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Soil organic carbon stocks in permafrost-affected soils in West Greenland

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ABSTRACT

Little is known about soil organic carbon (SOC) stocks in permafrost-affected soils in Greenland. Generally, occurrence and stocks of SOC in permafrost-affected soils of the Arctic were underestimated for many years. Compared to the assumed dimension of the influence of carbon dynamics on climate change this knowledge should be substantially widened. A total of 155 soil samples were used to get a better understanding about SOC stocks, depth function and spatial distribution of SOC in permafrost-affected soils in a characteristic deglaciated valley in West Greenland southeast of Kangerlussuaq. The valley is characterized by a high variability of active layer thickness and pedo-variance mainly caused by topography.

The average SOC stock of the Umimmalissuaq valley is 9.9 kg m^{-2} in the upper 30 cm and around 30 kg m^{-2} in the first meter, which is remarkably higher than regional predictions with 6–15.9 kg m^{-2} in the first 100 cm. To account for spatial heterogeneity landscape units are developed which are most useful for grouping and predicting SOC stocks. The SOC store measured 14.2 kg m^{-2} in the upper 30 cm, 11.5 kg m^{-2} on north-facing slopes, and 8.4 kg m^{-2} on south-facing slopes. Little SOC stocks with around 6 kg m^{-2} were found under abrasion fields particularly on hilltops and moraine ridges. Soils on south-facing slopes usually have very low SOC stocks in deeper soil horizons except of organic rich horizons in rarely occurring paleosols. North-facing soils and valley bottom slopes generally have high SOC stocks of around 19 kg SOC m^{-2} in soil horizons with a depth of 30–100 cm.

In general, the main influencing parameter on SOC stocks is the soil organic matter input from the vegetative cover. The vegetative cover is mainly a result of topographic position and aspect related to the ice margin and katabatic winds. Soil moisture and high active layer may influence SOC stocks positively.

1. Introduction

During the past century, warming of local Arctic climate (>60°N) was approximately 0.09 K per decade, whereas on the entire northern hemisphere temperatures rose only by 0.06 K per decade (McBean, 2005; Intergovernmental Panel on Climate Change (IPCC), 2001). Soil temperatures at the permafrost table of artic soils already increased by 3 K since the 1980s (Lemke et al., 2007; Vaughan and Comiso, 2013). Different models predict an increase of the global mean surface temperature of up to 4.8 K until 2100 (Intergovernmental Panel on Climate Change (IPCC), 2008). Especially winter air temperatures will be significantly warmer (4–7 K) across the terrestrial Arctic (>60°N Arctic Climate Impact Assessment, 2004). With rising air temperature, carbon sequestered in permafrost-affected soils is likely to decompose to a larger extend and become a source of atmospheric carbon. Increasing active-layer depth and altered hydrological conditions can lead to intensified mineralization, which is followed by higher carbon turnover rates and CO_2 emission from soils (Dutta et al., 2006; Koven et al., 2009; Schuur et al., 2008; Wagner and Liebner, 2009; Zimov, 2006). Permafrost-affected soils cover about 27% of the terrestrial surface above 50° N (Jones et al., 2009). A high amount of soil organic carbon (SOC) is currently frozen and thereby partially excluded from interactions with the atmosphere and biogeochemical cycles (Wagner and Liebner, 2009).

The SOC pool in the first 300 cm of arctic soils includes about 50% of the estimated global terrestrial belowground organic carbon, which makes about 1024 Pg C and up to 496 Pg (1 Pg = 10^{15} g) within the uppermost one meter (Tarnocai et al., 2009). Compared to the ocean and to the forest carbon pool the carbon pool of permafrost-affected soils is sensitive to warming and changing hydrological circumstances (Gruber et al., 2004; McGuire et al., 2009; Schuur et al., 2009). Especially lowlands in the permafrost zone, which are common in the Arctic and Subarctic, hold large amounts of labile SOC (e.g. approx. 60 kg m^{-2} 0–100 cm in bogs of the Usa Basin, Siberia, Hugelius and Kuhry, 2009) and are prone to change under climate warming (Christensen, 2004; Khvorostyanov et al., 2008).

