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## **The Significance of the Erosion-induced Terrestrial Carbon Sink**

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Estimating carbon (C) balance in erosional and depositional landscapes is complicated by the effects of soil redistribution on both net primary productivity (NPP) and decomposition. Recent studies are contradictory as to whether soil erosion does or does not constitute a C sink. Here we clarify the conceptual basis for how erosion can constitute a C sink. Specifically, the criterion for an erosional C sink is that dynamic replacement of eroded C, and reduced decomposition rates in depositional sites, must together more than compensate for erosional losses. This criterion is in fact met in many erosional settings, and thus erosion and deposition can make a net positive contribution to C sequestration. We show that, in a cultivated Mississippi watershed and a coastal California watershed, the magnitude of the erosion-induced C sink is likely to be on the order of 1% of NPP and 16% of eroded C. Although soil erosion has serious environmental impacts, the annual erosion-induced C sink offsets up to 10% of the global fossil fuel emissions of carbon dioxide for 2005.

**Keywords:** erosion, deposition, carbon sequestration, soil organic carbon deposition, soil organic carbon stabilization

Recently there has been an increased interest in the ability of soils to affect atmospheric concentrations of carbon dioxide (CO<sub>2</sub>; Schlesinger 1999, Sarmiento and Gruber 2002, Bellamy et al. 2005), because the soil and climate systems are closely coupled through the exchange of carbon (C) among the atmosphere, biosphere, and pedosphere (Houghton and Woodwell 1989, Keeling et al. 1996). The soil system is the third largest reservoir of C, next only to the lithosphere and oceans. Globally, soils store approximately 2400 petagrams (1 Pg = 10<sup>15</sup> grams [g]) soil organic carbon (SOC) in the top 2 meters (m) (Kirschbaum 2000). Part of this SOC is annually redistributed across landscapes by soil erosion and deposition. Whether the combined effect of SOC redistribution and associated changes to ecosystem productivity result in a net C sink for or source to atmospheric CO<sub>2</sub> remains unresolved. Resolution of this inconsistency will have significant implications for soil scientists, ecologists, and policymakers.

The objectives of this paper are to (a) clarify the conceptual basis for why and how erosion can constitute a C sink and (b) argue that protection of marginal lands can have significant implications for C sequestration. We provide pedologically and ecologically sound explanations for how erosion and deposition can constitute a C sink. Although inorganic C in soils (soil carbonate) is also subject to erosion, the scope of this paper is limited to SOC.

### **Motivation**

Land degradation—defined as the reduction or loss of land resource potential as a result of human activities, including deforestation, biomass burning, cultivation, and accelerated soil erosion (Blaikie and Brookfield 1987)—has a significant influence on the global C budget. The net amount of CO<sub>2</sub> released from the biosphere to the atmosphere as a result of land-use change over time is likely to be equivalent to about 75% of total fossil fuel C emissions. It is estimated that since the Industrial Revolution, land conversion and degradation have caused up to 200 Pg C that was originally in the biosphere to be released to the atmosphere (DeFries et al. 1999).

The possibility of erosion-induced C sequestration has received widespread interest from the scientific community and policymakers for three reasons. First, erosion is among the most pressing environmental problems facing the world today. Accelerated erosion by water and wind is responsible for one-half and one-quarter of all soil degradation, respectively (ISRIC and UNEP 1990, Daily 1995, Pimentel et al. 1995). Persistently high rates of soil erosion affect more than  $1.1 \times 10^9$  hectares of land annually (Jacinthe and Lal 2001, Berc et al. 2003), redistributing on the order of 75 Pg soil per year, with sediment transport leading to silting of reservoirs and eutrophication of lakes. Soil erosion from agricultural lands alone, which accounts for two-thirds of the total soil loss, has been estimated to result in more than US\$400 billion of damages annually (Pimentel et al. 1995). Second, projected changes in climate are expected to stimulate the hydrologic cycle, increasing the intensity, amount, and seasonality of precipitation in many parts of the world, and thus accelerating soil erosion (CGER 1999, Berc et al. 2003). Third, soil erosion is the only way that otherwise stable, mineral-associated SOC can be relocated in large quantities and its decomposition rate enhanced during transport or reduced after transport (Starr et al. 2000, Lyons et al. 2002).

## **Recent scientific developments and disagreements**

In 1998, R. F. Stallard used linked hydrologic–biogeochemical models to assess the contribution of soil erosion and terrestrial sedimentation to the global C cycle. Stallard concluded that anthropogenic acceleration of soil erosion and terrestrial sedimentation may result in ecosystem disequilibria (between net primary productivity [NPP] and decomposition at the watershed scale), with the unexpected benefit of promoting C sequestration (Stallard 1998). Stallard's novel attempt at linking the seemingly unrelated, complex phenomena of soil erosion and the global C cycle has since received additional support from more detailed empirical studies (Harden et al. 1999, Smith et al. 2001, McCarty and Ritchie 2002, Liu et al. 2003, Berhe et al. 2005, Yoo et al. 2005, Berhe 2006).

However, along with this support, there has been disagreement in the scientific literature over the relationship between soil erosion and C sequestration. Some studies have concluded that erosion can constitute a net C sink, while others state that it represents a source of atmospheric CO<sub>2</sub> (Lal 1995, 2003a, Post et al. 2004). Stallard's (1998) work demonstrated that the more than 70% eroded SOC deposited in different basins can constitute human-induced burial of up to 1.5 Pg C per year. Smith and colleagues (2001) similarly showed that accounting for the amount of soil C eroded and deposited terrestrially in C budgets can increase previous estimates of soil C sequestration in the United States by up to 47%. For the sake of simplicity, if one assumes comparable rates of erosion and deposition globally, the estimation of Smith and colleagues (2001) is equivalent to a global net sequestration of 1 Pg C per year from erosion. Harden and colleagues (1999) also showed that it is possible for erosion to constitute a C sink if NPP is maintained in eroding slopes (e.g., with fertilization), while the rate of decomposition in low-lying depositional basins is reduced by approximately 20% relative to upland sites. This study demonstrated that an equally plausible scenario exists for erosion to constitute a very large and unaccounted C source if the eroded C is not protected from decomposition. A follow-up modeling study by Liu and colleagues (2003) demonstrated that erosion can decrease CO<sub>2</sub> emission from dynamic landscapes by replacing surface soil with subsurface material that has low bulk SOC content and a higher recalcitrant SOC fraction. More recently, Yoo and colleagues (2005), combining a hillslope sediment transport model with empirical soil C measurements in an undisturbed watershed, found that downhill accumulation of SOC transported by bioturbation, along with burial of in situ photosynthate at the depositional sites, could constitute a sink of 1.9 g C per m<sup>2</sup> per year. Van Oost and colleagues (2005) similarly demonstrated that tillage erosion can result in the sequestration of 3 to 10 g C per m<sup>2</sup> per year on sloping arable lands in Denmark and the United Kingdom.

