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# The net exchange of methane with high Arctic landscapes during the summer growing season

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**Abstract.** High Arctic landscapes are essentially vast cold deserts interspersed with streams, ponds and wetlands. These landscapes may be important consumers and sources of the greenhouse gas methane ($CH_4$), though few measurements exist from this region. To quantify the flux of $CH_4$ ($F_{CH_4}$) between the atmosphere and high Arctic landscapes on northern Ellesmere Island, Canada, we made static chamber measurements over five and three growing seasons at a desert and wetland, respectively, and eddy covariance (EC) measurements at a wetland in 2012. Chamber measurements revealed that, during the growing season, desert soils consumed $CH_4$ ($-1.37 \pm 0.06$ mg-$CH_4$ $m^{-2}$ $d^{-1}$), whereas the wetland margin emitted $CH_4$ ($+0.22 \pm 0.14$ mg-$CH_4$ $m^{-2}$ $d^{-1}$). Desert $CH_4$ consumption rates were positively associated with soil temperature among years, and were similar to temperate locations, likely because of suitable landscape conditions for soil gas diffusion. Wetland $F_{CH_4}$ varied closely with stream discharge entering the wetland and hence extent of soil saturation. Landscape-scale $F_{CH_4}$ measured by EC was $+1.27 \pm 0.18$ mg-$CH_4$ $m^{-2}$ $d^{-1}$ and varied with soil temperature and carbon dioxide flux. $F_{CH_4}$ measured using EC was higher than using chambers because EC measurements incorporated a larger, more saturated footprint of the wetland. Using EC $F_{CH_4}$ and quantifying the mass of $CH_4$ entering and exiting the wetland in stream water, we determined that methanogenesis within wetland soils was the dominant source of $F_{CH_4}$. Low $F_{CH_4}$ at the wetland was likely due to a shallow organic soil layer, and thus limited carbon resources for methanogens. Considering the prevalence of dry soils in the high Arctic, our results suggest that these landscapes cannot be overlooked as important consumers of atmospheric $CH_4$.

## 1 Introduction

Rapid warming is altering polar regions at unprecedented rates (AMAP, 2012). Recent climate models suggest that Arctic mean annual temperatures will rise 2.5–7 °C by the end of the 21st century (Overland et al., 2011) but up to 9 °C in local regions such as the Canadian Arctic Archipelago (ACIA, 2005). Mean annual precipitation is also projected to increase throughout the Arctic, resulting from the capability of a warmer Arctic atmosphere to transport more water from low to high latitudes (Manabe and Stouffer, 1994). Warming and wetting of the Arctic has resulted in several environmental responses, including permafrost thaw (Froese et al., 2008), glacial and sea ice melt (Pfeffer et al., 2008), increased surface runoff (Peterson et al., 2002), increased primary productivity and vegetation cover (Walker et al., 2006), and enhanced cycling of greenhouse gases (GHGs), including the powerful GHG methane ($CH_4$; O'Connor et al., 2010), between the atmosphere and changing landscapes.

Both $CH_4$ production (methanogenesis) and consumption ($CH_4$ oxidation, or methanotrophy) occur in Arctic terrestrial, freshwater and marine ecosystems. Methanogenesis is carried out by obligate anaerobic microorganisms (except in ocean surface waters), whereas methanotrophy occurs primarily in oxic environments. In the low and high Arctic (as defined by AMAP, 1998), there are numerous sources of $CH_4$ to the atmosphere, most of which are predicted to

strengthen in a warming and increasingly ice-free environment. These sources include thermokarst lakes, peatlands, lake sediments, thawing permafrost, subglacial environments, $CH_4$ hydrates in marine sediments and $CH_4$ production in ocean surface waters (Roulet et al., 1994; O'Connor et al., 2010; Kort et al., 2012; Wadham et al., 2012). Far more attention has been bestowed on these sources of $CH_4$ to the atmosphere, with proportionally less attention given to numerous sinks of $CH_4$ in polar regions. Sinks of $CH_4$ include the oxic layer above the saturated zone in peatlands where $CH_4$ is produced, in oxygenated water columns of lakes and oceans, and in dry, desert tundra soils that make up a large portion of the high Arctic landscape (Whalen and Reeburgh, 1990). These sinks are equally important to understand and quantify because they can both prevent $CH_4$ from entering the atmosphere and directly consume atmospheric $CH_4$.

Currently, the average atmospheric concentration of $CH_4$ is just over 1800 parts per billion (ppb) in the Northern Hemisphere, compared to a background concentration of $\sim$600 ppb for the majority of the past 600 000 years (Kirschke et al., 2013). Although monitoring the rise of atmospheric $CH_4$ concentrations is extremely important for understanding net emissions of this powerful GHG, it is equally important to quantify how $CH_4$ is interacting with landscapes to understand processes driving concentration changes. For example, the flux of $CH_4$ ($F_{CH_4}$) between landscapes and the atmosphere is the balance between methanogenesis and $CH_4$ oxidation (consumption). When $F_{CH_4}$ is negative, the system is in a phase of net $CH_4$ consumption (or methanotrophy), and $CH_4$ is being removed from the atmosphere. When positive, the system is in a phase of net methanogenesis, and $CH_4$ is being added to the atmosphere. Thus, as climate changes, the state of $F_{CH_4}$ in any ecosystem can have a positive or negative feedback on the atmospheric pool of $CH_4$. Currently, there is a paucity of $F_{CH_4}$ measurements in high Arctic ecosystems (Olefeldt et al., 2013) and little is known about how its direction and magnitude will respond as climate and landscapes change in the future.

The goal of this research was to quantify $F_{CH_4}$ for these remote landscapes where very little is known regarding carbon cycling in general and $CH_4$ fluxes in particular. Between 2008 and 2012, we measured $F_{CH_4}$ near Lake Hazen in Quttinirpaaq National Park, Ellesmere Island, Canada (81.8° N, 71.4° W). Using static chamber measurements, eddy covariance (EC) measurements and a mass budget analysis, we examined spatial and temporal variations in $F_{CH_4}$ over this high Arctic landscape. We hypothesized that dry, unproductive polar desert landscapes would act as a $CH_4$ sink while wet, productive meadow wetlands would be a $CH_4$ source to the atmosphere. As elsewhere, soil moisture, and air and soil temperature were expected to be important drivers of $F_{CH_4}$. However, the high Arctic land area is substantial and represents the extremes of environmental conditions which are changing rapidly, making it a key ecosystem to examine in the context of global $CH_4$ cycling. To our knowledge, this study represents one of the longest records of $F_{CH_4}$ in the high Arctic, and the highest northern latitude EC $CH_4$ measurements collected to date.

