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Methane emission from Siberian arctic polygonal tundra: eddy covariance measurements and modeling

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Abstract

Eddy covariance measurements of methane flux were carried out in an arctic tundra landscape in the central Lena River Delta at 72°N. The measurements covered the seasonal course of mid-summer to early winter in 2003 and early spring to mid-summer in 2004, including the periods of spring thaw and autumnal freeze back. The study site is characterized by very cold and deep permafrost and a continental climate with a mean annual air temperature of −14.7 °C. The surface is characterized by wet polygonal tundra, with a micro-relief consisting of raised moderately dry sites, depressed wet sites, polygonal ponds, and lakes. We found relatively low fluxes of typically 30 mg CH_4 m^{-2} day^{-1} during mid-summer and identified soil temperature and near-surface atmospheric turbulence as the factors controlling methane emission. The influence of atmospheric turbulence was attributed to the high coverage of open water surfaces in the tundra. The soil thaw depth and water table position were found to have no clear effect on methane fluxes. The excess emission during spring thaw was estimated to be about 3% of the total flux measured during June–October. Winter emissions were modeled based on the functional relationships found in the measured data. The annual methane emission was estimated to be 3.15 g m^{-2}. This is low compared with values reported for similar ecosystems. Reason for this were thought to be the very low permafrost temperature in the study region, the sandy soil texture and low bio-availability of nutrients in the soils, and the high surface coverage of moist to dry micro-sites. The methane emission accounted for about 14% of the annual ecosystem carbon balance. Considering the global warming potential of methane, the methane emission turned the tundra into an effective greenhouse gas source.

Keywords: carbon balance, eddy covariance, methane, tundra

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Introduction

Approximately 24% of the Northern Hemisphere's exposed land area is underlain by permafrost (Zhang *et al.*, 1999). Permafrost is a globally significant carbon reservoir, although estimates of its size vary. Post *et al.* (1982) estimates arctic tundra environments to account for 190 Gt or 13–15% of the global soil organic carbon pool. More recent studies suggest a carbon content of 500 Gt in frozen yedoma sediments alone and an additional 400 Gt in nonyedoma permafrost excluding peatlands, which exceeds the carbon content of the atmosphere (730 Gt) and that of vegetation (650 Gt) (Zimov *et al.*, 2006). Because of the high sensitivity of high-latitude ecosystems to climate changes, as well as their large proportion of the earth surface, these landscapes are critically important for the Earth System, in particular for the global carbon cycle (Chapin *et al.*, 2000).

Northern wetlands and tundra are a major source of methane, contributing about 20% of the annual natural emissions (Fung *et al.*, 1991; Cao *et al.*, 1996; Christensen *et al.*, 1996). With growing concern about climate change and the need to quantify emissions on a large scale, the greenhouse gas (GHG) budget of arctic wetlands have

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come into the focus of attention. Because methane has a 25-fold global warming potential compared with carbon dioxide (time horizon of 100 years) (IPCC, 2007), it has a strong influence on the GHG budgets of these landscapes (Friborg *et al*., 2003; Corradi *et al*., 2005). Furthermore, global climate models rely on predictions of future GHG concentrations, which require the ability to accurately model sinks and sources of methane as a powerful GHG.

However, there is still much uncertainty about the source strength and the driving forces of methane flux of tundra landscapes. Existing studies of high-latitude methane fluxes were mostly based on the closed-chamber technique. Due to the high temporal and spatial variability of methane fluxes (Christensen *et al*., 1995, 2000; Wagner *et al*., 2003; Kutzbach *et al*., 2004), this technique alone does not give reliable information on landscape scale fluxes. In addition, during chamber measurements the soil surface is isolated from the atmosphere so that the coupling of atmosphere and methane emission cannot be studied. The eddy covariance technique provides nonintrusive spatially integrated flux data at the landscape scale. However, to our knowledge only three studies reported eddy covariance methane flux data from arctic tundra ecosystems, namely Fan *et al*. (1992) from Alaska, Friborg *et al*. (2000) from Greenland, and Hargreaves *et al*. (2001) from Finland.

Here, we present the first eddy covariance methane flux data from a Siberian arctic tundra landscape. The objective of this study was to quantify the methane emission over the full course of the 'active' season from early spring to early winter, to analyze the contribution of different parts of the vegetation period, particularly the little studied periods of spring thaw and soil refreeze, to identify the biological and physical parameters which control the methane fluxes, and to estimate the annual methane emission. Together with the fluxes of carbon dioxide, which were measured concurrently and analyzed elsewhere (Kutzbach *et al*., 2007), a comprehensive picture of the GHG budget of the tundra was gained.

Materials and methods

Study site

The investigation site was located on Samoylov Island in the Lena River Delta at 72°22′N, 126°30′E (Fig. 1). During the last years, Samoylov Island has been the focus of several studies in the field of microbiology, soil science, and surface–atmosphere fluxes of carbon, energy and water (Hubberten *et al*., 2006). The Lena River Delta is located in the zone of continuous permafrost

Fig. 1 Satellite image of the Lena River Delta (Landsat 7 ETM + GeoCover 2000, NASA); the location of the investigation area Samoylov Island is marked by a white square.

with permafrost temperatures between −11 and −13 °C (Kotlyakov & Khromova, 2002). Samoylov Island is situated in the southern central part of the river delta, approximately 120 km south of the Arctic Ocean. The central delta region has a dry continental arctic climate, which is characterized by very low temperatures and low precipitation. The 30-year (1961–1999) averages of annual air temperature and precipitation measured at the meteorological station in Tiksi about 110 km east of Samoylov Island are −13.6 °C and 319 mm, respectively (ROSHYDROMET, 2004). Data from the meteorological station on Samoylov Island from the period 1999–2005 show a mean annual air temperature of −14.7 °C and a highly variable total summer precipitation (rain) between 72 and 208 mm (mean 137 mm). Typically, the ground is snow-covered between the end of September and the beginning of June (Boike *et al*., 2008), and the growing season lasts from June to August. During spring, summer, and autumn, the weather in the central delta region is characterized by the rapid change between the advection of arctic cold and moist air masses from the north and continental warm and dry air masses from the south.

