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Short Communication
Potential Remobilization of Belowground Permafrost Carbon under Future Global Warming

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ABSTRACT

Research on permafrost carbon has dramatically increased in the past few years. A new estimate of 1672 Pg C of belowground organic carbon in the northern circumpolar permafrost region more than doubles the previous value and highlights the potential role of permafrost carbon in the Earth System. Uncertainties in this new estimate remain due to relatively few available pedon data for certain geographic sectors and the deeper cryoturbated soil horizons, and the large polygon size in the soil maps used for upscaling. The large permafrost carbon pool is not equally distributed across the landscape: peat deposits, cryoturbated soils and the loess-like deposits of the yedoma complex contain disproportionately large amounts of soil organic matter, often exhibiting a low degree of decomposition. Recent findings in Alaska and northern Sweden provide strong evidence that the deeper soil carbon in permafrost terrain is starting to be released, supporting previous reports from Siberia. The permafrost carbon pool is not yet fully integrated in climate and ecosystem models and an important objective should be to define typical pedons appropriate for model setups. The thawing permafrost carbon feedback needs to be included in model projections of future climate change.

INTRODUCTION

The total area of the northern circumpolar permafrost region, encompassing the zones of isolated, sporadic, discontinuous and continuous permafrost, is about 18.78 million km^2 or 16% of the global soil area. Recently, Tarnocai *et al.* (2009) provided a new estimate of 1672 Pg carbon (C) for the soil organic carbon (SOC) stored in the region, most of which is found in perennially frozen soils. This new estimate represents a significant increase compared with values previously reported, primarily as a result of quantifying deep frozen C deposits (>1 m), and was reported in Nature's Research Highlights (Ciais, 2009). The new estimate was also mentioned by Nobel Laureate Al Gore in his speech at COP 15 in Copenhagen (December 2009). A good barometer for how interest in permafrost carbon (C) has developed is the dramatic increase in the number of scientific papers that have been published on the topic in the past 5 years compared with previous decades (Figure 1).

Permafrost degradation has already been observed in parts of the northern circumpolar region and a significant portion of permafrost is expected to thaw out this century (ACIA, 2004). However, ground thermal regimes and ground ice content are highly variable at regional to landscape scales. Therefore, future thawing will progress at different rates and will result in uneven ground subsidence, even at very fine scales. A unique aspect of permafrost degradation is that gradual thawing of the ground with depth over time will be accompanied by more dramatic events, such as ground subsidence due to melting of buried ice bodies and lateral erosion along the edges of thaw lakes and Arctic coastlines. Thawing permafrost could result in remobilization of the previously frozen SOC pools and the release of large amounts of greenhouse gases (Schuur *et al.*, 2008). This is a positive feedback within the Earth System, as climate warming results in permafrost thawing that causes a further increase of greenhouse gases in the Earth's atmosphere, resulting in even more warming. This effect is not yet considered in climate model projections of future global warming.

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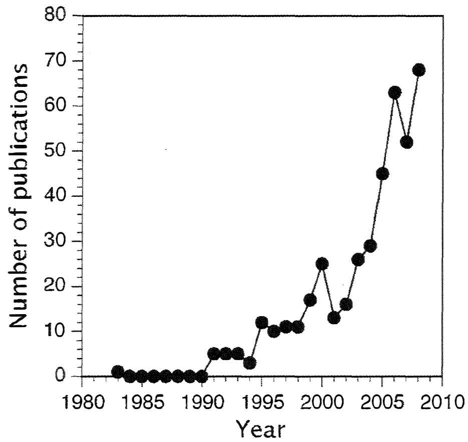

Figure 1 Number of scientific papers by year of publication returned from a Web of Science search for the terms permafrost + carbon.

