The mineralization and sequestration of organic carbon in relation to agricultural soil erosion

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ABSTRACT

The coupling of soil erosion (especially interrill erosion by water) and the dynamics of soil organic carbon (SOC) in agricultural landscapes has been widely studied over the past two decades. To date, however, the role of soil erosion in global C cycle remains a topic of debate. Numerous questions remain to be addressed before determining the C sink/source effect of soil erosion, especially for the mineralization and sequestration of eroded SOC upon erosion, transport and deposition. In this review, we provide a comprehensive cross-disciplinary review on SOC mineralization and sequestration at sites of erosion, along the transport pathway and at depositional sites. The current state of knowledge on the impacts of erosion-induced soil aggregate breakdown and formation, removal of SOC from eroding sites and deep burial of SOC at depositional sites on the mineralization and sequestration of SOC are presented. Furthermore, we provide an overview of the conceptual relations between soil biological properties (microbial abundance, species diversity, community composition and enzyme activity) and the mineralization and sequestration of SOC in eroded agricultural landscapes, which are often overlooked by previous research and reviews. The comprehensive understanding of physical, chemical and biological mechanisms affecting the mineralization and sequestration of eroded SOC provides important insights to balance the global carbon budget and finally holds the answer on the carbon sink/source controversy.

1. Introduction

Over the past 200 years, the emission of greenhouse gases (GHGs), particularly carbon dioxide (CO_2), has rapidly increased (IPCC, 2014). The global CO_2 concentration has risen from 280 to 382 ppm in 2007, with an increase rate of 0.88 ppm yr^{-1} (Canadell et al., 2007). This rapid increase in CO_2 concentration has raised concerns regarding the identification of sources and sinks of CO_2 (Lal, 2003). Although lots of significant efforts and works have been conducted over the past two decades, it is still difficult to quantify and balance the global carbon (C) budget according to currently known C fluxes (Stallard, 1998). The unknown C sinks are estimated at 2–4 Pg C yr^{-1} (Lal, 2003). Identification of unknown C sinks is important for reducing the enrichment rate of GHGs in the atmosphere and developing strategies for mitigating potential climate change. Actually, the current understanding of global

C budget is grossly inadequate because all CO_2 sources and sinks related to soil erosion are not accounted for (Lal, 2005).

Soil erosion by water is a natural geomorphic process and is one of the widespread forms of soil degradation (Doetterl et al., 2016; Liu et al., 2010). The area subjected to accelerated erosion (referring to water erosion and similarly hereinafter) has reached approximately 8.6% of the global land surface area (Lal, 2003). Soil erosion is a three-stage process: (i) detachment of soil particles; (ii) migration and redistribution of sediment over the landscape; and (iii) deposition in low-lying sites or aquatic ecosystems (Doetterl et al., 2016; Ritchie et al., 2007; Zhang et al., 2013). These processes have strong impacts on soil organic carbon (SOC) dynamics (Lal, 2005; Xiao et al., 2017a), with influencing mechanisms mainly involving: (i) removal and replacement of organic carbon (OC) at eroding sites; (ii) OC mineralization during transport; and (iii) deep burial of allochthonous and autochthonous OC
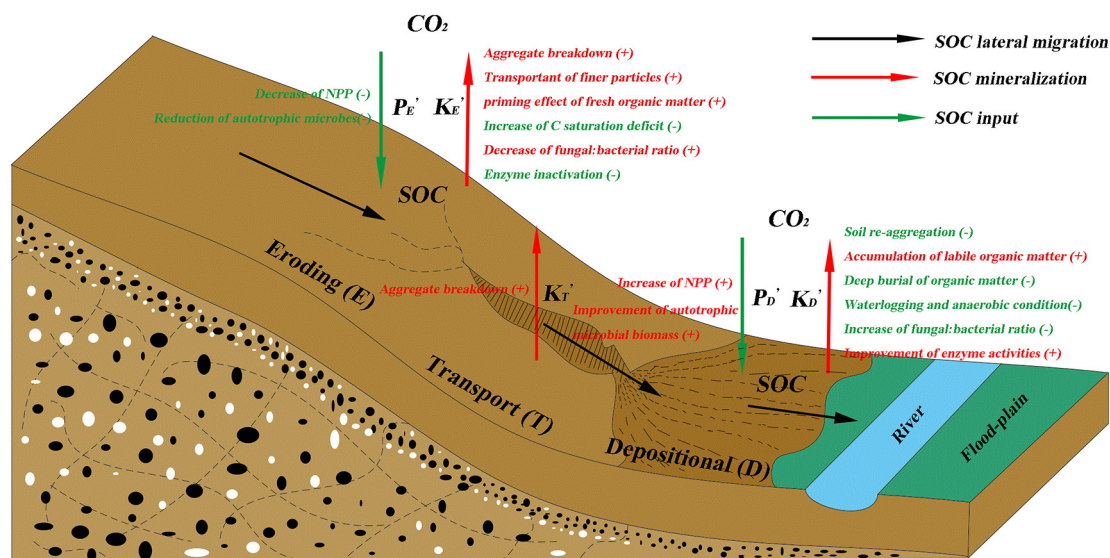
Fig. 1. Migration, mineralization and input of SOC in an eroded slope. P_E' and P_D' represent the post-erosion condition of the SOC input rate at eroding and depositional sites, respectively. K_E' and K_D' indicate the post-erosion condition of the SOC mineralization rate at eroding and depositional sites, respectively. K_T' stands for the SOC mineralization rate during transport. The red letter and " + " in brackets represent a positive effect on the SOC dynamic; the green letter and " − " in brackets indicate a negative effect on the SOC dynamic. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

