Methane fluxes during the initiation of a large-scale water table manipulation experiment in the Alaskan Arctic tundra

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[1] Much of the 191.8 Pg C in the upper 1 m of Arctic soil of Arctic soil organic mater is, or is at risk of, being released to the atmosphere as CO_2 and/or CH_4. Global warming will further alter the rate of emission of these gases to the atmosphere. Here we quantify the effect of major environmental variables affected by global climate change on CH_4 fluxes in the Alaskan Arctic. Soil temperature best predicts CH_4 fluxes and explained 89% of the variability in CH_4 emissions. Water table depth has a nonlinear impact on CH_4 efflux. Increasing water table height above the surface retards CH_4 efflux. Decreasing water table depth below the surface has a minor effect on CH_4 release once an aerobic layer is formed at the surface. In contrast with several other studies, we found that CH_4 emissions are not driven by net ecosystem exchange (NEE) and are not limited by labile carbon supply.

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1. Introduction

[2] Peatland regions of the world are sites of past accumulation of organic materials due to generally anaerobic conditions [*Harriss et al.*, 1985]. These regions could be significant current and future global sources of CH_4 and CO_2 release [*Ehhalt*, 1974; *Harriss et al.*, 1985; *Oechel et al.*, 1993; *Oechel et al.*, 1994; *Zimov et al.*, 1997; *Harazono et al.*, 2003]. The amount of carbon stored in the upper 1 m layer of Arctic tundra soil is approximately 191.8 Pg [*Post et al.*, 1982; *Schlesinger*, 1991; *Hobbie et al.*, 2000], 14% of the global soil organic carbon [*Post et al.*, 1982; *Billings*, 1987]. A more recent estimate reported the carbon stock in the first 1 m of cryosols in the northern circumpolar regions to about 268 Pg of organic carbon [*Tarnocai et al.*, 2003]. However, large amounts of soil organic carbon lie below 100 cm in both mineral and organic soils [*Sombroek et al.*, 1993; *Tarnocai*, 1994] and reserves of carbon in the upper 2 m of the world's peat soils is 679 Pg of C [*Batjes*, 1996]. The large carbon storage is attributed to low decomposition rates in saturated soils and to low temperatures [*Clymo*, 1984]. As the global climate warms, carbon stored in these regions could be released to the atmosphere [*Grulke et al.*, 1990; *Oechel et al.*, 1993; *Oechel et al.*, 1995], providing a strong positive feedback to global warming. The current release of CH_4 from northern peatlands is estimated to range from 14.1 Tg C-CH_4 [*Fung et al.*, 1991; *Bartlett and Harriss*, 1993] to 46 Tg C-CH_4 each year [*Gorham*, 1991].

[3] Recent evidence indicates a heterogeneous response of soil moisture in the Arctic to global warming. Soil drying in continuous permafrost has been reported and appears to be due to an increasing gap between potential summer evapotranspiration and summer precipitation [*Oechel et al.*, 1993; *Oechel et al.*, 2000; *Barber et al.*, 2000, *Dickson*, 2000; *Klein et al.*, 2005]. Drying in discontinuous permafrost is often due to increased drainage following permafrost degradation [*Yoshikawa and Hinzman*, 2003; *Smith et al.*, 2005; *Riordan et al.*, 2006]. On the other hand, some Arctic areas on continuous permafrost are showing increased lake numbers and/or extent [*Smith et al.*, 2005]. In addition to lake expansion, slumping and thermokarst erosion can cause new wet areas. These latter situations could result in extensive new areas of anaerobic soils. Therefore, even under scenarios of warming and drying of the Arctic, many regions underlain by continuous permafrost are likely to show increased water availability and anoxic condition in the soil in coming decades.

[4] The drying of currently wet and anaerobic areas of the Arctic will likely affect the net rates of CH_4 emission from soils [*Moore and Knowles*, 1989; *Freeman et al.*, 1993; *Funk et al.*, 1994; *Bubier*, 1995; *Nykänen et al.*, 1998]. In particular, the increase in surface soil aeration resulting from a drop in the water table can increase methanotrophy near the soil surface, and lead to the almost complete oxidation

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of CH_4 produced in deeper, anaerobic layers [*Conrad and Rothfuss*, 1991; *Updegraff et al.*, 1995; *Cao et al.*, 1996; *Bridgham et al.*, 1998]. Field and laboratory manipulations generally show a decrease in CH_4 emissions with the increase in drainage (or a lowering of the water table) [*Moore and Roulet*, 1993; *Silvola et al.*, 1996; *Oechel et al.*, 1998a; *Updegraff et al.*, 2001; *Whittington and Price*, 2006]. However, some other experiments showed contrasting results where sometimes CH_4 emissions were not affected by lowered water table [*Updegraff et al.*, 2001; *Chimner and Cooper*, 2003; *Strack and Waddington*, 2007]. These contradictory results make generalization and prediction of CH_4 release with changes in water table difficult.

[5] In addition to the effects of water table, and soil aeration, CH_4 fluxes may also be influenced by availability of labile C. Ecosystem productivity has been positively correlated with CH_4 release in a number of studies, including some in Arctic tundra [*Whiting and Chanton*, 1993; *Reeburgh*, 1996; *Nykänen et al.*, 2003]. This link could be explained by an increase in energy available to methanogens associated with increased plant productivity. Prediction of future CH_4 release requires a solid understanding of the controls on CH_4 production, including the role that substrate availability plays in limiting methanogenesis.

[6] Because trace gas fluxes are so tightly linked to soil moisture and water table in the Arctic, we initiated a large-scale manipulation in the Alaskan Arctic at the Barrow Environmental Observatory (BEO) in 2005 as part of the NSF Biocomplexity Program. The target of this manipulation includes increasing and decreasing the water table over large areas of tundra and observing the effects over the diverse microtopography of the region. Here, we present results from the initial year of the manipulation. The differences in water table reported are caused by the installation and initiation of the dikes, as well as by seasonal progression in water table, but do not reflect the full long-term effect of the planned manipulation. However, these results do show the impact of the observed variation and progression in water table on CH_4 fluxes. The Biocomplexity experiment described here is the first large-scale (1.2 km $\times$ 0.3 km) water table manipulation in the Arctic. The experiment was designed to investigate the link between CH_4 and CO_2 fluxes as affected by changes in soil water, over complex terrain.

[7] This paper focuses on the patterns and controls on CH_4 flux and the relationship between NEE and CH_4 fluxes during summer (June and July) 2007, the first year of manipulation. We also present the comparison of the two locations within the experimental area with different water table depth and soil moisture. We hypothesize that (1) water table and soil moisture are major predictors for CH_4 fluxes; an alternative hypothesis is that (2) NEE is a major predictor of CH_4 fluxes.

2. Site Description

[8] The investigation site was located in the wet sedge tundra in the northern part of the Arctic Coastal Plain which generally occurs across the northernmost region in Alaska and is distinguished by low elevation, gentle slopes, and proximity to the Arctic Ocean [*Brown*, 1967]. The area is characterized by continuous permafrost and by an average seasonal thaw (active layer) depth of about 37 cm [*Hinkel et al.*, 2001]. More than half of the surface is covered by polygonal ground [*Brown*, 1967]. Low center and high center polygons are common features of this area. The soils of the site are characterized by the presence of the main horizons: an organic-rich surface layer, a horizon of silty clay to silt loam textured mineral material, and an underlying perennially frozen organic-rich mineral layer [*Brown et al.*, 1980]. The soils are classified as Gelisols in the suborders Histels (organic soils with permafrost within 100 cm of the surface) [*Bockheim et al.*, 1999].

[9] The study site is located about 10 km east of the town of Barrow, Alaska. This 3,021 ha BEO reserve was set aside by the Ukpeagvik Iñupiat Corporation (UIC, the Barrow village native corporation), in perpetuity, for scientific research. The vegetation shows very high heterogeneity reflecting variation in microtopographic and other environmental factors and the presence of thaw lakes, frost boils and other patterned ground features [*Walker et al.*, 1998]. In particular lakes and revegetated drained lake basins represent about 20% and 50% respectively of the northern part of the Arctic Coastal Plain [*Hussey and Michelson*, 1966; *Hinkel et al.*, 2003; *Hinkel et al.*, 2005]. After the (usually) natural drainage of the lakes, as part of the thaw lake cycle [*Billings and Peterson*, 1980; *Hinkel et al.*, 2003], vegetation establishes (or reestablishes) on the exposed lacustrine sediment. The vegetation mainly consists of wet sedge meadow tundra [*Webber*, 1978; *Brown et al.*, 1980; *Billings et al.*, 1982]. In our site, a vegetated drained lake basin that drained 50–300 years ago [*Hinkel et al.*, 2003], mosses represent the major component of the living biomass (about 80% of the total living biomass in the second week of August 2006); the main vascular plant is represented by *Carex aquatilis* (average of 0.64 leaf area index), followed by *Eriophorum vaginatum* (0.04 leaf area index) and *Dupontia fisheri* (0.02 leaf area index) (P. C. Olivas and S. F. Oberbauer, unpublished data, 2006). The vegetated drained lake basin chosen for this study (1.6 km long and about 0.4 km wide at the widest point) was divided into three sections separated by two dikes (Figure 1). Dikes were built from water impermeable, interlocking rigid plastic barriers inserted in trenches dug in the frozen soil during the spring of 2007. Sand bags were placed at strategic locations to provide support for the barrier and impede any seepage. The dikes effectively slowed or prevented the water from moving from the north site to the south area toward the natural outlet for the basin and the associated drainage channel (Figure 1). The imposed treatments resulted in different water table heights between the various treatment areas, with the north area having the highest water table and the south the lowest.

