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Abstract

In contrast to good knowledge of dissolved organic matter (DOM) adsorption on mineral soils in temperate climate, the behavior of DOM in frozen mineral horizons located under peat soils of permafrost-affected regions remains poorly characterized. Yet, these regions contain sizeable and potentially highly labile pools of organic (peat) carbon (C) that may migrate downwards across mineral layers in case of massive thaw in frozen peatlands induced by on-going climate warming. To quantify these pools and the lability of DOM in permafrost peat soils, we performed experiments focusing on dissolved organic carbon (DOC) desorption from, and adsorption onto, mineral horizons (iron-poor and iron-rich sands as well as silt loam) from the largest frozen peatland in the world, the western Siberia Lowland (WSL). Desorbed DOC ranged between 0.1 and 0.6 mg C g_{soil}^{-1} depending on type of mineral substrate. The adsorption of peat leachate DOM ranged between 0.1 and 0.5 mg C g_{soil}^{-1} being highest in Al-Fe-rich mineral horizons.

Field measurements of C pools in peat and underlying mineral horizons over 1 m depth in the discontinuous permafrost zone yielded 47 and 15 kg C m^{-2}, respectively. The organic carbon (OC) adsorption capacity of the 1 m - thick mineral layers represented < 2% of total amount of OC containing in the 1 m - thick peat layer. However, this adsorption capacity is comparable to the amount of DOC that can be leached from overlaying peat horizons (18 %). On average, out of 1.38±0.13 kg C m^{-2} capable of being initially released from the upper 0-100 cm of peat, 0.25±0.19 kg C m^{-2} can be adsorbed by the underlying 100-200 cm of Fe- and Al-rich sands and clays. The remaining 1.13 kg C m^{-2} can be exported to lakes and rivers. Therefore, DOC released during peat thaw in upper soil horizons in permafrost regions can be sizably attenuated via adsorption on mineral layers. This should be taken into account when modeling the feedback of permafrost thaw on C export and CO_2 emissions.

1. Introduction

The fate of organic carbon (OC) stored in peat and mineral frozen deposits of the Northern Hemisphere under climate warming and permafrost thaw scenario is a central issue in modern Earth surface biogeochemistry (Feng et al., 2013; Kramer and Chadwick, 2018; Ciais et al., 2019; Opfergelt, 2020). Global C storage in peatlands is 530 ± 160 Pg C, and 80 % of this storage occurs in northern peatlands (Hugelius et al., 2020). The western Siberian Lowland (WSL) contains a vast amount of carbon (70 Pg, Sheng et al., 2004) in the form of massive 1-3 m thick peat deposits, most of which are currently frozen (Kremenetski et al., 2003; Smith et al., 2004). The frozen peat in the WSL is highly vulnerable to thawing; permafrost temperatures in these regions are close to zero and zones of sporadic to discontinuous, rather than continuous, permafrost prevail (Romanovsky et al., 2010). As a result, a sizeable proportion of OC stored in this peat can be exchanged with the hydrological network and the atmosphere. There are two main pathways of OC removal from the soil system. A part of OC is dissolved via peat leaching by soil water and removal to the surface waters, whereas another part can be mineralized by bacteria and emitted into the atmosphere (Schuur et al., 2008, 2009; Gentsch et al., 2015a; Serikova et al., 2018, 2019). However, frozen mineral horizons, which underlay peat layers, can act as important sinks of OC due to their high affinity to DOC (Gu et al., 1994; Kaiser, 1998; Kalbitz et al., 2005; Kothawala et al., 2012). For example, in the south of western Siberia, these mineral soils are known to decrease lateral export of DOM: the catchments of southern rivers exhibit extended unfrozen mineral horizons compared to those of northern rivers because the latter drain essentially frozen peat (Frey and McClelland, 2009; Kawahigashi et al., 2004; Pokrovsky et al., 2015). This mechanism of surface DOM retention by deep horizons of mineral soil is thought to be of importance for high latitude and boreal regions (Kaiser and Guggenberger, 2000) without an impermeable permafrost table, where infiltration of organic-rich surface waters to deep mineral layers and DOC sorption on clay minerals may occur causing a decrease the overall DOC export (Smedberg et al., 2006).

A number of factors are known to govern DOC retention in mineral soils [typically B horizon of podzols as noted by Zysset and Berggren (2001)] such as: 1) adsorption onto Al and Fe sesquioxides (Kaiser et al., 1996; Kaiser and Zech, 1998; Kleber et al., 2005; Kothawala et al., 2009; Moore et al., 1992; Pengerud et al., 2014), and 2) co-precipitation of DOM with Al and Fe oxy(hydr)oxides (Buurman, 1985; Kaiser, 1998; Nierop et al., 2002; Mu et al., 2016) and 3) DOM retention onto other mineral phases (i.e.,

clay minerals, Mn oxides, Zysset and Berggren, 2001; Jardin et al., 1989; Oosterwoud et al., 2010). In contrast to numerous works demonstrating a dominating role of mineral surfaces in DOM transport in non-permafrost soils (i.e. Groeneveld et al., 2020; Kaiser et al., 1996; Kaiser and Guggenberger, 2000; Kothawala et al., 2009; Porras et al., 2017; Ussiri and Johnson, 2004)), limited information is available on transport of DOM between organic (i.e., peat) and mineral (sand, clay) horizons in permafrost-affected soils. In the latter settings, the underlying DOC-poor mineral horizons can be separated from organic-rich upper soil/peat layers via an impermeable permafrost boundary. As a result, the water and OM residence time and DOM bioavailability/biodegradation potential in the mineral soil horizons of the permafrost zone can be drastically different from those of the non-permafrost zone, and even autochthonous subducted OM can be protected from decomposition via, for example, cryoturbation (i.e., Čapek et al., 2015).

A few experimental studies of DOC adsorption on the mineral fraction of soils in arctic and subarctic regions provided experimental parameters of DOC retention by mineral layers in the non-peat context (Dahlgren and Marrett, 1991; Kothawala et al., 2009; Pengerud et al., 2014). In these works, the authors performed laboratory assessment of adsorption constants, partitioning coefficients and reactive soil pool of C for a large number of mineral horizons. Several other studies demonstrated the importance of organo-mineral (Prater et al., 2020), and organo-ferric (Patzner et al., 2020) associations and bioavailability of mineral-associated organic matter in permafrost soils (Gentsch et al., 2015b). Of special note is a work on the role of soil clay fraction in DOM sorption across the forest-tundra gradient in Central Siberia (Kawahigashi et al., 2006). These authors demonstrated that large concentrations of DOM comprising large shares of hydrophilic fraction can pass mineral soils where the active layer is thin (such as Gelisols), and enter streams, whereas soils with deep active layer such as inceptisols release little DOM because of DOM adsorption onto their thick mineral horizons. Furthermore, Prokushkin et al. (2007) related the DOM adsorption capacity of the mineral soils to the lateral export/retention of DOM by rivers across the gradient of active layer thickness in the non-peatland soils of the Yenisey basin in Central Siberia and demonstrated that DOM produced in the active layer is retained and stabilized in underlying, unfrozen mineral soils. Sizable retention of DOM on unfrozen mineral horizons was later confirmed by quantitative estimations of DOM release from frozen peat layers and impact of DOM adsorption capacity by mineral horizons on DOM

export fluxes in rivers of western Siberia, draining through partially frozen peat deposits (Pokrovsky et al., 2015). However, quantification of parameters controlling DOM adsorption onto and release from frozen mineral horizons and quantitative estimation of reaction parameters in the peatland context remains very limited. A lack of this information precludes the predictive assessment of the link between the permafrost thaw (deepening of the active layer) and C emission and export in frozen peatlands of arctic and subarctic regions via quantitative modeling at the catchment scale (i.e., Jantze et al., 2013; Tang et al., 2018).

To bridge this knowledge gap, we estimated the soil organic carbon (SOC) pool in organic and mineral layers in both thawed and frozen horizons in typical soil environments of the permafrost peatlands in the western Siberia Lowland (WSL). We further performed laboratory experiments on DOC release from peat and adsorption onto various mineral soils that were sampled in permafrost peatlands of the region. Our experimentation goal was to answer 3 specific questions: 1) How much DOC issued from frozen peat can be retained by mineral horizons during thaw of peatlands? 2) Is there a strong control of Al- and Fe-oxides on DOC adsorption by mineral soils (as known for non-permafrost environments)? and 3) Is the magnitude of DOC cycling between organic and mineral horizons at the watershed scale comparable with lateral export of C from watershed soils? It is anticipated that, when combined with natural observations of SOC pools in peat and mineral horizons as well as with DOC riverine fluxes, these experimental results will allow possible prediction of changes in soil C stocks and fluxes that are induced by thawing of organic (peat) horizons and transport of DOC over the mineral layers.

2. Materials and methods

2.1. Site description and soil sampling

Peat core samples were collected in the northern part of west Siberian Lowland at the Khanymey INTERACT station (https://eu-interact.org/) in the discontinuous permafrost zone (forest-bog biome) and at the Tazovsky site in the continuous permafrost zone (polygonal tundra), **Fig. 1 and S1.** A detailed description of the sites is provided elsewhere (Loiko et al., 2019; Morgalev et al., 2017; Raudina et al., 2018, 2017). Peat and mineral horizons were sampled in the end of August 2017 when the active layer thickness (ALT) was at maximum. The peat (0.4-1.5 m thick on Khanymey) developed on sands with minor amounts

of clay and silt. The dominant soil catenae are flat-mound bogs along the watershed divides which comprise the Hemic Cryic Histosols on mounds (palsas) and Fibric Gelic Histosols on fens. The soil specifications from here forward are termed according to WRB IUSS (2015). Average active layer thickness is 41±5 cm at the top of mounds and between 1 and 2 m in depressions and hollows. For experiments described below, we collected the frozen peat from the peat mound and three types of sandy soil horizons (0.05-0.50 mm grain size) underlying peat deposits (E, Bs and Bh horizons). The mineral layers had C content from 1.4 to 40 mg g^{-1} and Fe content from 0.4 to 3.1 mg g^{-1} (**Table 1**). Typical soil profiles containing thawed and frozen peat and frozen mineral deposits from the discontinuous permafrost zone collected for leaching and adsorption experiments are shown in **Fig. 1**.

The second sampling site was located in the continuous permafrost zone, near town of Tazovsky, where polygonal tundra developed on thick (2-4 m) peat deposits underlain by silt/loam mineral deposits. Note that in the continuous permafrost zone (Tazovsky), we collected only frozen silt loam material (a mixture of < 0.05 and < 0.002 mm grain size underlying a 4 m-thick frozen peat) because such type of substrate is absent in Khanymey.

Core samples of frozen peat and mineral deposits were extracted using a motorized Russian peat corer (UKB-12/25 I, Russia) with a 4-cm diameter corer sterilized with 40% ethanol prior to each extraction. Thawed horizons were sampled laterally at discrete depths from a 1 x 1 m pit using a ceramic knife. The lateral stock of organic carbon in peat and underlying mineral deposits was estimated in the discontinuous permafrost zone, were we disposed 11 independent cores (7 to 15 horizons in each core), sampled from positive (mounds) and negative (hollows, depressions) micro-landscapes (**Table S1**). In order to evaluate the lateral storage of C in peat and mineral horizons, each 5 cm of both peat and mineral horizons of the drilled core were collected using a clean ceramic knife and stored frozen at -18°C before freeze-drying in the laboratory. The density and C concentration were measured on bulk sample with a resolution of 5 cm.

2.2. DOC leaching from peat

Peat leachate was prepared using the frozen peat horizon (55-75 cm), which was sampled below the bottom boundary of the active layer (~40 cm) of the frozen mound in the discontinuous permafrost site

(Khanymey). The frozen peat was reacted at 10 g_{dry} peat L^{-1} with Milli-Q water in a sterile glass bottle under gentle stirring over 24 h at room temperature. The obtained solution was filtered through a sterile Sartorius filtration unit (0.22 µm) and stored in the refrigerator prior to adsorption experiments.

The capacity of frozen peat to release DOC was studied by varying the concentration of peat in Milli-Q water (1, 2, 5, 10 20 and 100 g_{peat} L^{-1}) placed in 60-mL aerobic glass reactors and measuring the resulting DOC concentration (< 0.45 µm) at the steady-state (pseudo-equilibrium), typically achieved after 48 h of incubation. The kinetics of DOC leaching from frozen peat was studied in batch glass reactors at 1 g_{peat} L^{-1}, 25°C over 132 h of exposure with sampling after 1, 3 and 9 h during first day and after 48, 72 and 132 h the other 5 days. All peat leaching experiments were conducted in replicates, in two parallel reactors. At each sampling, the whole reactor was sacrificed.