Although the preceding studies found erosion to be a C sink in some locales and regions, the work of Lal and coworkers (Lal 1995, 2001, 2003a, 2003b, 2003c, Bajracharya et al. 2000, Jacinthe and Lal 2001, Jacinthe et al. 2001, Starr et al. 2001, Lal et al. 2004) suggests that soil erosion constitutes a source term in the global C budget. These studies conclude that erosion is currently releasing up to 1.14 Pg C per year to the atmosphere, as a result of aggregate breakdown by the energy of rain splash and the shearing forces of runoff (Starr et al. 2000). This conclusion is supported by Schlesinger (1990, 1995), who estimated that almost 100% of eroded C is decomposed during detachment and transport, leaving little or no opportunity for burial and protection of eroded C. Most past modeling efforts based on this latter concept have implicitly assumed that all eroded C is either deposited in the ocean or rapidly oxidized, and hence that its contribution to terrestrial C sequestration is negligible (Schlesinger 1990, DeFries et al. 1999, Houghton et al. 1999). Some of the SOC deposited in the ocean is also likely to be sequestered, but this article focuses on the contribution of soil erosion and deposition to terrestrial C sequestration. Clearly, this issue is far from straightforward.

## **Processes and controls**

Because of the dynamic nature of NPP, decomposition, C stabilization, erosion, and deposition, several complicating factors need to be considered when defining or quantifying an erosion-induced C sink.

## **Erosional redistribution of soil and associated soil organic carbon**

Soil erosion is traditionally conceived as a three-step process involving the detachment, transport, and deposition of soil particles. Detachment exposes SOC that is physically protected within aggregates and clay domains; subsequently, finer soil particles and associated SOC are preferentially transported away from eroding slopes to different low-lying depositional sites (Gregorich et al. 1998, Lal 2001, Starr et al. 2001). Following detachment and transport, burial usually is believed to protect SOC from decomposition, because there generally are enhanced and radiometrically old C stocks in the deep soils of agricultural lowlands and sedimentary basins (Stallard 1998, Harden et al. 1999, 2002). Most of the eroded topsoil (> 70%; Stallard 1998) remains within the adjacent topography and is stored in a variety of depositional basins, including wetlands, peat lands, estuaries, fluvial deltas, terrestrial

depressions (hollows), and reservoirs within the same or adjacent topography. The increased wetness and reduced aeration at the depositional basins (compared with eroding slopes) can slow down decomposition (Stallard 1998, Smith et al. 2001, McCarty and Ritchie 2002).

Stallard (1998), on the basis of past data (Meade et al. 1990) and model simulations, provided three major reasons why soil erosion should not necessarily represent loss of C from the terrestrial biosphere: First, soil redistribution downhill or downstream is usually accompanied by partial replacement of eroded upland C with new photosynthate. Second, a significant portion of eroded, C-rich topsoil is buried in different depositional settings, rather than flowing to the ocean. Erosion transports relatively fresh organic matter that is present at or near the soil surface (compared with deep soil organic matter [SOM]). After successive erosive events, the C- and nutrient-rich topsoil of the eroding slopes is buried in the depositional lowlands and becomes a subsoil horizon of the convergent slopes or plains (figure 1), probably reducing its rate of decomposition (compared with noneroded C on the contributing slopes). Third, the surface area for terrestrial deposition of eroded C has increased since the beginning of the Industrial Revolution (figure 2). The estimated 10- to 100-fold acceleration of erosion rates by anthropogenic activities in recent history has not been accompanied by a concurrent and proportional increase in sediment discharge to the ocean. The discharge of sediment and C to the ocean has remained approximately constant as a result of hydrologic projects on managed floodplains. Therefore, the recent increase in the rate of soil erosion has led to increased storage of eroded C in different types of depositional basins (Stallard 1998).

Figure 1. Soil and soil organic carbon transport from divergent slopes to convergent or flat depositional basins and erosion-facilitated inversion of a hillslope soil profile.

Figure 2. Global aerial coverage of different types of depositional basins and associated storage of carbon (C) in pre- and postindustrial times (adapted from Stallard 1998). Stallard (1998) assumed that human activities affected only one class of wetlands that were created from floodplains: rice paddies, which were created primarily in areas that were not already occupied by wetlands. The size of wetland area lost since industrialization is about the same as the size of the area where new paddies were created.

### **Erosion and watershed carbon balance**

Soil erosion results in drastic modifications to the structure as well as the biological and chemical properties of the soil matrix, affecting its productive capacity and ability to sequester atmospheric CO<sub>2</sub>. Erosion affects watershed-level C balance by changing the magnitude of opposing C fluxes of (a) C input rates and (b) decomposition and stabilization.

**Carbon input rates.** Generally, unless the soil is eroded beyond a critical level, NPP on eroding slopes continues, albeit at a reduced rate if nutrients or water becomes limiting (Onstad et al. 1984). The newly assimilated C at eroded sites replaces, at least partially, C that was transported by erosion. As demonstrated by Harden and colleagues (1999), this dynamic replacement of eroded SOC is an important variable in maintaining the watershed-level C balance. This is especially important if NPP could be enhanced in eroding slopes with the use of supplements or best management practices, such as fertilization, irrigation, crop rotation, and reduced tillage. In the depositional part of a watershed, the C input is derived not only from fresh plant residue growing in situ but also from deposition of laterally flowing, eroded C. The rate of NPP in depositional basins is likely to be high, because the deposited topsoil provides additional organic matter, essential nutrients, and water-holding capacity.

