Erosion, deposition and soil carbon: a review of process-level controls, experimental tools and models to address C cycling in dynamic landscapes

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Erosion, deposition and soil carbon: A review of process-level controls, experimental tools and models to address C cycling in dynamic landscapes

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A B S T R A C T

The role of soil erosion in terrestrial carbon (C) sequestration and release remains one of the most important uncertainties in our attempts to determine the potential of soils to mediate climate change. Despite its widely recognized importance for terrestrial C sequestration, to date, no Earth System Model (ESM) implements soil erosion effects on carbon cycling in sufficient detail. So far, available studies have mostly investigated the magnitude of erosional C transport and *in-situ* measurements of vertical C fluxes on the catchment or regional scale. Recognizing the need to adequately represent C erosion processes and controls in ESMs, we provide a comprehensive cross-disciplinary review on lateral C redistribution in the landscape and discuss the implications for biogeochemical cycling of carbon. We present current knowledge on the role of erosional C distribution in controlling the stabilization and release of C in soils, taking into consideration the important geomorphic, ecological, hydrologic, pedologic and micro-climatic processes and controls that affect soil organic carbon (SOC) stock, fluxes, and persistence in dynamic landscapes. Further, we provide an overview on latest experimental and modelling approaches that are being used to investigate the role of erosion in the carbon cycle. Finally, to advance our understanding of the role of soil redistribution in biogeochemical cycles of essential elements, we discuss the most promising topics for future research in this field.

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1. Introduction

1.1. Estimates of global soil and C fluxes

Soil organic carbon (SOC) is the largest terrestrial organic carbon pool. Global SOC assessments based on the Harmonized World Soil Database (FAO et al., 2009) reported estimates that soils contain 699 Pg C in the top 0.3 m and 1417 Pg C when that layer is extended to 1 m (Hiederer and Köchyl, 2012). Storing around four times more C than aboveground vegetation and three times more than the atmosphere (Jobbagy and Jackson, 2000; Sundquist, 1993; Schlesinger, 1990; Houghton, 2007), the important role of soils in the global C cycle particularly with respect to anticipated changes in climate and land use is increasingly acknowledged (Amundson et al., 2015). Under natural stable conditions, soil systems are considered to be sustainable, as the loss of soil material by erosion from a given area (i.e. the loss of soil, its redistribution over the landscape and its export to river systems) is approximately balanced by the production of soil as a result of weathering (Montgomery, 2007; Verheijen et al., 2009). However, the conversion from natural to agricultural land use removes the protective cover of natural vegetation and this typically increases soil erosion by one to two orders of magnitude (Montgomery, 2007), thereby accelerating physical erosion well over equilibrium levels.

A recent review on the topic by Kirkels et al. (2014) summarized important aspects of soil (C) redistribution, viewed from a geomorphologist's perspective and focusing on the physical aspects of soil fluxes. Human activities have not only drastically accelerated soil redistribution rates but have also altered patterns of erosion and deposition (i.e. Gregorich et al., 1998, Wilkinson and McElroy, 2007) (Table 1). For example, soil loss by tillage erosion can, especially under highly intensive and mechanized land management regimes, reach the same magnitude as soil loss by water erosion. This further increases the severity of harmful on- and offsite environmental effects (Van Oost et al., 2006), such as the loss of fertile soil on cropland and the increase of sediment load in river systems. While the highest rates of water erosion occur where water accumulates along drainage pathways, tillage erosion is highest at the uphill convex shoulder of the hillslope and back-slope positions. Most (70–90%) of the eroded topsoil material is ultimately deposited downhill within the same source or in adjacent watersheds (Stallard, 1998). Some of the eroded soil material fills concave drainage pathways and can be transported further

Table 1
Global rates of accelerated, agricultural erosion, as assessed by various methods and studies. (Plot = Plot scale extrapolations; SDR = Sediment delivery ratios; Model = Empirical and mechanistic modeling approaches.)

Method	Source	Global human-induced soil erosion (Pg y^{-1})
Plot	Myers, 1993	50
	Pimentel et al., 1995	74
	Hooke, 2000	120
SDR	Wilkinson and McElroy, 2007	60
	Berhe et al., 2007	75
Model	Stallard, 1998	24–65
	Lal, 2003	134
	Yang et al., 2003	81
	Ito, 2007	140
	Van Oost et al., 2007	30–40
	Doetterl et al., 2012b	12–34

downslope (Li et al., 2008). Anthropogenically accelerated erosion rates, hence, have resulted in substantial redistribution of SOC over the landscape and also in export to water bodies (Galy et al., 2015).

1.2. Erosion and terrestrial carbon sequestration: C sink or source?

In environments with high rates of soil redistribution, such as agroecosystems, soils can exhibit a large variability in SOC content due to removal of C-rich topsoil from eroding slope positions and burial in colluvial foothills (and similar depositional landform positions) (Berhe et al., 2008; De Gryze et al., 2008; Gregorich et al., 1998; Heckrath et al., 2006; Nadeu et al., 2012; Quine and Zhang, 2002; Schwanghart and Jarmer, 2011; Van Oost et al., 2005a; Vandenbygaart et al., 2012; Wang et al., 2010). The amount of SOC that is laterally distributed by soil erosion globally, predominantly through water erosion, is estimated to be between 0.3 and 5 Gt C y^{-1} (Berhe et al., 2007; Chappell et al., 2015). These estimates, based on spatially explicit data, are already significantly lower than older assessments based on plot extrapolations, which are often biased towards eroding landscape positions (Auerswald et al., 2009; Doetterl et al., 2012b). Recent studies give even lower estimates, about 0.5 Gt C y^{-1} (Quinton et al., 2010; Doetterl et al., 2012b), from arable land, which is the main source area for eroded sediment in a large part of the world. The dimensions of C loss through erosion can be dramatic: Doetterl et al. (2012b) estimate that for Europe, annual mobilization of SOC on agricultural fields is of the same magnitude as additional C sequestration induced through the use of fertilizers.

Soil erosion has, hence, traditionally been regarded as a process which leads to a loss of organic matter from soils through increased decomposition rates and export of organic matter from fields (Jacinthe et al., 2002; Lal, 2004, 2008; Ito, 2007). However, studies over the last two decades have sometimes shown otherwise. Traditional studies, which were mainly focused on pedon and plot scale inventories of C, generally concluded that erosion leads to net loss of SOC from the soil system. It was not until the final years of the last century that Stallard (1998) and Harden et al. (1999) showed that, when considered at a more appropriate watershed scale of analysis, soil erosion can induce a net terrestrial sink for atmospheric carbon dioxide (CO_2). Stallard (1998) and Harden et al. (1999) works illustrated the fundamental conditions under which erosion can contribute to terrestrial carbon sequestration: i.e. some of the eroded C is effectively buried at depositional sites and some is replaced at eroding landform positions. Until recently, this issue (as to whether the net impact of erosion on C cycling acts as a C source or sink) has been intensely debated in the literature, with divergent results on both sides on the order of 1 Pg C y^{-1} globally (Lal, 2004; Berhe et al., 2007; Harden et al., 2008; Smith et al., 2001; Van Oost et al., 2007) (Fig. 1). The discrepancies in quantifying the global sink/source term arise primarily from differences in the approaches used in each study (Table 1 and Fig. 1): The wide range of proposed values reflects on the one hand uncertainties associated with estimates of the magnitude of soil erosion (e.g. Boardman, 2006; Quinton et al., 2010), and on the other hand the fact that different elements of the soil landscape (soils vs. sediment deposits) (Kirkels et al., 2014) are analyzed separately or at different time scales (agricultural research: short-term, sedimentological research: long-term) (Hoffmann et al., 2013). Secondly, differences in sink/source terms also result from our incomplete understanding of the interactions between erosion and C cycling at the process level (e.g. Liu et al., 2003) and their variability in space and time. On this front, recent studies have provided new insights into how SOC stabilization changes throughout the different phases of soil erosion (detachment, transport, deposition) due to physical, chemical and biological factors as well as changing environmental conditions (e.g. increased moisture at depositional sites). Recently, Chappell et al. (2015) highlighted that, at least for Australia, omitting lateral fluxes can lead to an overestimation of C respiration from cropland of up to 40%. However, findings on different controls have often remained unconnected and a holistic ecosystem approach to address soil C cycling

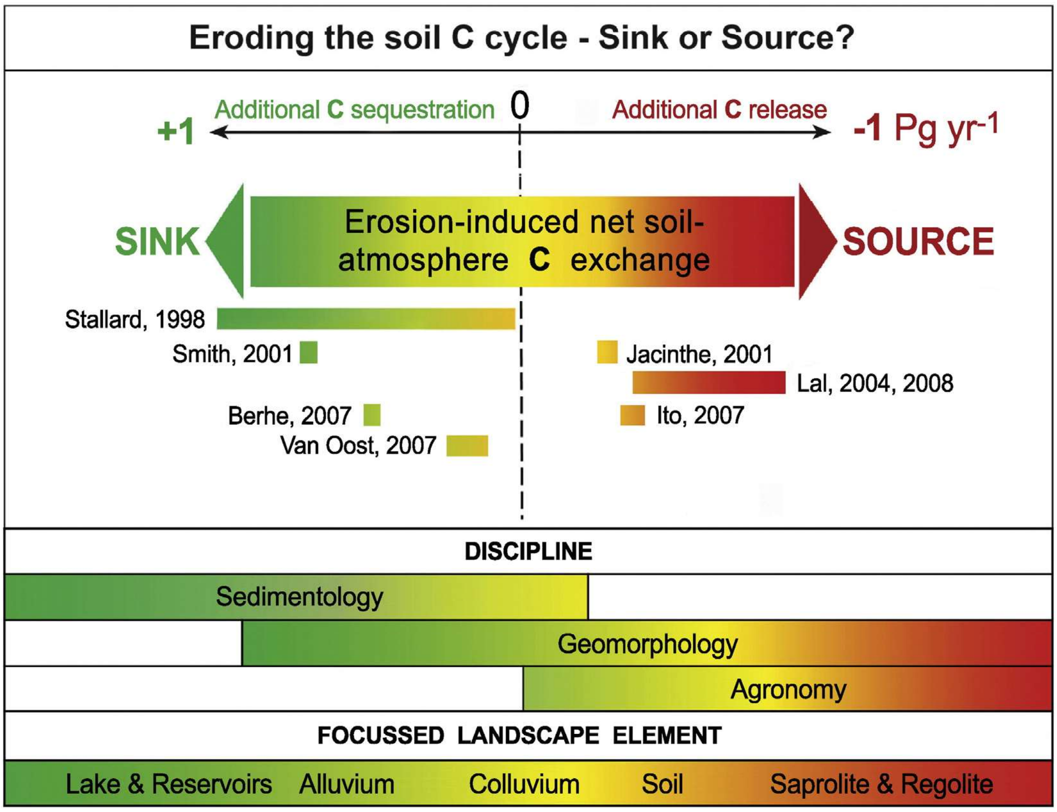

Fig. 1. Net effect of soil redistribution on C cycling as determined by current global scale studies. Colors and position along the x axis show the relationship between the disciplines and landscape elements focused on in these studies.

has yet to be developed. Due to an inadequate transfer of knowledge from smaller to larger scales, current ESMs do not consider lateral carbon fluxes and their effect on C cycling.

As a step in this direction, we present a synthesis of current knowledge on the processes regulating the interactions between soil erosion and the global C cycle and discuss the availability of suitable modeling and analytical tools for the study of C dynamics at the landscape scale. This review focuses on the terrestrial C cycle affected by water and tillage erosion processes, with an emphasis on arable land. While discussions on the effect of wind erosion or the coupling of terrestrial and aquatic cycles are beyond the focus of this review, we acknowledge that these related areas of inquiry have important implications for carbon cycling in the earth system. In subsequent sections, we summarize the effects of soil erosion on SOC dynamics from biogeochemical and physical, transport-related perspectives and illustrate opportunities to use both well-established and cutting edge experimental and modeling approaches in order to assess differences in C cycling between stable and dynamic landscapes. We place special emphasis on soil conditions at eroding and depositional sites and the conditions under which C is transported along the hillslope. We end the review with an outlook on the methodological and scientific challenges which need to be addressed in order to further our understanding of this important element of the terrestrial C cycle.

1.3. Defining key terms in SOC erosion research

There is still considerable confusion about the usage of key terms that describe critical process and variables in SOC erosion research (Berhe and Kleber, 2013; Kirkels et al., 2014). Therefore, before starting the review we provide the best available and latest definitions on the terms we consider to be most important when discussing SOC dynamics in eroding and depositional environments or settings. Detailed explanations of more complex concepts related to C cycling in eroding landscapes are given in Section 2.

