Available online at www.sciencedirect.com
SCIENCE DIRECT®

Environment International 29 (2003) 437–450

ENVIRONMENT
INTERNATIONAL

www.elsevier.com/locate/envint

Soil erosion and the global carbon budget

R. Lal*

School of Natural Resources, The Ohio State University, 2021 Coffey Road, Columbus, OH 43210-1085, USA

Received 18 September 2002; accepted 29 November 2002

Abstract

Soil erosion is the most widespread form of soil degradation. Land area globally affected by erosion is 1094 million ha (Mha) by water erosion, of which 751 Mha is severely affected, and 549 Mha by wind erosion, of which 296 Mha is severely affected. Whereas the effects of erosion on productivity and non-point source pollution are widely recognized, those on the C dynamics and attendant emission of greenhouse gases (GHGs) are not. Despite its global significance, erosion-induced carbon (C) emission into the atmosphere remains misunderstood and an unquantified component of the global carbon budget. Soil erosion is a four-stage process involving detachment, breakdown, transport/redistribution and deposition of sediments. The soil organic carbon (SOC) pool is influenced during all four stages. Being a selective process, erosion preferentially removes the light organic fraction of a low density of < 1.8 Mg/m^3. A combination of mineralization and C export by erosion causes a severe depletion of the SOC pool on eroded compared with uneroded or slightly eroded soils. In addition, the SOC redistributed over the landscape or deposited in depressional sites may be prone to mineralization because of breakdown of aggregates leading to exposure of hitherto encapsulated C to microbial processes among other reasons. Depending on the delivery ratio or the fraction of the sediment delivered to the river system, gross erosion by water may be 75 billion Mg, of which 15–20 billion Mg are transported by the rivers into the aquatic ecosystems and eventually into the ocean. The amount of total C displaced by erosion on the earth, assuming a delivery ratio of 10% and SOC content of 2–3%, may be 4.0–6.0 Pg/year. With 20% emission due to mineralization of the displaced C, erosion-induced emission may be 0.8–1.2 Pg C/year on the earth. Thus, soil erosion has a strong impact on the global C cycle and this component must be considered while assessing the global C budget. Adoption of conservation-effective measures may reduce the risks of C emission and sequester C in soil and biota.

Keywords: Global warming; Greenhouse effect; Soil organic matter; Soil degradation; Desertification; Global carbon cycle

1. Introduction

Soil erosion, displacement of soil from the place of its formation by causative agents (e.g., raindrop, runoff, wind, gravity, etc.) and its deposition at a depressional and/or protected site, is a natural geologic phenomenon. The slow geologic erosion is a constructive process, which has created vast tracts of fertile soils of alluvial flood plains and loess plateaus around the world. These soils, with built-in soil fertility renewal mechanisms, have supported ancient civilizations (e.g., in the valleys of the Nile, Euphrates, Indus, Yangtze) and thriving cultures for millennia. In contrast, the accelerated soil erosion, exacerbated by anthropogenic perturbations, is a destructive process. It depletes soil fertility, degrades soil structure, reduces the effective rooting depth and destroys the most basic of all natural resources. Numerous, once-thriving civilizations have vanished because of the degradation of the resource base on which they arose (Lowdermilk, 1953; Eckholm, 1976; Olson, 1981; Brown, 1981, 1984; Dregne, 1982).

The importance of protecting and restoring the soil resource is increasingly been recognized by the world community (Lal, 1998; Barford et al., 2001; Lal, 2001). Sustainable management of soil received strong support at the Rio summit in 1992 and its Agenda 21 (UNCED, 1992), UN Framework Convention on Climate Change (UNFCCC, 1992) and Articles 3.3 and 3.4 of the Kyoto Protocol (UNFCC, 1997), the 1994 UN Framework Convention to Combat Desertification (UNFCD, 1996). These conventions are indicative of the recognition by the world community of strong linkages between soil degradation and desertification on the one hand and loss of biodiversity, threat to food security, increase in poverty and risks of accelerated greenhouse effect and climate change on the other.

---

* Tel.: +1-614-292-9069; fax: +1-614-292-7432.
E-mail address: lal.1@osu.edu (R. Lal).

0160-4120/03/$ - see front matter © 2003 Elsevier Science Ltd. All rights reserved.
doi:10.1016/S0160-4120(02)00192-7

Whereas the adverse impacts of accelerated soil erosion on soil degradation (Bennett, 1939; Jacks and Whyte, 1939; UNEP, 1992) food security (Oldeman, 1998; Scherr, 1999) and water quality/sedimentation, etc. (Lal, 1998) have long been recognized, its effects on carbon (C) dynamics and emission of carbon dioxide (CO_2) and other greenhouse gases (GHGs) into the atmosphere have not been given the emphasis they deserve. Although transfer and redistribution of soil organic carbon (SOC) by erosion is evident and widely known, its impact on the global C budget is neither understood nor accounted for. Therefore, this manuscript reviews the state of the knowledge about the extent and severity of soil erosion, impact of erosion on soil C dynamics, fate of eroded soil C and the potential of erosion management and desertification control to sequester C and mitigate the accelerated greenhouse effect.

2. Global extent and severity of soil erosion

Accelerated soil erosion has plagued mankind ever since the dawn of settled agriculture, and the literature is replete with gloomy and catastrophic effects. UNEP (1986) estimated that 2 billion ha of land that was once biologically productive has been irreversibly degraded since 1000 AD. Rozanov et al. (1990) reported that more productive soil may have been irreversibly lost in the past 10,000 years than is currently under agricultural production [estimated at about 1500 million ha (Mha)]. Yet most available statistics on the extent and severity of soil erosion is subjective, qualitative, obsolete, crude and unreliable. There are several uncertainties in these available statistics on the extent and severity of soil erosion. The most widely used statistics on soil erosion is that by Oldeman (1994) (Table 1). Total land area affected by water erosion is 1094 Mha, of which 751 Mha is severely affected, and that by wind erosion is 549 Mha, of which 296 Mha is severely affected. There are also regional hot spots of erosion including the Himalayan–Tibetan ecosystem in South Asia, the Loess Plateau in China, the sub-humid and semiarid regions of sub-Saharan Africa, highlands of Central America, the Andean region, Haiti and the Caribbean (Scherr and Yadav, 1996).

Table 1
Global extent of severe (moderate + level) water and wind erosion (adapted from Oldeman, 1994; Scherr, 1999)

Region	Land area affected by severe erosion (moderate + level) (Mha)			Total as a percent of total land used
	Water erosion	Wind erosion	Total	
Africa	169	98	267	16
Asia	317	90	407	15
South America	77	16	93	6
Central America	45	5	50	25
North America	46	32	78	7
Europe	93	39	132	17
Oceania	4	16	20	3
World	751	296	1047	12

Table 2
The global pattern of suspended yield and estimates of gross erosion on continental scale (the data on suspended load is recalculated from Walling and Webb, 1983)

Region	Land area ($\times 10^9$ ha)	Sediment load ($\times 10^6$ Mg/year)	Gross erosion assuming delivery ratio of 10% ($\times 10^9$ Mg/year)	Mean erosion rate (Mg/ha/year)
Africa	3.01	3891.3	38.9	12.9
Asia	4.46	7401.3	74.0	16.6
South America	1.78	3942.6	39.4	22.1
North America	2.28	2808.6	28.1	12.3
Europe	0.98	1312.7	13.1	13.4
Oceania	0.80	763.8	7.6	9.5
Total	13.26	20,120.3	201.1	15.2

Similar to the land area affected, there are also uncertainties about the extent of sediment transported into the oceans. Estimates of global rates of soil erosion have been made at 75 billion Mg/year by Pimentel et al. (1995), assuming an average erosion rate of 100 Mg/ha on 751 Mha of area affected by severe erosion (Table 1). The first attempt at estimating the sediment transport by rivers was that by Fournier (1960). The next attempt by Strakhov (1967) differed from that by Fournier both in trends and sediment yield by several orders of magnitude. Since then, several attempts at estimating sediment transport have been made including those by Jansson (1982, 1988), Milliman and Meade (1983), Walling (1983, 1996, 1997), and Walling and Webb (1983, 1985, 1987, 1988, 1996). Milliman and Syvitski (1992) estimated that the annual sediment transport to the ocean is about 20 billion Mg, but the mass of sediment in motion is about 30 billion Mg/year. Most available literature shows that the annual sediment transport into the ocean by the world's rivers is 15–20 billion Mg (Walling and Webb, 1996).

Global maps of suspended sediment yield on continental basis have been compiled by Walling and Webb (1983) and Lvovich et al. (1991). Estimating the actual soil erosion using the sediment yield data requires information on the delivery ratio. It is defined as the fraction of the sediment originating over the land delivered to the major drainage or river system (Walling, 1983). Assuming a delivery ratio of 13–20%, estimates of gross erosion on continental scale are shown in Tables 2 and 3. The data in Table 2 is based on the suspended sediment map prepared by Walling and Webb and that in Table 3 by Lvovich et al. The data in Table 2 indicating global suspended sediment load of 20 billion Mg/year and global erosion rate of 7–11 Mg/ha/year seem more realistic than about twice the rates shown in Table 3. The global erosion rate of 197–303 billion Mg is about 2.6–4.0 times that reported by Pimentel et al. (1995). For further analyses, therefore, the

Table 3
The global pattern of suspended sediment yield and estimates of gross erosion rate at continental scale (the data on suspended load is recalculated from Lvovich et al., 1991)

Region	Land area ($\times 10^9$ ha)	Suspended load ($\times 10^6$ Mg)	Gross erosion assuming delivery ratio of 13–20% ($\times 10^9$ Mg/year)	Mean erosion rate (Mg/ha/year)
Africa	3.01	2342.1	23.4	7.8
Asia	4.46	22,088.3	220.9	49.5
South America	1.78	5599.7	56.0	31.5
North America	2.28	4534.2	45.3	19.9
Europe	0.98	3288.2	32.9	33.6
Oceania	0.80	1525.5	15.3	19.1
Total	13.26	39,378.0	393.8	26.7

calculations presented in Table 2 will be used in this manuscript.

