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Soil organic matter composition along a slope in an erosion-affected arable landscape in North East Germany

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ABSTRACT

In hummocky landscapes, soil erosion is forming truncated profiles at steep slope positions and colluvial soils in topographic depressions thereby affecting soil organic carbon (SOC) storage. However, the knowledge on the spatial distribution and composition of differently stable organic matter (OM) fractions in arable landscapes is still limited. Here, amount and composition of OM from top- and subsoil horizons at eroded, colluvic, and non-eroded slope positions were compared. The horizons were from a Luvisol at plateau (LV), an eroded Luvisol (eLV) at mid slope (6% slope gradient), a calcaric Regosol (caRG) at steep slope (13%), and a colluvic Regosol (coRG) at hollow position. Water soluble (OM-W) and pyrophosphate soluble (OM-PY) fractions were extracted sequentially. Soil samples, OM fractions, and extraction residues were analyzed with transmission Fourier transform infrared (FTIR) spectroscopy. The soluble fractions were 3% of SOC for OM-W and 15% of SOC for OM-PY. For topsoil samples, extraction rates were independent of slope position. The highest intensities of both, C—H (alkyl groups) and C=O (carboxyl groups) absorption band, were found in FTIR spectra of OM-PY from top and subsoil horizons at the steep slope position (caRG). The C—H/C=O ratio in OM-PY decreased with increasing contents of oxalate soluble Fe and Al oxides from steep slope (0.25 for caRG-Ap) towards plateau, and hollow position (0.09 for coRG-Ap) except for the Bt-horizons. This relation is reflecting that the downslope-deposited Ap material, which is higher in poorly crystalline Fe and Al oxides, consists of relatively stable OM. This OM is enriched in C=O groups that are known for their interaction with soil minerals. These OM-mineral interactions may help explaining C storage in arable soil landscapes.

1. Introduction

The organic matter content and composition in arable soils in hummocky landscapes is often affected by soil erosion, which is forming truncated profiles at steep slope positions and colluvial soils in topographic depressions. However, soil management effects on soil organic carbon (SOC) dynamics have mostly been studied on topsoil samples from relatively flat long-term agricultural field experiments (e.g., van den Bygaart et al., 2003). Rumpel and Kögel-Knabner (2011) identified research gaps in understanding the SOC dynamics and the potential storage of carbon in subsoil horizons. The erosion effect on SOC dynamics was found important for understanding of landscape-scale SOC-dynamics in cropland soils (e.g., Quine and van Oost, 2007).

During an erosion event, soil mineral particles and organic matter (OM) in form of particulate (POM) and soluble OM are removed from upslope positions, redistributed, and finally buried in topographic depressions (Fig. 1). In consequence, soil erosion by water and wind can have a profound effect on the lateral and vertical distribution of SOC in agricultural landscapes (van den Bygaart et al., 2012). While van Hemelryck et al. (2009) among others assume erosion to act as a carbon source because of an increase in OM decomposition, van den Bygaart et al. (2007) assume erosion to act as a carbon sink because SOC at depositional positions is protected from decomposition (van Oost et al., 2007). Thus, erosion and deposition has significant implications for understanding SOC dynamics in soil landscapes (e.g., Berhe et al., 2007). However, the understanding of SOC dynamics in eroding and topographically complex soil landscapes is still limited (e.g., van Oost et al., 2007).

At steep slope positions, erosion leads to a removal of topsoil, while soil minerals from subsoil horizons are incorporated into the topsoil by tillage (Berhe et al., 2012; Gerke and Hierold, 2012). The

Abbreviations: C—H, carbon hydrogen bonds (i.e., alkyl groups); C=O, carbon oxygen double bonds (i.e., carboxylic groups); C—H/C=O, ratio between CH absorption band and C=O absorption band intensity; caRG, calcaric Regosol; coRG, colluvic Regosol; eLV, eroded Luvisol; LV, Luvisol; OM, organic matter; OM-W, water soluble organic matter; OM-PY, sodium pyrophosphate soluble organic matter; SOC, soil organic carbon.

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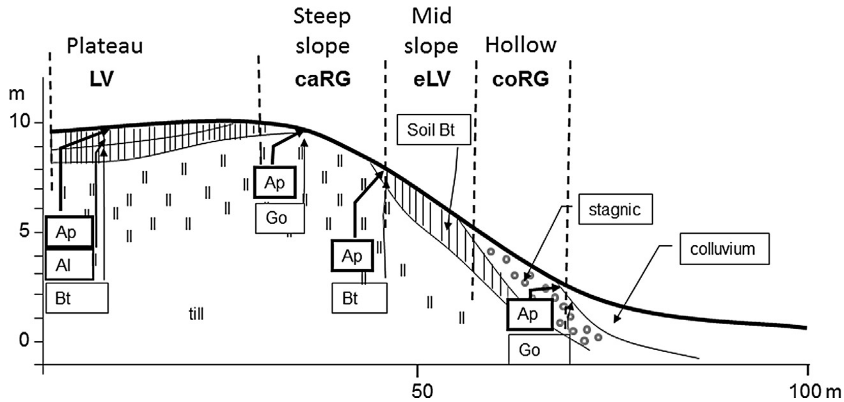

Fig. 1. Visualization of DOM and SOM transfer in an eroded soil landscape; adapted from Gerke and Hierold (2012).

subsoil is poor in SOC, and subsoil minerals rapidly interact with topsoil OM thereby forming OM-mineral associations (e.g., Berhe et al., 2007). Polyvalent cations (e.g., Ca^{++}) and soil minerals especially poorly crystalline iron oxides are known to stabilize OM (e.g., Torn et al., 1997; Masiello et al., 2004; Kögel-Knabner et al., 2008). Schwertmann (1973) suggested the oxalate soluble iron (Fe_{ox}) to represent the poorly crystalline iron oxides in soil.