**Decomposition and stabilization.** Soil erosion and deposition can speed or slow the decomposition of SOC at different parts of a watershed. At eroding slope positions, erosion can increase the rate of decomposition by breaking down aggregates (because of rain intensity or shearing during transport) and exposing organic matter that was previously encapsulated and physically protected from microbial and enzymatic degradation. On the other hand, removal of topsoil material from the eroded site exposes subsoil material, typically with less C content than topsoil, and therefore lowers the rate of decomposition. During transport, however, the decomposition of upland SOC can be enhanced, since the eroding material has the potential for further disturbance. For example, in arable lands, if transport rates are slow enough, eroded SOM can be decomposed through the breakdown of aggregates by tillage. Therefore, conceptually, the net impact on the CO<sub>2</sub> budget depends on the residence times of both the sediment and C (Harden et al. 1999). The extent to which soil erosion results in net enhancement of the SOC decay rate is still being debated. Previous estimates of the SOC fraction that is oxidized during erosion range from 0 to 100% (Beyer et al. 1993, Lal 1995, Schlesinger 1995, Jacinthe and Lal 2001, Smith et al. 2001, Oskarsson et al. 2004).

At depositional settings, the rate of decomposition of eroded SOC can be reduced by a combination of processes. Some of these processes are biochemical (recalcitrance of organic constituents), others physical (protection with

burial, aggregation, and changing water, air, and temperature conditions), and still others chemical (mineral–organic matter associations).

Regardless of the rate of SOM oxidation, detachment and transport of soil particles modify the biochemical makeup of the SOC that reaches the depositional basins. During transport, the labile SOC fraction decomposes quickly, leaving behind a larger fraction of relatively more recalcitrant SOC, compared with the SOC that originates from the eroding hillslope profiles. In addition, inevitable losses (due to, e.g., leaching and mineralization) further reduce the amount and, moreover, change the chemical recalcitrance of the deposited SOC after it arrives at the depositional settings. During intensive storm events, however, large loads of sediment can be moved from upper slopes directly to lower slope positions and streams. Indeed, it is possible that most of the stream sediment is moved during such events. With such rapid transport, it is likely that eroded C has little chance to be decomposed and reworked during transport, and that a significant fraction of labile C can enter depositional basins. In this scenario, eroded C remaining near the surface of lowlands could contribute to enhanced decomposition, while the decomposition rate of the eroded C that is buried at the depositional settings is likely to be reduced.

The role of burial during sedimentation is key to the sink-versus-source question for eroded C. Decomposition is generally accepted to be slower in the buried sediments of depositional basins than in the source profiles in the eroding slopes. This is partly because deposition of eroded C down-slope is often accompanied by increased water content, reduced oxygen availability, compaction, and physical protection within inter- or intra-aggregate spaces that collectively can retard the decomposition rate of buried SOC. Indeed, SOC may be preserved and have much longer residence times in anoxic or suboxic floodplains, riparian ecosystems, reservoirs, or peat lands, compared with aerobic soils in upper watershed positions. Postdeposition (diagenetic) remobilization and transformations also are reduced in wetter depositional basins, favoring SOC preservation over mineralization (Callender and Smith 1993, Gregorich et al. 1998, Stallard 1998, Harden et al. 1999, McCarty and Ritchie 2002), since anoxic or suboxic conditions reduce the rate at which soil micro-organisms decompose organic matter (Jacinthe et al. 2001).

Furthermore, burial facilitates chemical and mineralogical transformations that contribute to C stabilization. With time, newly weathered, precipitated, or transported reactive mineral particles come in contact with buried C. These mineral particles provide surface area for the chemical stabilization of buried C, allowing the physically protected, labile SOC to form stable or metastable complexes with the mineral surfaces, thereby further slowing down its turnover. Moreover, during deposition, low-lying native soils are buried by erosion, potentially resulting in a significant reduction of native SOC decomposition (Liu et al. 2003).

Consequently, burial (in most cases) represents a net C sink, because it constitutes transfer of SOC from more active components in plant biomass and topsoil with short mean residence time (typically less than a century) to more passive reservoirs in adjacent depositional basins (Smith et al. 2001), where C is physically protected from near-surface environments (Harden et al. 1999, Jacinthe et al. 2001). In summary, the foregoing discussions indicate that the increased C input and reduced decomposition (stabilization) usually result in increasing the overall C stock in a watershed with erosion and deposition (figure 3).

Figure 3. Effects of erosion on watershed-level carbon (C) balance. Soil erosion and deposition contribute directly to changing C balance (the sum of the opposing fluxes, where positive C balance indicates a C sink and negative C balance indicates a source term in the C balance). As illustrated here, erosion also contributes to changing C balance through its effect on local hydrology and the soil's physical and chemical properties, which in turn exert an indirect effect on C balance.

### **Criterion for an erosion-induced terrestrial carbon sink**

How and when does soil erosion result in a C sink in which C uptake rates outpace C release rates across landscapes? Whether the rates of C uptake are higher than the rates of C loss is determined by several factors: (1) the rate of C input from plant production at the eroded site; (2) the inherent recalcitrance of SOC; (3) the fraction of plant C that enters the SOC pool rather than being mineralized to CO<sub>2</sub> (the humification coefficient, or  $h_e$ ); (4) how much, and to what extent, eroded C is stored in more protected forms at the sites of deposition; and (5) the energy (C) cost of irrigation- and fertilization-supplemented production. In other words, what determines whether soil erosion and deposition constitute a C sink is the disequilibrium created by the replacement and stabilization of C in eroding and depositional parts of the watershed, respectively.

To derive the criterion for erosion to constitute a C sink using instantaneous phenomenological rate variables, consider two contiguous regions of a hillslope, upper eroding (E) and lower deposition (D). Let the pre-erosion condition be represented by PE, PD, kD, kE, and the post-erosion condition by P'E, P'D, k'D, k'E, where the

parameters  $k$  and  $P$  are spatial averages for rates of SOM loss and production (NPP), respectively, within each of those two regions (expressed in the same units, such as kilograms [kg] C per year). Then the criterion for sequestration due to erosion is simply.