The CAPP (Carbon Pools in Permafrost Regions) Project is an initiative of IPA (International Permafrost Association) and was a full project under IPY (International Polar Year). The CAPP Project was launched in 2005 with endorsement of the ESSP (Earth Systems Science Partnership) Global Carbon Project and the WCRP (World Climate Research Programme) Climate and Cryosphere Project. Its principal objective is to address the increased concern and awareness both within the international scientific community and the general public about the effects of global warming on frozen ground in the northern circumpolar region.

The aim of this short communication is to summarize recent progress in permafrost C research, in part carried out under CAPP, but also to identify remaining uncertainties and indicate directions for future investigations. There is an urgent need for more field studies, data syntheses across the northern circumpolar region, and full integration of permafrost C into modelling activities.

RECENT RESULTS

Total Storage

Carbon stored in the permafrost regions is one of the least understood, yet most significant, global terrestrial SOC pools. The spatial extent of permafrost soils and the amount of carbon they contain have been greatly underestimated. In addition, the specific soil processes leading to long-term carbon sequestration have not been taken into account.

Using the Northern Circumpolar Soil Carbon Database, or NCSCD (Tarnocai *et al.*, 2007), the carbon pools for all soils of the northern circumpolar permafrost region were determined for 0–30 cm and 0–100 cm depths. Soil C estimates were also made for the 0–300 cm depth and, separately, for the total depth (below 300 cm) of high carbon content deltaic and yedoma deposits.

The estimated carbon masses for all soils in the northern circumpolar permafrost region were 191 Pg for the 0–30 cm depth, 496 Pg for the 0–100 cm depth, and 1024 Pg for the 0–300 cm depth. In addition, the yedoma and deltaic deposits (below 300 cm) contain 407 and 241 Pg C, respectively (Tarnocai *et al.*, 2009). The distribution of carbon contents for the 0–100 cm depth of the various soils in the northern circumpolar permafrost region is shown in Figure 2.

In total, the northern permafrost region contains about 1672 Pg of organic C, of which approximately 1466 Pg C, or 88%, occurs in perennially frozen soils and deposits. This 1672 Pg C represents approximately 50% of the estimated global belowground organic C pool. The new estimate also indicates that the total belowground C pool in the northern circumpolar permafrost region is more than twice as great as the present atmospheric pool (ca. 750 Pg C) and more than three times as great as the total global forest biomass (ca. 450 Pg C).

The basic data used to generate these estimates were derived from numerous sources and, thus, the confidence values depend primarily on the number of pedons (soil data sets) used and the accuracy of the spatial extent they cover. Using confidence levels given in IPCC's Fourth Assessment, the carbon estimates for the North American portion had a medium to high confidence rating (66–80%) because this area had the greatest number of soil datasets (1169 pedons). The Eurasian soil C estimates had a low to medium confidence rating (33–66%) because of the small number of datasets (253 pedons). Finally, soil C estimates for deep carbon in yedoma and deltaic deposits had a very low to low confidence rating (<33%) since these values were determined using both a very small number of datasets and broad estimates for spatial extent.

Landscape and Soil Horizon Partitioning

It should be noted that the large permafrost C pool is not distributed evenly across the landscape, or with soil depth. This can be demonstrated by more detailed regional case studies.

A study from northeastern European Russia, covering a lowland taiga–tundra transition from the southern limit to continuous permafrost, indicated that the vulnerable permafrost C (large pool and high lability) was particularly found in palsa and peat plateau bog deposits (Hugelius and Kuhry, 2009). Peatlands accounted for 72% of SOC storage in the region investigated, with only 30% of the surface area. Of the perennially frozen SOC, 98% was found in peatlands, mainly peat plateaus and palsas in the northern forest–tundra and tundra regions. In these peat deposits, high C/N ratios in peat are associated with permafrost bogs, but not fens or thermokarst lake sediment (Hugelius and Kuhry, 2009). C/N ratios of peat might reflect differences in the original botanical constituents (e.g. Kuhry and Vitt, 1996; Vardy *et al.*, 2000) but are also a reflection of decomposition rates (e.g. Sannel and Kuhry, 2009). Here, permafrost bog deposits clearly represent the vulnerable soil C pool.

