Title

Soil Degradation and Global Change: Role of Soil Erosion and Deposition in Carbon Sequestration

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Role of soil erosion and deposition in carbon sequestration

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ABSTRACT
Soil erosion and terrestrial sedimentation are important variables in global change science. Erosion is estimated to transport more than 100 Gt soil yr^{-1}; 70 to 90-percent of which is deposited in depositional basins within the same or adjacent toposequence. Terrestrial sedimentation may constitute a sink of up to 1 Gt C yr^{-1} (missing Carbon (C)-sink = 1.8 (+/- 1.2) Gt C yr^{-1}), which would offset up to 15-percent of global fossil fuel emissions. Our study characterized the rates of input, storage and stability of soil organic matter in three positions of an eroding hillslope and two types of depositional basins of an undisturbed zero-order watershed in Tennessee Valley, CA. Our study provided experimental evidence that in this small, undisturbed watershed photosynthesis is able to replace eroded C and that the depositional basins contain twice as much C, with preliminary findings of three times longer turnover time, compared to the eroding hillslopes. Here we show that burial of eroded-C can promote a significant, formerly unaccounted, terrestrial C-sink in undisturbed landscapes that are not experiencing anthropogenically accelerated erosion.

INTRODUCTION

The soil environment is a principal component of the global carbon (C) cycle where key interactions between biotic and abiotic components take place to regulate the flow of materials to and from the pedosphere, atmosphere and hydrosphere. The pedosphere contain 1500 Gt soil organic carbon (SOC) and 750 Gt carbonate (inorganic) C in the top meter alone (Kirschbaum, 2000). In the absence of anthropogenic interferences, photosynthetic uptake balances release of carbon dioxide (CO_2) by respiration, globally. Human activities contribute to significant disequilibria in the global C cycle. Since the industrial revolution alone, it is estimated that up to 200Gt C that was originally in the biosphere has been released to the atmosphere due to land conversion and degradation (DeFries et al., 1999).

Soil degradation is defined as the decline in the physical, chemical and biological quality of the soil resource. Soil degradation is an important variable in global change studies because of its aerial extent and its effect on input, transport and stabilization of OM and other essential nutrients from or in the soil matrix. It is believed that about 43% of the world's vegetated land area has experienced some form of degradation over the last half of the 20th century alone. In the United States, anthropogenic activities have caused degradation of 90% of croplands and 54% of pasturelands. More than 12 million hectares of arable land, 0.8 percent of the land under cultivation, is degraded and abandoned every year. Soil erosion is the most prevalent form of degradation. Over 15% of the Earth's ice-free land surface or about 80% of the global arable land is experiencing some form of erosion. Accelerated soil erosion by water and wind are responsible for 56% and 28% of all soil degradation, respectively (Daily, 1995; GLASOD, 1990; Nadakavukaren, 1995; Pimentel et al., 1995).

Generally, soil degradation diminishes the productive ability of soils, which in turn affects the ability of soils to take out CO_2 from atmospheric circulation and its contribution to reduction of net fossil fuel emissions from terrestrial ecosystems. Soil organic matter (SOM), the organic component of the soil, is concentrated on the upper soil horizons of O and A, usually the top 35 cm of the soil, and experiences exponential decrease with depth. Therefore, the magnitudes of carbon stored in and lost from the soil are considerably affected by transport and/or degradation of topsoil (DeFries et al., 1999; Harden et al., 1999;

Reichle et al., 1999; Rillig et al., 1999; Schimel et al., 2000; Townsend et al., 1995).

Soil Carbon Redistribution – Erosion and Deposition

Processes regulating the inventory of soil C in an eroding landscape include plant photosynthesis (Net Primary Productivity, NPP), microbial decay of plant and animal remains, and a variety of erosional and depositional processes including bioturbation, soil creep, runoff and landsliding (Figure 1) (Amundson, 2001; Gregorich et al., 1998; McCarty and Ritchie, 2002; Stallard, 1998).

