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Angaben zur Veröffentlichung / Publication details:

Hoffmann, T., S. M. Mudd, K. van Oost, G. Verstraeten, G. Erkens, A. Lang, H. Middelkoop, et al. 2013. "Short Communication: Humans and the missing C-sink: erosion and burial of soil carbon through time." *Earth Surface Dynamics* 1 (1): 45–52. https://doi.org/10.5194/esurf-1-45-2013.

Earth Surf. Dynam., 1, 45–52, 2013
www.earth-surf-dynam.net/1/45/2013/
doi:10.5194/esurf-1-45-2013

Earth Surface Dynamics
Open Access

Short Communication: Humans and the missing C-sink: erosion and burial of soil carbon through time

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Received: 9 June 2013 – Published in Earth Surf. Dynam. Discuss.: 24 June 2013
Revised: 31 October 2013 – Accepted: 6 November 2013 – Published: 26 November 2013

Abstract. Is anthropogenic soil erosion a sink or source of atmospheric carbon? The answer depends on factors beyond hillslope erosion alone because the probable fate of mobilized soil carbon evolves as it traverses the fluvial system. The transit path, residence times, and the resulting mechanisms of C-loss or gain change significantly down-basin and are currently difficult to predict as soils erode and floodplains evolve – this should be a key focus of future research.

1 Introduction

A considerable fraction of human-induced carbon dioxide (CO_2) emissions due to fossil fuel combustion and land cover change is absorbed by the oceans and the terrestrial biosphere. Soils store about 80 % or 2500 Pg C of the total carbon (C) in the terrestrial biosphere, and thus contain three times more C than the atmosphere (Lal, 2004). Consequently, soil C represents a substantial and highly sensitive component within the global carbon cycle and small changes in the soil C may result in large changes of atmospheric CO_2 at timescales of 10^1 to 10^3 yr.

During the last decade, research has highlighted the importance of vertical exchanges of carbon between the atmosphere, biosphere and pedosphere (e.g. Ruddiman, 2003). Little attention was devoted to the full "life cycle" of the eroded minerals from soil to sea including the lateral fluxes of organic carbon (OC) associated with human-induced soil erosion. In many regions of the world, soils have increasingly been eroded (Wilkinson and McElroy, 2007), owing to increasing agricultural activity during the last few thousand years, with negative impacts on soil fertility and productivity as well as strongly increased lateral sediment-burden carbon fluxes. Therefore, sediment-burden carbon fluxes potentially provide an important, yet unknown, component in the global carbon cycle. As yet, lateral C-fluxes on hillslopes (Quinton et al., 2010; Starr et al., 2000; Van Oost et al., 2007) and in river channels (Aufdenkampe et al., 2011; Battin et al., 2008, 2009; Cole et al., 2007) (Fig. 1) have received little

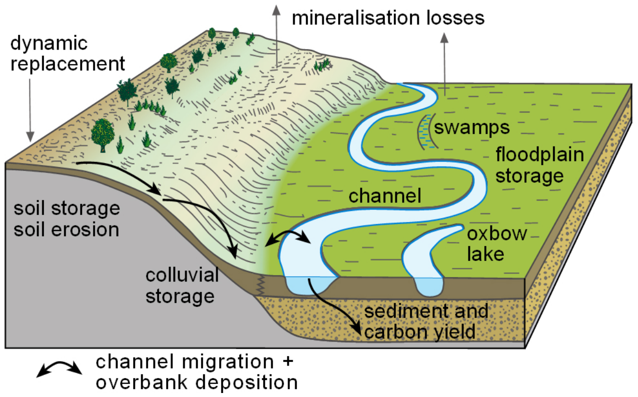

Figure 1. Storage compartments and lateral and vertical C-fluxes in agricultural landscapes.

attention, and their representation in contemporary carbon cycle models is rudimentary: soils are typically represented as spatially homogeneous and static entities, which does not reflect their dynamic nature in response to anthropogenic disturbance. Hillslopes are often regarded as simple engines that sequester or release carbon through soil formation and subsequent erosion. River channels are generally viewed as passive pipes that flush carbon from the hillslopes to the oceans (Battin et al., 2009), with floodplains that are mostly considered as wetlands that trap sediment and nutrients from the river, sustain productive riparian vegetation and hence induce higher CH_4 fluxes (Petrescu et al., 2010).

