Permafrost - physical aspects, carbon cycling, databases and uncertainties

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Abstract

Permafrost is defined as ground that remains below 0 °C for at least 2 consecutive years. About 24 % of the northern hemisphere land area is underlain by permafrost. The thawing of permafrost has the potential to influence the climate system through the release of carbon (C) from northern high latitude terrestrial ecosystems, but there is substantial uncertainty about the sensitivity of the C cycle to thawing permafrost. Soil C can be mobilized from permafrost in response to changes in air temperature, directional changes in water balance, fire, thermokarst, and flooding. Observation networks need to be implemented to understand responses of permafrost and C at a range of temporal and spatial scales. The understanding gained from these observation networks needs to be integrated into modeling frameworks capable of representing how the responses of permafrost C will influence the trajectory of climate in the future.

1 Permafrost: a phenomenon of global significance

Ongoing discussions on the degradation of permafrost are now reaching a broad scientific, public, and political audience since the consequences of permafrost degradation are expected to be felt not only locally (infrastructure) and regionally (water supply), but also globally due to the resulting greenhouse gas (GHG) emissions. Trenberth (2010) highlighted the need to include feedback mechanisms, such as GHG emissions from shrinking permafrost areas in climate models for the next Intergovernmental Panel on Climate Change (IPCC) assessment. Changes in carbon (C) cycling (the mobilization of frozen, formerly protected soil organic carbon (SOC) pools) and changes in surface energy partitioning (Chapin et al. 2005) are expected to be nonlinear and, once certain thresholds in a system have been reached, subsequent incremental changes have the potential to produce strong effects. However, changes to permafrost have not yet been taken into account in these "tipping elements" for the earth’s climate system, in which a small change in control parameters can have large consequences for some system variables (Lenton et al. 2008). Recent studies have emphasized the role of permafrost as a crucial factor affecting the global C budget. Thawing of permafrost has the potential to release large C reservoirs (Schuur et al. 2008), but there is a lot of uncertainty about the sensitivity of the C cycle to changes in the Arctic (McGuire et al. 2009). This uncertainty is due to a limited knowledge of the large Arctic area in general, as well as to small- scale variability and the complexity of processes.

Friedlingstein et al. (2006) have shown that the current generation of global climate models lead to very large uncertainties in the magnitude of feedbacks to global change from the high northern latitudes. There is a large inter-model spread, with estimates of cumulative carbon dioxide (CO_2) exchange from high-latitude land surfaces by the year 2100 ranging from -80 (-: sink) to +20 (+: source) Pg C (1 Pg = 1 billion metric tons = 10^{15} g = 1 Gt) under the IPCC's SRES A2 scenario. It is possible that the model mean could even have the wrong sign, simply because the dynamics and size of the permafrost C reservoir were not correctly represented in the models used for the last round of the Coupled Carbon Cycle Climate Model Intercomparison Project (C4MIP). Moreover, the partitioning between methane (CH_4) and carbon dioxide (CO_2) emissions, which can have a major effect on global-scale climate feedback, was not even analysed because most models simply did not consider CH_4 emissions. Recent modeling by Schaefer et al. (2011) projects that the Arctic permafrost areas will change from C sinks to C sources by the mid 2020’s, amounting to a cumulative C flux to the atmosphere of 190 ± 64 Pg C by 2200. Schneider von Deimling et al. (2011) predict that by the year 2300 more than half of the potentially vulnerable C in the upper 3 m of soil of the northern permafrost region (600–1000 Pg C) could be released as CO_2. Permafrost, therefore, needs to be included in all projections of future climate.

This chapter provides an overview of the current understanding of the distribution of permafrost, the physical processes, and the interactions between physical and biogeochemical cycles, C pools and fluxes. The overall objective is to discuss major gaps and uncertainties in the understanding of both negative and positive feedbacks between permafrost and climate.

2 Permafrost: definition, distribution and history

Permafrost is defined as ground that remains below 0 °C for at least 2 consecutive years (van Everdingen, 1998). About 24 % (22.8 million km²) of the northern hemisphere land area is underlain by permafrost, amounting to about one fifth of the world's total land area (Brown et al., 1997). The material forming permafrost can include bedrock, sediment of mineral and organic origin, and ice (or unfrozen water with a freezing point depression due to high salt content). Permafrost regions are classified as continuous (with permafrost underlying 90-100 % of the landscape), discontinuous (50-90 %), or sporadic (10-50 %) and isolated (0-10 %) permafrost, on the basis of areal distribution (Brown et al. 1997, Fig. 1a). The thickness of the permafrost layer varies from less than one meter to up to 1600 m (recorded in Siberia: Romanovskii et al. 2004). The extreme thickness is explained by large regions in Northeast Siberia and North Alaska having remained unglaciated during much of the Pleistocene, which allowed greater heat loss from the ground. In addition, loess deposition and syngenetic permafrost (permafrost that formed through a rise of the permafrost table during the deposition of additional sediments or other earth material on the ground surface; van Everdingen, 1998) resulted in the formation of permafrost to a depth of several hundreds of meters, with a high ice content sporadically (>70 vol % at depths of up to 60 m in Yedoma sediments: Romanovskii et al. 2004). These late Pleistocene, ice-rich sediments "survived" the Holocene temperature maximum (ca. 5-9 kyr before present), when temperatures were up to 3 °C higher than today. A large portion of these ice-rich sediments was degraded by thermokarst, and riverine and coastal erosion. The exact amount of the fraction that "survived" is unknown, but first-order estimates assume that 1 million km^2 still exist, compared to an estimated more than 2 million during the last glacial maximum (Zimov et al. 2006; Walter et al. 2007). In contrast, the southern discontinuous permafrost zone is largely prone to thawing with warmer temperatures (Zhang et al. 2008).

3 Physical factors affecting the permafrost thermal regime

3.1 Permafrost temperatures

The ground thermal regime of permafrost areas has an important influence on their stability and resistance to environmental change. Permafrost temperature is a result of the glaciation history, the large scale surface energy balance (past and present), and the thermal properties (conductivity and capacity) of surface and subsurface material (soil, snow, vegetation) and water bodies, including subsurface waters (Yershov, 2004). Below the permafrost base, temperatures are above the freezing point of water due to geothermal heat from the interior of the earth (Yershov, 2004).

Significant variations are observed in permafrost temperatures within the Arctic. High Arctic areas have permafrost temperatures in the -5° to -10 °C range, with some sites below -15 °C (Romanovsky et al. 2010). The permafrost thermal state and spatial variability for three sites are shown as examples in Fig. 2. Imnavait Creek (northern foothills of the Brooks range, Alaska) was glaciated during the Pleistocene and continuous permafrost extends to depths of about 240 m (Lachenbruch et al. 1982), with present-day mean annual air temperatures (MAAT) of -7.4 °C (Fig. 2a). Of the three sites, the coldest permafrost temperature is recorded from Eastern Siberia (Fig. 2b), resulting from the lack of glaciation during the Pleistocene together with extreme climatic conditions during the Late Pleistocene and still today, with a present-day MAAT of about -14.9 °C. The warmest high Arctic permafrost exists to depths of only 100 m in Svalbard (Fig. 2c) as a result of repeated glaciations during the Quaternary and a relatively “warm” maritime climate with a present-day MAAT of -5.5 °C.