3. Materials and Methods

3.1. CO_2, H_2O, and CH_4 Eddy Covariance Measurements

[10] Three eddy covariance towers for CO_2, H_2O vapor, and energy flux were installed in the vegetated drained lake in the BEO (Figure 1) in July 2005, one in each of the three

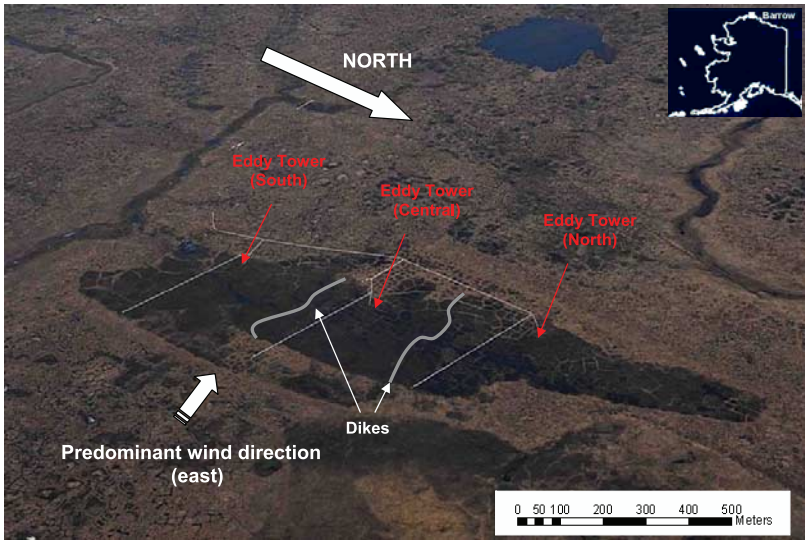

Figure 1. Site of the Manipulation experiment in the Barrow Environmental Observatory (BEO), Barrow, Alaska. Indicated are the orientation of the main axis of the vegetated drained lake (north) and the dominant wind direction (east). Two dikes (highlighted in gray) separate the three sites and prevent water from draining through the drainage channel in the southern part of the basin. Boardwalks run along and across the basin and provide access to the site avoiding disturbance.

manipulation sites (north 71°17′11.80″N 156°36′12.23″W, central 71°17′1.71″N 156°35′54.77″W and south 71°16′51.17″N 156°35′47.28″W). The eddy covariance towers operated continuously from July 2005 to the present. CO_2 and H_2O fluxes were measured through the use of an open path infrared analyzer (Li-7500, Li-COR, Lincoln, Nebraska, United States) with a sampling rate of 10Hz [*Oechel et al.*, 1998b; *Vourlitis and Oechel*, 1997, 1999; *Harazono and Miyata*, 1997]. The Li-7500 was positioned 10 cm from the center of the sonic anemometer. CO_2 and H_2O vapor were calibrated every 2 to 4 weeks using ultra-high-purity nitrogen as the H_2O and CO_2 zero and 729 ppm CO_2 in air standard gas (certified grade ± 1 ppm) (Matheson Gas Product, Montgomeryville, Pennsylvania, United States) for the CO_2 span. A dew point generator (Li-610, Li-COR, Lincoln, Nebraska, United States) was used to produce an air stream with a known water vapor dew point (typically 7°C lower than the ambient air temperature). A sonic anemometer (WindMasterPro, Gill Instruments, Limited, Lymington, Hampshire, United Kingdom) was used to measure the three wind velocity components and to determine the high-speed sonic temperature fluctuations.

[11] In addition, two closed-path DLT-100 fast response CH_4 analyzers (Los Gatos Research, Mountain View, California, United States) were added to the system in the summer 2007 for eddy covariance measurements of CH_4, one at each of the two sites studied in detail in 2007. The Los Gatos CH_4 analyzers deployed here are based on an off-axis integrated-cavity output spectroscopy using a high-flow-mode operation (100 l/min) and 10 Hz sampling rates. The Los Gatos analyzers were installed at the two towers with the largest differences in water table, the north and south towers. An electrical power line was installed to provide electrical power to the sites and to allow continuous, year-round, data collection. The two CH_4 analyzers were set up from the middle of June to the end of July. Air was sampled at the north and south site through 2.5 and 4 m long respectively 8 mm diameter Teflon™ tubing with the intake positioned 5 cm from the center of the sonic anemometer and at 1.6 m height from the ground. The different tube lengths were due to the slightly different geometry of the towers and position of the analyzer in the two sites, and a tube attenuation correction was applied in the analysis of the data from both towers. A stainless steel Swagelok™ filter (2 micron filter, SS-4FW4-2) was installed at the input of the sample line to prevent dust from entering the sample cell. A dry scroll pump (Iwata ISP-250B, Iwata ISP-500B) was used to drawn air through the system. Two meters of vacuum tubing was used to dampen the airflow and pressure in the air stream. The DLT-100 methane analyzers were not calibrated because the instruments do not require external calibration (see T. Owano and D. Baer, DLT-100 Fast Methane Analyzer Manual, Los Gatos Research, Mountain View, California, United States). The CH_4, CO_2 and H_2O raw signals were synchronized with the sonic anemometer data and sent through a wireless connection to a computer located in a protected control shelter 170 to 800 m from the eddy covariance tower sites.

3.2. Eddy Covariance Flux Computation and Gap Filling

[12] Fluxes of CH_4, CO_2 and H_2O vapor, sensible heat and momentum were calculated using the EdiRe program and software (version 1.4.3.1169, Robert Clement, University of Edinburgh). Time delays were calculated through the use of a cross-correlation function of the scalar fluctuation and the vertical wind velocity. A two component rotation was applied to set mean vertical (w) and lateral (v) velocity

components to zero. Correction for density change was applied to CO_2 and H_2O fluxes according to *Webb et al.* [1980]. We did not correct the CH_4 fluxes either for high-speed variation in sensible heat since the temperature fluctuations are dampened by passage of air through the tube [*Leuning and Moncrieff*, 1990; *Leuning and King*, 1992], nor for latent heat since it was a very small correction. Obvious data outliers (values more than 6 standard deviations from the 30 min mean for CO_2 and H_2O vapor and more than 10 deviations from the 30 min mean for the wind velocity components, u, v, and w) were removed. Data quality was assessed through the analysis of energy budget closure and by comparing cospectra of $\overline{w'T'}, \overline{w'CO_2'}, \overline{w'H_2O'}, \overline{w'CH_4'}$ [*Kaimal et al.*, 1972]. The data were filtered by wind direction, and only wind directions from 350 to 180 degrees were used, excluding all results when the winds were coming from the back of the tower and outside of the experimental area and the footprint of interest. A friction velocity (u*) threshold 0.25 m/s was used as a cut off, and data below this value were removed. The percentage of data loss for the closed path Los Gatos CH_4 analyzer was 15%, while for the open path LI-COR 7500 it was 32%. The reason for the loss was mainly due to rain, fog, or ice blocking the sonic transducers or the LI-COR 7500 mirrors, low turbulence condition or wind coming from the back of the tower. The gap filing procedures used were: linear interpolation for short gaps (from 1/2 h to 2 h) with approximately uniform environmental conditions [*Falge et al.*, 2001]. The mean diurnal of four to seven adjacent days (averaged for the same time period) was used to fill larger gaps in the data, up to 5 days long [*Falge et al.*, 2001]. For more details about the eddy covariance techniques see *Baldocchi* [2003].

4. Environmental Variables

[13] Micrometeorological variables were recorded continuously at each site. Soil moisture was measured at three depths (0–30 cm, 0–10 cm, 20–30 cm) in five different locations in proximity to the eddy covariance towers using Time Domain Reflectometry (TDR) (CS616 Campbell Scientific, Logan, Utah, United States) moisture probes. Soil temperature was recorded by type *T* thermocouples, (Omega Engineering, Stamford, Connecticut, United States) in nine different locations at 6 different depths (at surface, −1 cm, −5 cm, −10 cm, −20 cm and −30 cm depth). Surface temperature was recorded using an Apogee infrared sensor (Apogee Instruments, Incorporated™, Logan, Utah, United States) pointing into the footprint of the tower at an angle of about 45° with the ground (with field of view 22°, a height above surface of 2.31 m, corresponding to a footprint of about 1.82 m^2). Air temperature and relative humidity was recorded at three heights on the eddy tower structure (0.46, 1.6 and 2.95 m from the ground) using Vaisala HMP45C probes (Vaisala, Helsinki, Finland). Air pressure was measured with an electronic barometer (model PTB 101B, Vaisala, Helsinki, Finland). Incoming, reflected and surface photosynthetically active radiation (PAR) (400–700 nm) was recorded using quantum sensors (Li-190, Li-COR, Nebraska, United States). Net radiation was recorded using a net radiometer (REBS Q7, 0.25–60 μm), shortwave global solar radiation measurements in the spectral range from 310 to 2800 nm incoming and reflected from the ground was collected using two pyranometers (model CMP3, Kipp & Zonen, Delft, The Netherlands). The PAR sensors (except for a surface sensor placed in the vegetation layer), the net radiometer and the pyranometers were mounted on a tripod at about 1.5 m above the ground and at about 5 m to the side of the eddy covariance towers. To obtain a representative ground heat flux measurement, five heat flux plates (HFT3, REBS Incorporated, Seattle, Washington, United States) were installed near each tower positioned at 2 cm depth. Wind speed and wind direction were measured using a wind vane (RM Young Wind Sentry, R.M. Young Company, Traverse, Michigan, United States). Precipitation was recorded using tipping bucket rain gauges (TR-525M, Texas Electronics, Dallas, Texas, United States). All the instruments were connected to a data logger (model 23X, Campbell Scientific, Logan, Utah, United States) and each environmental variable was read once every 10 s and the 30 min averages were recorded.

