

CHAPTER 4

Soil Organic Matter: a Source of Atmospheric CO₂

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ABSTRACT

A careful analysis of the literature suggests that there is a pool of carbon held within the soils of the earth of about 1515×10^{15} g. Of this, human activities are causing a net release to the atmosphere of about 0.8×10^{15} g C annually. According to this analysis, the cumulative transfer of carbon to the atmosphere since prehistoric times may have reached 40×10^{15} g C.

4.1 INTRODUCTION

Much terrestrial plant debris accumulates as dead organic matter in soils. The total amount of carbon retained in this pool globally is large. Over the past century, human disturbances of soil layers have caused changes in the size of this pool; estimates suggest that between 10 and 40×10^{15} g C have been transferred to the atmosphere as CO₂ (Bolin, 1977). My purpose is to improve my earlier calculation of the size of the soil carbon pool (Schlesinger, 1977) and to estimate the cumulative transfer and current annual net loss of carbon to the atmosphere.

4.2 THE SOIL CARBON POOL

Previous estimates of the size of the soil carbon pool have differed by a factor of four (Table 4.1). The lowest estimate, 700×10^{15} g C, was based on the carbon content of nine types of US soils which were used to determine the world pool (Bolin, 1970). The highest estimate, 3000×10^{15} g C (Bohn, 1976), was based on an extrapolation from South American soil data to the world. Most recent values range from 1000 to 2000×10^{15} g C.

My 1977 estimate resulted from a search of the literature for data on the amount of carbon in surface organic matter and in underlying layers of soil. It was based on 82 values collected throughout the world. Unfortunately,

Table 4.1 Some estimates of the total carbon content of the soils of the world and net losses to the atmosphere. All values $\times 10^{15}$ g C

	Pool	Recent annual net loss	Cumulative historical loss
Bolin (1970, 1977)	700	0.1–0.5	10–40 ^a
Bazilevich (1974)	1 405		
Bohn (1976, 1978)	3 000	1–2	150 ^a
Baes <i>et al.</i> (1977)	1 080		
(cf. Olson <i>et al.</i> , 1978)			
Schlesinger (1977)	1 456		
Bolin <i>et al.</i> (1979)	1 672		
Ajtay <i>et al.</i> (1979)	2 205		
	1 636		
Buringh (1983)	1 477	4.6	537 ^b
Present paper (Table 4.2)	1 515	0.8	36 ^a

^a Since mid-1800s.^b Since pre-history.

agronomists seldom measure the undecomposed organic matter on the surface of the soil and ecologists frequently do not measure the mineral horizons. Studies by soil scientists often contain data on the percentage of carbon in soils but lack the measurements of soil bulk density to calculate the amount of soil carbon per unit land area. Mean values for carbon in soil profiles, multiplied by the mean area of world ecosystems (Whittaker and Likens, 1973), yielded an estimate of soil carbon for the world of 1456×10^{15} g C. The preliminary estimate was based on relatively limited data and many assumptions. The values for boreal forests, which store large amounts of carbon, and for tropical grasslands, which cover enormous areas of the world, were especially inadequate.

My revised estimate, shown in Table 4.2, utilizes 35 new values in addition to the 82 used in the 1977 estimate. Many of the new values came from studies of the International Biological Program (IBP). Despite a 43 per cent increase in the number of values, the estimate of the world's soil carbon pool changed only four per cent.

Among forest ecosystems, mean values for carbon in the soil profile (forest floor + mineral soil) increase from the lowland tropics to the boreal region. Lowland tropical forest soils, however, are not greatly different from temperate forest soils in terms of total carbon content (Sanchez and Buol, 1975; Sanchez, 1976). High rates of organic matter production in the tropics are accompanied by high rates of decomposition. Better data from tropical forests (Pécrot, 1959; Edwards and Grubb, 1977; Grubb, 1977) have permitted separate

consideration of lowland and montane rainforests, using the areas for each given by Olson *et al.* (1978). Montane tropical forests have large accumulations of soil organic matter (Table 4.2) due to the *Massenerhebung* effect (Grubb, 1971). Although production is lower at high elevations, large accumulations occur because the rate of decomposition is inhibited to a greater extent. Though less dramatic, a similar pattern is also seen on mountains in the temperate zone (Jenny, 1941; Whittaker *et al.*, 1968; Nakane, 1975; Reiners *et al.*, 1975; Franz, 1976; Hanawalt and Whittaker, 1976).