Almost all temperature measurements in permafrost show an increasing trend during the last 20 to 25 years of the 20th century, and into the first few years of the 21st century, caused by atmospheric warming and/or changes in snow depth (Romanovsky et al. 2010). For example, in Alaska the permafrost temperatures at 20 m depth increased by 1.5 °C over a period of 15 years (Osterkamp 2007; Hinzman et al. 2005). In warmer, discontinuous permafrost such warming will eventually convert permafrost ice to water and potentially mobilize previously frozen material containing C. Through the Global Terrestrial Network for Permafrost (GTN-P, http://www.gtnp.org/index_e.html), a network of permafrost observatories has been initiated since 1990 for detecting changes in permafrost temperature and active layer thickness (Circumpolar Active Layer Monitoring network, or CALM). However, these observatories are not distributed evenly; especially the Russian Arctic is an area of large data paucity.

3.2 Active layer dynamics

The upper soil layer in permafrost areas, characterized by seasonal thawing and freezing is called the *active layer*, and can range from several cm to several meters in thickness (van Everdingen 1998). The annual freeze-thaw cycle of the active layer is illustrated using temperature and soil moisture data over two years (2004, 2005; Fig. 3) from the polygonal tundra study site in Siberia (Fig. 4e). The annual temperature range is the highest at the surface, ranging between -30 and +15 °C, and it decreases with depth. During the short spring and summer period from June to September, the active layer progressively warms and thaws to a maximum depth of about 40 cm. Autumnal isothermal conditions (at 0 °C) and freeze back start in September but can take more than a month for the active layer to refreeze (Fig. 3). During winter, the ground temperature stays well below 0 °C and almost all water in the soil is frozen.

Only minor differences in thaw depth, soil temperature and volumetric water content occurred between 2004 and 2005, as a result of similar climatic and hydrologic conditions. Once the peat soil (with a porosity > 80 %) is thawed, the soil's volumetric water content remains at maximum (between 0.9 and 1 volume %). An exception to this are only drier years when evapotranspiration exceeds summer rainfall and the water level drops below the ground surface (Boike et al. 2008).

The active layer is of special significance for the C cycle since many of the biogeochemical processes take place in this layer and the release of C in the gas, liquid or particulate form occurs via this layer. Processes operating in the active layer include freezing and thawing and associated volume changes and sediment sorting due to frost heave and thaw subsidence, migration of water as a result of gradients (topographic, matric potential, temperature, vapour pressure, solute concentrations), and mechanical mixing through cryoturbation and cracking. These processes are complex and patterned ground results from a combination of many individual processes (Fig. 4; Washburn 1979). For example, non-sorted circles (Fig. 4a) are characterized by a circular, bare, almost flat soil surface (~1-3 m diameter) and lack a border of stones. These features form by differential frost heave and ice lens formation during winter (Washburn, 1979). Cryoturbation, i.e., the mechanical mixing of material as a result of seasonal freeze-thaw, mobilizes soil C both upwards and downwards (see examples in Fig. 5). Tarnocai et al. (2009) and Ping et al. (2008) have shown that cryoturbated soils generally have the highest mean SOC stocks. Koven et al. (2009) included soil organic matter (SOM) and a simplified vertical mixing in their C cycle model. The soil's thermal regime is changed by the change of the thermal properties (due to changes in SOC) which, in turn, affects the residence time of SOC. The effects of thermal insulation by organic matter (OM) and of cryoturbation lead to a significant increase in SOC stocks, in agreement with estimates of high-latitude soil C stocks (Tarnocai et al. 2009).

Predicting the vulnerability of permafrost C to climate change requires simulation of the active layer's annual dynamics coupled with the C cycle (Hollesen et al. 2010), as well as the soil water status which determines aerobic or anaerobic decomposition of OM. Predicting the hydrologic conditions of the surface is challenging due to difficulties associated with predicting regional changes of temperature, precipitation and drainage in global circulation models. To date, only a few small-scale local processes, such as cryoturbation, have been identified as globally significant for C cycling and incorporated into climate models.

3.3 Land cover

Land cover affects the biogeophysical and biogeochemical properties of the permafrost surface (hydrology, albedo, biomass, and vegetation type) which in turn determine the exchange of energy, water and C between the surface and the atmosphere (Bonan et al. 1995; Chapin et al. 2005). With a very low population density (0.32 per km^2), the anthropogenic influence on the land cover of permafrost areas is usually small and concentrated around towns or industrial areas in the south (Arctic Human Development Report: Einarsson et al. 2004). There is a limited use of land for cultivating crops, pasture, or forests. For example, less than 0.5 % of the total land area covered by the Lena river drainage basin in Siberia is used for forest harvest and agriculture (McGuire et al. 2010).

Fig. 1b shows the present-day distribution of ecozones in permafrost areas, covering a wide spectrum from barren ground surface in high Arctic areas to tundra and boreal forests in the south. The Circum Arctic Vegetation Map (CAVM 2003) by Walker et al. (2005) is based on 1 km resolution satellite data and provides a more detailed map of Arctic vegetation types. The CAVM is, however, restricted to the Arctic tundra, covering an area of about 5.05 x 10^6 km^2 (Walker et al. 2005) while forests cover an area of 14.6 x 10^6 km^2 (ACIA 2004), with large regions in central Siberia extending into the continuous permafrost zone. Boreal forests have been predicted to decrease and shift northwards; forest–steppe and steppe ecosystems have been predicted to be the dominant vegetation type rather than forests over half of Siberia in warmer and drier climate scenarios by 2080 (Tchebakova et al. 2009).

Water surfaces show a distinctly different surface energy balance than the surrounding vegetated or barren land surfaces (Gutowski et al. 2007; Rouse et al. 2007). Lakes and wetlands are abundant in permafrost landscapes (Fig. 1c). They show the greatest number and surface area compared to temperate and tropical lakes (Lehner and Döll 2004), but little difference in lake number and size has been found between areas of continuous, discontinuous, and/or sporadic permafrost (Smith et al. 2007). Current databases (such as the Global Lakes and Wetlands Database by Lehner and Döll 2004) only capture larger lakes (>10 ha), with smaller water bodies being not visible by coarse-resolution mapping (Grosse et al. 2008; Muster et al. 2011). Moreover, land cover classifications on similar scales show considerable diversity in Arctic regions, especially with respect to the extent of water bodies and wetlands (Frey and Smith 2007).

Thermokarst is a process that drastically alters the surface structures in permafrost terrains. The process of thermokarst involves the thawing of ice-rich permafrost, and subsidence of the ground surface (see example in Fig. 6). Thermokarst lakes and ponds are formed through water accumulating in the resulting depressions. In discontinuous permafrost areas, further thawing, however, may lead to the draining of lakes when thaw bulbs (taliks) underneath a lake completely penetrate the permafrost (Yoshikawa and Hinzman 2003). The quantification of thermokarst lake dynamics of three circumpolar peatland sites shows that lake drainage and new thermokarst lake formation is most pronounced in the sporadic permafrost zone (northern Sweden) compared to the continuous (central Canada) and discontinuous (northeastern European Russian) permafrost zone (Sannel and Kuhry, 2011).