Sediment mobilized in river catchments by the introduction of agriculture has important implications for lateral carbon fluxes in fluvial systems. Soil erosion and carbon emissions are tightly linked through OC-stabilization by organo–mineral associations, requiring a profound understanding of the coupling between sediment and OC-erosion, as well as transport and storage. Three key mechanisms determine the coupling between sediment and carbon fluxes: (i) OC fixation to surfaces of mineral particles eroded on hillslopes and transported in river systems, (ii) increased mineralization of OC due to aggregate breakdown during transport, and (iii) stabilization of OC through burial (Fig. 1) (Van Oost et al., 2007). While soil scientists, geomorphologists and geochemists generally agree these are all key mechanisms, markedly different assumptions have been made about their relative importance. This has resulted in an intense debate as the inter-relation of these key mechanisms decides the significance of erosion-induced terrestrial carbon as sink or source (Berhe et al., 2007; Lal and Pimentel, 2008; Van Oost et al., 2007): strong mineralization acts as an atmospheric CO_2 source, while OC-fixation and burial provide a major atmospheric CO_2 sink. Because erosion, transport and depositional processes change along the sediment and carbon transport path from hillslopes to large river systems, the relative contributions of erosion induced OC-fixation, mineralization and OC-protection through burial change over space and time. Recent studies have focused on decadal timescale perturbations to the C-cycle via soil erosion (Jacinthe and Lal, 2001; Ritchie and Rasmussen, 2000; Van Oost et al., 2007), yet it has emerged that sediments are stored on hillslopes and floodplains for several thousand years, meaning that human-induced changes to sediment-driven carbon fluxes can be delayed and altered over millennial timescales (Hoffmann et al., 2009a; Van Oost et al., 2012). Accounting for the non-steady-state C-dynamics along the flow path from hillslopes to river channels and then into the oceans is thus pertinent to understanding both the past and future global C-cycle. Here, we review the current understanding of sediment and carbon dynamics on hillslopes and in river channels and encourage an integrated view of long-term sediment-associated carbon dynamics.

2 Hillslope erosion and deposition

Hillslopes are major source areas of sediment and carbon. Estimates of global soil erosion on agricultural land vary between 28.1 and 150 $Pg\,a^{-1}$. Most recently Quinton et al. (2010) estimated the total contemporary erosion rate (including water, tillage and wind erosion) at $35 \pm 10\,Pg\,a^{-1}$, which corresponds to a carbon erosion rate of $0.5 \pm 0.15\,Pg\,a^{-1}$. These rates change through time as a result of population density, cultivation techniques and climate (Hoffmann et al., 2009a; Notebaert et al., 2011; Trimble, 1999). Eroded sites are usually characterised by lower soil fertility, crop productivity and C-contents than non-eroded sites supporting the notion that soil erosion is a major source of atmospheric CO_2 (Jacinthe and Lal, 2001; Lal, 2005; Victoria et al., 2012). Yet, OC-loss is partially balanced by OC-input through plants and fertilization and it may thus be argued that OC-fixation changes in concert with erosion rates at timescales covering the period of agriculture.