This inequality can be satisfied even if plant input to the entire watershed is reduced (i.e.,  $P'E + P'D < PE + PD$ ), provided the decomposition rate is reduced even more (i.e.,  $k'E + k'D \ll kE + kD$ ). According to equation 1, then, the eroding slope can have a C budget that is at steady state, even if some C is being eroded from its topsoil (Stallard 1998). For the purpose of this formulation, a closed watershed must include erosional and depositional sites wherein C dynamics must be generalized for the flux terms, and mass balance must be accounted for at both sites for the same time period, regardless of variations in episodicity. Equation 1 represents an instantaneous, short-term C sequestration and might not adequately represent the long-term, temporally integrated watershed C balance. In addition, estimation of  $k$  can be difficult where there is deposition of eroded C at different parts of the eroding slope or delivery of laterally flowing C to the lower depositional site from multiple points (for example, colluvial deposition and riverine sedimentation on a floodplain).

Note that, in addition to the above conditions, degradation or reduction of a soil's C stock would increase its sequestration potential, since soils are considered to have a certain intrinsic saturation point for storage of organic matter. As can be inferred from the differential model for fractional decomposition and constant input,  $dC/dt = I - kC$ , degraded soils are technically further from their intrinsic C saturation limit (Jenny 1941), and the newly deposited material may contain less SOC than the native soils, thereby creating new capacity for C sequestration in the depositional environments (Liu et al. 2003). Accordingly, compared with nondegraded, C-rich soils, degraded, C-depleted soils have more potential to accumulate C under favorable conditions that facilitate the delivery of organic matter from fresh plant residue.

In conclusion, the criterion that needs to be satisfied for soil erosion to constitute a C sink consists of two components: (1) at least partial replacement of eroded SOC by new photosynthate in the eroded site, and (2) preservation of SOC at the depositional site in more passive pools (with a longer residence time) than at the eroded site. In concert, the two components that make up the criterion must produce or stabilize sufficient SOC to compensate for the erosive loss of SOC from the upper part of the watershed.

### **Magnitude of the erosion-induced sink**

For a given watershed, assuming that NPP is maintained at eroding sites, the amount of C that is stabilized,  $C_s$  (the sink term in units of kg C per m<sup>2</sup> per year), relative to the non-eroding or steady-state system is calculated as

where  $S_e$  is the amount of eroded soil (kg soil per m<sup>2</sup> per year),  $f_{c,s}$  is the fraction of C in the eroded sediments (kg C per kg soil),  $f_{e,d}$  is the fraction of eroded C deposited in local depositional basins, and  $f_{d,o}$  is the fraction of deposited C that is oxidized or decomposed after it reaches the depositional basin. Equation 2 is formulated as such because the amount of eroded C ( $C_e$ , in kg C per year) is a product of  $S_e$  and  $f_{c,s}$ , while  $C_d = C_e \cdot f_{e,d}$  and  $C_s = C_d \cdot f_{d,o}$ . Here,  $C_d$  is the net amount of SOC deposited (kg C per year), not transported into oceans or oxidized during the detachment and transport of eroded soil particles, while  $C_s$  is the amount of C that is stabilized, indicating long-term C storage.

It is estimated that erosion redistributes about 75 Pg soil annually worldwide (Pimentel et al. 1995). Typically,  $f_{c,s}$  ranges between 1.5% and 5% (Lal 2001). Accordingly, 1.1 to 3.7 Pg SOC is redistributed by erosion annually. Moreover, it is generally accepted that 70% to 90% of eroded sediments are deposited within the same or an adjacent watershed (Stallard 1998). For lack of a better estimate and for simplicity, here we use an average value of 0.8 for  $f_{e,d}$ . Thus, the SOC deposited within the same watershed or adjacent watersheds ( $C_d$ ) can be 0.8 to 3.4 Pg. At the depositional site, the fraction of deposited SOC lost to decomposition and oxidation ( $f_{d,o}$ ) is estimated to be between 20% and 50% (Jacinthe et al. 2001). Taking the lowest value for  $f_{d,o}$  of 0.2, it is possible that between 0.4 and 2.8 Pg C can be stabilized in a variety of depositional basins.

This means that, using conservative estimates ( $S_e = 75$  Pg,  $f_{c,s} = 0.015$ ,  $f_{e,d} = 0.8$ , and  $f_{d,o} = 0.2$ ), it is plausible that at least 0.72 Pg C could be stabilized per year. This estimate falls within the range of Stallard's (1998) C sink estimate, 0.6 to 1.5 Pg. The magnitude of the erosion-induced C sink depends largely on how the rates of NPP and soil C input change in response to erosion and the physical setting of the depositional environment—for example, how quickly the deposited C is buried by subsequent deposition, and how moist the soils in the depositional environment are. As long as soil productivity is maintained (for example, by fertilization), an even higher C-sink value is possible with an increase in the aerial extent of depositional basins (figure 2), higher C content of eroded topsoil, and more effective preservation of C in aquatic or saturated depositional basins (for example, lakes and

impoundments versus colluvial footslopes). Moreover, as Liu and colleagues (2003) found, the ratio of erosional to depositional areas affects the magnitude of C sequestration such that, all other things being equal, concentrated deposition within a small area is more effective than redistribution of eroded C in a large depositional area.

Using values typical of different types of depositional basins can give considerably different values for the erosion-induced C sink. For example, Smith and colleagues (2001) use the rate of water and wind erosion, river discharge, and impoundment sedimentation. The C-sink size extrapolated from this work is about 1 Pg C per year. On the other hand, extrapolation of Yoo and others' (2005) data from an undisturbed watershed yields a global C sink estimate of 0.02 Pg C per year caused by soil erosion and deposition. Considering the aerial distribution of different types of depositional basins (figure 2), the actual value is probably closer to the estimates of Smith and colleagues (2001) and Stallard (1998).

