Sustained high magnitude erosional forcing generates an organic carbon sink: test and implications in the Loess Plateau, China.

Y. Li1*, T.A. Quine2, H.Q. Yu1, G. Govers4, J. Six4, D.Z. Gong1, Z. Wang3, Y.Z. Zhang1, K. Van Oost3

1Institute of Environment and Sustainable Development in Agriculture, Chinese Academy of Agricultural Sciences, 12 Zhongguancun South Street, 100081 Beijing, China.
2Department of Geography, College of Life and Environmental Sciences, University of Exeter, EX4 RJ Exeter, UK.
3Georges Lemaître centre for Earth & Climate Research, Earth & Life Institute, Université Catholique de Louvain, 1048 Louvain-la-Neuve, Belgium.
4Physical and Regional Geography Research Group, Katholieke Universiteit Leuven, 3001 Heverlee, Belgium.
5ETH Zürich Institut f. Agrarwissenschaften Tannenstrasse 1 8092 Zurich.

Abstract

Humans are now the most important geomorphic agent on the planet and accelerated erosion in agricultural landscapes results in high magnitude lateral organic carbon (OC) fluxes and significant perturbation of the land-ocean carbon flux. Nevertheless, the net effect of these lateral carbon fluxes on the C cycle is poorly constrained and there is no consensus as to whether they drive a net source or net sink of atmospheric CO2. Here, we test the hypothesis that, under sustained erosional forcing, soil carbon stocks on hillslopes reach a new equilibrium state in which all carbon exported with erosion is replaced; and, therefore, erosion results in a net sink for atmospheric CO2 at the scale of eroding hillslopes. The evidence from our study site, in the Loess Plateau of China, is consistent with this hypothesis. Despite net export of OC equivalent to ca 10% NPP, we found that all of the eroded OC was replaced and, therefore, that the sink strength was equal to the C export rate. This sets the upper limit of the erosion-induced sink term at the scale of whole watershed. The fate of the exported carbon in reservoirs, floodplains, riverbeds and the ocean ultimately controls the watershed-scale sink strength. Nevertheless, the full replacement observed here suggests that erosion does not induce a C
source, irrespective of the fate of the exported carbon, at least for high-input agricultural systems. Finally, we propose that assessment of the C cycle perturbation associated with erosion-induced lateral C fluxes must be made an integral part of accounting mechanisms for climate change mitigation strategies that are based on land use change and C sequestration in terrestrial environments.

Keywords: Carbon cycle, soil erosion, restoration

1. Introduction

It has been suggested that humans are now the most important geomorphic agent on the planet (Wilkinson, 2005) and our role in accelerating erosion in agricultural landscapes is widely acknowledged (Stallard, 1998). These geologically significant sediment fluxes result in high magnitude lateral carbon (C) fluxes and significant perturbation of the land-ocean carbon flux (Regnier et al., 2013). Nevertheless, the net effect of these lateral carbon fluxes on the C cycle is poorly constrained and there is no consensus as to whether they drive a net source or net sink of atmospheric CO₂.

Consideration of the key elements and mechanisms highlights the areas of uncertainty that underlie the current intense debate concerning the effect of high magnitude fluxes at the timescale of recent human perturbation (e.g. Berhe et al., 2007). There is broad agreement that human activities typically accelerate erosion 10- to 100-fold (Dearing and Jones, 2003) and that this exposes soil organic carbon (SOC) to lateral redistribution and burial. It is estimated that agricultural land use alone has increased rates of SOC erosion by 0.4-1.2 Pg C yr⁻¹ (Doetterl et al. 2012, Stallard 1998). Nevertheless, at present there is no consensus whether this SOC, eroded from agricultural soils, is rapidly metabolized and returned to the atmosphere as CO₂ without full replacement, resulting in an important net CO₂ source (e.g. Lal 2003), or instead is rapidly transferred to sedimentary basins and replaced at the site of erosion, resulting in effective sequestration and hence a net CO₂ sink (Stallard 1998). There is little
agreement regarding either the fate of SOC during and after transport (Hoffmann et al., 2013) or the
degree to which eroded SOC is replaced by new photosynthate at the site of erosion (Van Oost et al.,
2007). The proposition that new soil organic matter formation from vegetation inputs replaces some
or all of the exported SOC (Harden et al., 1999) underpins those studies that identify anthropogenic
erosion with a C sink. Variation in the degree of replacement of exported SOC has been explained in
conceptual terms by Van Oost et al (2007 SOM), who proposed that 2 distinct periods could be
identified following an erosion-inducing perturbation (e.g. change in land-use) and a subsequent
period of sustained erosional forcing. In the ‘transient’ period following the perturbation partial
replacement of lost SOC occurs at eroded sites and, in this period, these sites evidence a net loss of
SOC but this is compensated for by net gain of SOC at depositional sites. The albeit partial
replacement of SOC allows the whole system to operate as a net sink of atmospheric CO$_2$ (Quine and
Van Oost, 2007, Van Oost et al., 2007). In the second period identified by Van Oost et al (2007
SOM), SOC stocks at eroded sites have declined to a level that permits establishment of a new
dynamic equilibrium in which full replacement of eroded SOC occurs at sites of erosion. Under the
latter conditions, the eroding element (e.g. hillslope) in the system would exhibit constant SOC
content. Furthermore, the whole erosion-deposition system would either be in steady-state (if all
exported SOC is rapidly mineralized) or a sink (if some exported SOC is preserved) and the
maximum net sink magnitude would be determined by the SOC flux exported from the eroding
element. While this explanation is conceptually elegant, there remains a need to test the hypothesis
that, in agricultural systems subject to high erosional forcing, SOC stocks do in fact reach a new
dynamic equilibrium in which eroded SOC is replaced and the erosional export of flux SOC is in a
steady-state. The alternate hypothesis would be that continued erosional forcing would drive
deterioration in soil quality that prevents establishment of dynamic equilibrium and that SOC stocks
would continue to decline. This research question is relevant not only to the perturbation of the carbon
cycle by anthropogenically accelerated erosion, but to the debate concerning the interplay between the
biosphere, physical erosion and weathering at geological timescales (West et al., 2005; Dietrich and
Perron, 2006; Boucher & Gaillardet, 2014) and the implications for climate regulation (Raymo and Ruddiman, 1992; Willenbring and von Blanckenburg, 2010; Herman et al., 2013).