Table 5
The global C budget in 1990s (adapted from Schimel et al., 2001)

Activity	Flux (Pg C/year)
I. Sources	
1. Fossil fuel combustion and cement production	6.3 ± 0.4
2. Land use change	1.6 ± 0.8
II. Sinks	
1. Atmospheric increase	3.2 ± 0.1
2. Ocean uptake	1.7 ± 0.5
3. Terrestrial uptake	1.4 ± 0.7
III. Unknown residual sink	2–4

3. The global carbon budget and erosional processes

There has been a rapid increase in the atmospheric concentration of CO_2 and other greenhouse gases (GHGs) into the atmosphere. For example, the concentration of CO_2 has increased by 31% from 280 ppmv in 1750 to 367 ppmv in 1999, and is increasing at an average rate of 1.5 ppmv/year or 0.4%/year. This and similar increases in concentrations of CH_4, N_2O and other GHGs (Table 4) have raised concerns regarding the identification of sources and sinks of these gases. Attempts have also been made to prepare a global budget of C (Table 5), albeit with numerous uncertainties in the sources and sinks. While the emissions from fossil fuel combustion and cement manufacturing are reliable, those from land use change amount to sophisticated educated guesses even under the best case scenarios. Being non-point or diffused sources, it is difficult to precisely estimate local, regional and global emissions of CO_2 and other GHGs due to deforestation, biomass burning, soil cultivation, use of fertilizers and manures and other activities related to land use conversion. Because the C budget cannot be balanced, the so-called "missing sink" (Tans et al., 1990) or the "unknown residual sink" (Schimel et al., 2001) lumps all the uncertainties. In fact, the global C budget presented in Table 5 is grossly inadequate because all sources and sinks of CO_2 due to land use change and soil erosion are not accounted for. There has been a considerable interest in the land-based C sinks (Pacala et al., 2002; Hurtt et al., 2002; Myneni et al., 2001a,b). The unknown sinks are estimated at 2–4 Pg C/year (Schimel et al., 2001; Table 5). Identification of unknown sinks is important in developing strategies for mitigating potential climate change or at least reducing the rate of enrichment of GHGs in the atmosphere. The main challenge is to identify all the different pathways in the terrestrial and marine sub-cycles, and quantify and balance different fluxes to determine both the nature and magnitude of sources and sinks. Thus, credible sinks can be identified only if all sources (e.g., terrestrial including soils and vegetation, industrial, aquatic, biotic) of C are also known. It is in this connection that impact of soil erosion on C dynamics needs to be assessed objectively and quantitatively. Being located in the vicinity of the soil surface and of low density, soil organic carbon (SOC) is drastically impacted by erosional processes. Thus, importance of soil degradation in general and that of soil erosion in particular on C dynamics and possible emissions of GHGs cannot be overemphasized. Redistribution of eroded C over the landscape and its transport into aquatic ecosystems and oceans is an important process (Meybeck, 1993a; Hedges et al., 1997) that cannot be ignored in the context of the global C cycle.

4. Global carbon pools

There are five principal global C pools. The oceanic pool is the largest at 38,000 Pg (Pg = petagram = 10^{15} g = 1 billion metric ton), followed by the geologic pool at 5000 Pg comprising 4000 Pg of coal and 500 Pg each of oil and gas. The geologic pool as reported herein does not include carbonates in sedimentary rocks. The terrestrial C pool is the third largest pool comprising soil and the vegetation com-

Table 4
Anthropogenic increase in atmospheric concentration of several greenhouse gases (adapted from IPCC, 2001)

Gas	Change in concentration from pre-industrial to 1999	Increase in concentration		Radiative forcing (W/m^2)
		%	Current rate	
CO_2	280–367 ppm	31	1.5 ppm/year or 0.4%/year	1.46
CH_4	700–1760 ppb	151	7.0 ppb/year or 0.4/year	0.48
N_2O	270–314 ppm	16	0.8 ppb/year or 0.25%/year	0.15
CRC and others	0–268 ppt	?	− 1.4 ppt/year or − 0.5/year	0.34

ponents. The soil C pool comprises two components: the SOC pool estimated at 1550 Pg and the soil inorganic carbon (SIC) pool estimated at 750 Pg (Batjes, 1996; Eswaran et al., 1995), with a total soil C pool of about 2300 Pg to 1-m depth. The vegetation pool is estimated at 560 Pg, making the terrestrial C pool at about 2860 Pg. The atmospheric pool is 760 Pg and is increasing at the rate of 3.2 Pg C/year. Thus, the soil C pool of 2300 Pg is about 4.1 times the biotic/vegetational pool and about 3 times the atmospheric pool. In comparison, the terrestrial C pool of 2860 Pg is about 57% of the geologic pool and about 4 times the atmospheric pool. All these pools are interconnected. For example, 60 Pg C is exchanged in each direction between biota/vegetation and the atmosphere each year. Similarly, 90 Pg is emitted and 92 Pg is absorbed by the ocean each year (Schimel, 1995; Raich and Schlesinger, 1992). In comparison, only 6.3 Pg/year is emitted by fossil fuel combustion and 1.6–2.0 Pg/year by land use change (Table 5). Thus, enhancing photosynthetic fixation and sequestering even 5% of the photosynthetic C into the terrestrial ecosystems can drastically offset the industrial emissions.

Additional issues at hand are: which of the four pools is contributing to the increase of CO_2 concentration in the atmosphere, and which pool is a potential sink of atmospheric CO_2? Fossil fuel combustion or the geologic pool is an obvious and quantifiable source. Another obvious but not easily quantifiable source is tropical deforestation and the attendant biomass burning. Yet emission of CO_2 and other GHGs by soil degradation is an important but neither an obvious nor an easily quantifiable source. Each year, world soils release about 4% of their pool (60 Pg) into the atmosphere, which is 10 times the fossil fuel combustion. The exact magnitude of the loss is not known, and may in fact be much greater because of anthropogenic perturbations, and soil and ecosystem degradation. On the other hand, the so-called "missing C" may also be absorbed by soils and other terrestrial ecosystems. These issues need to be resolved.

5. Soil erosion and the global C cycle

Gaseous emissions from terrestrial ecosystems are exacerbated by soil degradation, of which accelerated soil erosion is the most predominant and widespread form. The economic and agronomic impacts of accelerated erosion are widely recognized (Pimentel et al., 1995; Lal, 1998; den Biggelaar et al., 2001). There has also been considerable research on the impact of erosion on water quality and nonpoint source pollution (Burwell et al., 1975; Owens et al., 1984; Sharpley et al., 1983; Zhu et al., 1989; Owens, 1994). Yet the research information is extremely scarce on soil C dynamics in relation to soil degradation in general and soil erosion in particular.

The steady state equilibrium of the SOC pool in undisturbed ecosystems is drastically disturbed by anthropogenic activities attendant upon conversion of natural to agricultural ecosystems. These activities often lead to a reduction in the amount of root and litter biomass returned to the soil, distribution in cycles of elements and water and change in the energy balance. Consequently, the SOC pool is diminished by an increase in the oxidation rate and losses due to erosion and leaching. The magnitude of depletion is exacerbated by ploughing, drainage and other drastic disturbances. Most soils may lose one-half to two-thirds of their SOC pool within 5 years in the tropics and 50 years in temperate regions. The new equilibrium may be attained after losing 20–50 Mg C/ha. There are several estimates of the historic loss of SOC pool at global scale due to land use change. The global loss has been estimated at 40 Pg by Houghton (1995), 55 Pg by IPCC (1995) and Schimel (1995), 66–90 Pg by Lal (1999), 150 Pg by Bohn (1978), 500 Pg by Wallace (1994) and 537 Pg by Buringh (1984). Rozanov et al. (1993) observed that world soils have lost humus (58% C) at a rate of 25.3 Tg/year ever since agriculture began 10,000 years ago, 300 million tons/year in the past 300 years and 760 million tons per year in the last 50 years. They estimated that nearly 16% of the original SOC pool may have been lost.

Soil erosion by water and wind, from detachment of particles to eventual deposition in depressional sites and aquatic/oceanic ecosystems, is a four-stage process. It involves (i) detachment of particles, (ii) breakdown of aggregates, (iii) transport and redistribution of sediments over the landscape and (iv) deposition in protected/depressional sites or aquatic ecosystems. Soil C is influenced during each of these four stages, often differently because of wide variations in soil moisture and temperature regimes, differences in quality of soil organic matter content (e.g., C/N ratio, lignin content), variation in soil properties (e.g., degree and strength of aggregation, clay content and mineralogy, differences in SOC concentration and relative extent of labile, intermediate and passive fractions), and changes in hydrodynamics at different landscape positions.