Dynamic landscapes:

A landscape in which biogeochemical properties and fluxes are affected by soil redistribution, i.e. the erosion of material upslope and deposition downslope.

SOC input:

A qualitative and/or quantitative flux of C molecules into the soil matrix.

Soil/Terrestrial C sequestration:

The process, often expressed as a flux rate, of transferring CO_2 molecules from the atmosphere into the soil system. Soil C sequestration implies long-term storage of C in soil, and is typically facilitated by C input to soil, usually in the form of plant and/or animal remains, and environmental conditions that slow down the rate of C loss (USDA, 1996).

SOC sequestration potential:

The amount of C that a given soil can theoretically store before reaching its C saturation limit (maximum amount of C it can store), where C saturation of a given soil is determined by its inherent physical and chemical properties (Stewart et al., 2007).

SOC storage:

SOC storage is defined as the mass of organic C stored in a defined volume of soil, also described as SOC stock.

SOC stability:

Refers to the tendency of organic compounds in soil to resist change and/or loss. Stability can signify thermodynamic or kinetic stability, where kinetic stability refers to how rapidly compounds can be transformed or lost and thermodynamic stability is strictly defined as the difference in Gibb's free energy (ΔG) between reactants and products in a given chemical reaction. The term stable should not be used to imply that a given organic compound is unreactive or less reactive than other compounds, since stability of chemical species can be dependent on their reaction partners and surrounding environmental conditions (IUPAC, 1997). In soil biogeochemistry, SOC stability refers to a situation where either SOC molecular composition, SOC concentration or both remain constant for extended periods of time (Berhe and Kleber, 2013).

SOC stabilization mechanisms:

Physical, geo-chemical and bio-chemical mechanisms with the ability to increase the residence time of a given C atom within a reservoir compared with a reference situation (modified after Berhe and Kleber, 2013).

SOC persistence:

Persistence of organic compounds in soil refers to their longevity in the soil system. As noted in Berhe and Kleber (2013) in the context of SOC biogeochemistry, persistence of SOC can have different connotations: "temporal (measured in time units, e.g. turnover rate or mean residence time); kinetic (measured as activation energy of a given reaction); thermodynamic (measured as change in enthalpy, entropy or combination thereof); or even geomorphic (C that is preserved because of its location within a landscape/watershed)".

SOC turnover (τ):

The length of time it would take for a given stock of SOC to be exhausted if the rate of input = 0. For example, when SOC decomposition can be approximated as an exponential decay function, then $\tau = 1/k$, where k is the decay constant, or rate of decay (modified after Jenkinson and Rayer, 1977).

SOC residence time:

Residence time is defined as the average amount of time that an individual C atom spends in a given reservoir, e.g. the soil matrix (modified after Jenkinson and Rayer, 1977). Typically in soil biogeochemistry, the term Mean Residence Time (MRT) is used to highlight the fact that the C atoms that make up SOC are associated with pools that turnover at different rates.

SOC burial efficiency:

Burial efficiency describes the net difference between C concentration in topsoils and subsoils at depositional sites and is related to the burial of former topsoil C. Soil organic carbon concentration in subsoils must be compared to a reference profile, not affected by soil redistribution, in order to correct for in-situ sequestered C and to not overestimate the burial efficiency. Burial efficiency can therefore be described as a percentage of C survival after a known time.

Erosion-induced C sink:

A condition that results from a combination of erosional and depositional (terrestrial sedimentation) processes that lead to higher C storage in a given watershed after or due to erosion, compared to non-eroding conditions. Simply defined, the criterion for an erosion-induced C sink (Berhe et al., 2007) states that one can occur if dynamic replacement of eroded C and strongly reduced decomposition rates in depositional settings, together temporarily compensate for erosional losses of C from a watershed (or a given geographical area).

2. Mechanisms of stabilization vs. decomposition of C in eroding systems

2.1. Organo-mineral interactions: the role of SOC chemistry and environmental conditions

To date, most of the work related to C dynamics in eroding landscapes has focused on quantitative accounting of how much C is transported with erosion, where in the landscape it is deposited, and what proportion of the eroded C remains within the same eroding catchment. However, when considering the long-term effect of erosion on soil C dynamics, it becomes as important to consider the fate of the mobilized C. A relatively small but growing number of recent studies have looked at the stability and stabilization mechanisms of C in eroding vs. depositional landscape positions. Despite the identification of distinct differences in chemical and physical properties related to stabilization mechanisms, the medium to long term stability (i.e. decadal to centennial scales) of SOC fractions remains highly uncertain (Trumbore and Czimczik, 2008; Kleber et al., 2011). In this section, we review some of the major findings of the relevant mechanistic studies.

Process-based research has clarified the mechanisms and timescales involved in stabilizing SOC at the plot scale, and this understanding has been implemented in models of SOC dynamics over the past decades (Coleman and Jenkinson, 2014; Parton et al., 1988). In both eroding and depositional landscapes, C in soil can be stabilized against decomposition by two major mechanisms: (a) chemical association of SOC with mineral surfaces, and (b) physical protection of SOC, either by spatial separation from decomposers or inaccessibility due to the encapsulation of SOC within soil aggregates (Tisdall and Oades, 1982; Sollins et al., 1996; Six et al., 2002; Von Luetzow et al., 2006; Schmidt et al., 2011;Berhe et al., 2012a; Berhe and Kleber, 2013; Kleber et al., 2014). Previously, it was assumed that chemical composition of SOC was the dominant factor in determining its recalcitrance in soil. It was thought that certain C compounds, such as fire-altered C (also known as black C (BC) or pyrogenic C (PyC)), lignin and lipids, were 'inherently recalcitrant' in soil and remained stable over centuries to millennia (Kelleher and Simpson, 2006; Mikutta et al., 2006; Marschner et al., 2008; Adair et al., 2008; Kleber, 2010; Kleber et al., 2011; Singh et al., 2012) due to their large size and/or complex molecular structure, but this paradigm has been challenged and discussed in light of new findings on the drivers for persistence of carbon in soils (Lehmann and Kleber, 2015). Recent studies indicate that chemical composition of SOC may only be effective in regulating the rate of decomposition over short timescales (< 10 a), while environmental variables are more important controls on longer time scales (Schmidt et al., 2011; Lehmann and Kleber, 2015). Most important among these variables are the physical and chemical protection of SOC from microbial decomposition, and the lateral and vertical mobilization of the C through erosion and leaching (Kleber et al., 2011; Schmidt et al., 2011; Dungait et al., 2012; Berhe et al., 2012b).

Chemically, one of the most important factors for the stabilization of C in soils is the availability of mineral surfaces for sorptive protection (Kleber et al., 2014). Among the factors that determine the availability of mineral surfaces for sorptive stabilization are soil texture (i.e. Feng et al., 2013) and soil mineralogy, which guide the chemical interactions that SOC forms with the mineral phase (Torn et al., 1997). Clay-sized phyllosilicates and hydro(oxy)oxides tend to be the most important mineral constituents of organo-mineral associations, given their high specific surface area and hydroxylated reactive surfaces (Schnitzer and Kodama, 1992; Kaiser and Guggenberger, 2003; Totsche et al., 2010). The reactivity of silicate minerals with SOC depends on their type (expandable 2:1 versus non-expandable 2:1 or 1:1 phyllosilicate clay minerals) and size (i.e. specific surface area). In addition, association of SOC with mineral surfaces can be facilitated through the presence of ions on the edges of mineral structures, pH-dependent surface charge, and intercalation of C compounds in the interlayer spaces of phyllosilicate clay minerals (Kleber et al., 2005, 2007; Kaiser and Guggenberger, 2007). In moderate to highly weathered soils, accumulation of reactive Fe-, Mn-, and Al-oxy-hydroxides, as a result of biogeochemical weathering processes, also leads to stabilization of SOC through complexation/coprecipitation and adsorption processes (Tipping, 1981; Torn et al., 1997; Wagai and Mayer, 2007; Eusterhues et al., 2008; Berhe and Kleber, 2013; Kleber et al., 2014).

Reactive soil minerals and availability of sorptive surfaces further contribute towards physical stabilization of SOC, through their effects on formation and stabilization of soil aggregates. When SOC is encapsulated inside aggregates, diffusion of oxygen, water, and enzymes needed for breakdown of organic matter is limited, and therefore SOC is physically protected (Pinheiro-Dick and Schwertmann, 1996; Duiker et al., 2003; Berhe et al., 2012b; Wang et al., 2014b).However, recent studies show that mineralization rates of SOC in soil are not always related to soil oxygen levels and that responses can vary depending on the type of organic matter (Wang et al., 2013 and references therein).

Given these mechanisms for C stabilization in soil, it is clear that the geochemical composition of soil is of great importance. Often, the soil mineralogy along a hillslope shows considerable variability, and thereby dictates the effectiveness of prevalent C stabilization mechanisms. For example, Berhe et al. (2012a) showed that the stocks of reactive soil minerals, especially Iron (Fe) and Aluminium (Al) oxy(hydroxides), tend to be higher in depositional sites compared to eroding slope profiles in a relatively undisturbed zero-order watershed in northern California that experiences mainly natural erosion as a result of gopher bioturbation and rain splash. Further, Doetterl et al. (2015) studied eroding and depositional profiles along a cropped hillslope in the Belgium loambelt and showed that the distribution of pedogenic oxides and clay minerals partly determine effectiveness of SOC stabilization.

Finally, environmental conditions set the stage on which all other interactions between SOC, microorganisms and soil minerals take place. For example, studies have shown that soil humidity (Thomsen et al., 2003; McHale et al., 2005), aeration (Arya et al., 1999; Zhang et al., 2011) and temperature (Davidson and Janssens, 2006) are key controls on microbial activity and hence SOC stability. Physical inaccessibility, as dictated by aggregation, burial in deep soil layers and spatial separation of decomposers from SOC in soils (Six et al., 1998, 2002; Von Luetzow et al., 2006; Holden and Fierer, 2005; Rumpel and Kögel-Knabner, 2011), are also important environmental controls on SOC persistence. In addition, it is likely that these variables interact in a complex manner affecting SOC stability (i.e. Salomé et al., 2010; Lehmann et al., 2007; Schmidt et al., 2011; Berhe and Kleber, 2013). For example, organo-mineral associations, which lead to higher energy requirements for decomposers to access and decompose C-based compounds, could

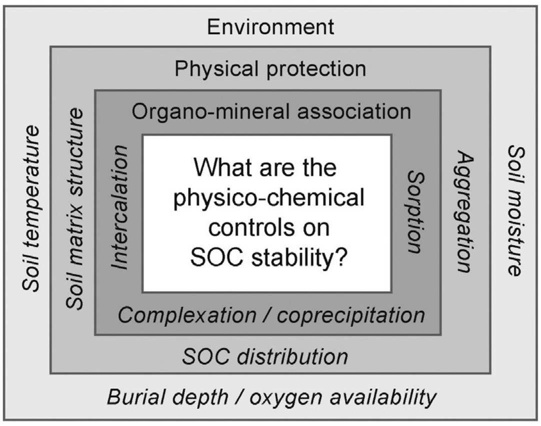

Fig. 2. Conceptual view on the most important geochemical, physical and abiotic environmental controls on SOC stability. See text for explanation.

also be physically separated from decomposers by being locked up in stable aggregates, or deposited in low oxygen environments (Fig. 2).

2.2. SOC in eroded soils – accelerated decomposition of old C

Two competing processes, the decomposition of old soil C and the sequestration and stabilization of fresh C inputs (Berhe et al., 2007; Berhe et al., 2014), take place simultaneously at eroding sites.

Generally, the turnover rate of deep SOC is slower than that of comparable topsoil SOC sources (Trumbore, 2009; Torn et al., 2009; Salomé et al., 2010; Rumpel and Kögel-Knabner, 2011; Schmidt et al., 2011). However, exposure of deep SOC by erosion of surface soil and associated changes in micro-climatic conditions are likely to increase the rate of deep SOC decomposition. For example, when erosion exposes subsoil layers they come into contact with fresh 'labile' C sources from plant growth on the eroding hillslope. Addition of 'labile' C to a soil C pool can stimulate the decomposition of soil C that was previously decomposing more slowly (Fontaine et al., 2007). Theoretically, this occurs when fresh SOC from topsoil horizons is mixed with formerly deep SOC, providing readily available energy sources for decomposers, which speeds up the decomposition rate of older, previously stable carbon. It is still not sufficiently clear how significant this 'priming' effect is on eroding hillslope soils, but it is likely to strongly affect the turnover rate of subsoil SOC previously protected by burial in oxygen limited subsoils, aggregation, or chemical association with soil minerals.

