Landscape controls of CH_4 fluxes and soil organic matter in a catchment of the forest tundra at the lower Yenissej

Heiner Flessa[1,*], Andrej Rodionov[2], Jens Dyckmans[3], and Georg Guggenberger[2]

[1]*Institute of Soil Science and Forest Nutrition, Georg-August-University Göttingen, Büsgenweg 2, 37077 Göttingen, Germany*
[2]*Institute for Soil Science and Plant Nutrition, Martin-Luther-University Halle-Wittenberg, 06114 Halle, Germany*
[3]*Centre for Stable Isotope Research and Analysis, Georg-August-University Göttingen, Büsgenweg 2, 37077 Göttingen, Germany*

[*]*Corresponding author: hflessa@gwdg.de*

1. INTRODUCTION

Soils of boreal and arctic ecosystems store an essential part of the total terrestrial carbon pool (Hobbie et al., 2000) and they affect atmospheric methane concentration by net-CH_4 exchange (Wahlen and Reeburgh, 1992). Climate warming will have a great impact on carbon storage and CH_4 production in high-latitude soils, especially in areas where melting of permafrost is induced. Permafrost soils can act as a sink for atmospheric CH_4 but they can also show CH_4 emission (Liblik et al., 1997). The dynamics of soil moisture as well as the amount, composition, and bioavailability of organic matter stored in permafrost soils are important factors determining the net-fluxes of CH_4 at the soil surface (Christensen, 1999). The knowledge how climate warming will change these factors and how these changes will influence the net-fluxes of CH_4 is of decisive importance for the prediction of the future role of these ecosystems in the global CH_4 cycle. However, little is known about soil organic matter stocks in permafrost soils and the characteristics of frozen carbon pools in particular. A better knowledge of organic substrate quality is required to predict the response of permafrost soils to changes in temperature and moisture (Hobbie et al., 2000).

The objectives of our study were i) to determine the temporal and spatial variability of net-CH_4 fluxes in a small catchment of the forest tundra in Siberia, ii) to analyze the soil and landscape controls of the net-exchange of CH_4, iii) to quantify soil organic matter stocks in the active layer and in the frozen subsoils of the catchment, and iv) to determine the potential bioavailability and DOC production of soil organic matter stored in this area.

2. MATERIALS AND METHODS

2.1 Study area

The study area was the Grawijka Creek catchment (size of 0.44 km^2) approximately 10 km north of the town of Igarka (Russian Federation) at 67°29.90' latitude and 86°25.26' longitude on the east shore of the Yenissej river. The region is situated in the transition zone from the moderately continental western Siberian climate to the strongly continental eastern Siberian climate. In nearby Igarka the mean annual air temperature is -7.4 °C and the mean annual precipitation is 510 mm with about half of it falling during the summer months July, August, and September. Snow cover usually extends from early October to late May.

Symptom of Environmental Change in Siberian Permafrost Region,
Eds. Hatano R and Guggenberger G, p 65-74, Hokkaido University Press, Sapporo, 2006

Parent materials in the catchment area are Quaternary sediments derived from the Karginskaya interstage (59 - 24 kyrs BP) of the last glaciation underlain by sediments from the Zyryansk stage (122 - 59 kyrs BP) (Sachs, 1948). The Karginskaya terrace is of glaciofluvial to glaciolimnic genesis and characterized by silty deposits. The site is located in the forest tundra ecotone. The vegetation is a mixed forest dominated by *Larix sibirica*, *Picea obovata*, *Betula pubescens*, and *Pinus sibirica*. Stands are relatively dense in areas with lacking or low permafrost, whereas the areas with high permafrost are characterised by open stands. Ground vegetation is characterized by *Vaccinium myrtillus*, *Vaccinium uliginosum*, *Rubus chamaemorus* and various grasses on well drained soils and is dominated by *Sphagnum* mosses at wet sites.