at depositional sites (Galy et al., 2015; Kirkels et al., 2014; Lal, 2003, 2005; Li et al., 2017). The influences of these mechanisms on SOC dynamics have been extensively studied in recent decades. Until now, however, there has been no consensus on whether soil erosion acts as a net source or sink of atmospheric CO_2 (Lal, 2003; Van Oost et al., 2007). Stallard (1998) applied hydrologic-biogeochemical models to evaluate the impact of sedimentation of eroded soil on the global C cycle and concluded that terrestrial sedimentation contributes to soil C sequestration with a capacity of 0.6 to 1.5 Pg C yr^{-1}. Similarly, Van Oost et al. (2007) by using caesium-137 and C inventories, estimated SOC erosion to result in a global C sink of 0.12 Pg C yr^{-1}. However, contrary to Stallard and Van Oost, the work of Jacinthe and Lal (2001) and Lal (2003, 2004, 2005) indicated that soil erosion constitutes a source of atmospheric CO_2. They concluded that soil erosion releases 0.37 to 1 Pg C yr^{-1} to the atmosphere due to the decrease in net primary production (NPP) and high mineralization rate of eroded SOC during transport.

The controversy over the role of erosion in global C cycle is largely attributed to different understanding on the fate of eroded SOC (predominately mineralized vs. sequestered), which is regulated by complex mechanisms (Berhe et al., 2008; Lal, 2003, 2005; Van Oost et al., 2007). For instance, the migration and mineralization of SOC during transport are closely related to aggregate stability. Chaplot and Cooper (2015) indicated that high aggregates stability induced low dissolved OC (DOC) and particulate OC (POC) losses (lateral migration) in a typical hillslope ($r = -0.81$ and -0.77, respectively) due to increased soil infiltration by water and reduced transport by runoff. By establishing runoff plots on sloping agricultural lands, Polyakov and Lal (2008) investigated the effects of interrill erosion on SOC transport and mineralization, and indicated that approximately 20% of eroded SOC could be lost to the atmosphere during transport due to the breakup of initial soil aggregates. Splash, collision and shear forces induced by raindrops and runoff can decrease soil aggregates stability and accelerate SOC mineralization (Wei et al., 2016). At present, there are considerable disagreements in the existing literatures regarding the magnitude of SOC migration and mineralization during transport (Lal, 2003; Novara et al., 2016). By analyzing empirical data from 240 runoff plots of the world, Müller-Nedebock and Chaplot (2015) indicated that the total amount of SOC displaced by interrill erosion from its source

approximately is 1.32 ± 0.20 Pg C yr^{-1}. Assuming a decomposition rate of 20% for displaced SOC by interrill erosion, the annual C emission to atmosphere would represent 0.26 ± 0.04 Pg C yr^{-1}. This is much lower, based on 20% decomposition rate, than the previous estimation by Lal (2003) of 0.8–1.2 Pg C yr^{-1}. Differences in study methods, temporal-spatial scales and hypotheses may be the main reasons for these disagreements. For example, the mean POC loss rate of 27.2 g C m^{-2} yr^{-1} was hypothesized by Müller-Nedebock and Chaplot (2015), while a SOC delivery ratio of 2–3% was considered by Lal (2003).

On the other hand, the mineralization and sequestration of SOC at eroding and depositional sites are closely related to changes in soil micro-environments (Huang et al., 2013; Nie et al., 2014; Park et al., 2014). Interrill erosion, as a selective process, preferentially transports fine soil particles and light organic matter fractions (density < 1.8 Mg m^{-3}) from eroding sites to the low-lying depositional sites (Lal, 2003), which alters the soil texture in eroded agricultural landscapes and results in a high C enrichment ratio (the ratio of the concentration of SOC in the eroded sediments to that in the original bulk soil) in eroded sediments (Mchunu et al., 2011). Boye and Albrecht (2006) observed in Kenya that eroded sediments (interrill erosion) from 22.2 m long agricultural plot were enriched in SOC by an enrichment ratio of 3.3. Over 80% of eroded sediments are redeposited at the adjacent topography (Chaplot and Poesen, 2012), thus the deep burial of SOC-rich sediment at depositional sites leads to large amounts of SOC being stored in limited oxygen and moisture environments, which inhibits the rapid release of eroded SOC (Vandenbygaart et al., 2015). Furthermore, accompanied with changes in soil structure and nutrient availability, soil microbial abundance, community composition and enzyme activity at eroding and depositional sites also exhibit significant variation. For instance, the loss of SOC and dispersion of soil aggregates at eroding sites can inhibit the rapid growth of soil microorganisms (especially for fungi) and alter the community composition of soil microorganisms (e.g., fungal:bacterial dominance), which have strong impacts on SOC mineralization and sequestration (Trivedi et al., 2013). These changes in soil micro-environments together determine the fate of SOC (mineralization vs. sequestration) at eroding and depositional sites (Kirkels et al., 2014). Therefore, to scientifically identify the C sink/source effect of soil erosion, a comprehensive investigation of the influencing

mechanisms concerning the fate of eroded SOC is required.

In this review, we aim to give an overview of different conceptual relations known to govern the mineralization and sequestration of SOC at sites of erosion, along the transport pathway and at depositional sites (Fig. 1). We discuss (i) criterion for erosion-induced C source or sink; (ii) impacts of erosion-induced changes in biogeochemical soil characteristics (such as soil aggregate, amount and quality of SOC and microbial community) on the mineralization and sequestration of SOC. The conceptual relations between biological soil properties (including microbial abundance, species diversity, community composition and enzyme activity) and the mineralization and sequestration of eroded SOC, often overlooked by previous research and reviews, are highlighted. Through this review, the understanding of the fate of eroded SOC is expected to be further improved.

