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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 desert and wetland landscapes on northern Ellesmere Island, Canada, we made static chamber measurements at both locations over five growing seasons and eddy covariance (EC) measurements at the wetland in 2012. Chamber measurements revealed that desert soils consumed CH_4 (-1.37 ± 0.10 mg-CH_4 m^{-2} d^{-1}) whereas the wetland emitted CH_4 ($+0.22 \pm 0.19$ mg-CH_4 m^{-2} d^{-1}). Desert CH_4 consumption rates were positively correlated 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 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 methanogenisis 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). For obvious reasons, 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 ~ 600 ppb for the majority of past 600 000 yr (Kirschke et al., 2013). Although monitor-

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ing 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 glaring 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 remote high Arctic 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 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 early June 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, 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 (∼ 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 (∼ 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.

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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 eddy co-
5   variance (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 closed path 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.,
10   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
15   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) $CO_2\backslash$ water vapour (H_2O) infrared gas analyzers (Fig. S1). In addition
20   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 (Table S1). Signals from all sensors were recorded as half-hour means on CSI CR3000-XT dataloggers. Thaw depth was monitored weekly at ten
25   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 and wetland sites (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* (Figs. 1 and 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 (Figs. 1 and S1). Each collar enclosed vegetation of similar type and cover as the rest of the wetland. Chambers were deployed at each site every five to seven 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 analyses 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. The coefficients of determination were typically strong for both desert and wetland measurements ($r^2 = 0.9 \pm 0.2$). 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} (see Results section).

2.2.3 Eddy covariance measurements

As described above, although EC technology is not new, only recently has a low-power, robust CH_4 analyzer become available. We had the opportunity during the 2012 growing season to deploy an LI-7700 open-path CH_4 gas analyzer 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 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 analyzer measurements using a covariance maximization approach. Because the LI-7700 is an open-path analyzer, 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, and when the friction velocity fell below $0.1 \, \mathrm{m \, s^{-1}}$, 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 ±3 SD of the growing season mean were also removed. These corrections resulted in 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 discuss 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 two to five days during the 2011 and 2012 growing seasons (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 analyze 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 analyses 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

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taken at each site at time of sampling using a YSI 556 MPS multi probe system. Water flow at each site was measured every two to three days using a Pygmy current meter. At each site, we chose a channelized section of 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 ($\pm1SE$) of 1.37 ± 0.10 mg CH_4 m^{-2} d^{-1} ($n = 27$), whereas the wetland site emitted 0.22 ± 0.19 mg CH_4 m^{-2} d^{-1} ($n = 18$) to the atmosphere (Fig. 2). Desert soils consistently consumed atmospheric CH_4 throughout the growing season, whereas wetland soils typically consumed atmospheric CH_4 during the first two weeks of July (0.40 ± 0.12 mg CH_4 m^{-2} d^{-1}) before transitioning to a source of CH_4 to the atmosphere (0.72 ± 0.22 mg CH_4 m^{-2} d^{-1}) (Fig. 3). These differences in F_{CH_4} between the desert and wetland were related to the large differences in soil moisture and soil temperature (Fig. 2). Daily mean soil moisture at 5 cm depth of the desert soils was consistently near 15.1 ± 1.0 % v/v during the measurement period, except during short rain events. Wetland soil moisture at the same depth was significantly higher (75.2 ± 3.2 %; two-sample t test, $t(43) = 20.9$, $p < 0.001$) than at the desert. Because the wetland was bowl-shaped, snow melt 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 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. Near-surface soils at the desert site were warmer

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through the growing season as described by the significant differences in 5 cm depth soil temperature ($14.4 \pm 0.5\,°C$-desert vs. $10.4 \pm 0.5\,°C$-wetland; $t(43) = 5.5$, $p < 0.001$), 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; $t(42) = 7.0$, $p < 0.001$) and thaw depths (0.88 ± 0.03 m-desert vs. 0.29 ± 0.01 m-wetland; $t(39) = 18.2$, $p < 0.001$; Fig. 2).

Between 2008 and 2012 at the desert site, mean growing season F_{CH_4} ranged between -0.91 and $-1.78\,mg\,CH_4\,m^{-2}\,d^{-1}$ (Table 1, Fig. 3). CH_4 consumption rates were positively correlated with soil temperatures between years ($r^2 = 0.99$; multiple correlation) but not influenced by changes in soil moisture ($r^2 = 0.03$). Consumption rates of CH_4 were not significantly different in chambers with or without vegetation (paired t test; $t(26) = 0.2$, $p = 0.83$). Associations between within-season F_{CH_4} and environmental factors were generally weak ($-0.28 < \rho < 0.07$; Spearman Rank Correlation; Table S2-A.). From 2010 to 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. 3). With only 3 yr 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 ($\rho = 0.72$; 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} ($\pm 1SE$) of $1.27 \pm 0.18\,mg\,CH_4\,m^{-2}\,d^{-1}$ at the wetland (Fig. 4a) with no discernible diurnal patterns (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 seven 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 and 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).

