Soil aggregation and the stabilization of organic carbon as affected by erosion and deposition

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ABSTRACT

The importance of soil aggregation in determining the dynamics of soil organic carbon (SOC) during erosion, transportation and deposition is poorly understood. Particularly, we do not know how aggregation contributes to the often-observed accumulation of SOC at depositional sites. Our objective was to assess how aggregation affects SOC stabilization in comparison to interactions of SOC with minerals. We determined and compared aggregate size distributions, SOC distribution in density fractions, and lignin-derived phenols from aggregated soil samples at both eroding and depositional sites. The stabilization effect of aggregation was quantified by comparing mineralization from intact and crushed macro-aggregates. Deposition of eroded soil material resulted in carbon (C) enrichment throughout the soil profile. Both macro-aggregate associated SOC and C associated with minerals (heavy fraction) increased in their importance from the eroding to the depositional site. In the uppermost topsoil (0–5 cm), SOC mineralization from intact aggregates was larger at the depositional site than at the eroding site, reflecting the large input of labile organic matter (plant residues) promoting aggregation. Contrastingly, in the subsoil, mineralization rates were lower at the depositional site because of effective stabilization by interactions with soil minerals. Aggregate crushing increased SOC mineralization by 10–80% at the eroding site, but not at the depositional site. The content of lignin-derived phenols did not differ between eroding and depositional sites in the topsoil (24.6–30.9 mg per g C) but was larger in the subsoil of the eroding site, which was accompanied by higher lignin oxidation. Lignin data indicated minor effects of soil erosion and deposition on the composition of SOC. We conclude that SOC is better protected in aggregates at the eroding than at the depositional site. During transport disaggregation and consequently SOC mineralization took place, while at the depositional site re-aggregation occurred mainly in the form of macro-aggregates. However, this macro-aggregation did not result in a direct stabilization of SOC. We propose that the occlusion of C inside aggregates serves as a pathway for the eroded C to be later stabilized by organo-mineral interaction.

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1. Introduction

Soil erosion influences terrestrial C dynamics through various processes (e.g. Harden et al., 1999; Lal, 2003; Berhe et al., 2007; Van Oost et al., 2007). Soil erosion disturbs carbon-rich topsoils and preferentially removes soil organic carbon (SOC) from upslope sites, resulting in (partial) mineralization, as well as in (partial) redistribution and burial of SOC in depositional environments (Lal, 2003; Berhe et al., 2007). The calculation of carbon enrichment ratios (CER), i.e. the enrichment (or depletion) of carbon at depositional landforms relative to eroding sites has been used as an effective way to illustrate effects of soil erosion on carbon redistribution (Juarez et al., 2011; Wang, 2011).

Based on the mechanisms proposed by Sollins et al. (1996), von Lützow et al. (2006) and Schmidt et al. (2011), Berhe et al. (2012) summarized three key mechanisms governing SOC stabilization related to soil erosion and deposition: (i) physical isolation of SOC resulting in organic matter inaccessible to soil microorganisms and fauna (e.g. Sollins et al., 1996; Six et al., 2000, 2002); (ii) chemical interaction of organic matter with the soil mineral matrix (Torn et al., 1997) and (iii) molecular composition of organic matter influenced by environmental factors (Berhe et al., 2007; Schmidt et al., 2011).

Physical isolation of SOC, which governs its stabilization, is closely related to aggregation (Six et al., 2000). Soil aggregation provides physical protection of SOC against rapid decomposition by

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soil microbes (Razafimbelo et al., 2008), and aggregate formation appears to be closely linked with soil C storage and stability (Golchin et al., 1995; Barreto et al., 2009; Salomé et al., 2010; Thevenot et al., 2010). Generally, aggregation is perceived as an indicator of soil stability and erodibility (Bryan, 1971; Morgan, 2005). However, the role of soil aggregation and disaggregation in SOC dynamics during erosion, transportation and deposition is far less well understood (e.g. Kuhn et al., 2009; Nadeu et al., 2011).

Most process level investigations of SOC cycling focus on non-sloping sites that experience minimal soil erosion and deposition (Sollins et al., 1996; Six et al., 2002; Schmidt et al., 2011; Berhe et al., 2012). Soil erosion can promote breakdown of aggregates at the eroding landform positions leading to exposure of previously encapsulated SOC (Lal, 2003; Doetterl et al., 2012). This exposure of formerly physically protected SOC increased the potential availability of SOC to mineralization (Berhe et al., 2012) and release of soluble compounds. Goebel et al. (2009) found smaller SOC mineralization from intact aggregates in comparison to crushed ones demonstrating the stabilization effect of aggregation.

Interactions of SOC with mineral surfaces are considered as the most effective stabilization mechanisms in many soils (Torn et al., 1997; von Lützow et al., 2006; Schmidt et al., 2011). Stabilization occurs through sorption to pedogenic oxides, clay minerals or by co-precipitation with polyvalent cations. Effects of erosion and deposition on this chemical stabilization of SOC are not well understood. Berhe et al. (2012; pp. 13) concluded that "the potential to form new sorptive mineral-organic matter interactions is not always higher in depositional, compared to eroding, landform positions". Stabilizing effects of organo-mineral interactions might be even more important than previously assumed because of the role of pedogenic oxides in aggregate formation and stabilization (Berhe et al., 2012), i.e. interactions between chemical and physical stabilization mechanisms.