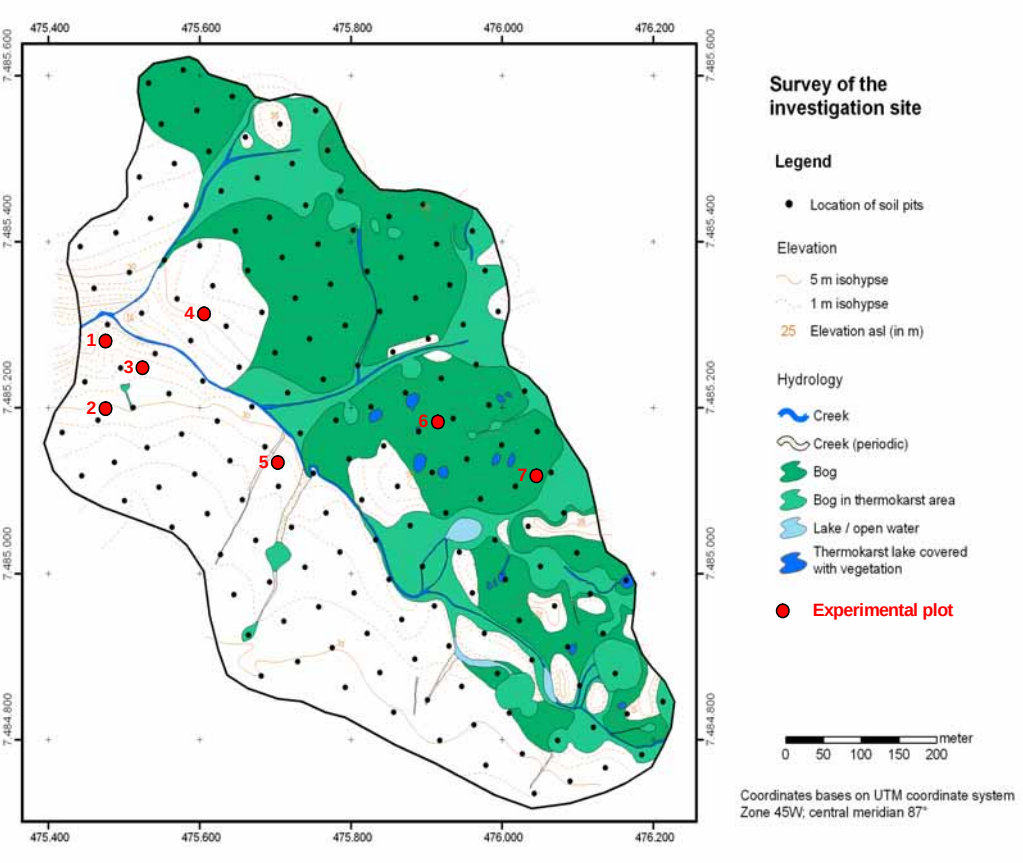

Fig. 1. The Grawijka Creek catchment: Land cover, soil sampling grid, and location of experimental plots for CH_4 flux measurements.

2.2 Soil sampling and analyses

Based on a 50 × 50 m grid mapping of soils, vegetation, and permafrost distribution (Rodionov et al., 2005) we established 7 experimental plots covering typical landscape units within the Grawijka Creek catchment (Fig. 1). The plots represent the typical regional variability of vegetation, soil type, exposition, and depth of seasonal thawing (Table 1). A detailed description of the experimental area is given by Rodionov et al. (2005). Soil samples were

Table 1. Topography and soils (soils type, thickness of active layer, soil organic carbon (SOC) and total nitrogen N_t) of selected experimental plots established within the Grawijka Creek catchment.

Plot	Exposition and slope	Soil type / aquatic system	Active layer* (cm)	SOC in 0-100 cm (kg m^{-2})	N_t in 0-100 cm (kg m^{-2})
P1	North slope	Dystrochryept	> 200	6.5	0.4
P3	North slope	Aquorthel	65	27.0	2.1
P4	South slope	Dystrochryept	> 200	7.9	0.6
P5	North slope	Aquiturbel	40	48.0	2.6
P6	Plateau within the bog area	Fibristel	40	135.7	6.9
P7	Depression within the bog area	Thermokarst lake			

* measured from the surface of the humus layer; at P6 and P5 it includes the layer of living moos

collected in summer 2003 from the active layer (by hand) and the frozen subsoils down to a depth of 2 m (using a gasoline-driven corer). Soil bulk density was determined using undisturbed soil cores. Soil organic carbon (SOC) and total nitrogen (N_t) were measured using a C/N analyzer (Vario EL, Elementar, Hanau, Germany). The ^{14}C acitivity of SOC was measured at the accelerator mass spectrometer of the Leibniz Labor, Kiel, Germany. The radiocarbon age in years B.P. was calculated according to Stuiver and Polach (1997).

2.3 CH_4 flux measurement

Net-fluxes of CH_4 were measured at each experimental plot (except plot 2) in four replicates using the closed chamber technique (Flessa et al., 1995). We employed dark chambers, each covering a surface of 706.5 cm^2. The accumulation time varied from 10 minutes on plots showing CH_4 emission to 1 hour for plots showing CH_4 uptake. Four gas samples were taken from the chamber air at equal sampling intervals using evacuated glass bottles. The gas samples were analysed using a gas chromatograph with a flame ionisation detector (Shimadzu GC-14A, Japan). The gas sampling procedure and the set up of the gas analysis system was described in detail by Loftfield et al. (1997). The CH_4 flux rate was calculated from the slope of the temporal change in CH_4 concentration within the closed soil cover. CH_4 fluxes were measured during the summer and the beginning winter in 2003. Flux rates were expressed as means with standard deviation (n = 4). Cumulative CH_4 fluxes calculated for the total experimental period were subjected to a Student-Newman-Keuls test after running a Kruskal-Wallis one way analysis of variance on ranks to demonstrate significant differences between the experimental plots.

