Impacts of soil redistribution on the transport and fate of organic carbon in loess soils

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Chapter 2

Stability of organic matter in soils of the Belgium Loess Belt upon erosion and deposition


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Chapter 2 Stability of organic matter upon erosion and deposition

Abstract

Soil erosion has significant impacts on terrestrial carbon (C) dynamics. It removes C rich topsoil and deposits it in lower areas, which might result in its stabilization against microbial decay. Subsequently, C poor deeper horizons will be exposed, which also affects C stabilization. We analyzed factors governing soil organic C (SOC) mineralization in topsoil (5–10 cm) and subsoil (75–100 and 160–200 cm) horizons from two contrasting sites (up-slope compared with down-slope) in the Belgian Loess Belt; we refer to these as eroding and depositional sites, respectively. Deposition of eroded soil material resulted in significantly increased SOC contents throughout the entire soil profile (2 m) and microbial biomass C in the topsoil. In a 28-day incubation experiment we studied effects of O$_2$ concentrations (0, 5 and 20%) and substrate (glucose) availability on C mineralization, soil microbial biomass and CaCl$_2$-extractable C. Carbon enrichment at the depositional site was accompanied by weak mineralization rates and small contents of water-extractable organic C. Addition of glucose stimulated microbial growth and enhanced respiration, particularly in the subsoil of the depositional site. Availability of O$_2$ showed the expected positive relationship with C mineralization in topsoils only. However, small O$_2$ concentrations did not decrease C mineralization in subsoils, indicating that controls on C dynamics were different in top- and subsoils. We conclude that reduced C mineralization contributed to C accumulation as observed at depositional sites, probably because of poor availability of C in subsoil horizons. Limited availability of O$_2$ in subsoils can be excluded as an important control of soil C accumulation. We hypothesize that the composition of the microbial community after burial of the organic-rich material might play a decisive role.
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2.1 Introduction

The biogeochemical cycling of carbon (C) in terrestrial ecosystems has received increasing attention worldwide, as CO₂ emission into the atmosphere plays a vital role in driving global warming. Most of the organic C in terrestrial ecosystems is stored in mineral soil horizons. Soil erosion and deposition can strongly influence the net C budget in a watershed firstly by removing C from the topsoil and depositing it at lower areas, and secondly by exposing C-poor subsoils and parent material to surface conditions at sites of erosion (Berhe et al., 2007; Harden et al., 1999; Quinton et al., 2010; van Oost et al., 2007). Burial of eroded carbon at depositional areas might result in less favourable conditions for C mineralization (lower temperature and O₂ availability), which might explain the long-term storage of the deposited C (Berhe et al., 2012; Lal, 2003). Exposure of subsoils to surface conditions could lead to increased sequestration of new photosynthates because of greater availability of non-C saturated reactive mineral surfaces capable of sorbing and stabilizing organic C (Berhe et al., 2007; van Oost et al., 2007). However, the exposure could also result in the opposite effect by providing better conditions for C mineralization (increased temperature and O₂ availability, drying and rewetting cycles), which might result in C losses (Davidson et al., 1998; Fierer et al., 2003).

For these reasons Lal (2003) assumed that soil erosion could increase CO₂ emissions. Harden et al. (1999) showed that in very eroded Mississippi loess soils, the net C balance and the extent of the CO₂ sink at the watershed scale are controlled by the rate at which eroded C is decomposed and replaced by new photosynthates. Berhe et al. (2007) reported that soil erosion and deposition probably results in stabilization of at least 0.72 Pg C per year globally. Thus, one of our current challenges is to identify the various mechanisms driving C mineralization in eroding and depositional areas and quantify their potential for C accumulation and stabilization (Berhe et al., 2012; Harden et al., 1999).

Mineralization of soil organic carbon (SOC) strongly depends on the size of the labile C pool, the activity and composition of the microbial community and abiotic environmental conditions such as temperature, moisture and oxygen availability (Davidson et al., 1998; Marschner and Kalbitz, 2003; Salome et al., 2010). The size of the labile pool is particularly important for short-term reactions and it is mainly composed of carbohydrates and N-rich compounds (Marschner and Kalbitz, 2003). Kemmitt et al. (2008) reported that SOC mineralization is independent of the microbial biomass size, structure or activity. They suggested that the rate-limiting step is governed by abiotic processes converting non-bioavailable SOC into bioavailable forms SOC such as dissolved organic C (DOC). Fontaine et al. (2007) pointed out that SOC in subsoil does not provide enough energy to sustain active microbial populations and the necessary production of enzymes, a phenomenon commonly referred to as the priming effect. This effect describes the extra mineralization of in situ C
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induced by addition of readily available C.

