Modern to millennium-old greenhouse gases emitted from ponds and lakes of the Eastern Canadian Arctic (Bylot Island, Nunavut)

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Abstract. Ponds and lakes are widespread across the rapidly changing permafrost environments. Aquatic systems play an important role in global biogeochemical cycles, especially in greenhouse gas (GHG) exchanges between terrestrial systems and the atmosphere. The source, speciation and emission rate of carbon released from permafrost landscapes are strongly influenced by local conditions, hindering pan-Arctic generalizations. This study reports on GHG ages and emission rates from aquatic systems located on Bylot Island, in the continuous permafrost zone of the Eastern Canadian Arctic. Dissolved and ebullition gas samples were collected during the summer season from different types of water bodies located in a highly dynamic periglacial valley: polygonal ponds, collapsed ice-wedge trough ponds, and larger lakes. The results showed strikingly different ages and fluxes depending on aquatic system types. Polygonal ponds were net sinks of dissolved CO₂, but variable sources of dissolved CH₄. They presented the highest ebullition fluxes, 1 or 2 or more magnitudes higher than from other ponds and lakes. Trough ponds appeared as substantial GHG sources, especially when their edges were actively eroding. Both types of ponds produced modern to hundreds of years old (< 550 yr BP) CO₂ and CH₄, even if trough ponds could contain much older carbon (> 2000 yr BP) derived from freshly eroded peat. Lakes had small dissolved and ebullition fluxes, however they released much older GHG, including millennium-old CH₄ (up to 3500 yr BP) from lake central areas. Acetoclastic methanogenesis dominated at all study sites and there was minimal, if any, methane oxidation in gas emitted through ebullition. These findings provide new insights on GHG emissions by permafrost aquatic systems and their potential positive feedback effect on climate.

1 Introduction

Permafrost stores large quantities of carbon compared to the atmosphere, although quantitative estimates are still under discussion (Tarnocai et al., 2009; Hugelius et al., 2014). Climate warming impacts Arctic landscapes through permafrost thawing and erosion (Romanovsky et al., 2010). This results in the release of both old and recent organic carbon to the atmosphere as greenhouse gases (GHG) (Zimov et al., 2006; Schuur et al., 2015). Widespread across permafrost environments, aquatic systems act as biogeochemical hotspots by releasing substantial amounts of carbon dioxide (CO₂) and methane (CH₄) (e.g., Walter et al., 2007; Laurion et al., 2010; Abnizova et al., 2012). It is generally considered that CH₄ ebullition is the main mechanism of GHG emissions from ponds and lakes, a transport mechanism highly heterogeneous in space and time (Wik et al., 2011). However, other processes, such as emissions through diffusion (Bastviken et al., 2008), plant-mediated transport and microbial oxidation (Bastviken et al., 2004; Liebner et al., 2011), also need to be considered in the specific context of the Arctic. Moreover, lateral inputs of CH₄ produced within the active layer or lateral export of permafrost carbon away from thaw sites via streams and rivers were recently demonstrated (vonk and
Overall, thermokarst (thaw) ponds and lakes represent a major landscape feature in permafrost-affected regions (Grosse et al., 2013), and there is a growing interest in defining the specific role of various types of freshwater ecosystems in global carbon dynamics associated to permafrost degradation, and how they may rapidly respond to environmental changes (see Vonk et al., 2015, and other articles in this special issue).

