Global Change Biology (2000) **6**, 317–328

Soil Carbon Sequestration and Land-Use Change: Processes and Potential

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Received 2 June 1999; resubmitted and accepted 6 August 1999

SUMMARY
When agricultural land is no longer used for cultivation and allowed to revert to natural vegetation or replanted to perennial vegetation, soil organic carbon can accumulate by processes that essentially reverse some of the effects responsible for soil organic carbon losses from when the land was converted from perennial vegetation. We discuss the essential elements of what is known about soil organic matter dynamics that may result in enhanced soil carbon sequestration with changes in land-use and soil management. We review literature that reports changes in soil organic carbon after changes in land-use that favor carbon accumulation. This data summary provides a guide to approximate rates of SOC sequestration that are possible with management, and indicates the relative importance of some factors that influence the rates of organic carbon sequestration in soil. There is a large amount of variation in rates and the length of time that carbon may accumulate in soil that are related to the productivity of the recovering vegetation, physical and biological conditions in the soil, and the past history of soil organic carbon inputs and physical disturbance. Maximum rates of C accumulation during the early aggrading stage of perennial vegetation growth, while substantial, are usually much less than 100 g C m^{-2} y^{-1}. Average rates of accumulation are similar for forest or grassland establishment: 33.8 g C m^{-2} y^{-1} and 33.2 g C m^{-2} y^{-1} respectively. These observed rates of soil organic C accumulation, when combined with the small amount of land area involved, are insufficient to account for a significant fraction of the missing C in the global carbon cycle as accumulating in the soils of formerly agricultural land.

Key words: soil carbon, land-use, reforestation, carbon sequestration

1 INTRODUCTION

1.1 Soil Organic Carbon Pools

In terrestrial ecosystems the amount of carbon in soil is usually greater than the amount in living vegetation. It is therefore important to understand the dynamics of soil carbon as well as its role in terrestrial ecosystem carbon balance and the global carbon cycle. The loss of soil organic carbon by conversion of natural vegetation to cultivated use is well known. Various land-uses result in very rapid declines in soil organic matter (Jenny 1941, Davidson and Ackerman 1993, Mann 1986, Schlesinger 1985, Post and Mann 1990). Much of this loss in soil organic carbon can be attributed to reduced inputs of organic matter, increased decomposability of crop residues, and tillage effects that decrease the amount of physical protection to decomposition. We assemble information concerning processes that regulate the amount and rate of change in SOC and use this information to interpret measurements of SOC accumulation when a soil is no longer cultivated and returns to supporting perennial vegetation.

Soil organic carbon includes plant, animal and microbial residues in all stages of decomposition. Many organic compounds in the soil are intimately associated with inorganic soil particles. The turnover rate of the different soil organic carbon compounds varies due to the complex interactions between biological, chemical, and physical processes in soil. Although there may be a continuum of soil organic carbon compounds in terms of their decomposability and turnover time, physical fractionation techniques are often used to define and delineate various relatively-discrete soil organic carbon pools. Physically defined fractions, while containing a diverse array of organic compounds, integrate structural and functional properties of soil organic carbon (Christensen 1996). Physical fractionation of soil emphasizes the role of soil minerals and soil structure in SOC turnover, and relates more directly to SOC dynamics *in situ* than classical wet chemical SOC fractions (Oades 1993, Elliott and Cambardella 1991, Christensen 1992). Figure 1 shows an outline of major physically separated SOC fractions that correspond to pools in many soil carbon turnover models used to