Dissolved organic C serves as a substrate for microbial respiration (Marschner and Kalbitz, 2003). Therefore, C mineralization is often related to DOC contents (Zhao et al., 2008). Additionally, DOC is a product of physical, chemical and biological breakdown of soil organic matter, which leads to a complicated relationship between DOC and microbial respiration (Park et al., 2002). A linear relationship between C mineralization and DOC contents could indicate poor C availability if DOC contents are small (Klotzbücher et al., 2011).

In the present study we investigated how C mineralizations at different depths in eroding and depositional sites are affected by the availability of oxygen and carbon. In detail, we hypothesized that there would be (i) smaller C mineralization at depositional sites than at eroding sites, (ii) decreasing mineralization with increasing soil depth for both eroding and depositional sites and (iii) decreasing availability of O2, which will reduce C mineralization to a larger extent in topsoils than in subsoils because of a microbial community adapted to such conditions in subsoils, and (iv) the addition of readily available substrates will increase C mineralization particularly in subsoils and this effect will be larger at sites with smaller C mineralization, such as the depositional site.

To test these hypotheses we incubated topsoils and subsoils from an eroding and a depositional site in the Belgian Loess Belt under different O2 concentrations and with addition of glucose and then determined C mineralization, amounts and composition of DOC and microbial biomass. Comparing SOC mineralization of soils from different depths from the two sites will also indicate whether larger SOC contents at depositional sites have been caused by the addition of material with more C or by the decreased mineralization after deposition. Although the composition of SOC at the eroding site is not affected by transport and deposition, the effect of depth-related environmental conditions on SOC can best be studied by comparing mineralization at the same depths of both sites.

2.2 Materials and methods

Soil sampling and sample preparation

Soil samples were collected from a terraced hill-slope in the Belgian Loess Belt near Leuven, Belgium, from two contrasting sites. One site was located in an upslope position (50°48′25″N, 4°35′07″E, 82 m above sea level) and the other in a downslope (50°48′25″N, 4°35′06″E, 79 m above sea level) depositional location (Figure 2.1). The upslope position was located on a gently sloping side of a hill spur, where severe erosion had previously taken place, as evidenced by the shallow underlying tertiary material (present within 1.70 m depth). The
down-slope location was situated in a shallow dry valley, directly connected to the up-slope location and bounded by a terrace downslope. The sediment deposition rate in this area was estimated to be 7.3 mm year$^{-1}$ using $^{137}$Cs as a tracer (Wang, 2011). Deposition of eroded upslope topsoil material has led to the build-up over time of more than 2 m of colluvium, which consists of transported loess material from the surrounding hillslopes. In the following text, we refer to the upslope site as the eroding site and the downslope site as the depositional site. Mean annual temperature is 9–10°C and annual precipitation is 800 mm.

The sample locations had soils that were classified as a Haplic Cambisol at the eroding site and a Colluvic Regosol at the deposition site (WRB, 2006). The main crops in this study area are wheat, maize ($\textit{Zea Mays}$ L.), sugar beet ($\textit{Beta vulgaris}$ L.), potato ($\textit{Solanum tuberosum}$ L.) and chicory ($\textit{Cichorium intybus}$ L.) (Evrard et al., 2007).

Sampling was carried out in March 2010 by collecting undisturbed in-line cores (diameter, 4.5 cm) in triplicate to a depth of 2 m at both sites. Mineral soil samples were obtained in the field at 5 cm increments for the 0–50 cm depth, at 10 cm increments over 50 cm to 1 m depth, and at 20 cm increments between 1 and 2 m depth. Soil samples were transported to the laboratory in plastic tubes and stored at 4°C until analysed. A part of the sample material was dried at 40°C and then sieved to <2 mm for further analysis.

Samples from all the depth increments were used to determine the contents of organic C and total nitrogen (N) in the soil profiles. For all the other analyses, we only used soils from three depths representing the whole range of differences in C content between the eroding and depositional sites, also in triplicate. These depths were near-surface soil (5–10 cm) with very
few differences, 45–70 cm deep soil, which had the maximum difference, and soil from 160 to 200 cm depth with intermediate difference in total organic C content (Table 2.1, Figure 2.2).