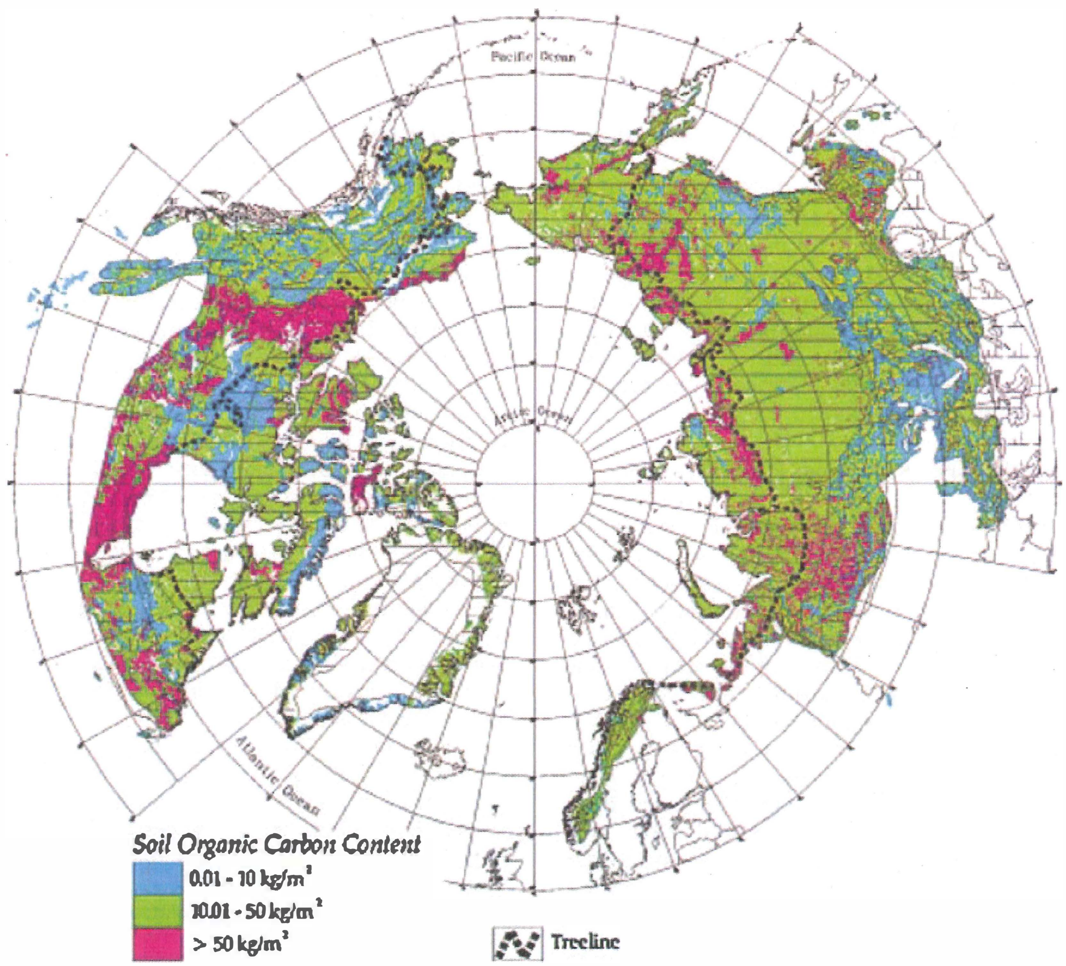

Figure 2 Distribution of SOC content (0–100 cm depth) in the northern circumpolar permafrost region (after Tarnocai *et al.*, 2009; © 2009, American geophysical Union; Reproduced by permission of American Geophysical Union).