Estimates of soil and carbon erosion rates on regional to global scales in agriculturally developed river systems either rely on erosion plot studies or on measured sediment discharges. Plot studies, however, measure the amount of soil that is moved on the fields, and thus overestimate the "loss" of soils to river channels, which is only a small proportion of the eroded soil. In contrast, sediment discharges at river gauging stations estimate the sediment or OC-yield (efflux) that is transported beyond the gauging station, but do not quantify any sediment deposition upstream. Major flux differences found between these two approaches indicate that a large fraction of the detached soil is transported only a limited distance (Parsons et al., 2006) and that a substantial amount of sediment is stored at the foot of hillslopes: i.e. the transition between the hillslopes and the channels (Houben, 2008; Verstraeten et al., 2009). Thus, considering hillslopes simply as sources of sediment and carbon neglects (i) the differences between soil erosion and sediment yield (Dotterweich, 2008; Trimble, 1999; Verstraeten et al., 2009) and (ii) the complexity of the internal sediment dynamics at the hillslope scale (Lang and Honscheidt, 1999; Cerdà et al., 2012). This simplification can account for the major discrepancies in the current estimates of the effects of soil erosion: Lal (2005) indicate that 20–30 % of eroded OC is released through erosion-induced mineralization, whereas Van Oost et al. (2007) derive an order of magnitude lower mineralization rates of only ~2 % based on ^{137}Cs inventories within fields.

During the last two decades, long-term sediment budgets (Hoffmann et al., 2010; Brown et al., 2009), which can account for some of the sources, pathways and sedimentary sinks, have increasingly shed light on the response of hillslope systems to human-induced soil erosion (Lang et al., 2003). These budgets provide an organizing framework to answer the following open research questions: How much eroded carbon is stored in colluvial sediments, and what are typical residence times of carbon and sediment on hillslopes? Sediment redistribution across hillslopes (as estimated by soil erosion plots) does not affect the downstream channel system, but does it have a significant impact on the global carbon cycle? How far must sediment be transported until it is depleted in carbon? Possible approaches to answer these questions are presented by Wang et al. (2010), who suggest OC-enrichment of exported sediment due to grain size sorting and the association of OC to fine grain fraction (e.g. clays), and by Van Oost et al. (2012) stating that 50 % of colluvial OC is decomposed within 500 yr. Yet, major uncertainties on the role of the hillslope sediment dynamics in the global carbon cycle remain (Aufdenkampe et al., 2011; Brantley et al., 2011; Yoo et al., 2011).

3 Fluvial transport and deposition

Recent studies have highlighted the role of rivers (including small streams, lakes, artificial reservoirs and wetlands) not only in transporting the C exported from terrestrial ecosystems but also in metabolising and burying significant amounts of C (Aufdenkampe et al., 2011; Battin et al., 2008, 2009; Cole et al., 2007; Tranvik et al., 2009). Globally, rivers receive about 2.9 Pg C each year, a quantity that represents the differences between global annual terrestrial production and respiration (Tranvik et al., 2009; Aufdenkampe et al., 2011). A major part of this carbon is associated with sediments mobilized through surface runoff. Only a fraction of this carbon is transported directly into the oceans (0.9 Pg C a^{-1}) while the majority is mineralized or outgassed to the atmosphere (1.4 Pg C a^{-1}), or buried in lakes and reservoirs (0.6 Pg a^{-1}) (Tranvik et al., 2009). Still, long-term C-burial in floodplains and subsequent outgassing are generally not considered (Battin et al., 2009; Cole et al., 2007; Tranvik et al., 2009) due to the limited availability of data on global floodplain extent, sedimentation rates, duration of inundation and gas exchange velocities between floodplains and the atmosphere.