Field water content was determined by drying soil subsamples at 105°C for 24 hours. Soil pH was measured in water at a 1:5 soil: water ratio. Total SOC and total nitrogen (TN) contents were determined with a C/N analyser (Elementar Vario EL, Hanau, Germany). No carbonates were found in the soil samples as tested by addition of 2 M HCl. Total content of pedogenic (hydr-) oxides was estimated as dithionite-citrate-bicarbonate extractable iron (Fe$_{d}$) using the method of (Le Mer and Roger, 1960) and atomic absorption spectroscopy (AAS, Perkin Elmer, Waltham Massachusetts, USA). Active iron and aluminum (hydr-) oxide (Fe$_{o}$ and Al$_{o}$) were extracted with 0.2 M ammonium oxalate at pH 3 and quantified by AAS.

Dissolved organic carbon was obtained by aqueous extraction of field moist soils with CaCl$_{2}$ as a surrogate for soil solution (Chantigny, 2007). This method has been used to demonstrate close relationships between DOC and C mineralization in agricultural loess soils (Zhao et al., 2008), although for forest soils, Rennert et al. (2007) recommended the use of K$_{2}$SO$_{4}$. We added 150 ml of 5 mmol l$^{-1}$ CaCl$_{2}$ to 30 g soil (oven-dried basis) and the vessels were shaken for 30 minutes.

After allowing larger particles to settle for 1 hour, the supernatants were filtered through pre-rinsed 0.45 μm cellulose-acetate filters (Zhao et al., 2008). Dissolved organic C in solution was determined by a TOC analyser (TOC-VC$^*_{CH},$ Shimadzu, Kyoto, Japan). Specific UV absorbance at 280 nm (UV280) was measured immediately using a Spectroquant Pharo 300 spectrometer (Merck KGaA, Darmstadt, Germany).

Soil incubation

In order to avoid disturbance effects, we did not sieve the samples but used subsamples directly from the core, which were mainly intact soil aggregates in an estimated size range of 2–16 mm. For each of the three selected depths (5–10, 45–70 and 160–200 cm) at each of the sites, we incubated 20 g moist soil (oven-dried basis) in 120-ml flasks at 20°C in the dark for 28 days in two sets of the three replicates. The water content of soil samples was adjusted to a water potential of −100 mbar (pF 2.0) prior to incubation with a pF tray. The flasks were sealed with rubber septa. The incubations were performed at three different oxygen levels, control (20% oxygen, air in the laboratory), no oxygen (N$_{2}$ atmosphere) and low oxygen (5% O$_{2}$ in a N$_{2}$ atmosphere), and with and without addition of readily available C, supplied as glucose. Following Hartley et al. (2010), 15 mg glucose-C per g soil C (1 ml solution) was added to half of the incubation flasks after the first 14 days of pre-incubation. The same amount of distilled water was added to the remaining flasks. The increase in SOC
mineralization by glucose addition can be used as a first indication of the availability of C in the soils (Hartley et al., 2010; Salome et al., 2010). However, we were not able to assess priming effects because we did not use labelled glucose.

We exchanged the headspace of the flasks after 1 and 3 weeks to maintain the desired level of O₂. The concentration of 5% O₂ was achieved by using 5% O₂ in N₂ (5.0 purity, 99.99% N₂, Praxair, Vlaardingen, the Netherlands). The headspace of the incubation flasks was sampled on days 1, 3, 7, 10, 14, 21 and 28 of the incubation period. Concentrations of CO₂ and CH₄ were determined by gas chromatography (Varian STAR 3600, Palo Alto, California, USA) after manual injection of 20 μl gas. Air pressure in the headspace was measured by a tensiometer (TC 1085, Tensio Technik, Geisenheim, Germany). Carbon mineralization was calculated with the general gas equation and related to the C content of the sample. After 28 days of incubation, soil samples were frozen and stored at −18°C.

Soil microbial biomass C (SMBC) was determined before and at the end of the incubation experiment, by the chloroform fumigation-extraction method (Chevallier et al., 2010; Joergensen, 1996). We used 0.05 m K₂SO₄ (Chevallier et al., 2010) after testing for the absence of significant differences between 0.5 and 0.05 m K₂SO₄ based extractions. Fumigated and non-fumigated soil samples (stored frozen) were subsequently extracted with 0.05 m K₂SO₄ (10 g in 50 ml) by shaking for 0.5 hours (170 rpm) at room temperature, followed by centrifugation (1300 g over 10 minutes) and filtration. Dissolved organic C and total N concentrations of the soil extracts were determined by a TOC analyser (TOC-VCPH, Shimadzu). Soil microbial biomass C was calculated by dividing the difference between the extracted OC in fumigated and non-fumigated soil samples by a conversion factor of 0.45, which is defined as the extractable part of the microbial biomass after fumigation (Joergensen, 1996).