## 2 Materials and methods

### 2.1 Research site

We conducted our research out of the Lake Hazen base camp in Quttinirpaaq National Park, Canada's most northerly and remote national park, on northern Ellesmere Island, Nunavut (Fig. 1). Fewer than 15 people typically visit the site each year. The lower reach of the lake's watershed is considered a high Arctic thermal oasis (France, 1993) because it is protected from coastal weather by the Grant Land Mountains and the Hazen Plateau adjacent to the lake. Much of the watershed is typical of the high Canadian Arctic, consisting of a dry, mineral soil landscape with intermittent meadow wetlands and ponds where water flows and collects. Following nine months of sub-0 °C temperatures, snowmelt commences in the watershed in late May and vegetation growth proceeds quickly to peak biomass in mid-July before senescence toward freezing conditions in September. Despite continuous daylight during the growing season, pronounced diurnal patterns in solar radiation exist.

We focussed our study on two common, contrasting landscape types in the high Arctic: a dry, unproductive polar desert (herein "desert") and a moist, productive meadow wetland (herein "wetland") (Fig. 1). Ground cover at the desert ($\sim$188 m a.m.s.l.) is classified as graminoid, prostrate dwarf-shrub forb tundra (Walker et al., 2005) consisting of cryptogamic crust (56.1 %), lichen (11.8 %), D. integrifolia (4.8 %), moss (1.9 %), Carex nardine/Kubresia myosuroides (1.3 %), Salix archea (0.6), litter (3.5 %) and bare ground (20.5 %; Tarnocai et al., 2001). Ground cover at the wetland ($\sim$231 m a.m.s.l.; 2.9 ha) is classified as sedge/grass, moss meadow wetland (Walker et al., 2005) consisting of Carex, Eriophorum and graminoids (Edlund, 1994). The wetland is part of the larger Skeleton Creek meadow wetland complex, consisting of permafrost seeps (PF sites), Skeleton Lake, shallow ponds (e.g. Pond 11) and a creek flowing through a wetland valley (Fig. 1). During a typical growing season, the creek flows into the wetland, saturates soils and exits downstream towards Lake Hazen.

### 2.2 Quantifying $F_{CH_4}$

#### 2.2.1 Measurement overview

$F_{CH_4}$ has overwhelmingly been measured throughout the Arctic using static chambers because of their simplicity and convenience (Parmentier et al., 2011). The EC technique (Baldocchi, 2003) for measuring $F_{CH_4}$ has only been used sporadically in the high Arctic (e.g. Friborg et al., 2000) because tunable diode laser detectors or other closed path

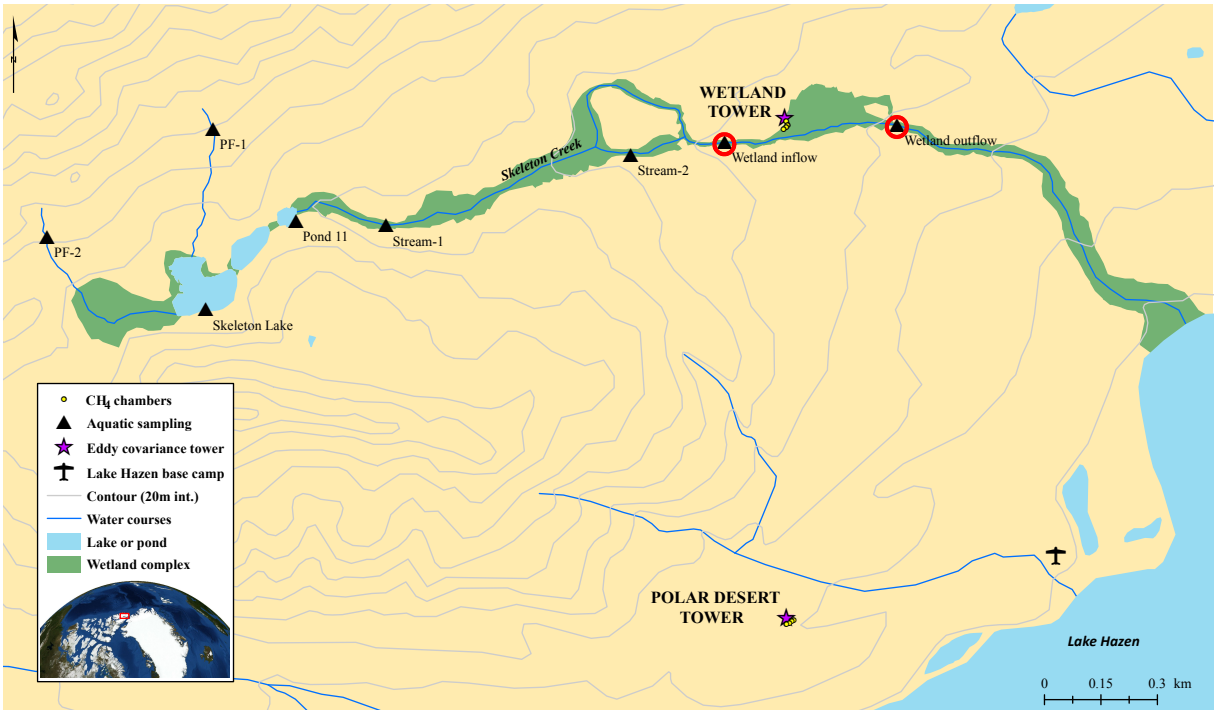


**Figure 1.** Lake Hazen base camp in Quttinirpaaq National Park, Nunavut, Canada (81.8° N, 71.4° W). Both the polar desert and meadow wetland study sites are shown with static chamber, eddy covariance and aquatic $CH_4$ sampling locations indicated. Emphasis added to aquatic sites upstream and downstream of the wetland. PF sites indicate permafrost seep streams and Stream sites indicate Skeleton Creek sites.

detectors require large quantities of power not readily available in remote high Arctic locations. Recently, a low power consuming open path $CH_4$ analyser (LI-7700; LI-COR, Lincoln, NE) has appeared on the market (McDermitt et al., 2011). EC provides $F_{CH_4}$ near continuously over short temporal scales (30 min) and large spatial scales (hectares) providing great potential to focus on ecosystem-scale exchanges with the atmosphere and the biotic and abiotic factors driving temporal variations across northern ecosystems. This study was part of a larger one in which we are quantifying the flux of the GHG carbon dioxide ($F_{CO_2}$; the balance between $CO_2$ uptake via photosynthesis and the release of $CO_2$ via ecosystem respiration) between the atmosphere and desert (2008–2012) and wetland (2010–2012) landscapes using EC flux towers. Towers were equipped with Campbell Scientific, Inc. (CSI; Logan, UT) CSAT3 sonic anemometers and LI-COR LI-7500 (open-path) and LI-7200 (enclosed-path) $CO_2$/water vapour ($H_2O$) infrared gas analysers (Supplement Fig. S1). In addition to $F_{CH_4}$ and $F_{CO_2}$, these tower-based EC systems quantified $H_2O$ and energy fluxes and were equipped with sensors to measure soil temperature and moisture at 5 cm depth (CS107B, CS616, CSI; 30 min mean each tower, each growing season), and other meteorological parameters (Supplement Table S1). Signals from all sensors were recorded as half-hour means on CSI CR3000-XT data loggers. Thaw depth was monitored weekly at 10 points along a transect at each site using a steel probe.