2.3. DOC release from and adsorption onto mineral horizons

Three fine sand horizons (E, Bs and Bh exhibiting different concentrations of Fe, Al and C_{org}) from the discontinuous permafrost zone and one silt loamy horizon (Bgh) from the continuous permafrost zone were used for DOC adsorption and release experiments (**Table 1**). The desorption of DOC from Al-Fe-rich sand was studied in 30-mL sterilized glass batch reactors with Milli-Q water, where different amounts of solid phase (0, 0.5, 1, 2, 5, 10, 20 and 100 g_{solid} L^{-1}) were added. The reactors were shaken on a ping-pong shaker at 25°C and the fluid was sampled after 2 days of exposure, filtered through a 0.45 µm filter unit and stored in the refrigerator for several days prior to analyses. These experiments were run in duplicates, and for each sampling, two entire reactors were sacrificed.

Preliminary DOC adsorption experiments were run in closed (batch) sterilized glass reactors containing 20 ml of peat leachate into which a different amount of mineral fraction (0, 0.02 and 0.2 g) was added. The peat leachate - mineral suspensions were shaken at room temperature over 24 h. These experiments allowed for defining optimal conditions for studying adsorption capacity in dynamic and static experiments. The DOC desorption experiments were also run in dynamic system using two parallel flow-through column reactors, well adapted for both mineral and organic substrates (i.e., Sanderman and Kramer, 2017). For these experiments, we used only the Bh horizon from the discontinuous permafrost zone because

this sample contained the highest amount of organic carbon. MilliQ water was delivered from the bottom to the top of a glass column reactor (diameter 1 cm) containing 3 g of mineral phase with a flow rate of 0.05±0.01 mL min^{-1}. The resulting 0.45 µm - filtered outlet solution was analyzed each day over the period of 10 days of time and the breakthrough curves of DOC concentration and UV absorbance vs elapsed time were compared between the control run (DOC-free for desorption and mineral-free for adsorption) and the experiment. These dynamic experiments were conducted in duplicates and allowed to constrain the optimal conditions of exposure time and the range of solid and DOC concentration suitable for reliable measurements of adsorption and desorption effects.

The main experiments were conducted for each of the 4 mineral horizons listed in **Table 1**, at 1 $g_{mineral\ soil}$ L^{-1} and 0, 6, 11, 20, 32 and 58 mg DOC L^{-1} concentration of peat leachate, using 30- mL sterile glass bottles that were gently stirred on a ping-pong shaker (100 rpm) at 25°C. After 2 days of exposure time, the supernatant was filtered through a 0.45 µm disposable filter and stored in the refrigerator for several days before the DOC analyses. For these experiments, we used three independent sub-samples from each soil horizons and each sub-sample was run in duplicates. This provided six identical replicates for each soil horizon. For the sampling, the entire vial was sacrificed.

In order to account for the release of 'native' DOC when treating the adsorption results, we used the initial mass isotherm approach which is standard in this field (Kaiser et al., 1996; Kothawala et al., 2008; Moore et al., 1992; Nodvin et al., 1986; Ussiri and Johnson, 2004; Vance and David, 1989):

$$RE = m \times [DOC]_{added} - b \quad (1)$$

where RE is the net DOC sorption (mg C g^{-1}) and b is the amount desorbed to a solution containing no DOC (mg C g^{-1}), corresponding to the release of native soil organic carbon (OC). The $[DOC]_{added}$ is the DOC concentration added to the soil suspension (mg C g^{-1}). The empirical partition coefficient *m* corresponds to the slope of this dependence and represents a fraction of total reactive DOC that is adsorbed by the soil. Further, following Nodvin et al. (1986) and Pengerud et al. (2014), we defined the reactive soil pool (RSP; mg C g^{-1}) as the amount of native soil OC that can readily exchange with DOC in solution under the given experimental conditions as:

$$RSP = b / (1 - m) \quad (2)$$

2.4. Analyses

The DOC was measured in filtered samples using a Shimadzu TOC-VCSN with an uncertainty of 3 % and a detection limit of 0.1 mg L^{-1}. The specific UV-absorbance (L mg^{-1} m^{-1}) was measured in 10 mm quartz cuvette at 254 nm using a CARY-50 UV-VIS spectrophotometer (Bruker, UK) and normalized to the DOC concentration in the sample.

The C and N concentrations in dry peat and mineral samples were measured by Cu-O catalysed dry combustion at 900 °C with ≤0.5% precision for standard substances (Thermo Flash 2000 CN Analyzer). The total major composition of mineral fractions (Si, Al, Fe, Ti, S, Na, Ca, Mg, Na, Mn, K, P, O) was measured by X-ray Fluorescence (Bruker AXS GmbH) in tablets of Li borate fused with mineral powder, with an uncertainty of 2% and a detection limit of 0.02% (see **Table S2**). Dithionite-extractable Fe (Fe_d) was measured for bulk soil samples using the method of Mehra and Jackson (1960). Oxalate-extractable Fe (Fe_o) and Al (Al_o) were quantified using standard techniques (van Reeuwijk, 1995). The extracts were measured using flame AAS (Perkin Elmer AAnalyst 300).

The BET specific surface area of mineral phases was measured by multi-point N_2 adsorption (Autosorb-1, Quantachrome Instr.). The specific surface area of samples ranged from 0.12 to 2.1 m^{-2} g for sands of the discontinuous permafrost zone and amounted to 25.3 m^{-2} g for the silt/loam horizon of the continuous permafrost zone. The size of soil particles distribution was assessed using a laser diffraction particle size analyzer (Beckman Coulter, USA). To separate the soil particles into their spherical diameter size classes as following: coarse sand (2-0.50 mm), fine sand (0.50-0.05 mm), silt (0.05-0.002 mm) and clay (<0.002 mm) using the USDA groupings.

3. Results

3.1. Field assessment of C stock in peat and underlying mineral horizons

The stock of SOC in peat and mineral horizons at the Khanymey site was estimated from measured C concentrations and bulk density of peat and mineral horizons. We integrated the C pool in mineral horizons located over 100-200 cm under the organic+mineral layer. In the latter, we considered the thickness till the

depth where the C concentration dropped down to 0.1 %. The C pool in the discontinuous permafrost zone over 2 m deep on peat mounds encompassed 40±20 cm of the active (thawed) peat layer, 60±20 cm of the frozen peat and 100 cm of the frozen mineral layer (**Fig. 1**). In sands of discontinuous permafrost zone, the OC content decreased from ca. 3% at the top of mineral horizon (0-20 cm) to 0.5-1.5% in the deeper parts of the core (40-80 cm), **Fig. 2A**. The lateral variability of C stock in the 11 sampled sites shown in **Fig. S1** is illustrated in **Table S1**. The average C stock in the peat and mineral deposit was 46.5±22.7 and 14.6±9.2 kg m^{-2}, respectively. It can be seen from this table that the active layer thickness (ALT) and peat thickness have no straightforward impact on C stock in mineral horizons (correlation coefficients are below -0.3, $p > 0.05$). The OC concentration in the silt/loam under peat of the continuous permafrost zone showed a sizeable maximum (6-10%) at 20-40 cm below the peat deposit (**Fig. 2 B**). Overall, the OC pool in 100 cm frozen mineral horizons is generally lower but not negligible compared to the upper 0-100 cm peat layer (**Fig. 2 C**).

3.2. Leaching of DOC from frozen peat

Kinetics of DOC leaching from frozen peat demonstrated that stable yield of DOC (5.7±0.5 mg C g_{peat}^{-1}) is achieved after 48 h of reaction; however, a 70% level of final DOC (4.6 mg C g_{peat}^{-1}) is reached over first 3 hrs (**Fig. 3A**)**.** The peat-mass normalized DOC concentration (5.7 mg C g_{peat}^{-1} at 1 g_{peat} L^{-1}) was consistent with the value obtained in batch reactors which were used for preparation of DOC stock solution for adsorption experiments (10 g_{peat} L^{-1} yielded 58 mg L^{-1} of DOC). The 2 day interaction of frozen peat with MilliQ water at different concentrations of peat demonstrated a general decrease in DOC with peat concentration increase from 1 to 20 g_{peat} L^{-1} following an increase at 100 g_{peat} L^{-1} (**Fig. 3B**). Overall, for the range of peat concentrations comparable to those of mineral phases in adsorption experiments (i.e., 1-5 g_{solid} L^{-1}; see section 3.3 below), an average value of 5.5±0.5 mg C g_{peat}^{-1} was obtained. The $SUVA_{254}$ also increased from 2.7 to ca. 5.3 L mg C^{-1} m^{-1} over 132 h of reaction (**Fig. 3A**), and decreased with peat concentration from 4.2 to 1.7 (**Fig. 3B**).

3.3. DOM adsorption onto and leaching from mineral horizons

At low concentrations of added DOC (typically < 5 mg L^{-1}), all mineral substrates released C with maximal desorption values equal to 0.49±0.23, 0.13±0.13, and 0.1±0.1 (mean ± s.d., n = 6) mg C g_{soil}^{-1} for Bh, Bs, and E horizons in the discontinuous permafrost zone (**Table 2**). The silt/loam from the continuous permafrost zone (Bgh) exhibited 0.09±0.02 mg C g_{soil}^{-1}. Additional desorption experiments were conducted with the Bh horizon (organo-Al-Fe-rich sand) from the discontinuous permafrost zone, at variable soil to water ratios over 48 h of exposure time. These experiments yielded an average value of released OC equaled to 1.20±0.95 mg C g_{soil}^{-1} (for 0.5 to 100 g L^{-1} of sand, n = 7, **Fig. 4 A**). The $SUVA_{254}$ of DOM released from the Bh horizon was low at 1 g_{sand} L^{-1} (1.2 L mg C^{-1} m^{-1}) but raised to approx. 4.9 L mg C^{-1} m^{-1} at higher concentrations of added mineral substrate (2 to 100 g_{sand} L^{-1}).

The release of DOC from the Bh horizon was also studied in dynamic mode using a column reactor. The breakthrough curves of DOC and $SUVA_{254}$ demonstrated a contrasting pattern (**Fig. 4 B**). The DOC decreased from ca. 260±40 mg L^{-1} to 17±5 mg L^{-1} over the first 0-10 h of reaction and was followed by a further decrease to ca. 8-4 mg L^{-1} after 100-200 h of leaching. Numerical integration of the breakthrough curve allowed assessing the overall amount of DOC released from 3 g of mineral soil, although insufficient temporal resolution at the beginning of reaction yielded sizable uncertainty. The integral value of DOC released over 200 h of reaction was 2.3±0.8 mg C g_{soil}^{-1}, which is somewhat higher than the value obtained in batch (static) reactors (1.2±0.6 mg C g_{soil}^{-1}). An elevated amount of DOC released in dynamic experiments compared to batch reactors could be due to longer (4 x) exposure time of the solid phase in the column-through reactor and enhanced mobilization of internal, strongly bound DOC from poorly accessible pores. Initial $SUVA_{254}$ strongly increased from approx. 0.8 L mg C^{-1} m^{-1} after the first 2 h of reaction to its maximum of 4 L mg C^{-1} m^{-1} after 8 h before gradually decreasing to 1.8 L mg C^{-1} m^{-1} between 10 and 200 h of leaching (**Fig. 4 B**).

3.4. Adsorption parameters

A plot of net adsorbed DOC as a function of added DOC concentration demonstrated linear (R^2 > 0.86, $p < 0.005$) dependences for both adsorption and desorption parts. The linear regression (Eqn. 1) with

R^2 = 0.86 to 1.0 ($p < 0.001$) allowed for determination of DOC adsorption parameters (**Fig. 5**). The *b* and *m* parameters of Eqn. 1 as well as the RSP value are listed in **Table 2**.

The partition coefficient *m* (Eqn. 1) and the maximal reactive soil pool (RSP, mg C g^{-1}) were highest for the Bs horizon (0.61±0.05 and 0.38±0.02, respectively; mean ± s.d., $n = 6$) and lowest for the E horizon (0.17±0.12 and 0.13±0.15, respectively; mean ± s.d., $n = 6$). Note quite high uncertainty on adsorption parameters obtained for the E and Bh horizons of the discontinuous permafrost zone. This uncertainty is mostly due to sizable release of DOC from these mineral samples prior the addition of peat leachate. The maximum amount of adsorbed C ranged from 0.15 to 0.63 mg C g_{soil}^{-1} when 58 mg C L^{-1} were added; however, no saturation of mineral surface for C was observed (**Fig. 5**). In the discontinuous permafrost zone, maximum adsorption capacities were observed for the Al-Fe-rich sand (Bs horizon) with minimum adsorption capacities observed in poor sand and organo-Al-Fe-rich sand (Bh horizon). The adsorption capacity of the silt/loam horizon from the continuous permafrost zone was intermediate between two sand horizons of the discontinuous permafrost zone. Note that the surface-area normalized adsorption was maximal for the poor sand E horizon (0.83±0.42 mg C m^{-2}) of the discontinuous permafrost zone and minimal for the Al-Fe-rich silt/loam (Bgh horizon) of the continuous permafrost zone (0.01±0.0004 mg C m^{-2}). Therefore, after normalization to mineral specific surface area, the order of DOC adsorption on mineral horizons was E > Bs > Bgh.