Changes in land cover (due to thermokarst or fires), earlier snow melt, and later snow cover result in a change in surface albedo which alters the radiative feedback from the surface to the atmosphere. Vegetation and organic soil layers may act as insulators, protecting permafrost from warmer temperatures. For example, the expansion of shrub cover observed on tundra in Alaska (Sturm et al. 2001; Tape et al. 2006), reduced the mean annual permafrost temperature by several degrees (Blok et al. 2010). This effect may be offset by an increase in snow cover associated with shrub expansion (Sturm et al. 2005) which insulates permafrost from cold winter temperatures (Blok et al. 2010).

Documenting the current state of land cover establishes a baseline for monitoring land cover changes due to climate warming. None of the existing land cover classifications are able to resolve the heterogeneity of Arctic land surfaces which can alter the surface properties on a scale of meters. Regional studies show promising results as they can process higher resolution data such as Landsat data (Schneider et al. 2009), but the lack of uniformity in classification approaches means that it is difficult to make comparisons across the Arctic (Walker et al. 2005).

The accuracy of available databases is largely compromised by their low resolution and limited field validation or ground truthing. This is especially evident in the case of small water bodies or patterned ground.

3.4 Surface energy balance

The permafrost's surface temperature is determined by the surface energy balance composed of radiation (short and long wave), turbulent heat fluxes (sensible and latent), and heat flux into the ground, snow, or water bodies. Several descriptive seasonal studies are available for Alaskan, Canadian, and Scandinavian sites, as well as a few for Siberian sites (Ohmura 1982; Eugster et al. 2000; Lloyd et al. 2001), but quantitative long-term studies are almost non-existent. Annual energy balance studies and studies from Siberia are particularly scarce but are of great importance for the validation of climate and permafrost surface schemes within climate models. In this section, the differences in annual surface energy budget are demonstrated for the Svalbard (Westermann et al. 2009) and Siberian sites (Langer et al. 2011a:b) introduced in Section 3.1. At both locations, the surface energy balance is determined largely by radiation, i.e., net short-wave radiation during summer as an energy source and net long-wave radiation during the winter as an energy sink (Fig. 7). The latent heat flux is a factor of two higher than the sensible heat flux at the wet tundra Siberian site while at the Svalbard site with drier surface conditions, the sensible and latent heat fluxes are nearly equal. The seasonal thawing of the active layer in July and August takes up to 20 % of the net radiation at the Siberian site. Of importance is the fact that the winter ground heat flux forms a significant component of the surface energy balance at the Siberia site, with a relative contribution of up to 60 %. The high contribution of this ground heat flux is due to the strong soil temperature gradient, the high ice content, and the large annual surface temperature range which is related to the extreme climatic conditions. During winter, the importance of water bodies, in this case small thermokarst ponds (~100 m^2: Fig. 4e) is twofold. First the energy storage close to the surface in the form of unfrozen water is greater than in the surrounding tundra so that the heat flux supplied to the atmosphere is higher by a factor of two. Second, ponds require substantially more time (several months) to fully refreeze so that biogeochemical processes are active for a much longer period of the year (Langer et al. 2011b). Predicting C emissions through permafrost and climate models requires the accurate representation of the energy balance, including freeze-thaw processes, as well as representation of sub-grid cell variability in the landcover, especially with regard to water bodies.

4 Carbon stocks and carbon mobilization

4.1 Carbon stocks of soils and deeper permafrost

Tarnocai et al. (2009) estimated that 495 Pg C are stored in the upper meter of soil, and 1024 Pg C in the upper three meters of soil in the northern circumpolar permafrost region (Fig. 1d). By including an 407 Pg C for deeper amounts stored in Yedoma (ice-rich and C-rich loess deposits formed in the Pleistocene) and 241 Pg C stored in alluvial deposits (Tarnocai et al. 2009) a total of 1672 Pg C was estimated. The organic soils (peatlands), cryoturbated soils, and Yedoma deposits being the soils with the highest mean organic C contents. The C content of Arctic permafrost areas is, therefore, significant when compared to global SOC stocks in the upper meter of soil of 1462 to 2344 Pg C (Batjes 1996; Denman et al. 2007).

Comparisons of soil organic carbon (SOC) stocks between regions are difficult because of the different methods and upscaling techniques used. Using the large scale circumpolar data set for cryoturbated soils (Turbels) for the Thule peninsula, Greenland, 0.018 Pg C was estimated. But applying a combined high resolution data set of field and satellite observations, only 0.004 Pg C was calculated (using a correlation between SOC from field profiles and high resolution satellite picture determined normalized difference vegetation index; Howarth Burnham and Sletten, 2010). For the central Canadian Arctic region, mean SOC storage estimates have been revised from previous estimates to show higher contents overall as well as large spatial variability. The peatlands (mainly bogs) comprised the highest SOC pool with 56 % of the total SOC, but cryoturbated soil pockets in turbic cryosols contributed 17 % to the total SOC stock (Hugelius et al. 2010).

The estimation of C stocks through upscaling attempts to compensate for limited data sets of SOC content, poor accuracy in geolocation of older datasets, low resolution of available soil maps, and high local variabilities in SOC. The estimation of permafrost C stocks is based on about 3530 pedons and confidence levels are higher for the North American regions to 1-m depth level data, while for the Eurasian sector they are low to medium (33–66 %) (Tarnocai et al. 2009). The lowest confidence (<33 %) is observed for deeper soil layers. It is, therefore, not surprising that these estimates need to be revised and updated. For example, a recent publication by Schirrmeister et al. (2011) suggested that the SOC stocks in deeper Siberian deposits may be smaller than that previously estimated based on revised bulk density data. Still, large uncertainties related to spatial distribution of soils, deeper stocks in Yedoma and peatland deposits render the total C stock highly uncertain (Tarnocai et al. 2009).

4.2 Carbon mobilization

Carbon in permafrost can be mobilized in response to pressure related disturbances (slow but persistent) and pulse disturbances (rapid but local) (Grosse et al. 2011). Pressure disturbances include changes in air temperature and directional changes in water balance, while pulse disturbances include fire, thermokarst, and flooding. Responses to these disturbances can result in paludification, aridification, top-down permafrost thaw, subsidence, changes in soil processes and vegetation cover, lake drainage, inundation, and erosion (Grosse et al. 2011; Kuhry et al. 2010). All of these disturbances can have effects on either vertical or horizontal C mobilization resulting in CO_2 or CH_4 exchange with the atmosphere resulting in a direct feedback on the climatic system or, in the case of lateral transport, for freshwater systems.

When permafrost thaws, large quantities of otherwise temperature-protected organic C becomes available for decomposition. The production and consumption of CO_2 and CH_4 within the pedosphere and biosphere have been documented in permafrost regions (Sachs et al. 2008, Friborg et al. 2000), but the role of storage and the timing between production, consumption and actual release of CO_2 and CH_4 is less clear. In addition to frequently reported spring emission peaks, Mastepanov et al. (2008) also reported high CH_4 emissions during autumn freeze-back that likely included significant amounts of CH_4 produced earlier during the growing season. Wagner et al. (2007), on the other hand, reported evidence of recent methanogenesis under *in situ* conditions in permafrost deposits at temperatures down to -6 °C. This microbial CH_4 production at subzero temperatures and recent data on spring and autumn emissions suggest that these traditionally understudied periods of the year may be particularly important in the context of the annual GHG budget of tundra ecosystems. Although it can be assumed that, over long time scales, the GHGs produced eventually reach the atmosphere, many process-based models now run on time steps where intermittent storage and short-term atmospheric processes may play an important role. For example, using eddy covariance methods, Sachs et al. (2008) and Wille et al. (2008) reported that atmospheric conditions (wind speed, pressure) play a dominant role in tundra-atmosphere CH_4 exchange running on daily time steps.