2.3. SOC in eroded soils – dynamic C replacement

Typically, unless erosion rates are large enough to significantly retard plant productivity in eroding hillslopes, some of the eroded C is replaced by new C inputs (Harden et al., 1999; Berhe et al., 2007). Harden et al. (1999) first coined the term 'dynamic replacement' to describe the continuous loss (by erosion) and replacement (mainly by production of new photosynthate) of C in eroding watersheds. Since then, the term dynamic replacement has been used to infer replacement of C lost by erosion through different processes.

First, the erosion of surface soil and exposure of subsoil layers can increase rates of chemical weathering of formerly deep soils or saprolite (with fluxes of CO_2 from carbonates, if they are present). Secondly, because there is typically less C in subsoils compared to near-surface layers, minerals in formerly deep soil layers are likely to have reactive sites on their surfaces that are not saturated with sorbed C compounds. Forming organo-mineral complexes of new C with these mineral surfaces can render the new C in the eroded soil profiles more stable (Harden et al., 1999). Finally, the erosional removal of weathered topsoil material also brings the zone of C accumulation closer to the mineral weathering front of a soil. In that sense, erosion leads to a "rejuvenation" of the soil by removing the most weathered soil material and leaving plants to grow in C depleted and less weathered substrate. This can favor SOC sequestration and supports the concept of dynamic replacement of eroded SOC, leading to a local net sink for atmospheric C: When the erosional soil loss is small to moderate, soil weathering/production can supply the soil with reactive soil minerals to stabilize C. It is likely, that the same processes can take place at depositional sites if C depleted material is mobilized and deposited. This is particularly relevant for systems with fast erosion rates, where C depleted former subsoil from hillslopes is mobilized before mineral surfaces can be saturated with new C.

In summary, soil erosion can have important implications for the rate of C input to the soil C pool (i.e. plant productivity), as well as for the accumulation and stability of organic matter in soil (Berhe and Kleber, 2013). Erosion removes topsoil material from sloping landscapes by a combination of natural and anthropogenically enhanced water, wind, and tillage erosion processes. At eroding sites, topsoil removal can, over the long-term and under high erosion rates, reduce plant productivity by depleting soils of their capacity to hold nutrients and water due to the removal of reactive minerals and the destruction of the physical soil matrix. For example, in many loess regions of the northern hemisphere, the periglacial aeolian loess deposits are a limited source for soil development of only centimeters-to-meters thickness, in which Luvisols have developed (e.g. Rommens et al., 2005). These regions are often intensely cultivated and subject to high rates of erosion. Underlying the loess are often less fertile substrates such as tertiary sands. Eroding the loess soils and substrates will then ultimately lead to dropping yields and hence reduced input of C to the soil. However, several studies (i.e. Rosenbloom et al., 2001; Doetterl et al., 2012a) found no significant differences in the SOC depth distribution and SOC stocks between stable and eroding soil profiles. This indicates that in some cases, where erosion rates are moderate, C sequestration takes place at the same rate as the removal of SOC enriched topsoil. It is debatable as to whether erosion could, in some soil systems, also lead to (temporary) yield increases. In tropical soil systems, for example, long lasting chemical weathering has led to the depletion of mineral nutrients from soils. The erosional removal of nutrient depleted layers, for example through landslides, now leads to the surfacing of less weathered, more fertile mineral substrate (i.e. Vitousek et al., 2003; Porder et al., 2005). These substrates might then stimulate higher plant productivity and yields. However, as these soils are very deeply weathered (several to 10s of meters), with an enrichment of Fe and Al oxides and kaolin types of clay minerals (IUSS Working Group WRB, 2014), it is unclear whether erosion can act at a pace where it is fast enough to reach the less weathered regolith, which contains the minerals with more reactive surfaces, while still allowing biomass production and C sequestration to remain at a high level.

2.4. SOC in transport – enhanced mineralization following erosion events

A large part of the controversy on the role of soil erosion in the C cycle relates to what happens during transport of eroded SOC. Studies to date have highlighted two apparently contradictory processes taking place during transport: (i) increased mineralization of C due to the breakdown of aggregates, and (ii) C enrichment in sediments relative to source soils due to selective transport and deposition of C.

The breakdown of macroaggregates is a key concern for understanding the magnitude of C mineralization during soil transport. Easily mineralizable C encapsulated within aggregates represents a large proportion of C in this fraction and can easily be released to the atmosphere upon breakdown of the aggregates (Six et al., 2000; Six et al., 2001; Six et al., 2004). The breakdown of aggregates, and hence increased CO_2 fluxes from mobilized C, depends predominately on the kinetic energy of raindrop impact. Several studies have quantified the increased SOC mineralization in the laboratory (Jacinthe et al., 2004; Polyakov and Lal, 2008) or directly in the field (Van Hemelryck et al., 2011). However, results on increased mineralization differ greatly, with between 0% and 100% increases relative to the fluxes of the reference source soil. In addition, the increase in CO_2 flux appears to be short-lived and limited to the days following the rainfall events (Jacinthe et al., 2002; Van Hemelryck et al., 2011). When observed over longer periods, CO_2 release from eroded soil eventually reaches a point where there is no observed difference compared to non-eroded controls (Van Hemelryck et al., 2010; Bremenfeld et al., 2013).

In addition, the gaseous fluxes of CO_2 from the soil to the atmosphere are strongly dependent on the physico-chemical characteristics (ex. soil texture and water content) of the soil and the nature of the C (amount and stability) in the mobilized material (Hu and Kuhn, 2015). A few studies have suggested that the respiration rate of small sediment particles could be lower than that of larger particle size classes and bulk soil (Polyakov and Lal, 2008; Hu and Kuhn, 2014), indicating that the size-distribution of sediment particles could be important when quantifying changes in the mineralization rate of SOC due to erosion. However, so far, it remains challenging to quantify additional CO_2 fluxes following the transport and breakdown of aggregates in the field and/or once eroded SOC leaves source catchments. The incorporation of this process into numerical models, which currently consider a constant percentage of transported SOC to be mineralized (Van Oost et al., 2005b), is an area that demands further research.

2.5. SOC in transport – particle size selectivity

Higher SOC concentrations in depositional profiles are usually related to the preferential detachment, downslope transport and hence deposition of soil constituents that are typically enriched in SOC relative to the bulk soil (e.g. Gregorich et al., 1998; Kuhn, 2007; Kuhn et al., 2009; McCorkle et al., 2016) and the source soils of the sediments (Sharpley, 1985; Palis et al., 1997). Studies conducted in areas with steep topography and forest cover show particularly significant C enrichment in eroded sediments (McCorkle et al., 2016; Stacy et al., 2015). This is related to the fact that the free light fraction, the POM fraction and clay and silt sized particles, which are commonly C enriched compared to coarser soil particles, are preferentially transported during erosion events. However, deposition can also be a selective process. Where transport capacity is high enough, the mobilized lighter fractions (POM or clay sized particles) might not be deposited but instead exported out of the catchment. In this case, no enrichment of C in locally-deposited sediment relative to the source material will take place.

The selective transport of certain particle-sizes in surface runoff is, hence, an important variable to determine C transport, and ultimately, sediment enrichment, with important consequences for C burial and the potential sink. Coarse particles are the first to be deposited under transport limited conditions. Therefore, at low surface runoff velocities, deposition starts immediately after detachment, and runoff becomes enriched in suspended smaller sediments and associated C (Jacinthe et al., 2004; Palis et al., 1997; Schiettecatte et al., 2008a, 2008b).

However, if C is stored in water-stable (micro-)aggregates, the particle size related fractionation of C enriched and C depleted material might be small. Stable aggregates consist predominantly of C rich fine silt, clay and POM (Six et al., 2000, 2001). As these aggregates have the size and the weight of fine sand or coarser silty particles, they will settle much earlier after detachment than smaller particles. Furthermore, sediments can also present lower C concentrations than their source soils due to the transport of mineral particles with very low C concentration (Nadeu et al., 2012).

In conclusion, C enrichment is not homogeneous along different particle-size groups (Hu and Kuhn, 2014) and can also be variable through time (Hu et al., 2013). It is possible that most of the easily decomposable 'labile' C could be mineralized before it reaches depositional profiles. Under other conditions, sediment and C erosion may not lead to any selective mobilization if aggregates, instead of single particles, are eroded and remain intact during the entire transport phase of soil erosion. This is supported by studies that observe only small differences in soil texture along an eroding hillslope toposequence (e.g. Steegen et al., 2001; Wang et al., 2010). Overall, the net effect of transport and deposition on eroded SOC depends on the amount and nature of the eroded C, soil texture, soil aggregation, the transport distance (Starr et al., 2000), the rate and nature of soil erosion, and terrain attributes such as slope gradient and surface roughness (Berhe and Kleber, 2013). Slow but long-range transport may lead to a higher degree of decomposition of mobilized C, while fast but short-range transport might lead to the burial of mobilized, potentially easily mineralizable C, with a lower degree of decomposition, at the depositional setting (Berhe and Kleber, 2013).

2.6. SOC burial at depositional landform positions

Relatively little is known about the effect of SOC burial in subsoil layers (below 30 cm depth) (Rumpel and Kögel-Knabner, 2011). This is a consequence of the fact that traditionally most SOC studies focused on mechanisms and controls of C storage and stabilization in topsoils, even though half of the organic C in the top one meter of soils is stored below 30 cm depth (Hiederer and Köchyl, 2012). This is undeniably a source of great uncertainty for estimating global SOC dynamics. On average, only 10–30% of the eroded topsoil material is subsequently transported into lakes and oceans via major river systems (Walling and Webb, 1996; De Vente et al., 2007). The rest is typically deposited downslope at footslopes, as well as colluvial valley bottoms and alluvial flood plains (Stallard, 1998). Moreover, tillage erosion only translocates material within single fields, creating its own patterns of buried soils. In depositional areas, SOC undergoes a fundamentally different dynamic than it was subjected to at its source. When assessing C stocks at larger scales, our understanding of physical and chemical mechanisms of C stabilization in subsoils is currently limited. In eroding/depositional landscapes, several studies have shown that, when C burial takes place, SOC stocks below 30 cm (i.e. below plow layers and the zone of intense mixing by mammals) can account for more than 80% of a soil profile's total SOC stock in the upper 2 m of soil (Berhe et al., 2008; Doetterl et al., 2012b; Wiaux et al., 2014b; Wang et al., 2015b). Thus, focusing on topsoil when investigating the erosion-induced C sink can lead to strongly skewed SOC stock assessments, as an overestimation of the amount of SOC lost by erosion disregards the large proportion of SOC buried at depositional sites.

Transported and deposited SOC can be protected from decomposition if efficiently buried in slow turn-over environments, leading to large C sinks in colluvial and alluvial sediments (Polyakov and Lal, 2004a, 2004b; Van Hemelryck et al., 2011; Wang et al., 2015a, 2015b; Vandenbygaart et al., 2015). Over timescales of decades to centuries, erosion and subsequent deposition can then result in large deposits of C rich soil material of varying quality. However, the amount of C buried over time will depend largely on the rate of burial, the time since burial, the nature and amount of mobilized C, and the environmental conditions that the buried C is exposed to after deposition. The capacity of this process to constrain SOC mineralization may be reduced over time and as a function of environmental factors and sedimentation rates (Van Oost et al., 2012; Wang et al., 2015b). For example, the protection of C within aggregates or by organo-mineral association might be weakened if aggregates lose stability and mineral weathering alters the chemical reactivity of minerals to C. Hence, changes in C stocks with depth at depositional sites can shed light on the underlying mechanisms for the persistence and turnover rates of buried C (i.e. Chaopricha and Marín-Spiotta, 2014; Doetterl et al., 2015). Deeper soil layers have generally lower C concentrations and more stable SOC pools than comparable topsoils (Rumpel and Kögel-Knabner, 2011; Berhe et al., 2012a). However, if buried topsoil experiences minimal change before burial (as would be expected when sediments are mobilized and buried in large erosive events), it can have a similar C concentration and composition as the current topsoil. In reality, the quality and quantity of C in buried topsoils has been observed to vary greatly. Buried C can be similar to C rich topsoils or can be more similar to C depleted subsoil horizons of profiles not affected by erosion and deposition (Doetterl et al., 2012a; Berhe et al., 2012a, Berhe, 2012b; Van Oost et al., 2012; Hoffmann et al., 2013; Helgason et al., 2014; Wiaux et al., 2014b; Vandenbygaart et al., 2015). While well aerated colluvial soils often show strong depth gradients of C concentrations, with lower C concentrations in the deepest, hence oldest, layers, water-saturated, and hence oxygen limited, alluvial soils can store C for centuries with no measurable changes (Van Oost et al., 2012; Hoffmann et al., 2013). Therefore, if burial is fast and conditions at depositional sites are favorable for C conservation, sediment deposition at foothills, valleys and floodplains can result in the burial of C enriched in labile compounds (Gregorich et al., 1998). If decomposition during sediment transport, deposition and burial has led to significant degradation of more easily decomposable C fractions (Wang et al., 2014a; Wang et al., 2015b), depositional sites can accumulate smaller but highly stable C stocks with long residence times. However, it is important to note that there currently exists no reliable tracer to distinguish between subsoil C that has been transported, deposited and buried at depositional sites and C that has been sequestered *in situ* and buried with deposited sediment. This ultimately leads to great uncertainties when assessing the long term effect of soil redistribution on depositional sites and their function as sinks for atmospheric C.