### 2.2.2 Chamber measurements

Static, non-steady state chambers were used to quantify $F_{CH_4}$ at the desert (2008–2012) and wetland sites (2010–2012; Fig. 1). At the start of each season, we set four 25 cm diameter white PVC collars 10–15 cm into the soil within 20 m of each tower (the same locations each year), where they remained for the rest of the field season. Two desert collars enclosed bare soil and two other collars enclosed $> 50\,\%$ vegetation cover consisting mostly of Dryas (Fig. 1, Supplement Fig. S1). At the wetland, four collars were placed along its margin because a boardwalk was not permitted in the National Park to access the centre of the wetland (Fig. 1, Supplement Fig. S1). Each collar enclosed vegetation of similar type and cover as the rest of the wetland. Chambers were deployed at each site every 5 to 7 days between June and August. On sampling days, between 10:00–16:00 LT, foil-covered 18 L plastic chambers with sampling lines were placed into a water-filled groove on the collars. At 0, 20, 40 and 60 min after deployment, air inside each chamber was mixed by syringe before chamber air was collected into an evacuated 35 mL Wheaton glass bottle. Ambient air pressure and temperature were recorded. All samples were stored in the dark at 4 °C until analysis at the University of Alberta. We used a Varian 3800 gas chromatograph (GC) with a flame-ionizing detector to measure the $CH_4$ concentration (in parts per million; ppm) of each gas sample from each

chamber. Three standard-grade gases (0, 1, 54 ppm-$CH_4$) were used to calibrate the GC, and all samples were analysed in duplicate. We then used the $CH_4$ concentration, the ideal gas law, chamber metrics, ambient pressure and temperature, and the gas constant to quantify the mass of $CH_4$ enclosed by each chamber at each sampling time. Linear regressions were used to fit relationships between sample times and total masses of $CH_4$ for each chamber, and root mean squared errors (RMSEs) were used to assess regression performance (Kutzbach et al., 2011). Regression estimates typically fit well to observed $CH_4$ masses in both desert (mean measured $\pm$ RMSE; $11.19 \pm 0.45$ µg; $n = 101$) and wetland ($13.23 \pm 0.47$ µg $CH_4$; $n = 66$) chambers. The slope of the regression line determined $F_{CH_4}$ (mg $CH_4$ $m^{-2}$ $hr^{-1}$) for each chamber. Fluxes from the four chambers were averaged to determine site daily means (mg $CH_4$ $m^{-2}$ $d^{-1}$) with the assumption that there would be little diurnal variation in $F_{CH_4}$ (supported by EC measurements; see Sect. 3.2).

#### 2.2.3 Eddy covariance measurements

As described above, although EC technology is not new, only recently has a low-power, robust $CH_4$ analyser become available. We had the opportunity during the 2012 growing season to deploy an LI-7700 open-path $CH_4$ gas analyser on one of our two EC towers. Because we could not obtain chamber measurements in the centre of the wetland, we deployed the LI-7700 on the wetland EC tower to attain more representative $CH_4$ fluxes from that ecosystem than provided by the chambers on the wetland's periphery. The wetland EC tower was positioned just outside the western margin of the wetland, leeward of the prevailing wind. Winds originated from the prevailing direction for 82 % of all half-hour measurements, and 90 % of all fluxes originated from within the wetland footprint using the Kljun et al. (2004) model. The LI-7700 was laterally positioned 25 cm from the sonic anemometer and 1.9 m above the vegetation canopy height. Measurements of $CH_4$ molar density, wind velocity in three coordinates, sonic temperature, ambient pressure, and $CO_2$ and $H_2O$ mixing ratios (LI-7200) were collected at 10 Hz and logged on a LI-COR LI-7550 interface unit.

We used EddyPro (LI-COR, v. 4.1) to calculate $CH_4$, $CO_2$ and $H_2O$ fluxes and to QA/QC data and remove outliers. Gas fluxes were calculated at half-hour intervals using a block averaging approach. To correct for anemometer tilt, a double rotation was performed to force mean vertical and lateral wind components to zero. $F_{CH_4}$ data were de-spiked and corrected for time lag between the anemometer and the gas analyser measurements using a covariance maximization approach. Because the LI-7700 is an open-path analyser, density fluctuations were corrected for using the Webb et al. (1980) approach. We used spectral corrections to adjust for flux loss at high and low frequencies (after Ibrom et al., 2007) and to correct for the spectroscopic effects of $H_2O$ (LI-COR, 2011). We removed half-hour fluxes when EC sensors malfunctioned, returned poor diagnostic values (e.g. during rare rain events), when wind did not pass over the wetland (17.8 % of all half-hour fluxes), and when the friction velocity fell below 0.1 m $s^{-1}$, similar to other studies (Wille et al., 2008). We also applied turbulence tests after Mauder and Foken (2006) to remove the poorest-quality fluxes (level 2) when they did occur. Half-hour fluxes that were beyond $\pm 3$ SD of the growing season mean were also removed. These corrections resulted in the removal of 43.8 % of total collected flux data. Measurement gaps occurred between 22 June and 1 July and between 31 July and 1 August when solar charging could not match power requirements. For both chamber and EC $F_{CH_4}$ measurements, positive values represented $CH_4$ emission to the atmosphere, whereas negative values represented $CH_4$ consumption in soils.

### 2.3 Wetland aquatic chemistry

We determined if there were significant dissolved $CH_4$ inputs by Skeleton Creek into the wetland so we could examine the potential for methanogenesis within the wetland soils. These measurements, in combination with EC flux tower measurements, would also allow us to construct a general $CH_4$ mass budget for the wetland. We collected surface water upstream and downstream of the wetland every 2 to 5 days during the 2012 growing season (Fig. 1). We measured the partial pressure of dissolved $CH_4$ by collecting surface water at each site into evacuated 160 mL Wheaton glass serum bottles with butyl rubber stoppers (after Kling et al., 1991). Each bottle contained 8.9 g of potassium chloride preservative, and 10 mL of ultra high-purity $N_2$ headspace. Samples were analysed on the same GC used to analyse the chamber samples, but using 0, 50, 350, and 900 ppm $CH_4$ standard gases. All samples were placed in a wrist-action shaker for 20 min to equilibrate headspace gas with the sample. 500 uL of headspace gas was extracted from each sample for analysis using a gas-tight syringe. Duplicate analyses were performed on all samples. We used the headspace $CH_4$ gas concentrations from each sample, ambient and laboratory temperature and pressure, and Henry's Law to determine the dissolved $CH_4$ concentration in the collected water sample. Water was also collected at each site for analyses of general water chemistry parameters, including concentrations of particulate and dissolved nutrients, ions, chlorophyll *a* and dissolved organic carbon. All samples were initially processed and preserved on-site in the Lake Hazen/Quttinirpaaq Polar Laboratory and subsequently analysed using standard methods at the University of Alberta's Biological Analytical Services Laboratory. In situ measurements including pH, dissolved oxygen, water temperature, oxidation-reduction potential and specific conductivity were also taken at each site at time of sampling using a YSI (YSI Environmental, Yellow Springs, OH) 556 MPS multi-probe system. Water flow at each site was measured every 2 to 3 days using a Pygmy current meter. At each site, we chose a channelized section of