However, quantification of the actual CO_2, and, in particular, CH_4 land-atmosphere exchange remains localized with very few and often clustered flux towers providing data for only a small part of the vast terrestrial ecosystems in the Arctic (Fan et al. 1992; Friborg et al. 2000; Kutzbach et al. 2007; Sachs et al. 2008, 2010; Wille et al. 2008). Most research is concentrated in the Alaskan Arctic with relatively few sites in the Canadian Arctic and even fewer in the remote northern regions of the Russian Federation. Year-round operation of flux observation sites remains an exception rather than the rule, despite the importance of the winter, spring and fall seasons in annual C budgets. Consequently, most large-scale assessments of

Arctic C fluxes rely on process-based models, which include models that use drivers derived from remote sensing, or atmospheric inverse modeling. McGuire et al. (2010) analysed the decadal C balance of the entire Arctic Basin from 1997 to 2006 using several model-based tools. The terrestrial and marine Arctic has been found to be a net CO_2 sink of 0.109 Pg C yr^{-1}, but that the Arctic Basin as a whole is a source of GHG radiative forcing due to the terrestrial net source of 0.042 Pg CH_4 yr^{-1}. There is also a concern about a possible weakening of the high-latitude terrestrial CO_2 sink due to enhanced decomposition and increased fire frequency (Hayes et al. 2011).

Fire emissions may account for much of the variability in C sources and sinks between Arctic watersheds from 1997 to 2006 (McGuire et al. 2010). Although the proportion of burned area was small (e.g., ~4 % for the Lena watershed), the actual area covered was large with a 25 year fire return interval. Increased fire frequency and intensity have been observed in the second half of the 20th Century in Canada, Alaska and northern Eurasia (McGuire et al. 2004, 2007). Wildfires can destroy the insulating organic surface layer and warm the soil, increasing the rates of permafrost thaw and the active layer thickness (Yoshikawa et al. 2003; Johnstone et al. 2010). With subsequent regrowth of mosses the soils cool again, but with a warmer and drier climate, increased fire frequency and intensity may trigger a positive feedback loop between the loss of SOC and subsequent warming and thawing of permafrost soils (O'Donnell et al. 2011; McGuire et al. 2007).

Inland waters (lakes, wetlands, streams, and rivers) also process large quantities of organic C. Thermokarst lakes have been identified as a major source of CO_2 and CH_4 in permafrost terrain, from which the highest CH_4 fluxes are reported in Yedoma lakes, in which older C stocks are being mobilized (Zimov et al. 1997, Walter et al. 2006). However, a recent thermokarst lake modeling study by van Huissteden et al. (2011) reduced these flux estimates by an order of magnitude due to some hydrological effects, and lake drainage in particular, that limit lake expansion. Small polygonal ponds and lakes can release large amounts of C, accounting for 80 % of landscape-scale net CO_2 emissions during September at a polygonal tundra landscape in Siberia, with roughly half of this CO_2 outgassing from ponds (Abnizova et al. 2011, submitted).

The major Arctic rivers drain C-rich peatlands and soils and (Fig. 1b) transport large quantities of organic C to the Arctic Ocean (McGuire et al. 2010). The most important period for dissolved organic carbon (DOC) export is during spring snowmelt, when 60 % of the annual DOC flux occurs, half of which is estimated to be only 1-5 years old (Raymond et al. 2007). Raymond and colleagues conservatively estimated the total input of DOC into the Arctic Ocean to be 0.025-0.026 Pg C, which is 2.5 times more when compared to estimates of river export of temperate watersheds with similar basin size and discharge.

4.3 Arctic coasts, subsea permafrost, and gas hydrates

The Arctic coastal system is defined as extending landwards as far as the influence of the marine realm, and seawards as far as terrestrial influences to the edge of the continental shelf (Lantuit et al. 2011). Where the Arctic coastline is eroding, this coastal zone is expanding landward, with inundation of onshore permafrost transforming it into offshore or subsea permafrost (Fleming et al. 1998).

Terrestrial permafrost formed on Arctic continental shelves that were sub aerially exposed due to a lowered sea level during the Pleistocene glacial stages and not covered by continental ice masses. Subsequent inundation during the deglacial and Holocene sea level rise turned much of the terrestrial shelf permafrost into subsea permafrost. Today, most of the shelf area potentially affected by permafrost (>80 %) lies on the broad and gently inclined shelves of the Laptev and East Siberian Seas (Fig. 1a). Various models have shown that permafrost on the Siberian coastal plains has probably existed without interruption for at least the last 400,000 years, with cyclic cooling and warming corresponding to glacial and interglacial periods (Nicolsky and Shakhova 2010; Romanovskii et al. 2004).

Through warming of permafrost by geothermal heat flux from below and atmospheric cooling at the surface, CH_4 and other volatiles may migrate into permafrost and be retained in pore space, either in gas form or as gas hydrates. The temperature-pressure fields within and below permafrost result in gas hydrate stability (Romanovskii et al. 2005). Significant deposits of GHGs, and especially CH_4, can thus be found on the Arctic shelf (Ginsburg and Soloviev 1995; McGuire et al. 2009). Warming of the shelf seabed may lead to thawing of the permafrost and the potential release of GHG into the water column and the atmosphere.

These sources of GHGs have not been well quantified, especially for the eastern Siberian shelf seas (the Laptev, East Siberian and Chukchi seas), and fluxes are not yet included in estimates of the Arctic C cycle (Semiletov and Pipko 2007). High concentrations of dissolved CH_4 in relatively shallow shelf sea waters, together with evidence of atmospheric venting, indicate that CH_4 flux from shelf sediments is high (Shakhova et al. 2010). Nonetheless, there remain uncertainties with regard to the amount of GHGs involved, what physical and chemical form it takes, and how changes to the shelf seas and underlying permafrost will affect its release.

Relatively stable sea levels in the Arctic slow the onshore to offshore permafrost transition, but coastal erosion continues through thermal retreat of the coastal cliffs (Overduin et al. 2007). The highest coastal erosion rates in the Arctic are observed where the modern coastline cuts through unconsolidated, ice-rich permafrost deposits. Since two-thirds of the Arctic coastline is composed of such material, a considerable length of coastline is affected (Lantuit et al. 2011). Historical data on coastal change in the Arctic is not as available as it is for more populated temperature regions, and the critically relevant question along the Arctic coast is to determine the current trajectory and rate of change in coastline position. Coastal erosion of Arctic permafrost may contribute as much as 0.0069 Pg C per year of particulate organic C (Rachold et al. 2003), which is of the same order of magnitude as the contribution from all Arctic rivers. This amount, however, needs to be re-evaluated because of the previous lack of data on the contribution of riverine and coastal dissolved organic carbon (DOC) and as new data for SOC content become available (see Section 4.1).