The flux measurements were carried out on the eastern part of Samoylov Island, which is characterized by wet polygonal tundra (Fig. 2). This part of the island represents the Late-Holocene river terrace which is one of the main geomorpholocical units in the Lena River Delta, occupying about 65% of the total delta area (Are & Reimnitz, 2000). The eastern part of Samoylov Island has a level surface with slope gradients <0.2%, and an elevation of 10–16 m a.s.l. It is not flooded during the annual spring flood. Larger elevation differences up to

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2.5 m occur only along the shorelines of the large lakes. However, the surface of the terrace is structured by a regular micro-relief with elevation differences of up to 0.5 m within a few meters distance, which is caused by the genesis of low-centered ice wedge polygons (Meyer, 2003). In the depressed polygon centers, drainage is impeded by the underlying permafrost, hence the soils are water-saturated and small ponds frequently occur. In contrast, the elevated polygon rims are characterized by a moderately moist water regime. The typical soil types are Typic Historthels in the polygon centers and Glacic or Typic Aquiturbels at the polygon rims. The vegetation in the polygon centers and at the edge of ponds is dominated by hydrophytic sedges (*Carex aquatilis*, *Carex chordorrhiza*, *Carex rariflora*) and mosses (e.g. *Limprichtia revolvens*, *Meesia longiseta*, *Aulacomnium turgidum*). The vegetation on polygon rims is dominated by mesophytic dwarf shrubs (e.g. *Dryas octopetala*, *Salix glauca*), forbs (e.g. *Astragalus frigidus*), and mosses (e.g. *Hylocomium splendens*, *Timmia austriaca*). Aerial photography in July 2003 and subsequent surface classification showed that the surface fraction taken by moderately moist to dry micro-sites, wet micro-sites, and open water bodies in the area surrounding the flux tower was about 60%, 10%, and 30%, respectively (Schneider *et al*., 2006).

Fig. 2 CORONA satellite image of Samoylov Island, taken during the spring flood in June 1964. The white star marks the position of the flux tower; the rectangle indicates the area displayed by the aerial photograph in Fig. 6.

Experimental setup

Eddy covariance measurements of methane flux were carried out in the periods July 19–October 22, 2003 (96 days), and June 1–July 21, 2004 (51 days). The eddy covariance system was set up at a central position of the eastern part of Samoylov Island (Fig. 2). Wet polygonal tundra of the river terrace extended for 600 m around the tower, with several large lakes protruding into the periphery of the otherwise homogeneous fetch area. The wind vector and sonic temperature were measured with a three-dimensional sonic anemometer (Solent R3, Gill Instruments Ltd, Lymington, UK) at a height of height 3.65 m. From a sample intake 15 cm below the anemometer measurement point, the sample air was drawn at a rate of 20 L min^{-1} through a CO_2/H_2O gas analyzer (LI-7000, LI-COR Inc., Lincoln, NE, USA), a membrane gas dryer (PD-200T-48SS, Perma Pure Inc., Toms River, NJ, USA), and the methane gas analyzer, all of which were housed in a temperature regulated case at the foot of the tower. The methane gas analyzer was a tunable diode laser spectrometer (TGA100, Campbell Scientific Inc., USA). The diode laser was cooled by a closed cycle cryogenic system (Cryo-Tiger, APD Cryogenics, USA). The TGA 100 required a constant flow of a reference gas (0.5% CH_4), which was supplied by calibrated gas bottles. The calibration of the TGA 100 was checked using a zero gas and a span gas (25 ppm CH_4) whenever the reference gas bottles had to be exchanged (about every 5 weeks) or the measurement parameters were adjusted. During the calibration intervals, we observed no appreciable span drift in the methane measurements. The data from the gas analyzers were sampled by the anemometer at a rate of 20 Hz and logged on a laptop PC running the software EDISOL (J. Massheder, University of Edinburgh, UK).

The tower was equipped with additional instruments for the measurement of air temperature and relative humidity (MP103A, ROTRONIC AG, Switzerland), incoming and outgoing solar and infrared radiation (CNR1, Kipp and Zonen B.V., the Netherlands), and barometric pressure (RPT410, Druck Messtechnik GmbH, Germany). Measurements of the water level were carried out at three points in the vicinity of the flux tower at intervals of 1–3 days. Precipitation, snow height, and soil temperature data were taken from the long-term monitoring station, which is situated about 700 m south-west of the flux tower (Boike *et al*., 2008). The thaw depth was measured by probing the soil with a steel rod at 150 regularly spaced grid points near the long-term monitoring station at intervals of 3–7 days.

Calculation and validation of fluxes

Data analysis was done using the software EDIRE (R. Clement, University of Edinburgh, UK). Two coordinate rotations were performed on the wind components measured by the sonic anemometer, so that the mean

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transverse and vertical wind components were reduced to zero for each averaging period (McMillen, 1988). The mean absolute value of the angle of the second rotation was $1.0 \pm 0.9°$, hence the error introduced to turbulent fluxes by the rotation should be well below 10% for most measurements (Foken & Wichura, 1996). The time lag between wind and methane concentration measurements was determined and removed for each averaging period.

The co-spectra of the fluxes of sensible heat and methane (Fig. 3) show the characteristic features of the surface-layer turbulence spectrum and closely follow the theoretical spectrum for sensible heat flux under unstable atmospheric conditions (Moore, 1986). The co-spectrum of sensible heat follows the expected power law in the inertial sub range, whereas in the methane flux co-spectra the attenuation of high frequencies due to the limited frequency response of the gas analyzer and experimental setup is visible. These observations agree well with reports of the performance of the TGA 100 gas analyzer by other investigators (Billesbach *et al.*, 1998; Laurila *et al.*, 2005).

During the measurement campaigns, a short-term drift in the methane concentration measured by the TGA 100 was observed frequently. This drift has been described before as a changing concentration offset caused by optical interference fringes (Billesbach *et al.*, 1998). The analysis of methane power spectra revealed a strong increase in signal intensity at frequencies $<10^{-2}$ Hz. As this feature was not observed in the spectra of the other scalar's time series, it was attributed to the concentration drift. In order to suppress the drift components in the signal, a recursive high pass filter with a filter constant of 10 s (cut-off wavelength 63 s) was applied to the methane concentration time series before the flux calculation.

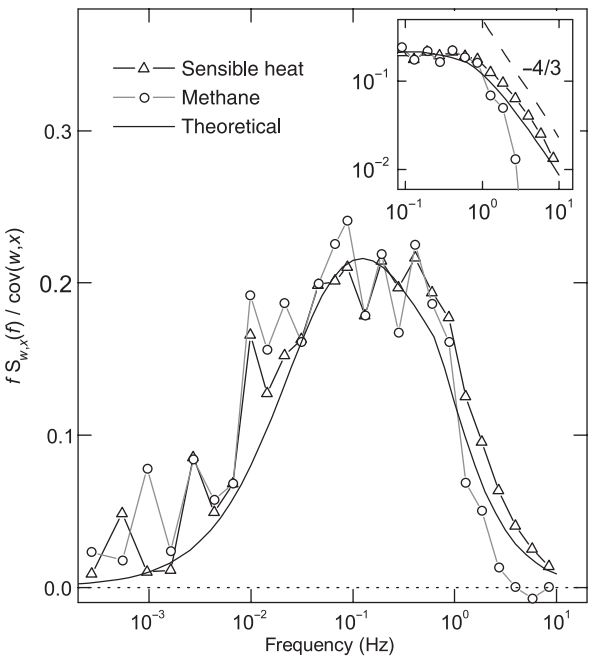

Fig. 3 Co-spectra of sensible heat and methane fluxes and theoretical spectrum. The co-spectra are averages of four hourly co-spectra (August 18, 2003, 10:00–14:00 LT) scaled by the corresponding covariance cov(w,x). The theoretical spectrum is the model of the sensible heat flux co-spectrum for the average meteorological conditions of the 4 h period (wind speed $u = 4.4\,\mathrm{m\,s^{-1}}$, stability $z/L = -0.019$). The inset shows the high frequency part of the spectra on a logarithmic scale ordinate.