The evolution of DOC quality which is adsorbed onto/desorbed from mineral horizons was monitored via analysis of the specific ultraviolet absorbance of the aqueous solution. The $SUVA_{254}$ of equilibrium solution exhibited a peculiar pattern depending on the amount of sorbed C (**Fig. 6 A-D**). In the discontinuous permafrost zone, the E horizon exhibited non-systematic variations in $SUVA_{254}$ between 1.5 and 6.0 L mg C^{-1} m^{-1}. The Bs horizon exhibited an increase in $SUVA_{254}$ with an increase in the amount of adsorbed C varying from 1.5 to 5.0 L mg C^{-1} m^{-1}. For the silt/loam Bgh horizon from the continuous permafrost zone, $SUVA_{254}$ slightly increased from 3.3 to 4.4 with an increase in adsorbed C (**Fig. 6 D**).

4. Discussion

4.1. Lateral C stock in mineral vs organic layers of WSL peatlands: comparison with other estimations in Siberia and across the permafrost zones

The stock of C in peat layer of the permafrost peatbogs of the discontinuous permafrost zone obtained in this study (47±23 kg C m^{-2} (mean±s.d.); median of 15 kg C m^{-2}, IQR from 9 to 22 kg C m^{-2} over 1 m depth, **Fig. 2 C**), is consistent with the range of 31 to 60 kg C m^{-2} for this territory (Sheng et al., 2004) and with the value of 44.8 kg C m^{-2} provided by Schepaschenko et al. (2013). The Northern Circumpolar Soil Carbon Database lists a value of 44.3 kg C m^{-2} for the 0-100 cm layer (Hugelius et al., 2013b). However, if we consider the C in mineral soil horizons (100-200 cm depth), the C stock increases by a factor of 1.4-1.5 and reaches 61±21 kg C m^{-2}. Similarly, consideration of mineral horizons in the NE European tundra peatlands increased overall C stock by 30% (Pastukhov and Kaverin, 2013) which is similar to our estimations for western Siberia. Note that lack of correlation between peat thickness and C stock in mineral horizons (according to our data, Table S1, section 3.1) suggests that accumulation of C in mineral layers most likely occurs via lateral influx of C-rich shallow subsurface waters such as supra-permafrost waters (Raudina et al., 2018) rather than via direct vertical downward transfer from the peat to the mineral layer.

The sizeable stock of OC in mineral horizons of frozen soils of WSL **Fig. 2 C)** is consistent with that reported in Central Siberia (Gentsch et al., 2015a), Northern Siberia (Palmtag et al., 2016), East Siberian Arctic (Gentsch et al., 2015b) and lower than in the yedoma of Eastern Siberia (Hugelius et al., 2013b) and Alaska and Canada (Michaelson et al., 1996; Tarnocai et al., 2009). Although the current estimates of OC stock in WSL comprises more than 2 m depth (70.2 Gt C for WSL, Sheng et al. (2004)) and up to 3 m depth (Hugelius et al., 2014, 2013a), the non-negligible stock of C in the underlying mineral horizon may increase this number by approx. 30 to 50% depending on local peat depth, mineralogical character of mineral deposits, and type of permafrost distribution. In this work, we did not aim to estimate C stock in mineral layers below 3 m depth. In a 7-m deep core from the Tazovsky region (continuous permafrost zone, WSL) that was comprised of 2 m of peat, it was demonstrated that the addition of 4 m of a mineral layer to the first 3 m of peat+mineral layer yields an increase in overall SOC stock by only 8.7% (**Fig. S2**). Further, these are silt and clay marine deposits strongly enriched in C compared to the organic-poor sands of the discontinuous permafrost zone. However, given the poor current assessment of C stock in mineral sediments underlying the

WSL peat deposits, and the high vulnerability of WSL permafrost with respect to climate warming (i.e., up to 4 m of frozen soils may be thawed over next 50 to 100 years (Vasiliev et al., 2020)), a detailed field-based assessment of OC stocks in mineral layers of this environmentally important region is clearly needed.

In this work, we focused exclusively on organic C in peat and mineral horizons. Another important nutrient in the permafrost soils is nitrogen (N). Thus, Meyer et al. (2006) demonstrated that in western Siberia, most of N mineralization occurs in mineral soil horizons rather than in overlaying peat. As such, study of organic N adsorption onto and release from frozen mineral layers in permafrost peatlands should be of high priority.

4.2. DOC release from peat and mineral horizons and DOC adsorption capacity

The efficiency of DOC leaching from frozen peat of western Siberia measured in this study (5.5±0.5 mg DOC g_{peat}^{-1}) is very similar to that of the peat from the NE European tundra (5 mg DOC g_{peat}^{-1}, Payandi-Rolland et al., 2020) but substantially higher than the values obtained for organic soil samples from Alaska (0.14-2.2 mg g_{soil}^{-1}; Gao et al., 2018). The DOC leaching efficiencies from organic-poor mineral horizons underlying peat deposits of the WSL (0.1 to 0.5 mg C/$g_{mineral\ soil}$) are comparable with those reported for mineral horizons in Canada (b = 0.05-0.1 mg g^{-1}, Kothawala et al., 2012) and Norway (b = 0.17±0.03 mg g^{-1}, Pengerud et al., 2014).

To assess the capacity of mineral horizons to process DOC from the overlaying peat, we present a histogram of soil mass-normalized adsorption/desorption yields (**Fig. 7 A**). Whereas poor sand exhibits comparable adsorption and desorption capacities, Al-Fe-rich sand of the discontinuous permafrost zone and silt/loam of the continuous permafrost zone can adsorb at least 5 times more C than they can release. In contrast, organo-Al-Fe-rich sand of the discontinuous permafrost zone presents the highest desorption capacities. This demonstrates a strong variability of DOC behavior in frozen mineral horizons that underlay peat deposits in western Siberia, and calls a need for a thorough region- and depth-specific inventory of C surface stock depending on underlying mineral deposit type. For example, in the discontinuous permafrost zone, thawing of peat and underlying mineral layers, notable in hollows and depressions, can saturate the mineral horizons in OC originated from peat and plant litter leaching in surface layers. Later, these horizons

(such as Bh) are capable of releasing the adsorbed DOC, if the hydrological conditions modify the lateral flow leading to DOC export along the permafrost boundary.

At the same time, experiments of this study demonstrate that adsorption of OC on mineral horizons cannot sizably increase SOC stock in these horizons. This increase was only 5 and 15% for the E and Bs horizons of the discontinuous permafrost zone. The Bh horizon exhibited only desorption thus decreasing its stock by 2 %. Overall this indicates the rather stabile non-exchangeable nature of C in these horizons, similar to what is reported for other clay soils (Avneri-Katz et al., 2017). Furthermore, the $SUVA_{254}$ of DOM released from E and Bs horizons was much lower than that of Bh and Bgh horizons (**Fig. 6**). This may reflect more refractory and less labile OM in the Bh and Bgh horizons that contain elevated proportions of clays, Fe, and Al oxides. These compounds are likely to retain aromatic and highly polymerized refractory C (He et al., 2016). Specifically, the long-term persistence of C in subsoil minerals is known to occur due to strong bonds between hydroxylated mineral surfaces and phenolic functional groups of DOM (Kramer et al., 2012).

To assess the factors controlling DOC release from and adsorption onto frozen mineral soils we tested a possible link between magnitude of C release/adsorption and basic physical and chemical properties (including Fe and Al content) of study soils. The latter are known to be responsible for DOM adsorption on non-permafrost soils (Kaiser et al., 1996; Karavanova et al., 2020; Nierop et al., 2002). The amount of C desorbed from mineral horizon and the RSP positively correlated ($p < 0.05$) with Al_0, N and C ($r = 0.93$, 0.92, 0.91, respectively). Maximal adsorption capacity positively correlated ($p < 0.05$) with Fe_d and Fe_o ($r =$ 0.75, 0.68, respectively). Such correlations are consistent with recent observations on strong interactions between iron and organic carbon in European permafrost peatlands (Patzner et al., 2020). At the same time, sizable desorption of DOC from the Bh horizon and high concentration of oxalate-extractable Al for this soil suggest that the majority of Al hydroxides sites are already saturated with organic matter. This strongly corroborates a recent result of Fukumasu et al. (2021) that Al-bearing reactive mineral phases are more important than clay content and Fe-bearing reactive mineral phases for SOC stabilization in topsoils of humid continental climates.

It is important to note that the leaching of DOC from the frozen peat and DOC adsorption onto mineral surfaces occur quite fast, within hours to the first day (**Fig. 3A**); a stable DOC concentration in contact with peat is attained within 1-2 days. This is consistent with the generally high reactivity of peat in aqueous solutions (Payandi-Rolland et al., 2020). Similarly, the adsorption pseudo-equilibrium was typically achieved within the first several hours of reaction. This suggests highly labile behavior of OC within the peat-mineral soil profile. As a result, the vertical and lateral water travel rate (from 8 до 80 cm per day, Loiko et al., 2019) rather than chemical reactions of leaching and adsorption will be the limiting step in DOC distribution within the soil profile. Fast reaction of DOC in laboratory experiments of this study further demonstrates that the response of frozen peat and mineral horizons to permafrost thawing will be quasi instantaneous, and that the fast leaching/adsorption reactions may prevent the modification of OC fluxes within the soils- surface waters continuum. It has been shown from incubation experiments that sizeable biodegradation and photo-degradation of DOC in soil and surface waters occurs within 7-40 days (i.e., Oleinikova et al., 2017b, 2017a; Shirokova et al., 2019; Vonk et al., 2015). In this regard, the temporal variability of DOC sources (leaching from peat or mineral horizons) and sinks (sorption to minerals) will be more important than the autochthonous transformation processes. At the same time, a part of mineral-fixed carbon, especially the labile fraction of DOM, can be decomposed by microorganisms over the time scale of days to month (Panneer Selvam et al., 2017). From the one hand, this may diminish the efficiency of long-term retention of the leached carbon by mineral layers, given quite long (thousands of years) residence time of OM in Siberian soils below 30-50 cm depth (Goryachkin et al., 2000) and very slow rate of DOC movement downward the peat profile (0.047 cm y^{-1}, Zarov et al., 2021). From the other hand, certain mineral horizons, such as clay/silt, are known to stabilize organic matter in soil (Six et al., 2002). Despite the fact that clay/silt soil had a lower capacity to C adsorption compared to Al- and Fe-rich soil (**Fig. 7 A**), this may enhance overall retention ability of C in the soil profile due to lower decomposition rate of OM in the Bgh horizon.

4.3. The impact of C mobility within peat-mineral soil profile at the watershed scale

Climate change in western Siberia is likely to cause a shift of the permafrost boundary northward and an increase in active layer thickness (Anisimov et al., 2013; Frey and Smith, 2003; Pavlov and Moskalenko,

2002; Romanovsky et al., 2010; Vasiliev et al., 2011). According to available scenarios of permafrost thawing in the continuous zone, ALT may reach 1 to 4 m in the western Siberia (Anisimov and Zimov, 2020; Anisimov and Kokorev, 2017) and will inevitably involve currently-frozen mineral horizons into OC release and adsorption. However, these scenarios have to take into account the difference in thawing of micro-landscapes of permafrost peatlands. These include positive (mounds in the discontinuous permafrost zone and polygons in the continuous zone) and negative (hollows/depressions in the discontinuous and cracks in the continuous zone) forms of relief. In the discontinuous permafrost zone, the ALT rarely exceeds 40-50 cm at the mounds and is entirely located within the peat layer. In contrast, the hollows (depressions) are thawed down to the mineral horizons and thus can represent to some extent, an extreme case of the permafrost thawing at mounds in the discontinuous permafrost zone. Based on experimentally-assessed C desorption from mineral horizons, we calculate that thawing of 1 m of frozen sand in the discontinuous permafrost zone may release between 0.05 and 0.5 mg C/g_{soil} m^{-2} (**Fig. 7 B**) whereas thawing of frozen silt/loam in the continuous permafrost zone will yield 0.1 mg C/g_{soil} m^{-2}. Further, the amount of OC originated from leaching of peat that can be adsorbed by unfrozen sand horizons of 1 m thick is between 0.3 and 0.4 kg C m^{-2} (**Fig. 7 B**). This is likely to occur in many sites of the discontinuous permafrost zone, where the active layer often extends into the mineral horizon, although it is dependent on the degree of sand coverage by Fe and Al hydroxides. Thus, at the Khanymey site, a transport of OC with seepage into the mineral horizons and the following adsorption onto Al hydroxides of the Bh layer can form organo-mineral spodic horizons.