In order to address this pressing research question, we employ a conceptual framework that builds on that presented by Berhe et al. (2007). This framework is illustrated in Figure 1 and can be summarized as follows. Erosion affects watershed-level SOC balance by changing the magnitude of two opposing C fluxes, namely SOC input through photosynthesis and SOC loss through respiration. The vertical soil-atmosphere C fluxes on the eroding hillslope are represented by \( I_h \) and \( R_h \), where \( I_h \) and \( R_h \) are temporally (over a full year) and spatially (over an eroding hillslope) integrated fluxes for rates of SOC production and SOC loss through respiration, respectively (expressed as t C ha\(^{-1}\) year\(^{-1}\), Figure 1). In a natural system \( I_h \) would equate to NPP but in the agricultural system studied here, \( I_h \) is NPP less the biomass exported at harvest (in this case, grain and stalk). If the hillslope SOC budget is in dynamic equilibrium then \( I_h - R_h = \text{C}_{\text{exp}} \), where \( \text{C}_{\text{exp}} \) is the erosional SOC flux from the hillslopes. Furthermore, if in dynamic equilibrium, the erosional SOC flux equates to the maximum C sink term associated with erosion (\( \text{C}_{\text{atm}} \)), i.e. \( \text{C}_{\text{atm}} = (I_h - R_h) = \text{C}_{\text{exp}} \). Note that this is the maximum C sink term at the scale of the eroding hillslope; the actual C sink at the scale of the whole watershed may be less depending on the fate of the exported SOC.

The terms \( I_h \) and \( R_h \) can be further defined by reference to non-eroding systems where \( i \) and \( r \) are the differences in the SOC production and respiration terms relative to a non-eroding system (subscript 0), respectively (i.e. \( I_h = I_0 + i \) and \( R_h = R_0 + r \) (Berhe et al., 2007). If it is accepted that a non-eroding system is in dynamic equilibrium between input and respiration (\( I_0 = R_0 \)), then \( \text{C}_{\text{atm}} \) for an eroding system will be equal to (i-r). Therefore, for a SOC budget in dynamic equilibrium at the scale of the eroding hillslope, erosion must result in a reduction in the respiration term (i.e. \( r < 0 \)) that balances the SOC export and any SOC input changes due to erosion, i.e. \( -r = C_{\text{exp}} - i \). The quantification of \( r \) must include both erosional hillslopes (\( r_e \)) and colluvial footslopes where eroded SOC is redeposited and
exposed to aerobic conditions \( (r_d) \) so that \( r = (r_e/Ae) + (r_d/Ad) \) (where \( Ae \) and \( Ad \) are the fractional areas of the hillslope experiencing erosion and deposition, respectively).

Here, we test the hypothesis that on agricultural hillslopes under sustained high magnitude erosional forcing, SOC stocks reach dynamic equilibrium in which eroded SOC is replaced, the export SOC flux is in steady state at decadal timescales and a potential SOC sink is created. We do so by quantifying the current component SOC fluxes \( C_{\text{exp}}, i, r_e \) and \( r_d \) in an in-situ measurement campaign in an intensively monitored field system. We consider our slope to exhibit SOC stocks in dynamic equilibrium and to be generating a steady-state SOC flux if \( C_{\text{exp}} = i - r \).

2. Materials and Methods

2.1 Study Site

Home to the Terracotta Army and once the cradle of Chinese civilization, the Loess Plateau of China has a long agricultural history and became one of the most degraded ecosystems in the world (Ren, 2006). Sediment yields from the eastern plateau (from Hekouzhen to Longmen) exceeded 4800 t km\(^{-2}\) year\(^{-1}\) and croplands were subject to erosion rates an order of magnitude higher than those typical in Europe and North America (Doetterl et al., 2012). Therefore, this study area exhibits the characteristics required to test our hypothesis and the wider implications for land management strategy if it is accepted.