Low temperatures retard decomposition; thus, tundra and boreal forest soils contain large accumulations of soil carbon. In two spruce forests in northern Manitoba, for example, Tarnocai (1972) measured soil carbon contents of 31 and 127 kg C/m², with an average of 80 per cent of this carbon in the permafrost layers. Because turnover of the surface layers is markedly slower at high latitudes (Lang and Forman, 1978), it is not surprising that litter on the soil surface increases from one per cent of the total detritus in tropical forests to 13 per cent in boreal forests (Schlesinger, 1977). My mean value for boreal forest is higher than that of Kononova (1966, 1975), probably because it includes some values for northern forested peatlands.

Table 4.2 Distribution of the world pool of carbon in soil, arranged by ecosystem types for recent (1950–1970) conditions and for 1860

Ecosystem type	Mean soil profile carbon (kg C/m ²)	Recent		1860	
		World area (10 ⁸ ha)	Portion of world soil carbon pool (10 ¹⁵ g C)	World area (10 ⁸ ha)	Portion of world soil carbon pool (10 ¹⁵ g C)
Tropical forest					
Lowland	9.8	22.0	216	23.3	228
Montane	28.7	2.5	72	2.5	72
Temperate forest	13.4	12.0	161	13.0	174
Boreal forest	20.6	12.0	247	12.3	253
Woodland	6.9	8.5	59	10.2	70
Tropical grassland	4.2	15.0	63	16.3	68
Temperate grassland	18.9	9.0	170	11.8	223
Tundra and alpine	20.4	8.0	163	8.0	163
Desert scrub	5.8	18.0	104	18.0	104
Extreme desert, rock and ice	0.17	24.0	4	24.0	4
Cultivated	7.9	14.0	111	5.0	40
Swamp and marsh	72.3	2.0	145	2.1	152
Total		147	1 515	147	1 551

Soils in temperate zone grasslands, such as those in parts of the USSR, contain very large amounts of organic matter (Kononova, 1966, 1975). While a large percentage of the detritus in temperate zone forests is derived from litter and is rapidly mineralized on the forest floor, a major portion of grassland detritus is found deep within the soil as a result of the death of roots. This may partially explain why there is generally more carbon in the soils of temperate zone grasslands than in those of forests of the same region. The mean value for tropical grasslands, 4.2 kg C/m^2 , is very low (Jones, 1973; Sanchez, 1976; Kadeba, 1978). The frequent fires may limit the amount of plant debris on the soil in these ecosystems.

Most of the values for the mean content of soil carbon in my updated estimate are slightly higher than in my 1977 review, reflecting the use of studies involving deeper sampling. Estimates for the mean content of soil carbon in tundra decreased as a result of new IBP data. The most marked change in the revised estimate, however, is due to the use of 7.9 kg C/m^2 as the typical value for cultivated soils. In 1977 the figure used was 12.7 kg C/m^2 , a high value that was biased by the assumption that most cultivated land originally had the same carbon content as temperate zone grasslands. The new value, 7.9 kg C/m^2 , is a weighted mean for cultivated soils worldwide that was derived from the estimates of Revelle and Munk (1977) for the area cultivated in each world ecosystem and from the assumption of a 40 per cent loss of soil carbon during cultivation.

The profile values in Table 4.2 suggest some areas in which Bohn's (1976) estimate seems too high. Bohn's value for chernozem soils, 40 kg C/m^2 , is about twice my mean value for temperate grasslands and is considerably larger than the mean value of 25.3 kg C/m^2 based on many samples of chernozem soils in the USSR (Kononova, 1966). If Bohn's chernozem value is reduced by one half, his world estimate decreases by $100 \times 10^{15} \text{ g C}$. More seriously, Bohn applies the datum from one South American dystic histosol, 200 kg C/m^2 , to much of the boreal region. My data for these regions are limited (Table 4.3), but no value approaches 200 kg C/m^2 . If Bohn's value is reduced to 50 kg C/m^2 , which is probably still too high, his world estimate is lowered by $760 \times 10^{15} \text{ g C}$. The values for carbon in the soil profiles of tropical forests (ferralsols = 12 kg C/m^2) and tundra (regosols = 20 kg C/m^2) given by Bohn are in good agreement with those in Table 4.2.