In a study from the continuous permafrost zone of the central Canadian Arctic, peatlands (mainly bogs) again constituted the main SOC pool, with 56% of the total SOC mass, but cryoturbated soil pockets in Turbic Cryosols also contributed significantly, with 17% of the total pool (Hugelius *et al.*, 2010).

Ping *et al.* (2008) quantified SOC (<1 m depth) in the North American Arctic Region, and found that soils in lowlands and hilly upland areas had the highest SOC stores. They further emphasized the role of cryoturbation in vertical redistribution of large masses of SOC from the surface organic horizon to the so-called ‘transient layer’ of permafrost table fluctuation (found between 60 and 120 cm in this study). Many recent studies have emphasized the role of patterned ground and cryoturbation in the redistribution of SOC (Bockheim, 2007; Bockheim and Hinkel, 2007; Horwath *et al.*, 2008), and it has also been shown that cryoturbation may significantly reduce soil organic matter (SOM) decomposition rates. Results from the Central Canadian Arctic suggest that SOM in deeper cryoturbated organic pockets is less decomposed than SOM in the surrounding mineral horizons (Hugelius *et al.*, 2010). In a study of Siberian Turbic Cryosols, Kaiser *et al.* (2007) showed that cryoturbation of A-horizons deeper into the active layer retarded decomposition of SOM. Xu *et al.* (2009) showed that cryoturbated SOM in Alaskan Turbic Cryosols was of high quality and vulnerable to remobilization.

The decomposability of thawed permafrost C pools depends on botanical origin but also on decomposition trajectories. Much of the permafrost C has been exposed to decay in the active layer prior to incorporation in perennially frozen layers. What is now permafrost C might have first accumulated under permafrost-free conditions. This is clearly the case with many subarctic palsa and peat plateau deposits which first developed as permafrost-free fens during the Holocene Hypsithermal (Kuhry and Turunen, 2006). As a result, peat deposits were exposed to anaerobic decay prior to permafrost aggradation. Cryoturbation provides a mechanism for quicker burial of relatively undecomposed organic material. Similarly, the loess deposition in yedoma rapidly incorporated SOM with low degree of decomposition into permafrost (Zimov *et al.*, 2006). However, both peat plateaus and yedoma might have been significantly affected by repeated thermokarst episodes that temporarily exposed SOM to increased decay.

An impressive effort is underway, with a hierarchy of increasingly sophisticated geochemical techniques and incubation experiments, to assess the lability of permafrost C (e.g. Weintraub and Schimel, 2003; Michaelson and Ping, 2003; Dutta *et al.*, 2006). However, a synthesis is so far lacking and no attempt has been made to define or map SOM lability at the northern circumpolar scale. It is safe to state that knowledge of the quality and decomposability of SOM stored in permafrost is still limited.

Deep Carbon Remobilization from Permafrost Soils

The feedback from permafrost C to climate change depends on the amount of carbon stored in permafrost soils and on the rate of release of CH_4 and CO_2 to the atmosphere via microbial decomposition of thawed organic matter (Schuur *et al.*, 2008). Additionally, any changes in the quantity or quality of C input to the soil, through changes in plant productivity or community composition due to climatic warming and permafrost thawing (Turetsky *et al.*, 2007), should also be taken into account as these may either reinforce or counteract the feedback from permafrost carbon alone (Figure 3). High-latitude soils contain large quantities of C with the bulk of this soil C pool contained in the lower part of the active layer and in the permafrost underneath (Ping *et al.*, 2008). It is thus important to know whether respiration from deep C is sensitive to increasing temperatures and deeper thaw depths, as this could cause these soils to act as a significant and potentially long-lasting positive feedback to climate change.