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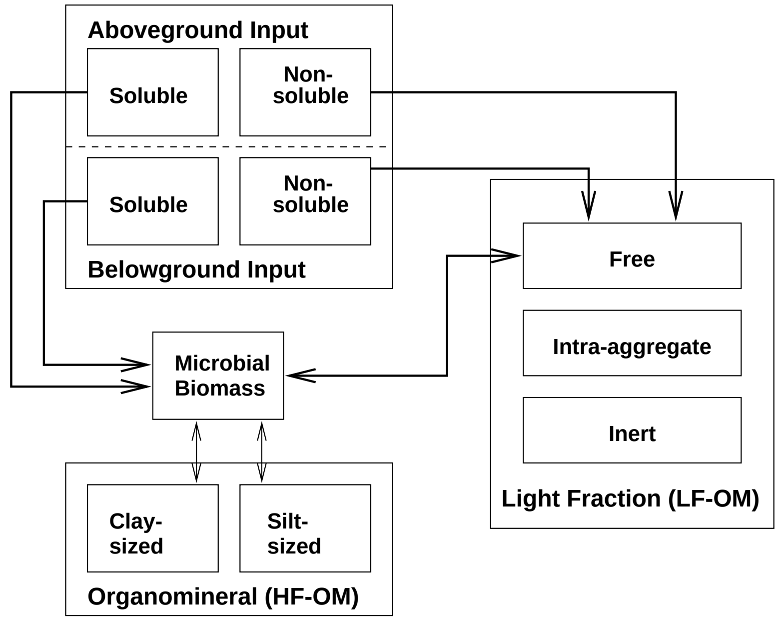

Figure 1. Mineralization and transfer of organic matter in soil. (Christensen 1996)

simulate long-term (decades to centuries) changes in soil carbon (Christensen 1996, Buyanovsky et al. 1994).

The light fraction organic carbon (LF-OC) is free (not complexed with mineral matter), particulate plant and animal residues undergoing decomposition (Spycher et al. 1983). Occasionally some of this material may be biologically resistant, such as charcoal (Skjemstad 1990). Part of the LF-OC can be physically stabilized in macroaggregates as intra-aggregate particulate carbon (Cambardella and Elliot 1992, 1993). Thick surface accumulations of LF-OC occur in boreal and tundra ecosystems where it persists due to low temperatures slowing decomposition. In ecosystems that are more commonly used for cultivation, accumulation of LF-OC can be quite high despite higher decomposition rates where there are significant returns of plant litter (forests and permanent grasslands). This fraction is highly decomposable and can show seasonal fluctuations and spatial variation with changes in litter inputs (Boone 1994). The turnover of LF-OC in such ecosystems is linked to macroaggregate formation and its amount is greatly impacted by cropping and tillage (Beare et al. 1994, Biederbeck et al. 1994, Bremer et al 1994). Short term shifts in SOC storage and turnover are in large part due to the dynamic nature of this pool which has a bulk turnover time measured in months to a few years.

SOC is transformed by bacterial action and stabilized in clay or silt sized organomineral complexes (HF-OC) where the majority of SOC is found (Figure 1). The highest concentrations of SOC are associated with < 5 μm mineral particles. Following the addition of simple substrates, new SOC is found to be associated with a range of mineral particle sizes. However, clay sized organomineral complexes often show greater accumulations and subsequently more rapid loss rates than in silt sized particles, indicating a higher stability of silt-SOC (Christensen 1996). Turnover times of the HF-OC are on the order of decades.

Microbial biomass, while a small portion of SOC, mediates the transfer of SOC among inputs, LF-OC and organomineral HF-OC. As a result, rates of transfer and transformation are influenced by biologically important factors including soil moisture and soil temperature. In addition, most models of SOC turnover postulate a pool of passive (old or stable) carbon with turnover times of 1,500 to 3,500 years or longer (Parton et al. 1988, Jenkinson 1990). The presence of such a pool with long turnover is necessary for consistency with ^{14}C measurements (Harrison et al 1993). This pool is not explicitly shown in Figure 1, since a physical method of isolating this passive SOC fraction is unknown. It is thought that passive SOC is comprised of a nearly inert LF-OC component, say charcoal, and some very chemically recalcitrant material in organomineral HF-OC complexes.