A correction was applied to the calculated methane flux to account for the mismatch of the turbulence frequency spectrum and the spectral response of the measurement system. In detail, the correction compensated for the effects of the spectral response of the gas analyzer, the separation of the anemometer and gas analyzer sampling points, the line attenuation in the sample tubing, and the detrending filter (Moore, 1986; Moncrieff *et al.*, 1997). On an average, 40% were added to the calculated flux, of which 25% and 13% were related to the effect of the spectral response of the gas analyzer and the high pass filtering of the methane signal, respectively.

As an estimate of the error associated with each individual flux data point, the standard deviation of the cross correlation function of vertical wind speed and methane concentration at time shifts 100–200 s was calculated. This method accounts for the Gaussian error of the individual measurements of wind speed and methane concentration, as well as the uncertainty in the stationarity during the averaging period (Kormann *et al.*, 2001). The 30 min flux time series (screened for instrument malfunctions) showed frequent large outliers and an overall signal-to-noise ratio (SNR) of only 3 (mean flux $12.1\,\mathrm{mg\,m^{-2}\,day^{-1}}$, mean flux error $4.0\,\mathrm{mg\,m^{-2}\,day^{-1}}$). The reasons for this were thought to be the generally low methane fluxes and high wind speeds at our site, and an insufficient suppression of concentration drift in the methane concentration signal. Stronger high pass filtering of the methane signal did not seem appropriate, because this would also attenuate signal components at the low frequency end of the turbulence spectrum. However, by increasing the averaging time, the statistics of the averaging process for the drift components in the methane signal could be improved (Billesbach *et al.*, 1998). Hence, the averaging interval for the flux calculation was set to 60 min. This measure led to an increase of the SNR to 4.5 (mean flux error $2.7\,\mathrm{mg\,m^{-2}\,day^{-1}}$). The 60 min flux time series (Fig. 4) still shows considerable scatter and some negative values. Considering the distribution of wet and dry micro-sites within the eddy covariance fetch and the

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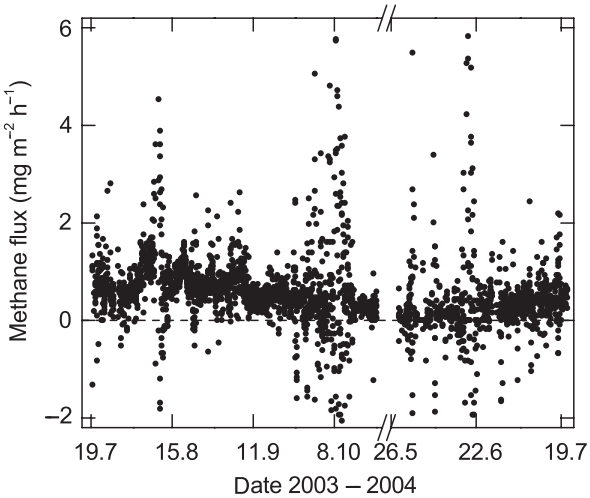

Fig. 4 Time series of hourly measured methane fluxes, screened for instrument malfunctions, but not SNR-screened ($N = 2823$).

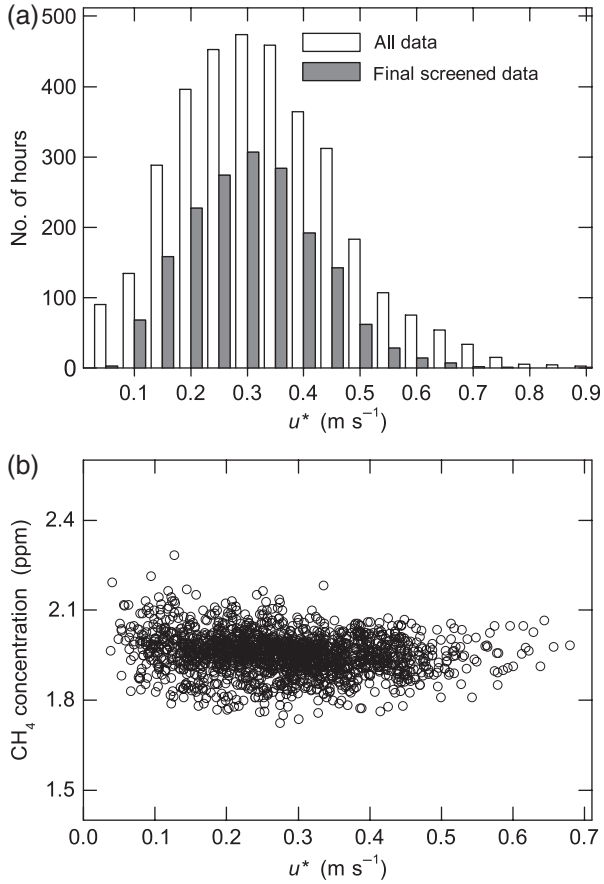

Fig. 5 Turbulence characteristics during measurements campaigns 2003 and 2004: (a) frequency distribution of hourly friction velocity u^* for all data points ($N = 3451$) and final screened data ($N = 1770$); (b) mean hourly methane concentration vs. friction velocity of final screened data.

results of previous static chamber methane flux measurements (Wagner *et al.*, 2003), large negative fluxes of methane are not anticipated in the studied tundra ecosystem. Other investigators have attributed observations of large negative methane fluxes to measurement uncertainties particularly during periods of low flux (Rinne *et al.*, 2007), and have ignored these data points or even included them in further analysis. We think that these negative flux values need to be removed by rigorous data screening based on metrological and micro-meteorological criteria. We used a filter based on the SNR of the flux calculation, which rejected all data points showing any signal larger then the flux peak in the cross correlation function of vertical wind and methane concentration. This filter effectively removed all negative fluxes and much of the scatter in the flux time series.