Statistical analysis

The statistical significance of differences between the SOC mineralization rates of the two sites and three soil depths were tested by one-way (cumulative C mineralized during the incubation period) and repeated measurements (time) ANOVA with PASW Statistics 17.0. We used a one-way ANOVA to test the effects of oxygen availability (0, 5 and 20% O₂) and substrate (glucose) addition on SOC mineralization. Differences between the three depths, 5–10, 45–70 and 160–200 cm, from eroding and depositional sites were tested with one-way ANOVA. In all cases, we considered differences as statistically significant at $P < 0.05$. All results are expressed as the mean of three replicates.
Table 2.1 Properties of the experimental sites used for the incubation experiment (mean ± SD, n = 3)

<table>
<thead>
<tr>
<th>Property</th>
<th>Eroding site 5–10</th>
<th>Depositional site 5–10</th>
<th>Eroding site 45–70</th>
<th>Depositional site 45–70</th>
<th>Eroding site 160–200</th>
<th>Depositional site 160–200</th>
</tr>
</thead>
<tbody>
<tr>
<td>Soil pH</td>
<td>5.5 (0.06)</td>
<td>7.0 (0.05)</td>
<td>6.6 (0.06)</td>
<td>6.7 (0.05)</td>
<td>6.5 (0.02)</td>
<td>7.3 (0.09)</td>
</tr>
<tr>
<td>Soil organic carbon/ g kg⁻¹</td>
<td>8.2 (1.01) bA</td>
<td>10.8 (0.73) aA</td>
<td>1.8 (0.13) bB</td>
<td>5.8 (0.51) aB</td>
<td>0.8 (0.24) bC</td>
<td>2.8 (0.71) aC</td>
</tr>
<tr>
<td>Total nitrogen/ g kg⁻¹</td>
<td>0.9 (0.17) aA</td>
<td>1.0 (0.08) aA</td>
<td>0.3 (0.04) bB</td>
<td>0.7 (0.05) aB</td>
<td>0.1 (0.02) bC</td>
<td>0.2 (0.02) aC</td>
</tr>
<tr>
<td>C/N</td>
<td>9.1</td>
<td>10.8</td>
<td>6.0</td>
<td>8.3</td>
<td>8.0</td>
<td>14.0</td>
</tr>
<tr>
<td>Fe_d/ g kg⁻¹</td>
<td>8.5 (0.19)</td>
<td>12.5 (0.27)</td>
<td>7.8 (0.16)</td>
<td>11.5 (0.22)</td>
<td>15.7 (0.41)</td>
<td>8.9 (0.14)</td>
</tr>
<tr>
<td>Fe_o/ g kg⁻¹</td>
<td>2.2 (0.23)</td>
<td>1.9 (0.06)</td>
<td>2.5 (0.24)</td>
<td>2.1 (0.06)</td>
<td>0.6 (0.04)</td>
<td>1.6 (0.12)</td>
</tr>
<tr>
<td>Al_o/ g kg⁻¹</td>
<td>0.5 (0.03)</td>
<td>0.6 (0.02)</td>
<td>0.5 (0.05)</td>
<td>0.6 (0.02)</td>
<td>0.5 (0.01)</td>
<td>0.6 (0.03)</td>
</tr>
<tr>
<td>Clay / %</td>
<td>10.0</td>
<td>11.2</td>
<td>11.2</td>
<td>12.0</td>
<td>15.6</td>
<td>11.7</td>
</tr>
<tr>
<td>Silt / %</td>
<td>75.1</td>
<td>69.4</td>
<td>80.9</td>
<td>72.9</td>
<td>79.4</td>
<td>78.4</td>
</tr>
<tr>
<td>Sand / %</td>
<td>14.9</td>
<td>19.5</td>
<td>7.9</td>
<td>15.1</td>
<td>5.1</td>
<td>10.0</td>
</tr>
</tbody>
</table>

* Fe_d: dithionite-extractable Fe

* Fe_o and Al_o: oxalate-extractable Fe and Al
2.3 Results

**Carbon contents in different soil profiles**

Soil organic C contents were significantly larger at the depositional site than at the eroding site throughout the entire soil profile (Figure 2.2). The differences between the two sites were smaller in the topsoil and increased with soil depth, resulting in three-fold larger SOC contents at the depositional site at 40 cm. At each depth, the C/N ratios of the depositional site were larger than those of the eroding site (Table 2.1). The value was greatest in the deepest soil at the depositional site, while the converse was observed at the eroding site, with the smallest C/N ratio in the deeper subsoil.