### **Clarifying the disagreement**

Given the discussion above, it is perhaps puzzling why the debate is still ongoing. We believe the debate continues because different researchers are addressing two facets of the same issue. Stallard (1998), Smith and colleagues (2001), and Yoo and colleagues (2005) calculated the proportion of transported C that arrives in the depositional basins and the potential for that C to be replaced by plant uptake and assimilation into soil and sediment. In these studies, most of the C in depositional basins is assumed to be protected after burial. In contrast, Lal and colleagues (Lal 1995, 2001, 2003a, 2003b, 2003c, Bajracharya et al. 2000, Jacinthe and Lal 2001, Jacinthe et al. 2001, Starr et al. 2001, Lal et al. 2004) and Schlesinger (1995) describe how most of the eroded C that would have entered the SOC pool of depositional basins decomposes during detachment, because of raindrop impact or transport.

These contradictions are amplified by the lack of information about the transfer of C from plant to soil and from soil to sediment SOC pools. For example, in noneroding cultivated croplands, it is estimated that, at most, 30% of the annual NPP enters the SOC pool ( $h_e = 0.3$ ; Six and Jastrow 2002). This value may be lower for marginal lands, such as eroding slopes and a variety of depositional basins, because of C losses associated with the detachment and transport of soil particles.

The preliminary results of Berhe and colleagues (2005) suggest that less than 10% of the annual NPP in eroding slopes and depositional settings of an undisturbed watershed enters the SOC pool; the rest of the annual aboveground NPP decomposes rather rapidly, never contributing to the SOC inventory of eroding and depositional sites. Furthermore, the nature of the erosion-induced C sink or source can vary over time, depending on the form of erosion and type of depositional basin. Multiple competing processes can also occur simultaneously at different positions of the same toposequence. It is generally believed that depositional basins at or near their moisture saturation points may promote C stabilization more than aerated colluvial deposits. At the same time, however, tillage and exposure of deep soil C to the surface can enhance the decomposition of otherwise stable forms of C. Harden and colleagues (1999) showed that, in highly eroding Mississippi loess soils, the net C balance of the watershed and strength of the CO<sub>2</sub> sink are dictated by how much inherent soil fertility or fertilization contributes to the perpetuation of C input, and how fast eroded C is decomposed and replaced by new photosynthate.

### **Erosion, deposition, and carbon sequestration: Two examples**

Despite differences in estimates in the literature of the erosion-induced C flux term (and whether it is a source or a sink of atmospheric CO<sub>2</sub>), experimental evidence strongly suggests that erosion and deposition constitute a net sink in the C balance of eroding–depositional watersheds, relative to noneroding basins. Here we illustrate the importance of C protection in depositional environments and the fraction of deposited C that is oxidized using two examples from sites with different erosion rates.

The first example is an annual grassland in an undisturbed, zero-order watershed in northern California (Tennessee Valley, Marin County), where erosion is mainly a result of the burrowing activity of pocket gophers and diffusive mass transport (Berhe et al. 2005, Yoo et al. 2005, Berhe 2006). This site has relatively low erosion rates of 0.1 kg soil per m<sup>2</sup> per year. The second example is a cropland in central Mississippi (Nelson Farm) that has been cultivated since 1870 and continues to experience high rates of erosion, 4.7 kg soil per m<sup>2</sup> per year (Harden et al. 1999). In both examples, the amount of annually eroded C is less than the production rate of new photosynthate (NPP) ( $P_e > C_e$ ). Primary data representing the two sites are given in table 1. Ignoring plant production in the depositional environment for a moment, under the given local conditions (table 1), and unless  $f_{d,0} = 1$  (the maximum value, with the entire eroded C oxidizing rapidly after reaching the depositional basin), some positive amount of C is stabilized (i.e., a net positive C sink term; figure 4). Note that considering only C dynamics in the eroding slope positions and

ignoring plant productivity in the depositional basins makes the amount of Cs estimated here an order of magnitude less than the C sink estimated by Yoo and colleagues (2005).

Table 1.

Values for variables used in calculating carbon sequestration in the illustrative examples.

Figure 4.

Amount of carbon (C) stabilized (Cs) as a function of the fraction of deposited C oxidized ( $f_{d,o}$ ) and the fraction of eroded soil deposited ( $f_{e,d}$ ) for (a) Tennessee Valley, California, where rates of C erosion are low, and (b) the Nelson Farm, Mississippi, where erosion rates are high. The circled area represents the most plausible range of C stabilization possible with the above assumptions. Abbreviations: Pg, petagram; yr, year.

Table 2.

### **Magnitude of eroded, deposited, and sequestered carbon.**

However, under conditions in which deposition of eroded SOC promotes enhancement of the decomposition rate in the depositional site,  $f_{d,o}$  would need to be greater than 1 to accurately account for all the loss terms in the C balance model of the depositional site. The net amount of uneroded SOC,  $C_u$  (in kg C per m<sup>2</sup> per year), in the upland soils is the difference between the humified NPP ( $P_e \cdot h_e$ ) and the eroded SOC ( $C_e$ ). Using the plausible values for the fraction of eroded C that is deposited locally ( $f_{e,d} = 0.8$ ) and oxidized in the depositional basins ( $f_{d,o} = 0.2$ ), one can easily compute the fraction of total annual plant production at the eroding site ( $P_e$ ) that is stabilized ( $P_{e,s} = C_s/P_e$ ) and the fraction that remains uneroded ( $P_{e,u} = C_u/P_e$ ). The values of  $P_{e,s}$  and  $P_{e,u}$ , along with  $C_e$ ,  $C_d$ ,  $C_s$ , and  $C_u$ , for the two sample sites are given in table 2.

In this example, the short-term field measurements and the calculations based on them (table 1) are assumed to be representative of the long-term equilibrium conditions at the sites. If the total land area of the eroding slopes (source slopes) and depositional environments (receiving basins) is known, the magnitude of the C sink can be computed by multiplying  $C_s$  by the area. Here,  $C_s$  is not likely to be greater than  $P_e$ , and these formulations are more appropriate for actively eroding sites such as ridgetops or agricultural fields, where most of the eroded C ( $C_e$ ) is derived from NPP—unlike mid-or foot-slope positions, where a significant amount of  $C_e$  can be derived from C that is eroded from upslope positions.