First estimates of global C-burial on hillslopes and floodplains (Aufdenkampe et al., 2011) range between 0.5–1.5 Pg C a^{-1}. These numbers are derived from the difference between global soil erosion (50–150 Pg C a^{-1}) (Wilkinson and McElroy, 2007) and sediment delivery to the ocean (12.6 Pg a^{-1}) (Syvitski et al., 2005), which is multiplied by an average C-concentration of eroded and exported sediment (~1 %) (Aufdenkampe et al., 2011). Although informative, these numbers are not direct measurements, and account for neither temporal storage nor remobilization within the fluvial system. More explicit representation of floodplains and their impact on the global carbon cycle is essential because floodplains represent a major depositional environment (Aalto et al., 2003; Hoffmann et al., 2007; Noe and Hupp, 2005; Verstraeten et al., 2009). Based on a Holocene sediment budget of the Rhine basin, Hoffmann et al. (2013) estimated that 1.1 ± 0.5 Pg C is stored in the floodplains of the non-alpine part of the Rhine basin (i.e. 125 000 km^2), equivalent to a long-term OC-sequestration rate of 5.3 to 17.7 g C m^{-2} a^{-1}. While these rates are time-integrated Holocene averages, several lines of evidence support a significant increase of organic-rich overbank deposition during the late Holocene (Hoffmann et al., 2009a; Verstraeten et al., 2009). Pre-human background rates of overbank deposition in the Rhine are ~0.5 mm a^{-1}, corresponding to a C-sequestration of 8.3 g C m^{-2} a^{-1}. In contrast, maximum sedimentation rates during the last 300 yr indicate at least an order of magnitude increase to 15 mm a^{-1} or 166 g C m^{-2} a^{-1}. These high rates generally coincide with increased hillslope erosion resulting from agricultural intensification (Notebaert and Verstraeten, 2010; Hoffmann et al., 2009a).

High sequestration rates of 100 g C m^{-2} a^{-1} associated with soil formation in freshly deposited overbank deposits (Zehetner et al., 2009) imply that a large fraction of floodplain OC is not the result of sediment-burden carbon fluxes, but represents in-situ OC-formation. These high rates are maintained during the initial 100 yr after sediment deposition, and thus strongly conditioned by the input of fresh sediments that provide abundant mineral surface area for complexation of OC on timescales > 100 yr. As shown by studies on OC-burial in marine fans and shelves (Hilton et al., 2008; Galy et al., 2007), high sediment input not only favours OC-sequestration, but also increases the burial efficiency of OC in marine deposits.

The residence time and the stability of OC in floodplains are strongly related to the geomorphological and hydrological floodplain dynamics, depending on the dominant grain size of floodplain sediments and the structure and age of riparian vegetation. High energy, non-cohesive floodplains of headwater streams are characterised by coarse sediments, mobile channels, open vegetation cover, and large groundwater fluctuations. The high sediment transport capacity causes a low storage potential, and thus short residence times of floodplain OC. In contrast, low energy, cohesive floodplains are characterised by stable channel banks with massive and cohesive overbank deposits, and dense vegetation coverage. These represent major sedimentary sinks that are able to store sediment-burden carbon for several thousand years (Battin et al., 2008). Further important links between geomorphic dynamics and floodplain OC include changes of the groundwater tables, hydrological connectivity and river incision caused by river engineering and land cover change (Hupp et al., 2009; Noe and Hupp, 2005; Osterkamp et al., 2012).

The comparison of the burial efficiency (given by the rate of OC-accumulation minus the oxidation of OC within the sink) in different depositional settings suggests that sequestration of OC within floodplains exceeds that of lakes, artificial impoundments and afforestation of catchments (Fig. 2). The ubiquitous prevalence of floodplains and their comparable burial efficiency to peats highlights both their importance and their insufficient representation within global C-budgets. Despite growing awareness of the multiple sources of floodplain OC and the feedbacks between sediment and vegetation dynamics in floodplains, quantitative understanding of the functioning of floodplains as C-sources or sinks remains elusive, with specific implications for the global C-cycle are hardly considered.