Finally, the data was screened using an integral turbulence characteristics test following Foken & Wichura (1996), which removed data points associated with disturbed and under-developed turbulence. After screening, there remain about 4% of data points with a friction velocity $u^* < 0.1$ (Fig. 5a). This data is often rejected in flux studies because under these conditions fluxes are deemed to be disturbed by under-developed turbulence or storage effects. Storage situations were observed a few times in 2003. However, the data associated with these events was rejected by the SNR and turbulence screening procedures. Consequently, in the screened data, an increase of methane concentration at low values of u^* which would indicate the influence of storage is not visible (Fig. 5b).

A footprint analysis following Schuepp *et al.* (1990) was carried out for the assessment of the fetch area size of the flux measurements (Fig. 6). During periods of $u^* < 0.1$, the 80% cumulative flux distance, (i.e. the upwind distance from which 80% of the observed methane flux originated), occasionally extended beyond 1000 m. Data points associated with these events were removed. Of the screened data, the 80% cumulative flux distance was on an average 477 m, and the distance of the point of origin of the maximum contribution to the measured flux was on an average 107 m. In 2003, altogether 33% of the data points were rejected by the screening procedures described above. In 2004, due to technical problems during the first half of the measurement campaign the rejection rate was 74%.

Results

Meteorological conditions

The summer and autumn of 2003 were characterized by above-average temperatures and precipitation (Fig. 7).

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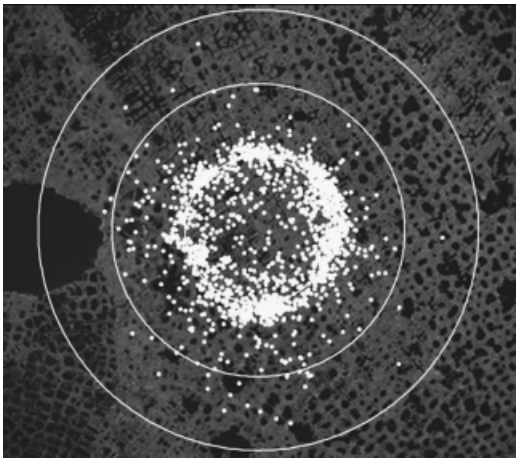

Fig. 6 Flux footprint climatology of 2003 and 2004 overlaid on an aerial image of Samoylov Island taken in July 2003. The white dots indicate the position of maximum flux contribution of screened hourly data points ($N = 1770$). The white circles indicate the distance to the flux tower in steps of 100 m.

The advection of warm continental air from the south lead to high air temperatures during the middle of July, at the beginning of August, and during large parts of September. The daily average soil temperature at 20 cm depth in a depressed, wet polygon center reached a maximum of 6.6 °C on 9 August. The isothermal state of the thawed layer was reached and refreeze of the soil began on 30 September. The soil thaw depth was 0.28 m on 15 July and increased to a maximum of 0.48 m at the beginning of September. Measurements of thaw depth stopped on 30 September due to the freezing of the top soil layer. However, the temperature profile measurements showed that the soil was not completely frozen until the middle of November. At 168 mm, the total amount of rainfall during the measurement period was exceptionally large. A great part of the rainfall occurred within 1 week at the end of July (94 mm), which caused the water table in the investigated polygons close to the eddy tower to rise well above the soil surface. Following a slow decrease, the water level stayed within ±1 cm of the soil surface after the end of August. Snow started to accumulate at the beginning of October. By the end of the measurement campaign, the snow cover had reached a height of 15–25 cm in the polygon centers and just a few centimeters on the polygon rims. The average wind speed during the measurement campaign 2003 was 4.7 m s^{-1}. There was no single predominant wind direction; however, wind directions east–north-east, south, and south-west occurred more frequently than other directions (data not shown).

When methane flux measurements started on June 1, 2004, the ground at the eddy tower site was completely covered with snow. The snow height had already started to decrease but was still 0.4–0.5 m in the polygon centers and about 0.1 m on the polygon rims. The daily average air temperature was in the range −5 to −2 °C, and the soil temperature in 20 cm depth was −12 °C. The snow thaw period started on 8 June with the occurrence of the first significant rainfall and the air and soil temperatures reaching 0 °C. The snow height decreased rapidly, and the polygon rims were largely free of snow after 2 days. Snow thaw in the polygon centers continued until 18 June. The thaw of large ice bodies of polygon ponds and lakes started towards the end of the snow thaw and lasted until about 25 June. At the end of the measurement campaign, daily average air and soil temperatures (20 cm depth) were around 8 and 1.5 °C, respectively, and the soil thaw depth had reached about 30 cm (linear extrapolation from measurements). The water table in the polygon centers was generally higher than in 2003 and never fell below the soil surface. The total rainfall up to 21 July was 60 mm. The average wind speed of the measurement period in 2004 was 4.7 m s^{-1}. Unlike 2003, there was a clear dominance of easterly winds, followed by winds from north-westerly directions (data not shown).

Methane flux

In the first week of measurements in 2003, methane fluxes were on an average 23 mg m^{-2} day^{-1} (Fig. 7). During the following cold and rainy period, the fluxes dropped markedly but subsequent warming and further thawing of soils lead to the highest fluxes of on average 30 mg m^{-2} day^{-1} being measured during the second week of August. During periods with high wind speed, the methane flux increased greatly compared with fluxes during calm periods directly before and after, as for instance on August 10, 2003. After the middle of August, the measured methane flux showed a general slow decreasing trend until the end of the measurement campaign. No marked influence of the freezing of the top soil layer at the end of September on the methane flux was visible. Average fluxes measured during the first week of October and during the last week of measurements, when snow had accumulated on the ground, were 13 and 7 mg m^{-2} day^{-1}, respectively.

The average methane flux from the beginning of measurements in 2004 until the end of the snow thaw on 18 June was 11 mg m^{-2} day^{-1}. However, the variation in the flux data during this time was large. Low flux values of about 4 mg m^{-2} day^{-1} occurred frequently throughout this period, but at the beginning of the snow melt methane fluxes of about 30 mg m^{-2} day^{-1} were repeatedly measured. The high variability of the