![Graph showing soil organic carbon content in different soil profiles](image)

Figure 2.2 Soil organic carbon content (g kg\(^{-1}\)) in the soil profiles of the eroding and depositional site (mean and standard deviation of four replicates).

**Soil organic carbon mineralization**

Cumulative C mineralization, as measured by incubation of the soil under different oxygen treatments, varied between sites and over depths (Figure 2.3). Cumulative C mineralization was significantly less at the depositional site than at the eroding site, except for the topsoil at 5% oxygen. Differences in C mineralization between the two sites increased with soil depth and this difference became evident after 3 days of incubation (Figure 2.3). In soils from both locations, C mineralization decreased significantly with increasing soil depth under aerobic conditions. This trend was less evident at 5 and 0% O\(_2\) in the headspace. Under these conditions C mineralization decreased from the topsoil to 45–70 cm and then increased again.
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under 0% oxygen in 160–200 cm samples at both sites (Figure 2.3). Effects of O₂ availability on C mineralization depended on soil depth. In the topsoil cumulative C mineralization was smaller with 0% O₂ atmosphere than with 5 and 20% O₂, regardless of the site (Figure 2.3). In the upper and deep subsoil, O₂ availability did not significantly affect C mineralization.

Figure 2.3 Accumulated CO₂-C mg g⁻¹ SOC during 28 days of incubation as affected by soil depth (in rows) and oxygen availability (in column) at the depositional and eroding sites (mean and standard error of three replicates).
Addition of glucose resulted in increased C mineralization in all soils and at all depths (Figure 2.4). Cumulative C mineralization was 1.2–3.4 times larger than that without glucose addition. Interestingly, the relative increase in C mineralization after addition of glucose was larger in the subsoil (both depths) from the depositional site than that from the eroding site independently of oxygen availability.

Figure 2.4 Cumulative CO$_2$-C emissions with and without glucose added to soil from three different depths at the depositional and eroding sites (mean and standard error of three replicates).
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**CH₄ emissions from different soil horizons**

Methane emissions from the soils were small and increased from topsoils (0.32 μg CH₄-C per g soil C per day) to subsoils (3.95 μg CH₄-C per g soil C per day) at both sites (Figure 2.5). As with CO₂, CH₄ emissions were significantly larger at the eroding than at the depositional site. The availability of oxygen did not affect greatly the CH₄ emissions from soil. Addition of glucose resulted in a similar pattern for CH₄ to that for CO₂, with a larger relative increase in CH₄ emission at the depositional site than at the eroding site for all depths.

![Graphs showing CH₄ emissions](image)

Figure 2.5 Accumulated CH₄-C emissions (μg C g⁻¹ SOC) from the top- and subsoil at the eroding and depositional sites during 28 days of incubation (mean and standard error of three replicates).
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**Soil microbial biomass C**

Before incubation, soil microbial biomass C was significantly larger at the depositional site than at the eroding site and decreased with soil depth (Table 2.2). Considering the large C mineralization at the eroding site, the metabolic quotients (respiration per unit microbial biomass) were much larger at the eroding than at the depositional site. In some subsoil samples, the content of microbial biomass C was below the detection limit. During incubation, microbial biomass C mostly remained constant or decreased, except for the upper subsoil of the depositional site, where it increased (with addition of glucose and 5% O₂). We also found an exceptional increase in microbial biomass C in the subsoil (160–200 cm) of the eroding site under 0% oxygen.