2. Criterion for erosion-induced C source or sink

Soil erosion has both positive and negative effects on the mineralization/sequestration of SOC (Doetterl et al., 2016). For instance, the mineralization of eroded SOC can be accelerated due to the breakdown of aggregates, while it can also be inhibited due to the deep burial of SOC. Then, how can we distinguish the C sink/source effect of soil erosion? That is, what is the criterion for us to determine whether soil erosion is acting as a net source or sink of atmospheric CO_2? A simple formula was proposed by Berhe et al. (2007) to determine erosion-induced C dynamics.

$$\{P_E' + P_D' - (K_E' + K_D')\} - \{P_E + P_D - (K_E + K_D)\} \qquad (1)$$

Two contiguous sites, upper eroding (E) and lower deposition (D), in an eroded agricultural land were considered. The parameters P and K represented the SOC input rate and mineralization rate, respectively. The pre-erosion condition of the eroding and the depositional sites was represented by P_E, P_D, K_E, and K_D, and the post-erosion condition was indicated by P_E', P_D', K_E', and K_D' (Fig. 1). In this formula, the SOC mineralization rate during transport is overlooked, which is likely to have important implications for the role of soil erosion within the global C cycle (discussed in Section 4.1). To comprehensively consider SOC dynamics during and after erosion, the formula was modified as follow:

$$\{P_E' + P_D' - (K_E' + K_D' + K_T')\} - \{P_E + P_D - (K_E + K_D)\} \qquad (2)$$

where the SOC mineralization rate during transport (T) was indicated by K_T' (expressed in the same units, such as kg C yr^{-1}) (Fig. 1). According to the description of Polyakov and Lal (2004), a soil erosion rate of 10 Mg ha^{-1} yr^{-1} is often considered as a threshold for maintaining soil productivity. Under the condition of long-term accelerated erosion (erosion rate > 10 Mg ha^{-1} yr^{-1}), NPP in eroded agricultural lands is decreased ($P_E' + P_D' < P_E + P_D$). Assuming that the OC mineralization rate is reduced even more due to effective soil re-aggregation and deep burial of redistributed and autochthonous OC at depositional sites ($K_E + K_D \gg K_E' + K_D' + K_T'$), soil erosion constitutes a C sink ($P_E' + P_D' - (K_E' + K_D' + K_T') > P_E + P_D - (K_E + K_D)$) (Berhe et al., 2007). However, when the OC mineralization rate is reduced less or increased due to the breakdown of soil aggregates (Lal, 2003), soil erosion constitutes a source of atmospheric CO_2 ($P_E' + P_D' - (K_E' + K_D' + K_T') < P_E + P_D - (K_E + K_D)$). In addition, under a low or moderate erosion rate (< 10 Mg ha^{-1} yr^{-1}), NPP in eroded agricultural lands is generally remaining steady ($P_E' + P_D' = P_E + P_D$) (Harden et al., 1999). The C sink/source effect of soil erosion is primarily determined by the mineralization rate of SOC at sites of erosion, along the transport pathway and at depositional sites. Thus, clarifying the controlling mechanisms for the mineralization and sequestration of eroded SOC is very important for resolving the C sink/source controversy.

3. Eroding sites

3.1. Dynamic loss and replacement of OC impact its mineralization and sequestration

The dynamic loss (lateral migration) and replacement of SOC and its mineralization and sequestration together determine the soil C budget at eroding sites (Lugato et al., 2016). In this dynamic C system, the mineralization and sequestration of SOC at eroding sites, to some extent, are affected by the dynamic loss and replacement of OC (Doetterl et al., 2016; Fontaine et al., 2007). Under the influence of long-term water erosion, the topsoil on eroded agricultural landscapes is gradually removed, and the subsoil exposed (Liu et al., 2010). Compared with topsoil, OC in subsoil is generally depleted but rather stable (Doetterl et al., 2012; Salomé et al., 2010). It is accepted that, unless the soil is eroded beyond a critical level, the newly assimilated OC at eroding sites should be able to replace, at least partially, OC that was transported by erosion. For example, by investigating SOC erosion, NPP and replacement of eroded C at different landform positions, Berhe et al. (2008) reported that the input of fresh organic matter at eroding sites is high enough to replace the eroded OC (2.75 to 5.5 g C m^{-2} yr^{-1}). Furthermore, Harden et al. (1999) described a case study in Mississippi loess soil area where the original SOC was lost in 127 years after land conversion, but fertilizers enabled 30% replacement after 1950 with rates of SOC recovery exceeding erosion losses. The input of fresh organic matter likely stimulates mineralization of stabilized OC in the former subsoil due to the so-called "priming effect" (Fontaine et al., 2007; Wang et al., 2013a). The phenomenon that soil samples from eroding sites respire substantially more C (µg C d^{-1} SOC g^{-1}) than those from non-eroding or depositional sites can be partly explained by this priming effect (Xiao et al., 2017a). Although the mineralization of stabilized OC in subsoil is accelerated, the significance of this positive feedback in eroded agricultural landscapes is still unclear due to the complex regulatory mechanisms and the difficulty in obtaining field-scale observations (Doetterl et al., 2016). The erosion intensity, vegetation type, organic matter quality and SOC replacement rate at eroding sites are all likely to affect the magnitude of this priming effect (Kuzyakov et al., 2000).

