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Carbon saturation drives spatial patterns of soil organic matter losses under long-term bare fallow

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ABSTRACT

Spatial controls of soil organic carbon (SOC) turnover are not well understood. We hypothesized that spatial patterns of SOC turnover are related to carbon (C) saturation rather than to the size of measurable SOC-pools such as particulate organic matter (POM), determined as SOC in particle-size fractions. Therefore, we repeatedly grid-sampled a field after one, three, seven, and eleven years under bare fallow management, which revealed a spatial gradient from high to low degrees of C saturation. We measured the contents of SOC and the contents of SOC in coarse sand-size (2000–250 μm, POM1), fine sand-size (250–53 μm, POM2), silt-size (53–20 μm, POM3), and fine silt to clay-size fractions (nonPOM, < 20 μm), calculated the degree of C saturation from textural properties and nonPOM contents, and related these parameters to SOC losses. In the first year of bare fallow, the soil contained on average 12.1 g SOC kg^{-1}, of which 0.6 g kg^{-1}, 1.7 g kg^{-1}, and 2.1 g kg^{-1} were lost after three, seven, and eleven years of bare fallow, respectively. The SOC losses within eleven years were spatially variable and varied between 1% and 46% relative to the initial SOC content. In support of our hypothesis, SOC losses were largest at subsites with largest degrees of C saturation (R^2 = 0.83). Although the POM fractions declined most drastically, they only comprised 4 to 9% of bulk SOC, and they did neither correlate with nor explain spatial patterns of SOC losses. We conclude that the concept of C saturation is superior to conventional physical fractionation approaches for predicting spatio-temporal patterns of SOC turnover at sites with a high degree of C saturation.

1. Introduction

Knowledge about soil organic carbon (SOC) turnover is of immense importance to understand and predict SOC storage and CO_2 emissions from the soil. For modeling approaches, SOC is usually conceptualized into pools, which differ from each other in terms of turnover times and stabilization mechanisms (Balesdent, 1996; Skjemstad et al., 2004). The conceptual assignment of SOC to rapidly mineralizable, slow, or passive/refractory pools, and their quantification, however, has engaged researchers ever since (e.g., Christensen, 2001; Smith et al., 2002; von Lützow et al., 2007).

A common way to approach different SOC pools is to fractionate soil into measurable size or density fractions. Particulate organic matter (POM-C) or light-fractions-C then represent a fast to intermediate pool, which is prone to rapid decay at a time-scale of months to very few decades (Dalal and Mayer, 1986; Derrien and Amelung, 2011; Skjemstad et al., 2004). Smaller, non-particulate organic matter (nonPOM-C) represents a slow pool with long turnover times linked to associations with clay or silt particles that protect nonPOM-C from mineralization (Six et al., 2002). Other researchers, such as Flessa et al. (2008) and Marschner et al. (2008), have questioned the stability of this SOC pool and Leifeld and Kögel-Knabner (2005) and Schiedung et al. (2017) suggested that the nonPOM fraction may also comprise considerable amounts of SOM with rather short turnover times. On the contrary, Flessa et al. (2008) and Kiem et al. (2002) reported that POM fractions may also contain refractory SOC compounds. This suggests that the fractionation approach may not be fully adequate to model SOC turnover (e.g., von Lützow et al., 2007). Other mechanisms might interfere with the correlation between physically separated size fractions and SOC turnover times.

It is well known that clay and silt particles promote an accumulation of SOC by protecting SOC through mineral associations (e.g., Hassink, 1997; Six et al., 2002). Thus, at sites with elevated contents of these fine fractions larger proportions of SOC could be protected from mineralization. Hassink (1997) even assumed that the capacity of soils to protect SOC from mineralization is limited by the proportion of clay and silt particles (fine fraction < 20 μm). The author concluded that once the amount of SOC in the soil approaches this protection capacity,

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little if any additional SOC can be effectively protected by mineral associations. In other words C saturation has been reached. According to the concept of C saturation, faster SOC turnover might be expected from sites that are close to C saturation, as considerable amounts of SOC are not protected efficiently. Several studies thus used this concept with the objective to define an upper limit of soils to sequester SOC (e.g., Beare et al., 2014; Hassink, 1997; West and Six, 2007), or to determine the potential of soils to sequester additional SOC, often referred to as saturation deficit (Angers et al., 2011; Beare et al., 2014; Wiesmeier et al., 2014, 2015). The latter has received particular interest in terms of strategies to mitigate climate change (Lal, 2011). An explicit consideration of C saturation effects on SOC stability and turnover rates has received far less attention though it seems reasonable to speculate that C saturation is an important driver of SOC stability and mineralization rates (e.g., Chung et al., 2010).

Long-term bare fallow experiments offer a unique opportunity to study the turnover of SOC and its regulating factors as all changes in SOC are directly linked to decomposition without being blurred by recent input of biomass (Barré et al., 2010; Kiem et al., 2000). Single field sites, even those uniformly managed over many years, frequently reveal a heterogeneous distribution of SOC and related soil properties such as texture or rock fragments (Bornemann et al., 2011; Hbirkou et al., 2012; Kölbl and Kögel-Knabner, 2004; Patzold et al., 2008). To our knowledge, the spatial heterogeneity of these soil properties has not been repeatedly assessed at fields under bare fallow management. This assessment would allow an evaluation of the soil parameters that regulate the losses of SOC, while keeping other factors such as climate, management, or former SOC input rather constant.