The site at which monitoring was undertaken is located near Nanwang village (35°03’ 18.45” N, 109°38’ 25.34” E), in Shaanxi province. The area is characterized by a semiarid continental monsoon climate, with a long-term mean annual temperature of 13.2 °C, an annual precipitation of 540 mm, and approximately two-thirds of the annual rainfall occurring from July to September. The slope had a length of 54 m, width of 45 m, angle of 3.1°-16.8°, and elevation of 704 to 713 above mean sea level (Figure 2). In common with the majority of the loess plateau, the soil was developed from loess parent material dominated by the silt fraction, in this case with a uniform soil texture (25% clay, 70% silt, and 5% sand). The soil at the site is classified as a Calciustept in the U.S. taxonomic classification.
system (Soil Survey Staff, 1999); this soil type represents 64% of the loess plateau surface. As is the case for much of the loess plateau, the study area has a long history of cultivation dating back more than 1000 yr. SOC concentrations for the plough layer in the Loess soils are low and range 0.4-0.6% (Li et al., 2006). In the study area, wheat (*Triticum aestivum* L.) and corn (*Zea mays* L.) are the major crops in the rotation with soybean (*Glycine max* (L.) Merr.), rape (*Brassica Napus* L.) and sweet potato (*Ipomoea batatas* L.). Typical soil depths in the eroding and depositional positions are, respectively, 30 and 45 cm.

In considering the degree to which a single field can be representative of an area as large as the Loess Plateau it is clear that it cannot be a statistically representative sample; however, it may display characteristics typical of the area. This has been highlighted above with respect to soil type and land use. With respect to topography, it is worth noting that much of the heavily eroded Loess Plateau is characterized by repeating landform assemblages of rounded convex ridges dissected by gully walls that collapse regularly as a result of mass movements (Zhang et al., 1998; Zhao and Cheng, 2014). Cultivation in these areas is predominantly located on the upper convex slopes and in floodplain and other depositional contexts (e.g. infilled check-dams) in the valley floor. The field site studied here is located on a typical upper convex slope and is characterized by slope angles and lengths in the modal class of the region, representing c. 60% and 90% of the cultivated land in 1986 and 2000, respectively (Chen et al., 2007). Although it is the case that the topography of the individual field studied determines the limited potential for deposition within the field, this is a common occurrence due to the recession of the gully walls (Quine et al., 1997; Zhang et al., 1998). Recent work on basin-wide sediment budgets reports that re-deposition on hillslopes of the Loess Plateau is a small term while sediment export rates are typically very high (92.3% +/- 18) (Ran et al., 2014). Finally, despite the diversity and scale of the loess plateau, prior estimates of cropland soil erosion rates have fallen within relatively narrow bands, 50 and 80 t ha⁻¹ yr⁻¹ (Zhang et al., 1994; 1997; 1998; 2002). Therefore, while statistical extrapolation of results to the wider loess plateau would be inappropriate, it is reasonable to propose that mechanisms operating at the chosen field site may be broadly representative of those operating over a wider area.
2.2 Quantification of erosion, deposition and export

Rates of soil erosion and deposition, representative of the period 1954-2007, were derived using the artificial fallout radionuclide caesium-137 ($^{137}\text{Cs}$) as outlined below. Since no significant change in land use or erosion intensity has occurred over the period 1954-2007, these long-term rates are considered to be reliable measures of contemporary erosion and deposition. Individual rates of erosion or deposition derived for each sample were integrated (with weighting for sample spacing) to establish the rate of soil export from the hillslope and used with soil C measurements (made in 2007) to derive current values of $C_{\text{exp}}$.

Using a hand-operated core sampler with a diameter of 8 cm, 53 soil cores were collected to a depth of 45 cm for the upper and middle slope (potential eroding sites) and to a depth of 60 cm at the lower slope positions (potential depositional sites). From these cores, soil samples were extracted at 15 cm intervals. These samples were used for determining the $^{137}\text{Cs}$ inventory and carbon content (see SOM). Reference sites for determining the local reference (accumulated fallout and decay) inventory of $^{137}\text{Cs}$ were established on uncultivated grassland near the cultivated site. The reference sites have remained undisturbed for more than 200 years. Reference soil cores were taken to a depth of 60 cm to ensure that the core had penetrated to the full depth of the $^{137}\text{Cs}$ profile. We measured the fresh soil moisture after sampling, and calculated the soil bulk density (BD; kg m$^{-3}$) using the sampling volume and oven-dry soil mass.

Using the methods described in Li et al. (2007), we estimated soil redistribution (erosion or deposition) and soil organic carbon (SOC) redistribution. $C_{\text{exp}}$ for the hillslope was then derived as the difference between total SOC erosion and total SOC deposition. We also collected soil samples at a depth of 0-15 cm on erosional (5 cores) and depositional (3 cores) sites, respectively, for determining the microbial biomass carbon (MBC). The MBC per unit of soil (mg kg$^{-1}$) was estimated using the methods described in (Vance et al., 1987).
2.3 Monitoring of in situ soil CO$_2$ fluxes, temperature and moisture

We derive values of $r_e$, $r_d$ and $r$ using measurements of soil respiration (in the absence of vegetation) collected at high temporal resolution along an erosion gradient based on 14 sites. In total, 13296 observations of the soil to atmosphere CO$_2$ flux were made through full diurnal cycles between April 2007 and September 2008. We investigated the factors controlling the magnitude of the soil to atmosphere C flux (due to in-situ soil respiration) including: soil temperature, moisture, soil erosion, and SOC stock and quality.