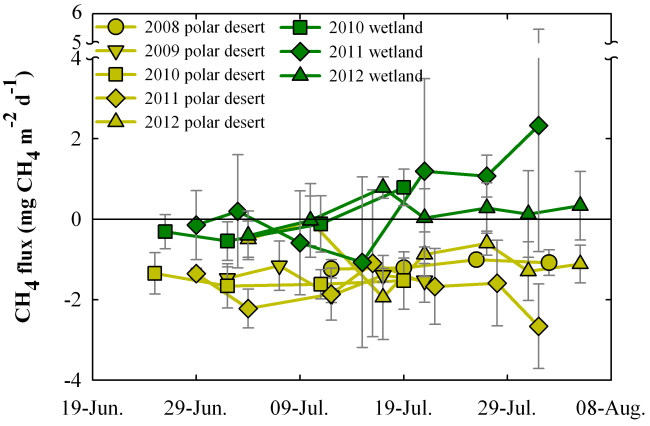


**Figure 2.** Mean $CH_4$ fluxes ($F_{CH_4}$; ±1SD) from four polar desert and four wetland static chambers during the 2008–2012 growing seasons.

stream and measured the water velocity at half-depth across 5 cm segments of stream. We then took the product of stream cross sectional area and mean velocity in each segment and summed all segments to quantify total stream flow.

## 3 Results

### 3.1 Chamber measurements

Over several growing seasons, soils at the desert site consumed atmospheric $CH_4$ at a mean rate (±1SE) of $-1.37 \pm 0.06$ mg $CH_4$ $m^{-2}$ $d^{-1}$ ($n = 4$, mean of 4 independent collars measured 27 times each between 2008 and 2012), whereas the wetland site emitted fluxes of $CH_4$ ($+0.22 \pm 0.14$ mg $CH_4$ $m^{-2}$ $d^{-1}$; $n = 4$, mean of 4 independent collars measured 18 times each between 2010 and 2012). Desert soils consistently consumed atmospheric $CH_4$ throughout the growing season, whereas wetland soils typically consumed atmospheric $CH_4$ during the first 2 weeks of July ($-0.34 \pm 0.12$ mg $CH_4$ $m^{-2}$ $d^{-1}$) before transitioning to a source of $CH_4$ to the atmosphere ($0.77 \pm 0.24$ mg $CH_4$ $m^{-2}$ $d^{-1}$) (Fig. 2). When comparing paired sampling dates from each site between 2010 and 2012, we found that the desert landscape consumed significantly more atmospheric $CH_4$ than the wetland (Repeated-measures ANOVA; $F_{(1,17)} = 92$, $p < 0.001$; Fig. 3). These site differences in $F_{CH_4}$ were related to the large differences in soil moisture and soil temperature (Fig. 3). Daily mean soil moisture at 5 cm depth of the desert soils was consistently near $15.1 \pm 1.0$ % v/v during the measurement period, except during short rain events. Wetland soil moisture at the same depth was considerably higher ($75.2 \pm 3.2$ %) than at the desert. Because the wetland was bowl-shaped, snowmelt and creek water saturated the centre of the wetland first before wetting the margins where the chamber collars were located. In 2012, the wetland gradually dried after snowmelt because creek

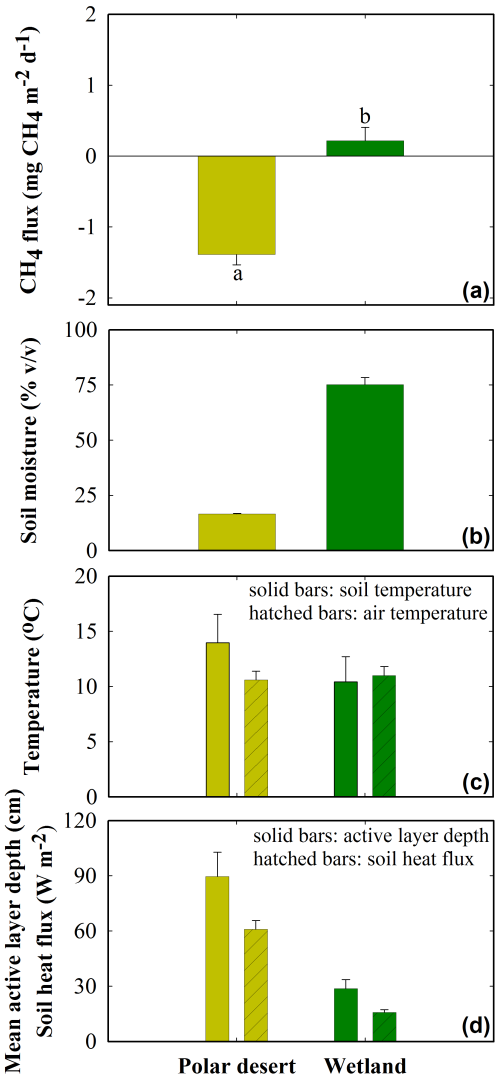


**Figure 3.** Comparison of 2010–2012 growing season mean $CH_4$ fluxes ($F_{CH_4}$; ±1SE) measured in chambers **(a)** and other environmental variables **(b–d)**, paired by site. The sampling period represented by each bar spans approximately late June to early August. Letters indicate if there were statistically significant differences of $F_{CH_4}$ between sites using a repeated-measures ANOVA.

flow ceased due to low water levels in ponds upstream. Once ponds returned to maximum storage, creek flow resumed on 16 July 2012 and eventually re-saturated the wetland margin soils to levels similar to other years. Throughout the chamber measurement period, the desert site, relative to the wetland, had higher 5 cm depth soil temperature ($14.4 \pm 0.5$ °C-desert vs. $10.4 \pm 0.5$ °C-wetland), higher soil heat flux at 5 cm depth ($52.3 \pm 4.2$ W $m^{-2}$-desert vs. $15.8 \pm 1.5$ W $m^{-2}$-wetland) and deeper thaw depths ($88 \pm 3$ cm-desert vs. $29 \pm 1$ cm-wetland; Fig. 3).