[14] Water table and thaw depth measurements were made every 3–4 days about every 13 m along first 200m downwind from the towers during summer 2007. Thaw depth was measured using a graduated, pointed metal rod approximately 6 mm in diameter. Water table was measured in 2.5 cm diameter PVC pipe water wells installed at 12 locations randomly selected within the first 200 m of an upwind transect at each site. Three boardwalks were installed across the vegetated drained lake basin to provide access for sampling while avoiding disturbance of the vegetation (Figure 1).

5. Statistical Analysis

[15] General linear modeling was used to identify the most important predictors of CH_4 fluxes in the north site (Systat version 10, Systat Software Incorporated, 2002). Both a single variable model and a forward stepwise multiple regression approach were used to discriminate between and rank the most important environmental variables correlated with CH_4 fluxes. The models were applied to CH_4 fluxes averaged both in half-hour and daily blocks. The data from the south site were not included in these models because we lacked a continuous seasonal data set for this site.

[16] Thaw depth was divided into two groups after examination of the position of the water table (Figure 2a). The variable "Logit thaw depth" was a binary variable with 0 for thaw depth of <15 cm and 1 for thaw depths of >15 cm. We considered this value as a threshold because after 7 July, when thaw depths exceeded 15 cm, the CH_4 fluxes are consistently higher. The water table in the north site (Figure 2a) likely never dropped below 15 cm, and as a consequence the soil layer below this depth never became aerobic during the whole summer season and it is probably the soil layer where the majority of the methanogens occur. This interpretation is in agreement with past measurement of CH_4 concentration profiles at different saturated soil

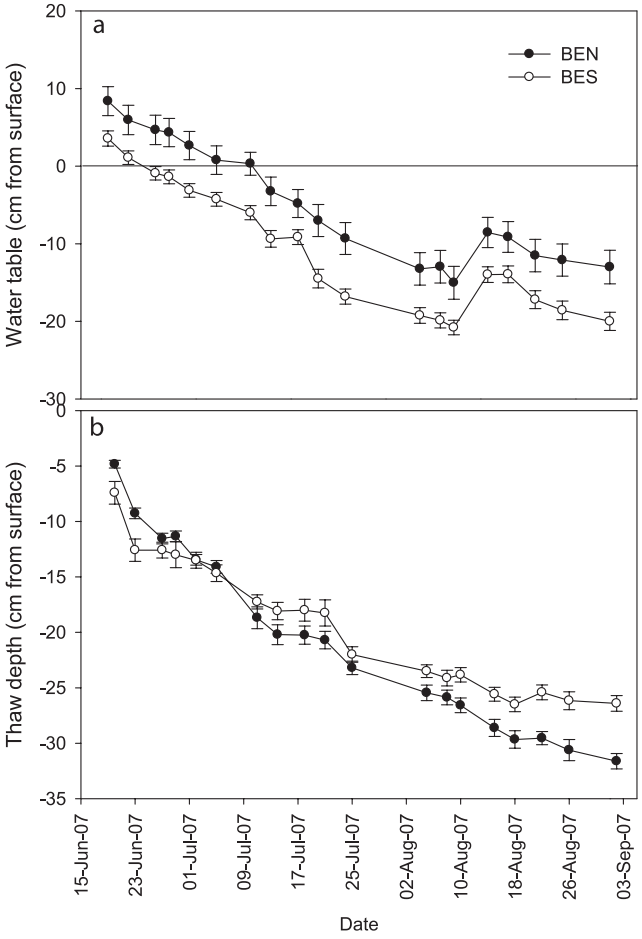

Figure 2. (a) Water table depths and (b) thaw depth (centimeters from surface, positive above surface, negative below surface) during summer 2007 in the north and south site. Each point is the average of 12 points along the first 200 m upwind of the towers; error bars represents standard errors of the mean.

depths that showed the maximum concentration between 15 and 25 cm depth [*Whalen et al.*, 1996].

[17] Water table was also divided in two groups and the variable "Logit water table" (water table above or at surface and below surface) was generated after we noticed an increase in CH_4 fluxes with the water table drop to the surface, probably in response to the decrease in resistance to CH_4 released through plants and soil. Unfortunately Logit thaw depth and Logit water table are identical and their colinearity makes it very difficult to tease apart their reciprocal contribution to the increase in CH_4 emission.

[18] A *t* test was used to test the significance of the difference between CH_4 fluxes in the north and south sites during the periods 18 to 22 June and 16 to 23 July. To model the differences in CH_4 fluxes as a function of the differences in the environmental variables between the two sites, new variables were created by calculating the difference in CH_4 fluxes and environmental parameters between south and north sites. After that, a general linear model procedure was performed to identify the main predictors of the differences in CH_4 fluxes between the two sites, averaged both in half-hour and daily blocks.

6. Results and Discussion

6.1. CH_4 Fluxes and Environmental Variables

[19] In the summer 2007 from visual inspection during daily visit to the sites, and from albedo measurements general snowmelt occurred on 10 June, however the north site showed a higher amount of snow and snowmelt at the tower occurred at 16 June at the north and 11 June at the south tower. The thaw depth was higher in the south site than in the north one until 2 July when the two sites showed the same thaw depth. After that date the north site had a larger thaw depth (Figure 2b). The water table was consistently higher in the north site than in the south during the entire summer (Figure 2a), and the average difference in water table between the two site was 6 cm. Summer 2007 was exceptionally dry and warm; the rainfall from mid-June to the end of July was 3.9 mm and the average air temperature for the same period was 4.7°C.

[20] Previous studies reported a wide range of estimates for CH_4 emission from the Arctic tundra. *Svensson et al.* [1975] and *Svensson and Rosswall* [1984] reported emissions from Swedish subarctic mire ranging from 0.34 to 950 mg CH_4 m^{-2} d^{-1}. *Sebacher et al.* [1986] and *Whalen and Reeburgh* [1990] measured CH_4 emission along a latitudinal gradient from the Arctic to sub-Arctic between −0.3 to 265 mg CH_4 m^{-2} d^{-1}. *Vourlitis and Oechel* [1997] estimated Arctic and subarctic CH_4 flux from 12 mg CH_4 m^{-2} d^{-1} from moist tundra to 100 mg CH_4 m^{-2} d^{-1} from wet sedge ecosystems. Wet sedge ecosystems appeared to be the dominant Arctic and subarctic sources for CH_4 and are estimated to emit between 30 to 128 mg CH_4 m^{-2} d^{-1} [*Vourlitis et al.*, 1993; *Vourlitis and Oechel*, 1997]. CH_4 emission was shown to vary along a microtopographic gradient of soil moisture, with the highest fluxes occurring in wet, vegetated areas [*Morrissey and Livingston*, 1992]. In the Barrow region flooded and wet sites are also characterized by the highest emission, with an average around 48 mg CH_4 m^{-2} d^{-1}, and dry sites are a small source (4.8 mg CH_4 m^{-2} d^{-1}) or sometimes a small sink for CH_4 [*Rhew et al.*, 2007]. Integration of CH_4 fluxes for different vegetation types in Greenland showed this area to be net source of CH_4 of 45.6 mg CH_4 m^{-2} d^{-1} [*Christensen et al.*, 2000] during the summer growing season. In our study we measured CH_4 emission on average of 24.6 mg CH_4 m^{-2} d^{-1} during June and July, which is consistent with previous estimates and very similar to the 20 mg CH_4 m^{-2} d^{-1} determined by *Fan et al.* [1992] for the wet meadow tundra using micrometeorological measurements.