2.4 Soil incubation

Selected soil samples from the active layer and from frozen subsoils were incubated at 18°C and well aerated conditions in the laboratory (three replicates per sample). Soil samples were air-dried to a soil moisture that allow sieving to < 2 mm, then they were rewetted and stored for 3 months at 5°C. Five to 34 g of soil (dry weight, depending on the C_{org} content of the

Table 2. Soil samples used in the incubation experiment. (n.a. means "not available")

Plot	Soil depth (cm)	Active layer or permafrost	SOC ($g\ kg^{-1}$)	N_t ($g\ kg^{-1}$)	^{14}C age (yr)
P1	0 - 10	Active layer	27.3	1.5	n.a.
P3	0 - 12	Active layer	63.8	4.8	905
P3	95 - 115	Permafrost	17.7	1.5	15731
P4	0-10	Active layer	37.7	2.4	n.a.
P5	0-20	Active layer	235.0	12.0	n.a.
P5	70 - 100	Permafrost	11.4	0.6	n.a.
P6	20 - 40	Active layer	352.0	23.0	2210
P6	70 - 100	Permafrost	445.0	18.6	2720

samples) were placed in 250 ml Nalgene filter units with upper and lower polycarbonate beakers and with a built-in filter unit consisting of a glassfibre prefilter and a 0.45µm nylon filter. These filtration units were used to collect soil leachate after the addition of de-ionized water and, after closing the upper chamber, to analyse CO_2 emission from soil. After a first leaching of the soil with 100 ml of de-ionized water, the samples were incubated for 192 days. Emission rates of CO_2 were measured five times during the incubation period. The upper beakers were closed for one day to allow for CO_2 accumulation and then a gas sample was taken from the headspace of the incubation beakers by a gas-tight syringe. Measurement of CO_2 concentration was performed on a gas chromatographic system (GC) as described by Loftfield et al. (1997). Emission of CO_2 was calculated from the gas concentration in the headspace of the beaker before and in the end of the accumulation period. Leaching of dissolved organic carbon (DOC) was determined four times during the incubation period by adding 100 ml of de-ionized water. DOC was determined by a TOC analyser (TOC 5050, Shimadzu, Japan). The selected soil samples used in this incubation experiment are described in Table 2.

3. RESULTS AND DISCUSSION

3.1 Permafrost distribution

The experimental area is characterized by a high variability of the seasonal thaw depth of the soils. About 40% of the catchment had soils with an active layer <90 cm (Fig. 2, Rodionov et al. (2005)). Depth of thawing was low in particular in the bog soils with thick layers of living and dead mosses and in soils of the NNE slope. In contrast, most SSW exposed soils and soils located at steeper inclinations showed a seasonal thaw depth >90 cm. The spatial variability of the thickness of the active layer in the catchment was influenced mainly by i) exposition (influencing insolation), ii) inclination (influencing insolation and drainage), and iii) the occurrence of an isolating moss layer (reducing heat flow into the soils) (Rodionov et al., 2005). Active layer thickness in general is a result of a complex interaction of regional (e.g. temperature, snow cover) and local factors (e.g. topography, vegetation, soil properties) (Stendel and Christensen, 2002; Tarnocai et al., 2004). Our results indicate that the permafrost regime in the Grawijka Creek catchment is highly sensitive to increasing temperature since small differences of exposition and inclination considerably influenced the thickness of the active layer.

The soils of the experimental plots differed in permafrost dynamics and drainage (Table 1). The soils of P1 and P4 completely thawed during the summer and were well drained. The soils of plot P3 and P5 had an active layer thickness of less than 70 cm. These soils had gleyic properties since the shallow permafrost prevented percolation. Hydromorphic properties were more distinct at P5 than at P3 because of the smaller depth of seasonal thawing at P5 (40 cm).