Between 2008 and 2012 at the desert site, mean growing season $F_{CH_4}$ ranged between $-0.91$ and $-1.78$ mg

**Table 1.** Mean (±1SE) daily $CH_4$ flux ($F_{CH_4}$) and environmental variables during the chamber measurement period of several growing seasons at the desert and wetland sites.

| | $F_{CH_4}$ (mg $CH_4$ $m^{-2}$ $d^{-1}$) | n (#) | $Air_T$ (°C) | PAR (µmol $m^{-2}$ $s^{-1}$) | SHF (W $m^{-2}$) | $Soil_M$ (% v/v) | $Soil_T$ (°C) | $n$ (daily) |
|---|---|---|---|---|---|---|---|---|
| Desert | | | | | | | | |
| 2008 | $-1.13 \pm 0.05$ | 4 | $7.3 \pm 0.6$ | $520 \pm 22$ | $14.6 \pm 2.7$ | $17.0 \pm 2.2$ | $11.4 \pm 0.4$ | 24 |
| 2009 | $-1.49 \pm 0.12$ | 5 | $10.0 \pm 0.7$ | $678 \pm 20$ | $26.9 \pm 1.1$ | $9.4 \pm 0.1$ | $12.9 \pm 0.5$ | 20 |
| 2010 | $-1.54 \pm 0.07$ | 4 | $9.8 \pm 0.6$ | $685 \pm 26$ | $28.8 \pm 1.9$ | $17.5 \pm 0.2$ | $12.7 \pm 0.5$ | 25 |
| 2011 | $-1.78 \pm 0.20$ | 7 | $9.5 \pm 0.3$ | $678 \pm 22$ | $32.1 \pm 1.4$ | $16.9 \pm 0.1$ | $13.7 \pm 0.3$ | 34 |
| 2012 | $-0.91 \pm 0.23$ | 7 | $8.1 \pm 0.4$ | $520 \pm 32$ | $26.5 \pm 2.0$ | $15.4 \pm 0.1$ | $11.1 \pm 0.4$ | 33 |
| Wetland | | | | | | | | |
| 2010 | $-0.05 \pm 0.29$ | 4 | $11.0 \pm 0.8$ | $652 \pm 24$ | $7.8 \pm 0.8$ | $79.0 \pm 1.3$ | $10.6 \pm 0.4$ | 22 |
| 2011 | $0.43 \pm 0.44$ | 7 | $10.7 \pm 0.4$ | $657 \pm 21$ | $6.4 \pm 0.6$ | $80.6 \pm 0.6$ | $10.8 \pm 0.2$ | 34 |
| 2012 | $0.16 \pm 0.14$ | 7 | $9.1 \pm 0.5$ | $507 \pm 31$ | $8.1 \pm 0.4$ | $58.2 \pm 1.3$ | $8.2 \pm 0.1$ | 33 |

$Air_T$: air temperature; PAR: photosynthetically active radiation; SHF: soil heat flux at 5 cm depth; $Soil_M$: volumetric soil moisture; $Soil_T$: soil temperature; $F_{CH_4}$ $n$ indicates the number of landscape mean measurements (of four chambers) taken during each growing season (also see Fig. 2).

$CH_4$ $m^{-2}$ $d^{-1}$ (Table 1, Fig. 2). $CH_4$ consumption rates were positively correlated with soil temperatures between years ($r^2 = 0.97$; $n = 5$; simple correlation) but not influenced by changes in soil moisture ($r^2 < 0.01$). Consumption rates of $CH_4$ were not significantly different in chambers with or without vegetation (RM-ANOVA; $F_{(1,26)} = 0.15$, $p = 0.76$). Associations between within-season $F_{CH_4}$ and environmental factors were generally weak ($-0.28 < \rho < 0.07$; Spearman Rank Correlation; Supplement Table S2-A). From 2010–2012 at the wetland site, mean growing season $F_{CH_4}$ ranged between $-0.05$ and $+0.43$ mg $CH_4$ $m^{-2}$ $d^{-1}$ (Table 1, Fig. 2). With only 3 years of data, trends between mean growing season $F_{CH_4}$ at the wetland site and explanatory variables were not meaningful. However, we do note that years with fairer weather (air pressure $r^2 = 0.95$) and warmer conditions (thaw depth $r^2 = 0.81$; soil heat flux at 5 cm depth $r^2 = 0.67$) seemed to be associated with greatest emissions at the wetland. Within-season wetland $F_{CH_4}$ was positively correlated with mean daily stream flow in Skeleton Creek ($\rho = 0.72$; Supplement Table S2-B.).

### 3.2 Eddy covariance measurements

$F_{CH_4}$, measured using the EC flux tower in 2012, was between $-0.84$ and $+2.73$ mg $CH_4$ $m^{-2}$ $d^{-1}$ with a mean daily $F_{CH_4}$ (±1SE) of $1.27 \pm 0.18$ mg $CH_4$ $m^{-2}$ $d^{-1}$ at the wetland (Fig. 4a) with no discernible diurnal patterns (Supplement Fig. S2). On days when net $CH_4$ consumption occurred, mean $F_{CH_4}$ was $-0.33 \pm 0.07$ mg $CH_4$ $m^{-2}$ $d^{-1}$ ($n = 9$) compared to $+1.76 \pm 0.14$ mg $CH_4$ $m^{-2}$ $d^{-1}$ ($n = 29$) when net $CH_4$ emission was occurring. Net uptake of $CH_4$ quickly changed to net emission just after wetland soils rapidly thawed. Soil temperature warmed from freezing conditions (−1.3 °C) to above 7 °C during the first 7 days of measurements (Fig. 4b). During that time, frozen moisture within soils and in snow covering the wetland thawed and saturated the wetland landscape (Fig. 4c). The increase in evaporative fluxes preceded the saturation of the 5 cm depth of the wetland margin, while $F_{CO_2}$ remained positive (net $CO_2$ emission) for another week after this initial thaw period (Fig. 4d). $CH_4$ emissions peaked during the first 2 weeks of July, similar to when net $CO_2$ uptake peaked. We did not observe significant changes in whole-wetland $CH_4$ emission rates when Skeleton Creek flow resumed during the third week of July, and soil moisture in the wetland margin returned to values similar to the post-thaw period (Fig. 4a, c). In contrast, $F_{CH_4}$ measured by static chambers increased through the summer with peak $CH_4$ emissions at the end of the season when Skeleton Creek flow was greatest. However, chamber-based $F_{CH_4}$ on the wetland margin was always lower than the fluxes measured by the EC technique, including a period in early July where average $F_{CH_4}$ indicated net $CH_4$ uptake (Fig. 4a). Overall, seasonal variations in $F_{CH_4}$ measured by EC associated strongest with $F_{CO_2}$ and soil temperature (Supplement Table S3).

### 3.3 Wetland aquatic chemistry

Flow-weighted mean dissolved $CH_4$ concentrations (±1 weighted SD) in Skeleton Creek water (Table 2) decreased from $0.005 \pm 0.004$ µmol $L^{-1}$ upstream of the wetland to $0.001 \pm 0.005$ µmol $L^{-1}$ downstream of the wetland between 3 July and 5 August, a decrease of 70 %. Ammonium concentrations increased downstream, while nitrate concentrations were below the analytical detection limit at both sites. Concentrations of dissolved organic nitrogen and carbon were higher in the wetland outflow than inflow (Table 2).