6.2. Environmental Controls on CH_4 Fluxes

[21] In the early season 2007, the south site shows a larger CH_4 release than the north site (Figure 3a). CO_2 uptake during this period is pronounced for the south site, with substantial midday uptake. At the same time, the north site shows a small CH_4 release, with very little evidence of midday CO_2 uptake. The earlier snowmelt at the south site is probably responsible for the earlier activation of the

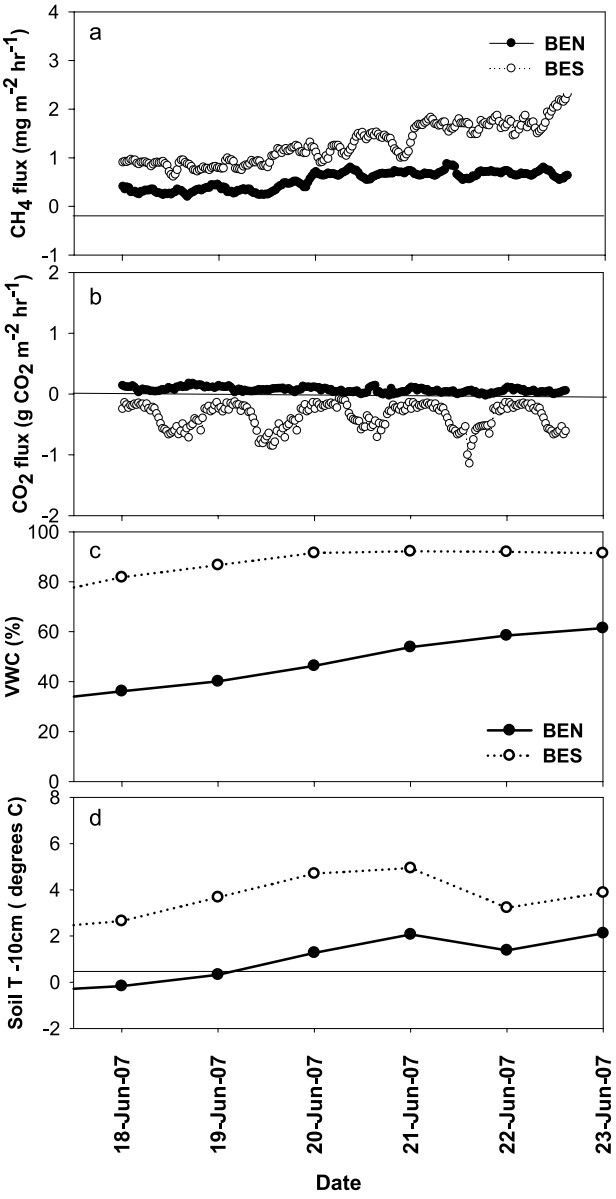

Figure 3. Half-hour (a) CH_4 fluxes and (b) CO_2 fluxes in the north and south sites in early season (18 to 22 June 2007); (c) daily averaged volumetric water content in the first 10 cm depth and (d) soil temperature at 10 cm depth.

ecosystem for both CH_4 and CO_2 fluxes. Because of the earlier snowmelt, the south site shows higher thaw depth (Figure 2b), higher soil temperature, and higher soil moisture (Figures 3c and 3d) in late June when compared to the north site. In fact the ground is still largely frozen in the north site and the water availability to the plants is lower (from 37 to 61% volumetric water content, VWC, in the soil layer 0–10 cm depth) compared to the south site (from 82 to 94% VWC) during the period from 18 to 22 June 2007.

[22] Later in the season (16 to 23 July) the south site shows CH_4 emission about 21% lower than the north site (Figure 4). At this time the north site has 20% higher soil moisture (at 0–10 cm depth), 1°C lower average soil temperature (at −10 cm) than the south site (Figures 4c and 4d); the water table depth is 6 cm lower in the south site than in the north, and thaw depth is about 2 cm lower in the north than in the south (Figure 2). The differences between CH_4 fluxes in the north and south sites during both early and late season proved to be highly significant (*t* test, Bonferroni adjusted probability of 0.009).

[23] The general linear model chosen to explain the differences in CH_4 fluxes between north and south sites, for the data averaged in daily blocks, included only the difference in soil moisture in the 0–10 cm layer between the two sites. This model explained 88% of the difference in

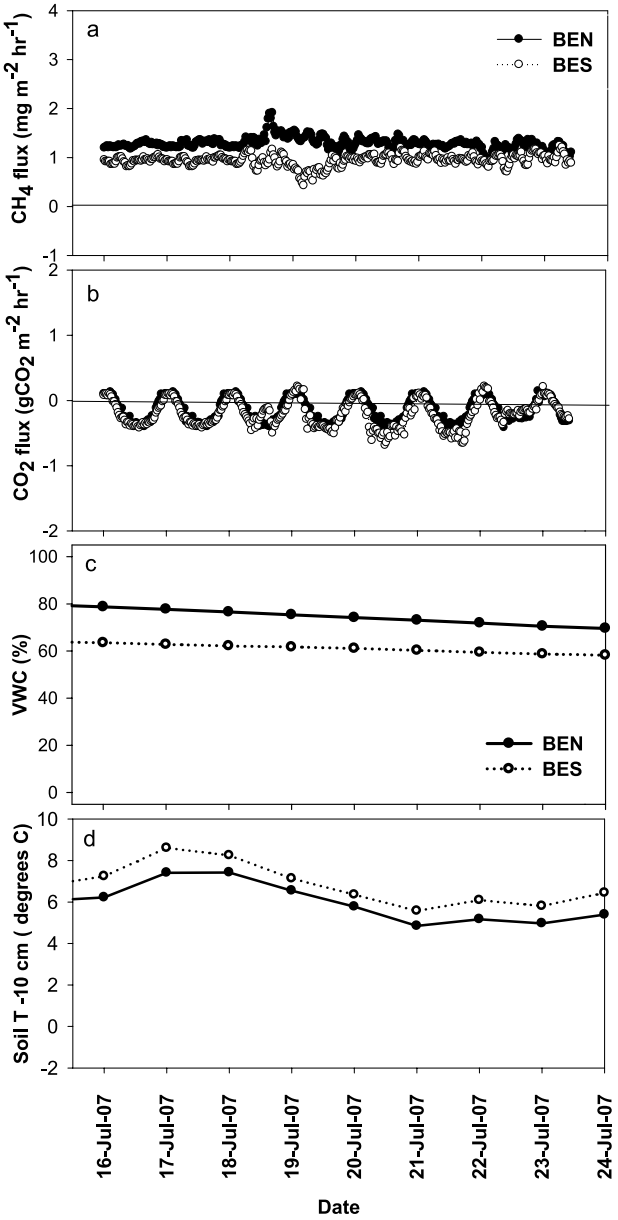

Figure 4. Half-hour (a) CH_4 fluxes and (b) CO_2 fluxes in the north and south site in late season (16 to 23 July 2007); (c) daily averaged volumetric water content in the first 10 cm depth and (d) soil temperature at 10 cm depth.

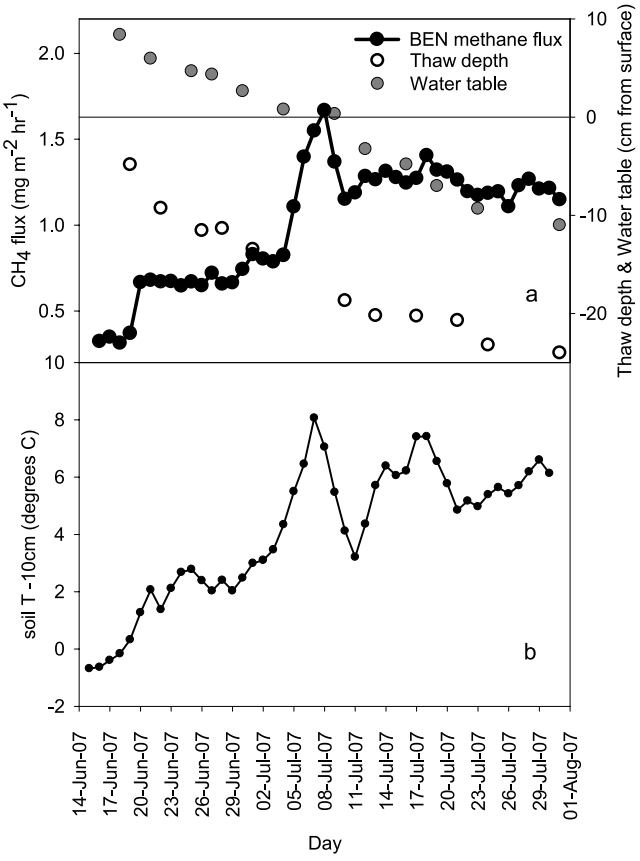

Figure 5. (a) Daily averaged CH_4 fluxes (mg m^{-2} h^{-1}), thaw depth, and water table and (b) soil temperature at 10 cm depth in the north site for the entire measurement period (17 June to 30 July 2007).

CH_4 fluxes between the two sites. Lower water table and soil moisture probably allow more oxygen penetration to the upper soil layers in the south site. This, in turn, would result in greater CH_4 oxidation and direct inhibition of methanogenesis. However, single variable models that included only the difference in thaw depth or the difference in water table between north and south explained about 86% of the variability in the difference in CH_4 fluxes between the two sites, very similar to the one that included difference in soil moisture (0–10 cm) between the two sites. CH_4 production most likely occurs in the deepest, most anaerobic layers, i.e., in the thawed zone below the water table. For this reason we created a new variable, the difference between thaw depth and water table. Once we modeled the difference in CH_4 emission between sites as a function of this variable, the single variable model was able to explain 90% of the variability in CH_4 fluxes. The deeper thaw depth and the higher water table in the north site are linked to a larger amount of anaerobic peat available for methanogenesis. However, the correlation and colinearity among the different environmental variables presents a major challenge in the selection of the best statistical model in explaining the differences in CH_4 fluxes. Better knowledge of the processes occurring in the soil could improve the analysis and avoid underestimation of the role of variables with slightly lower explanatory power.