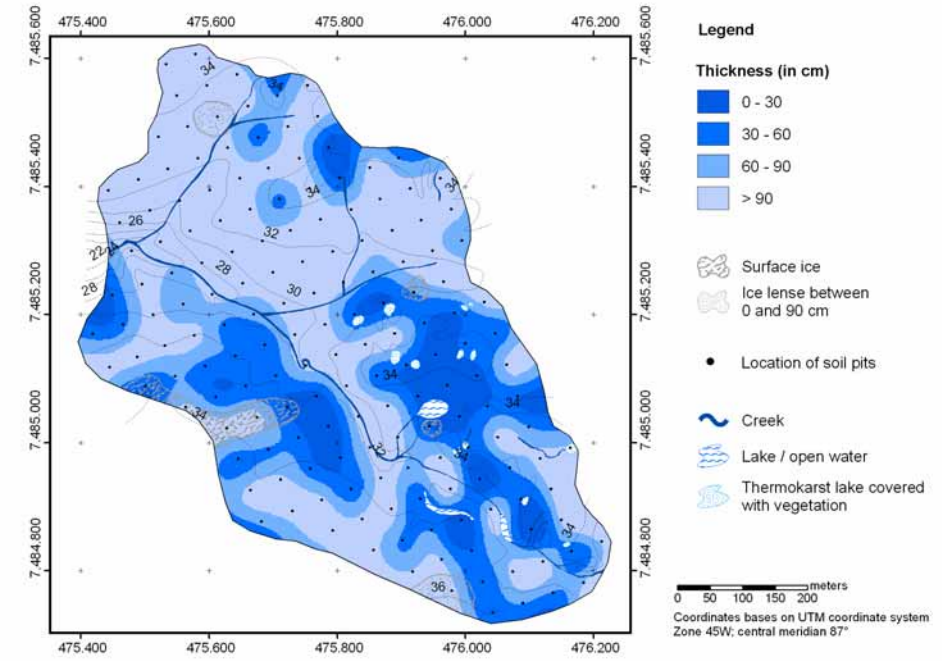

Fig. 2. Thickness of the active layer of the soils of the Grawijka Creek catchment (Figure from Rodionov et al. (2005)).

Plot P6 was located within a large raised bog area. Most of the bog soils had an active layer thickness <40 cm. Plot P7 represented a thermokarst lake. These small lakes in our experimental area were a result of permafrost degradation and formed small depressions within the bog area. Water depth at P7 was about 2 m. The lake was completely covered by a swimming vegetation layer.

3.2 CH_4 fluxes

In summer of 2003, nearly all soils of the catchment showed a net-CH_4 uptake. The uptake rates ranged from about 10 to 60 µg CH_4 m^{-2} h^{-1}. This range agrees well with the CH_4 uptake (0.5 – 2.0 mg CH_4 m^{-2} d^{-1}) reported for well-drained soils in various natural ecosystems (King, 1997). In all soils, CH_4 uptake was greatest in August when air and soil temperature were highest and soil moisture was lowest (Fig. 3 and 4). The CH_4 uptake activity decreased with decreasing temperature, however, it was still significant in October when air and soil temperatures dropped below 0°C. This indicates that the atmospheric CH_4 oxidizing microbial population was well adapted to the low temperature of this area.

CH_4 uptake rates were significantly higher in soils with an active layer >200 cm than in soils with permafrost in the upper meter of the soil profile (Fig. 3 and 4). The mean CH_4 uptake rate calculated for the whole study period was 37 µg m^{-2} h^{-1} for the soil of P1 and P4, 22 µg m^{-2} h^{-1} for P5, 11 µg m^{-2} h^{-1} for P3 , and 18 µg m^{-2} h^{-1} for the organic soil at P6. Higher CH_4 uptake was associated with lower soil moisture. In August, the mean gravimetric water content in the Ah horizon increased in the order P4 (16%) < P1 (40%) < P3 (48%) < P5 (105%). The results suggest that permafrost distribution and the related soil moisture dynamics are important factors controlling the net-uptake of CH_4 in the soils of this region. Shallow permafrost induced gleyic soil properties and reduced CH_4 diffusion into the soil. Additionally, the thick moss layer present on these wet permafrost soils may have retarded gas diffusion.

The cumulative CH_4 fluxes from August to November are shown in Fig. 5. The mean total CH_4 uptake in the soils without permafrost (P1, P4) was about twice the size as the mean CH_4 uptake in soils with a thawing depth <70 cm (P3, P5, P6). This difference between soils with deep and shallow permafrost was highly significant.