The amount, decomposability (represented in Figure 1 as proportions of soluble and non-soluble components), and placement of aboveground and belowground inputs differ greatly between ecosystem types and with land-use. In agricultural soils aboveground inputs and most roots are mechanically mixed in the surface layer. In permanently vegetated soils, aboveground residues are left on the surface to decompose or a portion may be transported or mixed into the soil by animal activity. Roots and root exudates enter the soil directly. These differences affect decomposition through moisture and temperature conditions, exposure to soil organisms, and degree of contact with mineral soil.

1.2 Soil Potential for Carbon Accumulation

The amount of organic carbon stored in soil results from the net balance between the rate of soil organic carbon inputs and rate of mineralization in each of the organic carbon pools described above. Schlesinger (1990) compiled data on long-term rates of soil organic carbon accumulation in Holocene age soils. He found a slow rate of carbon increase in soil even after thousands of years. This long-term increase represents accumulations of passive soil organic carbon fractions, which include charcoal and resistant compounds physically protected in organomineral complexes. Schlesinger (1990) documented long-term rates of carbon storage from 0.2 g C m^{-2} y^{-1} in some polar deserts to greater than 10 g C m^{-2} y^{-1} in some forest ecosystems, with an average rate of 2.4 g C m^{-2} y^{-1} over all ecosystems. Schlesinger (1990) indicates that faster rates of change over short time periods are possible as a result of changes in environmental conditions.

When natural vegetation is converted to cultivated crops, rapid declines in soil organic matter are partly due to a lower fraction of non-soluble material in the more readily decomposed crop residues. Tillage, in addition to mixing and stirring of soil, breaks up aggregates and exposes organo-mineral surfaces otherwise inaccessible to decomposers. This results in a reduction in the amounts of intra-aggregate LF-OC and some organomineral SOC. Losses of SOC of as much as 50% in surface soils (20 cm) have been observed after cultivation for 30 to 50 years. Reductions average around 30% of the original amount in the top 100-cm. The large and relatively rapid changes in SOC with cultivation indicates that there is considerable potential to enhance the rate of carbon sequestration in soil with management activities that reverse the effects of cultivation on SOC pools. The refilling of depleted fast turnover LC-OC pools and the active portions of the organomineral pools may result in much higher rates of SOC storage than the slow accumulation of passive soil carbon documented by Schlesinger (1990). Although the time period for high accumulation rates may be relatively short, years or decades, these accumulation rates are of significance for current soil sustainability and carbon management issues.

2 METHODS

We collected the available literature that report soil organic matter changes resulting from land conversion from cultivated to perennial vegetation. The information can be organized into two categories of land-use change (Tables 1 and 2). Table 1 reports rates of soil carbon change during forest or woody vegetation establishment after some period of agricultural use. Table 2 contains rates of soil carbon change after establishment of permanent pasture. For most studies, changes in SOC are estimated using paired plots. One or more plots were converted from agricultural use to forest or grassland, while adjacent plots or nearby plots with the same soil were treated as controls or initial conditions in calculating changes of organic soil carbon. At several sites, soil samples were collected periodically during the period of forest growth and were used for making SOC measurements (Jenkinson 1971, Richter et al. 1994, 1999). All studies have measurements that represent changes in soil carbon content during at least one time interval. These are reported in the "avg." column and represent the average rate of change in soil carbon over the time interval computed by taking the total change in carbon amount and dividing by the number of years in the time interval. The actual rate of soil carbon change may or may not be constant over the time interval. Several studies estimated rates of SOC change for two or more time intervals. For these, the maximum rate of change over any time interval is reported in the "max." column. The time course of SOC accumulation for each plot with multiple measurement intervals reported is not presented. There is considerable variation and insufficient replication so that these time courses are not uniform. Presentation of the "max." rates allows some evaluation of whether most of the accumulation occurred over a relatively short period of time (max. much greater than avg.) or whether the accumulation was fairly constant over the entire measurement period (max. approximately the same as avg.). The maximum was often observed during the first measurement interval.