5 Modeling permafrost and carbon cycling under a changing climate

5.1 Modeling permafrost and implementing physical permafrost processes in global models

Since even the most advanced subsurface schemes employed in Atmosphere-Ocean General Circulation Models (AOGCMs) rarely include explicit treatment of the ground below 10 m depth (Roeckner et al. 2003, Cox et al. 1999), the thermal regime of permafrost cannot be obtained directly from model outputs. Specialized permafrost models which use AOGCM output variables, such as near-surface air temperature and snowfall, are therefore, used to obtain predictions on the future state of permafrost. Empirical or semi-empirical methods (degree days of freezing and thawing, factors to account for temperature offset due to snow cover) have generally been used for pan-Arctic projections of permafrost (Anisimov and Nelson 1996), revealing projected wide-spread degradation during this century. Such approaches, however, do not allow any detailed insight in the dynamics of permafrost processes. The state-of-the-art in permafrost modeling is represented by transient models which numerically solve the one-dimensional heat transfer equation for temperature in a soil domain between the surface and a specific depth. The heat transfer equation contains two effective parameterizations of the material, the volumetric heat capacity and the thermal conductivity. They both depend on the water content and also on temperature since they implicitly represent the energetics of freezing and thawing. Furthermore, both the initial state of the system and the forcing data for the upper and lower boundary conditions must be defined. Despite the simplicity of the governing physical laws, modeling the thermal regime of permafrost is an intricate task since:

1. In most permafrost simulations, a constant (geothermal) heat flux is used as a lower boundary condition (Riseborough et al. 2008; Zhang et al. 2003). Thus, the model domain must extend to depths at which a stable temperature gradient exists during the target

period of the simulation. Therefore, the longer the time period to be covered by the model, the deeper the chosen soil domain must be. For example, for a modeling period of 100 years a soil domain of more than 100 m is chosen for many studies (Zhang et al. 2003).

2. Only in very few areas where deep boreholes are available can the initial condition be derived from direct measurements (Marchenko et al. 2008). Elsewhere, a sufficiently long time series of forcing data is required to calculate the recent temperature distribution in the ground, especially if a deep soil domain is chosen. The required time series of this “model spin-up” can exceed 100 years in permafrost modeling studies (Zhang et al. 2008).

3. The forcing data for permafrost models can be derived from different sources such as meteorological observations, satellite measurements, or the output of atmospheric models. However, using the output of AOGCMs is the only possibility for future projections. Deficiencies in the output of these models are therefore, directly reflected and possibly emphasized in the permafrost models (Chapman and Walsh 2007). Reproducing the seasonal and perennial insulating snow cover in an adequate way is a particularly challenging task but critical for the ground thermal regime.

4. The results of permafrost simulations strongly depend on the thermal properties assumed for the soil. Although diverse approaches exist for calculating these parameters for permafrost soils on the basis of soil composition (de Vries 1952; Farouki 1981), virtually no ground truthing is available. Another major uncertainty relates to the parameterization of vegetation cover which displays distinctively different thermal properties compared to the underlying soil (Wania et al. 2009 a:b).

5. The thermal properties of both soil and the surface cover can exhibit a strong spatial variability over distances ranging from meters to few kilometers which is considerably less than the grid-scales of typical gridded data sets for soil and surface properties as well as those of atmospheric models (Wilson and Henderson-Sellers 1985). Programs have been initiated to measure subgrid variability in permafrost areas, but these are not yet reflected in

current modeling approaches (Westermann et al. 2011; Langer et al. 2010).

Input data sets and the thermal properties of soil can generally be tuned on a local or point scale to yield a satisfactory to excellent agreement with measured data (Romanovsky and Osterkamp 1997; Nicolsky et al. 2009). On a larger scale, where validation data sets are generally sparse to non-existent, permafrost modeling must rely on gridded data sets but coarse features such as the annual temperature amplitude which can usually be sufficiently reproduced (Oelke and Zhang 2004). Most permafrost modeling has been accomplished on local to regional scales. Zhang et al. (2008) demonstrated the use of a transient permafrost model for Canada that is based on the output of six AOGCM-generated climate scenarios. They predicted a reduction in the permafrost area of up to 20 % and a significant thickening of the active layer in the remaining areas by 2100. Their permafrost model was initialized during the Little Ice Age at which time the permafrost was assumed to be in a steady state defined by the average atmospheric temperature and the geothermal gradient. The “model spin-up” for calculating the present temperature distribution in the ground was accomplished using time series interpolated from meteorological observations. A similar study has been performed for Alaska by Marchenko et al. (2008), who initialized their model using soil temperature measurements from boreholes. They modeled a widespread permafrost degradation until the end of the century, except in the most northerly areas. Both studies were operated on coarse spatial resolutions of 0.5° x 0.5°. To moderate the impact of resolution-related inaccuracies in the data sets used for forcing the model, Stendel et al. (2007) used dynamic downscaling with a Regional Climate Model (RCM). They demonstrated significant improvements in the precipitation (and thus snow depth) pattern for eastern Siberia, resulting in marked improvements in the modeling of the thermal regime of the permafrost.

While future permafrost conditions derived from modeling appear to be converging towards a permafrost reduction on the order of 25 % and a significant thickening of the active layer during the remainder of this century, there has, to date, been no comprehensive pan-Arctic study based on a transient permafrost model. Furthermore, future permafrost models could be improved by including up- and downscaling algorithms to take into account subgrid variability of the key input and output variables.

5.2 Permafrost-atmosphere feedback through a modified surface energy balance

Future predictions on permafrost are necessarily based on AOGCMs. The “European Centre/Hamburg Model” (ECHAM, Roeckner et al. 2003), for instance, contains a total of five soil layers to a depth of 10 m but does not include the freezing of soil water (Roeckner et al. 2003). Thus, a considerable proportion of the energy exchange between soil and atmosphere is not accounted for (Boike et al. 1998; Romanovsky and Osterkamp, 2000). This raises the question whether it is valid to base permafrost predictions on the output of a model that does not take into account the physics of frozen ground. In other words, is there a feedback between permafrost processes and atmospheric near-surface variables, which could influence/impact the regional climate?

A number of studies have pointed out the influence that selected soil parameters have on the results of atmospheric modeling (Peters-Lidard et al. 1998). Viterbo et al. (1999) observed that the inclusion of soil freezing leads to a significant improvement in modeled near-surface air temperatures for periods when a freezing front is close to the surface. Rinke et al. (2008) reported that including a low-conductivity organic layer for Arctic land masses has implications for modeled near-surface air temperatures, and even for the modeled regional circulation patterns. Furthermore, long-term monitoring of the surface energy balance in Siberia has demonstrated that the ground heat flux resulting from refreezing of the active layer and subsequent soil cooling compensates for more than half of the radiative losses during winter (Langer et al. 2011b). Permafrost processes must therefore be considered to be a driving force behind the wintertime surface energy balance in this area, and, thus, an accurate description in a land-surface scheme is highly desirable.

Since such evidence suggests a more active role for permafrost processes in the climate system, increased effort has been put into incorporating permafrost processes into land-surface schemes (Nicolsky et al. 2007), so that coupled runs with AOGCMs seem to be only a matter of time. However, such new schemes can only be considered a step forward if they are validated for sites across the entire range of climatic and ecological conditions found in permafrost areas.