The thermokarst lake (plot P7) was a strong source of atmospheric CH_4 (Fig. 5). The formation of thermokarst in this catchment is mainly driven by permafrost degradation within the bog area. The reasons of thermokarst formation are multiple and include fire, increasing temperature and precipitation (Harris, 2002; Agavonov et al., 2004). The emission of CH_4 from these aquatic systems is fueled by erosion of bog material into the thermokarst pond. We found a mean CH_4 emission of 6.5 g CH_4 m^{-2} during our study period. This indicates that an area of about 70 ha of well-drained soils with a thawing depth >100 cm or 160 ha of soils with an

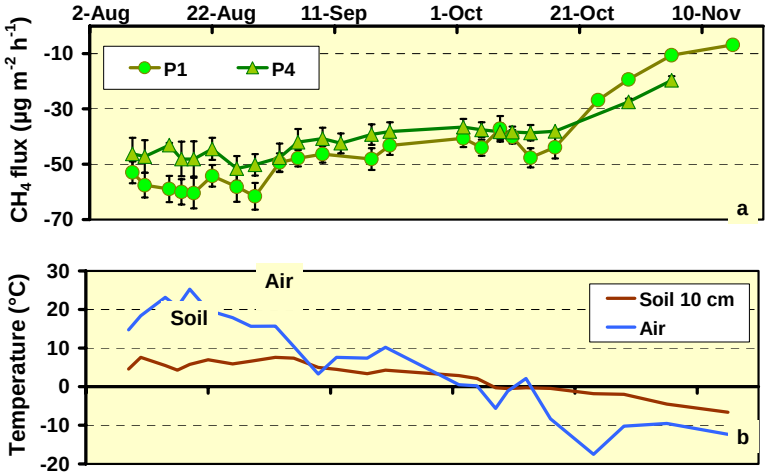

Fig. 3. Time course of a) the net-flux of CH_4 (mean and standard deviation, n=4) measured at two experimental plots (P1, P4) with an active layer >200 cm and b) the soil (depth of 10 cm) and air temperature measured during the flux measurements.

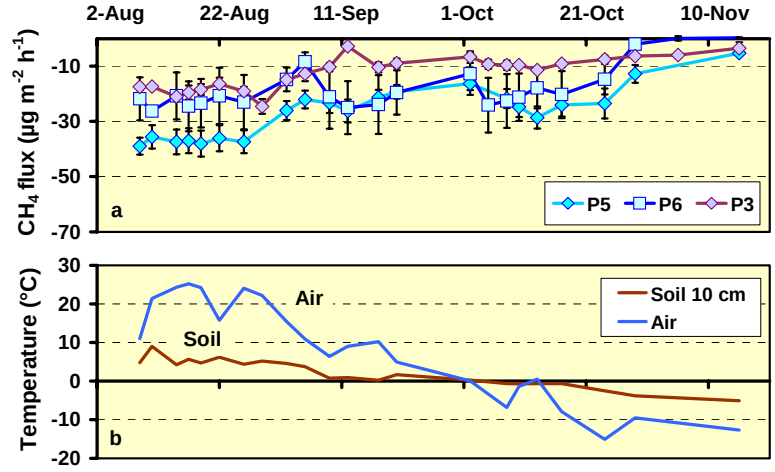

Fig. 4. Time course of a) the net-flux of CH_4 (mean and standard deviation, n=4) measured at three experimental plots (P3, P5, P6) with an active layer <70 cm and b) the soil (depth of 10 cm) and air temperature measured during the flux measurements.

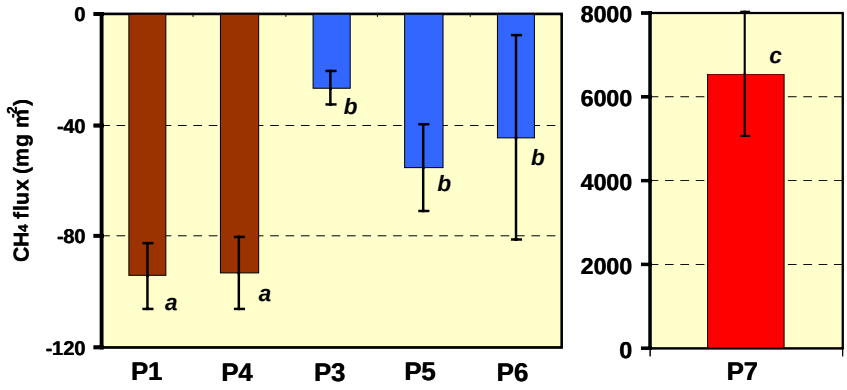

Fig. 5. Cumulative fluxes (means and standard deviation) of CH_4 from August to November 2003 on the experimental plots within the Grawijka Creek catchment. Different letters indicate significant differences ($P< 0.05$).