The values of OC adsorption capacity assessed in this study are non-negligible compared to the amount of DOC that can be leached from peat column by downward migrating fluids (1.38±0.13 kg C m^{-2}). This suggests a sizable control by mineral horizons on C migration across the peat - mineral soil profile in the permafrost region. In particular, in the discontinuous permafrost zone, especially at the hollows and depressions, this control is currently visible via release rather than accumulation of DOC in the already saturated Bh horizon. In the continuous permafrost zone, the manifestation of DOC adsorption on mineral layers is anticipated to occur within the next decades.

On average, out of 1.38±0.13 kg C m^{-2} capable of initial release from the upper 0-100 cm of peat and migrating downwards, 0.25±0.1 kg C m^{-2} can be adsorbed in the 100-200 cm underlying Fe- and Al-rich sand and silt with the remaining 1.13 kg C m^{-2} to be exported to lakes and rivers (**Fig. 7 C**). Assuming that, in the continuous permafrost zone, all 200±100 cm of frozen peat and underlying sand/silt will thaw over next 100 years, this yields an export potential of 6±3 t C km^{-2} y^{-1} which is higher than the current estimation of annual DOC export by peatland-draining Siberian rivers (around 3 t C km^{-2} y^{-1}, Pokrovsky et al., 2020). The results of this work demonstrate that, in the continuous permafrost zone, on a short-term scale, sizeable amounts of DOC can be liberated to soil waters due to frozen peat thawing and will be quickly captured by underlying mineral horizons if thawing progresses according to the proposed scenarios. In the discontinuous permafrost zone, this OC retention by mineral (Fe- and Al-rich sand horizons) is already occurring in depressions and hollows. Overall, the DOC retention via adsorption can greatly mitigate the "negative "impact of OC release from frozen peat and should be taken into account for modeling of OC transport between soils and rivers in permafrost-bearing landscapes, especially in the continuous permafrost zone where the peat is underlayed by clay/silt horizons, capable of strong DOC adsorption.

Conclusions

We revealed a sizeable stock of organic carbon in mineral horizons underlaying peat in the discontinuous and continuous permafrost zone of the western Siberian Lowland (WSL). Significant contributions to the C pool in mineral layer are provided by organo-Al-Fe-rich Bh horizons. These mineral horizons can adsorb part of the C released from peat. The Fe-rich sand of the discontinuous permafrost zone did not exhibit any DOC adsorption but instead released sizable amount of DOC. This horizon contained the highest amount of oxalate extractable Al which could act as strong adsorbent of DOC in the form of Al oxy(hydr)oxides.

The on-going climate warming and permafrost thaw will produce a deepening of the active layer into mineral horizons. Such thawing will liberate some amount of DOM from peat and also release of DOM from mineral horizons. However, a sizeable amount of DOC may be retained on mineral horizons via adsorption. This can mitigate the negative impact of climate warming on permafrost thaw and DOC release and thus

attenuate the positive feedback of permafrost thaw on CO_2 emissions. Further measurements of the C pool and adsorption capacity of mineral horizons under frozen peatlands of western Siberia, down to 5-10 m under variable micro-landscapes and biomes (perhaps most importantly under forests), are necessary for improved regional and global modeling of C cycle in permafrost peatlands.

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References

Anisimov, O., Kokorev, V., Zhil'tsova, Y., 2013. Temporal and spatial patterns of modern climatic warming: Case study of Northern Eurasia. Clim. Change 118, 871–883. https://doi.org/10.1007/s10584-013-0697-4

Anisimov, O., Zimov, S., 2020. Thawing permafrost and methane emission in Siberia: Synthesis of observations, reanalysis, and predictive modeling. Ambio 1–10. https://doi.org/10.1007/s13280-020-01392-y

Anisimov, O.A., Kokorev, V.A., 2017. Russian permafrost in the 21st century: model-based projections and analysis of uncertainties. Earth`s Cryosph. XXI, 3–9. https://doi.org/10.21782/ec2541-9994-2017-1(3-9)

Avneri-Katz, S., Young, R.B., McKenna, A.M., Chen, H., Corilo, Y.E., Polubesova, T., Borch, T., Chefetz, B., 2017. Adsorptive fractionation of dissolved organic matter (DOM) by mineral soil: Macroscale approach and molecular insight. Org. Geochem. 103, 113–124. https://doi.org/10.1016/j.orggeochem.2016.11.004

Buurman, P., 1985. Carbon/sesquioxide ratios in organic complexes and the transition albic-spodic horizon. J. Soil Sci. 36, 255–260. https://doi.org/10.1111/j.1365-2389.1985.tb00329.x

Čapek, P., Diáková, K., Dickopp, J.-E., Bárta, J., Wild, B., Schnecker, J., Alves, R.J.E., Aiglsdorfer, S., Guggenberger, G., Gentsch, N., Hugelius, G., Lashchinsky, N., Gittel, A., Schleper, C., Mikutta, R., Palmtag, J., Shibistova, O., Urich, ., Richter, A., Šantrůčková, H., 2015. The effect of warming on the vulnerability of subducted organic carbon in arctic soils. Soil Biol. Biochem. 90, 19-29. DOI: 10.1016/j.soilbio.2015.07.013.

Ciais, P., Tan, J., Wang, X., Roedenbeck, C., Chevallier, F., Piao, S.L., Moriarty, R., Broquet, G., Le Quéré, C., Canadell, J.G., Peng, S., Poulter, B., Liu, Z., Tans, P., 2019. Five decades of northern land carbon uptake revealed by the interhemispheric CO_2 gradient. Nature. https://doi.org/10.1038/s41586-019-1078-6

Dahlgren, R.A., Marrett, D.J., 1991. Organic carbon sorption in arctic and subalpine Spodosol B horizons. Soil Sci. Soc. Am. J. 55, 1382–1390. https://doi.org/10.2136/sssaj1991.03615995005500050030x

Feng, X., Vonk, J.E., Van Dongen, B.E., Gustafsson, Ö., Semiletov, I.P., Dudarev, O. V., Wang, Z.,

Montluçon, D.B., Wacker, L., Eglinton, T.I., 2013. Differential mobilization of terrestrial carbon pools in Eurasian Arctic river basins. Proc. Natl. Acad. Sci. U.S.A. 110, 14168–14173. https://doi.org/10.1073/pnas.1307031110

Frey, K.E., McClelland, J.W., 2009. Impacts of permafrost degradation on arctic river biogeochemistry. Hydrol. Process. 23, 169–182. https://doi.org/10.1002/hyp.7196

Frey, K.E., Smith, L.C., 2003. Recent temperature and precipitation increases in West Siberia and their association with the Arctic Oscillation. Polar Res. 22, 287–300. https://doi.org/10.1111/j.1751-8369.2003.tb00113.x

Fukumasu, J.,Poeplau, C.,Coucheney, E., Jarvis, N., Kloffel, T., Koestel, J., Katterer, T., Svensson, D.N., Wetterlind, J., Larsbo, M., 2021. Oxalate-extractable aluminum alongside carbon inputs may be a major determinant for organic carbon content in agricultural topsoils in humid continental climate. Geoderma, 402, Art No 115345.

Gao, L., Zhou, Z., Reyes, A. V., Guo, L., 2018. Yields and characterization of dissolved organic matter from different aged soils in Northern Alaska. J. Geophys. Res. Biogeosciences 123, 2035–2052. https://doi.org/10.1029/2018JG004408

Gentsch, N., Mikutta, R., Alves, R.J.E., Barta, J., Čapek, P., Gittel, A., Hugelius, G., Kuhry, P., Lashchinskiy, N., Palmtag, J., Richter, A., Šantrůčková, H., Schnecker, J., Shibistova, O., Urich, T., Wild, B., Guggenberger, G., 2015a. Storage and transformation of organic matter fractions in cryoturbated permafrost soils across the Siberian Arctic. Biogeosciences 12, 4525–4542. https://doi.org/10.5194/bg-12-4525-2015.

Gentsch, N., Mikutta, R., Shibistova, O., Wild, B., Schnecker, J., Richter, A., Urich, T., Gittel, A., Šantrůčková, H., Bárta, J., Lashchinskiy, N., Mueller, C.W., Fuß, R., Guggenberger, G., 2015b. Properties and bioavailability of particulate and mineral-associated organic matter in Arctic permafrost soils, Lower Kolyma Region, Russia. European J. Soil Sci., 66, 722-734. DOI: 10.1111/ejss.12269.

Groeneveld, M., Catalán, N., Attermeyer, K., Hawkes, J., Einarsdóttir, K., Kothawala, D., Bergquist, J., Tranvik, L., 2020. Selective adsorption of terrestrial dissolved organic matter to inorganic surfaces along a boreal inland water continuum. J. Geophys. Res. Biogeosciences 125, e2019JG005236. https://doi.org/10.1029/2019JG005236

Goryachkin S.V., Cherkinsky A.E., Chichagova O.A., 2000. The soil organic carbon dynamics in high latitudes of Eurasia using ^{14}C data and the impact of potential climate change. In: R. Lai, J.M. Kimble, B.A. Stewart (eds). Global Climate and Cold Regions Ecosystems. Lewis Publishers, p. 145-161.

Gu, B., Schmitt, J., Chen, Z., Liang, L., McCarthy, J.F., 1994. Adsorption and desorption of natural organic matter on iron oxide: mechanisms and models. Environ. Sci. Technol. 28, 38–46. https://doi.org/10.1021/es00050a007

He, W., Chen, M., Schlautman, M.A., Hur, J., 2016. Dynamic exchanges between DOM and POM pools in coastal and inland aquatic ecosystems: A review. Sci. Total Environ. https://doi.org/10.1016/j.scitotenv.2016.02.031

Hugelius, G., Bockheim, J.G., Camill, P., Elberling, B., Grosse, G., Harden, J.W., Johnson, K., Jorgenson, T., Koven, C.D., Kuhry, P., Michaelson, G., Mishra, U., Palmtag, J., O'Donnell, J., Schirrmeister, L., Schuur, E.A.G., Sheng, Y., Smith, L.C., Strauss, J., Yu, Z., 2013a. A new data set for estimating organic carbon storage to 3 m depth in soils of the northern circumpolar permafrost region. Earth Syst. Sci. Data 5, 393–402. https://doi.org/10.5194/essd-5-393-2013

Hugelius, G., Tarnocai, C., Broll, G., Canadell, J.G., Kuhry, P., Swanson, D.K., 2013b. The northern circumpolar soil carbon database: Spatially distributed datasets of soil coverage and soil carbon storage

564 in the northern permafrost regions. Earth Syst. Sci. Data 5, 3–13. https://doi.org/10.5194/essd-5-3-2013

565 Hugelius, G., Strauss, J., Zubrzycki, S., Harden, J.W., Schuur, E.A.G., Ping, C.L., Schirrmeister, L., Grosse,
566 G., Michaelson, G.J., Koven, C.D., O'Donnell, J.A., Elberling, B., Mishra, U., Camill, P., Yu, Z.,
567 Palmtag, J., Kuhry, P., 2014. Estimated stocks of circumpolar permafrost carbon with quantified
568 uncertainty ranges and identified data gaps. Biogeosciences 11, 6573–6593. https://doi.org/10.5194/bg-
569 11-6573-2014

570 Hugelius, G., Loisel, J., Chadburn, S., Jackson, R.B., Jones, M., MacDonald, G., Marushchak, M., Olefeldt,
571 D., Packalen, M., Siewert, M.B., Treat, C., Turetsky, M., Voigt, C., Yu, Z., 2020. Large stocks of
572 peatland carbon and nitrogen are vulnerable to permafrost thaw. Proc. Natl. Acad. Sci. U. S. A. 117,
573 20438–20446. https://doi.org/10.1073/pnas.1916387117

574 IUSS, 2015. International soil classification system for naming soils and creating legends for soil maps,
575 Update 2015. Rome, Italy.