For more than a decade, there has been debate about the long-term temperature sensitivity of deeper, low-quality SOC in general (e.g. Kirschbaum, 1995; Giardina and Ryan, 2000; Melillo *et al.*, 2002; Knorr *et al.*, 2005; Hartley *et al.*, 2008). But more recently, changes in the overall SOC pool size, such as when permafrost carbon is thawed and enters the active layer, have been recognized to be potentially of greater importance (Davidson and Janssens, 2006). Furthermore, experimental results from field studies on this topic have been quite rare (e.g. Goulden *et al.*, 1998; Christensen *et al.*, 1999). Recent findings in Alaska and northern Sweden provide strong evidence that the SOC in permafrost terrain is highly temperature sensitive and is starting to be released (Schuur *et al.*, 2009; Dorrepaal *et al.*, 2009), supporting earlier observations from Siberia (Walter *et al.*, 2006). These new studies have made use of C isotopes to determine how deep C contributes to changes in ecosystem respiration following permafrost thaw and experimental warming. By measuring the isotope value of total ecosystem respiration, of plant respiration and of the respiration of individual soil layers, isotopic differences provide a sensitive fingerprint for detecting changes in the respiration of carbon deep in the soil as permafrost thaws.

A study in a tundra site in the discontinuous permafrost zone of Alaska (63° 53′ N, 149° 15′ W) made use of a natural gradient of permafrost thaw in relatively well-drained upland moist acidic tundra that permitted quantification of changes in plant and soil C cycling processes as a function of time since permafrost thaw had begun. A long-term temperature record from a permafrost borehole at the site

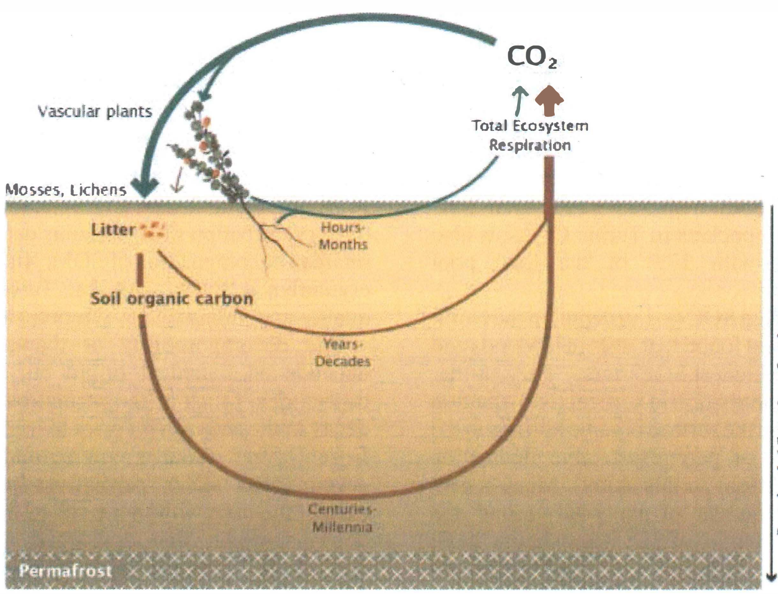

Figure 3 Carbon dioxide cycling in permafrost terrain. As the active layer deepens, old permafrost C becomes available for decomposition (adapted from Dorrepaal *et al.*, 2009).

demonstrated that permafrost thaw and ground subsidence had begun 15 years before the C cycle measurements. In combination with a nearby site where permafrost degraded more than 50 years ago and another with minimally altered permafrost, the gradient allowed quantification of changes in ecosystem C cycling over a decadal time scale of permafrost thaw. Increased C uptake by plants initially offset increased ecosystem respiration such that this area was a net C sink 15 years after the initiation of thaw (Schuur *et al.*, 2009). Using ^{14}C measurements of ecosystem respiration and of plant and soil respiratory components, it was demonstrated that respiration of deep C (defined as 15–35+ cm in this study) was higher than in the site with little permafrost degradation. Thus, the loss of deeper permafrost C was already occurring even though the site was a net C sink. Over more decades of thaw, plant growth rates remained high but increased old soil C losses from deep in the soil eventually offset increased C uptake. Deep C losses were 78% higher than the site with little permafrost degradation and this thermokarst became a net source of C to the atmosphere (Vogel *et al.*, 2009). The measured C emission rates suggested that 4.5–6.0 kg m^{-2}, or 9.5–13% of the SOM pool, could be lost on a century time scale (Schuur *et al.*, 2009).