active layer <70 are necessary to counterbalance the CH_4 emission from thermokarst lakes with a total area of 1 ha. The results suggest that the ongoing formation of thermokarst lakes induced by permafrost collapse in bog areas has probably changed this region from a sink to a source of atmospheric CH_4. An increase in CH_4 emission induced by permafrost melt was also observed in a peatland area in north-central Saskatchewan (Turetsky et al., 2002). It was found that permafrost degradation was associated with a 30-fold increase in CH_4 emission due to the creation of wet "internal lawns" which were strong sources of CH_4. High CH_4 emission from lakes in permafrost regions were also reported by Zimov et al. (1997). They found an annual emission from North Siberian lakes of 6.8 g CH_4 m^{-2} and showed that CH_4 emission was fueled by old Pleistocene organic carbon. Heikkinen et al. (2001) reported that summer CH_4 emission from thermocarst lakes in the Russian tundra of northeast Europe were about 3.2 g CH_4 m^{-2}. Similar results (~21 mg CH_4 m^{-2} d^{-1}) were reported for thermocarst lakes and ponds in artic Alaska by Whalen and Reeburg (1990). Walter et al. (2004) found even much higher emission rates from thermocarst lakes in northeast Siberia (up to 10 g CH_4 m^{-2} d^{-1}) indicating that the bioavailability of the thawed organic carbon stocks can vary considerably.

These results show that the consequences of climate change in permafrost regions and in particular in peatlands which have been accumulating organic carbon and nitrogen for several thousands of years may be dramatic. Permafrost degradation in peatlands alters local topography, hydrology, thermal regimes and plant communities and thus will change or even reverse the net-fluxes of CO_2 and CH_4 of these areas (Turetsky, 2004). However, the long-term development of thermokarst lakes in different areas and its effects on the net-exchange of carbon and nitrogen is still not clear.

3.3 Soil organic matter stocks

Large stocks of organic carbon and total nitrogen (N_t) were found in the soils of the catchment. Total SOC in the upper meter of the soil profile ranged from 6.5 to 129 kg m^{-2} and N_t stocks varied from 0.4 to 6.3 kg m^{-2} (Table 1). SOC and N_t storage was largest in the organic permafrost soil of P6 and smallest in the soils without permafrost at P1 and P4. The occurrence of permafrost influenced the preservation of SOC and N_t in the active layer and in the subsoil. In the permafrost itself, soil organic matter was preserved by persistent frost. In the active layer of soils with high permafrost, gleyic properties retarded organic matter decomposition. The frozen SOC stocks in the upper meter of the soils were 8 kg m^{-2} at P5, 11

kg m^{-2} at P3, and 103 kg m^{-2} in the bog soil at P6. The size of SOC stocks in the Grawijka Creek catchment fits well to studies on organic matter in mineral soils and peatlands of other permafrost regions (Ping et al., 1997; Michaelson et al., 1996; Bockheim et al., 1999). Differences of soil organic carbon and nitrogen storage between soils were even more pronounced when soils were sampled down to a depth of 2 m (data not shown). Related to this depth total SOC was about 330 kg m^{-2} in the bog soil (P6), about 60 kg m^{-2} in soils with shallow permafrost (P3, P5), and about 10 kg m^{-2} in soils without permafrost (P1, P4). The deep coring revealed that the major part of organic carbon stored in the Grawijka Creek catchment is located in the frozen subsoils. The results indicate, that SOC storage estimates in permafrost areas may be considerably underestimated if deep soil C is not taken into account. Since large areas exist where permafrost will completely disappear from soils (Stendel and Christensen, 2002), the frozen stocks of deep SOC may become bioavailable and influence the response of permafrost soils to changes in temperature and moisture with respect to gaseous C fluxes and leaching of nutrients and DOC.

3.4 Potential CO_2 and DOC release of soils

The origin and composition of organic matter stored in the active layer and the frozen subsoils varied considerably in our experimental area. Highly decomposed, humified OM with a low C/N ratio was found in the frozen subsoil of P3 (e.g. Table 2, P3, depth 95 to 115 cm). The low radiocarbon concentration revealed that this deep SOM formed in the late Pleistocene about 16 kyr ago. The frozen deep soil OM stocks in the bog (P6) consisted of less decomposed plant remains. Bog formation took place in the mid Holocene. The ^{14}C age of frozen organic