4.3 LOSSES OF SOIL CARBON

Each year organic matter enters the soil from plant debris derived from above and below ground sources. World-wide this input is estimated at $37.5 \times 10^{15} \text{ g C}$ of carbon per year (Reiners, 1973). At steady-state, decomposition should release an equal amount of carbon to the atmosphere as CO_2 . In fact, root and mycorrhizal respiration as well as the decomposition of

Table 4.3 Soil carbon content of boreal forest soils

Location/Reference	Vegetation type	Sites	Sampling depths (cm)	Soil profile carbon (kg C/m ²)	
				Range	Mean
Canada					
Gross (1946)	Aspen, birch, spruce	9	61–91	8.0–34.8	17.5
Tarnocai (1972)	Spruce	2	180, 319	31.0–127	79.1
Leahey (1947)	Spruce	2	56, 99	15.1–41.0	28.1
Alaska					
Van Cleve (unpublished) in Edmonds (1974)	Spruce	1	100	—	31.8
Crocker and Dickson (1957)	Spruce	1	61	—	9.0
Drew and Shanks (1965)	Spruce	1	38	—	14.6
Van Cleve (personal communication)	Aspen, birch, spruce	5	33–80	17.5–28.4	22.3
Sweden					
Romell (1932)	Spruce	2	100	13.9–16.0	15.0
Nykvist (1971)	Spruce	1	50	—	10.7
USSR					
Kononova (1966)	Mean for forests on podzolic soils	?	100	—	5.0
Suvorov (1974)	Spruce	1	120	—	7.9
Ovchinnikov <i>et al.</i> (1973)	Spruce and cedar	3	150–230	12.8–15.3	13.9
Zaydel'man and Narokova (1975)	Coniferous forests	5	130–185	10.5–15.5	13.5
Mean for boreal forest					20.6

large amounts of root detritus (e.g., Edwards and Harris, 1977) result in the production of additional CO₂ which is not usually measured as input. The release of CO₂ from world soils has been reviewed recently (Schlesinger, 1977); the total steady-state carbon output from soils may be as much as 75×10^{15} g C of carbon per year.

Soil organic matter exists in many forms, Kononova (1972, 1975) separates the various forms of soil organic matter into incompletely decomposed organic matter and humus. Humus is in turn subdivided into strictly humus substances (e.g., humic and fulvic acids) and other products synthesized by soil microbes. Some humus substances are very stable. If my present estimate of

1515×10^{15} g C is divided by Reiners' (1973) estimate of 37.5×10^{15} g C as the annual production of detritus worldwide, the mean turnover time is 40 years. It is certain, however, that specific fractions of the soil organic matter in various ecosystems are far more or far less refractory. It would be instructive to make an estimate of how much soil carbon turns over in 1, 10, 100 and 1000+ years.

The clearing of forests and the conversion of forested land to farmland often reduce the soil carbon content through reduced production of plant detritus and through increased rates of erosion and decomposition of organic matter. The separation of soil carbon into fractions with different turnover rates would therefore assume great significance in evaluating soil carbon losses due to cultivation. If most of the world pool of soil carbon consists of relatively refractory carbon compounds which are unlikely to be disturbed by man's activities, we may be overemphasizing the role of soil disturbance in contributing to the amount of CO_2 in the atmosphere.

Agricultural studies give one indication of the fraction of soil organic matter that is easily decomposed. When soils are cultivated, the surface layers (0–30 cm) often lose 20 to 50 per cent of their carbon content. The rate of loss is greatest in the first few years of disturbance and slows thereafter, following the pattern typical of decomposing litter (Olson, 1963; Minderman, 1968). Harcombe (1977) listed seven studies of tropical forests in which the loss of carbon ranged from 4 to 54 per cent within one to three years after forest removal (cf. Ayanaba *et al.*, 1976; Juo and Lal, 1977). In Sierra Leone, Brams (1971) found a 50 per cent loss of organic matter in five years of cultivation (Figure 4.1). Similarly, Jones (1973) found that the cultivated soils of tropical savannahs have about one half the carbon concentrations found in undisturbed areas. Studies in temperate zone grasslands have found similar losses of organic matter from the surface layer—25 per cent in 43 years (Haas *et al.*, 1957), 31 to 56 per cent in 50+ years (Meints and Peterson, 1977) and 50 per cent in 60 years (Martel and Paul, 1974). When temperate forests are cleared for farmland, the losses of soil carbon also range up to 50 per cent (Giddens, 1957; Rubilin and Dolotov, 1967). Losses from cultivated organic soils (histosols) can be particularly great (Browder and Volk, 1978). Large additions of crop residues or manure can offset the apparent losses. The loss is often attenuated by depth. One study reported an increase in the soil carbon in the 100 to 200 cm portion of cultivated soils, but the increase was small in terms of total profile content (Meints and Peterson, 1977).