Based on the $^{137}$Cs-derived erosion and deposition rates, we divided the sloping land into sites characterized by erosion (E) and deposition (D). Soil CO$_2$ emissions were measured using a LICOR LI-8100 Automated Soil CO2 Flux System (LI-8100, LI-COR, Lincoln, NB, USA). The soil chamber has an internal volume of 854.2 cm$^3$ with a circular contact area to soil of 83.7 cm$^2$, and was placed on PVC soil collars. Vegetation in the collars and the area of 1 m$^2$ around the collars was removed during the measurement period. Soil temperature (5cm depth) was monitored using mercury thermometers inserted into the soil close to the collars. Soil volumetric water content was measured close to the collars for the 0–10 cm layer with a portable hydro-sense system (TDR probe, Campbell, US). The CO$_2$ emission, soil temperature and moisture measurements were conducted for each sampling point five times during daytime (8am, 11am, 2pm, 4pm and 6pm) from April 20$^{th}$, 2007 to October 30$^{th}$, 2007, while the measurement frequency was reduced to four times a day (8am, 11am, 2pm and 4pm) from November 5$^{th}$, 2007 to March 22$^{nd}$, 2008. Nighttime measurements (from 8pm to 6am with a 2 hour interval) were conducted from May 24, 2008 onwards. Daytime measurements were conducted with intervals of two days and nighttime measurements were conducted with intervals of nine days. Correction procedures and the methods used to estimate time-integrated CO$_2$ fluxes and to identify the factors controlling soil respiration are described in Supplementary Online Material. The statistical software packages SAS 8.1 (SAS Institute, 1990) were used for Duncan's multiple range tests, at the 95% confidence level.
2.4 Estimation of carbon inputs

At this agricultural site, because stalk and grain are typically harvested and removed from the site, SOC production ($I_h$) is dominated by root biomass. The relative magnitude of the latter at erosional, stable and depositional was estimated, assuming that root biomass was proportional to above ground biomass (Berhe et al., 2008). Measurements of crop biomass (wheat) at three neighboring sloping sites were conducted in 2011 and 2012 in order to quantify differences in biomass, and by extension SOC, production between eroding and depositional sites. At each slope, three plots with a surface area of 1m$^2$ were established at eroded and deposited sites. The aboveground biomass of crop shoots and seeds from 18 plots (9 at eroded sites and 9 at deposited sites) were measured after harvest. Based on the assumption that the aboveground biomass is proportional to SOC production ($I_h$) (e.g. Berhe et al., 2008), we estimated the relative magnitude of C input at eroding and depositional sites.

3. Results

Erosion, deposition and Carbon export ($C_{exp}$)

Based on the $^{137}$Cs measurements, we estimate that 92% of the hillslope has experienced erosion. We obtained mean rates of soil erosion of $73\pm6$ t ha$^{-1}$yr$^{-1}$, equivalent to an annual soil loss of $5.8\pm0.3$mm, for the 49 profiles at eroding (E) sites. In contrast, only four profiles at the bottom of the slope, representing 8% of the hillslope experienced deposition (D) with a mean rate of $67\pm29$ t ha$^{-1}$ yr$^{-1}$ (Table 1). This implies that only 7.5% of the eroded soil is retained while 92.5% is exported from the cultivated slope. The export rate of sediment from the cultivated hillslope then equals $63\pm29$ t ha$^{-1}$ yr$^{-1}$. This rate is clearly specific to this field site and its particular configuration of slope angle and length and potential for deposition. Nevertheless, slope angles and lengths are in the modal class of the region and export observed rates are within the range reported for similar terrain in the Loess Plateau (Zhang et al., 1994; 1997; 1998; Li & Lindstrom, 2001; Li et al., 2003).
The spatial pattern of SOC storage is significantly related to the pattern of erosion and deposition, with correlations of 0.87 (p < 0.01) and 0.70 (p < 0.01) for the upper 0.45 m and 0.15 m of the soil, respectively. When considering the upper 0.45 m of the soil profile, E sites contain 28% less SOC than the D sites. By combining the $^{137}$Cs-derived erosion and deposition rates with SOC-stock measurements (Table 1) of the surface layer that is exposed to erosion (i.e. the 0.15 m soil layer), we estimate that the E sites experience SOC losses of $0.42 \pm 0.04$ t C ha$^{-1}$ yr$^{-1}$, while the D sites gain $0.37 \pm 0.16$ t C ha$^{-1}$ yr$^{-1}$. Due to the large difference in the spatial extent of both erosion and deposition, net SOC export by erosion ($C_{exp}$) from the hillslope is $0.36 \pm 0.16$ t C ha$^{-1}$ yr$^{-1}$. This amounts to an annual loss of SOC equivalent to 3% of the stock currently present in the upper 0.15 m of the soil profile.

Spatial & temporal patterns of in-situ soil respiration ($R_h$)

Recognizing the widely observed control on SOC mineralization exercised by soil moisture, temperature, soil organic carbon stocks and microbial activity (e.g. Davidson & Janssens, 2006; Wiaux et al., 2014), we analyzed the power of these parameters to explain the seasonal and spatial patterns in our respiration data. There was a significantly lower flux of soil CO$_2$ from E than D sites during all four seasons and this is reflected in a 39% lower annual soil to atmosphere C flux from the E sites (i.e., $2.3 \pm 0.1$ t CO$_2$ C ha$^{-1}$ yr$^{-1}$) than from the D sites (i.e., $3.8 \pm 0.1$ t CO$_2$ C ha$^{-1}$ yr$^{-1}$) (Table 1 and Figure 3). At all sites CO$_2$ fluxes were highest in the summer quarter, which accounted for more than 40% of annual totals (Table 1). Within individual time-series for the 14 monitoring sites, soil temperature and moisture were the key factors controlling soil respiration and, together, explained 31.5% of the observed variability of in-situ soil respiration. Nevertheless, these factors do not explain the observed (spatial) differences in soil respiration between E and D sites (Table 1, Fig. B.1).