Soil erosion affects SOC dynamics through its impact on the following processes: (i) slaking or disruption of aggregates, (ii) preferential removal of C in runoff water or dust storms, (iii) mineralization of soil organic matter on-site, (iv) mineralization of SOC displaced and redistribution over the landscape and transported in rivers and dust storms, (v) reaggregation of soil through formation of organo-mineral complexes at the depositional/protected sites and (vi) deep burial of C-enriched sediments in depositional sites, flood plains and reservoirs and ocean floor (Fig. 1). A brief description of each of these processes follows.

5.1. On-site depletion of soil organic carbon pool

Land use change from natural to agricultural ecosystems, ploughing and other soil disturbances deplete SOC pool on-site (De Jong et al., 1983; Janzen et al., 1997; Lal, 2000).

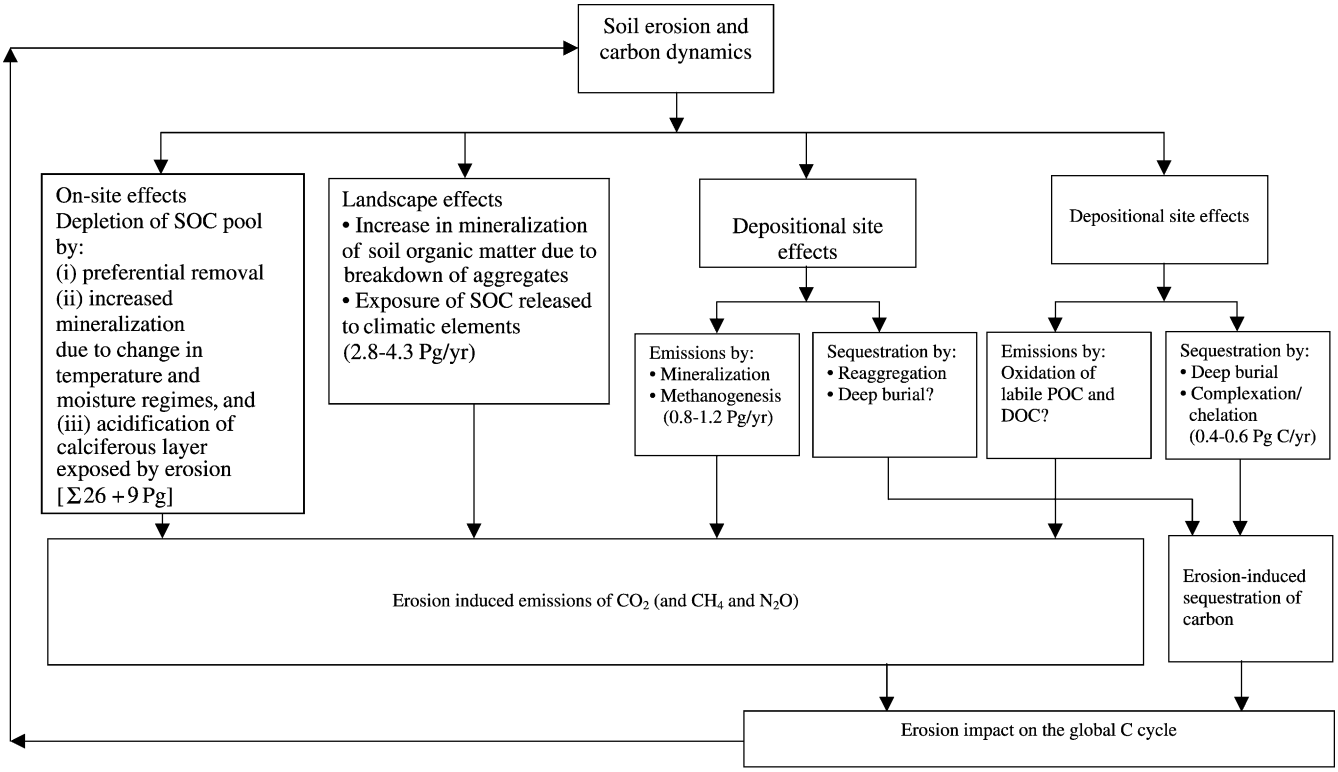


Fig. 1. Erosion impact on processes that impact the global C cycle.

The change in SOC due to change in land use can be represented by a mass-balance equation (Eq. 1):

$$\Delta\mathrm{SOC} = (\mathrm{SOC_a} + A) - (E + L + M) \qquad (1)$$

Where ΔSOC is the change in pool, $\mathrm{SOC_a}$ is the antecedent pool, A is accretion or input of C through biosolids, and losses due to erosion (E), leaching (L) and mineralization (M) are accentuated by anthropogenic perturbations. The process leads to increase in SOC pool if the magnitude of the terms $(\mathrm{SOC_a}+A) > (E+L+M)$, or profile truncation leads to high exposure of a buried A horizon with SOC concentration and depletion in SOC pool if $(\mathrm{SOC_a}+A) < (E+L+M)$.

Conversion of natural to managed ecosystems depletes SOC pool because C inputs in agricultural ecosystems are generally lower than in natural ecosystems (Voroney et al., 1981) and decomposition rates are higher (Gregorich and Anderson, 1985) due to differences in soil moisture and temperature regimes caused by ploughing, drainage, biomass burning or residue removal (Gregorich et al., 1998; Lal, 2000). The loss of SOC may be 20–50 Mg C/ha, which may occur within 5–50 years following conversion (Davidson and Ackerman, 1993), with the exact magnitude of loss depending on climate, soil, drainage, crop sequences, tillage methods and residue management (Lal, 2000). The magnitude of depletion is exacerbated by soil erosion and the attendant decline in soil quality.

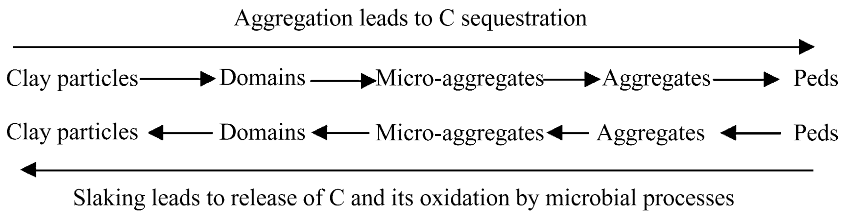
Translocation and redistribution of sediments and SOC by erosional processes is a pedogenic process, because eroded and depositional soils differ from the original soils (Blume, 1986; Beyer et al., 1993). The SOC is easily transported by runoff water or wind because it is of relatively low density (<1.8 Mg/m^3) and is concentrated in the vicinity of the soil surface. A study on wind erosion in southwest Niger showed that wind-borne material trapped at 2-m height contained 32 times more SOC relative to the antecedent topsoil (Sterk et al., 1996). Wind-borne sediments can differ in chemical characteristics compared to soil from which these are derived (Warn and Cox, 1951; Laprade, 1957; Cihacek et al., 1993). A high SOC enrichment ratio is also observed in water-borne sediments (Lal, 1976). In general, the magnitude of SOC transport in runoff water is more from bare fallow plots than those with vegetal cover (Lal, 1976; Lowrance and Williams, 1988). Soil erosion by water and wind depletes the SOC pool, and losses of SOC due to erosion on sloping land may be several times more than due to mineralization (Daniel and Langham, 1936; Slater and Carleton, 1938; Webber, 1964), but not so on flat terrain without wind or water erosion hazard soon after conversion from natural to agricultural ecosystems (Gregorich and Anderson, 1985). Consequently, the SOC pool in eroded soils is much lower than that on uneroded phases (Rhoton and Tyler, 1990; Fahnstock et al., 1996; Oskarsson et al., in press). On a chernozemic soil in Saskatchewan, Canada, Anderson et al. (1986) reported depletion of SOC pool by 70% on convex eroded slopes compared to 40% depletion on slightly eroded sites. De Jong and Kachanoski (1988) reported that about 50% of the total SOC loss occurred due to soil erosion. Pennock et al.

(1994) also reported high losses due to net soil export by erosion. In Iceland, Oskarsson et al. (in press) reported that 120–500 Tg SOC may have been lost due to accumulative historic soil erosion. The current rate of loss of SOC by erosion is estimated at 0.052–0.104 Tg C/year. Soils without erosional losses may attain stable level of SOC, while SOC content in those susceptible to erosion may continue to decrease over a long time (Voroney et al., 1981; Gregorich et al., 1998). Tiessen et al. (1982) observed that SOC losses due to erosion on a conventionally ploughed soil were evident for up to 90 years.

Over and above the lateral transport of light fraction, there are three other factors which exacerbate the depletion of SOC pool. One, the rate of on-site mineralization of the remaining SOC may increase due to changes in soil moisture and temperature regimes. For example, soil reflectance properties are influenced by erosion (Wagner-Ridle et al., 1996) which alter soil temperature regime. Two, erosion decreases soil productivity by reducing available water capacity, decreasing effective rooting depth, and reducing water and nutrient use efficiencies (Lal, 1998). Decline in productivity reduces the amount of plant residues (both shoot and root) returnable to the soil. Significant reductions in crop yields also occur in depressional sites (Monreal et al., 1997) with progressive decrease in amount of root and shoot biomass returned to the soil. Three, severe soil erosion by water or wind also leads to truncation of the soil profile by removal of the surface layer and exposure of the carbonate-rich sub-soil. Carbonates thus exposed may react with acidiferous material and release CO_2 to the atmosphere. In contrast, burial of the carbonaceous layer at depositional sites may reduce the risks of emission from the SIC pool.