[24] CH_4 fluxes in the north site were measured continuously for a large part of the growing season. The daily averaged CH_4 fluxes in the north site (Figure 5) showed an increase in CH_4 release from 17 June to 7 July 2007, followed by a decrease until 10 July with little reduction from 10 to 30 July. Several univariable and multivariable models were used to explain seasonal changes in CH_4 fluxes in the north site as a function of environmental factors (see Table 1 for the univariable models and Table 2 for the multivariate analysis). On the daily timescale, the single variable model with the highest explanatory power of averaged CH_4 fluxes identified soil temperature at −10 cm and explained 89% of the variability in CH_4 fluxes (Table 1). In contrast, the best single variable model on the data averaged for half-hour periods also used soil temperature (−10 cm) and had an R-square of 54%. Increased soil temperature not only increases the CH_4 production but also CH_4 diffusion through the soil [*Svensson and Rosswall*, 1984]. Our results suggest that a large part of the half-hour variability in CH_4 fluxes is not well explained by the short-term (half-hour) variation in the key environmental variables measured. In fact, since methanogenesis occurs across a range of soil depths which have different soil temperatures and different diurnal temperature lags, determining the relationship between short-term soil temperature patterns and CH_4 efflux is not straight forward. The multiple variable model of the daily averaged CH_4 flux includes soil temperature at −10 cm, Logit thaw depth (or Logit water table) and soil moisture at 20–30 cm depth, and explained 94% of the variability in CH_4 fluxes (Table 2). The other environmental variables increased the explanatory power of the model less than 1%, their p values were only slightly significant or not significant and, consequently they were not included in the model (Table 2). Also, the different variables were correlated with each other and their inclusion in the model could lead to colinearity problems.

[25] In the multivariable model, water table, used as a continuous variable, was not significant in predicting CH_4 fluxes, but this is probably due to the complexity and nonlinearity of its impact on CH_4 fluxes, with different effects at the beginning and late season. In fact the variable, Logit water table, was highly significant, as was Logit thaw depth (Table 2). The steep increase in CH_4 release until 7 July was correlated with increased soil temperature, increased depth of thaw and water table drop to the surface level (Figure 5). CH_4 emissions as a function of depth of thaw or water table show the same trend (Figures 6a and 6b) making it difficult to separate the contribution from these two environmental variables. When thaw depth is shallower than 15 cm CH_4 fluxes are lower and their rate of increase with increased thaw depth is steeper. Once the thaw depth reaches deeper than 15 cm, the CH_4 fluxes are higher and approach a plateau value. When the water table is at the surface the CH_4 has the lowest resistance to escape and the least probability to be oxidized as shown by the peak emission in Figure 6a. The increase in CH_4 emission with decreased water table until the surface is in agreement with

Table 1. Univariable Model Results for the Daily Average of CH_4 Fluxes as a Function of Several Environmental Variables at the North Site From 17 June to 30 July 2007[a]

Univariate	F	p Value	R Square (%)
Soil T − 10 cm	332.40	<0.001	88.8
Soil moist (0–30 cm)	113.30	<0.001	73.0
Logit thaw depth or Logit water table	103.20	<0.001	71.1
Soil T − 20 cm	86.10	<0.001	67.2
Water table	68.04	<0.001	62.0
Thaw depth	68.50	<0.001	62.0
Soil T − 1 cm	60.00	<0.001	60.0
fCO_2	50.40	<0.001	57.8
Soil T surface	42.00	<0.001	50.0
Soil T − 5 cm	37.00	<0.001	47.0
Air T	36.40	<0.001	46.4
Soil moist (0–10 cm)	22.33	<0.001	34.7
Soil heat flux	21.30	<0.001	34.3
Soil moist (20–30 cm)	21.50	<0.001	33.8
RH	0.80	0.37	1.9
PAR	0.70	0.39	1.8
Wind speed	0.14	0.71	0.3

[a]Reported are F ratio, p value, and R^2 of each of the variables in the model. The best single variable model included soil temperature at −10 cm and explained 89% of the variability in CH_4 fluxes. RH, relative humidity; PAR, photosynthetically active radiation. T, temperature; Units are mg m^{-2} h^{-1}.

the results of other researchers [e.g., *Jauhiainen et al.*, 2005; *Pelletier et al.*, 2007].

[26] Once the water table dropped to a level lower than the surface, the layer of dry soil would be expected to lead to CH_4 consumption [*Del Grosso et al.*, 2000] causing the observed decreases in CH_4 efflux. In fact, according to *Kelker and Chanton* [1997], CH_4 release from *Carex* may be affected by the water level if the water level covers the areas from which CH_4 exits from plant tissues (that is, at the location of the plant stem bases). If CH_4 could escape from any part of the leaf blades, changes in water levels would have a minor effect on CH_4 emission as the CH_4 would simply exit from farther up the leaf blades [*Kelker and Chanton*, 1997].

Table 2. Multivariable Model Results for the Daily Average of CH_4 Fluxes as a Function of Several Environmental Variables at the North Site From 17 June to 30 July 2007[a]

n	Multivariate	F	p Value	ΔR Square (%)
1	Soil T − 10 cm	332.43	<0.001	88.8
2	Logit thaw depth or Logit water table	103.24	<0.001	3.1
3	Soil moist (20–30 cm)	21.48	0.001	2.1
4	Soil moist (0–10 cm)	5.02	0.031	0.7
4	Soil moist (0–30 cm)	4.54	0.04	0.7
4	Water table	5.16	0.029	0.7
4	Thaw depth	5.02	0.031	0.7
4	Soil T − 1 cm	1.96	0.17	0.3
4	Soil T surface	1.99	0.167	0.3
4	PAR	0.79	0.381	0.2
4	Soil T − 5 cm	0.93	0.341	0.2
4	Air T	1.10	0.301	0.2
4	fCO_2	0.67	0.418	0.1
4	RH	0.49	0.487	0.1
4	Soil T − 20 cm	0.61	0.439	0.1
4	Wind speed	0.00	0.971	0
4	Soil heat flux	1.67	0.204	0

[a]Reported are F ratio, p value, and R^2 (soil T − 10 cm) or ΔR^2 of each of the variables in the model. The multivariable model chosen included soil temperature at −10 cm, thaw depth higher or lower than 15 cm and soil moisture at 20–30 cm and it was able to explain 94% of the variability in CH_4 fluxes. Units are mg m^{-2} h^{-1}.

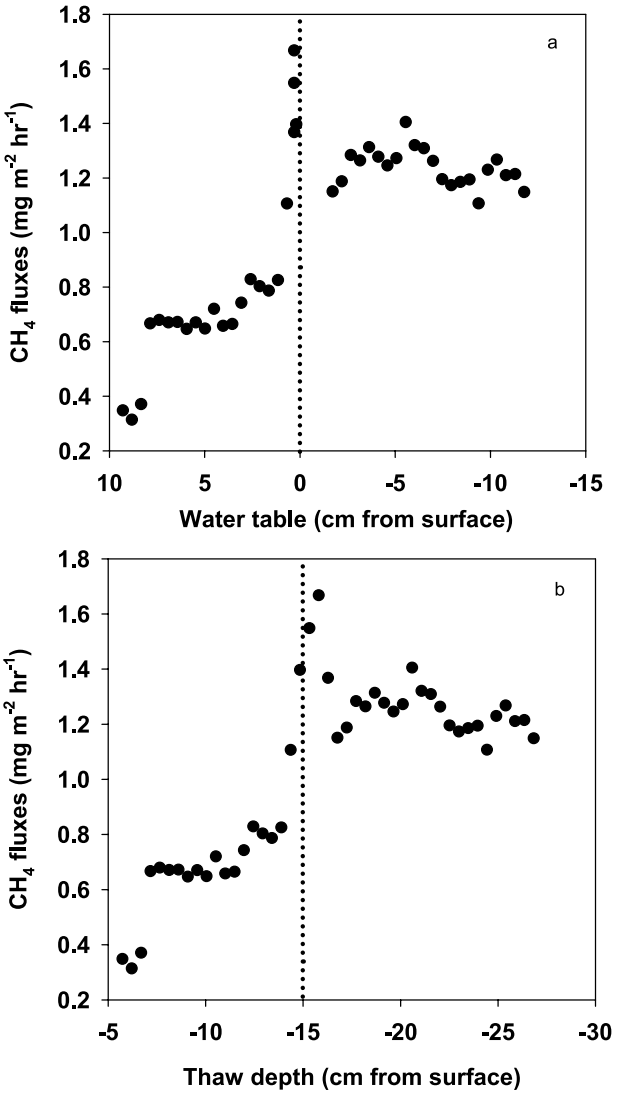

Figure 6. Daily averaged CH_4 fluxes (mg m^{-2} h^{-1}) in the north site (from 17 June to 30 July 2007) as a function of (a) water table and (b) depth of thaw. Noticeable is the peak increase in CH_4 release when water table drops to the surface level (Figure 6a), and the consistent higher CH_4 release after this value and when the thaw depth was deeper than 15 cm (Figure 6b).