In order to analyse further the spatial variation observed across the 14 monitoring sites, soil CO$_2$ fluxes for each site were integrated over the period of observation and relationships with ‘static’ soil variables were explored. The main factors correlated with time-averaged soil C fluxes were found to
be: (i) the SOC stock in the upper 0.15 and 0.45m of the soil profile ($r=0.66$, $p<0.05$ and $r = 0.94$, $p<0.01$ respectively); (ii) the microbial biomass C per unit soil in the upper 0.15m (MBC, $r=0.85$, $p<0.01$); (iii) the soil bulk density for the upper 15cm of the soil profile ($r = -0.87$, $p < 0.01$); and, (iv) the soil temperature in the upper 5cm ($r=-0.56$, $p<0.05$) and moisture in the upper 10cm ($r=-0.39$) (Table B.1). Linear regression and analysis of variance showed that, when the effect of the 0-0.45m SOC inventory was accounted for, no other variables significantly contributed to the explanation of variations in time-averaged respiration rates. This demonstrates the strong control on soil-atmosphere C fluxes exerted by the magnitude of the SOC inventory. The statistically stronger relationship between the C flux and the 0-0.45m SOC inventory, compared to that between C flux and the 0-0.15 m SOC inventory, suggests that subsoil SOC mineralization contributes significantly to the soil-atmosphere C flux and may reflect the role of deeper soil in regulating the soil moisture of the whole profile.

Our data allow only inference of the processes controlling the variations in C fluxes. As has been stated, the observed fluxes are proportional to the measured 0-0.45m SOC inventory, suggesting that the amount of SOC available is the main control on respiration rates. Reduction in SOC stocks at eroding sites limits the respiration potential, an effect that may be exacerbated by lower microbial population size (Table 1), both per unit SOC and per unit mass of soil. Conversely, greater respiration rates are found at the depositional sites where greater microbial population size (cf. MBC) is observed and where deposition of soil and SOC has increased the local SOC stock. The lower MBC observed at erosion sites is attributed to the advection, from deeper soil layers with ongoing truncation, of SOC that is less accessible to microbial populations. The latter may be due to organo-mineral association of the SOC (Berhe et al., 2012; Berhe and Kleber, 2013). The negative relationship between respiration and soil bulk density (BD c. 10% higher at eroding sites, $p<0.01$) may reflect the reduced CO₂ and O₂ diffusion through the soil profiles with higher bulk densities (Wan, 1998).

Erosion and deposition controls on respiration ($r_e$ & $r_d$)
The magnitude of SOC stocks, the key factor controlling the observed spatial variation in respiration, is related to the redistribution of SOC by erosion and deposition (see Supplementary Online Materials) and we found a strong functional relationship between SOC redistribution rates (negative values representing SOC erosion) and soil CO$_2$ flux rates from the 14 sites (Figure SOM 2.2). We use this relationship to estimate the change in respiration ($r$) due to erosion referred to above. The intercept of the relationship ($2.72 \pm 0.13$ t CO$_2$-C ha$^{-1}$ yr$^{-1}$) represents the CO$_2$ flux at stable (no erosion or deposition) locations on the hillslope; we use this to represent respiration in a non-eroding state (R0). This allows us to derive a functional relationship in Figure 4, in which the suppression of CO$_2$ respiration at eroding sites, relative to non-eroding conditions, is described by the relation $r_e = (0.99 \pm 0.13)C_e$; while the enhanced release at D sites is described by $r_d = (1.87 \pm 0.13) *C_d^{0.28}$, where $C_e$ and $C_d$ are SOC erosion (negative) and deposition (positive), respectively.

**Carbon Inputs ($I_h$ & $i$)**

The impact of erosion and deposition on C input ($i$) was assessed using measurements (made in 2011 and 2012, Table 2). We observed no significant differences in biomass production between erosional and depositional sites. These results suggest that the effect of erosion on yields is small when there are no constraints on root development and soils are fertilized. This is consistent with regional estimates (Huang et al., 2007). Similar to other regions in China, the NPP on cropland on the Loess Plateau increased by 245% between 1950 and 2000 (Huang et al., 2007). Despite elevated rates of erosion typical of the Loess Plateau, this suggests that reductions in crop yield in response to erosion are negligible, relative to the increase in NPP in response to fertilization and plant selection. Based on our direct observations and information derived from regional statistics, and the assumption that C inputs are proportionally related to above ground biomass, we therefore estimate that $i \approx 0$.