5.2. Slaking and disruption of aggregates

Being a work function, erosion causes slaking, disruption and breakdown of aggregates. The latter may happen by slaking caused by a quick immersion in water, disruption following the escape of compressed or entrapped air, impact of the raindrop related to its kinetic energy or momentum, shearing force of runoff or wind, or collision of aggregates against one another. Whereas the process of aggregation sequesters C, breakdown or dispersion of aggregates into soil separates releases hitherto encapsulated C and makes it vulnerable to decomposition by microbial processes. Based on the hierarchy model (Tisdall and Oades, 1982; Oades and Waters, 1991; Tisdall, 1996), processes of aggregation and slaking may be represented by the following schematics. In addition to exposure to microbial processes, the C thus released by slaking is also preferentially removed by water runoff or wind, redistributed over the landscape or is carried into depositional/protected sites.



5.3. Redistribution of sediments and carbon over the landscape

It is generally postulated that SOC transported with eroded sediment may be: (i) deposited into nearby intact soil, (ii) deposited into previously eroded and barren areas, (iii) carried into the fresh water bodies and (iv) oxidized and released into the atmosphere (Oskarsson et al., in press). Indeed, some of the SOC displaced is merely redistributed over the landscape, and may never reach a stream, reservoir or an aquatic ecosystem (Gregorich and Anderson, 1985). The distance to which particles are transported depends on the density/weight of the particles and velocity or carrying capacity of the fluid (wind or water). The light fractions (density of < 1.8 Mg/m^3) are carried longer distances than heavier particles and may enrich the depositional sites. The redistribution of SOC is enhanced over an eroded landscape, and the SOC thus translocated is highly susceptible to microbial processes and climatic elements. The net effect of the erosional process is to deplete the SOC pool on eroded sites and enhance it on depositional sites.

5.4. Mineralization of soil organic carbon redistributed over the landscape

The redistributed SOC is generally a light or labile fraction comprising particulate organic carbon (POC). This fraction is easily mineralized (Schlesinger, 1995), and the process is accentuated by changes in soil temperature and moisture regimes. Excessive runoff leads to the landscape becoming more arid, decreasing soil wetness and increasing soil temperature. The rate of mineralization increases with increase in soil temperature.

5.5. Fate of carbon in depressional sites

There are a series of complex and interacting processes involved in depressional sites: (i) deposition may bury the calciferous layer enriched in SIC content, (ii) deep burial of SOC/POC may decrease risks of mineralization, (iii) reaggregation of the dispersed clay and silt with POC/SOC transported into the depressional sites may also decrease mineralization (Gregorich et al., 1998) and (iv) the depres-

Table 6
Carbon contents of sediments in major rivers

River	Country	Carbon type	Concentration/flux	Reference
Amazon	Brazil	POC	1–2% of bed load (0.5–5.0 mg/l)	Richey et al. (1986, 1991)
		DOC	50% of TOC (2.7–4.7 mg/l)	
		DIC	485–1228 μmol/l	
		TSS	93–600 mg/l	
		TOC	30.9 Tg C/year (62% DOC) or 3.6–12.6 g/m^2/year (8.5 g/m^2/year)	
Mackenzie	Canada	POC	1.8 Tg C/year	Telang et al. (1991)
		DOC	1.3 Tg C/year	
		TOC	3.1 Tg C/year	
		TC	8.0 Tg C/year	
St. Lawrence	Canada	POC	0.3 Tg C/year	Telang et al. (1991)
		DOC	1.6 Tg C/year	
		TOC	1.9 Tg C/year	
		TC	8.9 Tg C/year	
Mississippi	USA	POC	0.9 Tg C/year	Telang et al. (1991)
		DOC	3.5 Tg C/year	
		TOC	3.6 Tg C/year	
Yukan	Mexico	POC	0.4 Tg C/year	Telang et al. (1991)
		DOC	2.4 Tg C/year	
		TOC	3.0 Tg C/year	
Orinoco	Venezuela	POC	1.9 Tg C/year	Depetris and Paolini (1991)
		DOC	3.2 Tg C/year	
		TOC	5.1 Tg C/year	
		TC	7.7 Tg C/year	
Zaire	Congo	DOC	1.02 Tg C/year	Martins and Probst (1991)
		POC	0.30 Tg C/year	
		TOC	1.30 Tg C/year	
Niger	Nigeria	DOC	48 Tg C/year	Martins and Probst (1991)
		POC	60 Tg C/year	
		TOC	108 Tg C/year	
Orange	South Africa	DOC	3 Tg C/year	Martins and Probst (1991)
		POC	1 Tg C/year	
		TOC	4 Tg C/year	
Indus	Pakistan	DOC	8.5 ppm	Subramanian and Ittekkot (1991)
		POC	2.2%	
Ganges	India/Bangladesh	DOC	2.8 ppm	Subramanian and Ittekkot (1991)
		POC	2.3%	
Brahmaputra	Bangladesh	DOC	2.5 ppm	Subramanian and Ittekkot (1991)
		POC	1.0 ppm	

sional sites may also have a high rate of mineralization (Bajracharya et al., 2000) because of a high fraction of the microbial biomass carbon (Beyer et al., 1993). Under anaerobic conditions, the SOC in depressional sites may also undergo methanogenesis (and release CH_4 with global warming potential of 21) and denitrification (and release N_2O and NO_x with global warming potential of 310).

5.6. Fate of carbon transported into rivers, coastal ecosystems and oceans

Sedimentologists argue that the so-called “missing sink” or “fugitive carbon dioxide” (Tans et al., 1990) is sequestered in aquatic ecosystems (Stallard, 1998; Smith et al., 2001). These claims of C sequestration are based on two

Table 7
Transport of carbon in world rivers (adapted from Degens et al., 1991)

Continent	DOC (Tg C/year)	POC (Tg C/year)	TOC (Tg C/year)	DIC (Tg C/year)	TC (Tg C/year)
South America	44.2	24.1	66.9	58.9	125.8
North America	33.8	14.6	41.8	120.9	162.7
Africa	24.7	8.3	33.1	9.1	42.2
Asia	94.0	128.1	169.1	158.1	327.2
Europe	–	–	23.6	0.4	24.0
World total (excluding Australia)	–	–	334.5	407.4	741.9

Table 8
In situ oxidation of labile POC in rivers and coastal regions

River	% oxidizable or labile C in the POC	Reference
Amazon	0	Richey et al. (1991)
Parana	Decomposition and remineralization in the flood plains	Depetris and Paolini (1991)
Global	30–75% of POC in rivers	Ittekkot and Arain (1986); Laane et al. (1987)
Global	20% in the tropics and 40% in temperate regions of POC in rivers and coastal	Ittekot and Laane (1991)

assumptions: (i) there is no mineralization of C en route from eroded sites to aquatic ecosystems, and (ii) all terrestrial C transported into aquatic ecosystems is buried with sediments. However, the fact remains that C being transported over the landscape and eventually carried into the marine ecosystems is subject to numerous processes, and its fate depends on climate, soil/sediment properties and nature of the carbon.

6. Flux of organic carbon in world rivers

The river water contains three types of C: (i) particulate organic carbon (POC) whose concentration decreases with increase in sediment load, (ii) dissolved organic carbon (DOC) whose concentration depends on photosynthetic activity in the riverine ecosystem and (iii) dissolved inorganic carbon (DIC). The combined POC and DOC is called total organic carbon (TOC). Meybeck (1982) reported most rivers carry TOC content of 1.6–6% of the suspended load, and that sediment transport in rivers is about 20 mg/l, and 2–3% of it may be POC (Meybeck, 1982). Miltner and Emeis (2000) estimated that concentration of TOC in sediments was 13 mg/g. Estimates of transport of TOC by rivers to the oceans range widely from 0.19 Pg/year by Kempe (1989), 0.03 Pg/year by Williams (1971) and Schlesinger and Melack (1981), 0.28 Pg/year by SCOPE (Degens et al., 1985), 0.33 Pg/year (Degens et al., 1991), 0.53 Pg/year (Stallard, 1998), 0.57 Pg/year (Lal, 1995) and 1.0 Pg/year by Richey et al. (1980). The data in Table 6 show transport of different forms of C for some rivers of the world. Transport of C by the world's rivers on continental scale is detailed by Degens et al. (1991) and is summarized in Table 7 and may exceed 0.75 Pg/year.

Table 9
Estimates of transport of terrestrial organic carbon to the oceans by world rivers

Reference	Organic carbon transport (Tg C/year)
Schlünz and Schneider (2000)	430
Smith et al. (2001)	400
Lal (1995)	570
Meybeck (1982)	378 (179 is POC)
Lal (2002) present study	400–600

Similar to the data on transport of sediments, there are also numerous uncertainties about the quantity of TOC transported to the coastal ecosystems and oceans. In most cases, the critical background information is lacking, and most published data are tentative at best (Degens et al., 1991). There are some important episodic events (floods

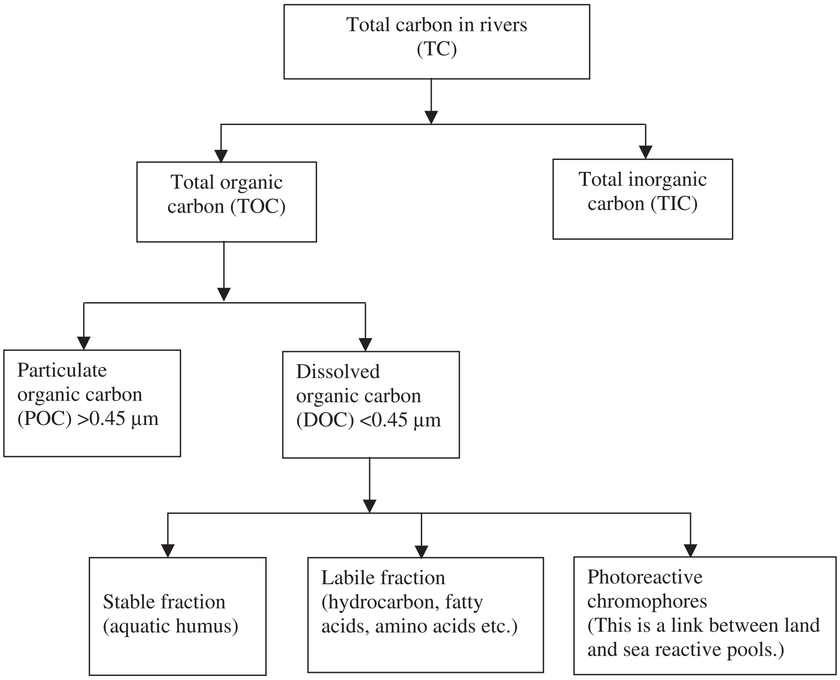


Fig. 2. Components of total C pool in rivers.