[27] A decrease in water table from a few centimeters below the soil surface to 11 cm below the surface seems to have little effect on CH_4 emissions (Figure 6a). This result is in opposition to the conclusions of other researchers where a drawdown of the water table was connected to decreased emission and/or to the sink activity for CH_4 in peatlands [*Harriss et al.*, 1982; *Whalen and Reeburgh*, 1990; *Conrad and Rothfuss*, 1991; *Updegraff et al.*, 1995; *Cao et al.*, 1996; *Bridgham et al.*, 1998]. Our interpretation is that the water table dropping to a few centimeters below

the surface leads to the formation of an oxic layer, which results in the oxidation of CH_4 diffusing through the soil to the atmosphere. A further decrease in water table does not affect CH_4 emissions. At this time thaw depth increases simultaneously with the drop in water table, thereby maintaining a constant active soil volume below the water table (the area most suitable for methanogenesis) and methane bypasses the oxidative zone in soil via plant aerenchyma [*Joabsson et al.*, 1999].

[28] The temporal pattern observed in the north site, that was described in section 6.1, probably also occurred in the south site, presumably due to similar processes. In fact in the south site there is a steady increase in methane release (Figure 3) with a concomitant decrease in water table (Figure 2a), until the water table reaches the surface on about 23 June. After that date the water table at the south site continues to drop and on 17 July it is −9 cm. However, the limited seasonal trends observed in the south site would make any further discussion unwarranted.

[29] Ecosystem carbon supply was highly correlated with CH_4 release in several studies [*Whiting and Chanton*, 1993; *Reeburgh*, 1996; *Nykänen et al.*, 2003]. In our experiment we could not find a significant correlation between NEE and CH_4 emission rates in a multiple regression analysis. The correlation between NEE and methane emission in a single variable model (Table 1) is probably due to dependence of these two variables on similar environmental parameters. It appears that CH_4 production in this system is not limited by available carbohydrates, or other labile carbon sources, and is therefore not likely to respond to increases in net CO_2 uptake from the atmosphere and is not expected to be stimulated by increased plant productivity or increased root exudation. It is well established that a significant fraction (∼20%) of C fixed by photosynthesis is exuded by roots, in the form of labile compounds like sugars and organic acids [*Hinsinger et al.*, 2005; *Jones et al.*, 2004], supporting a link between NEE and labile carbon in soil. Dissolved organic carbon (DOC) and CO_2 fluxes were reported to be strongly correlated for the northern latitude soils [*Neff and Hopper*, 2002] that would suggest a correlation between CH_4 release and available carbon pool in the soil. Because of the very high concentration of available carbohydrate for microbes in soils in our site (D. A. Lipson, unpublished data, 2007), CH_4 emission is probably not limited by available carbon. The upper 15 cm or so of these peat soils is entirely organic matter. The high levels of complex organic matter in these soils appear to support high fluxes of labile C as well: the average DOC concentration at our manipulation site is 43 mg C/L, of which 43% was found to be glucose; furthermore, the activities of polysaccharide-degrading enzymes, such as cellulases, appear to be close to saturation (D. A. Lipson, unpublished data, 2007). The interpretation that CH_4 production is not carbon limited in this system is supported by the results of the experiments performed by *von Fischer and Hedin* [2002], *von Fischer and Hedin* [2007], *Owens and von Fischer* [2007] and *von Fischer et al.* [2007] where acetate addition in a location very close to the site of the experiment described in this paper did not lead to any significant changes in CH_4 emission rates, indicating that the system is not carbon limited.

[30] In conclusion, our study shows that the effect of a decrease in water table is not necessarily a decrease in CH_4 release. An increase in water table above the surface could increase the diffusive resistance to CH_4 release. Additionally, a drop in water table below the surface may not decrease CH_4 emissions, because of the simultaneous increase in thaw depth, and therefore soil volume available for methanogenesis. This study was performed at a landscape level and some of the interpretations presented here should be tested with more mechanistic microscale studies. Moreover, in the long-term, vegetation and species composition may change in response to altered water level, which will mostly likely affect CH_4 fluxes. Better understanding of the complex, holocoenotic [*Billings*, 1952], and nonlinear controls on CH_4 flux is necessary to confidently predict future CH_4 feedbacks from the Arctic.

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References

Baldocchi, D. D. (2003), Assessing the eddy covariance technique for evaluating carbon dioxide exchange rates of ecosystems: Past, present and future, *Global Change Biol.*, *9*, 479–492, doi:10.1046/j.1365-2486.2003.00629.x.

Barber, V., G. P. Juday, and B. Finney (2000), Reduced growth of Alaskan white spruce in the twentieth century from temperature-induced drought stress, *Nature*, *405*, 668–673, doi:10.1038/35015049.

Bartlett, K. B., and R. C. Harriss (1993), Review and assessment of methane emissions from wetlands, *Chemosphere*, *26*, 261–320, doi:10.1016/0045-6535(93)90427-7.

Batjes, N. H. (1996), Total carbon and nitrogen in the soils of the world, *Eur. J. Soil Sci.*, *47*, 151–163, doi:10.1111/j.1365-2389.1996.tb01386.x.

Billings, W. D. (1952), The environmental complex in relation to plant growth and distribution, *Q. Rev. Biol.*, *27*, 251–265, doi:10.1086/399022.

Billings, W. D. (1987), Carbon balance of Alaskan tundra and taiga ecosystems: Past, present and future, *Quat. Sci. Rev.*, *6*, 165–177.

Billings, W. D., and K. M. Peterson (1980), Vegetational change and ice-wedge polygons through the thaw-lake cycle in Arctic Alaska, *Arct. Alp. Res.*, *12*, 413–432, doi:10.2307/1550492.

Billings, W. D., J. O. Luken, D. A. Mortensen, and K. M. Peterson (1982), Arctic tundra: A source or sink for atmospheric carbon dioxide in a changing environment?, *Oecologia*, *53*, 7–11, doi:10.1007/BF00377129.

Bockheim, J. G., L. R. Everett, K. M. Hinkel, F. E. Nelson, and J. Brown (1999), Soil organic carbon storage and distribution in Arctic tundra, Barrow, Alaska, *Soil Sci. Soc. Am. J.*, *63*, 934–940.

Bridgham, S. D., K. Updegraff, and J. Pastor (1998), Carbon, nitrogen and phosphorous mineralization in northern wetlands, *Ecology*, *79*, 1545–1561.

Brown, J. (1967), Tundra soils formed over ice wedges, northern Alaska, *Soil Sci. Soc. Am. Proc.*, *31*(5), 686–691.

Brown, J., K. R. Everett, P. J. Webber, S. F. MacLean Jr., and D. F. Murray (1980), The coastal tundra at Barrow, in *An Arctic Ecosystem: The Coastal Tundra at Barrow, Alaska*, edited by J. Brown et al., pp. 1–29, Dowden Hutchinson and Ross, Stroudsburg, Pa.

Bubier, J. L. (1995), The relationship of vegetation to methane emission and hydrochemical gradients in northern peatlands, *J. Ecol.*, *83*, 403–420, doi:10.2307/2261594.

Cao, M., S. Marshall, and K. Gregson (1996), Global carbon exchange and methane emissions from natural wetlands: Application of a process-based model, *J. Geophys. Res.*, *101*, 14,399–14,414, doi:10.1029/96JD00219.

Chimner, R. A., and D. J. Cooper (2003), Influence of water table levels on CO_2 emissions in a Colorado subalpine fen: An in situ microcosm study, *Soil Biol. Biochem.*, *35*, 345–351, doi:10.1016/S0038-0717(02)00284-5.

Christensen, T. R., T. Friborg, M. Sommerkorn, J. Kaplan, L. Illeris, H. Soegaard, C. Nordstroem, and S. Jonasson (2000), Trace gas exchange in a high-Arctic valley: 1. Variations in CO_2 and CH_4 flux between tundra vegetation types, *Global Biogeochem. Cycles*, *14*, 701–713, doi:10.1029/1999GB001134.

Clymo, R. S. (1984), The limits to peat bog growth, *Philos. Trans. R. Soc. B*, *303*, 605–654.

Conrad, R., and F. Rothfuss (1991), Methane oxidation in the soil surface layer of a flooded rice field and the effect of ammonium, *Biol. Fertil. Soils*, *12*, 28–32, doi:10.1007/BF00369384.

Del Grosso, S. J., W. J. Parton, A. R. Mosier, D. S. Ojima, and C. S. Potter (2000), General CH_4 oxidation model and comparisons of CH_4 oxidation in natural and managed systems, *Global Biogeochem. Cycles*, *14*, 999–1020, doi:10.1029/1999GB001226.

Dickson, R. R. (2000), The Arctic response to the North Atlantic Oscillation, *J. Clim.*, *13*, 2671–2696, doi:10.1175/1520-0442(2000)013<2671:TAORTT>2.0.CO;2.

Ehhalt, D. H. (1974), Sampling of stratospheric trace constituents, *Can. J. Chem.*, *52*, 1510–1518, doi:10.1139/v74-222.