Seasonal variations in F_{CH_4} correlated with F_{CO_2} ($\rho = -0.71$) and soil temperature (T_{SOIL}; $\rho = 0.65$) (Table S3). These associations were used to develop the following parabolic model:

$$F_{CH_4} = 23.4 + -0.5 \cdot F_{CO_2} + -5.4 \cdot T_{SOIL} + -0.1 \cdot F_{CO_2}^2 + 0.3 \cdot T_{SOIL}^2. \tag{1}$$

The proportion of variation in daily F_{CH_4} described by the model was 67 % (Fig. 5; Table S4). The model did not improve when a soil moisture term was included. Using Eq. (1) and F_{CO_2} and T_{SOIL} data from previous years, F_{CH_4} was predicted to have been much greater in 2010 and 2011 than in 2012 (Fig. 5). Chamber data also showed higher F_{CH_4} in 2011 compared to 2012 (Table 1), however too few chamber measurements were taken in 2010 to confidently evaluate its difference from other years.

3.3 Wetland aquatic chemistry

Flow-weighted mean dissolved CH_4 concentrations (± 1wSD) in Skeleton Creek water (Table 2) decreased from $0.029 \pm 0.018\,\mu mol\,L^{-1}$ upstream of the wetland to

$0.002 \pm 0.002 \, \mu mol \, L^{-1}$ downstream of the wetland between 28 June and 02 August in 2011, a decrease of 94 %. In 2012 between 03 July and 05 August, dissolved CH_4 concentrations decreased 70 % through the wetland from $0.005 \pm 0.004 \, \mu mol \, L^{-1}$ to $0.01 \pm 0.005 \, \mu mol \, L^{-1}$, though concentrations were low. Ammonium concentrations increased downstream each year while nitrate concentrations decreased or were below the analytical detection limit. Concentrations of dissolved organic nitrogen and carbon, and particulate matter 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 ($\sim$ late June to early August), we can calculate the net production of CH_4 (production-oxidation losses) in wetland soils using the following equation:

$$\Sigma(I_{CH_4} + NP_{CH_4})_{daily} = \Sigma(F_{CH_4} + O_{CH_4})_{daily} \qquad (2)$$

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. I_{CH_4} and O_{CH_4} were calculated using:

$$I_{CH_4} \text{ or } O_{CH_4}(g) = (([CH_{4t1}] + [CH_{4t2}])/2) \cdot V \qquad (3)$$

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 found that the net production of CH_4 in wetland soils was over twice as much in 2011 (4 612 g CH_4; 4.4 mg $CH_4 \, m^{-2} \, d^{-1}$) compared to 2012 (2 002 g CH_4; 2.0 mg $CH_4 \, m^{-2} \, d^{-1}$; Table 3). During 2011, soils were warmer and wetter with stronger F_{CO_2} compared to 2012 (Table 1). 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}; Table S5; 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 reside (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 soils. As water content increases, it replaces gas-filled pore spaces leading to reduced diffusivity 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. These results also suggested that there was no enhancement of methanogen activity in association with vegetation cover.

The significant CH_4 consumption rates at the desert through the measurement period in each year (Fig. 3) 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 (Fig. S3). This region experiences low cloud cover relative to much of the high Arctic (Thompson, 1994) resulting in high daily isolation (Fig. 4). 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 (-0.12 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; Table S5). 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 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., 2002a), and we suspect that was the case at our site.

Temperature influences CH_4 production and emission from wetlands, even 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 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 substrate. At our wetland site, vegetation cover included a substantial portion of Eriophrorum 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, 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, GR (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 (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. 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 (Table S6). Low CH_4 concentrations occurred in permafrost melt water (PF sites), high concentrations and emission rates were observed in productive lakes (Skeleton Creek, Pond 11; St. Louis, 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 2). 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 and particulate 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 methane-producing environments, possibly indicating that stream flow rates were too high to accumulate significant dissolved CH_4 and lower redox potentials.

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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 consumption rates at Lake Hazen and other regions (Lamb et al., 2011; Flessa et al., 2008) across the high Arctic, 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 melt water on the landscapes, at least in the short-term, will

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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 mass budget analysis indicated some differences in CH_4 emission rates between the contrasting hydrological and meteorological conditions of 2011 and 2012, but overall emission rates were still 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 GHG's at this high latitude.