As one of the major components of plant biomass and SOC, lignin has been used as an indicator of the fate of plant-derived organic matter and the origin and state of degradation of soil organic matter (e.g. Hedges et al., 1988; Guggenberger and Zech, 1994; Klotzbücher et al., 2011). Variations in the relative proportions of lignin-derived phenols provide information about microbial degradation and adsorption and desorption processes (Hernes et al., 2007; Thevenot et al., 2010). These compositional changes of SOC affect stabilization and sequestration of SOC in soils (Schmidt et al., 2011). Soil erosion potentially affects lignin probably because it preferentially removes organic matter not associated with minerals or aggregates, which is easily transported (Gregorich et al., 1998; Lal, 2003; Berhe et al., 2012). Therefore, the assessment of amount and composition of lignin might improve our understanding of how erosion influences the origin, composition and fate of SOC from upslope to downslope positions.

Physical fractionation by density has been proven useful to identify meaningful soil fractions, which can be related to different stability and stabilization processes (von Lützow et al., 2008). The free light fraction (fLF) comprises relatively undecomposed labile organic matter, i.e. unprotected from physical or chemical processes. The occluded light fraction (oLF) comprises organic matter stabilized by aggregation, whereas in the heavy fraction (HF) C is strongly associated with soil minerals (Golchin et al., 1994; Cerli et al., 2012). Therefore, density fractionation of SOC is a useful tool to study soil organic matter stabilization in the context of soil erosion and deposition (e.g. Gregorich et al., 1998; Berhe et al., 2012; Doetterl et al., 2012).

To obtain a better understanding of SOC stabilization at sites that experience soil erosion and deposition we wanted to know how aggregation and disaggregation affect SOC stabilization in comparison to interactions of SOC with minerals. Particularly, we wanted to reveal the unknown role of aggregation for the often-observed SOC accumulation at depositional sites. Additionally, we used lignin to trace the fate of organic matter during erosion and deposition. We therefore studied topsoils and subsoils sampled from eroding and depositional sites, carried out aggregate and density fractionation and studied amounts, composition (lignin) and stability of SOC.

2. Material and methods

2.1. Study site and sampling

Soil samples were collected from a terraced hillslope in the Belgian Loess Belt near Leuven, Belgium, with an upslope eroding position (Haplic Cambisol) and a downslope depositional location (Colluvic Regosol), referred in the text as the eroding and depositional sites (Wang et al., 2013). The soils had a similar soil texture (silt loam according to FAO, 2006). The deposition rate in this area was estimated as 0.73 cm/year with ^{137}Cs tracer techniques (Wang, 2011). Mean annual temperature is 9–10 °C and annual precipitation is 750–800 mm in this region. In this study, five soil cores at 0–200 cm depths were taken at each site. These soil cores were firstly separated by depth to determine C contents. Then the five cores were mixed per depth and site for further analyses. Parts of the samples were air-dried and the other part was stored under field moist conditions at 4 °C before analysis. More detailed information on the experimental area and the sampling methods was reported in Wang et al. (2013).

2.2. Soil analysis

The C and N contents of bulk soils, aggregate size fractions and density fractions were determined using a C and N analyzer (Elementar Vario EL, Hanau, Germany; Standard deviation (SD) <0.1% abs.). The samples were not pretreated with acid prior to C analysis. It was found that the soil was free of inorganic C because there was no reaction when 6 M HCl was added to the soil, hence total C was considered to be equivalent to SOC.

2.3. Isolation of soil fractions

2.3.1. Aggregate size distribution
Topsoil (0–5 and 5–10 cm) and subsoil (45–70 and 160–200 cm) samples from the eroding and depositional sites were subjected to aggregate size fractionation according to the dry sieving method (Cammeraat and Imeson, 1998). Briefly, 170–200 g of soil samples were air-dried at room temperature (20 °C) and then fractionated using nine mesh sieves (16, 8, 5, 4, 2, 1, 0.5, 0.25 and 0.125 mm) to obtain ten size fractions. Sieves were shaken for 5 min at 30 Hz using a horizontal shaking machine (Retsch, AS 200 basic, Haan, Germany). Each fraction was weighed to determine the proportion of aggregates in each size fraction. For each fraction, the C and N content were determined. Macro-aggregates were defined as aggregates being >2 mm. The average recovery of the total soil mass and C was 99.4 ± 3.7% (SD) and 127 ± 22%, respectively. We used smaller depth increments for the topsoil than the subsoil samples because of the exponential decrease in organic C with depth, i.e. the large decrease in organic C particularly in the topsoil (Fig. 2 in Wang et al., 2013).

2.3.2. Soil density fractionation
Density fractionation was applied to the topsoil (0–5 and 5–10 cm) and subsoil (45–70 and 160–200 cm) samples from eroding and depositional sites following the methods of Golchin et al. (1994) and Cerli et al. (2012). Twenty-five grams of air-dried