SOC stocks in arctic permafrost-affected soils, particularly Cryosols and Histosols, were underestimated for many years (Tarnocai et al., 2009; Burnham and Sletten, 2010) and the knowledge about SOC

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dynamics in periglacial landscapes is little compared to the assumed dimension of the influence of carbon dynamics on global warming. In addition to the underestimation of SOC stocks in permafrost-affected soils in the Arctic, which is mainly due to a lack of knowledge about specific soil processes, to limited depth consideration, and underestimation of landscape heterogeneity, our knowledge of the spatial distribution of SOC in periglacial landscapes is quite limited (Jobbagy and Jackson, 2000; Ping et al., 2008; Tarnocai et al., 2009). Since soils in general are in the "front line" (Schmidt et al., 2011) of global environmental change it is important to improve global estimations by conducting regional to local research and including i.e. deeper soil layers, and consideration and calculation of uncertainties. Local studies, e.g. from Alaska, Canada or Siberia, give a picture about SOC stock distribution in specific study areas, but this is not sufficient to create a global SOC distribution map. Ping et al. (2008) combined local SOC measurements with the distribution of landscape types and proved that this approach leads to a regional assessment of SOC stocks. To improve knowledge about local SOC distribution further studies are necessary, which also account for small-scale landscape heterogeneity and focus on other regions (cf. Kuhry et al., 2002).

In our study, we investigate the SOC stocks and their spatial and vertical distribution in the low mountain ranges of West Greenland adjacent to the inland ice margin. We chose a formerly glaciated valley southeast of Kangerlussuaq which exhibits the characteristic landscape structure, soils and sediments under periglacial climate conditions. The main objective of our investigation is to quantify the SOC stocks of permafrost-affected soils in relation to characteristic landscape units (valley bottom, crest, north-facing and south-facing slopes).

1.1. Study area

The study area is located in the Umimmalissuaq valley in West Greenland near Kangerlussuaq, where the ice sheet forms flat outlet glaciers onto a low mountain range. The distance from ice sheet to the coast is about 150 km. Kangerlussuaq area is dominated by an E–W and ENE–WSW oriented valley system, which leads the melt water to the sea. The Umimmalissuaq valley, about 30 km SE of Kangerlussuaq, stretches from the ice cap to the west for approximately five kilometers. The study area in the Umimmalissuaq valley expands from the terminal/lateral moraine of the Ørkendalen glacier (66°56′50″ N, 49°58′53″ W) to a melt water lake in the west (66°56′14″ N, 56°06′12″ W). In the north and south, it is bordered by mountain ranges with topographic differences of about 100 m and in the south partly by the lake Umimmalissuup Tasinngua (Fig. 1; names taken from Scholz and Grottenthaler, 1988; suppl. Mat. Soil_pits_JHenkner_2016_Geoderma.kmz).

The Kangerlussuaq Airport weather station shows an arctic continental climate with an annual mean air temperature of −5.7 °C and 149 mm annual mean precipitation (1976–1999 (Cappelen et al., 2001; Carstensen and Jørgensen, 2009)). Summer day temperatures are about 15 °C (June–August), highest precipitation occurs in August (33 mm). In summer, there are about 150 snow-free and 80 consecutive frost-free days (Bullard, 2011). The study area itself is likely to be slightly colder but sunnier, drier and with higher wind speed compared to Kangerlussuaq (situated around 35 km west of the ice margin), because of its location adjacent to the ice margin. The bimodal wind regime is dominated by katabatic winds from the ice sheet, channeling in the valley especially in winter, and westerly winds generated by Atlantic storms, which are less frequent and less strong (Dijkmans and Törnqvist, 1991). Recent climate data for the period 1979–2008 (Boas and Wang, 2011) already show warming and higher precipitation (mean annual air temperature − 4.8 °C, mean annual precipitation 257 mm) with a particular increase of winter temperature and summer precipitation.