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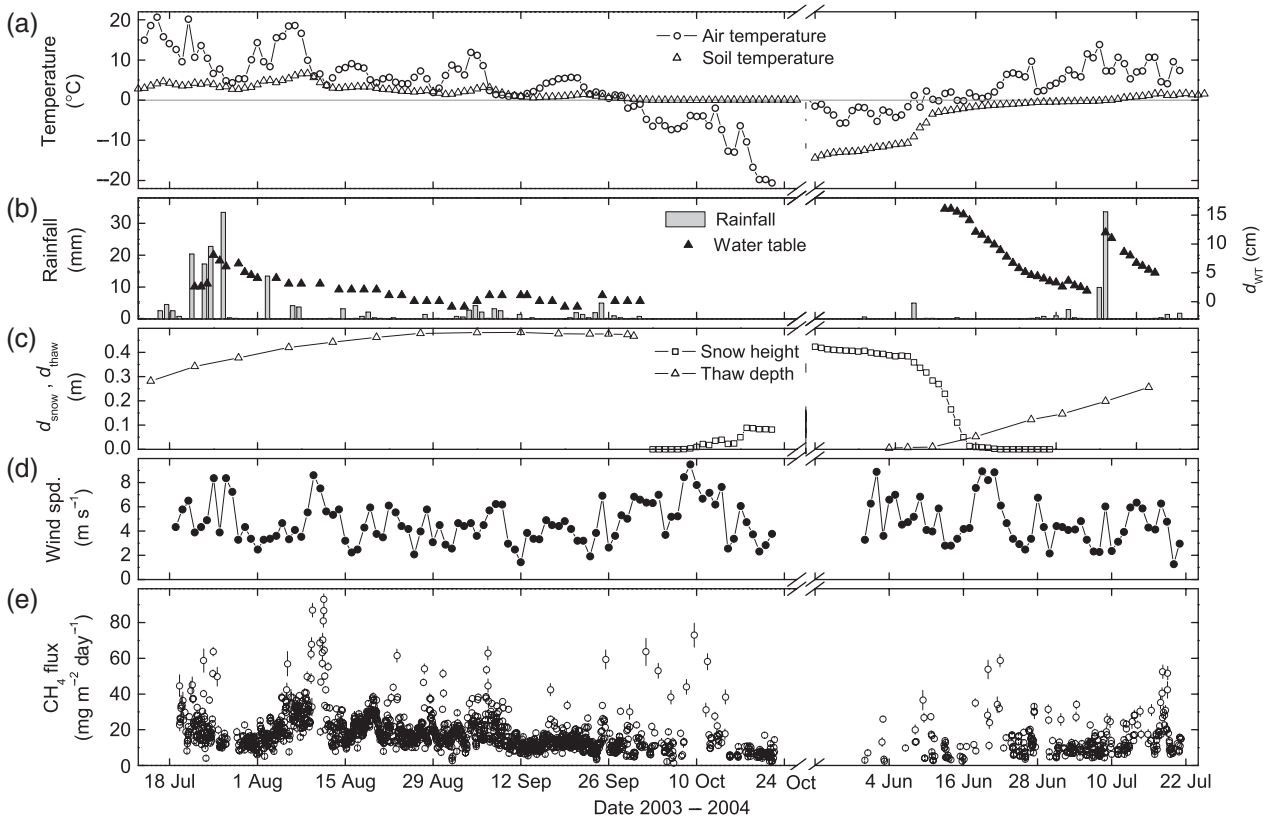

Fig. 7 Data of measurement campaigns July 19–October 22, 2003 and June 1–July 21, 2004: (a) air temperature, and soil temperature in a wet polygon center at 20 cm depth; (b) rainfall, and depth of water table with respect to soil surface in a depressed polygon center; (c) snow height in a depressed polygon center, and soil thaw depth; (d) wind speed from sonic anemometer measurements at 3.65 m height; (e) screened hourly methane flux as measured by eddy covariance ($N = 1770$).

fluxes continued until about 28 June. After this date, the fluxes stabilized at an average 10 mg m^{-2} day^{-1} and increased slowly to an average 17 mg m^{-2} day^{-1} during the last week of measurements.

In order to gain a functional relationship between methane flux and environmental drivers, which could be used for gap-filling and extrapolation, the correlation between calculated fluxes and environmental variables was studied. For this analysis, only days with a data coverage >33% were used ($N = 91$), and daily averages of methane flux were calculated, so that the variance induced by changes of fetch size and position was reduced. Errors of daily fluxes were calculated as standard errors of the mean. We found a strong correlation between methane flux and friction velocity ($r = 0.62$) and soil temperature in a wet polygon center at 20 cm depth ($r = 0.67$; Fig. 8). These two variables were included in a model following the approach of Friborg *et al*. (2000). A good agreement ($R^2 = 0.74$) with measured data was found when the methane flux was modeled using the equation

$$\mathrm{FCH_4} = a \times b^{[(T-T_{\mathrm{ref}})/10]} \times c^{(u^*-u^*_{\mathrm{ref}})}, \qquad (1)$$

where $\mathrm{FCH_4}$ is the measured methane flux, a, b, and c are the parameters determined by the fit process, T is the soil temperature, u^* is the friction velocity, and T_{ref} and u^*_{ref} are the mean values of the respective variables during the measurement period (Table 1). There was no correlation between methane flux and soil thaw depth or water table position, and expanding the model to include these variables did not improve the fit. Measured fluxes and those modeled by Eqn (1) agree well over the whole range of measured flux values (Fig. 9), except for the very windy day 11 August (mean wind speed 7.5 m s^{-1}), when the model significantly underestimates the measured flux.

Equation (1) was used for gap-filling of the daily flux time series (Fig. 10). The error of modeled daily fluxes was calculated as the root-mean-square of the fit residuals (3.91 mg m^{-2} day^{-1}). The error of cumulative fluxes was calculated by standard error propagation techniques for a 99% confidence limit. The resulting cumulative methane emission for the combined measurement period June 1–July 19, 2004 and July 20–October 22, 2003 was 2.38 ± 0.09 g m^{-2}.

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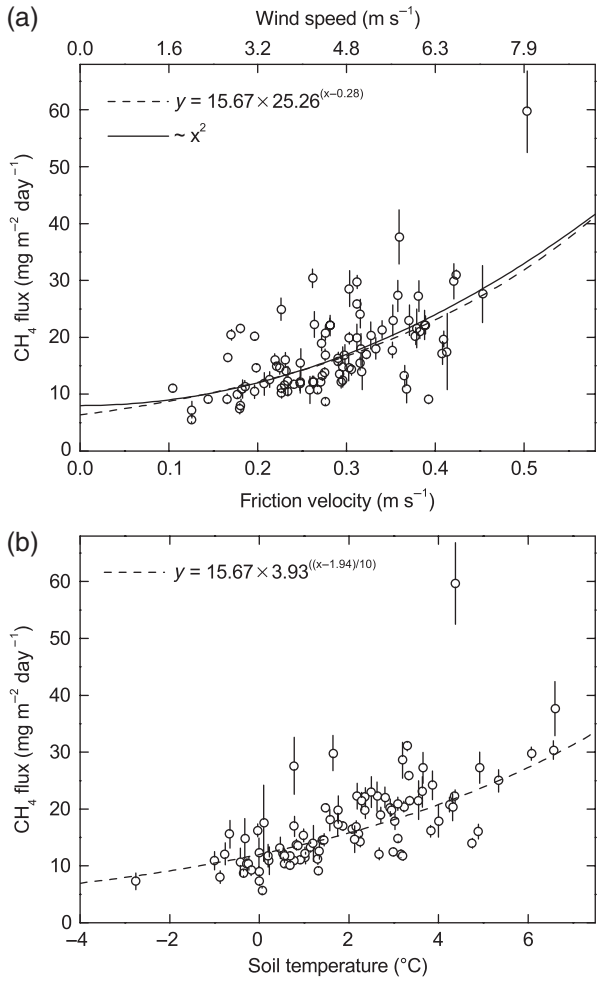

Fig. 8 Relationship between average methane flux and environmental parameters of days included in modeling ($N = 91$). (a) Methane flux vs. friction velocity. The top axis shows the approximate wind speed calculated from the linear regression of friction velocity u^* and wind speed u ($u = 15.84\ u^*$, $R^2 = 0.82$). (b) Methane flux vs. soil temperature at 20 cm depth in a wet polygon center. The dashed lines in both diagrams are the functional relationships derived by fitting Eqn (1) to measured fluxes and environmental data.