**Dissolved organic carbon and specific UV absorbance**

DOC contents (mg C per g soil C) increased with increasing soil depth at both sites and were larger at the eroding than at the depositional site (not significant for the topsoil, Figure 2.6). The difference between the two sites was largest in the deep mineral subsoil, where DOC contents were more than four times larger at the eroding site. The specific UV absorbance, used as a measure of the aromatic character of DOC, decreased with increasing soil depth and was smaller at the eroding than at the depositional site. However, differences were not statistically significant, except for the 160–200 cm depth at both sites.
Figure 2.6 Dissolved organic carbon (DOC) (mg C g\(^{-1}\) soil C\(^{-1}\)) and specific UV absorbance at 280 nm (UV280) for different soil depths at the depositional and eroding sites (mean and standard error of three replicates)
**Table 2.2** Soil microbial biomass C (mg C g\(^{-1}\) soil C\(^{-1}\)) before and after 28 days incubation (mean ± SD, n = 3).

<table>
<thead>
<tr>
<th>Locations</th>
<th>Depth/cm</th>
<th>0%</th>
<th>5%</th>
<th>20%</th>
<th>0%</th>
<th>5%</th>
<th>20%</th>
<th>0%</th>
<th>5%</th>
<th>20%</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>-Glucose</td>
<td>+Glucose</td>
<td></td>
<td>-Glucose</td>
<td>+Glucose</td>
<td></td>
<td>-Glucose</td>
<td>+Glucose</td>
<td></td>
</tr>
<tr>
<td>Deposional</td>
<td>5–10</td>
<td>21.7 (0.90)</td>
<td>19.3 (0.46)</td>
<td>19.4 (0.73)</td>
<td>31.6 (1.10)</td>
<td>24.4 (4.47)</td>
<td>25.0 (1.45)</td>
<td>0.10</td>
<td>0.24</td>
<td>0.30</td>
</tr>
<tr>
<td>site</td>
<td>45–70</td>
<td>7.4 (2.33)</td>
<td>5.9 (2.06)</td>
<td>4.3 (2.99)</td>
<td>8.6 (2.33)</td>
<td>9.9 (2.65)</td>
<td>9.0 (0.70)</td>
<td>0.15</td>
<td>0.22</td>
<td>0.23</td>
</tr>
<tr>
<td></td>
<td>160–200</td>
<td>2.3 (1.84)</td>
<td>4.1 (0.44)</td>
<td>ND</td>
<td>5.4 (1.93)</td>
<td>5.1 (3.05)</td>
<td>ND</td>
<td>3.55</td>
<td>3.70</td>
<td>2.00</td>
</tr>
<tr>
<td>Eroding</td>
<td>5–10</td>
<td>16.9 (1.50)</td>
<td>14.7 (2.85)</td>
<td>14.7 (0.51)</td>
<td>22.1 (1.71)</td>
<td>15.4 (1.74)</td>
<td>11.8 (1.48)</td>
<td>0.19</td>
<td>0.35</td>
<td>0.45</td>
</tr>
<tr>
<td>site</td>
<td>45–70</td>
<td>ND</td>
<td>4.5 (1.45)</td>
<td>11.2 (0.00)</td>
<td>3.5 (3.48)</td>
<td>ND</td>
<td>6.0 (0.00)</td>
<td>ND</td>
<td>ND</td>
<td>ND</td>
</tr>
<tr>
<td></td>
<td>160–200</td>
<td>20.63(12.07)</td>
<td>5.06(4.48)</td>
<td>ND</td>
<td>ND</td>
<td>ND</td>
<td>ND</td>
<td>ND</td>
<td>ND</td>
<td></td>
</tr>
</tbody>
</table>

*ND: not detectable

*-Glucose and +Glucose mean treatments without glucose addition and with glucose addition, respectively
2.4 Discussion

Decreasing C mineralization after deposition: a reason for C accumulation?

We confirmed our hypothesis that C mineralization was smaller at the depositional site than at the eroding site. Therefore, the large C enrichment of the soil at the depositional site probably results mainly from decreased C mineralization after deposition. Deposition of eroded C is often accompanied by increased water contents and physical protection of intra- or interaggregate organic C because of the rearrangement of transported aggregates during deposition. These conditions can, collectively, retard the decomposition rate of SOC (von Lützow et al., 2006). Furthermore, burial facilitates chemical transformations and reactions with the mineral phase, which both contribute to C stabilization as well (Berhe et al., 2007). However, differences in pedogenic oxides between the two sites are small and not consistent (Table 2.1). Therefore, they do not explain the larger SOC stabilization at the depositional site than at the eroding site.