If we assume no net storage of $CH_4$ in the wetland over a growing season when stream flow was occurring (∼ late June to early August), we can calculate the net production of $CH_4$ (production-oxidation losses) in wetland soils using the

**Table 2.** Flow-weighted mean concentrations (±1weighted SD) of several chemicals in Skeleton Creek water upstream and downstream of the wetland during the 2012 growing season. All chemicals are reported in µmol $L^{-1}$ except for water temperature (°C) and oxidation-reduction potential (mV).

| Parameter | 2012 | | |
|---|---|---|---|
| | Wetland inflow | Wetland outflow | % change |
| Dissolved $CH_4$ | $0.005 \pm 0.004$ | $0.001 \pm 0.005$ | −70 % |
| Dissolved $CO_2$ | $72 \pm 29$ | $65 \pm 10$ | −9 % |
| $NO_3^-$ | $0.04 \pm 0.00$ | $0.04 \pm 0.00$ | 0 % |
| $NH_4^+$ | $0.53 \pm 0.13$ | $0.76 \pm 0.07$ | 42 % |
| DON | $19.4 \pm 1.0$ | $22.0 \pm 0.5$ | 13 % |
| TDN | $20.0 \pm 0.9$ | $22.8 \pm 0.5$ | 14 % |
| DOC | $497 \pm 28$ | $549 \pm 19$ | 10 % |
| Water $_T$ | $8.1 \pm 1.6$ | $7.8 \pm 1.4$ | −4 % |
| ORP | $53 \pm 57$ | $21 \pm 17$ | −60 % |

$NO_3^-$: dissolved nitrate; $NH_4^+$: dissolved ammonium; DON: dissolved organic nitrogen; TDN: total dissolved nitrogen; DOC: dissolved organic carbon; $Water_T$: water temperature; ORP: oxidation-reduction potential.

**Table 3.** Wetland mass balance (Eq. 2) of $CH_4$ for the 2012 growing season (3 July to 5 August), including stream input ($I_{CH_4}$) and output ($O_{CH_4}$), flux of CH4 ($F_{CH_4}$) from the EC tower, and estimate of net $CH_4$ production within wetland soil ($NP_{CH_4}$).

| | Stream flow ($m^3$) | $n$ (#) | $CH_4$ transfer (g) |
|---|---|---|---|
| $I_{CH_4}$ | 6 578 | 8 | 0.5 |
| $O_{CH_4}$ | 5 451 | 6 | 0.1 |
| $F_{CH_4}$ | – | 34 | 2 002 |
| $NP_{CH_4}$ | – | – | 2 002 |

following equation:

$$\Sigma(I_{CH_4} + NP_{CH_4})_{daily} = \Sigma(F_{CH_4} + O_{CH_4})_{daily}\,, \quad (1)$$

where $I_{CH_4}$ and $O_{CH_4}$ are the daily masses of dissolved $CH_4$ entering and exiting the wetland, $NP_{CH_4}$ is the daily net production of $CH_4$ in soils scaled to 2.9 ha of the wetland, and $F_{CH_4}$ is the daily flux of $CH_4$ from the wetland surface (2.9 ha) as measured by the EC tower. Net storage of $CH_4$ in wetland soils during the growing season was clearly shown via burst events in autumn at another high Arctic location (Mastepanov et al., 2008). However, we suspect these events were less important at our site because of a substantially thinner organic layer (see Sect. 4.2), lack of a measurable $CH_4$ burst in spring, and the absence of an autumn $CO_2$ burst at our site (unpublished data), which is often coincident with $CH_4$ burst events (Mastepanov et al., 2013). $I_{CH_4}$ and $O_{CH_4}$ were calculated using

$$I_{CH_4} \text{ or } O_{CH_4}(g) = (([CH_{4t1}] + [CH_{4t2}])/2) \cdot V, \quad (2)$$

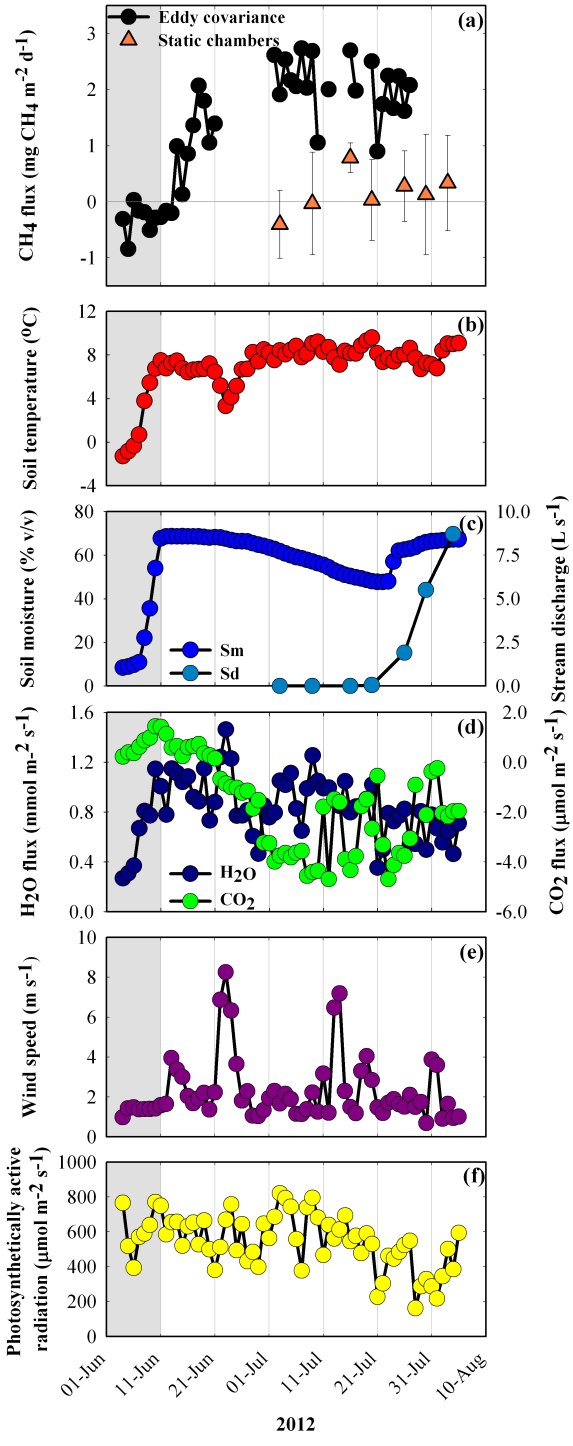


**Figure 4.** Comparison of mean daily eddy covariance and static chamber $CH_4$ fluxes **(a)** at the wetland and several mean daily environmental variables **(b–f)** during the 2012 growing season. Shaded bars highlight the period of rapid soil thaw.

where $[CH_{4t1}]$ and $[CH_{4t2}]$ were mean concentrations of dissolved $CH_4$ at two consecutive sampling times and V was the total volume of water that flowed through each station

between those times. Solving for $NP_{CH_4}$ in Eq. (2), we estimated that the net production of $CH_4$ in wetland soils was 2002 g $CH_4$ (2.0 mg $CH_4$ $m^{-2}$ $d^{-1}$; Table 3). We also found that even if dissolved $CH_4$ in Skeleton Creek was entirely evaded to the atmosphere in the wetland (i.e. not oxidized within soils), it was still a very small component (< 1 %) of $F_{CH_4}$ compared to net production in soils (Table 3).