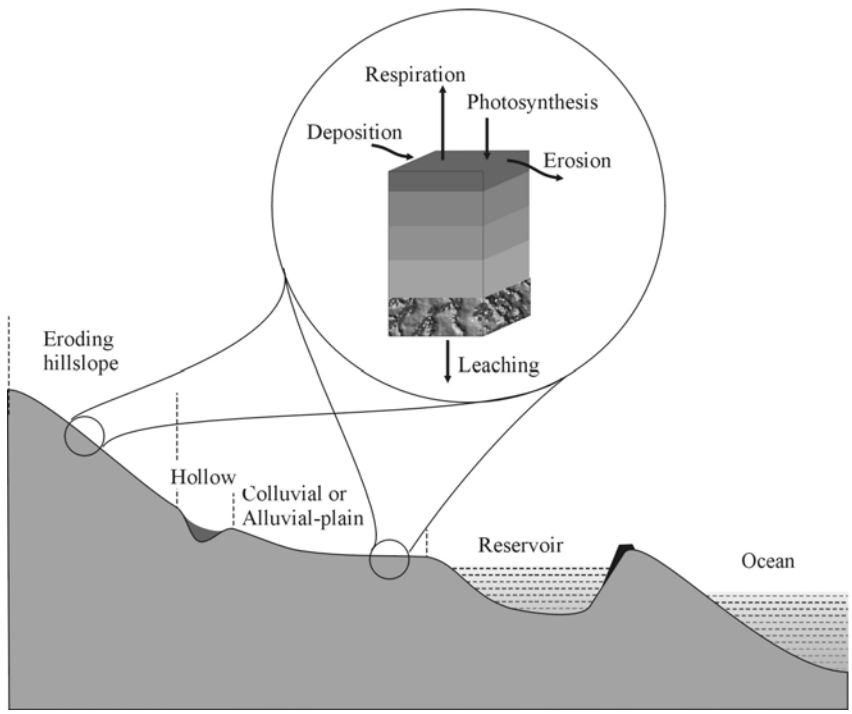

Figure 1. Longitudinal profile of toposequence in Tennessee Valley (not to scale) and description of major C fluxes in a typical depositional pedon. The different shades are supposed to show typical horizonation in a mature soil profile, including bedrock layer (does not necessarily apply to well mixed soils like those in Tennessee Valley). Our work was on the eroding slopes, hollow and alluvial/colluvial plains.

After the inclusion of soil erosion as an explicit term in the mass balance of soil C, the change in the soil carbon stock is modeled as:

$$dC/dt = I - (k_o + k_e)C \quad (1)$$

where, C= soil C stock (kg m^{-2}), I= zero-order introduction of soil carbon by plant growth (kg m^{-2} yr^{-1}), k_o = first order loss of slow carbon by oxidation/decomposition (yr^{-1}), k_e = first order loss or addition of slow carbon by erosion (negative k_e) or deposition (positive k_e) (yr^{-1}) (Stallard, 1998). Leaching is usually not included in the equation because its magnitude is believed to be small compared to the other fluxes.

Erosional transport of SOC

Erosional and depositional landscapes do not fit the steady state C storage concept that was previously assumed to be the case in most terrestrial C models. The processes of soil erosion and sedimentation redistribute more than 100 Gt soil yr^{-1}, along with up to 5Gtg C/yr (Figure 2). Soil erosion removes C from topsoil and continually exposes subsoil that has lower C content. Since around 80% of SOC is structurally bound to soil particles (Rosenbloom et al., 2001), it gets transported down slope along with soil mineral particles. In combination, erosion and cultivation can strip the soil mantle of its original organic carbon stock. In a Mississippi site that Harden et al (1999) studied, 100% of the original soil C was lost over 127 yrs of cultivation, where 80% of the C loss was attributed to erosion and 20% to mineralization.

Deposition of SOC downhill

Approximately 70-90% of soil and associated SOC that is transported from upland watersheds by erosion is stored in different depositional basins within the same or adjacent toposequence. It is not lost to the ocean as assumed in past C models (Stallard, 1998). Only 8% of the most stable C pool that is bound with fine silt and clay fractions is likely to be exported all the way to the ocean (Starr et al., 2000).