5.3 Modeling the permafrost-carbon feedback

The release of GHGs from permafrost soils is influenced by a number of factors, in particular the soil temperature, water content (as the aerobic/anaerobic state of the soil influences losses the types of GHG loss), SOC amount and quality, the availability of nutrients, and the composition of microbial communities. Models for soil C cycling based on these parameters are readily available (Jenkinson and Coleman 2008), although there is controversy about the classification of soil C stocks into pools with different turnover times, i.e., SOC quality (Davidson and Janssens 2006). For these modeling approaches to be applied in Arctic permafrost areas, field studies are required to validate their performance and to improve the mathematical representations between these factors and SOC dynamics. However, GHG emissions can only be successfully simulated if accurate data on the SOC concentration and distribution are available and if soil temperatures and soil water content are inferred from land-surface models. Furthermore, C cycling models are *a priori* formulations for a point in three-dimensional space, and application over large grid cells in terms of an area-averaged formulation may be problematic.

For these reasons, first estimates of the magnitude of permafrost C feedback are associated with considerable uncertainties. Off-line simulations with simplified climate scenarios (Schaefer et al. 2011) or simplified permafrost representation (Schneider von Deimling et al. 2011) have predicted a sizable permafrost C feedback after 2100, even under moderate warming scenarios. This lag is a consequence of the large thermal inertia of the frozen ground which delays thawing and the microbial degradation of the organic material.

A promising approach to a physically-based permafrost-C model has been presented by Khvorostyanov et al. (2008), but to date has only been applied offline for selected regions.

There remains a need for a fully coupled simulation with an AOGCM that includes permafrost C feedback. Furthermore, all the above-mentioned studies have pointed out that the applied methodology cannot account for crucial processes occurring on subgrid scale, such as the formation of thermokarst or modification of the local hydrological regime (Wania et al. 2009 a:b). The development of up- and downscaling techniques for the most important variables must, therefore, be considered a necessary prerequisite for reliable modeling of GHG release from permafrost areas.

6 Conclusions and recommendations

According to model projections, permafrost degradation will affect almost half of the current permafrost area in the northern hemisphere by 2100. Prediction of the sensitivity of the C cycle to climate change and permafrost thaw is complicated by complex interactions between hydrology, soil thermal regimes, and vegetation. These factors can result in both positive and negative feedbacks to permafrost and C exchange. Changes in land cover, such as vegetation type and distribution, or the areal extent of water bodies and drainage systems, will affect the vertical and horizontal fluxes of water, energy, and C. However, land-surface modules of most state-of-the-art coupled general circulation models include only considerably simplified descriptions of the thermal and hydrological effects of soil freezing and the related processes and properties (snow cover, high-latitude vegetation). They generally neglect the effects of sub-grid variability in landforms, soil types, etc., which have a strong influence on the large-scale effects of these processes.

Given the size of organic C stocks in permafrost, vulnerabilities associated with thawing and, decomposition of OM, and the production of GHG have the potential for strong positive feedbacks to the climate system. However, there are large uncertainties associated with estimates derived from the potential for both positive and negative climate feed backs as well as missing spatial data. In marked contrast to their recognized importance, comprehensive observations of Arctic soil, snow, and atmospheric quantities (climate and surface energy balance components) are extremely sparse.

For monitoring and modeling the present and future state of the permafrost, most approaches are designed to use averaged quantities for large grid cells. Such averaged quantities may, however, be inappropriate for some monitoring tasks. A spatial average of the active layer thickness is adequate for modeling distributed quantities like the background emissions of GHG. However, the monitoring of erosion or natural hazards due to permafrost degradation (which would initially occur at few localized "weak points" by capturing localized emission hotspots such as thermokarst lakes) is only feasible if information is available at the sub-grid level.

Implementing a potential feed-back mechanism by GHG-emission from warming permafrost in regional and global models requires the correct parameterization of a sufficient and critical set of thermal and hydrologic processes and parameters, such as the thermal properties of snow and soil. However, this would require: (i) accurate estimates of state variables such as SOC or thermal and hydraulic conductivities, (ii) both vertically and horizontally upscaled data on these state variables (Ciais 2010), (iii) an understanding of the key physical and biochemical processes in permafrost, and (iv) an understanding of the interaction and feedback mechanisms between permafrost and climate.

The primary objective of understanding permafrost and its role in the earth's climate system, including feed back mechanisms, requires new model developments and upscaling strategies:

1. Field studies should be conducted at representative sites to systematically monitor key parameters and processes over the long term (e.g. > 10 years), thereby improving the understanding of permafrost dynamics at a range of scales.

2. The development of conceptual and numerical permafrost landscape models is required, including suitable upscaling methods ranging from local to global scales, and
3. Remote sensing products must be used to test, validate and monitor (1) and (2) above.

All strategies require field-based knowledge of the surface characteristics, key processes and monitoring data for a few key parameters. Obtaining funds for long term process monitoring is often difficult due to political and technical limitations. Such funds are often cut from budgets since the benefits derived from field studies are often not apparent until after a longer period of time (Nisbet 2007). It is, therefore, urgent and timely to initiate efforts at various locations across the climatic and ecological gradients in permafrost areas, to eventually establish a pan-Arctic data base. Such a compilation would be of outstanding importance for improving the understanding of the sensitivities of permafrost and high-latitude ecosystems, and their susceptibility to climate change.

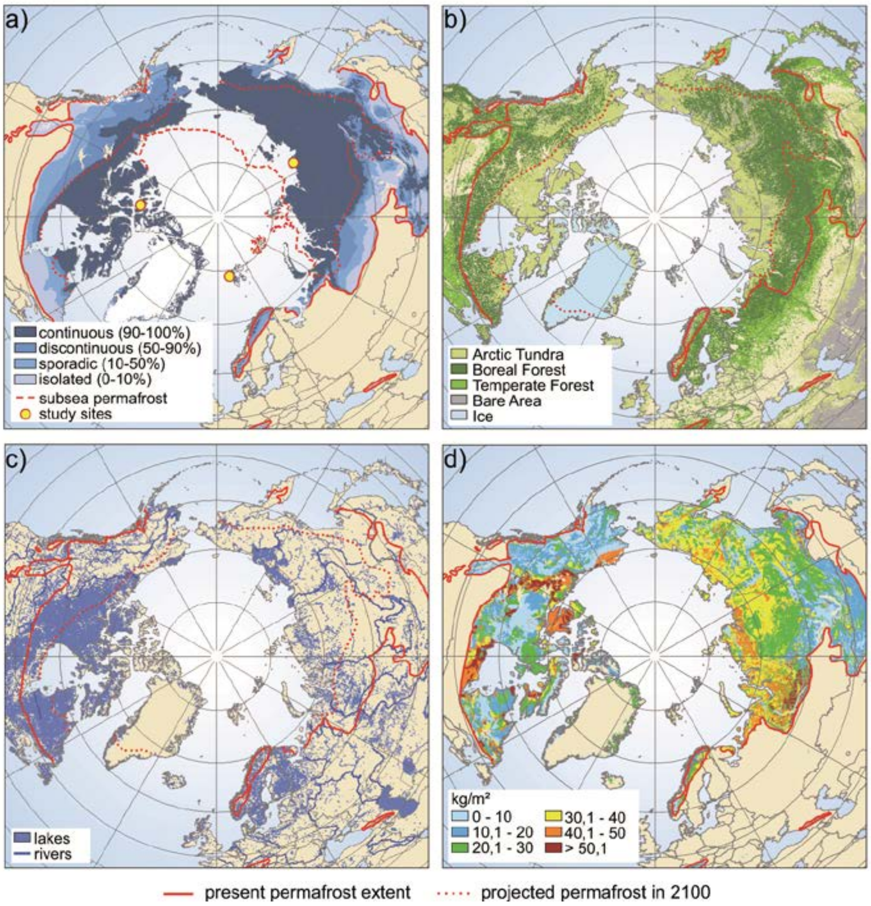

Fig. 1 a. Permafrost map, including subsea permafrost (Brown et al. 1997) and predicted decrease by 2100 (ACIA 2004). b. Major terrestrial ecozones (modified after Global Land Cover 2000 database). c. Lakes, wetlands and major rivers (modified after Lehner and Döll 2004). d. Organic carbon content map for the upper 100 cm of soils (Tarnocai et al. 2009). Sites introduced in Section 3 are shown as yellow dots in a.