Upscaling and modeling GHG emissions is challenging, and oversimplified assumptions can lead to large calculation errors (Stepanenko et al., 2011; van Huissteden et al., 2011; Gao et al., 2013). The gaps that need to be fulfilled to model future GHG emissions with more accuracy include defining the vertical distribution of carbon in permafrost soils across the Arctic, the interactions between permafrost thaw and surface hydrology, as well as distinguishing CH4 from CO2 emissions and gradual warming from abrupt thaw mechanisms (Schuur et al., 2015). Regarding thermokarst systems specifically, aspects that should be further investigated include physical (e.g., heat transfer, diffusive GHG exchange, daily storage flux) and hydrological (e.g., surface and groundwater flows) dynamics, as well as fluxes of particulate and dissolved organic carbon to these systems (Vonk et al., 2015). Another important yet rarely considered aspect is the age (old vs. modern) of the carbon that is processed and released by these biogeo systems, which is linked to their potential to generate a positive feedback on climate (Walter et al., 2006; Vonk et al., 2013; Mann et al., 2015). Large GHG emissions (especially CH4) from old (late Pleistocene-age) organic ice-rich loess permafrost (yedoma) have been reported from thermokarst lakes of Siberia and Alaska in regions that were not ice-covered during the last glaciation (Zimov et al., 1997; Brosius et al., 2012). In Canada, which accounts for a very large portion of circum-Arctic permafrost, these deposits are rare as the territory was almost entirely covered by ice sheets during that period (Dyke and Prest, 1987). The carbon trapped in permafrost is thus younger (Holocene-age) in this part of the Northern Hemisphere (Allard, 1996; Burn and Kokelj, 2009; Lauriol et al., 2010; Tremblay et al., 2014). It nevertheless represents an excess carbon stock that can contribute to accelerate climate warming via a positive feedback mechanism if released as GHG, compared to modern carbon that is used and recycled through short-term biogeochemical processes (photosynthetic fixation and microbial respiration).

Preliminary data on GHG radiocarbon age from small tundra ponds on Bylot Island (Nunavut) in the Eastern Canadian Arctic showed that the carbon released by these systems was generally modern (Negandhi et al., 2013). The objective of the present study was to further characterize GHG composition, production pathway, age and emission rates in ponds and lakes at this particular site. We analyzed dissolved and ebullition gas samples collected in July from ponds and lakes located within an organic-rich permafrost terrace of Late Holocene age (Fortier et al., 2006).

2 Study area

Bylot Island (Nunavut) is located in the Eastern Canadian Arctic, within the continuous permafrost zone (Fig. 1). The Byam Martin Mountains run southeast–northwest across the island, and the plains that stretch out on either side of the mountains belong to the Arctic Lowlands physiographic region (Bostock, 1970). The numerous valleys formed in the lowlands were shaped during the successive Pleistocene glaciations (Klassen, 1993). Since the Holocene, these valleys developed highly dynamic biogeo systems rich in permafrost ground ice, peat, and aquatic environments (Fortier and Allard, 2004). The study site (73°09′ N; 79°58′ W) is located in one such valley (glacier C-79) named Qarlikturvik, which has a NE-SW orientation and a surface area of ~65 km2 (~15 km-long × 4.5 km-wide). A terminal moraine, located about halfway between the actual glacier front and the seashore and sitting on marine clay, was 14C-dated to ~9.8 kyr BP (Allard, 1996). Glacial retreat, accompanied by a marine transgression phase, ended around 6 kyr BP. The clays were then covered by glacio-fluvial sand and gravels (Fortier and Allard, 2004). Today, a proglacial braided river runs through a glacio-fluvial outwash plain and drains glacier melt waters and sediments towards the Navy Board Inlet, where it forms a delta.

The outwash plain is bordered on both sides by a 3 to 5 m thick terrace, crisscrossed by networks of tundra polygons associated with the formation of syngenetic ice wedges (Figs. 1d and 2a). Along the southern bank of the river, the upper portion of the terrace is composed of alternating organic (peat) and mineral (wind-blown sand and silt) material, which started to accumulate over glacio-fluvial sands and gravels around 3700 years ago (Fortier and Allard, 2004). These peaty loess deposits contain excess pore ice (>100 % dry weight) and their gravimetric organic matter content can reach over 50 %. The active layer depth in such deposits generally ranges between 40 to 60 cm, and the maximum depth of permafrost on Bylot Island has been estimated to be over 400 m (Smith and Burgess, 2000). The terrace comprises abundant aquatic systems of different sizes and shapes (Fig. 2) that can act as effective biogeochemical hotspots (Lauriol et al., 2010; Negandhi et al., 2013). The hydrological network is mainly fed by rain and snowmelt runoff originating from gullies of the valley flanks or large snow banks on the lee side of hills. Most of water loss from ponds and lakes is through evaporation during the ice-free season (Negandhi, 2013).