Since a large portion of the soil carbon is usually found above a depth of 50 cm, these agricultural studies suggest that nearly half of the carbon in a typical soil profile is relatively labile. The remainder is found in the lower soil profile and is very stable. In tropical regions, carbon in the lower profile is often fixed on allophane; organic matter fixed in tropical soils is often red or colourless, a fact which has perpetuated the notion that these soils are low in organic content

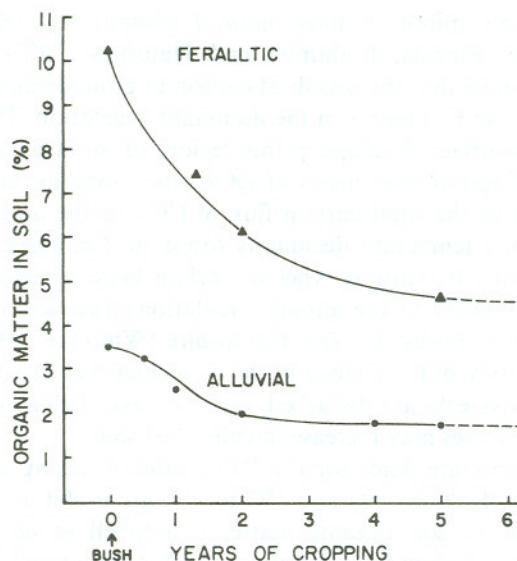


Figure 4.1 Organic matter diminution in tropical soils as a function of the number of years since clearing of native vegetation (from Brams, 1971, reproduced with permission).

(Sanchez, 1976). In other regions, crystalline clay minerals are important in complexing soil organic matter in the lower horizons (Birch and Friend, 1956; Kononova, 1966; Campbell *et al.*, 1967b; Jones 1973).

Most of the carbon lost from disturbed soils is probably lost through increased oxidation rather than erosion. Brink *et al.* (1977) report that average soil losses due to erosion in intensively farmed Wisconsin watersheds ranged from 1.2 to 1.9 kg/m² per year. These values are typical of all but the most careless of land clearing operations (Bormann *et al.*, 1974; Pimentel *et al.*, 1976). If we assume that these soils contain an average carbon content of two per cent, the resulting loss of carbon by erosion is rather small compared to the loss observed when a typical temperate soil is cultivated. Erosional removal of organic matter following forest cutting is probably unimportant as long as vegetation is allowed to regenerate immediately (Bormann *et al.*, 1974). Even in regrowing forests, however, the forest floor often continues to decline for a number of years after deforestation, since the young forest is not likely to produce detritus at a rate equal to the loss by decomposition. Covington (1976) measured a decline in the forest floor mass in New England for 15 years after cutting, but in subsequent years the mass increased to near steady state levels (see also Aber *et al.*, 1978).

Percolating water may transfer soil carbon to groundwater in the form of dissolved organic carbon (DOC), dissolved CO₂, HCO₃⁻ and CO₃²⁻. These

losses are probably minor in most natural communities (Schlesinger and Melack, 1981). In Florida, Rightmire and Hanshaw (1973) used $^{13}\text{C}/^{12}\text{C}$ isotope ratios to show that the dissolved carbon in groundwater did not reflect a change from C-3 to C-4 plants in the dominant vegetation. Thus, despite the importance of subsurface drainage in this region, subsurface carbon losses are probably slight. Groundwater losses of DOC and inorganic carbon were less than 0.3 per cent of the total carbon flux of CO_2 to the atmosphere due to decomposition in a temperate deciduous forest in Tennessee (Edwards and Harris, 1977). Similarly, various types of carbon losses in groundwater were extremely minor relative to the annual circulation of carbon in the Hubbard Brook Experimental Forest in New Hampshire (Whittaker, 1975). There has been almost no study of the change in the rate of carbon loss to groundwater when natural ecosystems are disturbed, but the data of Spalding *et al.* (1978) suggest that these losses may increase in cultivated soils.