Carbon can also be sequestered at eroding sites. Liu et al. (2003) demonstrated that erosion decreased CO$_2$ emission by stabilization of plant-derived organic matter at mineral surfaces. Subsurface material exposed to surface conditions has little SOC and therefore a larger proportion of mineral surfaces are available for organic matter sorption. The comparison of C mineralization in the subsoils of the two sites clearly showed that a large availability of mineral surfaces at the eroding site (smaller C contents and similar contents of pedogenic oxides) did not result in decreased C mineralization. Thus, decreased C mineralization after deposition should be responsible for SOC accumulation. Another explanation could relate to the preferential erosion and deposition of more stable organic matter (Berhe et al., 2012; Lal, 2003). However, preferential erosion of the relatively labile particulate organic matter fraction (Cerli et al., 2012) might be more likely (Kuhn et al., 2009).

The small differences in organic C contents of the topsoils of the two sites (2 mg C g$^{-1}$ soil) were accompanied by similarly small differences in their C mineralization and could potentially suggest a common origin of the C material in the topsoils of the two sites. Deeper in the profile (30–80 cm), the difference in organic C between the eroding and depositional site was largest (about 5 mg C g$^{-1}$ soil). This could result from the permanent smaller C mineralization at the depositional site over time, accompanied by the constant deposition of C-rich material. In the deep subsoil (>80 cm), the difference in organic C between the eroding and depositional sites was stable (about 2 mg C g$^{-1}$ soil), which might be because of long periods of small but different C mineralization rates at the two sites. Transport and deposition processes could also contribute indirectly to the decreased mineralization in the depositional environment (Lal, 2003; Berhe et al., 2007). There are indications of preferential erosion of labile organic material in comparison to protected and more stable organic matter in
aggregates (Kuhn et al., 2009). This labile fraction can be decomposed quickly during transport (Lal, 2003), leading to a relative enrichment of more stable organic matter in the deposited material.

Differences in DOC between the two sites resembled those of C mineralization, with larger contents at the eroding than at the depositional site (Figure 2.6). Therefore, we found further evidence for the hypothesis that DOC is a readily biodegradable fraction of soil organic matter and can be used to predict C mineralization from soils. Although not statistically significant, smaller UV absorbance of DOC from the eroding site than from the depositional site supports the concept of greater biodegradability of organic matter from the eroding sites because aromatic constituents of DOC are particularly stable against biodegradation (Kalbitz et al., 2003a).

**C mineralization in topsoil and subsoils**

There were clear differences in regulation of C cycling in the top- and subsoil. We did not find different depth patterns of SOC mineralization between the eroding and the depositional site. In our experiment, the decomposability of SOC decreased with depth, which was also reported in other studies (Fierer et al., 2003). In contrast, Salome et al. (2010) suggested that organic C in the subsoil was readily decomposable. The differences between the results obtained here (Figure 2.3) and those of Salome et al. (2010) might result from the rooting depths of the vegetation cover, SOC composition and soil structure, resulting in non-comparable types of organic matter inputs with depth.

Although C mineralization decreased with increasing depth, the initial DOC contents (per g SOC) increased in both soils (Figure 2.6). This degradability decrease was even more pronounced at the eroding site, as evidenced by a larger DOC content increase with depth. Specific UV absorbance, used as a measure of recalcitrant aromatic compounds, decreased with increasing soil depths and this seemed to contradict the reduced DOC degradability observed at both the sites. However, the degradability of DOC from different sources cannot be simply predicted on the base of its chemical composition (Schwesig et al., 2003). A small UV absorbance also indicates a large contribution of microbial-derived compounds to DOC, which also have a large stability (Kalbitz et al., 2003a). In subsoils, the contribution of microbial-derived compounds is usually larger than in topsoils, which might explain the small C mineralization rates with large DOC contents and a low specific UV absorbance.
Chapter 2 Stability of organic matter upon erosion and deposition

Effects of substrate addition and O₂ availability on C mineralization

Addition of readily available organic matter (glucose) did stimulate C mineralization, particularly in the subsoil of the depositional site, where DOC contents were small and characterized by greater UV absorbance. That meant that limited C availability restricted C mineralization in the subsoil at the depositional site. Enhanced mineralization after addition of glucose might reflect shifts in the composition of the microbial community towards r-strategists, which are adapted to a continuous input of readily available organic matter, as occurs at the depositional site (Paterson et al., 2009; Salome et al., 2010).