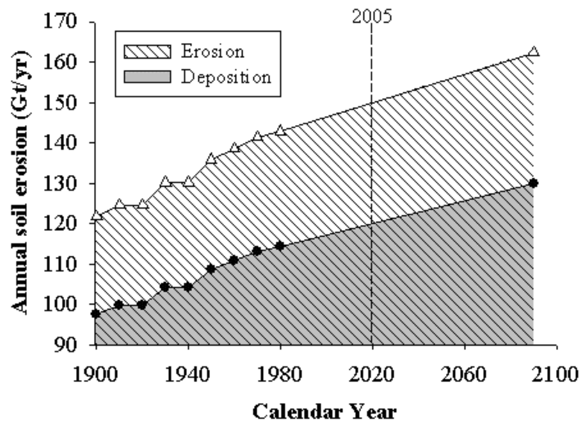

Figure 2. Global rates of soil erosion (Yang et al., 2003) and deposition (calculated as 80% of eroded C) according to (Meade et al., 1990; Stallard, 1998)).

The contribution of deposition, terrestrial sedimentation, to changes in soil properties at different lowland depositional sites depends on the nature of soil derived from upslope. The transported nutrient- and carbon-rich topsoil contributes an improvement in overall quality of the soil in depositional basins. Moreover, burial (after subsequent deposition) in different low lying depositional basins tends to be accompanied by higher proportion of fine soil particles (clay and fine silt), increase the proportion of water-filled pore-spaces that generally changes the redox potential, bioavailability of essential nutrients. In combination, the above conditions retard the activities of soil microbial organisms resulting in protection of SOC from decomposition (Canuel and Martens, 1996; Jacinthe et al., 2001; Krull et al., 2001).

Erosion associated C sink

The processes of soil erosion and terrestrial deposition can constitute a C sink if two important conditions are met: (1) there is at least partial replacement of eroded C at the eroding site by photosynthesis and that (2) eroded soil is deposited terrestrially (in different depositional basins such as flood plains or reservoirs) and protected from decomposition. Burial can promote C-sequestration locally because it removes organic C from more active components (plant biomass and topsoil) and stores them in potentially more passive reservoirs where the C is offered physical protection from near-surface environments

(Gregorich et al., 1998; Harden et al., 2002; Harden et al., 1999; Jacinthe and Lal, 2001; McCarty and Ritchie, 2002; Yoo et al., 2001). On a global scale, up to 2.3 Gt C yr^{-1} (on average 0.8 Gt C yr^{-1}) sequestration could be realized from soil erosion and terrestrial sedimentation (Stallard, 1998). Moreover, contrary to previous assumption of potentially significant oxidation of SOC during erosion (Schlesinger, 1995), Smith et al., (2001) found that the efflux of CO_2 accompanying soil erosion is insignificant (-0.001 $\pm$ 0.024Gt yr^{-1}) and that sedimentation is the dominant variable in C-budgets of eroding landscapes, compared to plant inputs and oxidation of SOC.

Objectives

Among the most important, but yet fully unanswered questions in this research area include: (1) What is the fate of eroded C (where does it end up and what kind of transformations ensue), (2) How much of the eroded soil C is replaced by photosynthesis and (3) Is buried C significantly more stable than C in other parts of the watershed? The objective of this study is to ascertain if the requirements for erosion to constitute a C sink are met in a small, undisturbed, zero-order watershed experiencing natural rates of erosion due to a combination of bioturbation by pocket gophers and diffusive mass transport. Here we present preliminary results of C input from aboveground net primary productivity (ANPP), storage, and turnover rate in summit (HSS) and back- and foot-slope (HSBF) of an eroding hillslope and two types of depositional basins - terrestrial depression/hollow (LW) and an alluvial/colluvial plain (A/CP). This project was motivated by the lack of adequate studies that describe the role of erosional and depositional processes on C sequestration in natural, anthropogenically undisturbed watersheds. After presenting results from our study site we will present a short discussion the phenomenon on disturbed landscapes, such as cultivated landscapes and the role of soil conservation practices.