When adjustments are made for the ^{14}C produced during atomic weapons testing, the ratios of ^{13}C to ^{12}C and ^{14}C to ^{12}C are useful in determining the age and turnover of soil organic matter (Campbell *et al.*, 1967a, 1967b; Scharpenseel, 1975; Jenkinson and Rayner, 1977; Bottner and Peyronel, 1977; O'Brien and Stout, 1978). Most of these studies have found that soil carbon consists of a small portion of relatively young material that is near the surface and turns over rapidly, and a larger amount of carbon dispersed throughout the profile with an older (500 to 2000 years) weighted mean age (Figure 4.2). The distribution of soil carbon declines exponentially with depth. O'Brien and Stout (1978) used the distribution of soil carbon enriched with bomb-derived ^{14}C to calculate a coefficient for downward diffusivity of $13 \text{ cm}^2/\text{y}$ for carbon in a New Zealand pasture soil. Nakane and Shinozaki (1978) derived downward velocities of movement in various ecosystems from a model of soil carbon, taking into account decomposition and transport processes (see also Nakane, 1976, 1978).

The turnover time calculated by dividing the total pool of carbon by annual input is often much more rapid than indicated by the weighted mean age of carbon in the entire profile as determined by ^{14}C dating (Jenkinson and Rayner, 1977). One should remember that fractions with old weighted mean ages (e.g., humic and fulvic acids) consist of both young and old resistant carbon. Cultivation presumably causes a loss of surface organic matter and the labile portion of the more resistant fractions. As a result, the weighted mean age of the remaining material increases. Martel and Paul (1974) measured a mean residence time of 710 years for the organic matter in cultivated soils in Canadian grasslands, whereas organic matter in nearby uncultivated grasslands had a mean residence time of 250 years. O'Brien and Stout (1978) used isotope ratios to find that 16 per cent of the organic matter in the soil of a New Zealand pasture was 'very old', that is, with a minimum age of 5700 years. 'Modern' carbon was concentrated near the surface and was mostly less than 100 years old.

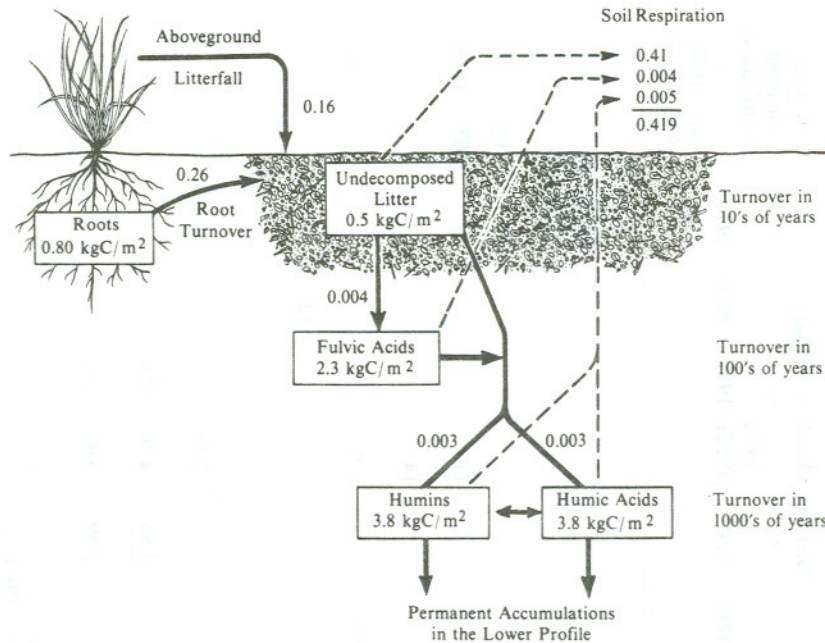


Figure 4.2 A hypothetical model for the soil carbon dynamics in the 0–20 cm layer of a chernozem grassland soil. Carbon pools (kg C/m^2) and annual transfers ($\text{kg C/m}^2/\text{y}$) are indicated. Total profile content to 20 cm is 10.4 kg C/m^2 with an overall turnover time of 25 years based on current annual input. Data are from many sources; see Schlesinger (1977)