4. Discussion:

4.1 SOC stocks in dynamic equilibrium and steady-state flux?
We proposed to test the hypothesis that under sustained high magnitude erosional forcing, SOC stocks reach dynamic equilibrium in which eroded SOC is replaced, the export SOC flux is in steady state and a potential SOC sink is created. We defined the condition for identifying dynamic equilibrium and a steady-state flux as $C_{\text{exp}} = i-r$. Our results indicate that there is a net reduction of in-situ soil CO$_2$ emissions induced by SOC erosion, such that $r_e = 0.43 \pm 0.15$ t CO$_2$-C ha$^{-1}$ yr$^{-1}$. When considering the whole hillslope, the integrated effect of the suppression of respiration at sites of erosion and the enhancement of respiration at sites of deposition, is a net reduction of in-situ soil CO$_2$ emission $(r = r_e \cdot A_e + r_d \cdot A_d)$, such that $r = 0.32 \pm 0.22$ t C ha$^{-1}$ yr$^{-1}$. We find evidence that $i \approx 0$ and, therefore, the net C exchange (i.e. $i-r$) between the soil and atmosphere as a result of erosion over the hillslope equates to a net sink of $0.32 \pm 0.22$ t C ha$^{-1}$ yr$^{-1}$. This sink is similar in magnitude to the independently derived value of $C_{\text{exp}}$ (C export from the hillslope) of $0.36 \pm 0.16$ t C ha$^{-1}$ yr$^{-1}$. Thus, $C_{\text{exp}} \approx i-r (=C_{\text{atm}})$ and the evidence is consistent with the SOC stocks (upper 0.45 m) of this unrestored eroding hillslope being in dynamic equilibrium, whereby continued plant input fully compensates for the erosional C losses. Furthermore, this evidence is consistent with the hypothesis that sustained high magnitude erosional forcing generates a steady state carbon flux. We note that our conclusions are also consistent with regional-scale carbon stock assessments on the croplands of the Loess Plateau in which no significant changes in SOC were detected between 1980 and 2000 (Yu et al., 2009).

Although our data shows that the eroding component of the hillslope is in steady-state (i.e. $C_e = r_e$), the net C balance for our site at the hillslope scale is strongly influenced by the fact that the depositional component of the hillslope, where respiration losses ($R_d$) are considerable higher (Fig 2), is only 10% of the hillslope area. Although the limited potential for re-deposition within the field is typical for the Loess Plateau (Ran et al., 2014), it is acknowledged that the sink term would be reduced if a greater proportion of SOC was deposited on the hillslope.

Significant fluxes of organic carbon have been identified from non-agricultural landscapes subject to natural erosion due to overland flow (Smith et al., 2013) and landslides across climatic environments from temperate New Zealand (Gomez et al., 2003, WRR; Hilton et al., 2008, GBC) to sub-tropical Taiwan (Kao and Liu, 2000, GBC; Hilton et al., 2012, GBC). The fluxes observed are typically of
order 1% of NPP and are implicitly assumed to be in steady-state, although this assumption remains
to be tested. In the system observed here, the export flux represents an order of magnitude greater
proportion of NPP (ca 10%, Table 2) and we have demonstrated that the SOC stocks on the slopes are
in dynamic equilibrium and, therefore, have strong evidence that the export flux is in steady state.

4.2 Broader implications

If the C dynamics of our study site are representative of the highly eroded Loess Plateau, and a
maximum C sink equal to C export from the slopes is the norm, then the broader implications of these
findings for land management strategies bear examination. During the last decades, rehabilitation
programs have reduced soil and C erosion rates from c. 5 million ha by one order of magnitude
through re-establishing a protective cover on the soil surface (Ren, 2006; Tang, 1993; El Kateb et al.,
2013). Conversion to planted forest and grassland reduces erosion by 83% and 76%, respectively
(Tang, 1993), while on steeper slopes, reduction factors are higher and are 99% and 91% for forest
and grassland, respectively (El Kateb et al., 2013). The efficacy of these rehabilitation measures,
alongside those to address sediment supply from gully walls, explains partly the reduction in the
suspended sediment load of the Huang He that is currently ca. 1/3 of that documented for the period
prior to the 1980s (Walling, 2009). Initially, these restoration programs focused on synergies between
soil conservation, food production and socio-economic welfare; however, more recently, the scope of
these programs has been extended to include sequestration of carbon (C) by soils and vegetation
(Chen et al., 2007; Lu et al., 2012). Implicit in this strategy is a desire to offset sequestration against
carbon-emitting activities. It has been suggested that the conversion of 4.8 million ha of the most
eroded cropland to forest and grassland has substantially reduced soil and C export from the plateau
and has led to a sequestration of c. 0.3 t C ha⁻¹ yr⁻¹ in soils of the Loess Plateau between 2000 and
2008 (Lu et al., 2012). While it is clear that these programs have enhanced ecosystem service delivery
and led to increases in soil C stocks, the quantification of the net C sink strength is associated with
considerable uncertainty because, implicit in the statement of sink strength, is the assumption that
there was no net C sink in the unrestored loess plateau. As is clear from the discussion above, this is
contrary to our findings. We can explore this further by reference to published soil C and erosion rate
data.