Table 1 Input and model parameters for the combined period 2003–2004 using Eqn (1)

T_{ref} (°C)	1.94
u^*_{ref} (m s^{-1})	0.28
a (mg m^{-2} day^{-1})	15.67 ± 0.46
b	3.93 ± 0.50
c	25.26 ± 7.23
R^2	0.74
P	<0.0001

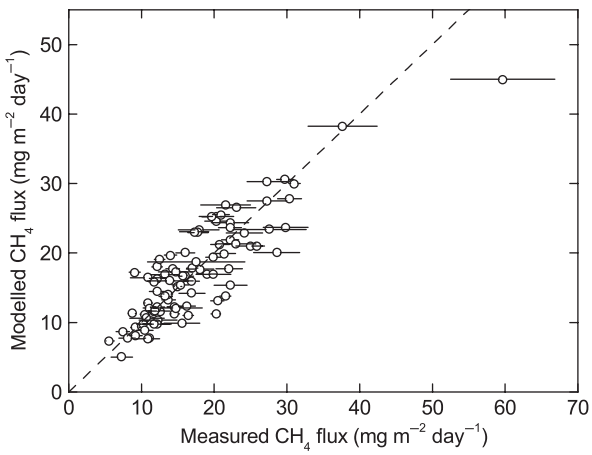

Fig. 9 Modeled flux using Eqn (1) vs. daily mean methane flux ($N = 91$).

Discussion

Drivers of methane flux

The combined measurement periods covered a whole vegetation period from spring thaw to refreeze of the soils and thus the most active period of methane emission. The wide range of environmental conditions covered allowed a detailed study of the driving forces of the methane flux. One of the two important parameters controlling methane emission was the soil temperature. The dependence of methane flux on soil temperature followed an exponential function. This reflects the fundamental dependence of soil microbiological activity on temperature (Conrad, 1989) and was confirmed by numerous studies of methane emission using closed-chamber or eddy covariance techniques (e.g. Nakano *et al.*, 2000; Christensen *et al.*, 2001; Hargreaves *et al.*, 2001).

The turbulence in the near-surface boundary layer, which correlates closely with wind speed, was the second important driving factor of the methane flux in the polygonal tundra. Fan *et al.* (1992) reported a similar effect from a mixed tundra landscape in Alaska, but only when there was a high surface coverage of lakes in the fetch area of the measurement. The exchange of gases between water bodies and the atmosphere proceeds by three primary pathways (MacIntyre *et al.*, 1995): (a) transport through emergent aquatic plants, (b) diffusive and turbulent transfer across the air–water interface, and (c) bubble ebullition. Plant-mediated transport of methane from soil layers or lake sediments to the atmosphere plays an important role in the gas exchange of wet tundra (e.g. Schütz *et al.*, 1991). Using the chamber technique, methane transport via *C. aquatilis*

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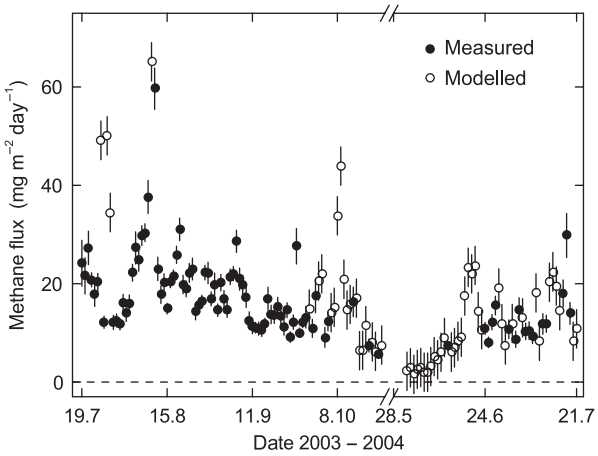

Fig. 10 Time series of measured and modeled daily averaged CH_4 fluxes during the periods July 19–October 22, 2003 and June 1–July 21, 2004.

was shown to account for between 27% and 66% of overall methane emissions from soils on Samoylov Island (Kutzbach *et al*., 2004). The same study suggested that the transport via this pathway was not limited by the above-soil diffusion resistances (leaf stomata, leaf surface boundary layer) but by the dense root exodermes. Hence, atmospheric turbulence is not likely to have a significant influence on plant-mediated transport of methane.

However, the diffusive and turbulent transfer across the water–air interface is well-known to be controlled by wind speed. The results of lake studies suggested a dependence of the gas transfer velocity on wind speed u proportional to $u^{1.6}$ (MacIntyre *et al*., 1995); studies of air–sea CO_2 exchange suggested a cubic relationship (Wanninkhof & McGillis, 1999). The flux–friction velocity relationship given by Eqn (1) is very similar to a quadratic function of u^* (Fig. 8). Hence, we hypothesize that the dependence of methane emission on atmospheric turbulence observed at our study site is in a large part due to the high surface coverage of water bodies in the polygonal tundra.

Bubble ebullition, which occurs in lakes and inundated soils, is also likely influenced by atmospheric turbulence. Gas bubbles which tend to adhere to surfaces under water could be released by wind-induced agitation of plants, wave action, and under-water turbulence. However, this process is not well understood. The observation of high methane flux during high wind speeds on August 11, 2003 may be due to turbulence-induced ebullition and indicates a possible threshold of wind speed for the triggering of this process. Ebullition could also be triggered by changes of air pressure. This mechanism has been suggested by other authors (e.g. Frolking & Crill, 1994), and was observed during a study of methane emission from lakes on Samoylov Island in 2002 (Spott, 2003). However, in our data, a relationship between air pressure and methane flux was not found.