Contrary to the positive priming effect in SOC mineralization, some studies indicated that exposure of C-depleted subsoil and input of fresh organic matter at eroding sites contribute to the sequestration of SOC (Kirkels et al., 2014; Wiesmeier et al., 2014). It has been hypothesized that each soil has a C saturation limit depending on soil texture, bulk density and mineralogy (Six et al., 2002). Decreased SOC sequestration potential with high C versus low C content across different treatments is consistent with the concept of soil C saturation (Bhattacharya et al., 2016; Chen et al., 2016). Soil C saturation deficit is defined as the difference between the theoretical soil C saturation level and the current C content. The C saturation theory manifests that soil C saturation deficit is positively correlated with soil C sequestration potential (Fig. 2a, b) (Stewart et al., 2008). Soil erosion transports C-rich topsoil from eroding sites to low-lying depositional sites and exposes the C-depleted subsoil, resulting in eroding sites having a high soil C saturation deficit (Li et al., 2016), and consequently a great capacity for SOC sequestration (Fig. 2a). Secondly, the removal of topsoil and exposure of subsoil at eroding sites increase the weathering rates of former subsoil leading to changes in soil mineralogy (Arnold, 1991; White et al., 1996). According to White et al. (1996), soil mineralogy in the central valley of California evolved from granitic sand to kaolinite and quartz with the increase of soil age, with a consequent increase in the specific surface area of soil minerals. The more reactive, C poor, mineral surfaces in weathered subsoil (current topsoil) with respect to the original subsoil at eroding sites, may rapidly adsorb organic C from fresh organic matter inputs, and limit the magnitude of the priming effect and SOC mineralization (Stewart et al., 2009; Wiesmeier et al., 2014). However, albeit the lower OC contents, soil texture at eroding
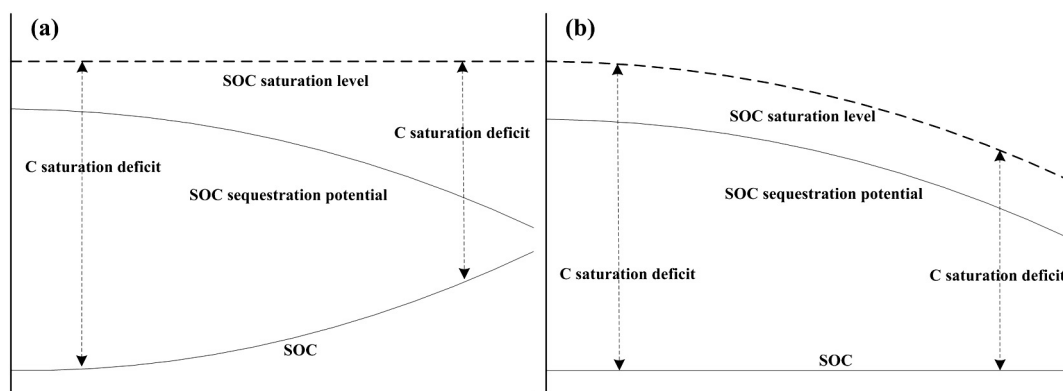
Fig. 2. Conceptual relationships between the SOC saturation limit, current SOC content and sequestration potential. (a) Changes in SOC sequestration potential and content; (b) Changes in SOC sequestration potential and saturation limit.

sites also coarsen with time (Lu et al., 1988; Nie et al., 2014) (discussed in Section 5.1), which may limit soil C sequestration potential due to the decrease in theoretical soil C saturation level (Fig. 2b) (Kimetu et al., 2009; Stewart et al., 2008).

3.2. Changes in microbial community impact SOC mineralization and sequestration

Serious soil erosion not only changes physicochemical soil properties but also strongly impacts the microbial communities in agricultural landscapes (Huang et al., 2013). Firstly, soil microbial communities at eroding sites can be directly impacted by hydrological processes (Huang et al., 2013; Xiao et al., 2017a). Splash, collision and shear forces induced by raindrops and runoff can directly induce the death of microorganisms on the surface soil, especially for microorganisms released from broken aggregates (Hou et al., 2014). Furthermore, erosion-induced changes in physicochemical soil properties, such as soil structure and organic matter contents, also strongly impact soil microbial communities at eroding sites (Dungait et al., 2013; Li et al., 2015). It has been proven that microbial abundance and species diversity decrease in eroded soils due to the dispersion of soil aggregates and nutrient loss (Mabuhay et al., 2004; Li et al., 2015). The reduction in soil nutrients and destruction of habitats limit the rapid growth of soil microorganisms at eroding sites (Huang et al., 2013).

Previously, it was assumed that SOC mineralization is primarily regulated by soil microbial community structure, and the decrease in microbial abundance and species diversity at eroding sites has a negative impact on mineralization rates (Don et al., 2017; Garcia-Pausas and Paterson, 2011). However, recently, an increasing number of studies indicate that erosion-induced decrease in microbial abundance and species diversity has no significant impacts on SOC mineralization (Wertz et al., 2006; Xiao et al., 2017a, 2017b). Microbial communities

in soil, being very diverse, appear to be characterized by a high functional redundancy (e.g., C mineralization, nitrification and denitrification) (Birge et al., 2015). That is, microbial abundance and species diversity in soil far exceed that of being requisite to maintain normal soil function. Although a decline in soil function may occur at a specific minimum value of microbial biomass, a moderate reduction in microbial abundance and species diversity induced by erosion has little effect on overall soil processes because the surviving microorganisms can offset their function (Kemmitt et al., 2008; Nannipieri et al., 2003). By simulating microbial diversity erosion, Wertz et al. (2006) demonstrated that the vast diversity of soil microbiota makes soil ecosystem functioning largely insensitive to biodiversity erosion. Furthermore, by investigating the relationships between SOC mineralization, and abiotic and biotic soil properties in eroded farmlands, Xiao et al. (2017a) indicated that the effects of erosion-induced changes in microbial abundance and species diversity on SOC mineralization were smaller than the potential effects of organic matter quality changes. Compared with microbial community properties, erosion-induced changes in labile organic matter had a more important effect on mineralization rate of SOC. A 'regulatory gate hypothesis' was proposed to further explain these phenomena (Kemmitt et al., 2008). The hypothesis considers the mineralization of soil organic matter as a two-stage process (Fig. 3). First, non-biodegradable soil organic matter is slowly altered to feed the available pool through abiotic processes (e.g., desorption, chemical oxidation and hydrolysis) that are independent of, and cannot be altered by, microbial community (step I). Second, organic matter slowly entering the available pool is rapidly assimilated and respired by excess microbial biomass (step II). The abiotic organic matter transformation process (step I) is the rate-limiting step of SOC mineralization, which is termed as the "regulatory gate" (Fig. 3). Despite soil is inhabited with a huge number of microorganisms, the limited bioavailable organic matter makes soil microorganisms show obvious functional redundancy
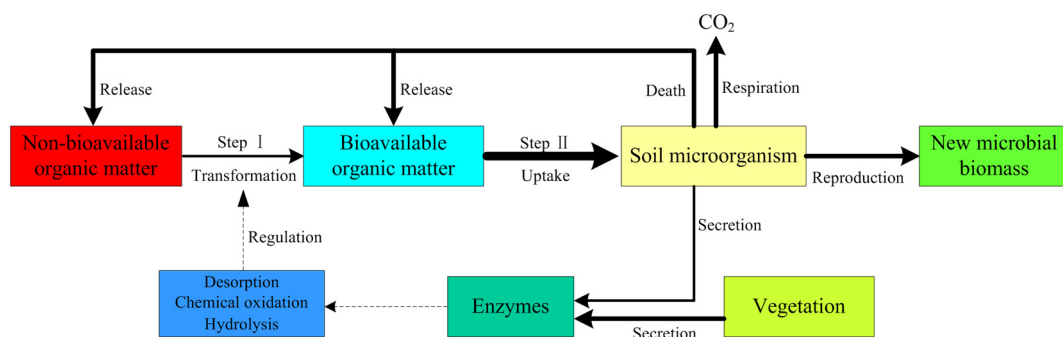
Fig. 3. Conceptual scheme of the regulatory gate hypothesis (modified and adapted from Kemmitt et al., 2008). Step I is the abiological transformation of non-bioavailable organic matter. Step II is the uptake and assimilation of bioavailable organic matter by microorganisms. Solid lines indicate flows of material, and the width of the solid lines represents the material flow rate. Dashed lines indicate regulation processes of C turnover.