3. Experimental approaches to capture soil redistribution effects on C dynamics

The challenge in experimentally evaluating the interactions between soil erosion and C dynamics results from (i) the short-term, event-driven nature of (water) erosion (Nearing et al., 1999) and (ii) the associated long-term spatio-temporal patterns arising from these processes. To tackle this challenge, two experimental pathways are generally followed: On the one hand, fluxes of sediment and SOC, as well as short-term vertical C fluxes, are measured or monitored on timescales of minutes to years. On the other hand soil, sediment and SOC stocks, representing the long-term cumulative effects of different fluxes over timescales of decades to millennia, are analyzed. In this section we review the experimental approaches available to analyze the effects of soil erosion on C dynamics and discuss the problems in linking the results from short-term flux and long-term stock measurements.

3.1. Event based measuring and monitoring – of sediment and associated lateral C fluxes

There is a long tradition in soil erosion research to use plots of differing size to determine water erosion under natural (Nearing et al., 1999; Wischmeier et al., 1958) or simulated (Kinnell, 2009; Sharpley and Kleinman, 2003;Fiener et al., 2011) rainfall. These types of plot studies are also widely used to determine erosion induced lateral C fluxes (Martinez-Mena et al., 2008; Schiettecatte et al., 2008a, 2008b; Hu and Kuhn, 2014) and associated short-term vertical C fluxes to the atmosphere (Bremenfeld et al., 2013; Van Hemelryck et al., 2010; Wang et al., 2014a). However, while plot studies are important for our understanding of single erosion events, they do not allow for an accurate evaluation of erosion and associated C fluxes on a landscape scale or over longer timescales. Due to the episodic nature of water erosion processes, a continuous and long-term catchment monitoring of sediment fluxes and dynamic catchment characteristics (e.g. land management) is required. Ideally, this can provide a more integrated view of the catchment response following erosion events. This is especially important when determining the effects (see Section 2) of erosion, transport and deposition on bulk SOC (Owens et al., 2002) or SOC fractions (Jacinthe et al., 2004) given different catchment characteristics.

However, such integrative measurements are costly and time consuming and, hence, catchment scale studies combining sediment and SOC delivery are rare (e.g. Wang et al., 2010; Boix-Fayos et al., 2015) and mostly focus on output measurements. The continuous monitoring of lateral C fluxes requires a large quantity of measurements and ideally distinguishes between mineral associated (MOC), particulate (POC) and dissolved organic carbon (DOC). Moreover, it requires sampling strategies which prevent any microbial activity and sample degradation in the sampled runoff. Techniques used range from time consuming event-based grab sampling to more sophisticated automated samplers (e.g. Fiener and Auerswald, 2003), which would ideally store samples in refrigerated containers. This allows for samples to be later analyzed for DOC, solid C (MOC + POC), C/N ratios, isotopic composition, etc. in order to inform the researcher on C quality, stabilization mechanism or origin of the transported C.

3.2. Event based measuring and monitoring – the use of tracers

The drawback of measurements at the catchment outlet is that catchment internal processes of soil and SOC redistribution and short-term vertical C fluxes cannot be analyzed. Therefore, some studies determine the event-based internal erosion processes based on after-event field campaigns, measuring the redistributed soil volume on linear erosion features (rills, ephemeral gullies) and deposits (Van Oost et al., 2005b). However, interrill erosion rates, especially important for the enrichment of C in transported soil (Kuhn, 2007), cannot accurately be quantified through visual inspections or measurements during field surveys (Auerswald and Weigand, 2000).

To overcome the difficulties in experimentally determining the spatial patterns of erosion rates due to single or episodic events, the use of a number of different tracers as proxies for soil redistribution has been established to track the transport patterns of mobilized soil. Rare earth elements are the most commonly used for tracking mass transport. These tracers can be applied at different landscape positions and their distribution pattern after erosion events can yield information about the strength and nature of these events (Deasy and Quinton, 2010; Polyakov et al., 2009; Polyakov and Nearing, 2004). Radionuclides are more common to analyze long-term erosion patterns (see below), but certain tracers, especially ^{7}Be, with a short half-life time of only 53.12 days, have the potential to provide information about single erosion events (Blake et al., 1999; Fitzgerald et al., 2001). Other studies use magnetic compounds (Parsons et al., 1993) or traceable particles like radio frequency identification (RFID) tags incorporated into topsoil (Parsons et al., 2014). Moreover, a wide range of biochemical fingerprinting techniques are used which, in principal, allow the origin of different materials in sampled runoff to be traced (Walling, 2013). Finally, isotopic labelling, for example of cutin and suberin biomarkers, has been successfully applied to disentangle and measure the respiration of shoot and root C respectively (Mendez-Millan et al., 2010, 2011). Assuming that soil erosion is selective towards transporting and depositing shoots over roots, or vice versa, this technique could also be used as a means of tracing the source of respired C, by marking C prior to mobilization and deposition.

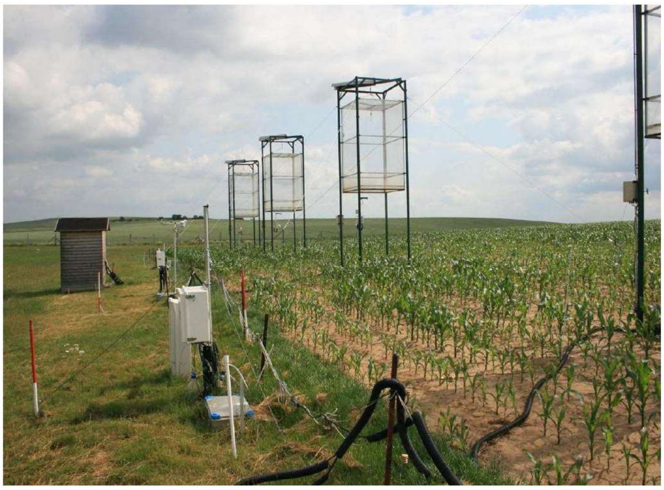

Fig. 3. Automatic chamber system at CarboZALF-D experimental site, situated on hummocky carbonate rich ground moraines near Dedelow (NE Germany). A long-term (>10 years) monitoring system for climate, soil water, temperature and redox dynamics, C and nutrient fluxes and plant biomass dynamics was established at a total of 9 experimental plots (Sommer et al., 2016). Plots were arrange in such a way as to cover representative soil and process domains and to minimize plot scale heterogeneity along a full gradient of soil erosion and deposition. Transparent chambers (height of 2.5 m, 2.5 m^2 base area) move up and down in 60 min intervals in order to periodically create a temporarily closed system. Dynamics of CO_2 concentrations (increase/decrease) are recorded by open path infrared gas analyzers upon closure of the chambers to calculate CO_2 net ecosystem exchange (NEE).

The attractiveness of using tracers comes from the fact that they can be used to determine the influence of both water and tillage erosion processes within fields or small catchments (Lobb et al., 1999; Van Muysen and Govers, 2002). However, while tillage translocation is not a selective process, potential bias when analyzing tracer patterns is introduced by the selective transport which occurs through water erosion. Uncertainties arising from this are discussed in a recent review of Guzmán et al. (2013).

3.3. Event based measuring and monitoring – erosion effects on vertical C fluxes

Measuring short-term, event based effects of erosion on vertical C fluxes at the landscape scale is challenging and subject to uncertainties based on the methodological approaches. First, C fluxes induced by aggregate breakdown during rainfall and the subsequent transport with water can only be determined indirectly, either by comparing respiration measurements on plots subjected to heavy rainfall with controls protected from raindrop impact (Van Hemelryck et al., 2010; Bremenfeld et al., 2013), or by performing incubation measurements using sediments sampled at watershed outlets (Jacinthe et al., 2004). Both approaches are somewhat problematic, as vertical C fluxes are not measured during the actual process (i.e. rainfall event).

Automatic chambers (Fig. 3) at erosional and depositional sites can be used to provide in-situ measurements, from which the relative effect of erosion and deposition on vertical C fluxes following an erosion event can be determined (Sommer et al., 2016). However, as such in-situ measurements at different landscape positions are affected by local environmental conditions, e.g. soil moisture and temperature, any up-scaling to the landscape scale is difficult and subject to large uncertainties. For example, the vertical C flux measured after a single deposition event might be biased by the variable underlying SOC stocks at different landscape positions.

One way to overcome these methodological constraints is to use manipulation experiments, creating conditions of erosion or deposition artificially and suppressing other processes, hence excluding further disturbance (like vegetation or temperature changes). At plot scale, such experiments were introduced starting some decades ago to quantify the impact of erosion on crop yields (Herzog and Kunze, 1976; Herzog, 1990; den Biggelaar et al., 2001; Bakker et al., 2004; Jagadomma et al., 2009; Larney et al., 2009).

For studies on C dynamics, such elaborate experimental setups and complex manipulations are still rare and prone to high cost, time and labor investment. To our knowledge, no such experiments have yet been conducted at the catchment scale. The CarboZALF-D experimental site (Fig. 3) is one of the rare examples of a successful application of a manipulation experiment to measure vertical C fluxes. At this site, automatic chamber systems were established to quantify net ecosystem C exchange (NEE) at different geomorphic settings along a full gradient of eroded and depositional soils, excluding subsequent soil translocations. CO_2 fluxes calculated from automatic chamber data over a 4-year period showed a significant net C sink of -75 g m^{-2} y^{-1} at eroding sites due to dynamic replacement of the eroded C, though there was a large interannual variation (Hoffmann et al., 2015b).

3.4. Estimating long-term soil erosion

The importance of soil erosion processes for long-term C cycling is generally determined by comparing the soil loss or gain by erosion or deposition with the associated SOC stocks at various landscape positions. At the field scale, monitoring techniques that measure CO_2 fluxes at different depths have been used in eroding landscapes. For example, Wiaux et al. (2015) studied an eroding hillslope to identify the origin of respired CO_2 and to determine the role of physical barriers in soils that may control the vertical diffusion of respired CO_2.

Many different approaches have been used to estimate the amount of eroded and deposited soil at different landscape positions or to calculate integrated erosion/deposition rates. Both erosional and SOC patterns arise from long-term processes, which maybe only partly linked to each other. Perhaps the most established approach to determine soil truncation at erosional sites is to compare the depth distribution of indicative soil horizons or properties (e.g. thickness of Bt horizons, depth of the weathering front, etc.) at reference sites unaffected by soil redistribution with other landscape positions (Rommens et al., 2005). The intrinsic weakness of this approach is that it assumes that soil development along eroding hillslopes is similar. This is questionable as, for example, the availability of water to drive chemical weathering differs strongly along the hillslope (Norton et al., 2014; Opolot et al., 2014; Hunt, 2015).

A more sophisticated approach uses naturally occurring or deliberately applied tracers. The most widely used erosion tracer, which allows for the determination of erosion and deposition rates over approximately the last 50 years, is the radionuclide ^{137}Cs (Kachanoski and De Jong, 1984; Parsons and Foster, 2011; Walling and Quine, 1990). 137-Cesium, which has a relatively short half-life of 30.17 years, was introduced into the stratosphere as a result of nuclear weapons testing in the 1950s, 60s and 70s. Its deposition pattern in soils reflects its atmospheric circulation (UNSCEAR, 1982, 1989). After deposition, ^{137}Cs is strongly absorbed by mineral particles in topsoils and hence any spatial patterns within agricultural landscapes can be attributed to lateral redistribution processes. Conversion models of varying complexity have been used to derive erosion rates from ^{137}Cs (for an overview see Walling et al., 2002). Another radionuclide associated with nuclear weapons testing is $^{239+240}$Pu (Litaor et al., 1995; Schimmack et al., 2001). The major advantages of these radionuclides is the much longer half-life time ($t_{½} = 24\,110$ and 6 564 years, respectively), which make $^{239+240}$Pu valuable tracers for future studies. Moreover, there is less recontamination of soils with $^{239+240}$Pu following nuclear power plant accidents compared to ^{137}Cs(e.g. Schimmack et al., 2001) and the ^{240}Pu/^{239}Pu ratio can give evidence of such contamination (Ketterer and Szechenyi, 2008). Apart from radionuclides resulting from human activities, ^{210}Pb ($t_{½} = 22.26$ years), which occurs as a natural radionuclide, is widely used for studies evaluating the erosion rates over approximately the last 100 years (Li et al., 2006; Walling et al., 1995).