Many studies identified the soil thaw depth of soils as an important predictor of methane emission, spatially and temporally (e.g. Friborg *et al*., 2000; Tsuyuzaki *et al*., 2001; van Huissteden *et al*., 2005). In our study, the thaw depth was not correlated with methane flux and, when added as an additional variable, did not improve the performance of flux model [Eqn (1)]. This indicates that during the warm period the majority of the methane emitted originated from the upper soil layers. The small contribution of deep soil layers to methane emissions could be explained by the temperature gradient in the thawed active layer and the temperature dependence of microbial activity. However, as microorganisms in deep soil layers were shown to be cold-adapted (Liebner & Wagner, 2006; Wagner *et al*., 2007), it is more likely caused by a substrate limitation of microbial activity due to a decreasing bio-availability of soil organic carbon with increasing depth in soils of Samoylov Island (Wagner *et al*., 2005).

The water table position is another environmental variable which was identified by many studies as a main factor controlling methane emission (e.g. Suyker *et al*., 1996; Friborg *et al*., 2000). This was explained with the regulation of the methane production/oxidation balance through the ratio of the aerobic/anaerobic soil column depth. In the spatial domain, this regulation was also observed at our study site. Concurrent measurements of methane emissions by the closed-chamber technique showed that the methane fluxes from waterlogged polygon centers were larger by a factor of 8–10 compared with emissions from elevated, moderately moist polygon rims at any time during the measurement campaigns 2003 and 2004 (unpublished data). However, temporally, no significant influence of water table on methane flux was detected. This can be explained with respect to the two micro-sites prevalent in our study area. Firstly, in the raised polygon rims, the water table was always well below the soil surface, so the ratio of aerobic/anaerobic soil column was always high. Furthermore, process studies have shown that oxidation activity in these soils is greatest near the aerobic–anaerobic interface where the substrate provision is at its optimum (Liebner & Wagner, 2006). Secondly, despite the variations in water table position, in most of the polygon centers the water table was distinctly above the soil surface during both measurement periods. Hence, at both micro-sites the change of water table position did not influence the methane production/oxidation balance significantly. However, extreme

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draught could lower the water level below the soil surface in many polygon centers and lead to increased oxidation and overall decreased methane flux. This 'on-off switch' effect (Christensen *et al.*, 2001) was observed for single polygon center sites on Samoylov Island during the dry summer of 1999 (Wagner *et al.*, 2003; Kutzbach *et al.*, 2004).

Seasonal dynamics of methane flux

Despite the low data coverage during spring 2004, a description of the processes during the thaw period can be given with reasonable confidence. Large methane fluxes were measured on several occasions during the first days of snow melt (June 8–13, 2004), which indicate the release of methane from the snow cover during the metamorphosis and settling of snow which is associated with the initial stages of the thaw process (Boike *et al.*, 2003). Furthermore, during a period with strong wind directly after the snow thaw (June 18–21, 2004), fluxes were observed which equaled those measured during the mid-summer period of 2003. These large fluxes were very likely caused by the escape of methane trapped in ice covers of ponds and lakes, which continued to thaw until at least June 25, 2004. Similar observations were made by Hargreaves *et al.* (2001) in a Finnish mire. By comparing the measured and modeled methane flux during the period June 8–25, 2004, the contribution of thaw related fluxes to the overall emission during the combined measurement period was estimated to be about 3%. This value is very similar to the spring pulse excess emission of 2.1% of annual emission estimated by Rinne *et al.* (2007) for a boreal fen.

The average methane emission of the polygonal tundra on Samoylov Island during the 'warm' months July, August, and September was 15.7, 22.3 and 15.2 $mg\,m^{-2}\,day^{-1}$, respectively. These values are at the lower end of summer emissions observed by other flux studies in arctic or sub arctic wetlands. Average methane fluxes observed by eddy covariance were 25 $mg\,m^{-2}\,day^{-1}$ during July–August at a tundra site in Alaska (61°N) (Fan *et al.*, 1992), and 38 $mg\,m^{-2}\,day^{-1}$ during August at a Finnish mire (69°N) (Hargreaves *et al.*, 2001). The total summer emission (June–August; 1.6 $g\,m^{-2}$) was about 60% lower than the value of 3.7 $g\,m^{-2}$ reported for a tundra site on Greenland (74°N) (Friborg *et al.*, 2000). The geographically closest study based on the closed-chamber method was conducted near Tiksi, about 120 km south-east of Samoylov Island, and reported a mean flux of 23 $mg\,m^{-2}\,day^{-1}$ from a tundra site during July and August (Nakano *et al.*, 2000). Generally, closed-chamber flux studies in far north-east Siberia observed significantly higher summer methane emissions: Between 105 and 196 $mg\,m^{-2}\,day^{-1}$ at floodplain sites near the Kolyma River (69°N) (Nakano *et al.*, 2000; Tsuyuzaki *et al.*, 2001; Corradi *et al.*, 2005), and 103 $mg\,m^{-2}\,day^{-1}$ at a river terrace polygonal tundra of the Indigirka River floodplain (71°N) (van Huissteden *et al.*, 2005).

There are a number of reasons which could explain the differences in methane emission observed. Soil temperature regime, vegetation cover, hydrology and texture of soils, as well as bio-availability of nutrients play an important role in determining microbial activity in the soil and the gas exchange between soils and the atmosphere. The tundra soils at our study site are characterized by a sandy texture. Sand is known to be an unfavorable habitat for microbes (e.g. Wagner *et al.*, 1999). Furthermore, the availability of nutrients is limited because the organic matter in the soils is only weakly decomposed, and there is no input of organic carbon by recent flooding. These conditions appear to impede microbial methane production at our study site compared with other sites in north-east Siberia.

During autumn and early winter, the methane emission of the tundra on Samoylov Island decreased slowly; drastic changes in response to the refreeze of the top soil layer were not observed. The emission of methane through the frozen top soil layer is thought to be facilitated by vascular plants (Hargreaves *et al.*, 2001). It was shown that transport of methane through *C. aquatilis* is independent of the phenological status of the plant tissues (Kutzbach *et al.*, 2004). Hence, although senescent, the plants keep providing a pathway for diffusion from deeper soil layers to the atmosphere. Moreover, also the accumulation of snow after the middle of October showed no marked influence on the methane flux. It was shown before that, during winter, the gas flux into the atmosphere is controlled by production in the soil, and that the snow cover acts as a passive layer only (Panikov & Dedysh, 2000; Corradi *et al.*, 2005). However, we expect the dependence of methane flux on turbulence, which was observed during snow-free periods to decrease with increasing snow cover, due to a de-coupling of the turbulence from vegetation, and soil and water surfaces. Though, owing to the lack of data, this hypothesis could not be verified.