576 Jantze, E.J., Lyon, S.W., Destouni, G., 2013. Subsurface release and transport of dissolved carbon in a
577 discontinuous permafrost region. Hydrol. Earth Syst. Sci. 17, 3827–3839. https://doi.org/10.5194/hess-
578 17-3827-2013

579 Jardine, P.M., McCarthy, J.F., Weber, N.L., 1989. Mechanisms of dissolved organic carbon adsorption on
580 soil. Soil Sci. Soc; Amer. J. 53(5), 1378-1385,
581 https://doi.org/10.2136/sssaj1989.03615995005300050013x

582 Kaiser, K., 1998. Fractionation of dissolved organic matter affected by polyvalent metal cations. Org.
583 Geochem. 28, 849–854. https://doi.org/10.1016/S0146-6380(98)00046-1

584 Kaiser, K., Guggenberger, G., 2000. The role of DOM sorption to mineral surfaces in the preservation of
585 organic matter in soils, in: Organic Geochemistry. pp. 711–725. https://doi.org/10.1016/S0146-
586 6380(00)00046-2

587 Kaiser, K., Guggenberger, G., Zech, W., 1996. Sorption of DOM and DOM fractions to forest soils.
588 Geoderma 74, 281–303. https://doi.org/10.1016/S0016-7061(96)00071-7

589 Kaiser, K., Zech, W., 1998. Soil dissolved organic matter sorption as influenced by organic and sesquioxide
590 coatings and sorbed sulfate. Soil Sci. Soc. Am. J. 62, 129–136.
591 https://doi.org/10.2136/sssaj1998.03615995006200010017x

592 Kalbitz, K., Schwesig, D., Rethemeyer, J., Matzner, E., 2005. Stabilization of dissolved organic matter by
593 sorption to the mineral soil. Soil Biol. Biochem. 37, 1319–1331.
594 https://doi.org/10.1016/j.soilbio.2004.11.028

595 Karavanova, E.I., Zolovkina, D.F., Stepanov, A.A., 2020. Interaction of the Water-Soluble Organic
596 Substances of Coniferous Litter with Minerals and Horizons of Podzolic Soil and Podzols. Eurasian
597 Soil Sci. 53, 1234–1246. https://doi.org/10.1134/S1064229320090070

598 Kawahigashi, M., Kaiser, K., Kalbitz, K., Rodionov, A., Guggenberger, G., 2004. Dissolved organic matter
599 in small streams along a gradient from discontinuous to continuous permafrost. Glob. Change Biol. 10,
600 1576–1586. https://doi.org/10.1111/j.1365-2486.2004.00827.x

601 Kawahigashi, M., Kaiser, K., Rodionov, A., Guggenberger, G., 2006. Sorption of dissolved organic matter
602 by mineral soils of the Siberian forest tundra. Glob. Change Biol. 12, 1868-1877.

603 Kleber, M., Mikutta, R., Torn, M.S., Jahn, R., 2005. Poorly crystalline mineral phases protect organic matter
604 in acid subsoil horizons. Eur. J. Soil Sci. 56(6), 717-725. https://doi.org/10.1111/j.1365-
605 2389.2005.00706.x

606 Kothawala, D.N., Moore, T.R., Hendershot, W.H., 2009a. Soil properties controlling the adsorption of

dissolved organic carbon to mineral soils. Soil Sci. Soc. Am. J. 73, 1831–1842. https://doi.org/10.2136/sssaj2008.0254

Kothawala, D.N., Moore, T.R., Hendershot, W.H., 2008. Adsorption of dissolved organic carbon to mineral soils: A comparison of four isotherm approaches. Geoderma 148, 43–50. https://doi.org/10.1016/j.geoderma.2008.09.004

Kothawala, D.N., Roehm, C., Blodau, C., Moore, T.R., 2012. Selective adsorption of dissolved organic matter to mineral soils. Geoderma 189–190, 334–342. https://doi.org/10.1016/j.geoderma.2012.07.001

Kramer, M.G., Chadwick, O.A., 2018. Climate-driven thresholds in reactive mineral retention of soil carbon at the global scale. Nat. Clim. Chang. 8, 1104–1108. https://doi.org/10.1038/s41558-018-0341-4

Kramer, M.G., Sanderman, J., Chadwick, O.A., Chorover, J., Vitousek, P.M., 2012. Long-term carbon storage through retention of dissolved aromatic acids by reactive particles in soil. Glob. Chang. Biol. 18, 2594–2605. https://doi.org/10.1111/j.1365-2486.2012.02681.x

Kremenetski, K. V., Velichko, A.A., Borisova, O.K., MacDonald, G.M., Smith, L.C., Frey, K.E., Orlova, L.A., 2003. Peatlands of the Western Siberian lowlands: Current knowledge on zonation, carbon content and Late Quaternary history. Quat. Sci. Rev. 22, 703–723. https://doi.org/10.1016/S0277-3791(02)00196-8

Loiko, S.V., Raudina, T.V., Lim, A.G., Kuzmina, D.M., Kulizhskiy, S.P., Pokrovsky, O.S., 2019. Microtopography controls of carbon and related elements distribution in the West Siberian frozen bogs. Geosciences 9, 291. https://doi.org/10.3390/geosciences9070291

Mehra, O.P., Jackson, M.L., 1960. Iron oxide removal from soils and clay by a dithionite-citrate system buffered with sodium bicarbonate, in: Swineford, A. (Ed.), Proceedings of the Seventh National Conference on Clays and Clay Minerals. pp. 317–327. https://doi.org/10.1016/C2013-0-01644-1

Meyer, H., Kaiser, C., Biasi, C., Hämmerle, R., Rusalimova, O., Lashchinsky, N., Baranyi, C., Daims, H., Barsukov, P., Richter, A., 2006. Soil carbon and nitrogen dynamics along a latitudinal transect in Western Siberia, Russia. Biogeochemistry 81(2), 239-252.

Michaelson, G.J., Ping, C.L., Kimble, J.M., 1996. Carbon storage and distribution in tundra soils of Arctic. Arct. Alp. Res. 28, 414–424. https://doi.org/10.1080/00040851.1996.12003194

Moore, T.R., de Souza, W., Koprivnjak, J.-F., 1992. Controls on the sorption of dissolved organic carbon by soils. Soil Sci. 154.

Morgalev, Y.N., Lushchaeva, I. V., Morgaleva, T.G., Kolesnichenko, L.G., Loiko, S. V., Krickov, I. V., Lim, A., Raudina, T. V., Volkova, I.I., Shirokova, L.S., Morgalev, S.Y., Vorobyev, S.N., Kirpotin, S.N., Pokrovsky, O.S., 2017. Bacteria primarily metabolize at the active layer/permafrost border in the peat core from a permafrost region in western Siberia. Polar Biol. 40, 1645–1659. https://doi.org/10.1007/s00300-017-2088-1.

Mu, C.C., Zhang, T.J., Zhao, Q., Guo, H., Zhong, W., Su, H., Wu, Q.B., 2016. Soil organic carbon stabilization by iron in permafrost regions of the Qinghai-Tibet Plateau, Geophys. Res. Lett., 43, 10,286– 10,294, doi:10.1002/2016GL070071.

Nierop, K.G.J., Jansen, B., Verstraten, J.M., 2002. Dissolved organic matter, aluminium and iron interactions: Precipitation induced by metal/carbon ratio, pH and competition. Sci. Total Environ. 300, 201–211. https://doi.org/10.1016/S0048-9697(02)00254-1.

Nodvin, S., Driscoll, C., Likens, G., 1986. Simple partitioning of anions and dissolved organic carbon in a forest soil. Soil Sci. 142, 27–35.

Oleinikova, O. V., Drozdova, O.Y., Lapitskiy, S.A., Demin, V. V., Bychkov, A.Y., Pokrovsky, O.S., 2017a.

Dissolved organic matter degradation by sunlight coagulates organo-mineral colloids and produces low-molecular weight fraction of metals in boreal humic waters. Geochim. Cosmochim. Acta 211, 97–114. https://doi.org/10.1016/j.gca.2017.05.023

Oleinikova, O. V., Shirokova, L.S., Gérard, E., Drozdova, O.Y., Lapitskiy, S.A., Bychkov, A.Y., Pokrovsky, O.S., 2017b. Transformation of organo-ferric peat colloids by a heterotrophic bacterium. Geochim. Cosmochim. Acta 205, 313–330. https://doi.org/10.1016/j.gca.2017.02.029

Oosterwoud, M. R., Temminghoff, E. J. M., and van der Zee, S. E. A. T. M., 2010. Quantification of DOC concentrations in relation with soil properties of soils in tundra and taiga of Northern European Russia. Biogeosciences Discuss., 7, 3189–3226, https://doi.org/10.5194/bgd-7-3189-2010.

Opfergelt, S., 2020. The next generation of climate model should account for the evolution of mineral-organic interactions with permafrost thaw. Environ. Res. Lett. 15, Art No 091003. https://doi.org/10.1088/1748-9326/ab9a6d

Palmtag, J., Ramage, J., Hugelius, G., Gentsch, N., Lashchinskiy, N., Richter, A., Kuhry, P., 2016. Controls on the storage of organic carbon in permafrost soil in northern Siberia. European J. Soil Sci., 67 (4), 478-491. DOI: 10.1111/ejss.12357.

Panneer Selvam, B., Lapierre, JF., Guillemette, F., Voigt, C., Lamprecht, R.E., Biasi, C., Christensen, T.R., Martikainen, P.J., Berggren, M., 2017. Degradation potentials of dissolved organic carbon (DOC) from thawed permafrost peat. Sci. Rep. 7, Art No 45811. https://doi.org/10.1038/srep45811.

Pastukhov, A. V., Kaverin, D.A., 2013. Soil carbon pools in tundra and taiga ecosystems of northeastern Europe. Eurasian Soil Sci. 46, 958–967. https://doi.org/10.1134/S1064229313070077

Patzner, M.S., Mueller, C.W., Malusova, M., Baur, M., Nikeleit, V., Scholten, T., Hoeschen, C., Byrne, J.M., Borch, T., Kappler, A., Bryce, C., 2020. Iron mineral dissolution releases iron and associated organic carbon during permafrost thaw. Nat. Commun. 11, 1–11. https://doi.org/10.1038/s41467-020-20102-6

Pavlov, A. V., Moskalenko, N.G., 2002. The thermal regime of soils in the north of Western Siberia. Permafr. Periglac. Process. 13, 43–51. https://doi.org/10.1002/ppp.409

Payandi-Rolland, D., Shirokova, L.S., Nakhle, P., Tesfa, M., Abdou, A., Causserand, C., Lartiges, B., Rols, J.L., Guérin, F., Bénézeth, P., Pokrovsky, O.S., 2020. Aerobic release and biodegradation of dissolved organic matter from frozen peat: Effects of temperature and heterotrophic bacteria. Chem. Geol. 536, 119448. https://doi.org/10.1016/j.chemgeo.2019.119448

Pengerud, A., Johnsen, L.K., Mulder, J., Strand, L.T., 2014. Potential adsorption of dissolved organic matter in poorly podzolised, high-latitude soils. Geoderma 226–227, 39–46. https://doi.org/10.1016/j.geoderma.2014.02.027

Pokrovsky, O.S., Manasypov, R.M., Kopysov, S.G., Krickov, I.V., Shirokova, L.S., Loiko, S.V., Lim, A.G., Kolesnichenko, L.G., Vorobyev, S.N., Kirpotin, S.N., 2020. Impact of permafrost thaw and climate warming on riverine export fluxes of carbon, nutrients and metals in Western Siberia. Water (Switzerland) 12. https://doi.org/10.3390/w12061817

Pokrovsky, O.S., Manasypov, R.M., Loiko, S., Shirokova, L.S., Krickov, I.A., Pokrovsky, B.G., Kolesnichenko, L.G., Kopysov, S.G., Zemtzov, V.A., Kulizhsky, S.P., Vorobyev, S.N., Kirpotin, S.N., 2015. Permafrost coverage, watershed area and season control of dissolved carbon and major elements in western Siberian rivers. Biogeosciences 12, 6301–6320. https://doi.org/10.5194/bg-12-6301-2015

Porras, R.C., Hicks Pries, C.E., McFarlane, K.J., Hanson, P.J., Torn, M.S., 2017. Association with pedogenic iron and aluminum: effects on soil organic carbon storage and stability in four temperate forest soils. Biogeochemistry 133, 333–345. https://doi.org/10.1007/s10533-017-0337-6

Prater, I., Zubrzycki, S., Buegger, F., Zoor-Füllgraff, L. C., Angst, G., Dannenmann, M., Mueller, C. W. 2020. From fibrous plant residues to mineral-associated organic carbon - the fate of organic matter in Arctic permafrost soils. Biogeosciences, 17, 3367-3383, https://doi.org/10.5194/bg-17-3367-2020.

Prokushkin, A., Gleixner, G., McDowell, W., Ruehlow, S., Schulze, E.D., 2007. Source- and substrate-specific export of dissolved organic matter from permafrost-dominated forested watershed in central Siberia. Global Biogeochemical Cycles, 21(4), 10.1029/2007GB002938.