The climate normal (1981–2010) is provided by a meteorological station located near the village of Pond Inlet (Mittimatik) (72°41′ N; 77°58′ W), about 85 km southeast from the study site (Fig. 1c). The region has a polar climate.
with a slight marine influence, a mean annual air temperature of $-14.6^\circ C$ (average daily temperatures ranging from $-33.4^\circ C$ in January to $6.6^\circ C$ in July) and total precipitations of 189 mm, of which 91 mm fall as rain between June and September (Environment Canada, 2015). Thawing and freezing degree-days are around 475 and 5735, respectively.
Winter (continuous daily mean air temperature $< 0^\circ \text{C}$) lasts from early September to mid-June, for an average total of 283 days per year. A station from the SILA network, operated since 2004 by the Center for Northern Studies (CEN) in the valley of glacier C-79, provides similar climate data (CEN, 2014).

The southwest plain of Bylot Island is a $\sim 1600 \text{km}^2$ low-lying wetland area of graminoid-moss tundra (Parks Canada, 2014). Local vegetation in the Qarlikturvik valley is dominated by sedges (e.g. Carex aquatilis var. stans, Eriophorum scheuchzeri), grasses (e.g. Arctagrostis latifolia, Diapontia fischeri, Pleurogon sabinei) and mosses (e.g. Drepanocladus spp., Aulocornium spp.) (Duclos, 2002; Ellis et al., 2008).

3 Materials and methods

3.1 Sampling sites

We selected and sampled different types of aquatic systems typical of the tundra polygon terrace of the valley (Fig. 2; Table 1): (1) polygonal ponds over low-centered ice wedge polygons; (2) elongated water channels over melting ice wedges (ponds formed in collapsed ice-wedge troughs, hereafter referred to as trough ponds); (3) lakes with underlying talik (unfrozen soil over permafrost), including a thermokarst (thaw) lake and a kettle (melted buried glacier ice) lake. A total of 23 ponds and lakes were sampled in June–July 2013, including 9 polygonal ponds, 12 trough ponds, and 2 lakes (1 thermokarst and 1 kettle lake). In July 2014, six water
Table 1. Limnological properties of ponds and lakes sampled in July 2013 and July 2014, including sampling depth, dissolved organic carbon (DOC), absorption coefficient of dissolved organic matter at 320 nm ($a_{320}$), total phosphorus (TP), soluble reactive phosphorus (SRP), total nitrogen (TN), and selected major ions (NO$_3$, SO$_4$, Fe). POL = polygonal pond; IWT = ice wedge trough pond; LAK = lake.

<table>
<thead>
<tr>
<th>Site</th>
<th>Type</th>
<th>Depth (m)</th>
<th>DOC (mg L$^{-1}$)</th>
<th>$a_{320}$ (m$^{-1}$)</th>
<th>TP (µg L$^{-1}$)</th>
<th>SRP (µg L$^{-1}$)</th>
<th>TN (mg L$^{-1}$)</th>
<th>NO$_3$ (mg L$^{-1}$)</th>
<th>SO$_4$ (mg L$^{-1}$)</th>
<th>Fe (mg L$^{-1}$)</th>
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</tr>
<tr>
<td>BYL30</td>
<td>POL</td>
<td>surf</td>
<td>8.7</td>
<td>17.8</td>
<td>14.8</td>
<td>NA</td>
<td>0.49</td>
<td>0.42</td>
<td>1.3</td>
<td>0.47</td>
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<td>surf</td>
<td>5.6</td>
<td>9.0</td>
<td>22.9</td>
<td>NA</td>
<td>0.42</td>
<td>0.09</td>
<td>1.3</td>
<td>0.25</td>
</tr>
<tr>
<td>Average POL ($n = 9$)</td>
<td>6.7</td>
<td>12.7</td>
<td>17.6</td>
<td>0.47</td>
<td>0.17</td>
<td>1.3</td>
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<td>surf</td>
<td>6.6</td>
<td>27.0</td>
<td>16.1</td>
<td>NA</td>
<td>0.29</td>
<td>0.37</td>
<td>4.3</td>
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<td>surf</td>
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<td>42.0</td>
<td>29.0</td>
<td>NA</td>
<td>0.58</td>
<td>0.07</td>
<td>6.2</td>
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<td>Average IWT ($n = 12$)</td>
<td>10.0</td>
<td>38.0</td>
<td>27.8</td>
<td>0.63</td>
<td>0.19</td>
<td>6.7</td>
<td>1.014</td>
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<td>BYL66</td>
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<td>surf</td>
<td>4.2</td>
<td>16.4</td>
<td>20.7</td>
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<td>surf</td>
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<tr>
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<td>surf</td>
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<td>NA</td>
<td>8.5</td>
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<td>1.32</td>
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<td>NA</td>
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<td>1.25</td>
<td>&lt; 0.2</td>
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<td>0.266</td>
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<td>IWT</td>
<td>surf</td>
<td>8.8</td>
<td>NA</td>
<td>23.7</td>
<td>1.28</td>
<td>1.02</td>
<td>0.30</td>
<td>1.3</td>
<td>1.549</td>
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<tr>
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<td>IWT</td>
<td>surf</td>
<td>12.1</td>
<td>NA</td>
<td>27.4</td>
<td>1.56</td>
<td>1.22</td>
<td>0.29</td>
<td>2.7</td>
<td>0.487</td>
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<td>LAK</td>
<td>surf</td>
<td>4.3</td>
<td>NA</td>
<td>9.8</td>
<td>&lt; 0.5</td>
<td>0.49</td>
<td>&lt; 0.2</td>
<td>2.4</td>
<td>2.949</td>
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<tr>
<td>BYL36</td>
<td>LAK</td>
<td>surf</td>
<td>4.3</td>
<td>NA</td>
<td>4.2</td>
<td>10.6</td>
<td>0.74</td>
<td>0.44</td>
<td>&lt; 0.2</td>
<td>2.5</td>
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</tbody>
</table>
| * 2011 data