Soil bulk density measurements are required to calculate a carbon amount from studies that report only carbon or organic matter concentrations. For studies where bulk density measurements were absent, we estimated bulk density (BD) using the Adams (1973) equation:

$$BD = \frac{100}{\frac{\%OM}{0.244} + \frac{100-\%OM}{MBD}}$$

where OM is organic matter and MBD is mineral bulk density. Mineral particle density is usually assumed to be the specific gravity of quartz (2.65 Mg m^{-3}). Actual MBD is considerably lower than rock bulf density since soil consists of irregularly shaped mineral particles that allow large voids between them. We used a typical value of 1.64 for MBD (Mann 1986). Soil depth was not adjusted to account for changes in bulk density unless the authors of the original data had already done so. Not adjusting soil depth may, in some cases, result in an underestimate of soil carbon gains.

3 RESULTS AND DISCUSSION

3.1 Forest establishment after agricultural use

Rates of SOC change under aggrading forest range from small losses under cool temperate-zone pine dominated natural succession to an increase of 300 g C m^{-2} y^{-1} in a subtropical wet forest plantation. In two sites large SOC losses were observed when subsequent organic carbon inputs during the early stages of forest growth were not large enough to replenish the decomposition losses. An average accumulation rate of soil carbon, including these two sites, is 33.8 g C m^{-2} y^{-1}. This is quite similar to the 30 g C m^{-2} y^{-1} estimated by Schlesinger (1990) as the rate of SOC accumulation in 40- to 50-year-old soils. There is a tendency for rates of SOC accumulation to increase from temperate regions to subtropical regions (Table 1). We infer from this trend that major factors determining the rate of accumulation are amounts of organic matter inputs which increase with temperature and moisture.

Bredja (1997) studied a subtropical thorn steppe system, where the vegetation was shifted from a grazed grassland to an ungrazed woodland. Growth of woody plants resulted in a decrease in SOC, despite the fact that woody plants produced a greater amount of more recalcitrant material. Woody plants may be less effective than perennial grasses in some environments at storing carbon in soil. Woody plants deposit a larger fraction of total inputs than both grasses and pastures on the surface where decomposition conditions are generally more favorable. This may be the case in this study since a site in a grass opening without woody shrubs showed a SOC increase when grazing was eliminated. Another example that indicates the significance of vertical placement of new carbon inputs is demonstrated in a wet tropical forest life zone where sugar cane fields were converted to fast-growing eucalyptus trees (Bashkin and Binkley 1998). After 10 to 13 years, soil carbon increased under eucalyptus by 1150 g C m^{-2} in the top 10 cm of soil, but decreased by 1010 g C m^{-2} in the 10 to 55 cm layer. By examining changes in ^{13}C concentrations Bashkin and Binkley (1998) were able to confirm that input rates of eucalyptus carbon into this deeper layer were small compared to previous sugar cane inputs.

There is considerable variation in accumulation rates of SOC in Table 1 that results from many factors and are not consistent among the studies. In addition to the differences in the quantity, quality and placement of organic carbon inputs mentioned above, there are differences in the degree that labile soil organic carbon pools, particularly the LF-OC pools, were depleted by cultivation prior to abandonment or land-use conversion. The rate of decomposition is usually well represented as a first order process where

Table 1. Rates of soil carbon accumulating during forest establishment after agricultural use.