These calculations (table 2) indicate that the highly eroding Mississippi site that has been receiving nutrient supplements stores at least 15 times more C than the naturally eroding California site. However, in both cases, the amount of C that is deposited and stabilized as a result of erosion and deposition is very small compared with the annual C input ( $P_e$ ) or the amount that is eroded ( $C_e$ ). Therefore, the discussions surrounding the potential for erosion to constitute a C sink should move beyond the question of how small a fraction of eroded C is deposited in local depositional basins ( $f_{e,d}$ ) to refocus on ascertaining the magnitude of the fraction of deposited C oxidized or decomposed after it reaches different types of depositional basins ( $f_{d,o}$ ) and its importance to the C balance in eroding and depositional basins.

### **Global ecopolitical significance**

The magnitude of the erosion-induced C-sink term (0.72 Pg C per year) and the dynamics of eroded C once it reaches different types of depositional basins ( $f_{d,o}$ ) are important for two reasons. First, this erosion-induced C sequestration may be equivalent to up to 10% of global fossil-fuel CO<sub>2</sub> emissions for 2005 (IPCC 2007). Second, erosion–deposition dynamics increase the impact of agriculture on atmospheric C exchange by threefold (McCarty and Ritchie 2002).

The erosion–deposition-induced C sink is significant at regional to global scales, but three factors complicate our ability to quantify its actual contribution to C sequestration: (1) the spatial variability in photosynthesis, soil fertility, fertilization (Jacinthe et al. 2002), and hydrological mitigation practices; (2) the spatial variability in decomposition, burial, and physical protection of C; and (3) the challenge of scaling information from local, site-specific studies to regional estimates. Moreover, in many parts of the world, land degradation further affects the input of organic matter to the soil system by contributing to a decrease in the amount of NPP that is returned to the soil as input to the SOC pool, for example, when farmers burn their crop residue as fuel.

### **Opportunity in marginal lands**



Improving agricultural and land-use policies in marginal lands, such as flood-prone agricultural lands, degraded soils, and other eroding landscapes, offers an enormous opportunity for enhancing C sequestration. Proper soil conservation practices that maintain vegetative cover and enhance plant productivity can promote higher SOC input and storage. Because soil C in eroded, marginal lands generally is depleted by a past history of erosion or intensive land use, rotation to minimum tillage or fallow conditions (with a cover of vegetation) is likely to increase the soil's potential to store C. For example, it is estimated that, in some regions, an increase in C storage of 0.2 to 2.2 metric tons per hectare per year may be observed with sustainable soil and water management (McCarty and Ritchie 2002). Realization of this potential would have significant benefits by reducing atmospheric buildup of CO<sub>2</sub>. Moreover, protecting depositional C from oxidation through minimal tillage increases the potential for sequestration.

The dependence of NPP and C sequestration on rates of erosion and deposition for sites with and without conservation measures is shown schematically in figure 5. If we consider eroding and depositional parts of a watershed separately, under a given erosion scenario, as soil erosion increases, NPP decreases; but the C sequestration potential of the soil increases, at least initially, because of the enhanced ability of the degraded upland soils to take up more C compared with undisturbed and undegraded sites (McCarty and Ritchie 2002). Similarly, at the depositional sites in a given scenario (for example, alluvial plain), actual C sequestration follows a pattern similar to what it was at the eroding site, but with a higher rate of sequestration and a smaller decline after the peak, because the depositional sites continue to receive C-rich eroded soil. The added input of nutrient-rich topsoil at the depositional sites contributes to the maintenance of higher NPP.

Figure 5.

Conceptual relationship between soil erosion and deposition, net primary productivity (NPP), and carbon sequestration (CS) under given erosion and deposition scenarios and when an appropriate level of soil conservation is applied, modified from the conceptual relationship McCarty and Ritchie (2002) developed for an eroding site not receiving any conservation measures.

In the erosion and deposition conditions shown in figure 2, proper soil and water conservation measures maintain or increase NPP. If NPP increases at the eroding site, C sequestration in the eroding soils is enhanced and maintained at a higher level, and the depositional sites have the capacity to increase C sequestration until a saturation point is reached. The absence of tillage and other anthropogenic disturbances is critical to achieving the conservation conditions described in figure 5.

On a cautionary note, marginal lands are vulnerable by definition, and C sequestration is not permanent. Most of the C stored in floodplains or reservoirs is protected physically by aggregation or burial, and potentially can be mineralized easily at a more rapid rate than it was accumulated if the depositional basins are disturbed by practices such as dredging or dismantling of impoundments.

### **Policy implications**

Regardless of how well deposition sequesters large amounts of C in some settings, erosion is not a sensible management strategy for C sequestration. The seriousness of the soil erosion problem in different parts of the world is well documented in the literature (e.g., Lal 1995e.g., Lal 2001, 2003a, 2003c, Troeh et al. 2004), including issues such as burning crop residue for fuel, fossil fuel burned to produce fertilizers, and reduction of water inputs to soil because of erosion. Since soil erosion reduces soil quality, its unexpected contribution to C sequestration should not be interpreted as advantageous. Advocating that marginal lands be kept marginal because they are far from C saturation and potentially can remain at their highest sequestration potential makes no ecological sense. Despite massive global efforts to reclaim them, it is still very hard to rehabilitate degraded lands; they stand as testimony to the limits of soil development, in terms of replacing fertility and corresponding C on the timescale of accelerated erosion. Proper management of erosional and depositional landscapes is crucial to C sequestration and to protecting these lands for future use.

In the policy arena, under the provisions of the Kyoto Protocol, accountable emission mitigation measures are limited to provable and accepted changes in C storage resulting from afforestation, reforestation, and deforestation (IPCC 2000). As it is currently formulated, this policy makes it difficult to realize the sequestration potential of protecting marginal lands from disturbance. The contribution of managing marginal lands should be considered, because tillage of C-rich alluvial soils, draining of floodplains for agriculture, and dredging of reservoirs all release large amounts of CO<sub>2</sub> that could otherwise be protected in land preserves.