METHODS

Study Area

The study is being conducted in Tennessee Valley, California (37^0 latitude, 122^0 longitude, inside the Golden Gate National Recreation Area). Soils in Tennessee Valley are derived from chert, greenstone, and sandstone bedrock of the Franciscan assemblage. Soil production mainly results from biogenic disturbance of weathered bedrock by pocket gophers (*Thomomys bottae*). Soil thickness varies across the toposequence, but generally soils are organic-rich, stony loam showing little or no horizon development and with a typically abrupt soil-bedrock boundary. Burrowed soil is transported downhill and deposited in unchanelled hollows. Evacuation of the depositional basins in Tennessee Valley occurs as a result of rains and exfiltrating surface flows that result saturated overland flow and landslides. Since the last documented major landslide event, 13000 years ago, the hollow and alluvial/colluvial plain have been infilling gradually by diffusive mass transport from the adjacent convex slopes. Rate of erosion in the backslope is about 50 g m^{-2} y^{-1} and as high as 130 m^{-2} y^{-1} at the shoulders (Dietrich et al., 1995; Heimsath et al., 1999; Yoo et al., 2003).

The highest position is the hillslope-summit, the higher and lower elevation of the two groups in HSBF are from backslope (B) and footslope (F), respectively (Figure 3). For convenience, the average values for B and F are presented as HSBF. The axis of the hollow covers (30m-elevation difference within horizontal distance of around 50 meters. The slope is very gentle in the alluvial/colluvial plain. The average elevation along with rates of ANPP, soil and C erosion and deposition are given in Table 1.

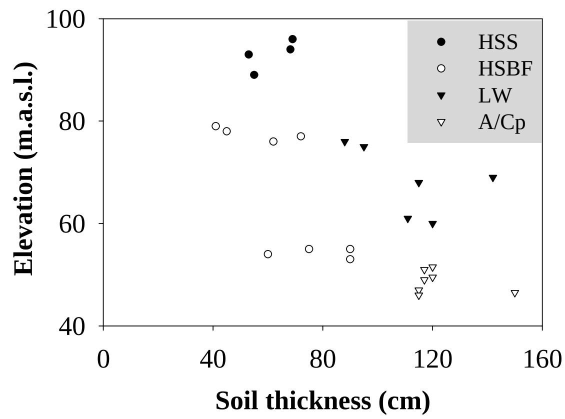

Figure 3. Change in soil thickness as a function of elevation in our study site. The highest position is the hillslope-summit (HSS), while the higher and lower elevations of the two groups in HSBF are from backslope (B) and footslope (F) respectively. The two depositional basins are the hollow (LW) and alluvial/colluvial plain (A/CP).

Table 1. Average elevation, rates of ANPP and average rates of soil and C erosion and deposition at the different positions in the toposequence Soil and C erosion rates are derived from (Heimsath et al., 1997; Yoo, 2003).

	Average Elevation (m.a.s.l.)	*Average ANPP (g C m^{-2} yr^{-1})*	*Soil erosion (-ve)/ deposition (+ve) (g m^{-2} yr^{-1})*	*C erosion (-ve)/ deposition (+ve) (g m^{-2} yr^{-1})*
HSS	105	144	-130	-2.05
HSBF	50-80	188	-50	-2.05
LW	60	242	90	1.25
A/CP	45	467	120	1.25

The annual rates of precipitation in the valley range from 600 to 900 mm, with dominant vegetation cover of Mediterranean annual grasses and *Baccharis* shrubs, and Holocene forest cover. There has been no documented grazing since 1972 (Heimsath et al., 1999; Yoo et al., 2003).