Additional evidence for the high sensitivity of soil C located just above the permafrost to a rise in temperature above freezing comes from a parallel study conducted in subarctic Sweden (68° 21′ N, 18° 49′ E). The effect of small increases in temperature on soil C losses by exposing parts of a relatively dry, ombrotrophic peatland site underlain by permafrost to passive experimental warming were directly quantified using transparent open-top chambers. During the 8-year study, increasing air and soil temperatures by as little as 1 °C during the spring and summer seasons (without changing the active layer depth) enhanced total ecosystem respiration rates on average by 60% and 52%, respectively (Dorrepaal *et al.*, 2009). Biomass production of the dominant peat moss and shrubs was, however, not stimulated to the same extent by experimental warming during the growing season (Dorrepaal *et al.*, 2004, 2006). No decline was observed in the increase in ecosystem respiration rates over the 8-year period of the experiment. Comparisons of the ^{13}C signatures of the respiration in the field warming experiment with those of respiration of individual soil layers during laboratory incubation were used to elucidate the sources of the respiration increase. At least 69% of the observed increase in respiration rates in the warming treatments appeared to originate from deep C, defined in this study as the C in the active layer below 25 cm depth and just above the permafrost (Dorrepaal *et al.*, 2009). The high temperature sensitivity of respiration at greater depth in permafrost soils is most likely related to a greater lability of the SOC near the permafrost rather than to a higher activity or sensitivity of the microbial community (Mangelsdorf *et al.*, 2009; Waldrop *et al.*, 2010). The large SOC reservoir at greater depth in permafrost soils thus has a large potential to sustain increased carbon losses to the atmosphere when active layer temperatures increase.

Schuur *et al.* (2009) indicated that if carbon losses from the Alaskan tundra exposed to thaw were a typical response, future annual net C emissions from widespread thaw across the permafrost zone could be similar in scale to current biospheric emissions from land use change. However, significant challenges remain in upscaling point measurements to the northern circumpolar region and modelling efforts at large scales are just beginning.

Upscaling Approaches and Comparisons Across Scales

Tarnocai *et al.* (2009) made use of an array of soil maps with large polygon sizes in the NCSCD and a generalized yedoma map to arrive at their new estimate of the SOC pool in the northern circumpolar permafrost region. Whereas these soil and yedoma maps are not expected to improve in resolution within the next years, global land cover products with high resolution (1 × 1 km, or higher) covering the northern circumpolar region are becoming available. Kuhry *et al.* (2002) showed for the Usa Basin (northeastern European Russia) that linking pedon data to land cover classes results in similar total storage values compared with upscaling using a regional soil map, even though the SOC pool according to land cover classes showed greater variances than the one using soil classes. Here we compare recent estimates of SOC storage based on pedon data upscaled using land cover classifications derived from Landsat satellite imagery (30 × 30 m raster resolution) for the northern Usa Basin, northeastern European Russia (Hugelius and Kuhry, 2009) and for an area near Tulemalu Lake, central Canadian Arctic (Hugelius *et al.*, 2010) with corresponding estimates for the same study areas calculated by Tarnocai *et al.* (2009) using the NCSCD (calculations for 0–100 cm depth, lakes not included).