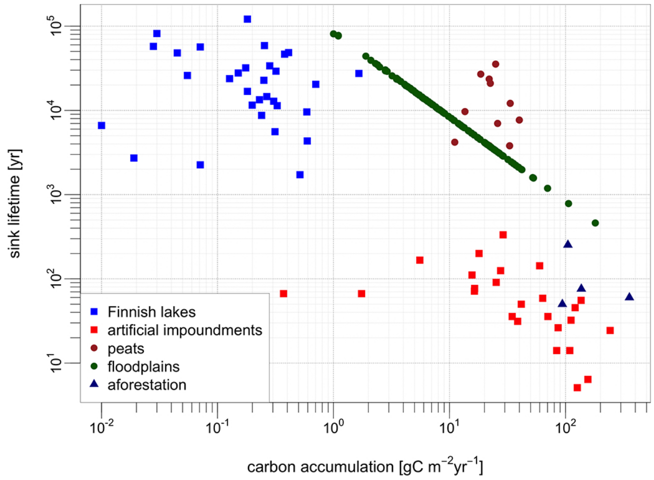

Figure 2. Lifetime versus sequestration rate of selected carbon pools. For details on data and calculations see Supplement.

4 Towards integrated biogeochemical and geomorphological approaches

Recently, major progress has been made in understanding land use and climate impacts on sediment dynamics through the study of Holocene sediment budgets (Hoffmann et al., 2010). These studies provided essential information on (i) the sensitivity of hillslopes and channel systems to environmental change (Notebaert and Verstraeten, 2010; Verstraeten et al., 2009); (ii) the storage, residence time and remobilization of sediment along the flow path (Houben et al., 2009; Hoffmann et al., 2007; Aalto and Nittrouer, 2012); (iii) the connectivity between hillslopes and channels (Verstraeten et al., 2009; Lang et al., 2003); and (iv) the resulting non-linear dynamics between soil erosion and sediment yield (Erkens et al., 2011; Van De Wiel and Coulthard, 2010). While long-term budgets portrayed changing sediment dynamics through time and are of great value to reconstruct variable sediment burden OC-fluxes, their potential is not yet fully exploited. This becomes apparent when comparing estimates of sediment-associated OC-erosion during the last 50 yr with long-term OC-burial studies. In contrast to the limited impact of soil erosion on atmospheric-C during the last 50 yr (Van Oost et al., 2007), terrestrial sediment storage presents an important long-term atmospheric C-sink (Fig. 3) (Hoffmann et al., 2009b; Van Oost et al., 2012). The discrepancies between short- and long-term OC-budgets and their implications highlight that mechanisms associated with changing conditions of OC-fixation and stabilization need to be considered more explicitly. These include the effects of changing hydrology, ecology (Osterkamp et al., 2012), and anthropogenically induced land cover change ALCC (Lechterbeck et al., 2009) during the Holocene. For instance, severe soil erosion in Mediterranean landscapes during the last Millennia changed soil covered regions into landscape of bare rock over large areas and thus resulted in almost irreversible changes of the ecological conditions (Fuchs et al., 2004; Lowdermilk, 1948). Dusar et al. (2011) and Marselli and Trincardi (2013), for instance, suggest that sediment yields in many Mediterranean landscapes declined during the last 1–2 ka as a consequence of the widespread soil depletion. Of major importance with respect to future global C-cycle is the declining capacity of the remaining soil to replace eroded OC as the extent and severity of soil degradation and desertification increases (Lal, 2009). In addition, the decay of buried OC in depositional settings is decreasing under drier climatic conditions as projected for the future decades (Solomon et al., 2007). Furthermore, deforestation increases not only soil erosion and sediment flux into sedimentary sinks, but also transforms the morphology of river channels (Walter and Merritts, 2008). Changing channel morphology involves, for example, (i) transformation of floodplains from stable channels in cohesive deposits to coarse sediments and mobile channels, and (ii) transitions from higher to low groundwater levels with corresponding changes from stabilized floodplain OC to destabilization and emission of large amounts of organic C to the atmosphere.