Experimental approaches

Samples for characterizing SOC in hillslope (summit, backslope and footslope) and depositional profiles of the hollow and alluvial plain were collected by digging soil pits that reached the soil-bedrock boundary. Samples for C and N analyses were ground using mortar and pestle and passed through a 64μm sieve before analysis. Total C and N were measured with a Carlo Erba elemental analyzer. Carbon inventory for the entire depth profile was calculated as:

$$C_{inv} = \sum_{i=0cm}^{N} \Delta Z_i \cdot \rho_i \cdot (1 - R_i) \cdot [\%OC_i \cdot 100] \qquad (2)$$

where C_{inv}= carbon inventory over the entire depth profile (g C m^{-2}), ΔZ_i = thickness of each soil layer (cm), ρ_i = bulk density (g cm^{-3}), R_i= fraction of rock and $\%OC_i$ = C concentration in <2mm fraction of sampled soil layer. Above ground NPP was measured by destructive harvest method using a 0.25 x 0.25m grid.

To determine dynamics of C our study watershed, we used a first-order model of decomposition [Equation (1)]. The solution for Equations (1) is given as:

$$C(t) = \frac{I}{k_o + k_e}\left[1 - \left(1 - C_o \frac{k_o + k_e}{I}\right) e^{-(k_o + k_e)t}\right] \qquad (3)$$

where C_0= the carbon inventory at t=0, in this case the carbon inventory at the bedrock interface; Carbon inventory at time *t*, *C(t)*, was calculated from Equation (2). Since C_0=0 for both the eroding hillslopes (at the time of first soil development) and the depositional basins (due to complete evacuation of the sediment during the last major landslide event), equation (3) is reduced to:

$$C(t) = \frac{I}{k_o + k_e}\left[1 - \left(e^{-(k_o + k_e)t}\right)\right] \qquad (4)$$

Before the C stock calculation was performed, the age of carbon at a few representative depths was determined with ^{14}C (incomplete data set), the age of C in depths between the surface and the dated layers was approximated by linear interpolation. The stock of carbon that accumulated between successive time stages was determined by adding carbon inventory above the different depths. After plotting C (t) versus time before present (yrs) values for *Input* and *k* were determined by best fit using a solver function in Excel (Microsoft Office 2000). Since k_e was previously determined, k_o is computed as the difference between model *k* and k_e. This method gives the average turnover time of organic matter over (= $1/k_o$) during the entire history of the profile (modified from Clymo, 1984) (see Trumbore et al., 1999 for use in peat cores).

RESULTS AND DISCUSSIONS

C input and storage

On average, the rate of input of SOC from above ground NPP (ANPP) in the depositional basins was found to be twice as much as that on the eroding slope positions. Although ANPP increases by more than 300% from HSS and A/CP, in contrast the difference between HSBF and LW is small (Figure 4).

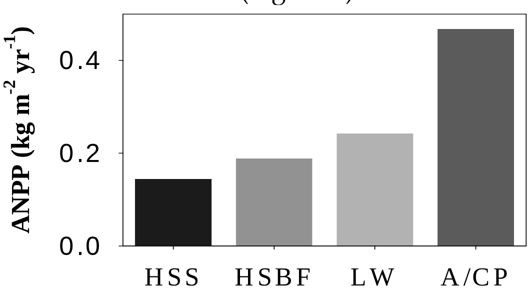

Figure 4. Above ground net primary productivity (ANPP) at different sites in the toposequence.

The C concentration (%C in the <2mm soil fraction) is not necessarily higher in the soil depth profiles of the depositional basins than in the eroding slopes (Figure 5). A/CP generally has high C content (profile average of 3%) compared to the other parts of the watershed. HSS, on the other hand, has highest C content in the surface but shows sharpest decline with depth. We found that the C concentration in HSBF and LW roughly the same and follows similar pattern with depth.