Supplementary material related to this article is available online at http://www.biogeosciences-discuss.net/11/1673/2014/ bgd-11-1673-2014-supplement.pdf.

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Table 1. Mean (±1SE) daily methane 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 ± 0.05	4	7.3 ± 0.6	520 ± 22	14.6 ± 2.7	17.0 ± 2.2	11.4 ± 0.4	24
2009	−1.49 ± 0.12	5	10.0 ± 0.7	678 ± 20	26.9 ± 1.1	9.4 ± 0.1	12.9 ± 0.5	20
2010	−1.54 ± 0.07	4	9.8 ± 0.6	685 ± 26	28.8 ± 1.9	17.5 ± 0.2	12.7 ± 0.5	25
2011	−1.78 ± 0.20	7	9.5 ± 0.3	678 ± 22	32.1 ± 1.4	16.9 ± 0.1	13.7 ± 0.3	34
2012	−0.91 ± 0.23	7	8.1 ± 0.4	520 ± 32	26.5 ± 2.0	15.4 ± 0.1	11.1 ± 0.4	33
Wetland								
2010	−0.05 ± 0.29	4	11.0 ± 0.8	652 ± 24	7.8 ± 0.8	79.0 ± 1.3	10.6 ± 0.4	22
2011	0.43 ± 0.44	7	10.7 ± 0.4	657 ± 21	6.4 ± 0.6	80.6 ± 0.6	10.8 ± 0.2	34
2012	0.16 ± 0.14	7	9.1 ± 0.5	507 ± 31	8.1 ± 0.4	58.2 ± 1.3	8.2 ± 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

Table 2. Flow-weighted mean concentrations (±1wSD) of several chemicals in Skeleton Creek water upstream and downstream of the wetland during the 2011 and 2012 growing seasons. All chemicals are reported in µmol L^{-1} except for water temperature (°C) and oxidation-reduction potential (mV).

Parameter	2011			2012		
	Wetland Inflow	Wetland Outflow	% change	Wetland Inflow	Wetland Outflow	% change
Dissolved CH$_4$	0.029 ± 0.018	0.002 ± 0.002	−94 %	0.005 ± 0.004	0.001 ± 0.005	−70 %
Dissolved CO$_2$	52 ± 17	61 ± 16	17 %	72 ± 29	65 ± 10	−9 %
NO$_3^-$	0.10 ± 0.07	0.06 ± 0.03	−42 %	0.04 ± 0.00	0.04 ± 0.00	0 %
NH$_4^+$	0.99 ± 0.25	1.01 ± 0.26	1 %	0.53 ± 0.13	0.76 ± 0.07	42 %
DON	17.5 ± 0.5	20.6 ± 1.3	17 %	19.4 ± 1.0	22.0 ± 0.5	13 %
TDN	18.5 ± 0.7	21.7 ± 1.3	17 %	20.0 ± 0.9	22.8 ± 0.5	14 %
DOC	421 ± 21	479 ± 27	14 %	497 ± 28	549 ± 19	10 %
Particles (PN)	0.41 ± 0.09	1.07 ± 0.68	160 %	n/a	n/a	n/a
Water $_T$	10.8 ± 0.9	11.2 ± 0.9	3 %	8.1 ± 1.6	7.8 ± 1.4	−4 %
ORP	n/a	n/a	n/a	53 ± 57	21 ± 17	−60 %

NO$_3^-$: dissolved nitrate; NH$_4^+$: dissolved ammonium; DON: dissolved organic nitrogen; TDN: total dissolved nitrogen; DOC: dissolved organic carbon; PN: particle-bound nitrogen; Water$_T$: water temperature; ORP: oxidation-reduction potential.

Table 3. Wetland mass balance (Eq. 2) of CH_4 for the 2011 and 2012 growing seasons including stream input (I_{CH_4}) and output (O_{CH_4}), flux of methane (F_{CH_4}) and estimates of net methane production within wetland soils (NP_{CH_4}). Asterisk indicates F_{CH_4} to the atmosphere was modelled.