(OC mineralization). Thus, the negative impacts of soil erosion on microbial abundance and species diversity have limited effects on SOC mineralization.

Apart from total microbial abundance and diversity, agricultural erosion-induced changes in microbial community composition were also observed (Huang et al., 2013; Park et al., 2014). Huang et al. (2013) investigated the impacts of interrill erosion on microbial communities in an agricultural land and found that fungal abundance decreased to a greater extent than bacterial abundance. The ability of bacterial and fungal communities to resist soil erosion is different (Six et al., 2006). The diameters of bacteria and fungi are typically 0.5–1 μm and 3–8 μm, respectively. Bacteria with a smaller size are generally protected in micropores and aggregate interiors, while fungi reside in macropores and on the surfaces of soil aggregates (Nannipieri et al., 2003). Thus, it is widely accepted that fungi are more sensitive to soil erosion due to the lower degree of protection (Trivedi et al., 2013). Compared with undisturbed soils, a lower fungal:bacterial dominance is generally observed in eroded soils, which is supported by Huang et al. (2013), Trivedi et al. (2013) and Xiao et al. (2017b). Further, the fungal:bacterial dominance of a site is related to its C sequestration potential. A greater potential is always observed in soils with higher fungal:bacterial dominance due to the high C-use efficiency of fungi and low biodegradability of its cell wall (Bailey and Smith, 2002; Stockmann et al., 2013; Strickland and Rousk, 2010). Therefore, decrease in fungal:bacterial dominance at eroding sites is expected to increase the turnover rate of C, which has a negative effect on soil C sequestration.

Although various studies, focusing on erosion-induced changes in microbial communities and associated SOC dynamics, have been conducted over the past decades (Huang et al., 2013; Li et al., 2015; Park et al., 2014; Xiao et al., 2017a), the regulatory mechanisms of soil microorganisms on the mineralization and sequestration of eroded SOC are not well understood so far. For example, erosion-induced changes in some important functional microorganisms (e.g., autotrophic microorganisms) and associated SOC dynamics are overlooked. Autotrophic microorganisms are known to contribute significantly to CO_2 assimilation in aquatic systems (Cannon et al., 2001). Recently, an increasing number of studies have found that autotrophic microorganisms also play an important role in soil C sequestration (Ge et al., 2012; Wu et al., 2015; Yuan et al., 2012). By combining a ^{14}C isotope tracer technique and incubation experiment, Ge et al. (2012) investigated the C sequestration potential of autotrophic microorganisms in paddy soils and indicated that the amount of SOC sequestered by autotrophic microorganisms ranged from 114.3 to 348.2 mg C kg^{-1}, accounting for 0.73% to 1.99% of total SOC. Assuming a terrestrial area of 1.4×10^8 km^2, autotrophic microorganisms have a global C sequestration potential of 0.6 to 4.9 Pg C yr^{-1} (Yuan et al., 2012). Even though autotrophic bacteria play a key role in soil C sequestration (Yuan et al., 2012), the responses of autotrophic bacterial community

and associated C sequestration potential to soil erosion still lacks understanding. Erosion-induced changes in quality and quantity of soil organic matter may have strong impacts on autotrophic microbial community composition (i.e., the proportion of obligate autotrophs and facultative autotrophs in autotrophic microbial community), which can regulate soil C sequestration potential in eroded agricultural landscapes due to different CO_2 assimilation rates between obligate and facultative autotrophs (Badger and Bek, 2008). Clearly, further research on this aspect is definitely required and essential to precisely evaluate the net effect of changes in microbial communities on soil-atmosphere C exchange in eroded agricultural landscapes.

With the rapid development of molecular biology, high-throughput sequencing and metagenomics have recently been developed and widely applied to investigate microbial community composition and function (Trivedi et al., 2013; Xiao et al., 2017b). These molecular approaches enable us to accurately evaluate erosion-induced changes in soil microbial communities at genus and species level. Furthermore, the combination of molecular approaches and isotopic tracer techniques (e.g., ^{13}C, ^{15}N) is highly recommended to evaluate the roles of erosion-induced changes in microbial communities in SOC mineralization and sequestration due to their high accuracy and efficiency (Long et al., 2015; Novara et al., 2016).