## 4 Discussion

### 4.1 Factors driving $CH_4$ consumption within polar desert soils

The range in mean growing season $F_{CH_4}$ at our desert site during five growing seasons (−0.9 to −1.8 mg $CH_4$ $m^{-2}$ $d^{-1}$) was similar to $F_{CH_4}$ measured at other dry soils in Arctic and temperate ecosystems (∼ 0 to −3.5 mg $CH_4$ $m^{-2}$ $d^{-1}$; Supplement Table S4; King et al., 1997; Smith et al., 2000; Olefeldt et al., 2013). Methanotrophs use $CH_4$ as their primary carbon and energy source for metabolism and in dry soils, rates of methanotrophy are controlled by factors that (1) deliver $CH_4$ and oxygen into soils (Benstead and King, 1997; Flessa et al., 2008); (2) allow passage and replenishment of these gases where methanotrophs reside (Moosavi and Crill, 1998); and (3) facilitate heat transfer and increase soil temperatures where methanotrophs inhabit (Christensen et al., 1999).

The bulk density and gas diffusivity of upper soil horizons affect diffusion rates of atmospheric gases into soils (Smith et al., 2000). No vegetation canopy, high wind speeds and surface roughness promote the exchange of gases between the soil surface and the atmosphere by increasing the concentration gradient for $CH_4$ and oxygen from the soil to the atmosphere. The barren and flat terrain with large fetch and some surface roughness (1.5 cm) at the polar desert site would have promoted sustained gas exchange in this way.

Soil moisture is also a crucial factor influencing methanotrophy within dry soils. As water content increases, it replaces gas-filled pore spaces, leading to reduced diffusivity and thereby restricting oxygen (and $CH_4$) replenishment required for microbial metabolism. Whalen and Reeburgh (1996) found that methanotrophic rates peaked near 20 % soil moisture (v/v) in boreal soils before decreasing substantially towards saturation. At our desert site, soils were sandy, well-drained, and typically between 9–16 % v/v at 5 cm below the surface. We found little association between within-season $F_{CH_4}$ rates and soil moisture, suggesting that desert soil moisture content was well below a threshold where moisture restricted gas availability for methanotrophs, and above the threshold where desiccation restricted microbial activity. This conclusion was further supported by the chamber results where $F_{CH_4}$ was similar in chambers with and without vascular vegetation, suggesting that the moister vegetated soils were still within a moisture range that sustained methanotrophs without restricting gas transport.

The significant $CH_4$ consumption rates at the desert through the measurement period in each year (Fig. 2) were also a function of relatively warm soil temperatures (Table 1) since methanotrophy is a microbial metabolic process. Despite the high latitude of our site, near-surface soils were warm with little variation through the measurement period (Supplement Fig. S3). This region experiences low cloud cover relative to much of the high Arctic (Thompson, 1994), resulting in high daily isolation. The deep, narrow valley structure of the watershed also retains heat more efficiently than other Arctic locales (Thompson, 1994).

### 4.2 Factors driving $CH_4$ emission from meadow wetland soils

$F_{CH_4}$ measured in our wetland margin chambers (−0.05 to +0.43 mg $CH_4$ $m^{-2}$ $d^{-1}$) and using the EC technique (−0.84 to +2.73 mg $CH_4$ $m^{-2}$ $d^{-1}$) were considerably lower than other low Arctic and sub-Arctic wetlands (Olefeldt et al., 2013; Supplement Table S4). $CH_4$ is produced by methanogenic bacteria as a by-product of carbon metabolism in anaerobic soil environments and several factors control its production and release to the atmosphere, including (1) soil moisture/water table position (Moosavi and Crill, 1997; Christensen et al., 2000); (2) soil temperature (Christensen et al., 1995; Nakano et al., 2000; Ström et al., 2012); (3) vegetation species composition and primary productivity rates (Christensen et al., 1999; Ström et al., 2012); and (4) substrate availability (Ström et al., 2012).

Saturated, poorly draining soils may sustain anaerobic conditions crucial for methanogens and also reduce habitat for $CH_4$-consuming methanotrophs above water tables. Soils in our wetland margin collars switched abruptly from net $CH_4$ consumption to net $CH_4$ emission when Skeleton Creek water saturated the previously dry organic soils. However, the EC flux tower measurement (near-constant $CH_4$ emission) integrated the full wetland area, suggesting that a significant portion of the wetland within the flux footprint was constantly near or at saturation following the rapid thaw period. This also may explain the lack of correlation between the soil moisture measured at the wetland margin and tower $F_{CH_4}$. Other studies have shown that $F_{CH_4}$ may cease to relate to soil moisture once saturation occurs (Heikkinen et al., 2002), and we suspect that was the case at our site.

Temperature influences $CH_4$ production and emission from wetlands in cold environments (van Huissteden et al., 2005). Soil temperature was strongly associated with $F_{CH_4}$ measured by the EC tower primarily as a consequence of the switch from $CH_4$ uptake to loss during soil thaw. After this period, soil temperatures were relatively stable, as discussed above. Without more variation in soil temperature during the growing season, it is difficult to assess the sensitivity of $F_{CH_4}$ at these higher soil temperatures (> 8–12 °C). Although soil

temperatures at our wetland were generally lower than at wetlands which emit large amounts of $CH_4$ in low Arctic regions (e.g. Parmentier et al., 2011), we found that other high Arctic wetlands with similar soil temperatures still emitted significantly more $CH_4$ to the atmosphere (Christensen et al., 1995; Friborg et al., 2000; Tagesson et al., 2012) than at our site. Soil temperatures, therefore, did not appear to fully explain the low $CH_4$ fluxes at our wetland.

Several Arctic studies have demonstrated the importance of plant structures and root exudates to the emission of $CH_4$ from wetlands (e.g. Ström et al., 2012). Certain aerenchymous plants are known to be important conduits of $CH_4$ to the atmosphere (e.g. Eriophorum, Carex; Ström et al., 2003). Plants also release carbon and nutrient-rich exudates from roots during growth, supplying methanogenic communities with key substrates. At our wetland site, vegetation cover included a substantial portion of Eriophorum and Carex species (Edlund, 1994) similar to other high Arctic wetlands (e.g. Ström et al., 2003). Since $F_{CH_4}$ measured by the EC tower correlated best with $F_{CO_2}$ during thaw and through the growing season (Fig. 4, Supplement Table S3), this suggests either plant productivity and/or plant-mediated transport of $CH_4$ may have been important in driving the seasonal variations in $F_{CH_4}$ at the wetland. However, $F_{CO_2}$ rates at our wetland were comparable to other high Arctic wetlands (Friborg et al., 1997; Tagesson et al., 2012) and others much further south (e.g. Lafleur et al., 2012, Humphreys and Lafleur, 2011), suggesting that plant productivity also cannot explain low rates of $CH_4$ emission from our wetland.