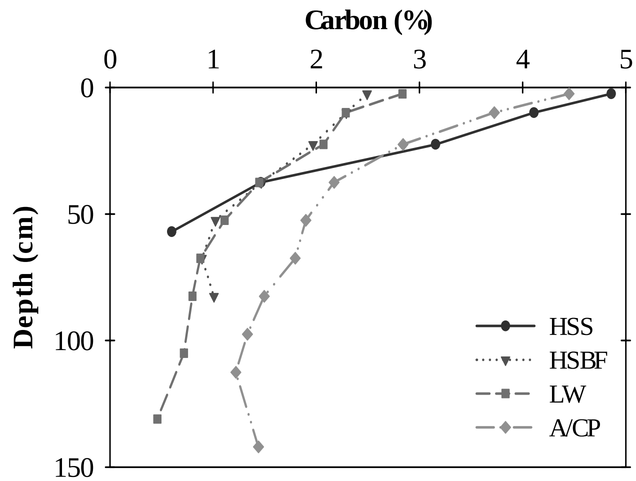

Figure 5. C concentration in the eroding slopes and depositional basins.

Despite the lack of significant difference in C concentration between the eroding slopes and depositional basins, the up to double soil thickness in the depositional basins gives rise to significant differences in their C inventories (Figure 6). There is almost twice as much C stored in the depositional basins compared to the eroding slopes. Depositional sites have higher C inventory as a result of input of C from erosion, higher rate of NPP, and reduced decomposition due to burial (see section on *storage effectiveness*) and possibly other environmental conditions (notably increased moisture) in the lower slope positions. Depositional basins are also likely to contain relatively higher fraction of their C in labile pools since these can be transported easily with erosion (Davidson and Ackerman, 1993; Gregorich et al., 1998).

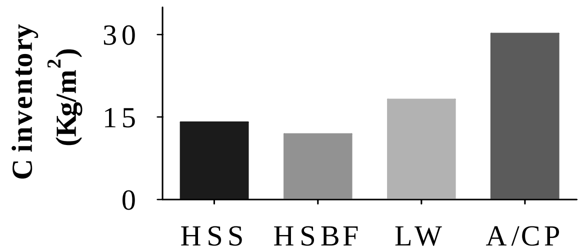

Figure 6. C inventory in the different positions of the watershed.

The C:N of OM in HSS, HSBF and LW decreases with depth, indicating that the OM gets progressively more humified with depth (Figure 7). In the A/CP, however, the trend is reversed. The C:N of dominant vegetation types (from leaves of *Baccharis* and foliage of the grasses) growing in A/CP is higher (average value of 64, compared to 41 in the rest of the watershed). Other possible explanations for the unique trend in A/CP could be higher soil moisture content throughout the year, anaerobiosis, and accumulation of finer soil particles in A/CP that can reduce rate of humification after burial or possibly elevated rates of denitrification at depth.

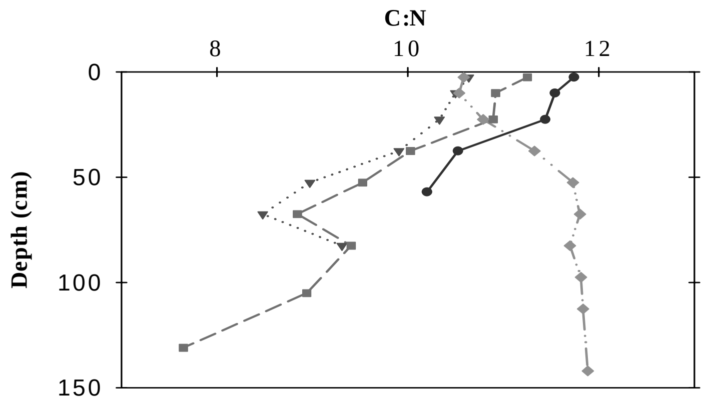

Figure 7. C:N at the different locations in the eroding/depositional toposequence.

Storage effectiveness

Initially, in the first few hundred years of our model simulations, the rate of C accumulation was found to be faster in the eroding slopes (Figure 8). However, the eroding slopes approach steady state at a much faster rate, within the first two thousand years of C accumulation. In comparison, the depositional basins are observed to accumulate a lot more C at a consistently fast rate, especially in the A/CP. Our model fits indicate that the depositional basins, especially A/CP, are still actively accumulating C (are not likely to be nearing steady state) since the combined rate of C input from deposition and NPP are still much larger than the rate they are loosing C by combined processes of decomposition and leaching.