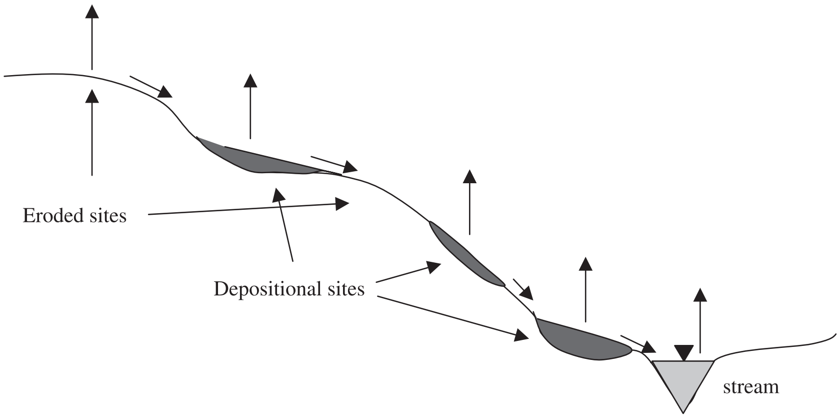


Fig. 3. Transport, redistribution and deposition of soil organic carbon on an eroded landscape.

with long return periods) which are often not monitored, and monitoring procedures used are not standardized and uniform across different gauging stations.

In addition to the uncertainties about the quantity, the fate of TOC transported in rivers is also speculative. Rivers may be a source or sink of atmospheric CO_2 depending on the P^{CO_2}, which in turn depends on the concentration of mineralizable C and temperature. In addition to CO_2 evolved within the river ecosystem, terrestrial soil CO_2 may also be prevalent during some seasons. In most cases, P^{CO_2} rises above the ambient due to prevailing respiration with attendant emission of CO_2 into the atmosphere, and making rivers as a source of atmospheric CO_2. Respiration is high when POC concentration and prevailing temperatures are high (e.g., summer and spring). The partial pressure of CO_2 in rivers can be as much as 2400 ppmv compared to 370 ppmv in the atmosphere, and there exists a strong correlation between P^{CO_2} and DOC concentration (Martins and Probst, 1991). The P^{CO_2} in some rivers may be 20 times that in the atmosphere (Degens et al., 1991). The reverse of respiration or photosynthesis is also an important factor in riverine ecosystems which determines the concentration of DOC. Rivers receiving large volumes of agricultural and urban effluent may have a high concentration of DOC. The global DOC transport may be as much as 0.3 Pg C/year. The data in Table 8 show estimates of the oxidizable C in some rivers of the world, and the value range from 0% to 75%. While the exact value may differ among climates and watershed characteristics, it is apparent that some of the C transported into rivers is oxidized and emitted into the atmosphere as CO_2.

Organic matter in rivers is highly diverse and covers a spectrum of molecular scale to large mass of detritus or animal remains. Components of total C pool are outlined in Fig. 2 and have been described by Thurman (1985), Zafiriou (1985) and Spitzy and Leenheer (1991). The labile fraction is easily mineralized within the riverine environment and is responsible for a higher partial pressure of CO_2 than ambient atmosphere. As much as 75% of the POC in some rivers may be mineralizable (Table 8). The median DOC concentration in world rivers is 5 mg/l with a range of 2–20 mg/l. The DOC is mostly leached out of the soil profile and is carried into the aquatic ecosystems. The POC concentration is between 1% and 10% of the total suspended sediments (TSS). In addition, black C (charcoal) may also be a major component of riverine C exported to the ocean. The annual flux of black C from the Mississippi River is estimated at 5×10^{-4} Pg, 27% of which was derived from fossil fuel combustion (Mitra et al., 2002).

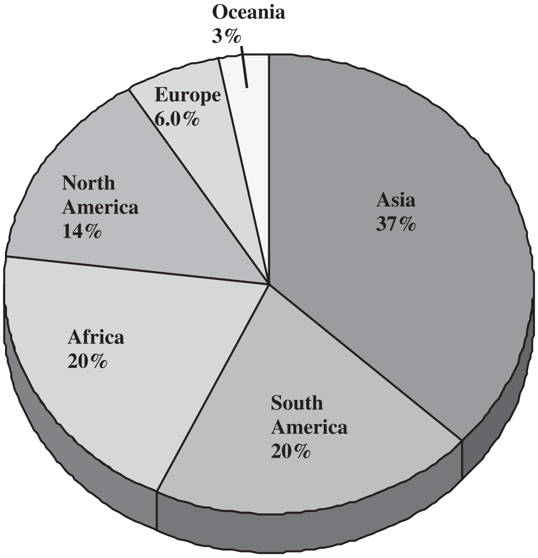
Then, there is a question of SOC transported and eventually buried in marine sediments. On a global scale, the organic carbon content in coastal ocean sediments is not solely of terrestrial origin but a mixture of nearly equal amounts of marine and terrestrial organic carbon (Schlünz and Schneider, 2000). In addition, estimates of C transported into the ocean also vary widely, <0.5 Pg C/year by

Table 10
Fraction of the soil organic carbon oxidized during the erosional process

Fraction oxidized (%)	Reference
0	Smith et al. (2001)
20	Lal (1995)
25–30	Jacinthe and Lal (2001)
50	Oskarsson et al. (in press)
70	Beyer et al. (1993)
100	Schlesinger (1995)

Table 11
Estimate of erosion-induced carbon emission on continental basis

Continent	Gross erosion ($\times 10^9$ Mg/year)	SOC displaced by erosion (2–3% of the sediment)	Emission (20% of the SOC displaced) (Pg C/year)
Africa	38.9	0.8–1.2	0.16–0.24
Asia	74.0	1.5–2.2	0.30–0.44
South America	39.4	0.8–1.2	0.16–0.24
North America	28.1	0.6–0.8	0.12–0.16
Europe	13.1	0.2–0.4	0.04–0.08
Oceania	7.6	0.1–0.2	0.02–0.04
Total	201.1	4.0–6.0	0.8–1.2



Total emission = 0.8-1.2 Pg C/yr

Fig. 4. Relative emission of erosion-induced carbon on continental basis. Erosion management and restoration of eroded soils is to be high priority in Asia, South America and Africa.

Ritchie (1989), Dean and Gorham (1998) and Mullholland and Elwood (1982); 0.57 Pg C/year by Lal (1995); 0.1 Pg C/year by Schlünz and Schneider (2000); 0.334 Pg C/year by Degens et al. (1991); 0.368 Pg C/year by Meybeck (1993b); 0.38 Pg C/year by Ludwig et al. (1996); 0.6–1.5 Pg C/year by Stallard (1998); 0.4 Pg C/year by Smith et al. (2001) (Table 9). Lyons et al. (2002) estimated the POC flux from high-standing islands in the southwest Pacific to be 48 Tg C/year. Schlüter et al. (2000) argue that 2–4% of the organic C produced in the photic zone may be transported to the sea floor in the northern North Atlantic. This may, however, be an overestimate at the global scale. In comparison, Meybeck (1982) reported that the TOC transport by rivers is 1–2% of the net primary production of the terrestrial ecosystem for any kind of environment.

The fate of C transported into the ocean and coastal ecosystems is also not known. Some of the organic C in these ecosystems may be labile, and others recalcitrant. The charcoal C transported in rivers may be an important part of the recalcitrant component (Mitra et al., 2002). Schlünz and Schneider (2000) estimated that a major part of the terrestrial C that enters the oceans by rivers (0.4 Pg) may be oxidized, and some of it may be dispersed throughout the oceans and accumulates in pelagic sediments. The present study estimates that C transported as marine sediments at 0.4–0.6 Pg C/year (Table 9); however, the fate of this C is not known. In fact, terrestrially derived C is not a strong component of the seawater or marine sediments (Meyer-Schultz and Hedges, 1986; Opshal and Benner, 1997). Therefore, a large proportion of terrestrially derived C may have been mineralized (Durrieu de Madron et al., 2000; Goñi et al., 1998). In addition to the nature of the organic material, the degree of decomposition also depends on climatic factors (Romankevich et al., 1999; Raymond and Bauer, 2001).