In our study, there was no significant difference between topsoil cumulative SOC mineralization at 5 and 20% O₂. However, topsoil C mineralization at both sites was significantly more than that under anaerobic conditions (0% O₂, Figure 2.3). Obviously, the 5% treatment was sufficient to avoid it becoming O₂-limiting, at least for the duration of the incubation (Salome et al., 2010). Gudasz et al. (2010) suggested that SOC burial efficiency in lake sediments was related to oxygen exposure. Freeman et al. (2001) demonstrated that in waterlogged soils the limitation of oxygen prevented the action of polyphenol oxidase, which acted as a ‘latch’ on the mineralization of organic matter. When the limitation was removed, rates of organic matter mineralization increased (Kemmitt et al., 2008).

The most interesting finding of our study is that C mineralization was not significantly different under 20% O₂ from that at 0 or 5% O₂ environments in the two subsoil horizons of both sites (Figure 2.3). This is in contrast to many other studies, which suggested that C mineralization slowed under anaerobic conditions (Freeman et al., 2001). Obviously, mineralization under anaerobic conditions can be as large as under aerobic conditions in subsoils, pointing to the presence of a highly adapted microbial community. At our site, pedogenic Fe oxides are available for reduction of Fe³⁺ (Table 2.1), and the soil was heavily affected by intensive agriculture. Therefore, we can assume that alternative electron acceptors (nitrate, iron and sulphate) were available to fuel anaerobic respiration. There are no indications that other factors limited the activity of microorganisms or superseded other effects of limited O₂ availability. Addition of glucose resulted in increasing C mineralization independently from O₂ availability (Figure 2.4).

Methane production was observed, particularly in the subsoil. Methanotrophic microorganisms can use CH₄ as a source of both energy and C (Galbally et al., 2010; Le Mer and Roger, 2001). In our study, the amount of CH₄ emission observed varied in different soil horizons (Figure 2.5). However, CH₄ concentrations in the headspace were small, never exceeding 15.4 μg l⁻¹, even in subsoil horizons. Furthermore, the smaller CH₄ emissions at the depositional site than those at the eroding site confirmed the poor availability of labile C at the depositional site.
Soil microbial biomass as a controlling factor of C mineralization

Kemmitt et al. (2008) hypothesized that SOC mineralization is independent of biomass, structure and activity of soil microorganisms. They also suggested that the rate limiting step is governed by abiotic processes, which convert non-bioavailable into bioavailable SOC and which cannot be affected by the microbial population. The data of our experiment did not completely support this hypothesis. We found clear indications that the structure of the microbial community affects C mineralization. Although the microbial biomass was significantly smaller at the eroding site than at the depositional site, C mineralization was equal (in the topsoil) or even larger at the depositional site. Therefore, metabolic quotients were significantly larger at the eroding site. This could be interpreted as a stress reaction but also as a different composition of the microbial community (Paterson et al., 2009; Salome et al., 2010). Secondly, the different responses of C mineralization to O₂ treatments in top- and subsoils, with the unexpected large C mineralization under anaerobic conditions in subsoils, indicated a different composition of the microbial community.

2.5 Conclusions

Soil erosion and deposition significantly contributed to a redistribution of organic matter with C accumulation after deposition. This was the result of translocation of organic-rich topsoil material followed by decreased mineralization. In subsoil horizons, stabilization of deposited organic C (decreased mineralization) was more important for accumulated C than translocation. We did not find any indications that soil erosion would result in limited C availability at the eroded sites, even with a large availability of mineral surface areas, which could stabilize organic C.

Carbon cycling was differently regulated in the topsoil than in the subsoils. We hypothesize that the composition of the microbial community was the major reason for this difference. In contrast to the hypothesis of Kemmitt et al. (2008), we think that the structure of the microbial community is critical for C mineralization. In subsoils, microorganisms are susceptible to O₂, which results in large C mineralization under anaerobic conditions. Dissolved organic C is not only a measure of readily available organic C, resembling differences in C mineralization between eroding and depositional sites, but also is a product of microbial activity. The degradability of DOC decreased with increasing soil depth. This decrease depended on microbial activity (and thus it will be larger with increased activity), indicating that relatively stable microbial compounds also contributed to DOC.

Our results support the view that soil erosion contributes to C sequestration. The processes responsible for C stabilization after deposition of the eroded material are still not fully
understood. We can exclude the limited availability of O$_2$ in subsoils as an important controller of C accumulation. There are indications that the small availability of C in colluvial subsoil horizons limits mineralization of deposited C.