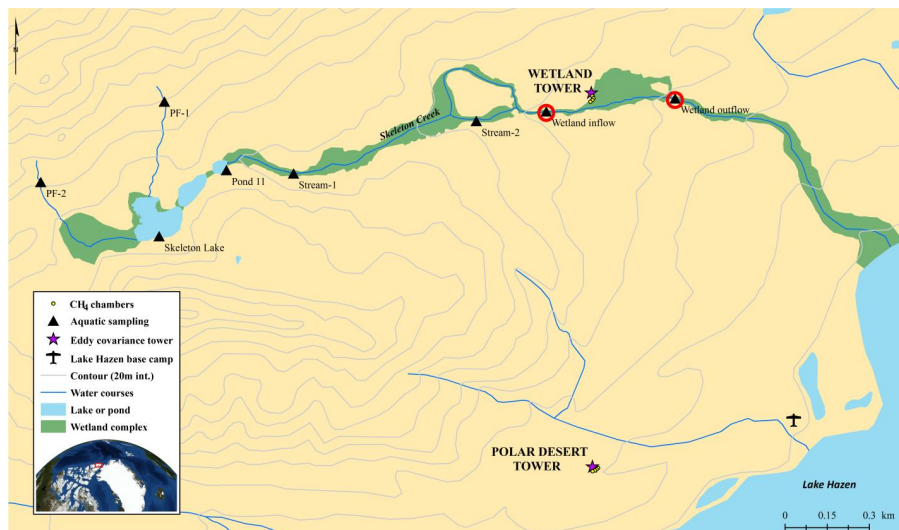|	Stream flow (m^3)	n (#)	CH$_4$ transfer (g)
2011 (36 days from 28 Jun to 2 Aug)			
I_{CH_4}	29 216	17	13.8
O_{CH_4}	25 580	17	0.0
F_{CH_4}*	–	n/a	4625
NP_{CH_4} (4.4 mg CH_4 m^{-2} d^{-1})	–	–	4612
2012 (34 days from 3 Jul to 5 Aug)			
I_{CH_4}	6578	8	0.5
O_{CH_4}	5451	6	0.1
F_{CH_4}	–	34	2002
NP_{CH_4} (2.0 mg CH_4 m^{-2} d^{-1})	–	–	2002

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Fig. 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 CH4 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.

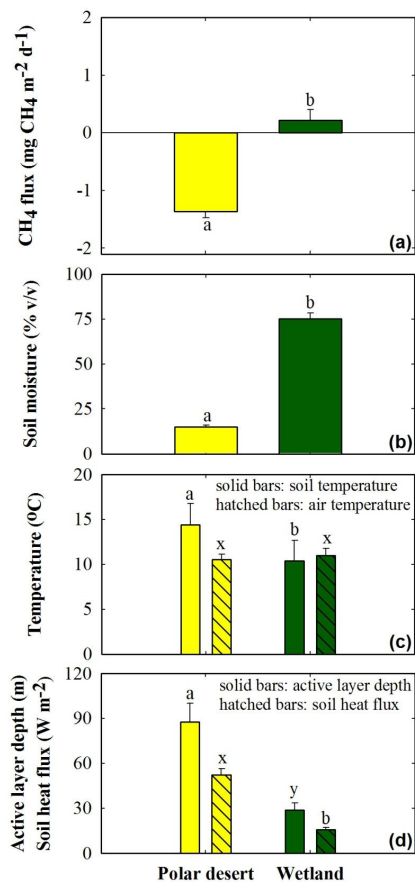**Fig. 2.** Comparison of 2008–2012 mean methane flux (F_{CH_4}; ±1SE) measured in chambers **(a)** and other environmental variables **(b–d)** between the desert ($n = 27$) and wetland ($n = 18$) sites. Letters indicate if there were statistically significant differences between sites.

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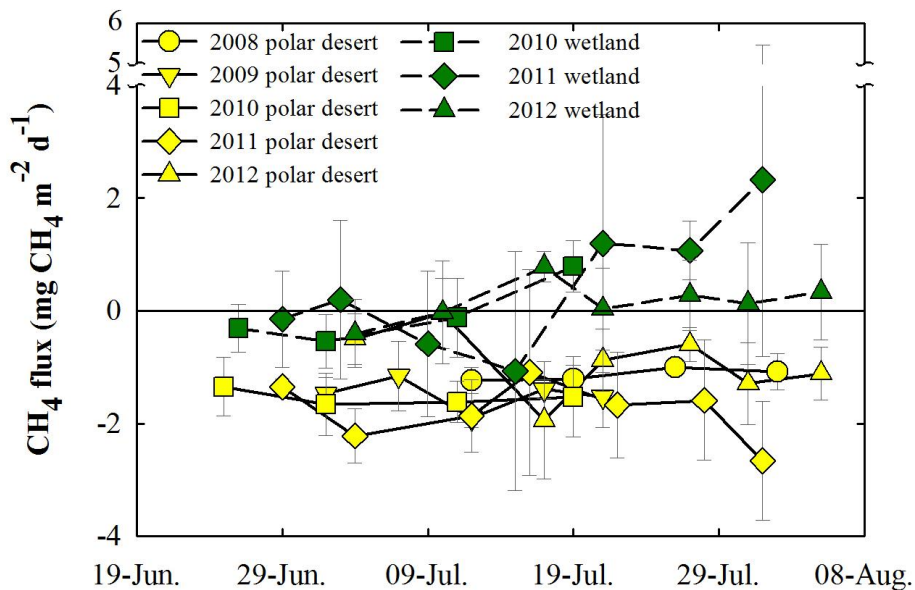**Fig. 3.** Mean methane fluxes (F_{CH_4}; ±1SD) from four polar desert and four wetland static chambers during the 2008 to 2012 growing seasons.