7. The fate of carbon translocated by erosion

The fate of C translocated by erosional processes is determined by a series of complex and interacting processes. It is difficult to model and generalize the impact of all processes involved. Being a dynamic process, it is difficult to assess the impact through a “snap-shot” approach. A simplified approach to assess the SOC budget over a watershed is shown in Eq. 2.

$$\mathrm{SOC_f} = (\mathrm{SOC_a}) - (O_\mathrm{i} + O_\mathrm{l} + O_\mathrm{d} + O_\mathrm{r}) + (D_\mathrm{b} + B_\mathrm{a}) \qquad (2)$$

Where $\mathrm{SOC_f}$ is final SOC pool in the watershed, $\mathrm{SOC_a}$ is the antecedent pool, O_i is oxidation in situ, O_l is oxidation of the

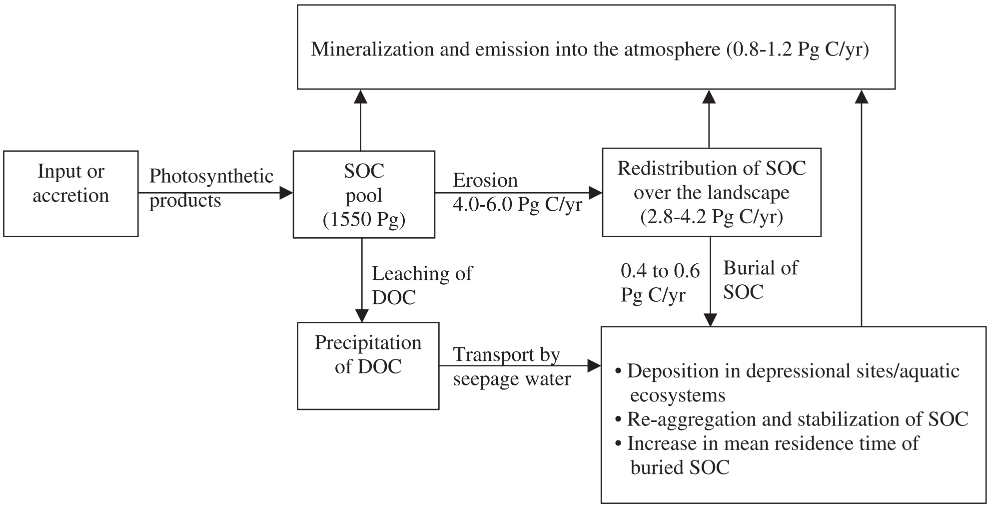


Fig. 5. Erosion-induced changes in SOC budget at the global scale (DOC is dissolved organic carbon).

displaced C over the landscape, O_d is oxidation (and methanogenesis) in depositional sites, O_r is oxidation in river and water bodies, D_b is deep burial in protected and depressional sites and B_a is transport and eventual burial into the aquatic ecosystems.

A principal question is "what fraction of the erosion-transported SOC is oxidized" en route to deep burial with aquatic or marine sediments. Published estimates of oxidizable fraction range from 0% to 100%. Schlesinger (1995) argues that SOC transported during erosion is largely oxidized. In contrast, Smith et al. (2001) assumed that the oxidation flux of C in eroded material is insignificant. Oskarsson et al. (in press) estimated that about 50% of the SOC eroded is mineralized. The fact remains that SOC and riverine C are subject to mineralization at all stages of the transport (Fig. 3). The magnitude of oxidizable material depends on the stability and relative proportion of the humic substances being eroded (labile, intermediate, passive or recalcitrant) (Table 10).

Assuming delivery ratio of 10% and sediment load of 20 billion Mg (Table 2), and SOC content of 2–3%, total SOC displaced by erosion is 4.0–6.0 Pg C (Table 11). Assuming 20% is emitted into the atmosphere (as O_i, O_e, O_d and O_r in Eq. 2 above), erosion-induced emission is 0.8–1.2 Pg C/year. Relative emissions at the continental scale are outlined in Fig. 4. Evidently, the largest contribution from Asia (37%) and the lowest from Oceania (3%) are proportional to the severity of soil erosion in the Asian continent (Fig. 4).

A schematic of the erosion-induced changes in SOC at the global scale are outlined in Fig. 5. Of the 4.0–6.0 Pg C/year translocated by water erosion, 2.8–4.2 Pg C/year is redistributed over the landscape and transferred to depressional sites, 0.4–0.6 Pg C/year is transported into the ocean and may be mineralized or buried with marine sediments and coastal ecosystems and 0.8–1.2 Pg C/year is emitted into the atmosphere. Lal (1995) estimated that 1.14 Pg C/year is emitted into the atmosphere by water erosion.

References

Anderson DW, de Jong E, Verity GE, Gregorich EG. The effects of cultivation on the organic matter of soils of the Canadian prairies. Trans XIII Cong Int Soc Soil Sci Hamburg, vol. 7. Hamburg: International Society of Soil Science; 1986. p. 1344–5.

Bajracharya R, Lal R, Kimble J. Erosion phase effects on CO_2 concentration and CO_2 flux from an Alfisol. Soil Sci Soc Am J 2000;64: 694–700.

Barford CC, Wofsy SC, Goulden ML, Munger JW, Pyle EH, Urbanski SP, et al. Factors controlling long- and short-term sequestration of atmospheric CO_2 in a mid-latitude forest. Science 2001;294:688–1691.

Batjes NH. The total C and N in soils of the world. Eur J Soil Sci 1996; 47:151–63.

Bennett HH. Soil conservation. New York: McGraw-Hill; 1939. 993 pp.

Beyer L, Köbbemann C, Finnern J, Elsner D, Schluß W. Colluvisols under cultivation in Schleswig–Holstein: 1. Genesis, definition and geo-ecological significance. J Plant Nutr Soil Sci 1993;156:197–202.

Blume HP, editor. Guidebook: soils and landscapes in schleswig–holstein. Mitt Dtsch Bodenkdl Ges, vol. 51, 1986. 132 pp.

Bohn H. On organic soil C and CO_2. Tellus 1978;30:472–5.

Brown LR. Eroding the base of civilization. J Soil Water Conserv 1981;36:255–60.

Brown LR. The global loss of topsoil. J Soil Water Conserv 1984;39: 162–5.

Buringh P. Organic carbon in soils of the world. In: Woodwell GM, editor. The role of terrestrial vegetation in the global carbon cycle. SCOPE, vol. 23. Chichester: Wiley; 1984. p. 41–109.

Burwell RE, Timmons DR, Holt RF. Nutrient transport in surface runoff as influenced by soil cover and seasonal periods. Soil Sci Soc Am Proc 1975;39:523–8.

den Biggelaar C, Lal R, Wiebe K, Breneman V. Impact of soil erosion on crop yields in North America. Adv Agron 2001;72:1–52.

Cihacek LJ, Sweeney MD, Deibert EJ. Characterization of wind-erosion sediments in the Red River Valley of North Dakota. J Environ Qual 1993;22:305–10.

Daniel HA, Langham WH. The effect of wind erosion and cultivation on the total nitrogen and organic matter content of the soil in the Southern High Plains. J Am Soc Agron 1936;28:587–96.

Davidson EA, Ackerman IL. Changes in soil carbon inventories following cultivation of previously untilled soils. Biogeochemistry 1993;20: 161–93.

Dean WE, Gorham E. Magnitude and significance of carbon burial in lakes, reservoirs and peatlands. Geology 1998;26:535–8.

Degens ET, Kempe S, Herrera R, editors. Transport of carbon and minerals in major world rivers. Pt. 3. Mitt Geol—Paläontol Inst Univ Hamburg, SCOPE/UNEP Sonderbd, vol. 48, 1985. 645 pp.

Degens ET, Kempe S, Richey JE. Summary: biogeochemistry of major world rivers. In: Deggens ET, Kempe S, Richey JE, editors. Biogeochemisty of major world rivers. Chichester: Wiley; 1991. p. 323–47.

De Jong E, Kachanoski RG. The importance of erosion in the carbon balance of prairie soils. Can J Soil Sci 1988;68:111–9.

De Jong E, Begg CBM, Kachanoski RG. Estimates of soil erosion and deposition for some Saskatchewan soils. Can J Soil Sci 1983;63: 607–17.

Depetris PJ, Paolini JE. Biological aspects of South American rivers: the Parana and the Orinoco. In: Degens ET, Kempe S, Richey JE, editors. Biogeochemistry of major world rivers. SCOPE, vol. 42. Chichester: Wiley; 1991. p. 105–25.

Dregne HE. Historical perspective of accelerated erosion and effect on world civilization. Determinants of soil loss tolerance. ASA Spec Publ, vol. 45. 1982. p. 1–14. Madison, WI.

Durrieu de Madron X, Abassi A, Heussner S, Monaco A, Aloisi JC, Radakovitch O, et al. Particulate matter and organic carbon budget for the Gulf of Lions (NW Mediterranean). Oceanol. Acta 2000;23:717–30.

Eckholm EP. Losing ground. New York: Norton; 1976.

Eswaran H, Van den Berg E, Reich P, Kimble J. Global soil carbon resources. In: Lal R, Kimble JM, Levine E, Stewart BA, editors. Soils and global change. Boca Raton: CRC/Lewis; 1995. p. 27–43.

Fahnstock P, Lal R, Hall GF. Land use and erosional effects on two Ohio Alfisols. I. Soil properties. J Sust Agric 1996;7:85–100.

Fournier F. Climat et erosion. Paris: Presses Universitaires de France; 1960.

Goñi MA, Ruttenberg KC, Eglinton TI. A reassessment of the sources and importance of land-derived organic matter in surface sediments from the Gulf of Mexico. Geochim Cosmochim Acta 1998;62:3055–75.