New steadystate conditions can be achieved despite the lowered production of plant debris typical of agricultural lands, but the total amount of carbon in the soil is also lower. In conjunction with ^{14}C studies, Jenkinson and Rayner (1977) modelled the carbon dynamics in the 0–23 cm layer of a cultivated soil with an input of 100 g C/m^2 per year. Their model predicted a soil content of 2.4 kg C/m^2 , about 35 per cent less than in a cultivated soil receiving 150 g C/m^2 per year and presumably less than in either area before clearing. Thus, while there are relatively few radiocarbon studies, the current evidence confirms empirical studies which show carbon losses of 20 to 50 per cent when soils are brought under long-term cultivation. I have used a 40 per cent loss factor in my estimates.

4.4 LAND CULTIVATION

Some estimates of the amount of land disturbed by human activities during the last century are shown in Table 4.4. Because many of the estimates are

Table 4.4 Estimates of the area of cultivated and disturbed land at various times in history

Source	Category	Pre-historic	Disturbed land (× 10 ⁸ ha)							Rate of clearing (× 10 ⁸ ha/y)			
			1860	1882	1910	1950		1959	1966	1970	1860–1970	1950–1970	
						(1952)							
World													
Olson <i>et al.</i> (1978)	Crops	0	5							12			
	Fringe	1	3							7			
	Buildings	0	1							3			
	Total	1	9							22	0.12		
Revelle and Munk (1977)	Agricultural land		5			11				14	0.07	0.15	
Whittaker and Likens (1973)	Cultivated					14 ^a							
Doane in Borgstrom (1969)	Tilled land			8.6		(11)							
Borgstrom from 1966 FAO Data (1969)	Tilled land								14				
North America													
Revelle and Munk (1977)	Agricultural land		0.58			2.35				2.35			
Hart (1968)	Total farmland (US only)				3.56	4.70	4.54						
	Cleared farmland (US only)				2.59	3.43	3.31						

^a Value presumably includes all land disturbed by man; thus higher than others for 1950–1952.

based on data from the Food and Agricultural Organization (FAO), it is not surprising that there is some similarity in the values for cropland, but the estimate by Olson *et al.* (1978) strongly suggests that the total area disturbed by people is much greater than the agricultural land alone. Nevertheless, it is likely that from the time people began cultivating land to some point between 1950 and 1970, 14×10^8 ha have been converted to agriculture. This is the value used in Table 4.2.

The considerable variations in Table 4.4 are due to problems with defining cultivated land and estimating its area. One might expect some of the best data to exist for North America, yet the FAO estimates for North America (Revelle and Munk, 1977) are considerably lower than an independent estimate for the United States alone (Hart, 1968).

Revelle and Munk (1977) present a breakdown of land cultivation for each ecosystem type from 1860 to 1970, during which time they estimate that 8.5×10^8 ha were disturbed. When their data are applied to present estimates of ecosystem areas, the land area for each ecosystem in 1860 and hence the soil carbon pool at that time can be estimated (Table 4.2). On a worldwide basis, this approach suggests that approximately 36×10^{15} g C have been released from soils from 1860 to the present. Similarly, the recent rate of new cultivation (1950–1970) can be calculated as 15×10^6 ha/y (Table 4.4). If these soils contain a weighted mean of 13.1 kg C/m^2 , which drops 40 per cent to 7.9 kg C/m^2 after disturbance, the present annual net release is on the order of 0.8×10^{15} g C. This estimate is higher than the range (0.1 to 0.5×10^{15} g C/yr) of potential rates of carbon release from soils suggested by Bolin (1977), but less than the value of 2×10^{15} g C/yr estimated by Woodwell *et al.* (1978). A net loss of 0.8×10^{15} g C/yr represents only a one per cent increase in the steady-state annual rate of CO₂ release from soils worldwide due to decomposition.