The aim of this study was to identify factors that regulate spatial patterns of SOC turnover. Therefore, we monitored the spatial patterns of SOC losses from a former arable field site that was eleven years under continuous bare fallow management, and which was suspected to reveal spatial variations in C saturation (see Bornemann et al., 2011). We hypothesize that SOC losses occur especially at sites with large SOC contents but low protection capacity, i.e., at sites which are close to C saturation. In addition we performed a soil particle-size fractionation to test the use of the C saturation concept against that of conventional C-pool analyses for explaining spatial patterns of SOC losses.

2. Materials and methods

2.1. Study area

The study was conducted at a 60 × 190 m field site near the village of Selhausen (50°52′09.34″N; 6°27′00.58″E), located in the Lower Rhine Embayment, Germany. The mean annual temperature is 9.8 °C and the mean annual precipitation is 694 mm. The site belongs to the TERENO long-term observatory sites (Zacharias et al., 2011), and it is integrated within the frame of the collaborative research centre CRC/TR32 to analyze spatial patterns in soil-vegetation-atmosphere systems (Simmer et al., 2015). Hence, we had access to a large field site under long-term monitoring, allowing us to resample the field at a high spatial resolution.

The original site heterogeneity prior to prolonged fallow management was reported by Bornemann et al. (2010, 2011). As outlined in this work, the field is weakly inclined (< 3.5°) over ~180 m in east-west direction (Fig. 1d). The Rhine, the Meuse, and the Rur rivers provided the underlying fluvial deposits, which were covered by aeolian sediments during the Pleistocene. At the downslope area aeolian substrate completely covered the underlying sediments, with only small contents of rock fragments mixed into the topsoil by postglacial soil erosion. Fluvial deposits with larger proportions of rock fragments penetrated the Pleistocene loess layer towards the summit, where a gravel stripe crossed the test site in northwestern direction (Bornemann et al., 2011).

Due to the spatial differences in substrate, soil properties were highly heterogeneous and followed a gradient from northeastern upslope positions to southwestern downslope positions. Downslope, the texture was defined by a slightly gravelly silt loam with < 10% rock fragments (> 2 mm) and > 70% silt within the fine earth. Towards the northeastern upslope parts, contents of sand and rock fragments increased gradually towards a very gravelly loam with > 45% rock fragments and < 60% silt within the fine earth (Fig. 1a, see also Bornemann et al., 2010, 2011). Soils ranged from Stagnic Luvisols downslope to Orthic Luvisols on the slope to Dystric Leptosols on upslope parts (Bornemann et al., 2011; WRB, 2007). The field had been under arable management for at least 100 years (mainly sugar beet and winter wheat were cultivated) and was converted to continuous bare fallow after the vegetation season towards the end of 2005.

The field was kept bare for this purpose by application of herbicides, which were regularly applied as soon as there was visible evidence of weeds, as well as by mechanical tillage. Tillage refers to a 5 cm deep ploughing, which was practiced once per year between 2005 and 2010. Although it was not possible to prevent any C inputs (here, few herbs, particularly thistles regrew in the bare-fallow field that had to be removed by regular surface spraying with glyphosate), we assume these inputs to be negligible, in line with Barré et al. (2010).

It should be noted that we cannot fully exclude effects of soil erosion on SOC patterns. Auerswald et al. (2009) summarized that the average soil erosion rate on German bare fallow sites was 80 t ha^{-1} for sites standardized to 5.1° slope gradient and 200 m flow path length. At our field site, the slope was inclined on average by 1.7° over a length of ~100 m (thereafter the slope gradient was < 1°, see Herbst et al., 2012 for details). In line with values reported by Auerswald et al. (2009), we may expect that soil erosion did not exceed 13 t soil h^{-1} year^{-1}. This would correspond to ~1 cm soil gain at deposition sites within the considered period of 11 years. Assuming that in worst case eroded soil is enriched in C by a factor of 2 (e.g. 20 g C kg soil^{-1}) and deposited to a downslope soil with SOC contents of 10 g C kg soil^{-1}, this would increase the average SOC content in a sample taken from 0 to 25 depth from 10.0 g SOC kg to 10.4 g SOC kg soil^{-1}. The actual amount is likely lower as (1) parts of the deposited SOC are mineralized in the time between deposition and sampling and (2) because the large proportion of rock fragments creates a rough surface, which might mitigate the extent of soil erosion (Poesen et al., 1994). In light of the large range of observed SOC losses (ranging from 0 to 6.6 g C kg soil^{-1}), we assume that even this theoretical worst-case impact of 0.4 g C kg soil^{-1} would not put our conclusion into risk.

Further, we are aware that variable proportions of rock fragments at this field site may induce spatial variations in soil temperature and soil water content. However, we are confident that the relation between C saturation and SOC loss is not blurred by spatial variations in temperature and soil moisture. Herbst et al. (2012) investigated soil temperature at this field site and described a “lack in variability in soil temperature at field scale”. In contrast to temperature, however, Herbst et al. (2012) described that soil moisture increased from upslope positions to downslope positions and that water deficiency might temporally limit soil respiration at upslope positions. However, there are two reasons why we believe that spatial variations in soil moisture have no major effect on our conclusion. First, the risk that the soil dries out is reduced under bare fallow conditions due to lacking transpiration (Auerswald et al., 1996). Second, if SOC turnover was temporally limited by water deficiency at upslope positions, it would even reinforce our conclusion: Largest SOC loss occurred at C saturated upslope positions despite a possible lack of water.

2.2. Experimental design and soil sampling

The field was sampled for a regular grid of 10 × 10 m (n = 64) (Fig. 1a). Soil samples were collected in 2006 (Bornemann et al., 2010), and, using an identical procedure, again in 2008, 2012, and 2016, corresponding to one, three, seven, and eleven years after conversion to