3.5. The use of tracers in long-term assessments

Most of the tracers introduced above were used to determine event based erosion and deposition rates along hillslopes or small catchments on shorter time scales. For investigating long term soil and C erosion and deposition dynamics, tracer techniques are applied to determine deposition rates in colluvial, alluvial or lake deposits. The most widespread technique to determine the age of deposits is ^{14}C dating of buried materials (e.g. charcoal etc.) or of SOC directly (an overview is given in Trumbore, 2009). The major advantage of working with ^{14}C dating is that, at least in the case of analyzing SOC, the object of interest is analyzed rather than sediments (Dlugoß et al., 2012). However, this advantage can be also considered a drawback, as C might be introduced via other processes, not just through soil deposition into sediments. Another well-established technique is optically stimulated luminescence (OSL) dating. OSL utilizes dosimetric properties of quartz and feldspar in sediments to determine the time of burial (Duller and Wintle, 2012; Lang et al., 2003; Notebaert et al., 2011). Apart from these widespread dating techniques, a tremendous number of other, often site-specific, tracers have been used to determine sediment ages. For example, Horák and Hejcman (2013) showed the potential of using heavy metals found in alluvial deposits which result from known historical mining upstream.

While the use of tracers for tracking sediment flows is well established, no such technique is yet available for qualitatively and quantitatively tracking SOC fluxes. The use of a series of biomarkers, such as lignin or plant waxes, is currently discussed as a means to potentially reconstruct land use change (Fisher et al., 2003; Sikes et al., 2009; Mendez-Millan et al., 2011; Bush and Mclnerney, 2013; Zech et al., 2013), climatic changes and to determine the origin of deposited C. However, a critical assessment is necessary when using these biomarkers for the study of dynamic landscapes affected by soil redistribution. For example, measuring the depletion and enrichment of various lignin phenols is a well-established approach for estimating C degradation in stable landscapes. However, Hernes et al. (2007) showed that these lignin phenols are strongly affected by selective sorption to minerals, which makes this biomarker very difficult to interpret in a landscape affected by soil redistribution.

3.6. Measurement and importance of DOC fluxes

While a fair number of studies on soil redistribution address MOC and POC fluxes, there is a substantial lack of studies and observations to elaborate the link between erosion and its effect on DOC fluxes. Although a relatively small fraction of total SOC (less than 10%), DOC export from soils can link terrestrial with aquatic C cycling from local (Jansen et al., 2014) to global scales (Regnier et al., 2013). In a comparative study on vertical DOC fluxes from Chernozems, Brye et al. (2001) found increased fluxes for arable land (3–4 g C m^{-2} y^{-1}) compared to a restored prairie (2 g C m^{-2} y^{-1}). Kindler et al. (2011) reported similar DOC fluxes of 3–5 g C m^{-2} y^{-1} from three arable sites with Cambi-, Luvi-, and Stagnosols. However, most data on DOC export have been collected in forested ecosystems of the northern hemisphere, mainly unaffected by erosion. The number of studies on vertical or lateral DOC fluxes in agricultural systems, where soil redistribution rates are much higher, is rather limited, and to this point, a direct linkage between the erosional status of catchments and DOC exports has not been established. Currently, plot-scale studies on DOC losses from arable land report these losses as vertical fluxes. Distinguishing between lateral (DOC in surface, interflow or groundwater runoff) and vertical DOC loss is unresolved but crucial when trying to understand the effect of erosion on DOC. Gerke et al. (2015) showed that water flow paths and fluxes exert the strongest control on DOC fluxes in a catchment with a strong impact of erosion. In the same catchment, Rieckh et al. (2014) modeled the vertical DOC export from eroding sites as 1 g C m^{-2} y^{-1}. Colluvial soils influenced by groundwater showed a vertical DOC import of 0.4 g C m^{-2} y^{-1}.

It may be possible to deduce the magnitude of lateral DOC fluxes from headwater catchment studies. DOC exports from agricultural catchments range from 0.2–2.0 g Cm^{-2} y^{-1} (Royer and David, 2005; Dalzell et al., 2007, 2011; Graeber et al., 2012). However, catchment scale studies face the methodological problem of undefined, variable source areas, like riparian zones, as hot spots for DOC export. In addition to the above mentioned spatial aspects, further methodological constraints are related to DOC flux quantification: (i) Soil water can be extracted by different methods, such as by using equilibrium-tension lysimeters, zero-tension ceramic plates or suction cups with variable vacuum settings. For example, when suction cups are used, depending on the suction strength, soil pores of different diameter will be drained, i.e. smaller soil pores are drained at higher suction. In consequence, the extracted water can differ in both local residence time and DOC concentration. (ii) Furthermore, in regions with a strong winter season, most sampling campaigns exclude the winter period due to freezing problems in the sampling devices. This can result in systematic bias and in an underestimation of DOC fluxes if non-monitored water fluxes occur during this period. (iii) Finally, as DOC fluxes are strongly dependent on the magnitude of water flow, differences in measurements could result from the use of different approaches to calculate water budgets or flows in soils, e.g., a simple bucket approach vs. dynamic process models, like HYDRUS (Rieckh et al., 2014).

3.7. Laboratory experiments

Although very much needed, it is not always feasible to undertake extensive field experiments on erosion and C dynamics. For example, it can be hard to monitor or maintain equipment installed on-site in remote areas. Some areas might be affected by disturbance or may be only temporarily accessible. In other regions, the variability in soil and topographic conditions will be too high to justify a proper installment of expensive monitoring equipment. In all these examples, laboratory experiments under controlled conditions might be a valuable alternative to replace or compliment time and cost intensive field scale experiments. They could also be useful in identifying criteria to help justify the installation of field equipment in a later phase of a study.

Laboratory based experiments can be ideal tools to study isolated processes from a mechanistic point of view and can often help to improve our basic understanding of real world processes. Some of the most prominent examples are rainfall simulations (for example Hignett et al., 1995; Lascelles et al., 2000; Schindewolf and Schmidt, 2012) and flume experiments (for example Giménez and Govers, 2002; Giménez et al., 2004; Govers et al., 2007), which analyze the effects of water erosion on soil structure, aggregate stability or the movement of soil material. One of the major advantages of these types of laboratory experiments is the level of control and comparability that is possible (i.e. replication of the exact same event) and the variety of manipulations (i.e. simulating an identical erosion event on different soils, with variable status of pre-wetting, aggregate structure, plant coverage etc.) that can be performed. As aggregate stability is one of the key parameters defining soil transportability via water erosion, many studies performed manipulations on different types of aggregates under various conditions (Wilson et al., 1947; Mbagwu and Bazzoffi, 1989; Gollany et al., 1991; Rawlins et al., 2015). During the last decades a rising number of studies are also explicitly addressing aggregate stability in relation to lateral C fluxes and SOC dynamics (i.e. Denef et al., 2001; Van Hemelryck et al., 2010; Saha et al., 2011; Bremenfeld et al., 2013). These are just a few examples of the hundreds of laboratory studies that have been performed, addressing erosion from either a geomorphological, soil transport or soil disruption point of view. A large number have also analyzed the enrichment of carbon in transported soil (Sharpley, 1985; Schiettecatte et al., 2008b; Jin et al., 2009). Incubation studies, either on field fresh, legacy or artificial samples (i.e. taken after a rainfall simulation), provide information on the respiration characteristics of sediments(Van Hemelryck et al., 2010; Bremenfeld et al., 2013) and can be complimentary to fractionation and C turnover studies. Incubation can be especially helpful to distinguish between potential C fluxes along a soil column, through the incubation of isolated samples, or to study the stability of C under various conditions. For example, experiments on the respiration from buried soils have yielded important information on the controlling factors for the stability of C (Wang et al., 2014b; Wang et al., 2014c; Wiaux et al., 2014a; Doetterl et al., 2012a). In combination with soil fractionation or isotopic analysis of C and N molecules, incubations can yield important qualitative information on the vulnerability of certain compartments of the soil C pool (Doetterl et al., 2015).

Although microbial communities and their respiration behavior in the laboratory are never the same as in field, and while it is known that field respiration measurements have high coefficients of variation compared to laboratory incubations, these experiments can still yield qualitative information on the potential respiration of soil samples under specific conditions (e.g. at particular moisture, temperature or oxygen levels). In addition, conditions that cannot be observed in the field can be simulated in the laboratory, a fact that is especially important for estimating the effects of future change on soils. At the same time, one must not forget that, at best, laboratory experiments are miniature versions of processes carried out under model conditions and often do not yield results that are reproducible under real-time field conditions (for an overview see Subke et al., 2009).

4. Modeling C dynamics in eroding landscapes

4.1. Process-oriented modeling of C redistribution

An ideal model for SOC redistribution should close the spatial and temporal gap between short-term, local processes and long-term effects on the landscape scale (see Fig. 6). However, a model addressing all potentially important processes on a large temporal and spatial scale does not currently exist, due to limitations in the availability of appropriate input data, problems with the generalization of model parameters, missing process understanding (see Section 2) and, last but not least, immense computing power requirements.

Several process-oriented water erosion models do exist, and are able to simulate event-based erosion with a temporal resolution of minutes and a spatial, mostly raster-based resolution of meters (for a review of erosion models see Jetten et al., 2003; Nearing, 2006; Smith et al., 2010). Especially those process-oriented erosion models simulating the grain size-specific redistribution of different texture classes using selective interrill erosion and selective settling of particles during deposition (e.g. WEPP, Nearing et al., 1989; LISEM, De Roo et al., 1996; EROSION-3D, Schmidt et al., 1999; MCST, Van Oost et al., 2004; Fiener et al., 2008) are particularly well suited to model event based SOC redistribution, including C enrichment and depletion processes.

The main challenges in adapting these erosion models for SOC redistribution simulation are (i) to estimate soil aggregation and aggregate stability, (ii) to implement not only texture but also aggregate classes in sediment transport, and (iii) to allocate C to the appropriate texture and aggregate classes. Due to the complexity of the involved processes, SOC redistribution is implemented in process-oriented models using a simple particle size distribution of the mobilized soil material. Mobilized material is routed through the catchment based on its primary texture class and organic matter content. Sedimentation is considered to be controlled by particle size and density (e.g. WEPP; Foster et al., 1995). Short-term effects of SOC redistribution on CO_2 fluxes following erosion events, as described in Section 2 (e.g. Van Hemelryck et al., 2010), are to our knowledge not implemented in any process-oriented erosion models.

It would be rather simple to couple high resolution, process-oriented SOC redistribution models with long-term SOC turnover models to analyze the interaction between SOC redistribution and turnover. However, a successful application of such coupled models is hardly possible due to the huge data requirements. Hence, more conceptual erosion modelling approaches, using greater time steps (≥1 year), seem to be the only realistic alternative to couple SOC redistribution and turnover models, especially on larger scales. Therefore, to date, coupling erosion to SOC turnover models is in most cases based on USLE (Universal Soil Loss Equation; Wischmeier and Smith, 1978) technology.

The major advantage of these models is that yearly or long-term mean erosion rates are estimated based on a relatively small number of inputs, which are widely available through environmental agencies using USLE technology for soil conservation efforts (e.g. US Department of Agriculture). Based on this data, the spatial distribution of SOC is then updated with yearly time steps and fed into coupled SOC turnover models (e.g. Van Oost et al., 2005a, 2005b). The most commonly used SOC turnover models in this context are ICBM (Andren and Katterer, 1997) and Century (Parton et al., 1987) (Table 2).

At the small catchment scale, models (overview in Table 2) combining mean water (and sometimes tillage) erosion and SOC dynamics are now able to reproduce the spatial heterogeneity of measured SOC stock in fields with different land use types (Liu et al., 2003; Rosenbloom et al., 2001; Rosenbloom et al., 2006; Van Oost et al., 2005a, 2005b; Yoo et al., 2005, 2006; Dlugoß et al., 2012). Based on positive validation results, these types of models have also been applied on regional scales (e.g. Nadeu et al., 2015) and even more parsimonious versions on global scales (Van Oost et al., 2007), but with much greater uncertainty.