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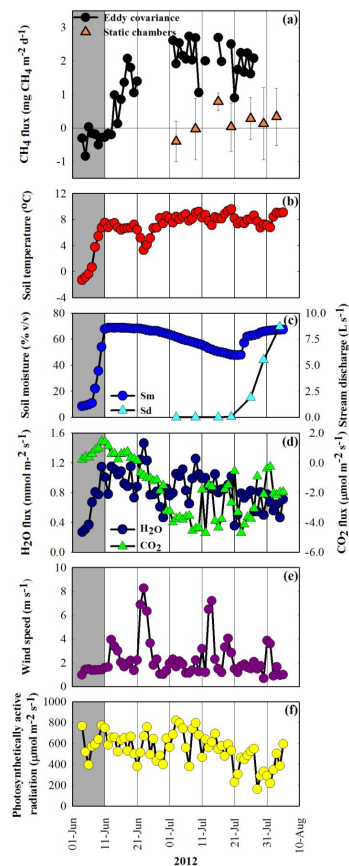
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Fig. 4. Comparison of mean daily eddy covariance methane flux (F_{CH_4}); **(a)** at the wetland and several mean daily environmental variables **(b–f)** during the 2012 growing season. Shaded bars highlight the period of rapid soil thaw.
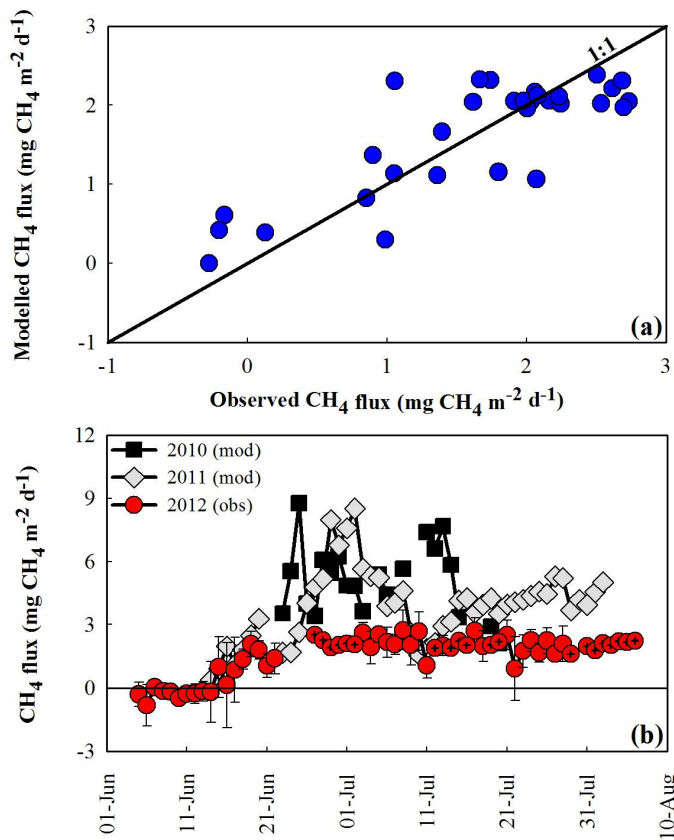
Fig. 5. Observed mean daily methane fluxes (F_{CH_4}) measured by eddy covariance at the wetland during the 2012 growing season vs. modelled 2012 F_{CH_4} as described in Eq. (1) **(a)**. Modelled data were not statistically different from observed data (paired t test; $p = 0.98$). Observed, gap-filled (cross-hairs) F_{CH_4} (±95 % CI) during 2012 and modelled fluxes during 2010 and 2011 are shown in **(b)**. Modelled F_{CH_4} using LI-7500 CO_2 fluxes are shown.

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