Falge, E., et al. (2001), Gap filling strategies for defensible annual sums of net ecosystem exchange, *Agric. For. Meteorol.*, *107*, 43–69, doi:10.1016/S0168-1923(00)00225-2.

Fan, F. C., S. C. Wofsy, P. S. Bakwin, D. J. Jacob, S. M. Anderson, P. L. Kebabian, J. B. McManus, C. E. Kolb, and D. R. Fitzjarrald (1992), Micrometeorological measurements of CH_4 and CO_2 exchange between the atmosphere and sub-Arctic tundra, *J. Geophys. Res.*, *97*, 16,627–16,643.

Freeman, C., M. A. Lock, and B. Reynolds (1993), Fluxes of CO_2, CH_4, and N_2O from a Welsh peatland following simulation of water table draw-down: Potential feedback to climatic change, *Biogeochemistry*, *19*, 51–60, doi:10.1007/BF00000574.

Fung, I., J. John, J. Lerner, E. Matthews, M. Prather, L. P. Steele, and P. J. Fraser (1991), Three-dimensional model synthesis of the global methane cycle, *J. Geophys. Res.*, *96*, 13,033–13,065, doi:10.1029/91JD01247.

Funk, D. W., E. R. Pullman, K. M. Peterson, P. M. Crill, and W. D. Billings (1994), Influence of water table on carbon dioxide, carbon monoxide, and methane fluxes from taiga bog microcosms, *Global Biogeochem. Cycles*, *8*, 271–278, doi:10.1029/94GB01229.

Gorham, E. (1991), Northern peatlands: Role in carbon cycle and probably response to climate warming, *Ecol. Appl.*, *1*, 182–195, doi:10.2307/1941811.

Grulke, N. E., G. H. Reichers, W. C. Oechel, U. Hjelm, and C. Jaeger (1990), Carbon balance in tussock tundra under ambient and elevated atmospheric CO_2, *Oecologia*, *83*, 485–494, doi:10.1007/BF00317199.

Harazono, Y., and A. Miyata (1997), Evaluation of greenhouse gas fluxes over agricultural and natural ecosystems by means of micrometeorological methods, *J. Agric. Meteorol.*, *52*, 477–480.

Harazono, Y., M. Mano, A. Miyata, R. C. Zulueta, and W. C. Oechel (2003), Inter-annual carbon dioxide uptake of a wet sedge tundra ecosystem in the Arctic, *Tellus, Ser. B*, *55*, 215–231, doi:10.1034/j.1600-0889.2003.00012.x.

Harriss, R. C., D. I. Sebacher, and F. Day (1982), Methane flux in the Great Dismal Swamp, *Nature*, *297*, 673–674, doi:10.1038/297673a0.

Harriss, R. C., E. Gorham, D. I. Sebacher, K. B. Bartlett, and P. A. Flebbe (1985), Methane flux from northern peatlands, *Nature*, *315*, 652–654, doi:10.1038/315652a0.

Hinkel, K. M., R. F. Paetzold, F. E. Nelson, and J. G. Bockheim (2001), Patterns of soil temperature and moisture in the active layer and upper permafrost at Barrow, Alaska: 1993–1999, *Global Planet. Change*, *29*, 293–309, doi:10.1016/S0921-8181(01)00096-0.

Hinkel, K. M., W. R. Eisner, J. G. Bockheim, F. E. Nelson, K. M. Peterson, and X. Dai (2003), Spatial extent, age, and carbon stocks in drained thaw lake basins on the Barrow Peninsula, Alaska, *Arct. Antarct. Alp. Res.*, *35*(3), 291–300, doi:10.1657/1523-0430(2003)035[0291:SEAACS]2.0.CO;2.

Hinkel, K. M., R. C. Frohn, F. E. Nelson, W. R. Eisner, and R. A. Beck (2005), Morphometric and spatial analysis of thaw lakes and drained thaw lake basins in the western Arctic Coastal Plain, Alaska, *Permafrost Periglacial Processes*, *16*(4), 327–341, doi:10.1002/ppp.532.

Hinsinger, P., G. R. Gobran, P. J. Gregory, and W. W. Wenzel (2005), Rhizosphere geometry and heterogeneity arising from root mediated physical and chemical processes, *New Phytol.*, *168*, 293–303, doi:10.1111/j.1469-8137.2005.01512.x.

Hobbie, S. E., J. P. Schimel, S. E. Trumbore, and J. Randerson (2000), Controls over carbon storage and turnover in high latitude soils, *Global Change Biol.*, *6*, 196–210, doi:10.1046/j.1365-2486.2000.06021.x.

Hussey, K. M., and R. W. Michelson (1966), Tundra relief features near Point Barrow, Alaska, *Arctic*, *19*, 162–184.

Jauhiainen, J. I., H. Takahashi, J. E. P. Heikkinen, P. J. Martikainen, and H. Vasander (2005), Carbon fluxes from a tropical peat swamp forest floor, *Global Change Biol.*, *11*, 1788–1797, doi:10.1111/j.1365-2486.2005.001031.x.

Joabsson, A., T. R. Christensen, and B. Wallén (1999), Vascular plant controls on methane emissions from northern peat forming wetlands, *Trends Ecol. Evol.*, *14*(10), 385–388, doi:10.1016/S0169-5347(99)01649-3.

Jones, D. L., A. Hodge, and Y. Kuzyakov (2004), Plant and mycorrhizal regulation of rhizodeposition, *New Phytol.*, *163*, 459–480, doi:10.1111/j.1469-8137.2004.01130.x.

Kaimal, J. C., J. C. Wyngard, Y. Izumi, and O. R. Cote (1972), Spectral characteristics of surface-layer turbulence, *Q. J. R. Meteorol. Soc.*, *98*, 563–589, doi:10.1002/qj.49709841707.

Kelker, D., and J. Chanton (1997), The effect of clipping on methane emissions from Carex, *Biogeochemistry*, *39*, 37–44, doi:10.1023/A:1005866403120.

Klein, E., E. E. Berg, and R. Dial (2005), Wetland drying and succession across the Kenai Peninsula Lowlands, south-central Alaska, *Can. J. For. Res.*, *35*, 1931–1941, doi:10.1139/x05-129.

Leuning, R., and K. M. King (1992), Comparison of eddy-covariance measurements of CO_2 fluxes by open- and closed-path CO_2 analyzers, *Boundary Layer Meteorol.*, *59*, 298–311.

Leuning, R., and J. Moncrieff (1990), Eddy-covariance CO_2 flux measurements using open-and closed-path CO2 analysers: Corrections for analyser water vapour sensitivity and damping of fluctuations in air sampling tubes, *Boundary Layer Meteorol.*, *53*, 63–76, doi:10.1007/BF00122463.

Moore, T. R., and R. Knowles (1989), The influence of water table levels on methane and carbon dioxide emissions from peatland soil, *Can. J. Soil Sci.*, *69*, 33–38.

Moore, T. R., and N. T. Roulet (1993), Methane flux: Water table relations in northern wetlands, *Geophys. Res. Lett.*, *20*, 587–590, doi:10.1029/93GL00208.

Morrissey, L. A., and G. P. Livingston (1992), Methane Flux from Tundra Ecosystems in Arctic Alaska: An Assessment of Local Spatial Variability, *J. Geophys. Res.*, *19*, 16,661–16,670.

Neff, J. C., and D. U. Hopper (2002), Vegetation and climate controls on potential CO_2, DOC and DON production in northern latitude soils, *Global Change Biol.*, *8*, 872–884, doi:10.1046/j.1365-2486.2002.00517.x.

Nykänen, H., J. Alm, J. Silvola, K. Tolonen, and P. J. Martikainen (1998), Methane fluxes on boreal peatlands of different fertility and the effect of long-term experimental lowering of the water table on flux rates, *Global Biogeochem. Cycles*, *12*, 53–69, doi:10.1029/97GB02732.

Nykänen, H., J. E. P. Heikkinen, L. Pirinen, K. Tiilikainen, and P. J. Martikainen (2003), Annual CO_2 exchange and CH_4 fluxes on a subarctic palsa mire during climatically different years, *Global Biogeochem. Cycles*, *17*(1), 1018, doi:10.1029/2002GB001861.

Oechel, W. C., S. J. Hastings, G. L. Vourlitis, M. Jenkins, G. Riechers, and N. Grulke (1993), Recent Change of Arctic tundra ecosystems from a net carbon dioxide sink to a source, *Nature*, *361*, 520–523, doi:10.1038/361520a0.

Oechel, W. C., S. Cowles, N. Grulke, S. J. Hastings, B. Lawrence, T. Prudhomme, G. Riechers, B. Strain, D. Tissue, and G. Vourlitis (1994), Transient nature of CO_2 fertilization in Arctic tundra, *Nature*, *371*, 500–503, doi:10.1038/371500a0.

Oechel, W. C., G. L. Vourlitis, S. J. Hastings, and S. A. Bochkarev (1995), Change in carbon dioxide flux at the U.S. Tundra International Biological Program Sites at Barrow, AK, *Ecol. Appl.*, *5*, 846–855, doi:10.2307/1941992.