Raudina, T. V., Loiko, S. V., Lim, A., Manasypov, R.M., Shirokova, L.S., Istigechev, G.I., Kuzmina, D.M., Kulizhsky, S.P., Vorobyev, S.N., Pokrovsky, O.S., 2018. Permafrost thaw and climate warming may decrease the CO_2, carbon, and metal concentration in peat soil waters of the Western Siberia Lowland. Sci. Total Environ. 634, 1004–1023. https://doi.org/10.1016/j.scitotenv.2018.04.059

Raudina, T. V., Loiko, S. V., Lim, A.G., Krickov, I. V., Shirokova, L.S., Istigechev, G.I., Kuzmina, D.M., Kulizhsky, S.P., Vorobyev, S.N., Pokrovsky, O.S., 2017. Dissolved organic carbon and major and trace elements in peat porewater of sporadic, discontinuous, and continuous permafrost zones of western Siberia. Biogeosciences 14, 3561–3584. https://doi.org/10.5194/bg-14-3561-2017

Romanovsky, V.E., Drozdov, D.S., Oberman, N.G., Malkova, G. V., Kholodov, A.L., Marchenko, S.S., Moskalenko, N.G., Sergeev, D.O., Ukraintseva, N.G., Abramov, A.A., Gilichinsky, D.A., Vasiliev, A.A., 2010. Thermal state of permafrost in Russia. Permafr. Periglac. Process. 21, 136–155. https://doi.org/10.1002/ppp.683

Romanovsky, V.E., Kholodov, A.L., Marchenko, S.S., Oberman, N.G., Drozdov, D.S., Malkova, G. V, Moskalenko, N.G., Vasiliev, A.A., Sergeev, D.O., Zheleznyak, M.N., n.d. Thermal State and Fate of Permafrost in Russia: First Results of IPY.

Sanderman, J., Kramer, M.G., 2017. Dissolved organic matter retention in volcanic soils with contrasting mineralogy: a column sorption experiment. Biogeochemistry 135, 293–306. https://doi.org/10.1007/s10533-017-0374-1

Schepaschenko, D.G., Mukhortova, L. V., Shvidenko, A.Z., Vedrova, E.F., 2013. The pool of organic carbon in the soils of Russia. Eurasian Soil Sci. 46, 107–116. https://doi.org/10.1134/S1064229313020129

Schuur, E.A.G., Bockheim, J., Canadell, J.G., Euskirchen, E., Field, C.B., Goryachkin, S. V., Hagemann, S., Kuhry, P., Lafleur, P.M., Lee, H., Mazhitova, G., Nelson, F.E., Rinke, A., Romanovsky, V.E., Shiklomanov, N., Tarnocai, C., Venevsky, S., Vogel, J.G., Zimov, S.A., 2008. Vulnerability of permafrost carbon to climate change: Implications for the global carbon cycle. Bioscience 58, 701–714. https://doi.org/10.1641/b580807

Schuur, E.A.G., Vogel, J.G., Crummer, K.G., Lee, H., Sickman, J.O., Osterkamp, T.E., 2009. The effect of permafrost thaw on old carbon release and net carbon exchange from tundra. Nature 459, 556–559. https://doi.org/10.1038/nature08031

Serikova, S., Pokrovsky, O.S., Ala-Aho, P., Kazantsev, V., Kirpotin, S.N., Kopysov, S.G., Krickov, I. V., Laudon, H., Manasypov, R.M., Shirokova, L.S., Soulsby, C., Tetzlaff, D., Karlsson, J., 2018. High riverine CO_2 emissions at the permafrost boundary of Western Siberia. Nat. Geosci. 11, 825–829. https://doi.org/10.1038/s41561-018-0218-1

Serikova, S., Pokrovsky, O.S., Laudon, H., Krickov, I. V., Lim, A.G., Manasypov, R.M., Karlsson, J., 2019. High carbon emissions from thermokarst lakes of Western Siberia. Nat. Commun. 10, Art No 1552, https://doi.org/10.1038/s41467-019-09592-1.

Sheng, Y., Smith, L.C., MacDonald, G.M., Kremenetski, K. V., Frey, K.E., Velichko, A.A., Lee, M., Beilman, D.W., Dubinin, P., 2004. A high-resolution GIS-based inventory of the west Siberian peat carbon pool. Global Biogeochem. Cycles 18. https://doi.org/10.1029/2003GB002190.

Shirokova, L.S., Chupakov, A., Zabelina, S., Neverova, N., Payandi-Rolland, D., Causserand, C., Karlsson, J., Pokrovsky, O.S., 2019. Humic surface waters of frozen peat bogs (permafrost zone) are highly resistant to bio- and photodegradation. Biogeosciences 16, 2511–2526. https://doi.org/10.5194/bg-16-2511-2019

Schnecker, J., Wild, B., Takriti, M., Eloy Alves, R.J., Gentsch, N., Gittel, A., Hofer, A., Klaus, K., Knoltsch, A., Lashchinskiy, N., Mikutta, R., Richter, A., 2015. Microbial community composition shapes enzyme patterns in topsoil and subsoil horizons along a latitudinal transect in Western Siberia. Soil Biol. Biochem. 83, 106-115. DOI: 10.1016/j.soilbio.2015.01.016.

Six, J., Conant, R.T., Paul, E., Paustian, K., 2002. Stabilization mechanisms of soil organic matter: Implications for C-saturation of soils. Plant and Soil 241(2), 155-176, DOI:10.1023/A:1016125726789

Smedberg, E., Mörth, C.-M., Swaney, D.P., Humborg, C., 2006. Modeling hydrology and silicon-carbon interactions in taiga and tundra biomes from a landscape perspective: Implications for global warming feedbacks. Global Biogeochem. Cycles 20(2), https://doi.org/10.1029/2005GB002567

Smith, L.C., MacDonald, G.M., Velichko, A.A., Beilman, D.W., Borisova, O.K., Frey, K.E., Kremenetski, K. V., Sheng, Y., 2004. Siberian Peatlands a Net Carbon Sink and Global Methane Source since the Early Holocene. Science 303(5656), 353–356. https://doi.org/10.1126/science.1090553.

Tang, J., Yurova, A.Y., Schurgers, G., Miller, P.A., Olin, S., Smith, B., Siewert, M.B., Olefeldt, D., Pilesjö, P., Poska, A., 2018. Drivers of dissolved organic carbon export in a subarctic catchment: Importance of microbial decomposition, sorption-desorption, peatland and lateral flow. Sci. Total Environ. 622–623, 260–274. https://doi.org/10.1016/j.scitotenv.2017.11.252

Tarnocai, C., Canadell, J.G., Schuur, E.A.G., Kuhry, P., Mazhitova, G., Zimov, S., 2009. Soil organic carbon pools in the northern circumpolar permafrost region. Global Biogeochem. Cycles 23, 1–11. https://doi.org/10.1029/2008GB003327

Ussiri, D.A.N., Johnson, C.E., 2004. Sorption of organic carbon fractions by spodosol mineral horizons. Soil Sci. Soc. Am. J. 68, 253–262. https://doi.org/10.2136/sssaj2004.2530

van Reeuwijk, L.P., 1995. Procedures for soil analysis, Rep. 9, International Soil Reference and Information Centre. Wageningen, The Netherlands.

Vance, G.F., David, M.B., 1989. Effect of acid treatment on dissolved organic carbon retention by a spodic horizon. Soil Sci. Soc. Am. J. 53, 1242–1247. https://doi.org/10.2136/sssaj1989.03615995005300040042x

Vasiliev, A.A., Drozdov, D.S., Gravis, A.G., Malkova, G. V, Nyland, K.E., Streletskiy, D.A., 2020. Permafrost degradation in the Western Russian Arctic. Environ. Res. Lett. 15, 045001. https://doi.org/10.1088/1748-9326/ab6f12

Vasiliev, A.A., Streletskaya, I.D., Shirokov, R.S., Oblogov, G.E., 2011. Evolution of cryolithozone of coastal zone of western Yamal during climate change. Kriosf. Zemli 2, 56–64.

Vonk, J.E., Tank, S.E., Bowden, W.B., Laurion, I., Vincent, W.F., Alekseychik, P., Amyot, M., Billet, M.F., Canário, J., Cory, R.M., Deshpande, B.N., Helbig, M., Jammet, M., Karlsson, J., Larouche, J., MacMillan, G., Rautio, M., Walter Anthony, K.M., Wickland, K.P., 2015. Reviews and syntheses: Effects of permafrost thaw on Arctic aquatic ecosystems. Biogeosciences 12, 7129–7167. https://doi.org/10.5194/bg-12-7129-2015

Zarov, E.A., Lapshina, E.D., Kuhlmann, I., and Schulze, E.-D., 2021. Dissolved organic carbon vertical movement and carbon accumulation in West Siberian peatlands, Biogeosciences Discuss. [preprint], https://doi.org/10.5194/bg-2021-211, in review.

Zysset, M., Berggren, D., 2001. Retention and release of dissolved organic matter in Podzol B horizons. Eur. J. Soil Sci. 52, 409–421. https://doi.org/10.1046/j.1365-2389.2001.00399.x

Table 1. Physico-chemical characteristics and sorption properties of the 4 mineral horizons from the discontinuous (E, Bs and Bh) and continuous (Bgh) permafrost zones.

Soil Horizon	**E (n=3)**	**Bs (n=3)**	**Bh (n=3)**	**Bgh (n=3)**
Permafrost zone	Discontinuous (Khanymey)			Continuous
Description	poor white sand	poor brunic sand	Al-Fe-rich sand	Silt loam
Material/horizon	albic	brunic	spodic	cryic
pH_{H2O}	5.1±0.4	5.0±0.4	4.9±0.8	5.9±0.7
Sand, %	95	98	82	45
Silt, %	4.3	1.4	16	48
Clay, %	0.7	0.6	2	7
C, mg g^{-1}	2±1.4	3.8±1.4	32.3±13.5	6.1±1.7
N, mg g^{-1}	0.12±0.03	0.13±0.05	0.89±0.42	0.2±0.01
Surface area, m^{-2} g	0.12	2.1	1.05	25.3
Total SiO_2, %	97±1.2	96±2.1	86±5.1	68.78
Total Al_2O_3, %	0.8±0.1	1±0.1	3.1±0.5	11.66
Total Fe_2O_3, %	0.1±0.03	0.2±0.1	0.4±0.1	4.1
$Al_{oxalate}$, mg g^{-1}	6.6±3.5	11±5.7	79±29	-
$Fe_{oxalate}$, mg g^{-1}	1.41±0.32	1.35±0.22	2.27±0.48	-
$Fe_{dithionite}$, mg g^{-1}	2.69±0.94	2.45±0.58	4.86±1.55	-

Table 2. Sorption characteristics (Eqn. 1) of the studied mineral horizons.

Soil Horizon	E (n=3)	Bs (n=3)	Bh (n=3)	Bgh (n=3)
Permafrost zone	Discontinuous			Continuous
Adsorbed OC, mg g^{-1}	0.1±0.05	0.55±0.09	-0.21±0.26	0.35±0.01
Adsorbed OC, mg m^{-2}	0.83±0.41	0.26±0.04	-0.20±0.25	0.014±0.0004
b, mg g^{-1} (Desorbed OC)	0.1±0.1	0.13±0.13	0.49±0.23	0.09±0.02
m	0.17±0.12	0.61±0.05	0.2±0.05	0.38±0.02
RSP, mg g^{-1}	0.13±0.15, y=0.17x-0.12, R^2=0.86	0.37±0.41, y=0.61x-0.14, R^2=1	0.6±0.25, y=0.20x-0.43, R^2=0.69	0.15±0.03, y=0.38x-0.09, R^2=1