The bedrock in the Umimmalissuaq valley is mainly Precambrian gneiss, extensively covered by glacial, glaciofluvial or aeolian material (Scholz and Grottenthaler, 1988; Henriksen, 2008; Ozols and Broll, 2003). Non-calcareous aeolian sand and silt forms the uppermost parent material of soils, underlying glacial material contains larger rocks. Soils are mostly dry, water saturated conditions were measured on the north-facing slope only a few days in summer (Hanebeck, 2012). Although mapped as continuous permafrost regime (Maurer, 2007), the soils in the Umimmalissuaq valley show characteristics of a discontinuous permafrost regime depending on topography. Permafrost is present within about 1 m from the soil surface on north-facing slopes. The active layer depth varies throughout the valley and could not be assessed in the valley bottom or south-facing slopes.

The study area of Umimmalissuaq valley is located within the moraines of the Umîvît/Keglen (according to Ten Brink, 1975) dated to around 7300 cal b2k (UtC-1987, UtC-1990, after Van Tatenhove et al., 1996). Small moraines stretch across the valley from the north to the south, a larger one, reaching up to 20 m above the valley bottom, stretches northeast to southwest and west (Fig. 1). Accordingly, deglaciation in this area began about 7300 years ago, which gives a maximum age for the beginning of the accumulation of SOC. The SOC stocks from 0 to 30 cm depth in crest positions/south facing slopes accumulated within around 2000 years (AMS^{14}C dates of bulk SOM from profile C39 in 36–40 cm depth: 19 cal BCE to 233 cal CE (Erl-16,616) and in 49–53 cm depth 7–244 cal CE). This reflects a slow accumulation and decomposition rate of organic matter (cf. Wagner et al., 2009).

Deflation forms and sediment accumulation occur due to active aeolian sediment transport seaward along the valley (Müller et al., 2016). Further geomorphologic features are ice-wedge polygons at north-facing slopes and ice-wedge pseudomorphs and earth hummocks at the valley bottom (Figs. 1–5). Vegetation is mainly controlled by microtopography, depth to bedrock or permafrost, and the strong continental climate, especially wind speed and direction (Ozols and Broll, 2003).

2. Methods

Field work was done in summer 2009 and 2011. All in all 39 soil profiles were described according to WRB 2006 (FAO, 2006; IUSS Working Group WRB, 2015) and the German classification system (Ad-hoc Arbeitsgruppe Boden, 2005) and sampled. A categorical assessment of current soil moisture was done (FAO, 2006). The soil profiles are located along four catenas reaching from the top of a north-facing hill through the valley to the opposing south-facing hill (Fig. 1).

Bulk soil samples were taken horizonwise of all genetic horizons. Additionally, systematic volumetric soil samples (3×100 cm^3) were taken from depth increments (0–5, 5–10, 10–20, and 20–30 cm) to get a comparable systematic sample set. Total carbon and total nitrogen content [mass%] were analyzed using oxidative heat combustion at 1150 °C in a helium atmosphere (element analyzer "vario EL III", Elementar Analysesysteme GmbH, Germany, in CNS mode). Since there is no geological source of inorganic carbon (Henriksen, 2008) and pHCaCl$_2$ is about 5.5, it is assumed that total carbon equals SOC. Soil-pH was determined using a soil to solution (0.01 M $CaCl_2$, H_2O, with Sentix 81, WTW, pH 340) ratio of 1:2.5 (Blume et al., 2002). Bulk density [g cm^{-3}] was gravimetrically determined (Eq. (1); cf. Don et al., 2007). SOC stocks [kg m^{-2}] were calculated using $BD_{finesoil,i}$ [g cm^{-3}] and SOC content ($SOC_{conc,i}$ [%]) of each depth increment with a certain thickness ($depth_i$ [cm]) (Eq. (2); Dörfer et al., 2013, cf. Hugelius et al., 2010; Tarnocai et al., 2009; Goidts et al., 2009).

$$BD_{finesoil} = \frac{(mass_{sample} - mass_{stones})}{volume_{sample}} \quad (1)$$

$$SOC_{stocks} = \sum_{l=1}^{n} 0.1 \times depth, i \times BD_{finesoil,i} \times SOC_{conc,i} \quad (2)$$