Gregorich EG, Anderson DW. The effects of cultivation and erosion on soils of four toposequences in the Canadian prairies. Geoderma 1985; 36:343–54.

Gregorich EG, Greer KJ, Anderson DW, Liang BC. Carbon distribution and losses: erosion and deposition effects. Soil Tillage Res 1998;47: 291–302.

Hedges JI, Keil RG, Benner R. What happens to terrestrial organic matter in the ocean? Org Geochem 1997;27:195–212.

Houghton RA. Changes in the storage of industrial carbon since 1850. In: Lal R, Kimble JM, Levine E, Stewart BA, editors. Soils and global change. Boca Raton: CRC/Lewis; 1995. p. 45–65.

Hurtt GC, Pacala SW, Moorcroft PR, Caspersen J, Shevliakova E, Houghton RA, et al. Projecting the future of the U.S. carbon sink. Proc Natl Acad Sci U S A 2002;99:1389–94.

IPCC. Climate change 1995. Working group 1. IPCC. Cambridge: Cambridge Univ Press; 1995.

IPCC. Climate change 2001: the scientific basis. Intergovernmental panel on climate change. UK: Cambridge Univ Press; 2001.

Ittekkot V, Arain R. Nature of particulate organic matter in the river Indus, Pakistan. Geochim Cosmochim Acta 1986;50:1643–53.

Ittekkot V, Laane RWPM. Fate of riverine particulate organic matter. In: Deggens ET, Kempe S, Richey JE, editors. Biogeochemisty of major world rivers. Chichester: Wiley; 1991. p. 233–43.

Jacinthe P, Lal R. A mass balance approach to assess carbon dioxide evolution during erosional events. Land Degrad Dev 2001;12:329–39.

Jacks GV, Whyte RO. The rape of the earth. New York: Arno Press; 1939. 139 pp.

Jansson MB. Land erosion by water in different climates. UNGI report #57. Dept. Physical Geography 57, Uppsala Univ; 1982.

Jansson MB. A global survey of sediment yield. Geogr Ann Ser A 1988;70:81–98.

Janzen HH, Campbell CA, Gregorich EG, Ellert BH. Soil carbon dynamics in Canadian agroecosystems. In: Lal R, Kimble J, Follett R, editors. Soil processes and the carbon cycle. Adv Soil Sci, Boca Raton, FL: Lewis Publishers, CRC Press; 1997.

Kempe S. Reservoir case study: Lake Nasser. In: Meybeck M, Chapman D, Helmer R, editors. Global freshwater quality, a first assessment. GEMS/WMO/UNEP. Oxford: Blackwell; 1989. p. 243–52.

Laane RWPM, Etcheber H, Relexans JC. Particulate organic matter in estuaries and its ecological implications for macrobenthos. In: Deggens ET, Kempe S, Weibin G, editors. Transport of C and minerals in major world rivers. Pt. 4. Mitt Geol—Paläontol Inst Univ Hamburg, SCOPE/UNEP, Sonderbd, vol. 64. 1987. p. 71–91.

Lal R. Soil erosion problems on Alfisols in western Nigeria and their control. IITA Monogr 1976;1. Ibadan, Nigeria, 208 pp.

Lal R. Global soil erosion by water and carbon dynamics. In: Lal R, Kimble JM, Levine E, Stewart BA, editors. Soils and global change. Boca Raton: CRC/Lewis Publishers; 1995. p. 131–41.

Lal R. Soil erosion impact on agronomic productivity and environment quality. Crit Rev Plant Sci 1998;17:319–464.

Lal R. Soil management and restoration for C sequestration to mitigate the accelerated greenhouse effect. Prog Environ Sci 1999;1:307–26.

Lal R. World cropland soils as a source or sink for atmospheric C. Adv Agron 2000;71:145–91.

Lal R. World cropland soils as source or sink for atmospheric carbon. Adv Agron 2001;71:145–91.

Laprade KE. Dust-storm sediments of Lubbock area, Texas. AAPG Bull 1957;41:709–26.

Lowdermilk WC. Conquest of the land through seven thousand years. SCS Agric Inform Bull, vol. 99, 1953. Washington, DC, 30 pp.

Lowrance R, Williams RG. Carbon movement in runoff and erosion under simulated rainfall conditions. Soil Sci Soc Am J 1988;52:1445–8.

Ludwig W, Probst JL, Kempe S. Predicting the oceanic input of organic carbon by continental erosion. Global Biogeochem. Cycles 1996;10:23–41.

Lvovich MI, Karasik GY, Bratseva NL, Medvedeva GP, Maleshko AV. Contemporary intensity of the world land intercontinentl erosion. Moscow: USSR Academy of Sciences; 1991.

Lyons WB, Nezat CA, Carey AE, Hicks DM. Organic carbon fluxes to the ocean from high-standing islands. Geology 2002;30:443–6.

Martins O, Probst J-L. Biogeochemistry of major African rivers. In: Deggens ET, Kempe S, Richey JE, editors. Biogeochemisty of major world rivers. Chichester, UK: Wiley; 1991. p. 127–55.

Meybeck M. Carbon, nitrogen and phosphorus transport by world rivers. Am J Sci 1982;282:401–50.

Meybeck M. Riverine transport of atmospheric carbon: sources, global typology and budget. Water Air Soil Pollut 1993a;70:443–63.

Meybeck M. C, N, P and S in rivers: from sources to global inputs. In: Wollast R, Mackenzie FT, Chou L, editors. Interactions of C, N, P and S: biogeochemical cycles and global change. Berlin: Springer; 1993b. p. 163–93.

Meyer-Schultz KJ, Hedges JI. Molecular evidence for a terrestrial component of organic matter dissolved in ocean water. Nature 1986;321:61–3.

Milliman JD, Meade R. Worldwide delivery of river sediment to the oceans. J Geol 1983;91:1–21.

Milliman JD, Syvitski JPM. Geomorphic/tectonic control of sediment discharge to the ocean: the importance of small mountainous rivers. J Geol 1992;100:325–44.

Miltner A, Emeis K-C. Terrestrial organic matter in surface sediments of the Baltic Sea, northwest Europe, as determined by CUO oxidation. Geochim Cosmochim Acta 2000;65:1285–99.

Mitra S, Bianchi TS, McKee BA, Stutula M. Black carbon from the Mississippi River: quantities, sources and potential implications for the global C cycle. Environ Sci Technol 2002;36:2296–302.

Monreal CM, Zentner RP, Robinon JA. An analyses of soil organic matter dynamics in relation to management, erosion and yield of wheat in long-term crop rotation plots. Can J Soil Sci 1997;77:553–63.

Mullholland PJ, Elwood JW. The role of lake and reservoir sediments as sinks in the perturbed global cycle. Tellus 1982;34:490–9.

Myneni RB, Dong J, Tucker CJ, Kaufmann RK, Kauppi PE, Liski J, et al. A large carbon ink in the woody biomass of northern forests. Proc Natl Acad Sci U S A 2001a;98:14784–9.

Myneni RB, Dong J, Tucker CJ, Kaufmann RK, Kauppi PE, Liski J, et al. A large carbon sink in the woody biomass of northern forests. Proc Natl Acad Sci U S A 2001b;10.

Oades JM, Waters AG. Aggregate hierarchy in soils. Aust J Soil Res 1991;29:815–28.

Oldeman LR. The global extent of soil degradation. In: Greenland DJ, Szabolcs I, editors. Soil resilience and sustainable land use. Wallingford: CAB International; 1994. p. 99–118.

Oldeman LR. Soil degradation: a threat to food security. Wageningen, Holland: ISRIC; 1998.

Olson GW. Archaeology: lessons on future soil use. J Soil Water Conserv 1981;36:261–4.

Opshal S, Benner R. Distribution and cycling of terrigenous dissolved organic matter in the ocean. Nature 1997;386:482–6.

Oskarsson H, Arnalds O, Gudmundsson J, Gudbergsson G. Organic carbon in Icelandic Andosols: geographical variation and impact of erosion. Catena [in press].

Owens LB. Impact of soil N management on the quality of surface and subsurface water. In: Lal R, Stewart BA, editors. Soil processes and water quality. Boca Raton: CRC/Lewis Publishers; 1994. p. 137–62.

Owens LB, Edwards WM, Van Keuren RF. Peak nitrate nitrogen values in surface runoff from fertilized pastures. J Environ Qual 1984;13:310–2.

Pacala SW, Hurtt GC, Baker D, et al. Consistent land- and atmosphere-based U.S. carbon sink estimates. Science 2002;292:2316–23.

Pennock DJ, Anderson DW, de Jong E. Landscape-scale changes in indicators of soil quality due to cultivation in Saskatchewan, Canada. Geoderma 1994;64:1–19.

Pimentel D, Harvey C, Resosudarmo P, Sinclair K, Kurz D, McNair M, et al. Environmental and economic costs of soil erosion and conservation benefits. Science 1995;267:1117–23.

Raich JW, Schlesinger WH. The global carbon dioxide flux in soil respiration and its relationship to vegetation and climate. Tellus Ser B 1992;44:81–99.

Raymond PA, Bauer JE. Riverine export of aged terrestrial organic matter to the North Atlantic Ocean. Nature 2001;409:497–500.

Rhoton FE, Tyler DD. Erosion-induced changes in the properties of a Fragipan soil. Soil Sci Soc Am J 1990;54:223–8.