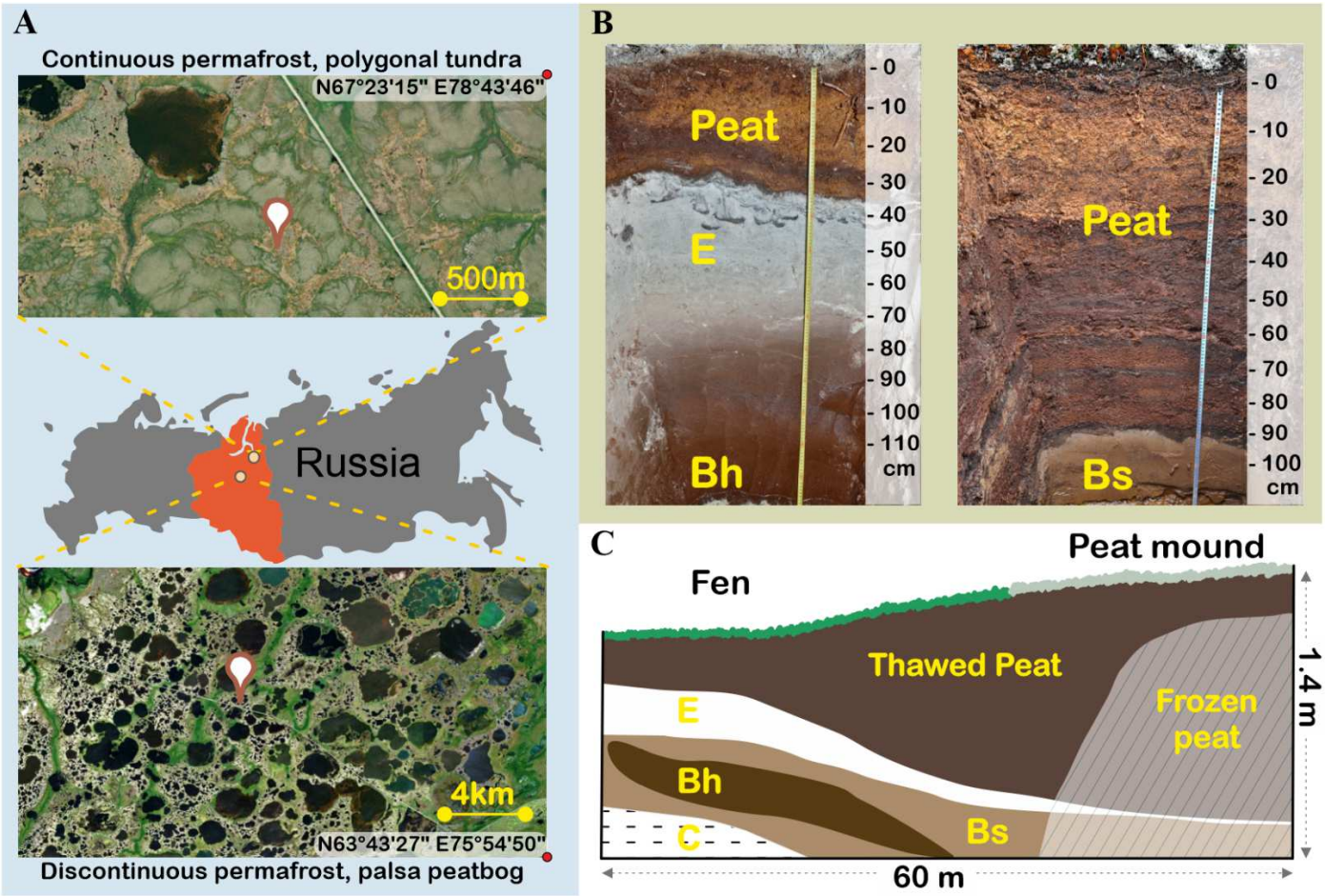

Fig 1. Map of the two study sites (A), a photo of soil profiles with peat (used for leachate preparation) and underlying mineral horizons (B), and a schematic representation of soil profile (C).

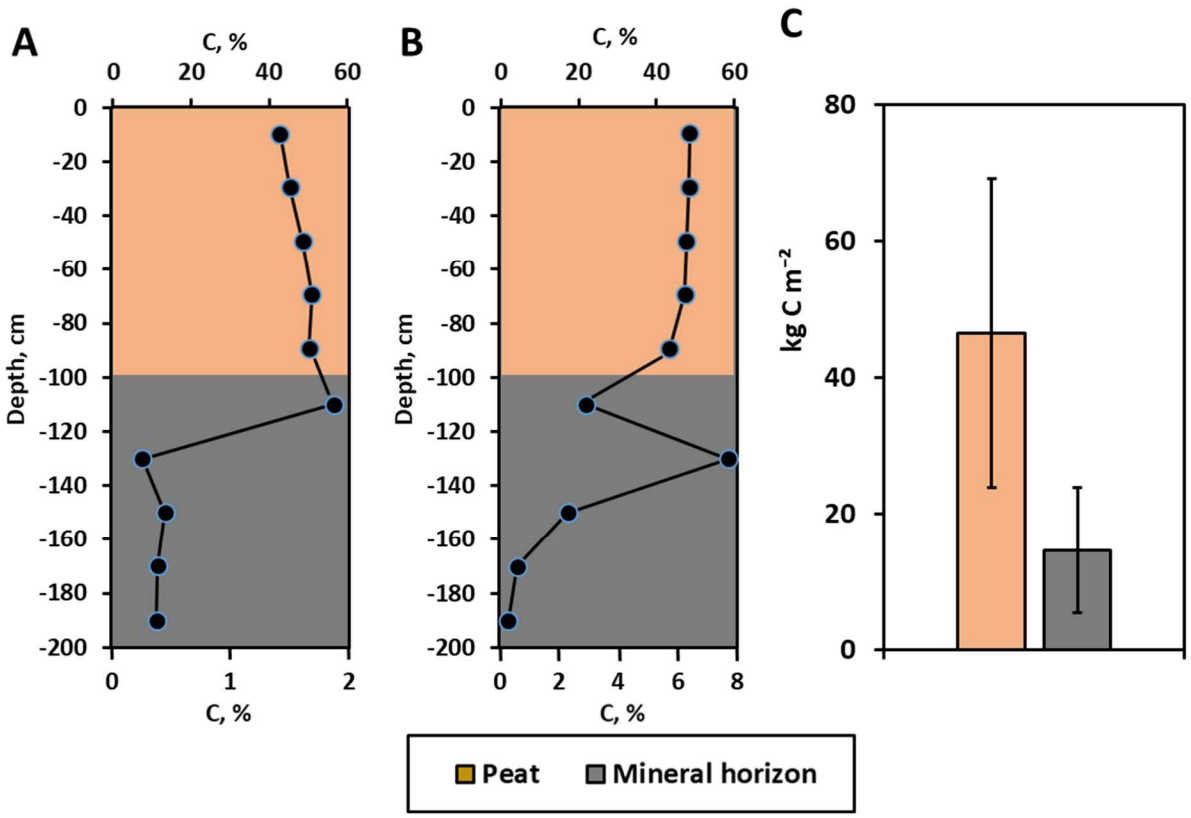

Fig. 2. Examples of depth profile of C concentration in the discontinuous permafrost zone (Khanymey, **A**) and continuous permafrost zone (Tazovsky, **B**) and lateral pools of C in Khanymey, averaged over 11 cores (C). Note that C concentration in peat and the mineral part of the core (A and B) are shown in the upper and bottom x-axes, respectively.

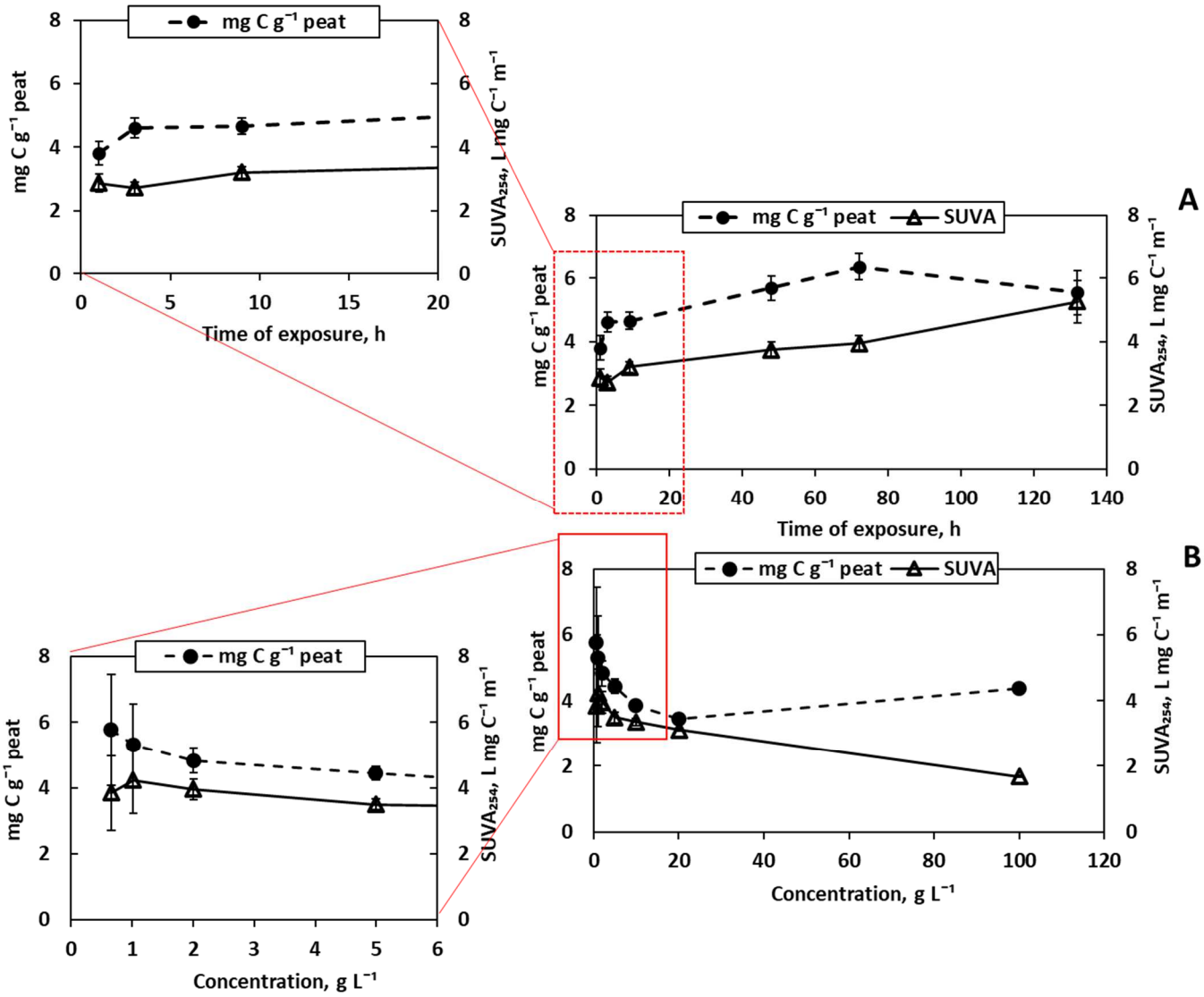

Fig. 3. Patterns of DOC (solid circles and lines, left y-axis) and $SUVA_{254}$ (open triangles and dashed lines, right y-axis) leaching from frozen peat (55-75 cm depth) in batch reactors as a function of exposure time at 1 g peat L^{-1} (**A**) and as a function of added peat concentration over 48 h of exposure (**B**). The error bars represent uncertainties of duplicates unless smaller than symbol size . Inserts illustrate the leaching during short exposure time (top left) and low concentration of peat (bottom left).

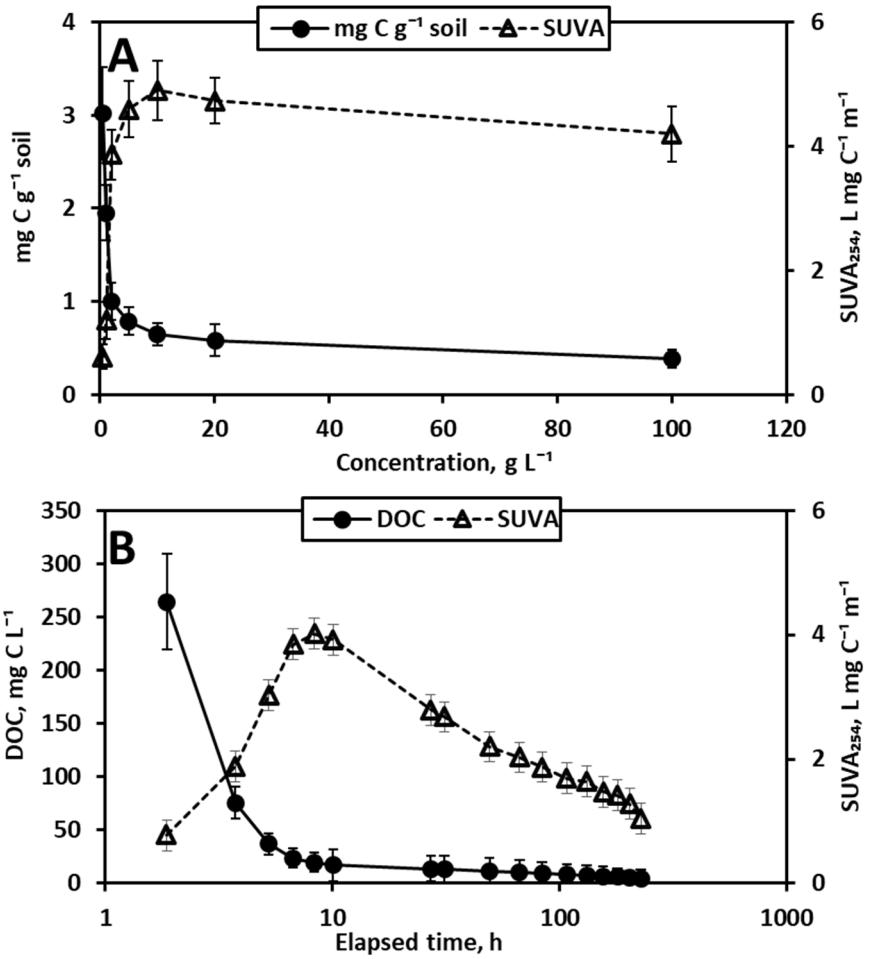

Fig. 4. Leaching of DOM from organo-Al-Fe-rich sand at different concentration of solid phase, after 48 h of equilibration time **(A)** and the breakthrough curves of DOC in a column reactor filled with 3 g of organo-Al-Fe-rich sand (Bh horizon), with a constant discharge of 3 mL h^{-1}. The error bars represent the uncertainties of duplicates (unless smaller than the symbol size). **(B)**. The evolution of DOC concentration is shown by solid circles and the $SUVA_{254}$ is represented by open triangles and dashed lines. The uncertainties are estimated from the propagation error and analytical procedure.