Similarly, due to a lack of data, methane emissions during winter, (i.e. during the period of completely frozen soils and low sub-zero soil temperatures are uncertain). Many studies have stressed the importance of cold season fluxes in boreal wetlands. The contribution of cold season flux to the annual flux was reported to be 4–21% at a Minnesota peatland at 47°N (Dise, 1992), 3.5–11% at a west Siberian peat bog at 57°N (Panikov & Dedysh, 2000), 5–33% at Finnish bogs and

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fens at 62–65°N (Alm *et al*., 1999), and 23% at a Finnish mire at 69°N (Hargreaves *et al*., 2001). However, there exist few studies in permafrost regions, which address the question of winter fluxes. Whalen & Reeburgh (1992) observed methane emission during winter which accounted for about 40% of the annual flux, but which they attributed to physical processes during the soil freeze rather than microbial activity. Laboratory experiments have shown only recently that methanogenesis takes places in soils at sub-zero temperatures (Rivkina *et al*., 2004; Wagner *et al*., 2007). Based on these findings, we assume that the relationship between soil temperature and methane emission derived from measurements in 2003–2004 can be extrapolated for the estimation of winter CH_4 emission. A similar approach was chosen by Corradi *et al*. (2005) for the estimation of winter soil respiration. Using the soil temperature record from 20 cm depth and Eqn (1), but omitting the u^*-term because of the expected de-coupling of methane flux and turbulence by the snow cover, the cumulative flux of the period October 23, 2003–May 31, 2004 (222 days) was estimated to be $0.77 \pm 0.15\,g\,m^{-2}$ (Fig. 11). Based on this estimate, the integrated emission of the cold season October–May was $1.1\,mg\,m^{-2}$, and its contribution to the annual emission was 35%. This estimate is at the upper end of the range of results discussed above which can be explained with the seasonal distribution of fluxes. During summer, methane emission from the tundra of Samoylov Island is generally low compared with other wet tundra sites, while substantial emission continues well into early winter. Considering this observation, the method of estimating cold season flux as the product of measured winter flux and number of days with sub-zero air temperature, as used (e.g. by Panikov & Dedysh, 2000), is likely to systematically underestimate the contribution of the cold season to the annual flux.

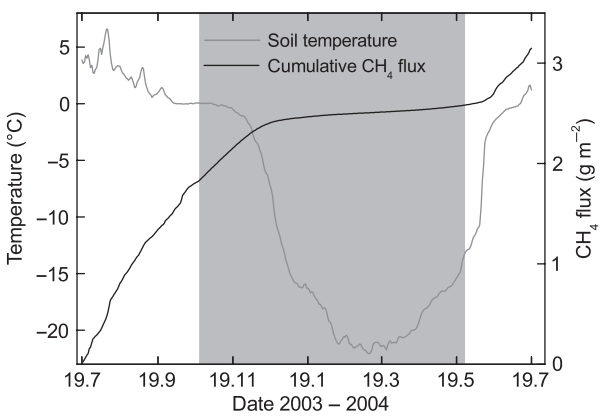

Fig. 11 Record of soil temperature at 20 cm depth at a polygon center and cumulative methane flux for the period July 2003–July 2004. The period with modeled data is shaded.

Annual carbon fluxes and GHG budget

Using the estimate of cold season flux, the annual methane flux of the tundra ecosystem during the period July 20, 2003–July 19, 2004 was $3.15 \pm 0.17\,g\,m^{-2}$. This value is about 40% lower than annual emissions given for wetlands at similar latitudes in Finland (Hargreaves *et al*., 2001) and Alaska (Reeburgh *et al*., 1998). As discussed with respect to seasonal fluxes, we hypothesize that the main reasons for the low annual flux observed are the low coverage of wet micro-habitats, the low soil temperature, and the low bio-availability of nutrients in the tundra soils of Samoylov Island.

Measurement and modeling of the fluxes of carbon dioxide showed that the tundra was a CO_2 sink of $72\,g\,m^{-2}$ during the period July 20, 2003–July 19, 2004 (Kutzbach *et al*., 2007). Thus, the overall carbon balance of the tundra was $-17.3\,g\,C\,m^{-2}$, and the methane emission accounted for about 14% of the carbon balance. A similar value of 19% was given by Friborg *et al*. (2003) for a west Siberian peat bog, which had a carbon exchange about five times as high as the tundra on Samoylov Island. The high value of 25% reported by Corradi *et al*. (2005) for a north-east Siberian tussock tundra was due to the high methane emission ($10\,g\,C\,m^{-2}$ during 60 days in summer) compared with a moderate annual carbon uptake of $-38\,mg\,C\,m^{-2}$. Considering the global warming potential of methane compared with carbon dioxide (factor 25 per unit mass for a time horizon of 100 years) (IPCC, 2007), the GHG balance of the tundra in units of CO_2 equivalents was $+6.8 \pm 4.4\,g\,m^{-2}$. Thus, although the methane emission had only a small influence on the tundra's capacity as a carbon sink, it offset the CO_2 GHG sink strength of the tundra and even turned it into a small source of greenhouse gases. Other Siberian wetlands were found to be strong GHG sources due to their emission of methane (Friborg *et al*., 2003; Corradi *et al*., 2005). Further observation of the tundra ecosystems in the Lena River Delta will be necessary to determine which way their GHG balance will go with the impact of climate change.

Conclusions

- The methane emission at the wet polygonal tundra studied was low regarding daily summer fluxes (typically $30\,mg\,m^{-2}\,day^{-1}$), as well as the annual flux ($3.15\,g\,m^{-2}$). Reason for this were thought to be the very low permafrost temperature in the study

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region, the sandy soil texture and low bio-availability of nutrients in the soils, and the high surface coverage (>50%) of moist to dry micro-sites in the polygonal tundra.

- The soil temperature and near-surface atmospheric turbulence were identified as the main factors controlling methane emission. The dependence of CH_4 fluxes on atmospheric turbulence was attributed to the high coverage of open water bodies in the polygonal tundra and demonstrates the close coupling of the soil and atmosphere systems. The variables soil thaw depth and water table position, which were often identified as (spatial) flux predictors by short-term flux studies using the closed-chamber technique were found to have no significant effect in the temporal domain.
- The relationship between methane flux and soil temperature found during the period spring–early winter was extrapolated to estimate the methane emission during the winter. At 35%, the contribution of the winter period to the annual flux was very large. This was due to the slow freezing of the tundra soils in early winter, the long cold period (October–May), and generally moderate fluxes during summer.
- During the period July 2003–July 2004, the tundra was a carbon sink of $17.3\,g\,C\,m^{-2}$, and the methane emission accounted for about 14% of the ecosystem carbon balance. Considering the global warming potential of methane compared with carbon dioxide, the GHG balance of the tundra in units of CO_2 equivalents was $+6.8\,g\,m^{-2}$. Thus, although the methane emission had only a small influence on the tundra's capacity as a carbon sink, it turned the tundra into an effective greenhouse gas source.

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