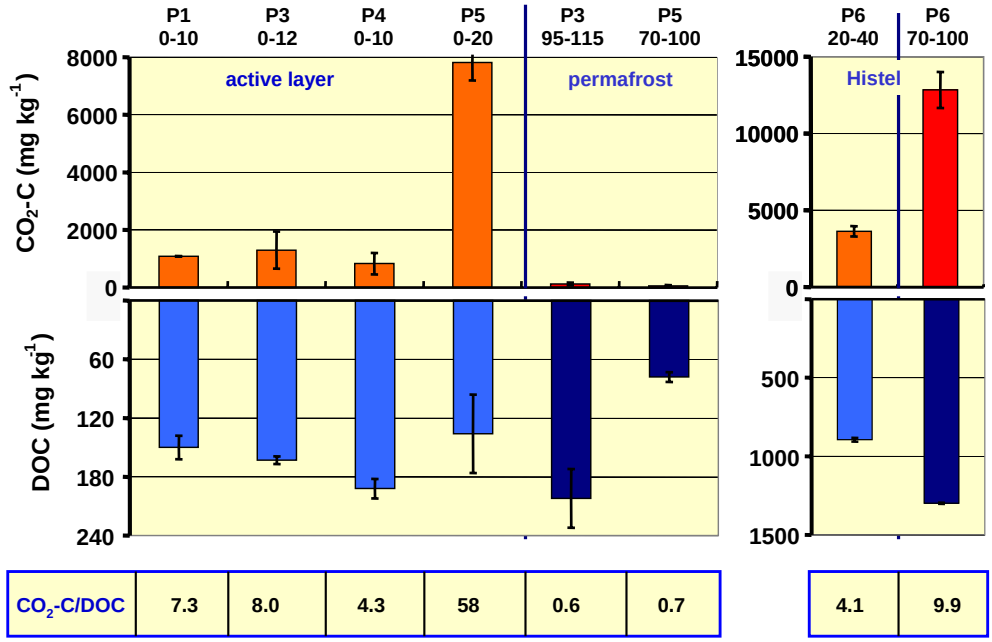

Fig. 6. Cumulative emission of CO_2 and leaching of DOC (means and standard deviation, n=3) of selected soil samples from the experimental plots of the Grawijka Creek catchment during an incubation of 192 days at 18 °C and the calculated ratio CO_2-C/DOC (properties of the soil samples are listed in Table 2). The samples were taken from the active layer and the frozen subsoils of the experimental plots.

carbon in the depth of 70 to100 cm was 2.7 kyr (Table 2, P6, depth 70 to 100 cm).

We analyzed the potential SOC mineralization and DOC leaching of soil samples taken from the active layer and from frozen subsoils of the Grawijka Creek catchment (Table 2). The samples were incubated at 18°C for 192 days. Potential soil respiration was highest for the bog subsoil and for the SOC-rich A horizon of the gleyic permafrost soil of P5 (Fig. 6). In contrast, the soil samples from the mineral subsoils of P3 and P5 showed the lowest CO_2 emission. The emission of CO_2 was related with the SOC concentration, however, the specific bioavailability of SOC (CO_2 emission per unit SOC) was significantly higher in the A horizons than in the mineral subsoils of P3 and P5. The cumulative CO_2-C emission indicated that about 2.0 to 5.5% of the initial SOC stock was mineralized in the A horizons, whereas it was only 0.5 to 0.7% in the mineral subsoils. In the bog soil however, specific bioavailability of SOC was higher in the subsoil (2.9% of SOC) than in the active layer (1.6% of SOC).

Leaching of DOC from the mineral soil samples was less variable than soil respiration (Fig. 6). It ranged from 78 to 202 mg DOC kg^{-1} (total DOC flux during the incubation). Much higher DOC leaching (894 and 1290 mg DOC kg^{-1}) occurred from the bog soil samples of P6.

The ratio of CO_2 production to DOC leaching varied considerable among soil samples (Fig. 6). DOC fluxes were about 13% to 24% of the total CO_2-C flux for soil samples from the A horizons (except for the A horizon of P5 where DOC/CO_2-C was lower) and for samples from the bog soil (Figure 6). In contrast, samples from the frozen subsoils of P3 and P5 showed higher DOC release than CO_2-C emission. This suggests that the frozen mineral soils of the experimental area contain large stocks of dissolved organic carbon with relatively low bioavailability.

The results show that quality of organic matter in the active layer and in the frozen subsoils and its potential for CO_2 and DOC production at higher temperatures under well drained conditions vary widely in our experimental area. Thus, it can be expected that the soils of the catchment differ with respect to their response to climate warming. Permafrost collapse in the bog area will result in a fast release of vast amounts of easily available organic matter that will quickly lead to high losses of CO_2, CH_4 and DOC. The old organic matter stocks in the permafrost of the mineral soils will initially release large amounts of stored DOC but the mineralization of these SOC stocks will be much slower than in the bog soils.

4. CONCLUSION

The results show that the sink strength of soils for atmospheric CH_4 was influenced by permafrost distribution and the related soil moisture regime. Additionally, high CH_4 emissions from thermokarst lakes suggest that the Grawijka Creek catchment has changed from a net-sink to a net-source of atmospheric CH_4. Permafrost degradation will probably increase CH_4 uptake rates in mineral soils but the total landscape flux will be dominated by CH_4 emission fueled by anoxic degradation of bog material in thermokarst lakes. Thus, the formation of thermokarst lakes and the further development of these aquatic systems are key factors determining the role of this area in the global CH_4 cycle.