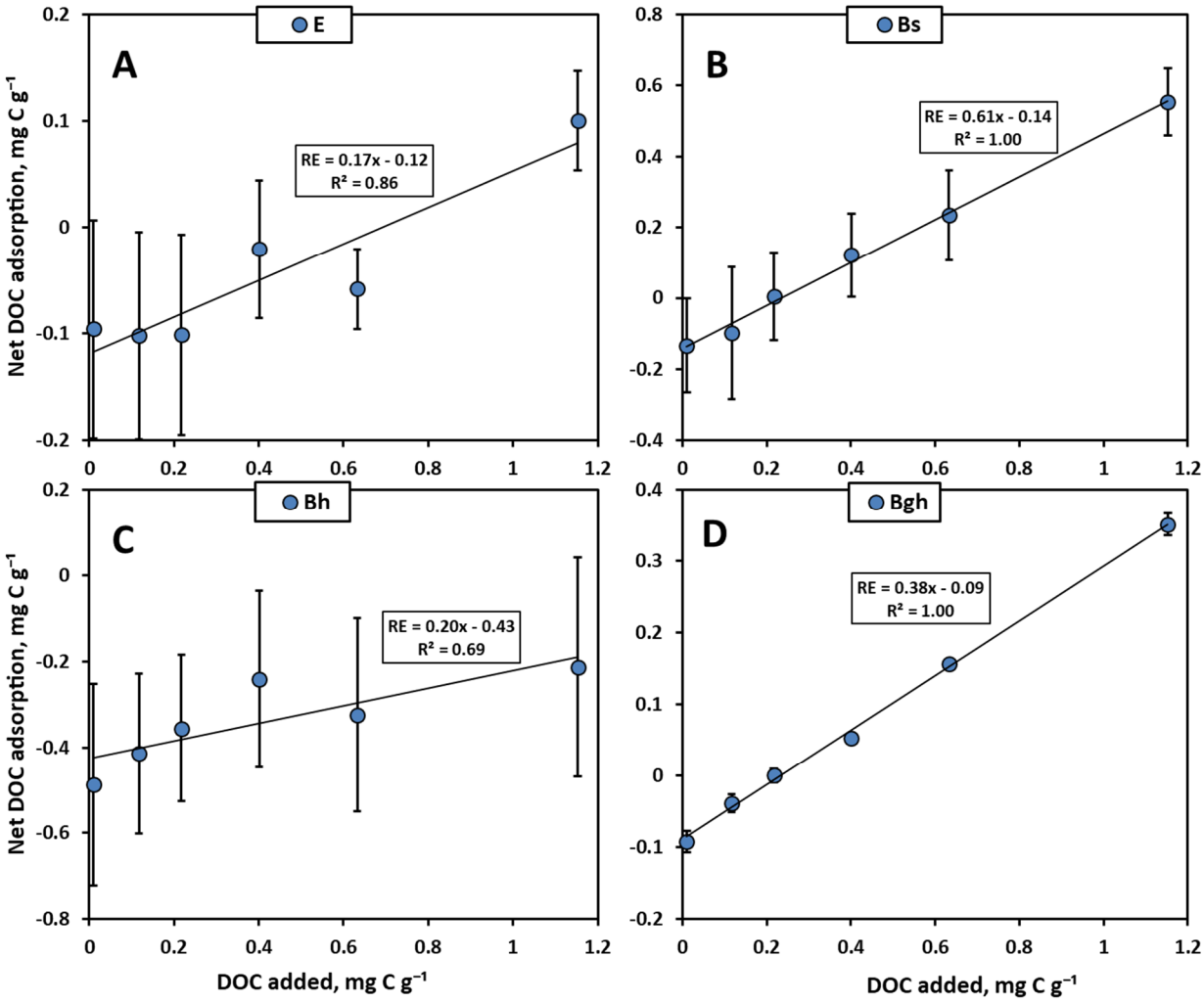

Fig. 5. Amount of DOC adsorbed or desorbed as a function of initial DOC concentration added to the reactor for mineral soil horizons of the discontinuous (A, B, C) and continuous (D) permafrost zone (1 g L^{-1}, 48 h of exposure). The linear regressions represent initial mass isotherms (Eqn. 1). Symbols represent the mean along with s.d. of experimental replicates (2 parallel reactors for 3 subsamples of each soil horizon, n = 6). Note that for the Bhg sample, the s.d. are within the symbol size.

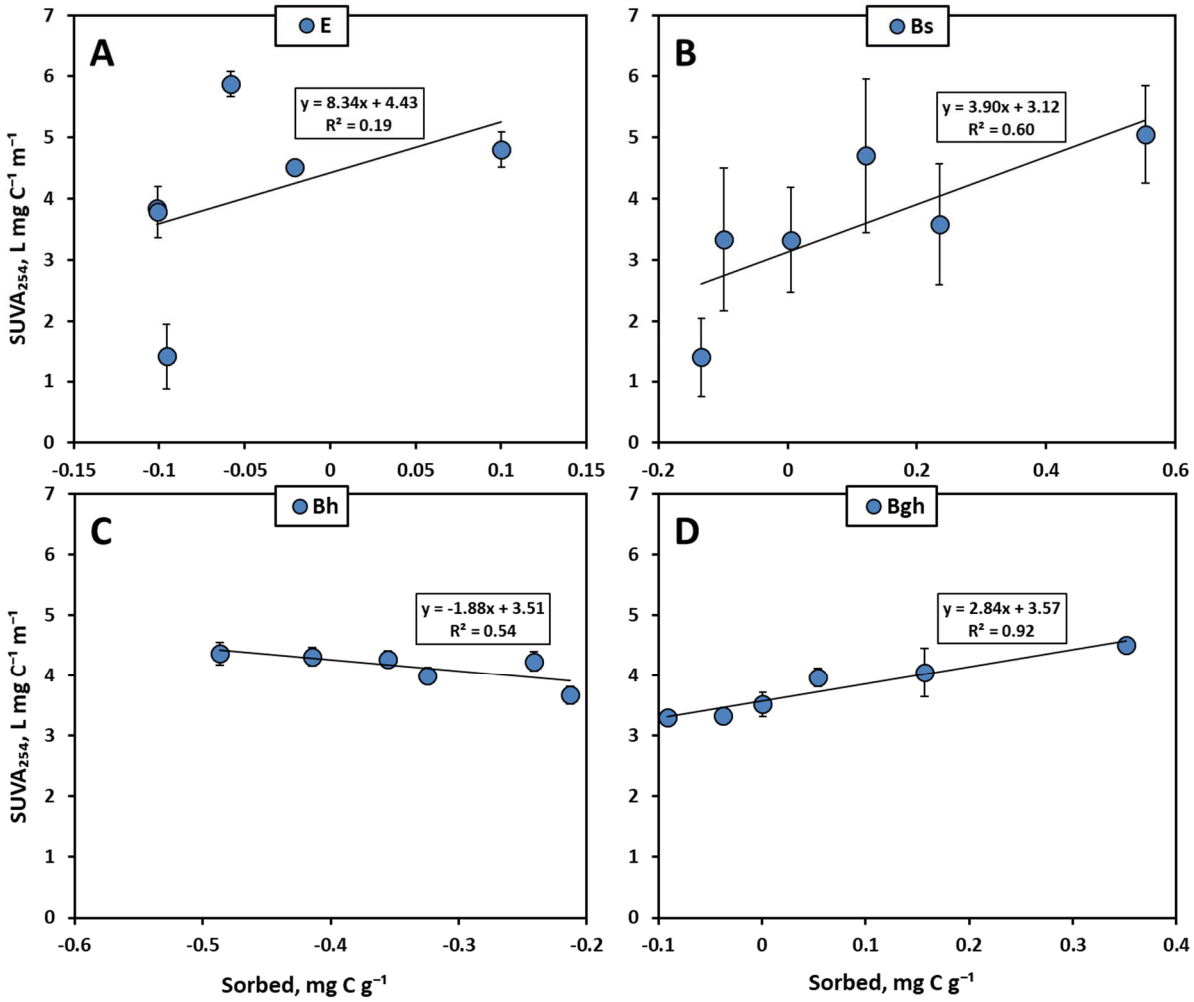

Fig. 6. Evolution of $SUVA_{254}$ (L mg C^{-1} m^{-1} , mean ± s.d., *n* = 6) as a function of sorbed DOC concentration for mineral soil horizons of the discontinuous (A, B, C) and continuous (D) permafrost zone (1 g L^{-1} of soil, 48 h exposure time).

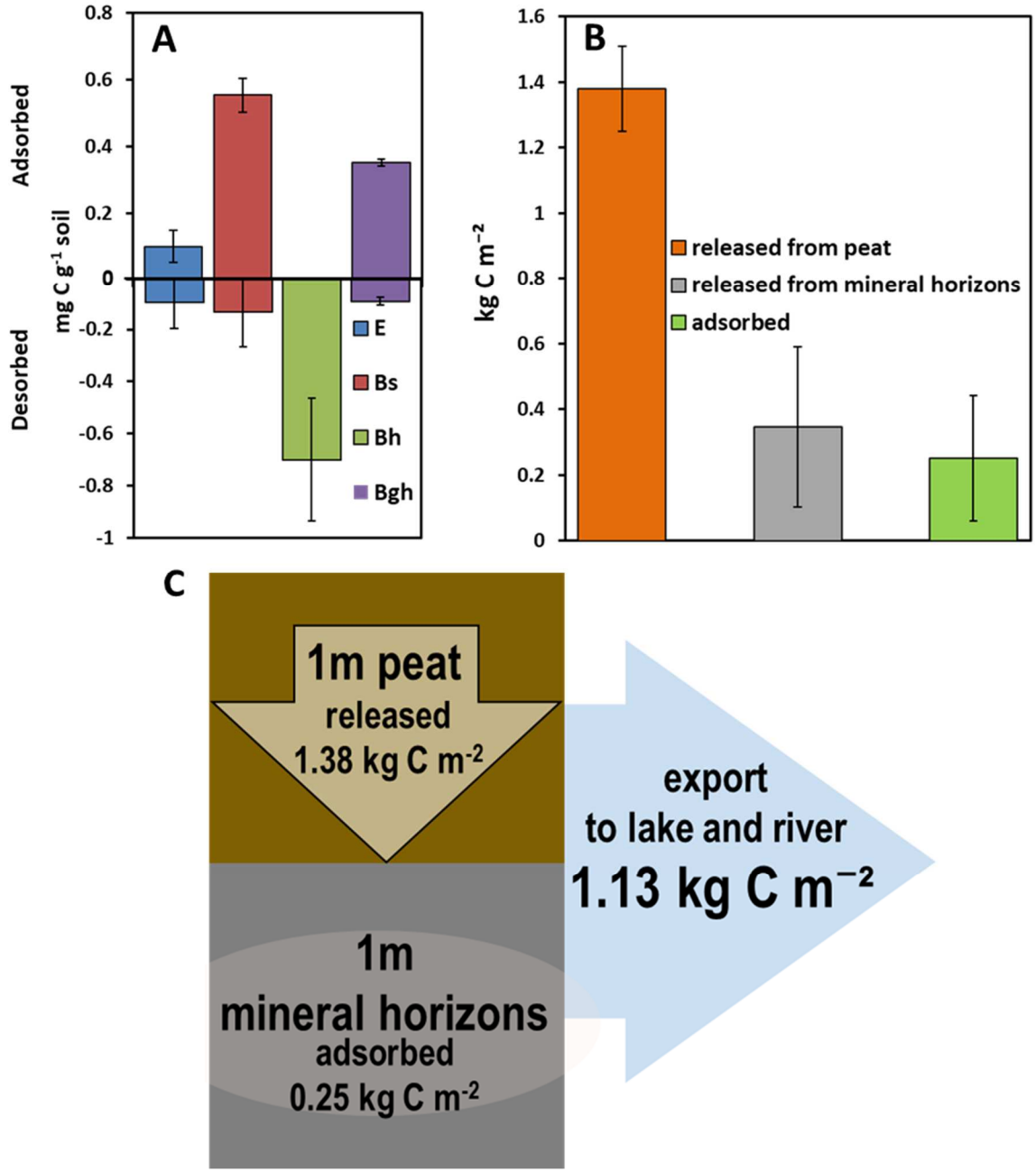

Fig. 7. Soil mass-normalized adsorbed and desorbed C concentration for the 4 mineral horizons **(A)**, the lateral pool of mobile C (over 1 m depth) that can be released from peat, mineral horizons and adsorbed onto minerals (**B**) and a mass balance illustration of C release from peat, adsorption on mineral horizons and lateral export to hydrological network (**C**).