## Conclusions

Soil erosion and deposition most likely stabilize at least 0.72 Pg C per year globally. At the watershed level, this amounts to between 0.2% and 2.2% of NPP and about 16% of eroded C. Regardless of the small magnitude of C stabilization compared with NPP or with the C erosion rate, partial replacement of eroded C by new photosynthate, and stabilization of deposited C, very likely offsets up to 10% of the 2005 global fossil fuel CO<sub>2</sub> emissions.

Minimizing soil erosion is vital to protecting natural resources, because accelerated erosion reduces soil quality and depletes the soil resource. Similarly, proper management of already degraded, marginal areas could ensure the environmental benefits of C sequestration resulting from burial and partial replacement of eroded SOC. Clearly, this benefit of erosion constituting a C sink is not a reason to relax erosion prevention measures; rather, it is an impetus for enhanced management of marginal (erosional and depositional) lands. Indeed, any such provision must ensure against incentives that might foster soil erosion. Erosion creates significant, formerly uncounted C sinks; thus there is a major need to better understand the C-sink potential of erosion in buried soils, and the processes that slow the turnover of buried C in alluvial and colluvial soils in different regions. The scientific uncertainty surrounding the fate and dynamics of eroded SOC after terrestrial sedimentation (in buried colluvial deposits and in aerated and waterlogged, submerged alluvial deposits), coupled with the potential for active management of these marginal systems, makes this a high-priority research area for global and regional C-cycle studies.

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## References cited

1. Bajracharya RM, Lal R, Kimble JM. 2000. Erosion effects on carbon dioxide concentration and carbon flux from an Ohio Alfisol. *Soil Science Society of America Journal*. 64: 694–700.
2. Bellamy PH, Loveland PJ, Bradley RI, Lark RM, Kirk GJD. 2005. Carbon losses from all soils across England and Wales 1978–2003. *Nature*. 437: 245–248.
3. Berc J. , et al 2003. Conservation Implications of Climate Change: Soil Erosion and Runoff from Croplands: A Report from the Soil and Water Conservation Society. Ankeny (IA): Soil and Water Conservation Society.
4. Berhe AA. 2006. Storage, replacement, stabilization, and destabilization of soil organic carbon in eroding and depositional settings. PhD dissertation University of California. Berkeley.
5. Berhe AA, Harden JW, Harte J, Torn MS. 2005. Soil degradation and global change: Role of soil erosion and deposition in carbon sequestration. (4 February 2007; <http://repositories.cdlib.org/ucias/breslauer/2>).
6. Beyer L, Köbbemann C, Finner J, Elsner D, Schluß W. 1993. Colluvisols under cultivation in Schleswig-Holstein, 1: Genesis, definition and geocological significance. *Journal of Plant Nutrition and Soil Science*. 156: 197–202.
7. Blaikie PM, Brookfield H. 1987. *Land Degradation and Society*. London: Methuen.
8. Callender E, Smith RA. 1993. Deposition of organic carbon in upper Missouri River reservoirs. *Mitteilungen des Geologisch–Paläontologischen Instituts der Universität Hamburg Heft*. 74: 65–79.
9. [CGER] Commission on Geosciences, Environment and Resources. 1999. *Hydrologic Science Priorities for the U.S. Global Change Research Program: An Initial Assessment*. Washington (DC): National Academy Press.
10. Daily GC. 1995. Restoring value to the world's degraded lands. *Science*. 269: 350–354.
11. DeFries RS, Field CB, Fung I, Collatz GJ, Bounoua L. 1999. Combining satellite data and biogeochemical models to estimate global effects of human-induced land cover change on carbon emissions and primary productivity. *Global Biogeochemical Cycles*. 13: 803–815.
12. Gregorich EG, Greer KJ, Anderson DW, Liang BC. 1998. Carbon distribution and losses—erosion and deposition effects. *Soil and Tillage Research*. 47: 291–302.