The local variability of soil organic matter stocks and their different potential to serve as substrate for oxic and anoxic microbial activity or DOC leaching will strongly influence the response of this area to climate change. More information especially about organic matter in the upper permafrost and the development of soil hydrology is necessary to elucidate the future role of this area as a possible source of CO_2, CH_4 and DOC.

Acknowledgement. This study was funded by the Deutsche Forschungsgemeinschaft (DFG).

REFERENCES

Agafonov L, Strunk H, Nuber T (2004) Thermokarst dynamics in western Siberia: insight from dendrochronological research. *Palaeogeogr. Palaeoclimatol. Palaeoceol.* **209**, 183-196.

Bockheim JG, Everett LR, Hinkel KM, Nelson FE, Brown J (1999) Soil organic carbon storage and distribution in Arctic tundra, Barrow, Alaska. *Soil Sci. Soc. Am. J.* **63**, 934-940.

Christensen TR (1999) Potential and actual trace gas fluxes in arctic terrestrial ecosystems. *Polar Res.* **18**, 199-206.

Flessa H, Dörsch P, Beese F (1995) Seasonal variation of N_2O and CH_4 fluxes in differently managed arable soils in Southern Germany. *J. Geophys. Res.* **100**, 23115-23124.

Harris SA (2002) Causes and consequences of rapid thermokarst development in permafrost or glacial terrain. *Permafrost Periglac. Process.* **13**, 237-242.

Hobbie SE, Schimel JP, Trumbore SE (2000) Controls over carbon storage and turnover in hgh-latitude soils. *Global Change Biol.* **6**, 196-210.

King GM (1997) Responses of atmospheric methane consumption by soils to global climate change. *Global Change Biol.* **3**, 351-362.

Liblik LK, Moore TR, Bubier JL, Robinson SD (1997) Methane emissions from wetlands in the zone of discontinuous permafrost: Fort Simpson, Northwest Territories, Canada. *Global Biogeochem. Cycles* **11**, 485-494.

Loftfield N, Flessa H, Augustin J, Beese F (1997) Automated gas chromatographic system for rapid analysis of the atmospheric trace gases CH_4, CO_2 and N_2O. *J. Environ. Qual.* **26**, 560-564.

Michaelson GJ, Ping CL, Kimble JM (1996) Carbon storage and distribution in tundra soils of Arctic Alaska, USA. *Arct. Alp. Res.* **28**, 414-424.

Ping CL, Michaelson GJ, Kimble JM (1997) Carbon storage along alatitudinal transect in Alaska. *Nutr. Cyc. Agroecosys.* **49**, 235-242.

Rodionov A, Flessa H, Grabe M, Mayer T, Bussemer S, Kasansky O, Guggenberger G (2005) Controls of organic carbon and total nitrogen variability in permafrost-affected soils at the forest tundra ecotone (lower Yenissej, Siberia). *Ecosystems*, submitted

Sachs VN (1948) The Quaternary in the Soviet Arctic. Trudy Arkicheskogo Nauch.-issled. Instituta 201 (in Russian).

Stendel M, Christensen JH (2002) Impact of global warming on permafrost conditions in a coupled GCM. *Geophys. Res. Letters* **29** (13), 10.

Tarnocai C, Nixon FM, Kutny L (2004) Circumpolar-Active-Layer-Monitoring (CALM) sites in the Mackenzie valley, northwest Canada. *Permafrost Periglac. Process.* **15**, 141-153.

Turetsky MR (2004) Decomposition and organic matter quality in continental peatlands: the ghost of permafrost past. *Ecosystems* **7**, 740-750.

Turetsky MR, Wieder RK, Vitt DH (2002) Boreal peatland C fluxes under varying permafrost regimes. *Soil Biol. Biochem.* **34**, 907-912.

Walter K, Draluk DA, Zimov SA, Chanton JP, Chapin FS (2004) Global warming feedbacks of methane bubbling along expanding north siberian lake margins. p.202, abstract in: *Climate Disturbance Interactions in Boreal Forest Ecosystems, Fairbanks, Alaska, May 2004*.

Whalen SC, Reeburgh WS (1992) Interannual variationsin tundra methane emissions: A four-year time series at fixed sites. *Global Biogeochem. Cycles* **16**, 181-211.

Zimov SA, Voropaev YV, Semiletov IP, Davidov SP, Prosiannikov SF, Chapin FS, Chapin MC, Trumbore S, Tyler S (1997) North Siberian lakes: A methane source fuelled by Pleistocene carbon. *Science* **277**, 800-802.