Prior to restoration, cropland soil erosion rates observed on steep lands in the Loess Plateau ranged
between 50 and 80 t ha\(^{-1}\) yr\(^{-1}\), and, typically, 60-100\% of this eroded sediment was exported from the
hillslopes (Zhang et al., 1994; 1997; 1998; 2002). Net soil export from the unrestored cropland,
therefore, ranged between 30 and 80 t ha\(^{-1}\) yr\(^{-1}\) and, assuming a mean SOC content of 0.4-0.6\% for the
plough layer (Chen et al. 2007; Li et al., 2007), this equates to a flux of C from the croplands of c.
0.12 - 0.48 t C ha\(^{-1}\) yr\(^{-1}\). This is of similar magnitude to the C sequestration flux achieved through
restoration. The erosional C export flux on unrestored cropland is therefore substantial. The stead-
state C export flux identified here implies that the restoration programs, by reducing C erosion, also
‘switched off’ the erosion-induced C sink discussed above. Hence, the net soil C sink benefit from the
restoration program, estimated at c. 0.3 t C ha\(^{-1}\) yr\(^{-1}\) (Lu et al., 2012), may have been substantially
overestimated at the scale of eroding hillslopes. Naturally, the actual C sink magnitude associated
with erosion prior to restoration was also controlled by the fate of the exported SOC and this in turn is
dependent on the degree of protection of the SOC, its composition, fluvial transfer times
(Aufdenkampe et al. 2010), burial rates and oxygen availability (Galy et al., 2007) and the
environment of deposition (Ran et al., 2014). In the absence of watershed-wide monitoring schemes it
is not possible to comment on these factors; however, there is evidence of preservation of buried C in
check dams (Wang et al., 2008), floodplain and riverbank sediments (Qin et al., 2007) and delta
deposits (Wang et al., 2012). Therefore, it is reasonable to consider the possibility that at least a
fraction of the exported C can be preserved (Ran et al., 2014). If this has been the case, the actual sink
strength of the restoration programmes may be smaller than former estimates. Therefore, while we
celebrate the success of the ecological programs in delivering effective C sequestration in hillslope
soils and the many co-benefits this brings for soil quality and delivery of ecosystem services, we urge
cautions in using the C sequestration in Loess Plateau soils to offset C-emitting activities because of
the uncertainty that exists with respect to long-term net gain in terrestrial C storage.
5. Conclusions

Our analysis provides the first empirical evidence in support of the hypothesis that under sustained high magnitude erosional forcing, SOC stocks reach dynamic equilibrium in which eroded SOC is replaced, the export SOC flux is in steady state and a net sink for atmospheric CO$_2$ is created at the scale of eroding hillslopes. At our study site, we found that all of the eroded organic C was replaced and that the sink strength was equal to the C export rate. This sets the upper limit of the erosion-induced sink term at the scale of whole watershed. The fate of the exported carbon in reservoirs, floodplains, riverbeds and the ocean will ultimately control the watershed-scale sink strength. However, the full replacement of eroded SOC observed here suggests that erosion does not induce a C source, irrespective of the fate of the exported carbon, at least for high-input agricultural systems (ignoring C emissions embedded in fertilizers). Finally, we propose that assessment of the C cycle perturbation associated with erosion-induced lateral C fluxes must be made an integral part of accounting mechanisms for climate change mitigation strategies that are based on land use change and C sequestration in terrestrial environments (Lu et al., 2012; Chen et al., 2007; Piao et al., 2009; IPCC, 2006; Chappell et al., 2012; Tang & Nan, 2013).


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Figure 1. Conceptual framework of C flux components $C_e$, $C_d$ and $C_{\text{exp}}$ (t C ha$^{-1}$ yr$^{-1}$) and differences in fluxes with respect to stable conditions $i$, $r_e$ and $r_d$ (t C ha$^{-1}$ yr$^{-1}$). $I$ and $R$ indicate C input and respiration for stable (subscript 0) and eroding (subscript h) conditions. For the differences with respect to stable conditions, $i$ indicates difference in annual crop C input to soil related to erosion; $r_e$ and $r_d$ indicate, respectively, differences in net annual rates of soil CO$_2$ respiration from eroded sites and deposited sites. For the respiration values, the negative value indicates suppression and the positive value indicates enhancement, suggesting the reduction of soil CO$_2$ respiration and the positive suggesting increase of soil respiration, relative to no erosion or deposition sites; For the lateral C fluxes, $C_e$ is the mean SOC export from the eroding part of the slope, $C_d$ is SOC accumulation at the depositional footslope; $C_{\text{exp}}$ is the net SOC export by erosion from the hillslope. $A_e$ and $A_d$ are the fractional areas of the hillslope experiencing erosion and deposition, respectively.
Figure 2. Topographical map of the study site (altitude in m above sea level). The black points indicate the sampling sites for SOC and $^{137}$Cs, red points indicate the positions of the soil respiration monitoring, and the black circles indicate the observation sites of CO$_2$ removed from analysis.
Figure 3. Effect of soil redistribution on cumulative in situ CO$_2$ fluxes for two classes (i.e. eroding (E) and depositional (D) sites). Error bars indicate the spatial variability of observed fluxes from E (n=10) and D (n=4) sites. Further site information is provided in Table 1.
Figure 4. Relationship between the annual in-situ soil CO$_2$ flux, relative to the flux at non-eroding/non-depositional sites (r; note separate relationships shows for re and rd), and SOC redistribution (i.e. Ce and Cd) from the 14 monitoring sites. The error bars represent the standard errors. The areas between the two dashed lines indicate the 95% confidence prediction range.

\[ y = 1.87x^{0.28} \]

\[ R^2 = 0.90, P<0.05 \]

\[ y = 0.99x \]

\[ R^2 = 0.63, P<0.01 \]
Table 1. Summarized statistics of CO₂ fluxes, $^{137}$Cs and SOC redistribution rates.