4. Transport

4.1. Mineralization of SOC during transport

Apart from in-situ mineralization of SOC at eroding and depositional sites, off-site transport of SOC also contributes to its depletion (Olson et al., 2016; Stockmann et al., 2013). The breakdown of soil aggregates during transport exposes the former encapsulated SOC making it susceptible to mineralization by microorganisms (Goebel et al., 2009; Wei et al., 2017). Mass balance approaches and laboratory simulations were widely used in previous studies to estimate the mineralization magnitude of SOC during transport (Lal, 1995; Smith et al., 2001). Assuming that the SOC level in agricultural land is at steady-state equilibrium (that is, crop residue addition and organic amendment compensate for SOC losses which occur through mineralization, leaching and surface translocation with runoff), Jacinthe and Lal (2001) indirectly estimated 0.37 Pg C yr^{-1} was released to the atmosphere during transport at global level. Furthermore, applying a laboratory simulation experiment, Van Hemelryck et al. (2010) quantified erosion-induced changes in C exchange between soil and atmosphere for different initial soil conditions and erosion rates, and showed that soil redistribution process contributes to an additional emission of 2 to 12% of total C contained in eroded sediment (Table 1).

SOC mineralization induced by this disturbance was widely explored over past decades, however, the mineralization magnitude is still controversially discussed (Doetterl et al., 2016; Lal, 2003). Smith et al.

Table 1
The mineralization of eroded SOC during transport.

Fraction mineralized (%)	Measuring methods	Soils	Monitoring time (days)	References
Near 0	Mass balance model	–	–	Smith et al. (2001)
1.6	Field monitoring (natural rainfall + incubation)	Agricultural soil (silt loam)	112 (May to September)	Van Hemelryck et al. (2011)
< 5	Mass balance model	–	–	Van Oost et al. (2005)
2–12	Laboratory simulation (simulated rainfall + incubation)	Agricultural soil (silt loam)	98	Van Hemelryck et al. (2010)
15	Field monitoring (natural rainfall + incubation)	Agricultural soil (silty clay loam)	68 (July to September)	Polyakov and Lal (2008)
20	Mass balance model	–	–	Lal (1995)
20	Mass balance model	–	–	Jacinthe and Lal (2001)
43	Field monitoring (natural rainfall + incubation)	Agricultural soil (sandy clay loam)	120 (November to February)	Novara et al. (2016)

(2001) and Van Hemelryck et al. (2011) indicated that erosion indeed led to increased mineralization of SOC as a result of aggregate breakdown and exposure of previously encapsulated SOC during transport, but this effect was relatively minor (< 5% of eroded SOC) (Table 1). However, the rationale behind this was questioned by Lal and colleagues. They argued that erosion-induced breakdown of soil aggregates greatly promoted the mineralization of SOC and that approximately 20% of eroded SOC was oxidized during transport (Lal, 1995; Polyakov and Lal, 2008). Novara et al. (2016) even suggested that 43% of eroded SOC was mineralized during transport (Table 1). The main limitation of this issue is the difficulty in comparing different studies due to differences in measuring methods and environmental conditions. For example, field CO_2 monitoring was conducted by Novara et al. (2016) to investigate the SOC mineralization during erosion events, while laboratory rainfall simulation and incubation experiment were conducted by Van Hemelryck et al. (2010). In addition, the SOC mineralization magnitude during transport is strongly related to rainfall intensity, soil type and transport distance (Berhe and Kleber, 2013). A high CO_2 release rate may be observed in erosion events with slow and long-distance transport. In contrast, fast but short-range transport generally leads to a low degree of SOC mineralization (Kirkels et al., 2014). On the other hand, field monitoring, simulation and extrapolation of SOC mineralization during transport exist inherent defects. For SOC mineralization in-situ monitoring, the instantaneous mineralization rates probably merely represent a spike of SOC mineralization during erosion (Hu and Kuhn, 2014). Large time scales extrapolation of SOC mineralization is unreasonable. Furthermore, the upper limits of the mineral particles size used in current erosion models are often smaller than the actual aggregate sizes (Morgan et al., 1998; Van Oost et al., 2004). For instance, the largest class in Van Oost et al., 2004 is only 90 μm. However, most primary soil particles are aggregated in field, which reduces the transport distance of eroded SOC (Hu and Kuhn, 2014). The magnitude of SOC mineralization during transport may be overestimated from these aspects.

The mineralization magnitude of eroded SOC during transport has important implications for the role of soil erosion within the global C budget (Berhe et al., 2007; Billings et al., 2010; Lal, 2003). Current models exploring the magnitude of soil erosion influence on soil-atmosphere C fluxes primarily depend on detailed parameterization of relevant SOC mineralization and production pathways during and after erosion (Lugato et al., 2016). Varying these parameters may generate a wide range of C source and sink estimates. By assuming 0% and 100% mineralization of eroded SOC during transport, Billings et al. (2010) indicated that the role of soil erosion changed from a net C sink (3.1 Pg C yr^{-1}) to a net source (1.1 Pg C yr^{-1}). Jacinthe and Lal (2001) assumed 20% of the eroded SOC to be emitted during transport and indicated that soil erosion constitutes a source of atmospheric CO_2 (6–52 g C m^{-2} yr^{-1}), while in the study of Van Oost et al. (2005), 5% of the eroded SOC was assumed to be mineralized and they reported that eroded agricultural landscapes exhibit a net gain of SOC in the order of 3–10 g C m^{-2} yr^{-1}. Clearly, the controversy over the role of soil erosion within the global C cycle is closely related to the assumption on the mineralization magnitude of eroded SOC during transport. In future work, standardization of procedures to measure soil C dynamics, long-term in situ monitoring of SOC mineralization and building of global database regarding erosion-induced soil C dynamics, which would contribute to the accurate estimation of SOC mineralization during transport, are strongly recommended.