Estimates of SOC storage for the northern Usa Basin (forest–tundra and tundra zones, ca. 55 000 km^2) vary from 27.6 kg C m^{-2} (regional database) to 40.1 kg C m^{-2} (NCSCD). The mean size of the NCSCD polygons intersecting the northern Usa Basin is 211 ± 372 km^2 ($n = 407$, median size = 99 km^2). The databases mainly differ in the representation of peat deposits and cryoturbated permafrost soils. The mean spatial coverage and SOC content for peat deposits are 34.0% and 67.3 kg C m^{-2} in the NCSCD, while corresponding numbers for the regional database are 31.8% and 55.7 kg C m^{-2}. Further, Hugelius and Kuhry (2009) found no evidence for significant burial of SOC through cryoturbation, while the NCSCD has a high coverage of SOC-rich, cryoturbated soils (46% of total area and 36% of SOC 0–100 cm).

The Tulemalu Lake area (403 km^2) studied by Hugelius *et al.* (2010) intersects four polygons in the NCSCD (mean area 5426 ± 2510 km^2), all of which are described as Turbels (74 ± 4%) and rockland (26 ± 4%). The mean SOC content estimated by the NCSCD is 21.4 kg C m^{-2}, while Hugelius *et al.* (2010) estimated SOC storage to be 33.8 kg C m^{-2}. The estimates for mean SOC storage in cryoturbated permafrost soils are similar between the two data sets: 27.8 kg C m^{-2} for the NCSCD and 25.7 kg C m^{-2} for the regional database. The clear difference between the databases lies in the absence of peat deposits in the NCSCD, while Hugelius *et al.* (2010) estimate that peatlands hold 56% of the SOC pool (covering 37% of total area).

Comparisons of estimated mean SOC storage between the NCSCD and regional scale studies in Russia and Canada revealed some significant differences, in one case the NCSCD estimate being 45% higher (Usa Basin) and in the other case 37% lower (Tulemalu) than the respective regional estimates. This direct comparison of estimates at widely different scales is interesting because it can provide information on where to focus future research efforts. In the northern Usa Basin, the discrepancy was due to differences in the representation of peatlands and cryoturbated soils, and at Tulemalu Lake the difference was caused by differences in peatland representation. While estimates of mean SOC content of non-cryoturbated upland soils are relatively well constrained between studies, there are difficulties with estimating SOC storage in cryoturbated soils and peat deposits. Burial of surface organic material through cryoturbation processes and the depth of peat deposits are spatially highly variable and mapping efforts are labour and cost-intensive.

CONCLUSIONS AND RECOMMENDATIONS

Recent findings on permafrost carbon research were discussed during the Second CAPP Workshop held in Stockholm from 3 to 5 June 2009, which was planned to summarize progress at the end of the IPY years. A number of key issues were raised that require further attention.

Whereas a new comprehensive update of the NCSCD is only reasonable when significant numbers of new pedons have become available, rapid updates on deep cryoturbated and yedoma permafrost carbon are considered urgent as they are at present so poorly constrained. Additionally, new field studies are needed especially for the underrepresented Eurasian and High Arctic sectors.

Regional pedon datasets from across the northern circumpolar region are needed to identify which combination of climatic, permafrost, vegetation and soil conditions lead to sequestration of large amounts of highly labile permafrost carbon and which factors could affect its subsequent remobilization.

A synthesis on permafrost SOM quality is so far lacking. Ongoing research needs to pinpoint relatively simple geochemical indicators to characterize the decomposability of permafrost SOM. Generalized classes should be identified to map SOM lability at a northern circumpolar scale.

A future aim should be to assess if land cover classifications, which have much higher resolution than soil maps, can be reliably used to estimate SOC pools at the northern circumpolar scale.

Periglacial processes such as differential ground subsidence, which alters surface hydrology, and thermokarst/coastal erosion should be better understood and quantified in order to assess the potential future remobilization of permafrost carbon.

A constant dialogue should be maintained with the climate and ecosystem modelling communities in order to compile and aggregate the highly variable permafrost carbon pool data into meaningful pedons applicable to different model setups, with vertical distribution of soil C quantity and quality (mean and range), for all the land cover and/or soil classes differentiated according to permafrost zone.

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