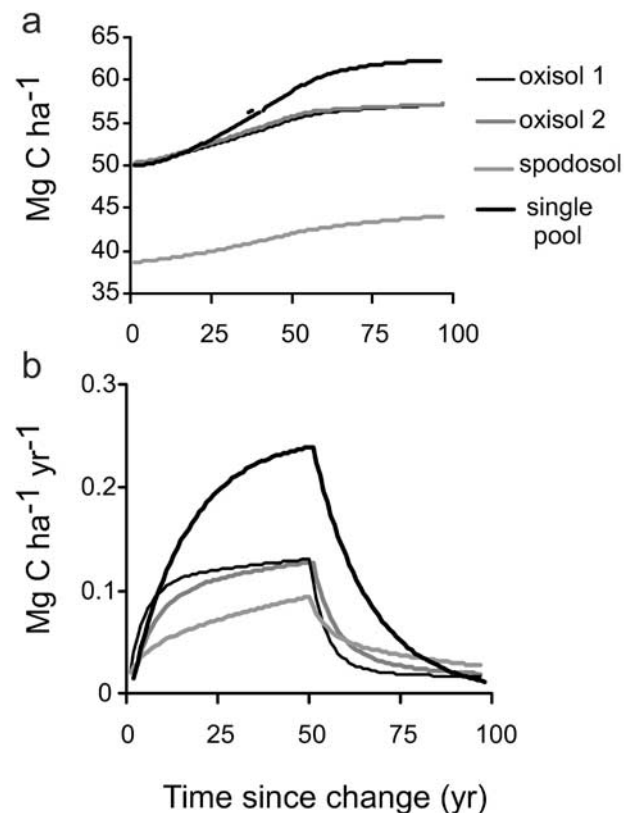
Figure 7. Estimated response of soil organic matter pool to a 25% increase in addition of C inputs to soils over a 50-year period, assuming no change in turnover time. (a) Response in MgC ha^{-1}. (b) Rate of response in MgC ha^{-1} yr^{-1}.
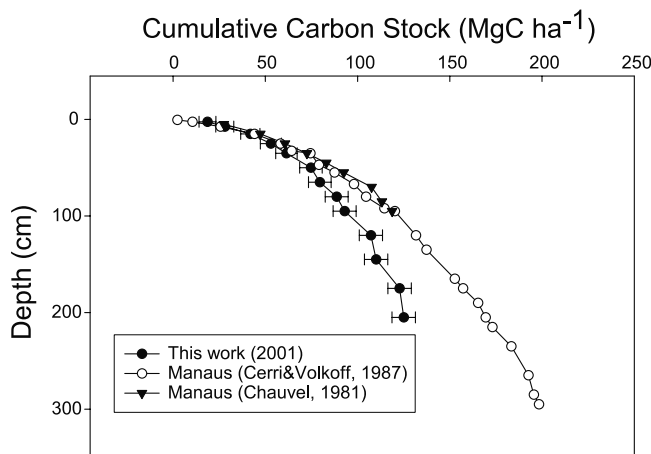
Figure 8. Comparison of cumulative carbon stocks calculated using data from the 1980s and this work. All samples were taken in the ZF2 reserve near Manaus.

with a 10% CO_2 atmosphere is 0.08 MgC ha^{-1} yr^{-1}. These estimates, for terra firme forests (Oxisols), indicate that export of C is not a key factor in determining stand level C balance for a given year. The region of Paragominas, with deep-rooting forests, may have less recharge than areas in the central Amazon; however, even if the amount of water export increased to 40% of precipitation (or 800 cm/yr), the predicted export from terra firme forests would be only ~0.3 MgC ha^{-1} yr^{-1}. Export from inundated soils, like the Spodosol we sampled in Manaus, is likely to be much higher, but remains to be investigated.

5.6. Total Storage Capacity of Terra Firme Tropical Forest

[52] Combining our results with those of *Chambers et al.* [2001] who modeled the effect of a similar increase of NPP on woody biomass of ~0.5 MgC ha^{-1} yr^{-1}, we conclude that the capacity for C storage given increase in NPP of 0.5%/year for the next few decades in biomass and soil is ~0.7 MgC ha^{-1} yr^{-1}, less than estimates suggested for the Manaus ZF2 site by eddy covariance studies [e.g., *Malhi et al.*, 1999]. These calculations assume a sustained small increase in NPP. In scenarios with more rapid NPP change (for example with one year having 25% more production than the previous one), the potential for storage or loss of C from soils is higher, approaching 1 MgC ha^{-1} yr^{-1}. These high rates are not sustained for more than a few years, but point to the potential importance of interannual variability in C storage by tropical forests.

6. Conclusions

[53] Comparing estimates of soil C stocks for a soil in Manaus measured over the past two decades in Oxisols near Manaus, we see no evidence to support C accumulation in terra firme soils at the rates that have been proposed to account for eddy covariance estimates. Using understanding of C dynamics in soils developed from stable isotopes and radiocarbon in soil organic matter, estimates of likely carbon storage in litter and soils in the first decade follow-

ing a sustained increase in ecosystem NPP of 0.5%/year are <0.2 Mg C ha^{-1} yr^{-1}. These values represent a potential response of the soil to an increase in NPP change; we see no evidence that such a change has occurred in the last 20 years. Other factors, such as changes in C allocation by plants, and quality of leaf and root litter materials, are likely to accompany any change in NPP and are not accounted for in our estimates.

[54] Estimates of carbon storage rates do not differ much between soils with different texture, largely because the amount of rapidly cycling carbon remains approximately constant. Soils with more clay contain more soil organic matter, but the majority of it is in forms that do not exchange with atmospheric C on timescales less than decades (in the 0–10 cm depth interval) to millennia (20 cm to 100 cm). Correlation of $\delta^{13}C$ and $\Delta^{14}C$ values in bulk organic matter allows us to separate carbon into fractions that vary in C residence time. Fractions defined in this manner are in accord with independent estimates based on changes in the amount of bomb ^{14}C with time and physical and chemical fractionation methods.

[55] Roots (<2 mm) make up 20–30% of the total carbon stored in the upper 10 cm and 2–12% of the carbon in the upper meter of primary forest soils. Both root and leaf detritus have radiocarbon values that are consistent with carbon fixed from the atmosphere 2–8 years ago. Leaf and root litter decompose relatively rapidly, hence most of the residence time of carbon in these reservoirs reflects time spent in living plant tissue. Models attempting to predict turnover based solely on radiocarbon measurements must account for these lags so as not to underestimate decomposition rates.

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