Richey JE, Brock JT, Naiman RJ, Wissman RC, Stallard RF. Organic carbon: oxidation and transport in the Amazon River. Science 1980;207:1348–50.

Richey JE, Meade RH, Salati E, Devol AH, Nordin CF, dos Santos U.

Water discharge and suspended sediment concentrations in the Amazon River: 1982–1984. Water Resour Res 1986;22:756–64.

Richey JE, Victoria RL, Salati E, Forsberg BR. The biogeochemistry of a major river system: the Amazon case study. In: Deggens ET, Kempe S, Richey JE, editors. Biogeochemisty of major world rivers. Chichester, UK: Wiley; 1991. p. 57–73.

Ritchie JC. Carbon content of sediments of small reservoirs. Water Resour Bull 1989;25:301–8.

Romankevich EA, Vetrov AA, Vinogradov ME, Vedernikov VI. Carbon cycling components in the Arctic Seas of Russia: carbon fluxes from land, carbon in the bottom sediments, elements of balance. Oceanology 1999;40:335–44.

Rozanov BG, Targulian V, Orlov DS. Soils. In: Turner II BL, Clark WC, Kates RW, Richards JF, Mathews JT, Meyer WB, editors. The earth as transformed by humans action: global and regional changes in the biosphere over the past 300 years. Cambridge: Cambridge Univ Press; 1993. p. 203–14.

Scherr S. Soil degradation: a threat to developing country food security by 2020. IFRI Food, Agric. and Environment Discussion Paper 27, Washington, DC; 1999. 63 pp.

Scherr SJ, Yadav S. Land degradation in the developing world: implications for food, agriculture and the environment to 2020. IFPRI, Food, Agric. and the Environment Discussion Paper 14, Washington, DC; 1996. 36 pp.

Schimel DS. Terrestrial ecosystems and the carbon cycle. Global Change Biol 1995;1:77–91.

Schimel DS, House JI, Hibbard KA, et al. Recent patterns and mechanisms of carbon exchange by terrestrial ecosystems. Nature 2001;414: 169–72.

Schlesinger WH. Soil respiration and changes in soil carbon stocks. In: Woodwell GM, Mackenzie GM, editors. Biotic feedbacks in the global climatic system: will the warming feed the warming? New York: Oxford Univ. Press; 1995. p. 159–68.

Schlesinger WH, Melack JM. Transport of organic carbon in world's rivers. Tellus 1981;33:172–87.

Schlünz B, Schneider RR. Transport of terrestrial organic carbon to the oceans by rivers: re-estimating flux and burial rates. Int J Earth Sci 2000;88:599–606.

Schlüter M, Sauter EJ, Schäfer A. Spatial budget of organic carbon flux to the seafloor of the northern North Atlantic (60°N–80°N). Global Biogeochem Cycles 2000;14:329–40.

Sharpley AN, Syers JK, Tillman RW. Transport of ammonium and nitrate N in surface runoff from pasture as influenced by urea application. Water Air Soil Pollut 1983;20:425–30.

Slater CS, Carleton EA. The effect of erosion on losses of soil organic matter. Soil Sci Soc Am Proc 1938;3:123–8.

Smith SV, Renwick WH, Buddemeier RW, Crossland CJ. Budgets of soil erosion and deposition for sediments and sedimentary organic carbon across the conterminous United States. Global Biogeochem Cycles 2001;15:697–707.

Spitzy A, Leenheer J. Dissolved organic carbon in rivers. In: Degens ET, Kempe S, Richey JE, editors. Biogeochemistry of major rivers. SCOPE vol. 42. Chichester: Wiley; 1991. p. 214–32.

Stallard RF. Terrestrial sedimentation and carbon cycle: coupling weathering and erosion to carbon burial. Global Biogeochem Cycles 1998; 12:231–57.

Sterk G, Herrmann L, Bation A. Wind blown nutrient transport and soil productivity changes in southwest Niger. Land Degrad Dev 1996;7: 325–35.

Strakhov NM. Principles of lithogenesis, vol. 1. Edinburgh, (UK): Oliver and Boyd; 1967.

Subramanian V, Ittekkot V. Carbon transport by the Himalayan rivers. In: Deggens ET, Kempe S, Richey JE, editors. Biogeochemisty of major world rivers. Chichester: Wiley; 1991. p. 157–68.

Tans PP, Fung IY, Takahashi T. Observational constraints on the global atmospheric CO_2 budget. Science 1990;247:1431–8.

Telang SA, Pocklington R, Naidu AS, Romankevich EA, Gitelson II, Gladyshev MI. Carbon and mineral transport in major North American, Russian, Arctic, and Siberian rivers. In: Deggens ET, Kempe S, Richey JE, editors. Biogeochemistry of major world rivers. Chichester: Wiley; 1991. p. 75–104.

Thurman EM. Organic geochemistry of natural waters. Boston (MA): Martinus Nijhoff/Dr W Junk; 1985. 497 pp.

Tiessen H, Stewart JWB, Betany JR. Cultivation effects on the amount and concentration of carbon, nitrogen and phosphorus in grassland soils. Agron J 1982;74:831–4.

Tisdall JM. Formation of soil aggregates and accumulation of soil organic matter. In: Carter MR, Stewart BA, editors. Structure and organic matter storage in agricultural soils. Boca Raton: CRC/Lewis Publishers; 1996. p. 57–96.

Tisdall JM, Oades JM. Organic matter and water stable aggregates in soil. J Soil Sci 1982;33:141–63.

UNCED, Agenda 21: programme of action for sustainable development, rio declaration on environment and development, statement of principles. Final text of agreement negotiated by governments at the United Nations Conference on Environment and Development (UNCED), 3–14 June 1992, Rio de Janeiro, Brazil, UNDP, New York.

UNEP, Sands of change: why land becomes desert and what can be done about it. UNEP Brief #2, United Nations Environment Programme, Nairobi, Kenya; 1986. 8 pp.

UNEP. World atlas of desertification. Nairobi, (Kenya): Edward Arnald Seven Oaks, UNEP; 1992.

UNFCCC. United Nations framework convention on climate change. Bonn, Germany: UNFCC; 1992.

UNFCCC. Kyoto Protocol to the United Nations Framework Convention on Climate Change. Bonn, Germany: UNFCC; 1997.

UNFCD. United Nations framework convention to combat desertification. Bonn, Germany; 1996.

Voroney RP, van Veen JA, Paul EA. Organic C dynamics in grassland soils: 2. Model validation and simulation of the long-term effects of cultivation and rainfall erosion. Can J Soil Sci 1981;61:211–24.

Wagner-Ridle C, Gillepsi TJ, Swanton CJ. Rye mulch characterization for the purpose of micro-climatic modelling. Agric For Meteorol 1996;78: 67–81.

Wallace A. Soil organic matter must be restored to near original levels. Commun Soil Sci Plant Anal 1994;25:29–35.

Walling DE. The sediment delivery problem. J Hydrol 1983;65:209–37.

Walling DE. Erosion and sediment yield in a changing environment. In: Branson J, Brown AG, Gregory KJ, editors. Global continental changes: the context of paleohydrology. Geol Soc Spec Publ, vol. 115. 1996. p. 43–56. London, UK.

Walling DE. The response of sediment yield to environmental change. In: Walling DE, Probst J-L, editors. Human impact on erosion and sedimentation. IAHS Publ, vol. 245. 1997. p. 77–89. Wallingford, UK.

Walling DE, Webb BW. Patterns of sediment yield. In: Gregory KJ, editor. Background to paleohydrology. Chichester: Wiley; 1983. p. 69–100.

Walling DE, Webb BW. Estimating the discharge of contaminants to coastal waters by rivers: some cautionary comments. Mar Pollut Bull 1985;16:488–92.

Walling DE, Webb BW. Material transport by the world's rivers: evolving perspectives. In: Rodda JC, Matales NC, editors. Water for the future: hydrology in perspective. Proc Rome Symp, April 1987. IAHS Publ, vol. 164. 1987. p. 313–29.

Walling DE, Webb BW. The reliability of rating curve estimates of suspended sediment yield: some further comments. In: Bordas MP, Walling DE, editors. Proc. Porto Alegra Symp, December 1988. IAHS Publ, vol. 174. 1988. p. 337–50.

Walling DE, Webb BW. Erosion and sediment yield: a global overview. Erosion and sediment yield: global and regional perspectives. Proc. Exeter Symp, July 1996. IAHS Publ, vol. 236. 1996. p. 3–19.

Warn GF, Cox WH. A sedimentary study of dust storm in the vicinity of Lubbock, TX. Am J Sci 1951;249:553–68.

Webber LR. Soil physical properties and erosion control. J Soil Water Conserv 1964;19:28–30.

Williams PM. The distribution and cycling of organic matter in the ocean. In: Faust SD, Hunter JW, editors. Organic Compounds in Aquatic Environment. New York: Marcel Dekker; 1971. p. 145–63.

Zafiriou OC. Atmospheric, oceanic and interfacial photochemistry as factors influencing air–sea exchange fluxes and processes. In: Buat-Menard P, editor. The role of air sea exchange in geochemical cycling. NATO ASI Ser, Ser C, Math Phys Sci, vol. 185. Dordrecht: D. Reidel; 1985. p. 185–207.

Zhu JC, Gantzer CJ, Anderson SH, Alberts EE, Beuselinck PR. Runoff, soil and dissolved nutrient losses from no-till soybean with winter cover crops. Soil Sci Soc Am J 1989;53:1210–4.