Aquino et al. (2007) found that carboxyl (C=O) groups in OM are preferably involved in the OM-mineral interactions. Such interactions between OM and soil minerals should be more effective in a soil with more reactive minerals (i.e., at plateau and hollow positions) than in soils depleted in such minerals (i.e., at steep and mid slope positions). Due to erosive downhill transport of particles, relatively stable OM fractions are eventually buried in the hollow position (e.g., Berhe et al., 2007). For erosion affected landscapes it is still unclear how OM-mineral interactions are depending on landscape and soil related differences in OM stabilization and OM composition (Berhe et al., 2012). These authors found a relation between erosion effects and the spatial distribution of OM composition, such as for instance, the relative alkyl (C—H) content in soil OM was found to decrease with depth at depositional positions and to increase with depth at eroding positions.

The OM composition has also been reported to affect soil properties such as the cation exchange capacity (e.g., Kaiser et al., 2008) and wettability (e.g., Capriel, 1997; Ellerbrock et al., 2005). Thus, in addition to the SOC amount, the spatial distribution of OM composition is highly relevant for soil landscape analyses with respect to plant nutrient availability and OM stabilization. The pyrophosphate soluble OM fraction (OM-PY) was frequently described as a measure for OM that interacted with soil minerals or cations (e.g., Masiello et al., 2004; Kögel-Knabner et al., 2008). It could be hypothesized that OM-PY reflected spatial variations in OM composition that are related to erosion-induced modifications in the soil mineral composition more directly than water soluble organic matter (OM-W) and SOM.

The objectives of this study are to characterize amount and composition of SOM and OM fractions along a hillslope, and to identify those OM fractions that are most strongly reflecting effects of soil erosion. For the OM characterization of cultivated topsoil and subsoil horizons, samples from steep slope (13%), mid slope (6%), plateau (2%), and hollow (3.4%) positions along an eroded hillslope in arable post-glacial hummocky soil landscape were analyzed.

2. Material and methods

2.1. Site and soil description and sampling

The CarboZALF experimental site is located in north-eastern Germany near the village of Holzendorf, northwest of the city of Prenzlau (53°23′N, 13°47′E). This site is located at a hummocky arable ground moraine landscape with a total slope length of 208 m at elevations of 51 m to 58 m asl. The slopes vary from flat summit and depression locations with a gradient of approximately 2%, long slopes with a medium gradient of 6%, to shorter and steeper slopes with a gradient of 13%.

The soils developed from glacial till and were under agricultural use for more than 700 years (since the 13th century). From 1950 to 1990 crop rotation was: wheat (*Triticum aestivum*), wheat, rape (*Brassica napus*) except for the hollows, which were temporarily used as grassland in wet years. From 1991 to 2008, the field was cropped with rape (*Brassica napus*), wheat (*Triticum aestivum*), and sugar beet (*Beta vulgaris*) (G. Verch personal communication, July 2015). Soil cultivation was based on conventional tillage using a moldboard plow.

For the period between 1992 and 2012, an average annual rainfall of 485 mm, and a mean annual air temperature of 8.6 °C (at 2 m above the soil surface) were recorded by the Dedelow Experimental Field Station of the Leibniz Centre for Agricultural Landscape Research (ZALF) Müncheberg (www.zalf.de). The average rainfall induced erosion rate for a 10-yr period (1986–1995) at a nearby experimental field was between 0.3 and 1.2 $Mg\,ha^{-1}\,yr^{-1}$ for crop rotations consisting of alfalfa (*Medicago sativa L.*), winter rye (*Secale cereale*), and maize (*Zea maize*) as monoculture or in rotation (Deumlich et al., 2010). The 100-years erosion event caused by rainfall in that region was characterized by the erosivity value of 71 $N\,h^{-1}$ (Deumlich, 1999). However, much stronger rain events may occur locally: For example on 5th June 2007, the region at Holzendorf site was affected by a rainstorm with an erosivity value of 408 $N\,h^{-1}$ (Vogel et al., 2015), which exceeded the 100-years event for almost 6-times.

The combined effect of water and tillage erosion resulted in the development of Calcaric Regosols at steep slopes while Colluvic Regosols developed at foot-slope and hollow positions. Within the 6 ha CarboZALF experimental field, the soil types were characterized by exploratory drilling (i.e., Pürkhauer auger) to select sampling sites at the following four soil types (FAO classification, IUSS Working Group WRB, 2006): Albic Cutanic Luvisol (LV) located at a plateau (slope gradient of 2%), eroded Calcic Cutanic Luvisol (eLV) at the mid slope (6%), Calcaric Regosol (caRG) at an exposed steep slope (13%), and Endogleyic Colluvic Regosol (coRG) at the hollow position (3%) in the topographic depression (Fig. 1). At each slope position, 2 m deep soil profiles of about 2 m width were excavated. Soil samples of about 1 kg mass were taken from the soil horizons. Note that the thickness and order of the subsoil horizons was spatially highly variable such that the sampling depths of the same horizons at the different landscape positions did not always correspond (Table 1). For the Ap horizon at the caRG profile, only the upper part of 0–15 cm depth was studied here in