13. Harden JW, Sharpe JM, Parton WJ, Ojima DS, Fries TL, Huntington TG, Dabney SM. 1999. Dynamic replacement and loss of soil carbon on eroding cropland. *Global Biogeochemical Cycles*. 13: 885–901.
14. Harden JW, Fries TL, Pavich MJ. 2002. Cycling of beryllium and carbon through hillslope soils in Iowa. *Biogeochemistry*. 60: 317–335.
15. Houghton RA, Woodwell GM. 1989. Global climatic change. *Scientific American*. 260: 36–47.
16. Houghton RA, Hackler JL, Lawrence KT. 1999. The U.S. carbon budget: Contributions from land-use change. *Science*. 285: 574–578.
17. [IPCC] Intergovernmental Panel on Climate Change. 2000. *Land Use, Land-Use Change and Forestry*. Cambridge (United Kingdom): IPCC.
18. [IPCC] Intergovernmental Panel on Climate Change. 2007. *Climate Change 2007: The Physical Science Basis. Summary for Policymakers*. Geneva: IPCC. (20 March 2007; [www.ipcc.ch/SPM2feb07.pdf](http://www.ipcc.ch/SPM2feb07.pdf)).
19. [ISRIC and UNEP] International Soil Reference and Information Centre and United Nations Environment Programme. 1990. *Global Assessment of Human-Induced Soil Degradation*. (25 February 2007; [www.isric.org/UK/About+ISRIC/Projects/Track+Record/GLASOD.htm](http://www.isric.org/UK/About+ISRIC/Projects/Track+Record/GLASOD.htm)).
20. Jacinthe PA, Lal R. 2001. A mass balance approach to assess carbon dioxide evolution during erosional events. *Land Degradation and Development*. 12: 329–339.
21. Jacinthe PA, Lal R, Kimble JM. 2001. Assessing water erosion impacts on soil carbon pools and fluxes. Pages 427–450 in Kimble JM, Follett RF, Stewart BA. eds *Assessment Methods for Soil Carbon*. Boca Raton (FL): CRC.
22. Jacinthe PA, Lal R, Kimble JM. 2002. Carbon budget and seasonal carbon dioxide emission from a central Ohio Luvisol as influenced by wheat residue amendment. *Soil and Tillage Research*. 67: 147–157.
23. Jenny H. 1941. *Factors of Soil Formation*. New York: McGraw-Hill.
24. Keeling RF, Piper SC, Heimann M. 1996. Global and hemispheric CO<sub>2</sub> sinks deduced from changes in atmospheric O<sub>2</sub> concentration. *Nature*. 381: 218–220.
25. Kirschbaum MUF. 2000. Will changes in SOC act as a positive or negative feedback on global warming? *Biogeochemistry*. 48: 21–51.
26. Lal R. 1995. Global soil erosion by water and carbon dynamics. Pages 131–141 in Kimble JM, Levine ER, Stewart BA. eds *Soils and Global Change*. Boca Raton (FL): CRC.
27. Lal R. 2001. Soil conservation for C-sequestration. Pages 459–465 in Steinhardt GC. ed *Sustaining the Global Farm: Selected Papers from the 10th International Soil Conservation Organization Meeting held May 24–29, 1999 at Purdue University and the USDA-ARS National Soil Erosion Research Laboratory*. (6 March 2007; <http://topsoil.nserl.purdue.edu/fpadmin/isco99/pdf/ISCOdisc/SustainingTheGlobalFarm/K010-R%20Lal.pdf>).
28. Lal R. 2003a. Soil erosion and the global carbon budget. *Environment International*. 29: 437–450.
29. Lal R. 2003b. Global potential of soil carbon sequestration to mitigate the greenhouse effect. *Critical Reviews in Plant Sciences*. 22: 151–184.
30. Lal R. 2003c. Offsetting global CO<sub>2</sub> emissions by restoration of degraded soils and intensification of world agriculture and forestry. *Land Degradation and Development*. 14: 309–322.
31. Lal R, Griffin M, Apt J, Lave L, Morgan MG. 2004. Ecology: Managing soil carbon. *Science*. 304: 393.
32. Liu S, Bliss N, Sundquist E, Huntington TG. 2003. Modeling carbon dynamics in vegetation and soil under the impact of soil erosion and deposition. *Global Biogeochemical Cycles*. 17: 1074 doi:10.1029/2002GB002010.
33. Lyons WB, Nezat CA, Carey AE, Hicks DM. 2002. Organic carbon fluxes to the ocean from high-standing islands. *Geology*. 30: 443–446.
34. McCarty GW, Ritchie JC. 2002. Impact of soil movement on carbon sequestration in agricultural ecosystems. *Environmental Pollution*. 116: 423–430.

35. Meade RH, Yuzyk TR, Day TJ. 1990. Movement and storage of sediment in rivers of the United States and Canada. Pages 255–280 in Wolman MG, Riggs HC. eds *Surface Water Hydrology*, vol. 0–1: *The Geology of North America*. Boulder (CO): Geological Society of America.
36. Onstad CA, Pierce FJ, Dowdy RH, Larson WE. 1984. *Proceedings of the National Symposium on Erosion and Soil Productivity*, December 10–11 1984. New Orleans (LA): American Society of Agricultural Engineers.
37. Oskarsson H, Arnalds O, Gudmundsson J, Gudbergsson G. 2004. Organic carbon in Icelandic Andosols: Geographical variation and impact of erosion. *Catena*. 56: 225–238.
38. Pimentel D. , et al 1995. Environmental and economic costs of soil erosion and conservation benefits. *Science*. 267: 1117–1123.
39. Post WM, Cesar Izaurralde R, Jastrow JD, McCarl BA, Amonette JE, Bailey VL, Jardine PM, West TO, Zhou J. 2004. Enhancement of carbon sequestration in US soils. *BioScience*. 54: 895–908.
40. Sarmiento JL, Gruber N. 2002. Sinks for anthropogenic carbon. *Physics Today*. 55: 30–36.
41. Schlesinger WH. 1990. Evidence from chronosequence studies for a low carbon-storage potential of soils. *Nature*. 348: 232–234.
42. Schlesinger WH. 1995. Soil respiration and changes in soil carbon stocks. Pages 159–168 in Woodwell GM, Mackenzie FT. eds *Biotic Feedbacks in the Global Climate System: Will the Warming Feed the Warming?* New York: Oxford University Press.
43. Schlesinger WH. 1999. Carbon sequestration in soils. *Science*. 284: 2095.
44. Six J, Jastrow JD. 2002. Soil organic matter turnover. Pages 936–942 in Lal R. ed *Encyclopedia of Soil Science*. New York: Marcel Dekker.
45. Smith SV, Renwick WH, Buddemeier RW, Crossland CJ. 2001. Budgets of soil erosion and deposition for sediments and sedimentary organic carbon across the conterminous United States. *Global Biogeochemical Cycles*. 15: 697–707.
46. Stallard RF. 1998. Terrestrial sedimentation and the carbon cycle: Coupling weathering and erosion to carbon burial. *Global Biogeochemical Cycles*. 12: 231–257.
47. Starr GC, Lal R, Malone R, Hothem D, Owens L, Kimble J. 2000. Modeling soil carbon transported by water erosion processes. *Land Degradation and Development*. 11: 83–91.
48. Starr GC, Lal R, Kimble JM, Owens L. 2001. Assessing the impact of erosion on soil organic carbon pools and fluxes. Pages 417–426 in Kimble JM, Follett RF, Stewart BA. eds *Assessment Methods for Soil Carbon*. Boca Raton (FL): CRC.
49. Troeh FR, Hobbs JA, Donahue RL. 2004. *Soil and Water Conservation for Productivity and Environmental Protection*. 4th ed. Upper Saddle River (NJ): Prentice Hall.
50. Van Oost K, Govers G, Quine TA, Heckrath G, Olesen JE, De Gryze S, Merckx R. 2005. Landscape-scale modeling of carbon cycling under the impact of soil redistribution: The role of tillage erosion. *Global Biogeochemical Cycles*. 19: GB4014 doi:10.1029/2005GB002471.
51. Yoo K, Amundson R, Heimsath AM, Dietrich WE. 2005. Erosion of upland hillslope soil organic carbon: Coupling field measurements with a sediment transport model. *Global Biogeochemical Cycles*. 19: GB3003.