Oechel, W. C., G. L. Vourlitis, S. J. Hastings, R. P. Ault Jr., and P. Bryant (1998a), The effect of water table manipulation and elevated temperature on the net CO_2 flux of wet sedge tundra ecosystem, *Global Change Biol.*, *4*, 77–90, doi:10.1046/j.1365-2486.1998.00110.x.

Oechel, W. C., G. L. Vourlitis, S. Brooks, T. L. Crawford, and E. Dumas (1998b), Intercomparison among chamber, tower, and aircraft net CO_2 and energy fluxes measured during the Arctic Systems Science Land-Atmosphere-Ice Interactions (ARCSS-LAII) flux study, *J. Geophys. Res.*, *103*, 28,993–29,004, doi:10.1029/1998JD200015.

Oechel, W. C., G. L. Vourlitis, S. J. Hastings, R. C. Zulueta, L. Hinzman, and D. Kane (2000), Acclimation of ecosystem CO_2 exchange in the Alaskan Arctic in response to decadal climate warming, *Nature*, *406*, 978–981, doi:10.1038/35023137.

Owens, S. M., and J. C. von Fischer (2007), Exploring the relationship between wetland methane emissions and net ecosystem productivity using experimental shading and labile carbon additions, *EOS Trans. AGU*, *88*(52), Fall Meet. Suppl., Abstract B53A-0935.

Pelletier, L., J. Loisel, and M. Garneau (2007), Late-Holocene reconstruction of methane fluxes for the last 3000 years based on testate amoebae assemblages: Application to a Canadian boreal peat bog, *EOS Trans. AGU*, *88*(52), Fall Meet. Suppl., Abstract B52C-07.

Post, W. M., W. R. Emanuel, P. J. Zinke, and A. G. Stangenberger (1982), Soil carbon pools and world life zones, *Nature*, *298*, 156–159, doi:10.1038/298156a0.

Reeburgh, W. S. (1996), "Soft Spots" in the global methane budget, in *8th International Symposium on Microbial Growth on C-1 Compounds*, edited by M. E. Lidstrom and F. R. Tabita, pp. 334–342, Kluwer, Dordrecht, Netherlands.

Rhew, R. C., Y. A. Teh, and T. Abel (2007), Methyl halide and methane fluxes in the northern Alaskan coastal tundra, *J. Geophys. Res.*, *112*, G02009, doi:10.1029/2006JG000314.

Riordan, B., D. Verbyla, and A. D. McGuire (2006), Shrinking ponds in subarctic Alaska based on 1950–2002 remotely sensed images, *J. Geophys. Res.*, *111*, G04002, doi:10.1029/2005JG000150.

Schlesinger, W. H. (1991), *Biogeochemistry: An Analysis of Global Change*, 443 pp., Academic, San Diego, Calif.

Sebacher, D. I., R. C. Harriss, K. B. Bartlett, S. M. Sebacher, and S. S. Grice (1986), Atmospheric methane sources: Alaskan tundra bogs, an alpine fen, and a subarctic boreal marsh, *Tellus*, *38*, 1–10.

Silvola, J., J. Alm, U. Ahlholm, H. Nykanen, and P. J. Martikainen (1996), CO_2 Fluxes from Peat in Boreal Mires under Varying Temperature and Moisture Conditions, *J. Ecol.*, *84*, 219–228, doi:10.2307/2261357.

Smith, L. C., Y. Sheng, G. M. MacDonald, and L. D. Hinzman (2005), Disappearing Arctic lakes, *Science*, *308*, 1429, doi:10.1126/science.1108142.

Sombroek, W. G., F. O. Nachtergaele, and A. Hebel (1993), Amounts, dynamics and sequestrations of carbon in tropical and subtropical soils, *Ambio*, *22*, 417–426.

Strack, M., and J. M. Waddington (2007), Response of peatland carbon dioxide and methane fluxes to a water table experiment, *Global Biogeochem. Cycles*, *21*, GB1007, doi:10.1029/2006GB002715.

Svensson, B. H., and T. Rosswall (1984), In situ methane production from acid peat in plant communities with different moisture regimes, *Oikos*, *43*, 341–350, doi:10.2307/3544151.

Svensson, B. H., A. K. Verum, and S. Kjelvik (1975), Carbon losses from tundra soils, in *Fennoscandian Tundra Ecosystems*, edited by F. E. Wielgolaski, pp. 279–286, Springer, Berlin.

Tarnocai, C. (1994), Amount of organic carbon in Canadian soils, in *Transactions of the 15th World Congress of Soil Science*, vol. 6a, pp. 67–82, Int. Soc. of Soil Sci., Acapulco, Mexico.

Tarnocai, C., C.-L. Ping, and J. Kimble (2003), Determining carbon stocks in cryosols using the northern and mid latitudes soil database, in *Permafrost*, edited by M. Philips, S. Springman, and L. U. Arenson, vol. 2, pp. 1129–1134, Swets and Zeitlinger, Lisse, Netherlands.

Updegraff, K., J. Pastor, S. D. Bridgham, and C. A. Johnston (1995), Environmental and substrate controls over carbon and nitrogen mineralization in northern wetlands, *Ecol. Appl.*, *5*, 151–163, doi:10.2307/1942060.

Updegraff, K., S. D. Bridgham, J. Pastor, P. Weishampel, and C. Harth (2001), Response of CO_2 and CH_4 Emissions From Peatlands to Warming And Water Table Manipulation, *Ecol. Appl.*, *11*, 311–326.

von Fischer, J. C., and L. O. Hedin (2002), Separating methane production and consumption with a field-based isotope pool dilution technique, *Global Biogeochem. Cycles*, *16*(3), 1034, doi:10.1029/2001GB001448.

von Fischer, J. C., and L. O. Hedin (2007), Controls on soil methane fluxes: Tests of biophysical mechanisms using stable isotope tracers, *Global Biogeochem. Cycles*, *21*, GB2007, doi:10.1029/2006GB002687.

von Fischer, J. C., G. Ames, R. Rhew, and W. C. Oechel (2007), Methane emission rates from the Arctic coastal tundra at Barrow: Temporal and spatial variability and response to an experimental carbon addition, *EOS Trans. AGU*, *88*(52), Fall Meet. Suppl., Abstract B51C-0609.

Vourlitis, G. L., and W. C. Oechel (1997), Landscape-scale CO_2, H_2O vapor, and energy flux of moist-wet coastal tundra ecosystems over two growing-seasons, *J. Ecol.*, *85*, 575–590, doi:10.2307/2960529.

Vourlitis, G. L., and W. C. Oechel (1999), Eddy covariance measurements of net CO_2 flux and energy balance of an Alaskan moist-tussock tundra ecosystem, *Ecology*, *80*, 686–701.

Vourlitis, G. L., W. C. Oechel, S. J. Hastings, and M. A. Jenkins (1993), The effect of soil moisture and thaw depth on CH_4 flux from wet coastal tundra ecosystem on the north slope of Alaska, *Chemosphere*, *26*, 329–338, doi:10.1016/0045-6535(93)90429-9.

Walker, D. A., et al. (1998), Energy and trace-gas fluxes across a soil pH boundary in the Arctic, *Nature*, *394*, 469–472, doi:10.1038/28839.

Webb, E. K., G. I. Pearman, and R. Leuning (1980), Correction of flux measurements for density effects due to heat and water vapour transfer, *Q. J. R. Meteorol. Soc.*, *106*, 85–100, doi:10.1002/qj.49710644707.

Webber, P. J. (1978), Spatial and temporal variation of the vegetation and its productivity, Barrow, Alaska, in *Vegetation and Production Ecology of an Alaskan Arctic Tundra*, edited by L. L. Tieszen, pp. 37–112, Springer, New York.

Whalen, S. C., and W. S. Reeburgh (1990), Consumption of atmospheric methane by tundra soils, *Nature*, *346*, 160–162, doi:10.1038/346160a0.

Whalen, S. C., W. S. Reeburgh, and C. E. Reimers (1996), Control of methane emissions by microbial oxidation, in *Landscape Function and Disturbance in Arctic Tundra*, *Ecol. Stud.*, vol. 120, edited by J. F. Reynolds and J. D. Tenhunen, pp. 257–274, Springer, Berlin.

Whiting, G. J., and J. P. Chanton (1993), Primary production control of methane emission from wetlands, *Nature*, *364*, 794–795, doi:10.1038/364794a0.

Whittington, P. N., and J. S. Price (2006), The effects of water table drawdown (as a surrogate for climate change) on the hydrology of a fen peatland, Canada, *Hydrol. Process.*, *20*, 3589–3600, doi:10.1002/hyp.6376.

Yoshikawa, K., and L. D. Hinzman (2003), Shrinking thermokarst ponds and groundwater dynamics in discontinuous permafrost near Council, Alaska, *Permafrost Periglacial Processes*, *14*, 151–169, doi:10.1002/ppp.451.

Zimov, S. A., Y. V. Voropaev, I. P. Semiletov, S. P. Davidov, S. F. Prosiannikov, F. S. Chapin III, M. C. Chapin, S. Trumbore, and S. Tyler (1997), North Siberian lakes: A methane source fueled by Pleistocene carbon, *Science*, *277*, 800–802, doi:10.1126/science.277.5327.800.

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