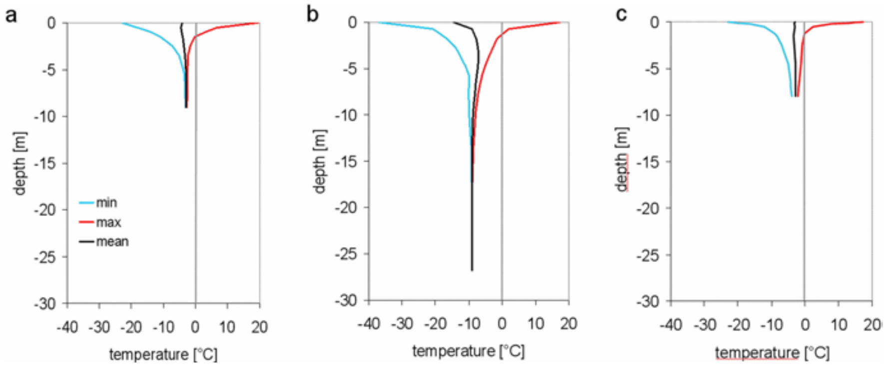

Fig. 2 Mean, maximum, and minimum permafrost temperatures recorded in selected boreholes in Alaska (a, Imnavait Creek, northern foothills of the Brooks Range), Siberia (b, Samoylov, Lena Delta), and Svalbard (c, Bayelva). For locations see Figure 1a. The same depth and temperature scales are used to demonstrate the differences between warm (Svalbard) and cold (Siberia) permafrost. Analysis based on August 2009 - August 2010 (Svalbard), August 2007 - August 2008 (Siberia) and February 2006 - February 2007 (Alaska). Data for the Alaska site courtesy of L. Hinzman (Hinzman et al. 2008).

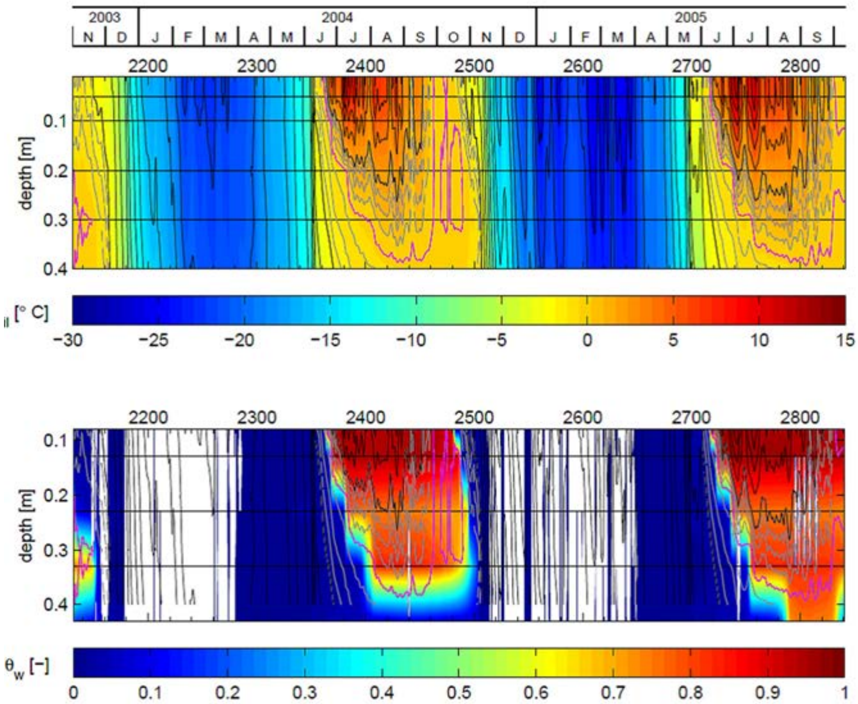

Fig. 3 Example of thermal and hydrologic dynamic of the active layer at the Siberian permafrost site (Lena Delta, Fig. 4e). Profile plots of soil temperature (upper plate) and volumetric water content using Time Domain Reflectometry (lower plate) over two years for the polygonal tundra site. The straight, horizontal black lines indicate the probe depths while the curved black lines are isothermal lines at intervals of 2 ºC The 0 ºC isotherm is shown as a magenta line, with intermediate isotherms (at intervals of 0.5 °C) shown as gray lines.