These conceptual considerations highlight the influence of indirect links between geomorphological processes and OC-fluxes that act on century-to-Holocene timescales. We believe that the on-going discussion about whether anthropogenic soil erosion is a sink or source of atmospheric carbon will not be solved until we synthesize biogeochemical

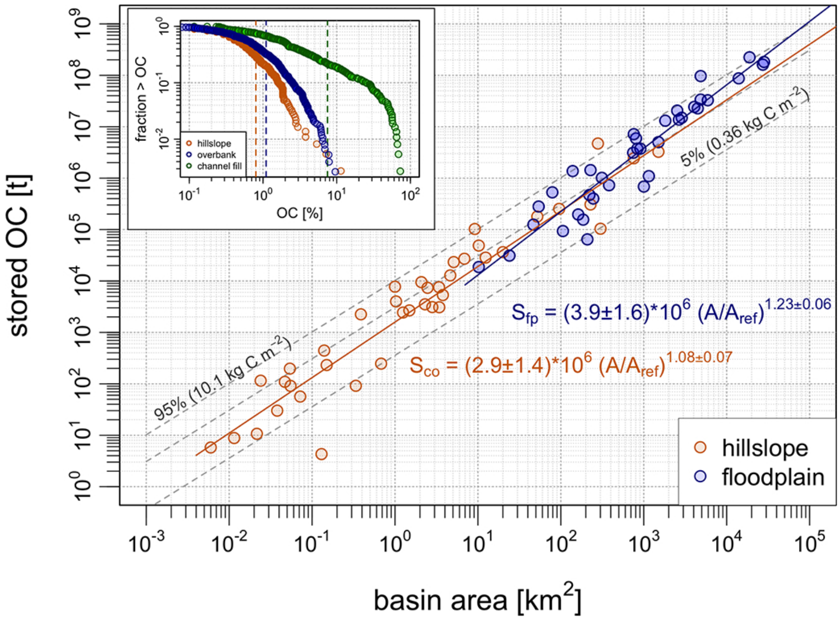

Figure 3. Holocene carbon storage on hillslopes and in floodplains as a function of basin size (Hoffmann et al., 2013). Lines represent predicted hillslope (S_{co}) and floodplain (S_{fp}) storage using power law regression with normalized basin area: $S = \alpha \times (A \times A_{\mathrm{ref}}^{-1})^{\beta}$. The scaling coefficient α represent OC-storage [t] at the reference basin area ($A_{\mathrm{ref}} = 1000\,\mathrm{km}^2$) and mainly reflect the greater OC-concentration of floodplain deposits (i.e. overbank and channel fills) compared to hillslope deposits (see inset: cumulative frequency distributions of OC-inventory from hillslope, and floodplain sediments). The scaling exponent of floodplain storage $\beta_{\mathrm{fp}} = 1.23 \pm 0.06$ indicates a greater increase of floodplain OC with basin than hillslope OC, which is given by $\beta_{\mathrm{co}} = 1.08 \pm 0.07$ slightly larger than one. These results are in accordance with an increasing accommodation space for floodplain storage with basin size. For details see Supplement.

and geomorphological approaches, thereby properly considering the fate of mobilized OC as it travels from the hillslopes through the river network at corresponding timescales of 10^0 to 10^4 yr. Such detailed studies are increasingly feasible using improved techniques for high-resolution geochronology of colluvial and fluvial deposits over timescales for decades to millennia (Aalto and Nittrouer, 2012; Chiverrell et al., 2008; Hobo et al., 2010).