Table 2
Overview of coupled soil erosion –SOC turnover models. For each model temporal and spatial scales, main processes modelled, input parameters and land use for which the model was developed is included for comparison.

Model	Extent		Resolution		Modelled process					Input parameters		Focused Land use			Authors
					Erosion		SOC								
	temporal	spatial	temporal	spatial	water	tillage	enrichment/depletion	turnover	delivery	erosion	C Input	arable	grass	forest	
SPEROS-C	annual to decadal	field to regional	annual	5×5 m to 20×20 m	×	×	×	×	×		×	×			Van Oost et al., 2005a, 2007; Wang et al., 2015a, Dlugoß et al., 2012; Nadeu et al., 2015; Fiener et al., 2015
CREEP	decadal to millennial	field to catchment	decadal	10×10 m	×			×	×				×		Rosenbloom et al., 2001, 2006
CENTURY/EDCM	decadal	soil profile	annual	soil profile	×			×			×	×			Liu et al., 2003
SOrCERO	annual	soil profile	annual	soil profile				×		×	×				Billings et al., 2010
(Geo)WEPP/CENTURY	annual	catchment	annual	10×10 m	×		×	×	×			×	×	×	Yadav and Malanson, 2009
CENTURY	decadal	catchment	annual	soil profile				×	×	×		×	×		Harden et al., 1999
CENTURY	decadal	soil profile	annual	soil profile				×		×		×			Gregorich et al., 1998
No name given	millennial	soil profile	annual	soil profile	×			×					×		Yoo et al., 2005

4.2. Shortcomings of current models coupling erosion and C turnover

Coupled erosion – SOC turnover models have a number of shortcomings, primarily associated with their low temporal (and spatial) resolution, and missing process representation of SOC redistribution and the associated C turnover. Moreover, there are still gaps in our understanding of all processes which should be modelled to address C stabilization and release (see Section 2). But most importantly, the data required for rigid model testing and large scale application is currently not available.

SOC redistribution:

Using mean erosion rates, which is the common approach in current SOC redistribution models, it is difficult to implement many of the processes affecting SOC distribution that have been investigated and described on an experimental level. For example, C enrichment due to selective interrill erosion and deposition cannot be integrated very well into conceptual models, as these processes differ in length and intensity from event to event. First attempts to implement empiric relations between mean erosion/deposition rates and C enrichment into SPEROS-C (Fiener et al., 2015) improved the modeling of SOC redistribution significantly and further underlined the importance of the C enrichment process for the overall catchment C balance. In another study, Dlugoß et al. (2012) used SPEROS-C (see Table 2) to show that burial efficiency is substantially increased when it is assumed that the modeled mean erosion is not equally distributed in each year. Hence, ignoring the effects of intensified deposition during single events may lead to an underestimation of the SOC burial efficiency.

SOC turnover related to SOC redistribution:

C turnover models (e.g. RothC (Coleman and Jenkinson, 2014), ICBM (Andren and Katterer, 1997) and Century (Parton et al., 1987)) were designed for stable landscapes and originally developed to explain vertical C fluxes and C sequestration patterns in relation to the soil environment, climate and land cover/management. Changes in the soil environment through lateral fluxes were not included in these models. Soil organic carbon is split into multiple pools, with rate coefficients used to represent the variability in turnover time between the different pools. First attempts have been made to incorporate these models into more complex models of soil development (Yu et al., 2013) or to link the modeled pools with measurable fractions (Zimmermann et al., 2007). While these approaches are highly sophisticated and are able to model the soil C cycle with increasing accuracy in a range of environments, it is difficult to transfer the model concepts to eroding landscapes and to incorporate lateral C fluxes. One difficulty is that the SOC depth distributions at eroding sites (subsoils C depleted) and depositional sites (subsoils C enriched) are different from those at stable sites unaffected by soil redistribution. Second, on top of changing the SOC stock distribution, the removal or deposition of soils also constantly changes the environmental parameters under which C sequestration and release takes place. Third, the lateral flux of SOC from sites of erosion to deposition leads to qualitative changes in C, which need to be accounted for. In summary, lateral soil fluxes create a strikingly different soil setting compared to stable sites, for which C turnover models have been developed.

Soil redistribution not only affects SOC quantity and quality at different landscape positions, but is also directly or indirectly interwoven with other environmental parameters affecting C fluxes from or into soils at different landscape positions. For example, flat areas with large local catchments tend to show higher soil moisture content (e.g. Western and Grayson, 1998; Wilson et al., 2005), than depositional sites. In consequence, a different moisture regime will affect SOC turnover in these areas. To account for spatial variability of soil moisture, Dlugoß et al. (2012) used the wetness index (Beven and Kirkby, 1979) as a modifier of annual SOC turnover. This improved the performance of SPEROS-C in a small catchment in Western Germany, but a generalization to other environmental settings with different moisture and temperature regimes would need further model testing.

Plant growth, and the associated C input to soils via plant residues and roots, is another process which is interlinked with erosion. As erosion changes the physical (e.g. texture) and chemical properties (e.g. via intermixing of subsoil into topsoil or transport of nutrients) of soil, it is obvious that plant growth conditions are directly affected. Without a dynamic representation of the relationship between plant growth and erosion related changes in soil properties, models might overestimate dynamic C replacement (Harden et al., 1999) at eroding sites, since potential changes in plant productivity will not be taken into account. Similarly, in-situ C inputs at depositional sites might be underestimated if improved growth conditions caused by increased water availability or nutrient deposition are not taken into account. Furthermore, when moving from decadal to centennial or millennial time scales, soil weathering and mineral alteration also need to be considered. As minerals change, stabilization mechanisms of C are weakened (Doetterl et al., 2015) and mineral associated C might become more accessible to decomposers again. Models of soil weathering and soil genesis (Opolot et al., 2014) that include SOC turnover for different pools are currently being tested in a range of environments for their robustness. However, only rudimentary attempts have so far been made to include lateral fluxes, and the associated changes in the conditions under which weathering takes place, into long-term coupled soil erosion – SOC turnover models (e.g. Yoo et al., 2005).

4.3. Quo Vadis, C flux modeling?

Currently, to explore SOC dynamics under eroding conditions, a range of different approaches are taken. Some of the first models to investigate SOC dynamics in eroding landscapes used erosion or sedimentation as terms to uplift subsoil and remove topsoil (erosion) or bury and add soil material (sedimentation) in SOC profile models. For example, Harden et al. (1999) and Billings et al. (2010) investigated the balance between the lateral soil loss and in situ replacement of SOC in eroding soil profiles, and highlighted the essential role of management practices in controlling the SOC balance. Wang et al. (2015b) demonstrated, with a similar approach, that sedimentation rates play an important role in determining the long-term fate of buried SOC in colluvial soils. In another study, Wang et al. (2015a) investigated the key factors controlling SOC dynamics in an eroding landscape on the hillslope scale. Functional SOC pools along the hillslope were measured and used to constrain a coupled soil erosion – SOC dynamic model. Spatially explicit models simulating the interaction between geomorphology and ecology were then applied to investigate the effects of various factors, such as soil tillage method and land use pattern, on the lateral and vertical fluxes induced by soil redistribution (Dlugoß et al., 2012; Nadeu et al., 2014; Nadeu et al., 2015; Yadav and Malanson, 2009). Current C redistribution models allow for a reasonable quantification of lateral C fluxes on the field scale. However, SOC in these models is seen as an indifferent component of the mobilized sediments, meaning that carbon is considered to erode in exactly the same way as the soil in which it is stored. For this reason, soil erosion models do not capture the differences in turnover time between C fractions or the qualitative changes in transported C that occur with erosion. First attempts (Wang et al., 2015a), based on empirical modeling, have tried to overcome this by modeling separate SOC fractions during transport and burial. This approach, which takes into account C quality, shows promising improvements in estimating C cycling in dynamic landscapes.

Recent field data enables us to include the interactions and feedback of spatially varying factors, such as C turnover (Doetterl et al., 2012a; Berhe et al., 2014), soil mobilization (Wang et al., 2010, 2013), C stabilization mechanisms (Wiaux et al., 2014a; Wiaux et al., 2014b; Doetterl et al., 2015) or the response of microbial decomposer communities (Park et al., 2014), into new models and to better understand the controls on SOC dynamics at the field scale. However, large data gaps

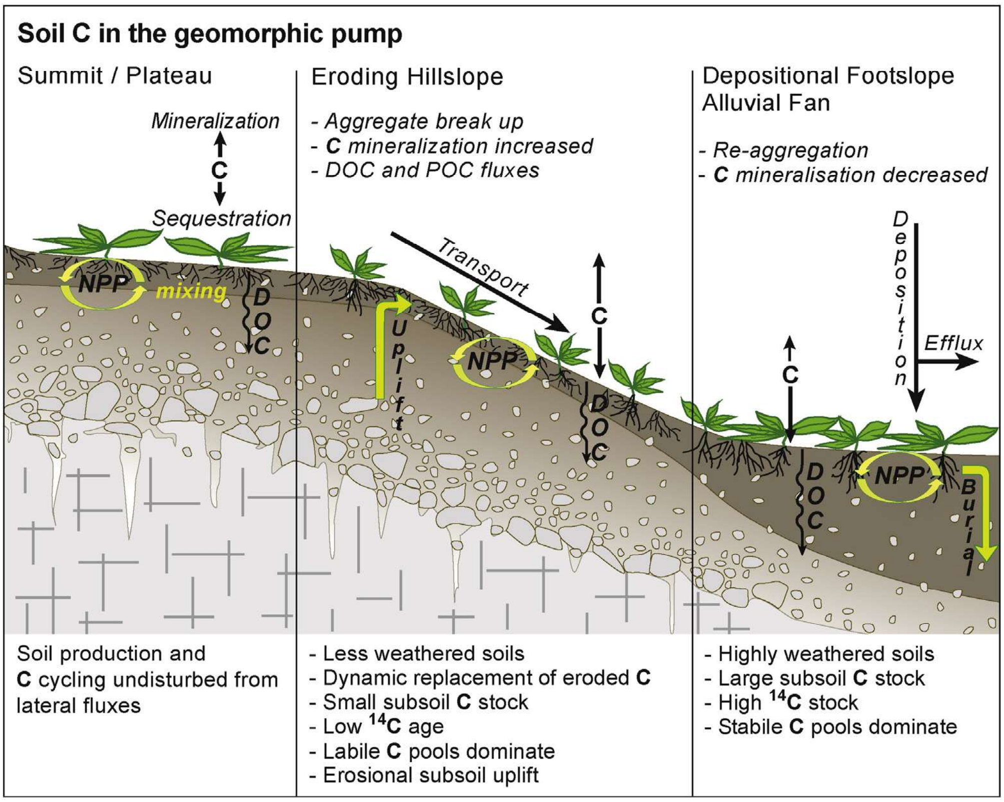

Fig. 4. Overview on soil C cycling in dynamic landscapes and the key mechanisms involved in C sequestration and release from soils. Processes indicated at the plateau/summit position take place along the whole hillslope, but are affected by soil redistribution as described in the text.

remain. For example, soil properties such as soil texture and aggregation, or environmental factors such as moisture and temperature, are found to be spatially and temporally varied at the field scale (Bajracharya et al., 2000; Doetterl et al., 2012a; Rosenbloom et al., 2001; Van Hemelryck et al., 2011; Vanwalleghem et al., 2013). These factors would not only affect SOC dynamics in the soil, but also crop productivity and therefore C input. However, these factors are generally treated as spatially homogeneous in current models of C cycling in eroding landscapes. As a result, physical soil redistribution is identified as the only driver for the spatial variability of SOC.

5. Discussion

5.1. Assessing SOC dynamics in eroding landscapes

The interactions between soil erosion and SOC dynamics are still not completely unraveled, but substantial progress has been made on our process understanding of involved mechanisms (Fig. 4) and how we represent them in models (see references in Table 2).