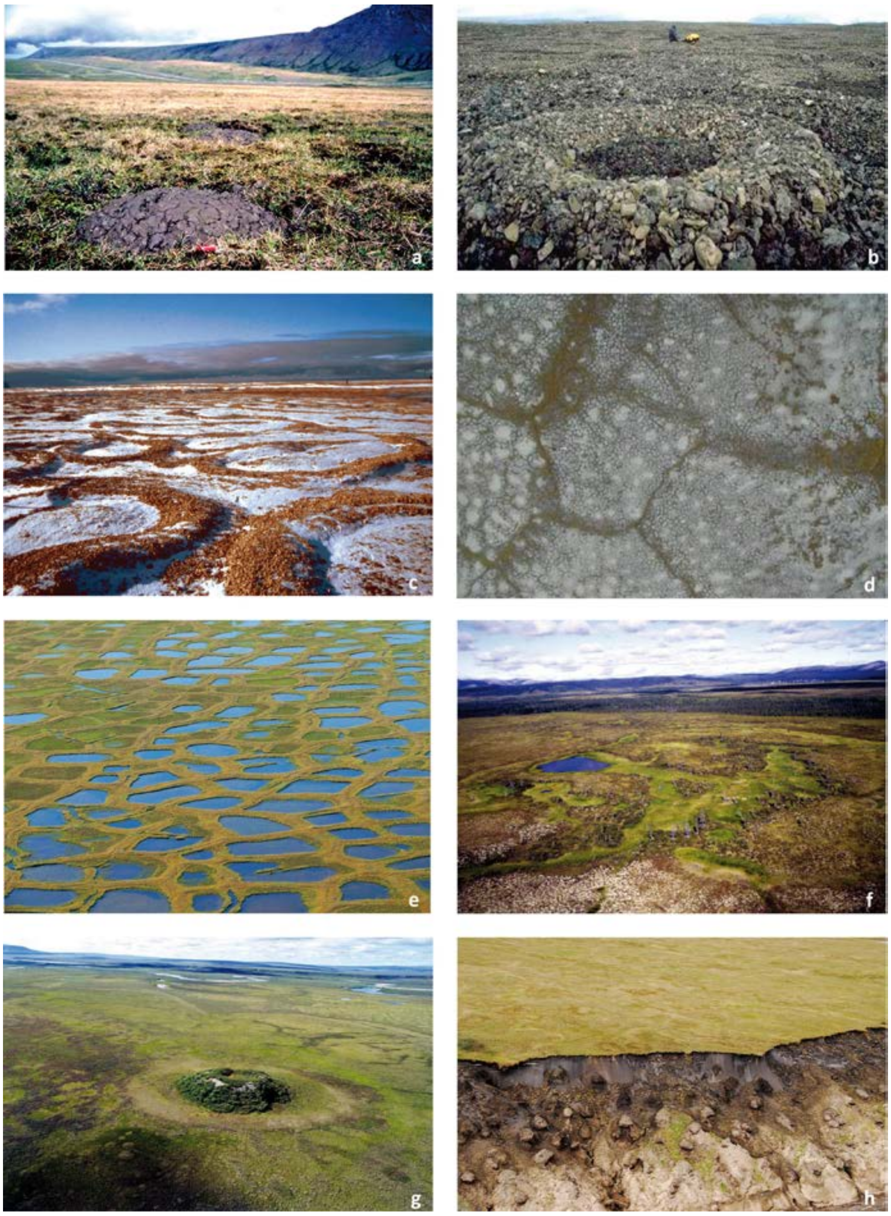

Fig. 4 Examples of permafrost patterned ground, photographed at various sites. The size of the patterns increases from meters (top row) to several tens and hundreds of meters (bottom row). a: non-sorted circles (Galbraith Lake, North Slope, Alaska); b, c: circles and polygons (Brøgger Peninsula, Svalbard); d: circles and polygons (Howe Island, Alaska); e: water-filled polygons (Lena Delta, Siberia); f: thermokarst lake; g: pingo (Seward Peninsula, Alaska); h: Yedoma sediments with high ground ice content, exposed by erosion (Lena Delta, Siberia).

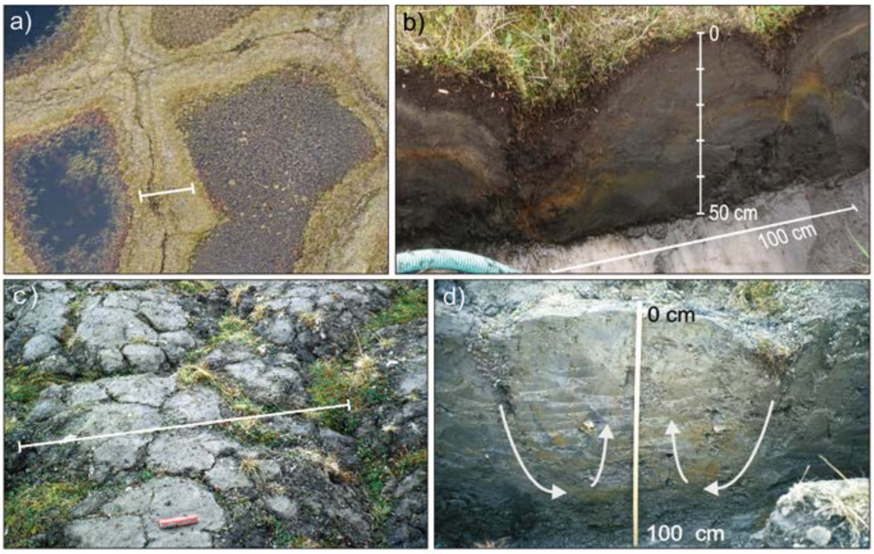

Fig. 5 Examples of the effect of cryoturbation on the surface and subsurface. a. tundra site with ice wedge polygonal pattern (Samoylov, Lena Delta, Siberia). The polygons are about 10-20 meters across. b. soil profile across a polygonal rim showing thawed soil and frozen ice wedge below. The effect of cryoturbation is visible by the organic material pulled downwards and the distortion of the horizontal soil layers. The blue tube is used to drain water from the otherwise water-logged profile. c. non-sorted circles (Bayelva, Svalbard). The diameter of the circles is about 1 meter. d. soil profile across a sorted circle. Cryoturbation results in downward migration of material at the edges of the circle and upward migration in the center.

Fig. 6 Example of drastic landscape changes caused by thawing permafrost (thermokarst) on Alaska's North Slope during August 2004. a. helicopter aerial photo (courtesy of L. Hinzman), b.-e. ground-based photos. It is not clear what caused the thaw and subsequent collapse of the surface, which was followed by retrogressive erosion of the stream. In the short term, sediment and solute transport was significantly enhanced into the nearby Toolik lake (~10 km to the north) affecting fish habitats (Bowden et al. 2008). Over the long term, the landscape surface and drainage characteristics were changed.

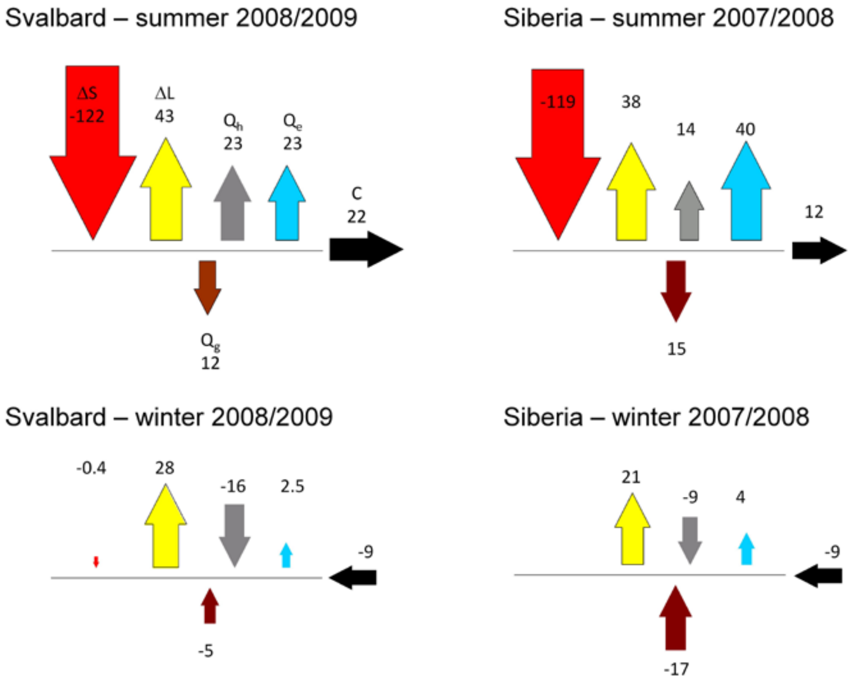

Fig. 7 Comparison of typical mean summer (upper two figures) and winter (lower two figures) energy fluxes (W m^{-2}) for the Svalbard and Siberian sites. Fluxes are scaled relative to each other. ΔS: net shortwave radiation, ΔL: net longwave radiation, Q_h: sensible heat flux; Q_e: latent heat flux; Q_g: ground heat flux; C: closure term (Databases: Westermann et al. 2009; Langer et al. 2011 a:b).

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