Site History	Years since Agriculture	Soil sample Depth (cm)	Rate of C Change ($g\ m^{-2} yr^{-1}$) max.	avg.	Reference
Cool temperate moist forest					
cultivated to pine plantation	42-88	0-10	-8.56	-4.44	Pregitzer and Brian 1977
	42-88	30-40	-5.27	-2.78	
old field succession to northern hardwoods	1-60	10	16.03	15.06	Zak et al. 1990
long-term agriculture to oak forest	83	68.6	61.7	59.60	Jenkinson 1971
long-term agriculture to oak forest(P amended)	82	68.6	33.3	28.0	Jenkinson 1971
old field succession to mixed oak stand	$>$250	15.0	28.3	9.4	Robertson & Vitousek 1981
old field succession to northern hardwoods	$>$100	10.0	17.3	11.6	Robertson & Tiedje 1984
old field to managed pine plantation	10-50	15.4		65.66	Wilde 1964
abandoned field to mixed forest	2-65.5	42.8	23.9	2.15	Hamburg 1984
constructed dike to forest	100	12		26.3	Beke 1990
mine spoil to forest					
Harrison #1	21	22.9		50.09	Leisman 1957
Warren	32	22.9		19.9	Leisman 1957
Silver	41	22.9		38.4	Leisman 1957
Kinney	51	22.9		38.4	Leisman 1957
Warm temperate thorn steppe					
grazing exclusion - shrub live oak	18	3.8		-9.44	Brejda 1997
grazing exclusion - mountain mahogany	18	3.8		-6.11	Brejda 1997
grazing exclusion - shrubless openings	8	3.8		12.78	Brejda 1997
Warm temperate dry forest					
new parent material to chaparral - oak	41	100		70.87	Ulery 1995
new parent material to chaparral - pine	41	100		59.93	Ulery 1995
new parent material to chaparral - chamise	41	100		25.43	Ulery 1995
new parent material to chaparral - ceabothus	41	100		22.00	Ulery 1995
Warm temperate moist forest					
old field to pine to hardwood succession	200	3.0	11.2	2.4	Switzer et al. 1979
	120-180	10.0	12.9	4.5	Montes & Christensen 1986
old field to pine-natural succession	40-60	5.0		2.94	Christensen & MacAller 1985
				-4.41	Christensen & MacAller 1985
				-14.12	Christensen & MacAller 1985
	$\approx$50	33.0		28.4	Schiffman & Johnson 1989
	50-70	17.8	22.6	11.81	Hosner & Graney 1970
	110	68.5-91.4	31.7	5.9	Billings 1938
old field to managed pine plantation	$\approx$50	33.0		24.8	Schiffman & Johnson 1989
	70	12.7	52.85	25.56	Coile 1940
	40	60		3.6	Richter et al. 1999
Subtropical dry forest					
abandoned pasture	25	38.1	-20.36	-13.08	Smith et al. 1951
long-term agriculture to secondary forest	35	25.0		80.0	Brown & Lugo 1990
long-term agriculture to mahogany plantation	50	25.0		38.0	Brown & Lugo 1990
Subtropical moist forest					
long-term agriculture to secondary forest	35	23.0		28.0	Weaver et al. 1987
long-term agriculture to secondary forest	100	50.0	300.0	105.0	Brown & Lugo 1990
coffee plantation to abandoned coffee shade	20	23.0		99.0	Weaver et al. 1987
forest plantation with intensive site preparation	2-34	43.5	566.71	-51.49	Gholz & Fisher 1982
Subtropical wet forest					
10 yr old crop fields to secondary forest	38-47	50.0	566.7	148.8	Brown & Lugo 1990
long-term agriculture to secondary forest	$\approx$35	23.0		98.7	Weaver et al. 1987
10 yr old crop fields to mahogany plantation	51	50.0		310.0	Brown & Lugo 1990
coffee plantation to abandoned coffee shade	30	23.0		10.3	Weaver et al. 1987
Tropical moist forest					
cultivated field to Eucalyptus plantation	11.5	55		12.17	Bashkin et al. 1998
1 year clearing to forest plantation	10.5	100	-1569.6	-47.13	Sanchez et al. 1985
swidden agriculture (forest fallow period)	10	10	143.3	68.9	Aweto 1981
short-term cropping to forest fallow	50	40	740.0	61.2	Ramakrishnan & Toky 1981
Tropical wet forest					
primary succession (P1)	126	1.5	3.36	1.18	Vitousek. et al. 1983