5. Depositional sites

5.1. Soil reaggregation impacts SOC mineralization and sequestration

Relative to eroding and undisturbed sites, a higher degree of soil aggregation is always observed in depositional profiles due to the enrichment in SOC and clay-sized particles (Wagner et al., 2007; Wang

et al., 2014). Sediment leaving an eroding site is a combination of primary soil particles (sand, silt and clay), free particulate organic matter and aggregated soil materials (Nie et al., 2014). Owing to limited erosive energy of interrill overland flow, the light and fine particles (e.g., clay, silt and light organic matter fraction) can be preferentially removed from eroding sites, especially in coarse textured soils (Leguédois and Bissonnais, 2004; Miller and Baharuddin, 1987; Müller-Nedebock et al., 2016). By applying simulated rainfall on silt loam and clay loam soils, Leguédois and Bissonnais (2004) investigated the sediment sorting characteristics during interrill erosion, and showed that silt and clay fractions of eroded aggregates would be preferentially transported. Mchunu et al. (2011) investigated interrill erosion induced soil and SOC losses in a small-scale agricultural community (sandy soil), and indicated that SOC enrichment ratio of eroded materials was about 4.5. Furthermore, also deposition is a selective process. Coarse particles (e.g., sand and macro-aggregates) are the first to be deposited under limited transport conditions (Walling and Moorehead, 1989). At low overland flow velocities, deposition of coarse particles starts instantly after detachment, and fine particles and light organic matter fractions are gradually enriched in overland flow (Lu et al., 1988). Therefore, the selective transport of light and fine particles during interrill erosion and the preferential deposition of the coarse fractions during the transport and delivery of sediment (especially in the low-rain intensity and long-distance transport) jointly result in the accumulation of light fraction organic matter, clay- and silt-sized particles at depositional sites (Doetterl et al., 2012; Leguédois and Bissonnais, 2004; Walling and Moorehead, 1989).

Clay particles and organic matter that serves as a binding agent, are the primary constituents of aggregates in soil (Bronick and Lal, 2005). Previous studies indicated that the formation of water-stable aggregates was promoted by large inputs of organic matter and clay (Rahman et al., 2017; Tang et al., 2009; Wagner et al., 2007). Wagner et al. (2007) investigated the influences of both clay and organic matter (barley straw) on the development of water-stable aggregates in agricultural soil and suggested that the incorporation of barley straw was most effective for the formation of water-stable aggregates in clay-rich soils (over 34% clay). The enrichment of soil organic matter and clay-sized particle at depositional sites, thus, contributes to the reaggregation of dispersed mineral particles, which in turn provides physical protection of organic matter against rapid decomposition by soil microbes (Nie et al., 2018; Wang et al., 2014).

5.2. Deep burial of SOC at depositional sites

On average, only 10–30% of eroded sediment is subsequently transported to aquatic ecosystems (e.g., rivers and oceans). Average 80% eroded sediment is deposited at the adjacent topography and stored in various depositional sites, such as colluvial and alluvial environments (Chaplot and Poesen, 2012; Stallard, 1998). The deposition of eroded sediments leads to a gradual increase in surface elevation and burial of the former topsoil at depositional sites, which disconnects the former topsoil from atmospheric conditions and decreases the turnover rate of buried SOC (Smith et al., 2005). The preservation of undecomposed wood dated thousands of years old in deep soil provides supporting evidence that burial prolongs the mean residence time of OC (Chaopricha and Marín-Spiotta, 2014). Once buried below the plow layer (30 cm depth), eroded SOC is thought to contribute to a passive C pool (Stallard, 1998; Van Oost et al., 2005). By comparing recent global estimates of modern fluxes of 12.6 Pg yr^{-1} of sediment delivered to the oceans (Syvitski et al., 2005) with estimates of modern erosion fluxes of 50–150 Pg yr^{-1} of soil (Wilkinson and McElroy, 2007), Hoffmann et al. (2013) indicated that the global C burial induced by erosion ranged from 0.5 to 1.5 Pg C yr^{-1} (assuming the OC content of eroded sediment is 1%). Stallard (1998) estimated that 0.6–1.6 Pg C is buried every year as a result of anthropogenic erosion. Erosion-induced deep burial of SOC is considered to have important implications for SOC sequestration

(Vandenbygaart et al., 2015; Van Oost et al., 2005).

The deep burial of eroded SOC indeed prolongs its mean residence time in soil, however, the long-term stability of buried SOC at depositional sites is poorly understood (Chaopricha and Marín-Spiotta, 2014; Wang et al., 2013b). The capacity of this process to constrain SOC mineralization may be reduced over time (Doetterl et al., 2016). For instance, the protection of OC within aggregates may be weakened if aggregates lose stability (Van Oost et al., 2006). Furthermore, the infiltration of DOC into deep soil can to some extent accelerate SOC mineralization due to the priming effect (Fontaine et al., 2007; Kirkels et al., 2014). By combining soil incubation and radiocarbon dating, Fontaine et al. (2007) investigated the stability of OC in deep soil layers and found that the supply of fresh plant-derived C to subsoil (0.6–0.8 m depth) stimulated the microbial mineralization of SOC with thousands of years. If buried topsoil experienced minimal change, it should have a similar C content and composition with the current topsoil. However, in reality, deep soils generally have lower C content and more stable SOC pools than comparable topsoil (Salomé et al., 2010). The alluvial soils often show a strong depth gradient of OC content, with the lowest OC content in the deepest horizons (Rumpel and Kögel-Knabner, 2010), suggesting that labile fractions of buried OC may nonetheless be gradually mineralized by soil microbes.