It has become clear that not accounting for lateral fluxes of soil material, the burial of C and the effects of soil removal on C sequestration and release will lead to large uncertainties when attempting to account for SOC at the landscape scale, and when predicting future C dynamics in these landscapes (Chappell et al., 2015, Nadeu et al., 2015). In order to improve predictions of SOC storage, dynamics and turnover in eroding landscapes, making high resolution data available on larger scales will be a priority in the upcoming decades. This requires further development and implementation of new tools (e.g. FT-IR or NMR spectroscopy, Multi-spectral airborne spectroscopy etc.) (McBratney et al., 2006; O'Rourke and Holden, 2011) in order to gather large qualitative and quantitative SOC datasets at high precision, lower cost and with lower time demands than conventional methods. In addition, an intensification of current networking efforts (e.g. European Soil Data Centre (ESDAC) at JRC) is necessary in order to gather and combine existing data and knowledge from the different SOC and erosion research communities. Furthermore, a critical assessment of established methods is necessary when studying dynamic landscapes affected by soil fluxes (see Section 3.5 and Hernes et al., 2007).

5.2. Critical quantifications

At an appropriate spatial (watersheds or hillslopes) and temporal scale (annual/decadal), soil redistribution processes can constitute a net sink with respect to atmospheric CO_2. However, the erosional sink term is tied to (i) the ability of eroded soils to support C inputs and to stabilize these inputs in order to replace eroded C (Billings et al., 2010, Harden et al., 1999), (ii) the burial efficiency of transported and deposited C, taking into account ongoing alteration of the soil mineral matrix and the changing effectiveness of C stabilization mechanisms (Van Oost et al., 2012; Wang et al., 2015b, Wang et al., 2014b), and (iii) the fate of exported SOC (Mayorga et al., 2005;Smith et al., 2005; Raymond et al., 2013). However, to our knowledge, no direct experimental quantifications of the full C balance at the catchment scale, including the complete net ecosystem (C) exchange, exist. For example, little is known about the coupling between terrestrial and aquatic carbon cycles (i.e. Galy and Eglington, 2010, Aufdenkampe et al., 2011). On the one hand, eroded terrestrial carbon might be stored more efficiently under anaerobic conditions, supporting a C sink. Additional nutrients supplied by terrestrial erosion could also support the growth of more biomass in aquatic systems. On the other hand, fresh terrestrial nutrient sources could induce priming of aquatic C, leading to increased C loss from the aquatic environment, or in other words, an additional C source. The inorganic C leached from eroding sites on carbonate containing parent material is another often overlooked issue in the sink/source discussion. This is

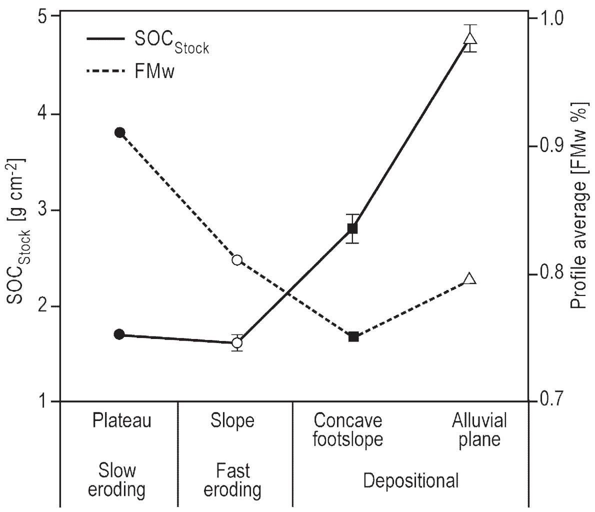

Fig. 5. SOC stocks and weighted fraction modern C (FMw), based on ^{14}C bulk soil data in eroding and depositional profiles that were analyzed down to the soil-saprolite boundary. The study was conducted in a naturally eroding grassland system within a zero-order catchment in northern California (Berhe et al., 2008, 2012a). The carbon stock is twice as large in depositional landform positions compared to eroding ones. Depositional soil profiles had low FM_w values, suggesting that they are effective at storing C and that they store a lot of old C, as determined by radiocarbon dating. On the other hand, soil profiles at eroding landform positions had high FM_w values, indicating that most, if not all of the soil C in these profiles, tends to be younger. Error bars for the SOC stocks (SOC_{Stock}) represent standard errors for n = 3 in the plateau and concave footslope and for n = 4 in the slope and alluvial plain.

of great concern, as many of the most productive agricultural systems are developed on soils from carbonate-rich loess, typically containing between 4 and 5% inorganic C (see for example Rommens et al., 2005). This commonly exceeds the organic C stored in agricultural soils of a similar soil volume. Through soil removal, the loess substrate at eroding sites is brought closer to surface and into a zone of faster soil weathering, where it becomes decalcified (Doetterl et al., 2015). However, the dissolution of C and its re-precipitation elsewhere is a CO_2 neutral process (Gislason, 2005). Ultimately, the sink or source function of agricultural erosion will depend on whether erosion stimulates the conversion of CO_2 into more SOC (i.e. through dynamic replacement) while maintaining high biomass production and whether this C remains in soils as a stable fraction (i.e. through burial) over longer periods.

5.3. Implications of C burial and release for vertical soil C flux on a global scale

Erosion is an important mechanism for C burial and stabilization at the landscape scale (Berhe et al., 2012a; Chaopricha and Marín-Spiotta, 2014; Marin-Spiotta et al., 2014; Doetterl et al., 2015; Vandenbygaart et al., 2015). Soils along a catena or toposequence typically have a varying amount of C in deep soil layers (Dlugoß et al., 2010; Mabit et al., 2008; Ritchie et al., 2007) and also strong variations in C turnover (i.e. Hu and Kuhn, 2015). For example, Berhe et al. (2012a) found that depositional landform positions not only have large SOC stocks (due to depositional C input and high rates of in-situ plant productivity), but also lower Weighted Fraction Modern (FMw) values, suggesting that a higher proportion of C cycles more slowly and is more effectively stored at depositional compared to eroding positions (Fig. 5).

The importance of C accumulation at depositional sites in cropland cannot be overstated (Quinton et al., 2010). Carbon burial at depositional sites plays a central role in the erosion-induced C sink. In extreme cases, particularly when mineralization is additionally constrained, accumulation of SOC at depositional sites in cropland can reach up to 2–4 Mg C ha^{-1} y^{-1} (Vandenbygaart et al., 2012).However, under anaerobic conditions the production of methane may counteract some of the potential C sink, especially if the depositional site is a wetland (for example Mitsch et al., 2012). In addition, there is growing evidence of a connection between terrestrial and aquatic C cycling, with terrestrial C inputs potentially priming C decomposition and disturbing/altering carbon cycling in aquatic systems (Bianchi, 2011; Bauer et al., 2013; Regnier et al., 2013). Lastly, it should be noted that the potential sink function (of depositional sites) might be largely temporally and spatially limited (Van Oost et al., 2012; Doetterl et al., 2015). Due to continuous weathering and hence weakening of various protection mechanisms, and as a result of potential changes in micro-climatic conditions at depositional sites, buried C is not expected to indefinitely remain stable, but will instead turn-over and, slowly but steadily, be released back into the atmosphere with a delay of decades to centuries. By delaying the release of C to the atmosphere, soil erosion and the deposition of C can constitute a net C sink when considering decadal to centennial timescales. The management of sites where C is present in deeper soil layers

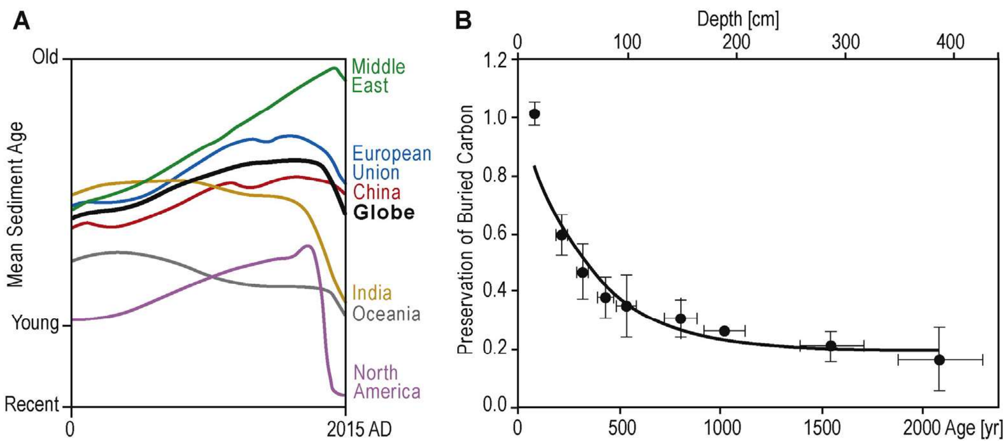

Fig. 6. Left panel (A) shows the relative age of the total mobilized sediment package (i.e. the sum of all sediment mobilized in an area through agricultural water and tillage erosion) in different world regions for the last 2015 years (extrapolations based on Doetterl et al., 2012b and Van Oost et al., 2012). A high sediment age (OLD) indicates strong mobilization and deposition of soil through erosion in the past, but not in recent times. A low sediment age (YOUNG) indicates increasing mobilization in recent times, in greater volumes than in phases of historical soil redistribution. Recent sediment age (RECENT) indicates that the total sediment deposit is dominated with material mobilized within the last 200 years. Right panel (B) shows the decrease of C in buried sediment over time in an exemplary central European catchment (Dyle river) with long lasting soil deposition at foothills through agricultural erosion (Van Oost et al., 2012). The relative ages of C buried within the sediments shown in panel A have resulted from the decay of buried C, following similar patterns to that which is shown in panel B. Old sediments will have lost a good portion of originally buried C, with the remaining fraction being very stable. Vice-versa for young sediments.

is crucial for maintaining this temporary C sink. For example, the destruction of agricultural terraces, which suddenly exposes buried C to surface conditions, or the draining of water saturated alluvial plains can both increase decomposition rates of formerly stabilized C. Information on the time since mobilization and burial can help with assessing the amount of C stored in buried soils and the vulnerability of this buried C to future release back into the atmosphere (Fig. 6) (Van Oost et al., 2012). The time since deposition is therefore a crucial variable in this regard. For example, sediments deposited in the Archaic Period (here defined as the Period of the first rise of agricultural civilizations in the Mediterranean, Middle East and China) can be expected to be depleted in C relative to comparable younger sediments, as the C buried in the Archaic Period has been facing a slow but long period of decomposition. Little C release to the atmosphere is hereafter expected from these soils (Fig. 6), as the more labile C fractions have already been decomposed. Hence, SOC stocks in old sediments (i.e. the Middle East or Southern Europe) might represent a fairly insignificant source for future C release from buried soils, while sediments from modern times (i.e. Oceania or North-America, where most erosion by agricultural starts with the arrival of European settlers) might still hold a significant portion of the originally buried C. These young sediments are therefore more prone to release C back to the atmosphere, since the buried C is less decomposed and presumably more labile than the C remaining in older sediments.

5.4. The interrelationship of controls on topography related SOC patterns

The last decade has seen a growing number of studies dealing with the control of topography on SOC distribution patterns and these studies have delivered some key insights (Stallard, 1998; Harden et al., 1999; Van Oost et al., 2007; Smith et al., 2007; Berhe et al., 2008). The investigation of topographic control on SOC patterns primarily considers the effect of hillslope position on SOC. Some studies have tried to clarify the effect of erosion/deposition on vertical C fluxes at different landscape positions through in-situ measurements. Some of these are solely focused on the effects of soil redistribution on autotrophic and heterotrophic soil respiration (e.g. Fiener et al., 2012), but there are also first attempts to measure the effects of soil redistribution on the full continuum of C assimilation and mineralization (Hoffmann et al., 2015a). However, soil redistribution is in most cases not the only process determining the spatial distribution of SOC, even within single fields with strong topographic gradients (Quine and Van Oost, 2007).

To understand the overall effect of soil erosion on C fluxes, the spatial pattern of other controls on SOC distribution patterns need to be analyzed too. Generally, these controls can be divided into (i) factors affecting the spatial distribution of SOC stabilization/decomposition processes (see Section 2) and (ii) factors affecting plant C assimilation and hence C input to soils. Both of these controls are often indirectly related to soil erosion and surface runoff, since these processes create different site conditions (e.g. soil moisture and nutrient status) along hillslopes. In addition, it is important to highlight that, while at the plot level a preferential transport of SOC over mineral particles may be essentially related to soil texture and rainfall characteristics (Schiettecatte et al., 2008b), at coarser spatial scales preferential transport of C will be affected by topography, runoff paths and land use structure (Fiener et al., 2008, Schiettecatte et al., 2008a). Hence, identifying the effect of soil redistribution on SOC patterns will be strongly dependent on the spatial and temporal scale of observation and the methods used to assess these patterns.