4.5 PROBLEMS AND APPROACHES

One of the basic tasks of soil science is to describe the physical and chemical properties of the various horizons that form a complete soil profile. The literature of soil science contains an enormous amount of these data from various parts of the world. A small subset of these data is the source of many of the profile values used in my world estimate. Further field sampling is not likely to improve our understanding of the soil carbon pool in most parts of the world, but boreal forest and tropical grassland regions are exceptions. There has been relatively little field work in these large areas, and field studies in these regions should receive high priority in the future. Soil scientists should routinely measure soil bulk density so that their measures of the percentage of carbon in the soil may be used to calculate the carbon content on an area basis with greater accuracy than is currently possible.

Rather arbitrarily, this paper's estimate of soil carbon is based on the

estimates of Whittaker and Likens (1973) of the land area for world ecosystems (Table 4.2) and Revelle and Munk's (1977) estimates of disturbance (Table 4.4). As Golley (1972) has clearly indicated, there is considerable disagreement about the areal extent of world systems. Given the large variation in the estimates of disturbed land area, great improvement in our ability to estimate the release of carbon from soils could be made if the land areas were systematically studied. If we knew the extent of the world's ecosystems as identified by vegetation, and recent changes in each, we would be able to improve our estimates of the soil carbon accumulated or released in areas with different types of vegetation. Remote sensing from satellites such as LANDSAT is an obvious possibility for achieving this goal (Adrien and Baumgardner, 1977).

Alternatively, it may be possible through the use of LANDSAT to improve the mapping of soil areas at a scale appropriate for world estimates (Lewis *et al.*, 1975; Westin and Frazee, 1976; Weismiller *et al.*, 1977; Valentine, 1978). By combining data on the areal extent of soil groups with field data on soil profiles, we might be able to estimate the world pool of soil carbon with great accuracy. However, it will probably be more difficult to use this approach to estimate recent changes in the soil pool since vegetation is more visibly affected by man's activities.

The science of measuring the soil carbon content of world regions directly from satellite photographs is still in its infancy. Baumgardner *et al.* (1970) were some of the first to suggest the mapping of the organic content of surface layers of soils from aerial photographs, but subsequent studies have not always been encouraging. Evans *et al.* (1976) found that photo-tone was significantly correlated ($r=0.57$) with the soil organic matter content of the surface layers in particular cultivated fields in Great Britain, but when the same approach was applied to a broader area of 60 km², the correlation dropped to disappointing levels ($r=0.31$). Cihlar and Protz (1973) also found that the correlation between photo-density values and soil properties was poor. There is apparently little possibility at the present time of estimating soil carbon content beneath areas of intact vegetation. The problem is complicated by the fact that the radiance waveband (0.62–0.66 μ m) correlated with soil organic content (Baumgardner *et al.*, 1970) is also that of chlorophyll absorption (Adrien and Baumgardner, 1977). These studies have examined the surface layer of cultivated soils, but more have evaluated the feasibility of measuring the carbon content in an entire soil profile.

There are several ways in which soils may be serving as a sink for atmospheric CO₂. When marginal soils are irrigated and fertilized, for example, soil organic matter may increase. Hart (1968) noted that 27×10^6 ha of farmland had been abandoned in the eastern United States, and that 16.7×10^6 ha of this land had been returned to forest between 1910 and 1959. In these areas, presumably, soil carbon increases to pre-agricultural levels. In

desert regions the deposition of calcium carbonate in soils may fix CO₂, although this process is likely to be minor on a global basis (Schlesinger, 1982).

Chancellor and Goss (1976) predict that 20×10^8 ha of land will be tilled by the year 2000. This would mean that 0.17×10^8 ha of land would be cleared each year for the rest of this century. On the basis of their estimate one could estimate the annual release of carbon from soils to be about 0.9×10^{15} g C per year, resulting in an additional 30×10^{15} g C being transferred to the atmosphere between now and the end of the century. Loomis (1978) presents a more apocalyptic view. Using Jenny's (1930, 1941) equations, which predict the content of soil nitrogen and organic matter in a region as a function of the mean annual temperature and a 'moisture' factor, Loomis suggests that 300 to 400×10^{15} g C might be released to the atmosphere with each 1.0 °C increase in global temperature. Presumably, this release would occur predominantly in cold regions with large carbon accumulations (e.g., boreal forest and tundra). Further atmospheric warming might then occur. Because Jenny's equations have little applicability on a broad geographic basis (Schlesinger, 1977), Loomis's calculation is rather crude, but the large value obtained suggests the possible upper limit of carbon release from soils.

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