OC persistence in buried soil is also affected by the timescale of burial process (Chaopricha and Marín-Spiotta, 2014; Olson et al., 2016). The burial of a surface soil by eroded sediments can happen instantaneously, such as during floods and mudslides, or gradually over long periods of time, such as the loess deposition induced by interrill erosion (Feggestad et al., 2004). The type of burial process has a strong influence on the preservation of eroded SOC at depositional sites (Chaopricha and Marín-Spiotta, 2014). At high sediment deposition rate, SOC in eroded sediment and former topsoil is rapidly isolated from the atmospheric environment and buried beneath several meters of soil, thus the SOC is in a good state of preservation (Hoffmann et al., 2013). In contrast, at low deposition rate, the eroded OC is situated in the surface soil (a few centimeters to a few dozen centimeters of soil depth) of depositional profiles. This OC would therefore be prone to accelerated mineralization due to the priming effect induced by labile OC inputs (e.g., DOC) and repeated wetting-drying cycles (Lamparter et al., 2009; Wang et al., 2013a).

5.3. Changes in biological properties impact SOC mineralization and sequestration

Contrary to eroding sites, the enhancement of soil nutrient level and aggregate stability at depositional sites effectively improves the microbial habitat (Park et al., 2014). Thus, higher microbial abundance and species diversity are always observed at depositional sites relative to comparable eroding and undisturbed sites (Li et al., 2015). Further, the deposition of nutrient-rich sediments (interrill erosion) also increases the proportion of fungi in the microbial community, and leads to a higher fungal:bacterial dominance at depositional sites (Huang et al., 2013). Erosion-induced moderate variations in total microbial biomass and species diversity may have a limited impact on the mineralization and sequestration of SOC due to the so-called 'microbial functional redundancy' (Kemmitt et al., 2008), while the increase in fungal:bacterial dominance at depositional sites can contribute to the sequestration of SOC (Six et al., 2006) (discussed in Section 3.2).

Enzymes also play key roles in the biochemical functioning of soil (Makoi and Ndakidemi, 2008). Polymerized organic matters in soil must be hydrolyzed by enzymes to form small molecular compounds before they can be mineralized by soil microorganisms (Caldwell, 2005). Relative to comparable eroding and undisturbed sites, higher soil enzyme activities are always observed at depositional sites (Li et al., 2015; Park et al., 2014; Shi, 2011). This phenomenon can be explained from three different aspects. The selective nature of interrill erosion causes a preferentially transport of fine and light fractions (e.g., clay,

silt and light organic matter fraction) resulting in a fine-texture and organic matter-enriched depositional environment (discussed in Section 5.1). Enzymes exhibit a strong affinity for organic matter and mineral colloids, especially for clay (Duan et al., 2018; Shi, 2011). Therefore, the enrichment of clay-sized particle and organic matter at depositional sites contributes to the formation of enzyme-organic matter and enzyme-mineral complexes. Immobilization of enzymes on mineral colloids and organic matters provides a better protection for the enzyme protein molecules against denaturation and prolongs their lifetime (Naidja et al., 2000). Furthermore, organic matter is an important energy source for soil microorganisms. The enrichment of organic matter at depositional sites can stimulate the release of enzymes from soil microorganisms (Shi, 2011). For example, Li et al. (2015) investigated the impacts of interrill erosion on soil enzyme activities in agricultural lands and indicated that the accumulation of organic matter at depositional site significantly increased the activities of urease and catalase. Apart from organic matter content, soil moisture level at depositional sites is also increased (Li et al., 2015; Park et al., 2014). Moderate increase of soil moisture, especially in arid and semi-arid areas, can increase the connection between soil pores and thus the availability of different resources for isolated populations, which contributes to increase the activity of soil microorganisms and the production of enzymes (Guenet et al., 2012). The increases in soil enzyme activities at depositional sites can improve the microbial availability of organic matter and are believed to have negative impacts on soil C sequestration (Guenet et al., 2012; Shi, 2011).

In certain depositional sites (e.g., flood plain and reservoir), however, the surface soil can be subjected to long-term water saturation. Under such oxygen limited conditions, the mineralization of SOC is inhibited partly due to the decrease in microbial activity (aerobic microorganisms) (Trivedi et al., 2013). Furthermore, limited oxygen availability in depositional profiles negatively impacts phenol oxidase activity in soil, and consequently slows down the degradation of phenolic compounds (Freeman et al., 2001). The consequent accumulation of toxic phenolic compounds can further limit the production of hydrolytic enzymes required for organic matter degradation, therefore contributing to soil C sequestration (Shi, 2011). The accumulation of organic matter at depositional sites contributes to the sequestration of SOC when soil moisture level is sufficient to inhibit microbial and enzyme activities.

6. Conclusion and framework for future research

Soil erosion is a complex geomorphic process, which not only leads to soil loss but also has strong influences on SOC dynamics. The mineralization and sequestration of SOC at sites of erosion, along the transport pathway and at depositional sites are strongly associated with changes in biogeochemical soil properties, such as soil aggregate breakdown and formation, SOC redistribution, deep burial of eroded SOC and shifts in microbial community composition. In this review, we provide an in-depth discussion of the conceptual relations between biological soil properties, and the mineralization and sequestration of SOC in eroded agricultural landscapes, and highlight the importance of erosion-induced changes in microbial community composition on SOC dynamics.

In future work, we should pay more attention to the mineralization magnitude of eroded SOC during transport, priming effect at eroding sites induced by SOC dynamic replacement, long-term stability of buried SOC at depositional sites and impacts of changes in microbial properties (especially for autotrophic microorganisms) on the mineralization and sequestration of eroded SOC, which all have strong consequences on soil-atmosphere C exchange but so far are not clearly quantified or understood. These knowledge gaps in combination with the discrepancy in erosion landform, erosion process, spatial-temporal scale, experimental method and basic assumption have resulted in divergent interpretations and enable continuation of the sink/source

controversy over last decades by opposing researchers. Further investigation of erosion-induced SOC dynamics at broad spatial-temporal scales, long-term in-situ monitoring, application of advanced technologies (such as, C isotope tracer and remote sensing) and building of global database regarding soil erosion and associated C dynamics are highly suggested. Furthermore, consistent use of units and standardization of procedures to measure C dynamics will allow better comparison between different studies.

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