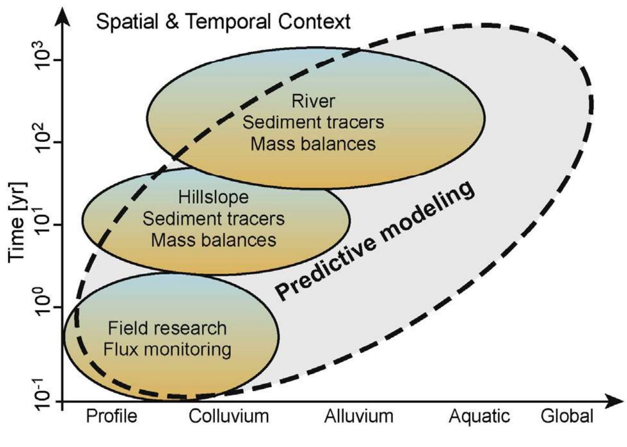

Fig. 7. Schematic diagram of typical temporal (Y axis) and spatial (X axis) extent and resolution of experimental studies analyzing soil erosion effects on SOC dynamics.

5.5. Linking short-term C fluxes with long term SOC pools

As indicated above by the large number of uncertainties associated with different assessment methods and the interrelationship of key control factors, researchers still struggle to address the processes at play on all necessary spatial and temporal scales. The main challenges are that (i) short-term (minutes to years) sediment and C flux measurements need to be related to long-term (decades to millennia) soil redistribution patterns and SOC stock assessments and that (ii) plot scale findings need to be integrated or extrapolated to the landscape or large catchment level (Fig. 7). To our knowledge, there are no studies measuring high resolution C fluxes long enough to see SOC stocks evolve. Nevertheless, we need to close this gap in temporal extent and resolution to gain more insight into the involved processes. A similar dilemma exists regarding the spatial coverage of existing experiments. To understand the event-based nature of erosion processes on SOC redistribution, spatially explicit measurements of SOC transport are taken from plots or small catchments (i.e. Schiettecatte et al., 2008a, 2008b; Wang et al., 2010). However, for larger catchments spatially explicit monitoring of this sort does not exist and therefore the effects of erosion on SOC redistribution is instead studied by investigating only the integrated SOC output from the catchment at its outlet.

Overall, the most promising approach to explore the erosion and C source/sink function of a catchment is to use a combined monitoring, inventorying and modeling approach, which uses insights from the landscape scale flux measurements to improve our models but also uses the long-term SOC stocks for model validation (i.e. Dlugoß et al., 2012). This requires linking process monitoring with new techniques, such as molecular fingerprinting and/or the use of biomarkers and tracer proxies, in order to spatially distribute lateral and vertical C flux measurements. In other words, this approach combines isotopic, spectroscopic and traditional wet chemistry methods in order to determine how and to what extent eroded C is decomposed/lost after erosional redistribution (i.e. Stacy et al., 2015).

5.6. Predicting and modeling dynamic landscapes

Without exception, current models of SOC turnover which include lateral fluxes have provided strong support for the assertion that lateral fluxes exert an important control on C dynamics at the landscape scale (Table 2). Studies (Nadeu et al., 2015; Sanderman and Chappell, 2013) have shown that neglecting effects related to lateral soil fluxes induced by soil erosion could result in significant uncertainties in SOC dynamics at various scales. Neglecting lateral soil fluxes may also amplify or dampen SOC turnover at different landscape positions and soil depths. Modeling the spatial and temporal dynamics of the environment (e.g. soil moisture, plant growth, nutrient dynamics) in which SOC sequestration, burial and mineralization take place is the next logical step to advance C modeling.

Finally, current state-of-the-art soil/ecosystem models are capable of simulating changes in SOC cycling and soil dynamics in response to management practices or environmental/climate change, at both the soil profile and regional scale (Gottschalk et al., 2012; Vleeshouwers and Verhagen, 2002). However, soil degradation processes like erosion are usually not represented in these models. Additionally, in order to incorporate the effects of lateral soil movement on C cycling into earth-system models, we need to develop ways to take changing conditions related to management practices or future climate change into account, adding additional complexity to these models (Nadeu et al., 2015). In conclusion, we encourage the development of coherent approaches that capture the interactions between erosion and SOC cycling processes, particularly on larger scales, in order to transfer knowledge on SOC dynamics and soil erosion to scales that are relevant for policy makers and for earth-system models.

5.7. The link to earth system models

Earth System models (ESMs) are increasingly used to understand and predict the interactions between different compartments of the earth system. Land cover change has received considerable attention, as it has driven changes in climate, global bio-geochemical cycles and biophysical processes, and these processes are now well represented in state of the art Earth System Models (IPCC, 2014). More recently, research has been targeted towards improving the implementation of land management (e.g. fertilizer use, pesticides; Luyssaert et al., 2014). Although the erosional disturbance introduced by human activities is well documented (see Section 1.1), lateral transfers of sediment and associated C (plus related feedback mechanisms) have not been implemented in ESMs.

Recent work demonstrates that lateral fluxes of carbon at the terrestrial-aquatic interface have profound ramifications for the terrestrial carbon balance and the anthropogenic CO_2 budget (e.g. Regnier et al., 2013; Chappell et al., 2015) and that physical erosion, lateral transport and burial of particulate organic carbon (POC) plays a key role (Galy et al., 2015). The focus of this work, however, is mainly on the transformation of carbon in inland waters, estuaries and the coastal ocean and on the effects of lateral transport on inland C (Aufdenkampe et al., 2011; Battin et al., 2009; Cole et al., 2007). While these are important factors for global C cycling that need to be implemented into ESMs, much research is still needed to finally understand how lateral soil fluxes connect terrestrial and aquatic C cycling (Luo et al., 2015).

Terrestrial C cycling and soil erosion, however, have been researched for decades, both separately and more recently in combination, and could likely be implemented into ESMs much more easily. Given that fluxes related to human-induced erosion in upland soils are large, and that a large fraction of the mobilized material does not reach inland fresh-waters (Walling, 1990; Collins and Walling, 2004; De Vente and Poesen, 2005), ESMs should also have a representation of upland erosional fluxes of carbon and its biogeochemical transformation. Erosion and lateral C processes have been studied at the scale of small hillslopes and watersheds, but their magnitude remains poorly constrained at the scale of regions, continents and the globe. A large part of the uncertainty is derived from the lack of global-scale mechanistic models (Doetterl et al., 2012b). Addressing this issue will require new erosion models and representations of carbon feedback mechanisms that are compatible with the scale and resolution of ESMs.

6. Conclusions

6.1. Where do we stand?

The increase in knowledge on SOC during the last two decades has been substantial, and we are now able to predict and model SOC dynamics in large variety of settings and scenarios, including under changing climate, land use and land management. It is now understood that SOC stability is more closely tied to a variety of environmental factors than it is to its intrinsic molecular structure. This shift in paradigm is leading to more and more interdisciplinary research approaches, increasing the awareness of the importance of soils for Earth's ecosystems.

Data mining and the creation of global datasets such as, for example, WISE 3.1 (Batjes, 2009) allow large scale studies to be conducted and validated. The rise of new technologies in remote sensing enables us to produce high quality soil data in large quantities even for remote areas. As a result, the global body of data on soil properties is growing faster than ever before and leading to the development of a new set of tools to be used for detailed soil analysis on a large number of profiles distributed throughout the landscape (McBratney et al., 2006; Janik et al., 2007; Calderón et al., 2011). A recent regional scale modeling study on lateral C fluxes (Nadeu et al., 2015) indicates that while lateral export of C from cropland may have approximately the same magnitude as additional C sequestration in C depleted eroded soils, the ultimate impact of erosion on the landscape C balance is also determined by the fate of the exported C. This requires integrated research on terrestrial and aquatic systems, which is only just beginning.

The study of SOC redistribution is currently profiting from the long history of soil erosion and sedimentological research at different scales. Various models have been developed, from the profile to the landscape scale, that are well capable of representing observed SOC stock variations as well as mechanisms like SOC dynamic replacement at eroding sites and SOC protection at depositional sites. Coupled soil redistribution and soil organic carbon turnover models, which account for water and tillage erosion, already show great potential to study erosion induced C fluxes, especially on arable land. Field to landscape scale manipulation experiments offer the chance to test our understanding of the soil-plant system with respect to annual to decadal CO_2 sink-source relations and to come to a better understanding of the drivers behind full C balances. Some of the latest and most advanced works in this area are those that investigate the fate of eroded C using a combination of isotopic, spectroscopic and traditional wet chemistry approaches in order to determine how and to what extent eroded C can be decomposed/lost after erosional redistribution. Part of the answer to this question lies in determining how erosion transports different pools of C from different types of ecosystems. Recently, a variety of approaches, including catchment scale C flux monitoring studies, the use of tracers, the acquisition of large spatial datasets that provide knowledge on erosion-SOC interactions and the calibration and validation of coupled soil erosion and C dynamics models, have provided us with the insights needed to work towards the implementation of lateral SOC fluxes into Earth System Models. Success in this endeavor will lead to better predictions of soil response and a better understanding of the role of soils in a changing world.

Table 3
Summary of research priorities, as seen by the authors.

Category	The way ahead
Scale	Prioritize catchment scale research (internal processes)
ESM	Integrate erosion-induced changes in soil carbon dynamics
SOC stabilization	Focus on changes in physical protection of SOC and changes in environmental and geochemical (mineral) variables at eroding and depositional sites
Land use	Expand to other land uses
Processes	Include wind erosion
Nutrients	Combine the study of SOC with that of other nutrients (N, P)
Methods	Development of remote sensing tools to increase detailed measurements over larger spatial scales
Scope	Integrate terrestrial and aquatic systems
Models	Integrate selectivity on transport and changing environmental conditions (e.g. moisture)
Disciplines	Integrate biogeochemical and microbial with geomorphological and climatic variables

6.2. The way ahead

Soil redistribution through accelerated erosion is a global phenomenon that is affecting all kinds of land uses and undulating land forms. Nevertheless, current Earth System Models fail to represent these processes when it comes to C dynamics, leaving large uncertainties for estimating the future development of the C cycle. In some ways we are still at the very beginning of our journey to fully understand the dynamics of CO_2 sinks and sources related to soil redistribution. Research is still needed to clarify the role that lateral soil redistribution and related factors, such as bioavailable nutrient distribution, play in determining the stability of organic matter and also to enhance our understanding of the interactions between terrestrial and aquatic C cycles. This is of special importance because the biogeochemical cycles of C, N and P are closely coupled to each other, but so far in erosion related research, the focus has been on terrestrial C dynamics alone, neglecting potential interaction effects with N and P. Hence, future research will require the combined (long term) study of biogeochemical and microbial soil parameters as well as geomorphological and climatic variables (Berhe et al., 2012c; Park et al., 2014; Doetterl et al., 2015) across a large temporal and spatial continuum. This requires more studies at the catchment scale, where catchment internal processes can be investigated. To accomplish this, there is a strong need for an international network of manipulation experiments, with a unique methodological and technical design.

On the modelling front, there are still some major challenges to model full C balances: At the field scale, models should be developed which not only account for the effect of physical soil redistribution by erosion, but also explicitly describe the effect of other topography-dependent factors, such as temperature and moisture, in order to fully represent the spatial variation of SOC dynamics. At regional and global scales, models that account for various natural and anthropogenic factors affecting lateral SOC fluxes are needed. In addition, modelling routines which integrate C enrichment during erosion need to be developed, spatially variable environmental factors affecting C input and SOC turnover(e.g. dynamic soil moisture) need to be included and existing models, which are mostly focused on arable land, need to account for other land uses.

Finally, at the continental or global scale, it is essential to couple these terrestrial models with C turnover models, taking aquatic systems into account to ultimately complete and improve the Earth System Models (Table 3) and to also integrate wind erosion into ESM's. There is evidence that elemental cycling and C dynamics in tropical rainforests are connected and partly driven by global scale dust transport (Boy and Wilcke, 2008). Furthermore, there is indication for a coupling between terrestrial and marine biogeosystems through long-distance dust transport (Shao et al., 2011) (Harrison, 2000). Dust emissions from arable land has been modelled to account for 25% of the global dust (Ginoux et al., 2012) and its relevance, which is due to the related carbon and nutrient fluxes, should not be neglected in future Earth System research.

With this review, we highlight the advances made methodologically and in process-understanding to advance from profile based SOC dynamics to a landscape scale understanding of the interplay of lateral and vertical C fluxes. In order to ingrate lateral C fluxes and their significant dependency on environmental and management constraints, it will be crucial to use the upcoming years to close the gap between small scale process understanding and large scale representation of lateral C fluxes in global models.

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