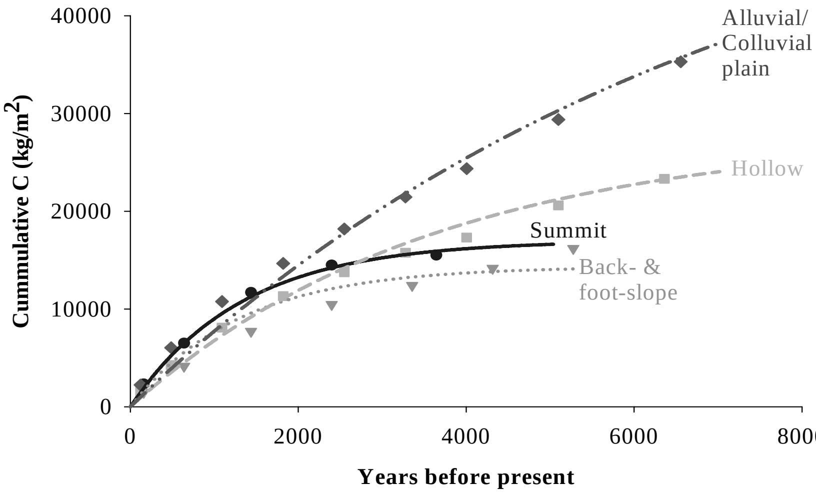

Figure 8. Rate of C accumulation in the different positions of the watershed.

The modeled average turnover was found to be up to three times slower in the depositional basins compared to the eroding slopes. The average turnover of the soil profiles increased with decreasing elevation and the average turnover time in the depositional basins was more than 4 times slower than that on the eroding slopes. This small difference in turnover time indicates that not only are the depositional basins accumulating a lot of C but they are also providing protection from decomposition (Figure 9). High inventory of SOC in depositional basins was found to be closely associated with longer average turnover times of organic matter.

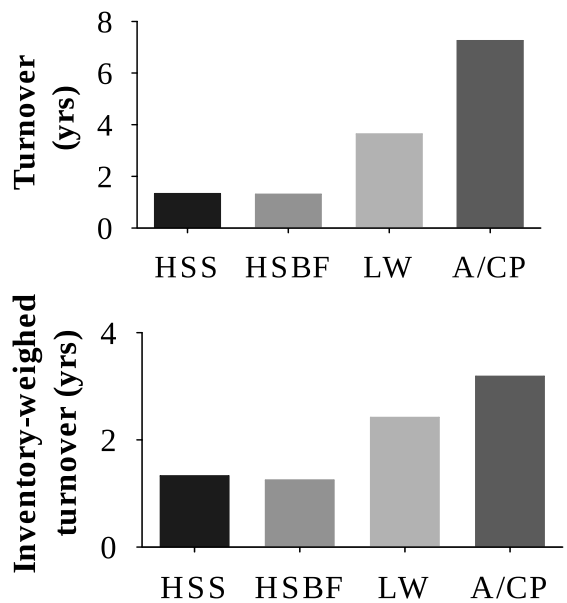

Figure 9. Profile average turnover and Inventory weighed turnover.

Inventory weighed turnover (=turnover/scaled Cinv, where Cinv for each site was scaled to HSS-Cinv) shows that overall the depositional basins are effective in C storage, meaning they have accumulated a lot more old C (than the eroding slopes). Relative storage effectiveness was highest in A/CP>LW>HSS>HSBF. This indicates that the depositional basins are stabilizing C by providing protection from decomposition (accumulation of large amounts of old C) and storing a lot of C overall.