the amount of decomposition per unit time depends on the amount of material subject to decomposition times the rate constant for the environmental conditions and type of material. The amount of material in each decomposition class at the initial time in each study depends on the previous management history, which is generally unknown. As a result, initial decomposition rates may be low (SOC pools relatively depleted) or high (SOC pools large) relative to those that can be maintained if SOC pools were in equilibrium with current input rates of organic matter. When considering management activity to sequester carbon in soil, knowledge of site history such as cultivation duration is important. In two sites large SOC losses were observed. In both cases, the prior period of agricultural use or a bare fallow was short - only one year (Sanchez et al. 1985, Gholz and Fisher 1982). In these cases, the short period of disturbance did not deplete the rapid turnover pools before forest planting.

In approximately half the studies that have multiple measurement intervals there is a large difference between the maximum and average accumulation (or loss) rate. This indicates a slowing in accumulation rate as new steady state amounts of SOC are established and decomposition rates more closely match input rates. Some of these may be artifacts that arise from spatial heterogeneity and insufficient sampling. Additional errors arise with studies of chronosequences and paired plots. The cultivated plots may not

Table 2. Rates of soil carbon accumulating during pasture establishment.

Site History	Years since Agriculture	Soil sample Depth (cm)	Rate of C Change ($g\ m^{-2} yr^{-1}$) max.	avg.	Reference
Cool temperate steppe					
cultivated to perennial grass	12	300		110.0	Gebhart et al. 1994
cultivated to abandoned field	50	10		3.1	Burke et al 1995
cultivated to seeded grass	6	5		0.0	Robles and Burke 1998
cultivated to improved pasture					White et al. 1976
Russian wildrye	8	7		6.86	
crested wheatgrass	8	7		18.87	
B-I-ALF(full)	8	7		14.01	
B-I-ALF(short)	8	7		34.15	
mine tailing to grass-forb meadow	5-80	10	60.0	4.01	Titlyanova et al. 1988
coal mine spoil to dry grassland	28-40	120		28.2	Anderson 1977
Subtropical moist forest					
cultivated to pasture					Lugo et al. 1986
Atlantic	37	18		-16.22	
Caonilas	37	18		-48.65	
Culebrinas	37	18		100.0	
Northwest	37	18		8.11	
West	37	18		37.84	
East	37	18		35.14	
Southeast	37	18		10.81	
Southwest	37	18		67.75	
South	37	18		113.51	
Turabo	37	18		24.32	
Tropical dry forest					
forest to unimproved pasture	23	10		-17.4	Trumbore et al. 1995
forest to improved pasture	23	10		-13.0	Trumbore et al. 1995
Tropical moist forest					
native forest to pasture	10	40		-30.0	Desjardins et al. 1994
mature forest cleared to pasture					Neill et al. 1997
Purto Velho	7	10		83.18	
Calcaulandia	8	10		-4.02	
Nova Vida-1	81	10	342.72	15.23	
Nova Vida-2	20	10	174.81	24.10	
Ouro Preto-Benjamin	20	10	115.54	39.27	
Ouro Preto-lenk	20	10	-84.13	2.37	
Vilhena	12	10	114.83	91.91	
Purto Velho	7	30		-14.29	
Calcaulandia	8	30		-90.00	
Nova Vida-1	81	30	460.0	21.85	
Nova Vida-2	20	30	410.0	59.0	
Ouro Preto-Benjamin	20	30	110.0	74.5	
Ouro Preto-lenk	20	30	49.17	17.0	
Vilhena	12	30	134.0	32.5	
Tropical wet forest					
native forest cleared for pasture					Veldkamp 1994
Eutric Hapludand	25	50		-87.2	
Oxic Humitropept	25	50		-6.0	
native forest cleared for pasture					van Dam et al. 1997
Andic Humitropept	18	60	142.03	34.93	

have been in equilibrium and therefore lost additional SOC since paired perennial vegetation plots were established resulting in an overestimate of the initial rates of SOC accumulation. In 2 studies the max. rates greatly exceed the avg. ones (Gholz and Fisher 1982, Sanchez et al 1985). In both these studies site preparation and mixing of harvest residue into the soil greatly increased the amount of relatively undecomposed residue included in the first time interval measurements.