To put our estimates of hillslope- and floodplain-OC storage into perspective (Fig. 3), we extracted C-emissions as a result of ALCC for the Rhine basin from the global modelling scenarios described in Kaplan et al. (2011, 2012) (Fig. 4). Our sediment storage data indicates that hillslopes and floodplains could have sequestered an amount of OC similar to the cumulative C-emissions from anthropogenic land cover change during the last 8000 yr. Thus, in this region, soil erosion associated with human-induced land-cover changes potentially offsets the effects of C-emissions caused by long-term ALCC. Furthermore, high accumulation rates and long residence times indicate that floodplains and peats represent the dominant terrestrial OC-sink, while OC-uptake in lakes, reservoirs and forests is limited due to either lower accumulation rates or shorter residence times in these pools (Fig. 2). Yet, the timescales of protection of eroded organic soil material are insufficiently understood (including the complex mechanisms of ensuing reburial and then further destabilization due to changing environmental conditions on hillslopes and in floodplains). Thus, whether anthropogenic soil erosion is a sink or source of atmospheric carbon is not dependent on hillslopes alone. Instead, the fate of mobilized soil carbon changes as it travels through the river network. The transit path and the resulting C-loss or gain dynamically changes and is complicated to predict as soils erode and floodplains evolve. In fact, hillslopes and floodplains are important components of the "boundless carbon cycle" (Battin et al., 2009), that introduce legacy effects into contemporary C-dynamics (Van Oost et al., 2012). Their functioning within the Holocene's boundless carbon cycle is heavily dependent on ALCC and catchment management, both of which require an appropriate understanding of changing environmental conditions on mineral sediments and associated carbon fluxes across spatial and temporal scales. A better accounting of hillslopes and floodplains (Fig. 3) as factors of global change demands integrated geomorphological and biogeochemical studies on long-term sediment-burden carbon fluxes under different environmental conditions (e.g. old vs. new world and temperate vs. semi-arid climate). Erosion-induced soil degradation and decline in biomass production should also be considered when evaluating the OC-budget, particularly in regions of the world where long-term land degradation is observed. Such research is now technically possible. Potential benefits from the study

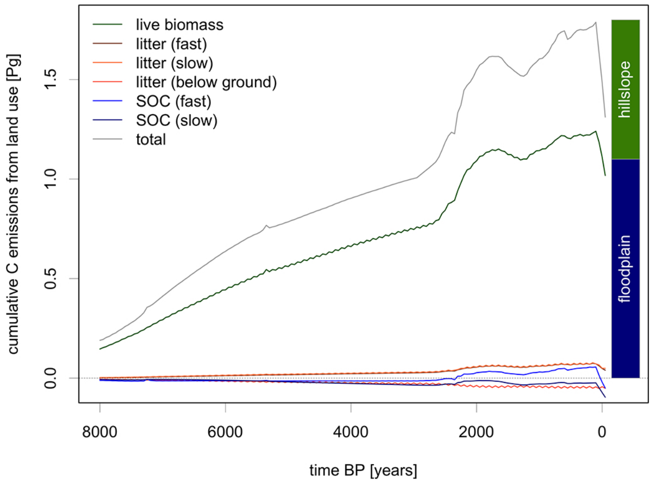

Figure 4. Cumulative carbon emissions as a result of anthropogenic land cover change in the non-alpine Rhine Basin. Emissions were calculated using the LPJ Dynamic Global Vegetation Model in the standard simulation using the KK10 anthropogenic land cover change described in Kaplan et al. (2011, 2012). Emissions accumulated steadily as a result of anthropogenic deforestation throughout most of the Holocene, with accelerated deforestation occurring during the Iron Age and Roman period around 500 BC–AD 500, followed by a period of land abandonment during the Migration Period, and accelerating deforestation during Medieval times. The period AD 1500–1900 is marked by relative stability in land use with little new deforestation emissions; during the 20th century land abandonment and afforestation lead to a strong uptake of carbon. Hillslope and floodplain storage are taken from Hoffmann et al. (2013) (compare Fig. 3).

of geomorphically coupled biogeochemical cycles would be substantial and help to (i) evaluate the wide-ranging implications of changing erosion and sediment dynamics for the global carbon budget, and (ii) assist policymakers to incorporate hillslopes and floodplains into catchment management strategies to mitigate climate change.

Supplementary material related to this article is available online at http://www.earth-surf-dynam.net/1/45/2013/esurf-1-45-2013-supplement.pdf.

Acknowledgements. This paper is a result of the open IGBP-PAGES-PHAROS workshop "Sediment and carbon fluxes under human impact and climate change", which was held in July 2011 in Bern, Switzerland. We would like to thank IGBP-PAGES for the financial support of the workshop. Furthermore, funding was provided for T. Hoffmann through the DFG-SFB 806 "Our way to Europe".

Edited by: A. Cerdà

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