Implications of soil degradation and reclamation of degraded landscapes

The patterns we observed in our study site are consistent with other studies on cultivated or anthropogenically disturbed landscapes that show that in hillslope settings C inventory is smallest in steeply sloping mid-slope positions and highest in lower slope positions or other depositional settings. Upper shoulder and summit positions usually show higher C inventory mainly because of slower rate of erosion, compared to steep mid-slope positions (see Figure 10) (Gregorich et al., 1998). But, the cultivated systems have lower C concentration and inventory than native grassland sites due to a combination of different factors including: (1) the input of C that returned to the soil tends to be lower, compared to native grasslands or forests, (2) enhanced rate of decomposition by tillage, mixing and breaking up of aggregates that contribute to provision of moisture, aeration and temperature conditions that are conducive for biological activity of decomposing microorganism, (3) redistribution of disturbed soil by accelerated erosion and deposition. Consequently, it is estimated that 20 to 30% of the soil's C stock can be lost due to cultivation, with most of the loss occurring within the first five years of cultivation. Erosion becomes a dominant loss mechanism in the long-term (Gregorich et al., 1998).

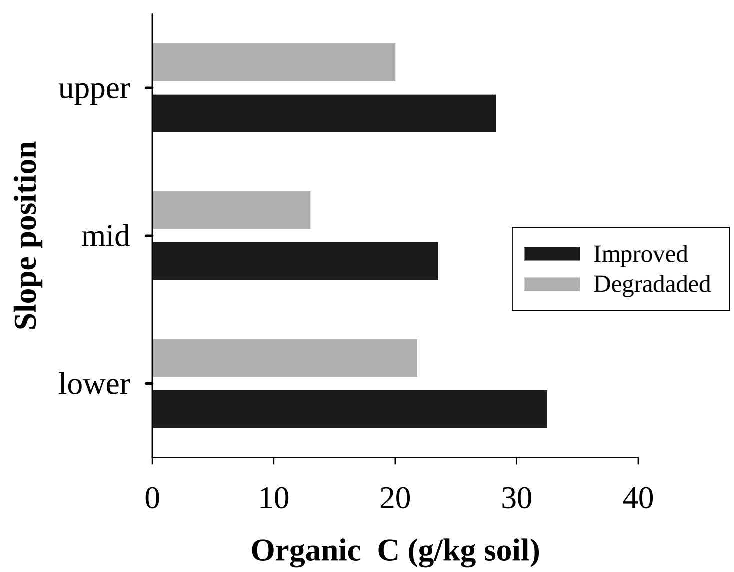

Figure 10. The stock of organic in three hillslope positions under degraded and improved conditions in an Ontario landscape (Gregorich et al., 1998).

Management practices such as no (or reduced) till or crop rotations have strong influence on C balance in cultivated and/or degraded systems. As shown in Figure 10, improvement of degraded landscapes by reducing rates of erosion, maintenance of soil cover, and returning plant residue to the soil tend to increase C inventory. Therefore, processes of erosion and decomposition have potential to constitute C sink, while reclamation of degraded landscapes can further improve C storage and the magnitude of the C sink.

CONCLUSION

In this study we observed processes by which soil erosion and terrestrial sedimentation of eroded SOC could constitute a C sink mechanism when compared to a stable landscape that is not experiencing soil redistribution. This study provided evidence that in this small, undisturbed watershed soil erosion affects soil carbon budgets by removal of topsoil, incorporation of subsoil material, deposition of transported material downhill, and protection of SOC from decomposition in the depositional basins.

Usually soils in depositional areas are typically made up of smaller and lighter density fractions as these are easily transported by agents of erosion. The accumulation of light fraction materials, coupled with the higher C:N at depth in depositional basins could mean that there is high proportion of C that is undecomposed (in labile fractions) or that significant part of the OM could be stabilized physically by burial under high moisture conditions.

Despite the potentially significant contribution of soil erosion in soil carbon sequestration, it should not be overlooked that soil erosion can also represent an output term if it reduces plant productivity or if sediment is delivered to an area where turnover is more rapid than the eroded landscapes. Furthermore, the amount of C that is returned to soil in the eroding slopes affects the strength of the C sink/source. It is critically important to protect and rehabilitate degraded soils because, as is shown in Figure 10, proper rehabilitation of degraded lands could further improve the sequestration potential of hillslope soils.

ACKNOWLEDGEMENT

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