Many studies have similar max. and avg. rates of SOC accumulation. This indicates that there are conditions where accumulation rate may be fairly constant over periods as long as 50 to 100 years. The clearest examples are from the long-term plots at Rothamsted, England (Jenkinson 1971). These studies avoid most effects of spatial heterogeneity introduced by paired plot studies since samples were taken 3 times from the same plots over the first 80 years of the experiments. The soils at the beginning of the experiments, had been continuously cultivated since Roman occupation nearly 2,000 years ago without any modern production enhancing amendments resulting in very low SOC amounts. The Broadbalk and Geescroft Wildernesses have shown constant rates of SOC accumulation, 60 g C $m^{-2}y^{-1}$ and 30 g C $m^{-2}y^{-1}$ respectively, for over 80 years since cultivation was halted and oak forest appeared through natural succession.

3.2 Permanent grassland establishment

Substantial gains in SOC are also possible with conversion of cropland to grassland, particularly with management for high grass productivity (Table 2). A study of the grassland Conservation Reserve Program (CRP) for a productive portion of U.S. Central Plains (Texas, Kansas, and Nebraska, Gebhart et al. 1994) indicates that SOC may accumulate at an average rate of 110.0 g C m^{-2} y^{-1} in the surface 300 cm. These are probably maximum rates that will decline over time. Results from subtropical moist forest life zones demonstrate a potential for SOC gains when row crops are replaced with managed pasture (Lugo et al. 1986). The mean accumulation rate from this study is 33.2 g C m^{-2} y^{-1}. SOC accumulation rates are much lower under more arid conditions. White et al. (1976)

found lower values in South Dakota - an average of 21 g C m^{-2} y^{-1}, where the rates show considerable variation among plots with different species of plants established. In the shortgrass steppe of Colorado, Burke et al. (1995) found very low rates of SOC accumulation on unimproved abandoned crop fields. They report an accumulation rate of 3.1 g C m^{-2} y^{-1} over a 50 year period. The actual accumulation rate may even be lower if additional losses, including erosion of C from the paired cultivated fields over the last 50 years are taken into account. Robles and Burke (1998) did not find significant soil carbon gains in CRP land soils 6 years following cessation of cultivation in a semiarid grassland (Wyoming), but they did find significant increases in mineralizable and coarse LF-OC. These facts suggest that longer time periods are required for pronounced increases in total SOC under conditions of low productivity.

SOC is likely to increase when cultivated soil is planted with permanent grasses. Conversion of woody vegetation to grasses may also affect SOC storage. Replacement of native tropical savanna with productive, deep-rooting exotic grasses (Fisher et al. 1994) can also result in significant SOC increases for several years (800 to 1300 g C m^{-2} y^{-1} over the first 3 to 6 years). When forest is cleared for pasture establishment, considerable aboveground carbon in vegetation is lost, but it is not necessary that there be declines in SOC. At least for tropical moist and wet forest life zones, it is also possible for SOC to increase when native mature forest is cleared and converted to pasture. In a study by Neil et al. (1997) eleven of fourteen pasture conversion sites studied in Brazil showed increases in soil carbon. All sites in pasture for at least 10 years showed increases, with rates as high as 74.0 g C m^{-2} y^{-1} over 20 years. Two out of three sites studied in Costa Rica found declines in SOC when native forest was cleared for pasture (Veldkamp 1994). However, on one rich volcanic soil with andic properties (soils with high activity clay minerals) conversion to pasture resulted in significant increases in SOC (van Dam et al. 1997). Tropical dry forest showed decreases in SOC when converted to pasture (Trumbore et al. 1995) although site preparation methods complicate interpretation of the results in this study.