<table>
<thead>
<tr>
<th></th>
<th>E</th>
<th>D</th>
<th>Entire slope</th>
</tr>
</thead>
<tbody>
<tr>
<td>CO₂ fluxes/soil moisture</td>
<td>10518/7740</td>
<td>2778/2045</td>
<td>13,296/9785</td>
</tr>
<tr>
<td>soil temperature</td>
<td>10518</td>
<td>2778</td>
<td>13,296</td>
</tr>
<tr>
<td>$^{137}$Cs/SOC</td>
<td>49/49</td>
<td>4/4</td>
<td>53</td>
</tr>
<tr>
<td>MBC</td>
<td>6</td>
<td>3</td>
<td>9</td>
</tr>
<tr>
<td>$^{137}$Cs residuals (%)</td>
<td>-79.25±1.86b</td>
<td>63.16±26.30a</td>
<td>-68.50±26.37</td>
</tr>
<tr>
<td>Soil E/D (t ha⁻¹ yr⁻¹)</td>
<td>-73.12±5.84b</td>
<td>66.81±28.66a</td>
<td>-62.56±29.25</td>
</tr>
<tr>
<td>SOC E/D (t C ha⁻¹ yr⁻¹)</td>
<td>-0.42±0.04b</td>
<td>0.37±0.16a</td>
<td>-0.36±0.16</td>
</tr>
<tr>
<td>CO₂ fluxes</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Spring (t CO₂ C ha⁻¹)</td>
<td>-0.50±0.04b</td>
<td>-0.69±0.09a</td>
<td>-0.52±0.10</td>
</tr>
<tr>
<td>Summer (t CO₂ C ha⁻¹)</td>
<td>-0.93±0.02b</td>
<td>-1.65±0.09a</td>
<td>-0.98±0.09</td>
</tr>
<tr>
<td>Autumn (t CO₂ C ha⁻¹)</td>
<td>-0.53±0.01b</td>
<td>-0.93±0.08a</td>
<td>-0.56±0.08</td>
</tr>
<tr>
<td>Winter (t CO₂ C ha⁻¹)</td>
<td>-0.33±0.02b</td>
<td>-0.51±0.12a</td>
<td>-0.34±0.12</td>
</tr>
<tr>
<td>Annual (t CO₂ C ha⁻¹)</td>
<td>-2.29±0.08b</td>
<td>-3.78±0.09a</td>
<td>-2.40±0.12</td>
</tr>
<tr>
<td>Temperature (℃)</td>
<td>18.52±0.14 a</td>
<td>18.24±0.26 a</td>
<td>18.55±0.30</td>
</tr>
<tr>
<td>Moisture (%)</td>
<td>16.11±0.12a</td>
<td>15.96±0.21a</td>
<td>16.08±0.24</td>
</tr>
<tr>
<td>SOC_t0-15 cm</td>
<td>11.54±0.29 b</td>
<td>14.07±0.39 a</td>
<td>11.74±0.49</td>
</tr>
<tr>
<td>SOC_t0-45 cm</td>
<td>22.69±0.70b</td>
<td>30.49±0.30a</td>
<td>22.73±0.76</td>
</tr>
<tr>
<td>MBC (mg kg⁻¹) t-15 cm</td>
<td>159.99±36.95b</td>
<td>348.59±26.35a</td>
<td>222.86±45.38</td>
</tr>
</tbody>
</table>

§ E and D mean erosion and deposition, respectively.

† Spring, Summer, Autumn and Winter indicate the periods of March to May, June to August, September to November, and December to February during the April 2007 to September 2008, respectively.

F Figures followed by the same letters within a row are not significantly different at p = 0.05 probability based on one-way analysis of variance (ANOVA).
Table 2: Mean and standard errors for biomass of crop (wheat), npp and carbon input measured in 2011 and 2012.

<table>
<thead>
<tr>
<th>Slope</th>
<th>Erosion (E)/Deposition (D)</th>
<th>Biomass (t ha⁻¹ yr⁻¹)</th>
<th>NPP (t C ha⁻¹ yr⁻¹)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>E</td>
<td>5.39 ±1.60a*</td>
<td>4.31 ±1.28</td>
</tr>
<tr>
<td></td>
<td>D</td>
<td>4.52 ±0.82a</td>
<td>3.62 ±0.66</td>
</tr>
<tr>
<td>2</td>
<td>E</td>
<td>3.98 ±0.86a</td>
<td>3.18 ±0.69</td>
</tr>
<tr>
<td></td>
<td>D</td>
<td>5.47 ±1.35a</td>
<td>4.37 ±1.08</td>
</tr>
<tr>
<td>3</td>
<td>E</td>
<td>5.50 ±1.18a</td>
<td>4.40 ±0.94</td>
</tr>
<tr>
<td></td>
<td>D</td>
<td>5.61 ±1.39a</td>
<td>4.49 ±1.11</td>
</tr>
</tbody>
</table>

*Figures followed by the same letters within a column are not significantly different between E and D sites for the same slope at p = 0.05 probability based on one-way analysis of variance (ANOVA).
*Highlights (for review)*

- SOC respiration from eroded soils is suppressed in proportion to soil loss
- Respiration suppression due to erosion is of equal magnitude to SOC export
- The SOC content of eroded Loess Plateau soils is in dynamic equilibrium
- Unless entirely mineralised during transport exported SOC can create a net C sink
- The potential erosional C sink is of similar magnitude to the restoration C sink
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