Substrate quantity and quality are key factors supporting microbial viability in soils. Peat accumulates in cold wetland environments because cold temperatures restrict microbial decomposition of fresh litter, while saturation limits more efficient aerobic degradation pathways. Peat can be a high-quality carbon source for microbes in Arctic wetlands because of its high labile carbon content (Updegraff et al., 1995). At Zackenburg, GL (74° N, 20° W), where soil temperatures and $CO_2$ fluxes were similar to our site, peat depths extended to over 30 cm, encompassing most of the active layer during the growing season (Christensen et al., 2000). At the centre of our wetland, the organic layer was only 7 cm thick with a sharp transition to mineral soil (Supplement Fig. S4). Therefore, approximately one-quarter to two-thirds of the wetland active layer was comprised of organic-poor mineral soils likely not ideal for substantial microbial activity. Further, the shallow mineral soils and flow-through nature of our wetland may have made strong oxidizing species more available for microbial communities, and thus restricting methanogen activity (Lipson et al., 2012). Therefore, this wetland site, and presumably its low $CH_4$ emissions, was distinguished from other high Arctic wetlands. The reason for a roughly 7–10 cm deep accumulation of organic materials at the centre of this wetland, despite $CO_2$ uptake rates comparable to other high Arctic wetlands, may be due to its young age or could be due to other factors, such as redox conditions, that limit carbon accumulation and do not support methanogenesis.

### 4.3 $CH_4$ transport and transformations through a high Arctic wetland

The Skeleton Creek wetland complex is a typical meadow wetland within the Lake Hazen watershed and includes soils and productive lakes which are potentially important $CH_4$ emission sources to the atmosphere. Chemistry sampling of six aquatic sites upstream of the wetland (Fig. 1) showed significant changes in dissolved $CH_4$ concentrations (Supplement Table S5). Low $CH_4$ concentrations occurred in permafrost meltwater (PF sites), high concentrations and emission rates were observed in productive lakes (Skeleton Lake, Pond 11; unpublished data), and concentrations declined downstream in the creek (stream sites) and wetland areas (wetland inflow, outflow) due to a combination of evasion and/or oxidation. These results suggest that wetland complexes in the watershed are comprised of potential "hot-spots" of $CH_4$ production and emission with very little lateral transfer of $CH_4$ between these systems and to Lake Hazen. This model of $CH_4$ flow differs from similar studies in the south which showed greater importance of lateral $CH_4$ transport in streams (e.g. Dinsmore et al., 2010).

One of these "hot-spots" of $CH_4$ production and emission was at our wetland site. Because Skeleton Creek delivered only small amounts of $CH_4$ to the wetland (despite high $CH_4$ concentrations draining from Skeleton Lake), we found that the majority of $CH_4$ emitted by the wetland was from $CH_4$ produced within its soils (Table 3). Although we did not measure pore water $CH_4$ within the wetland, these results suggest that $CH_4$ emissions in the wetland were due to in situ production exceeding oxidation, even if we assume all creek $CH_4$ was evaded and included in $F_{CH_4}$ measurements. Bacterial production of $CH_4$ in wetland soils was further supported by chemistry results downstream of the wetland which showed signatures of anaerobic microbial activity in the form of (1) increased $NH_4^+ : NO_3^-$ ratios; (2) increases in dissolved organic matter; and (3) decreases in oxidation-reduction potential. However, it is unclear if fast stream flow velocity and short water residence times in the wetland affected ultimate concentrations and redox potentials measured in stream water exiting the wetland. For example, redox potential measurements in stream water exiting the wetland ($\sim +20$ mV) were generally higher than expected for $CH_4$-producing environments, possibly indicating that stream flow rates were too high to accumulate significant dissolved $CH_4$ and lower redox potentials.

### 4.4 $CH_4$ fluxes in the high Arctic and future climate

Most $CH_4$ studies on Arctic landscapes focus on emission sources to the atmosphere, such as peatlands and wetlands, because of their considerable coverage in the low- and

sub-Arctic, and their important role in global $CH_4$ budgets (O'Connor et al., 2010; Kirschke et al., 2013). Results from the Lake Hazen watershed suggest that $CH_4$ consumption, not emission, is the larger, more consistent pattern of $F_{CH_4}$ in the high Arctic because of limited wetland and pond coverage (Lehner and Döll, 2004). The $CH_4$ consumption rates at Lake Hazen and other locations across the high Arctic (Flessa et al., 2008; Lamb et al., 2011) suggest that this region cannot be overlooked as an important consumer of atmospheric $CH_4$. For example, within Quttinirpaaq National Park, approximately 99 % of the plant-habitable zone in the Park (22 672 $km^2$) is considered to have moderate- to well-drained soils (Edlund, 1994) compared to only 1 % classified as saturated or poorly drained soils. Considering the extensive area of dry, upland landscapes in the broader high Arctic, substantially more $CH_4$ measurements on dry soils are required to better delineate areas of $CH_4$ consumption and ultimately support more robust Arctic $CH_4$ models.

Future changes in soil temperature and moisture (ACIA, 2005) are expected to have landscape-level effects in the Arctic, with some models predicting 18 % of polar desert regions being replaced with southern tundra species by 2080, relative to 1960 (Sitch et al., 2003). Results from our contrasting high Arctic landscapes suggest that soil moisture, soil temperature, and substrate quantity are key factors determining the magnitude and direction of $F_{CH_4}$ for these landscapes. However, future changes within each ecosystem will likely result in different $F_{CH_4}$ responses. Polar desert soils are mostly well-drained mineral soils with pockets of cryoturbated organic matter (Tarnocai et al., 2001). We found that $CH_4$ consumption rates were affected by soil temperature, but not vegetation cover. Therefore, we may expect that warming temperatures and longer growing seasons may increase $CH_4$ consumption rates. Predicted increases in precipitation and permafrost meltwater on the landscapes, at least in the short term, will likely not affect $CH_4$ consumption rates substantially because of this coarse-textured soil's poor ability to retain water. Until the soils develop greater organic matter content capable of retaining more water (to the point of limiting diffusivity), these soils should continue to consume $CH_4$ in a warmer and wetter climate. In the wetland, our EC measurements and mass budget analysis indicated that $CH_4$ emission rates to the atmosphere were very low. Although warming air temperatures and permafrost thaw should support methanogenic activity in the future, until substantial organic carbon accumulation occurs in this system, methanogenesis and thus $CH_4$ emission to the atmosphere will likely continue to be limited in poorly draining soils in the Lake Hazen watershed. The rate at which landscapes can change is an important unknown for the future cycling of GHGs at this high latitude.

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