4 CONCLUSIONS

There are many factors and processes that determine the direction and rate of change in SOC content when vegetation and soil management practices are changed. Ones that may be important for increasing SOC storage include (1) increasing the input rates of organic matter, (2) changing the decomposability of organic matter inputs that increase LF-OC in particular, (3) placing organic matter deeper in the soil either directly by increasing belowground inputs or indirectly by enhancing surface mixing by soil organisms, and (4) enhancing physical protection through either intra-aggregate or organomineral complexes. Conditions favoring these processes generally occur when soils are converted from cultivated use to permanent perennial vegetation. We observe variation in the rates of SOC change due to the differences in the influences of one or more of these factors even with data collected from similar studies. Additional variation in the data presented in Tables 1 and 2 can be attributed to a lack of consistent initial conditions resulting from differences in cultivation history and spatial heterogeneity.

To obtain a higher precision predictive capability of detecting changes in SOC, additional empirical studies are needed combined with a better understanding of the biological and physical processes involved. Long term agricultural trial have been valuable for understanding soil carbon dynamics under agriculture (Jenkinson 1991). Additional long term experiments that address SOC dynamics when land is converted from cultivation with known management histories to perennial vegetation would be valuable in improving our understanding and increase our predictive capability over short and long time scales.

While there is not enough data currently available to precisely determine the amount of carbon accumulating in any large region or even some particular plot of land, we have enough information to infer the order of magnitude of the soil carbon sequestration rate. We can use this information to investigate some aspects of soil carbon fluxes in the present global carbon cycle. Various investigations have inferred indirectly from atmospheric CO_2 measurements that the terrestrial ecosystems of the Northern Hemisphere have been taking up and storing carbon in both vegetation and soil at a rate of 1 to 2 Pg C y^{-1} (1 Pg = 10^{15} g) over the past several decades (Tans et al. 1989, Ciais et al. 1995, Fan et al. 1998). Various hypotheses have been proposed concerning the processes involved and regions where various processes may have greater or lesser effects on the rate of net carbon uptake. One hypothesis that the information presented here can address is that the inferred northern hemisphere terrestrial sink is related to land-use change over the past 50 years. Hart (1968) reported that over the decade of the 1950's that 27 x 10^6 ha of farmland was abandoned in the eastern United States. The amount of land in the U.S. reforested or afforested in the decade of the 1980's is reported in Winjum et al. (1990) as 12 x 10^6 ha. Taking the larger rate of land-use change reported by Hart (1968), and assuming that this continued from 1950 to present, there is a cumulative potential of 135 x 10^6 ha that support aggrading forest. Applying the average rate of SOC accumulation to this entire area, the present rate of SOC sequestration would be .05 Pg C y^{-1}. Only a small fraction of the infered northern hemisphere terrestrial sink can be explained in terms of SOC accumulations in aggrading forests of the U.S. King et al. (1990) indicate that 154 x 10^6 ha (120 x 10^6 ha in China and India alone) were reforested or afforested in the Northern Hemisphere in the 1980's. Assuming that all this land was all afforested and not just replanted after a harvest and that this rate continued through the 1990's then the rate of SOC accumulation would potentially be 0.11 Pg C y^{-1}. While not insignificant, this rate is a very small fraction of the estimated rate of carbon sequestration in the Northern Hemisphere. If 1 to 2 Pg C y^{-1} are currently accumulating in aggrading forests then nearly all of this C would be increases in biomass and surface litter, not SOC.

5 ACKNOWLEDGMENTS

Research sponsored by the U.S. Department of Energy, Carbon Dioxide Research Program, Environmental Sciences Division, Office of Biological and Environmental Research, under contract DE-AC05-96OR22464 with Lockheed Martin Energy Research Corp. We thank Anthony King, Robert Luxmoore, Steve Pacala, and 2 anonymous reviewers for suggestions that lead to clarifications of various aspects in an earlier version of the manuscript.

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