bodies (two polygonal ponds, two trough ponds, and two lakes including one thermokarst and one kettle lake) were selected and studied more intensively, including morphological measurements of ponds (depth, width and length) and lakes (bathymetry with a portable sonar as in Bouchard et al., 2015), and limnological profiles (see below).

3.2 Limnology

We measured a suite of limnological characteristics during both years, including temperature, dissolved oxygen, and concentrations of dissolved organic carbon (DOC), chromophoric fraction of dissolved organic matter (CDOM), nutrients (phosphorus, nitrogen) and major ions. Temperature and dissolved oxygen profiles were recorded with a ProODO handheld meter (YSI Inc.). Water samples were filtered through 0.2 µm pre-rinsed cellulose acetate filters (2013) or pre-combusted GF/F filters (2014, nominal porosity 0.7 µm) to analyze DOC and major ions. Cations were fixed with HNO$_3$ (0.15 % final concentration) while anions and DOC were not fixed but kept in dark and cold. DOC concentrations were measured with a Shimadzu TOC-5000A carbon analyzer calibrated with potassium biphthalate, and CDOM was quantified (in 2013 only) with the absorption coefficient of DOM at 320 nm ($a_{320}$) obtained on a Cary 300 (Varian; methodological details in Laurion and Mladenov, 2013). Major anions were quantified by ionic chromatography (Dionex ICS-2000), whereas major cations by inductively coupled plasma–optical emission spectrometry (ICP-OES, Varian VISTA AX). Total phosphorus (TP) and total nitrogen (TN) were quantified from unfiltered water samples fixed with H$_2$SO$_4$ (0.15 % final concentration) as described by Stainton et al. (1977). Finally, the thermal structure of one trough pond (BYL27) was assessed during a full year (July 2013–July 2014) by recording water temperature at two depths (0 and 50 cm) at a 15 min interval using two submersible data loggers (Vemco Minilog-II-T, accuracy ±0.1 °C, resolution ±0.01 °C) installed on a mooring line. The line was not moored at the deepest point of the pond (~1 m) as found upon its retrieval, but the data still provide a clear picture of the thermal stratification establishing in this type of humic ponds.

3.3 Ebullition flux of greenhouse gases

Ebullition gas samples were collected using submerged funnels (as in Wik et al., 2013) equipped with a 140 mL plastic syringe (Fig. A1 in Appendix A) and deployed for a
period of 1 h to 19 days depending on the flux. The samples trapped in the syringe were transferred into 50 mL glass bottles with butyl rubber stoppers (bottles acid-washed, pre-combusted, helium flushed and vacuumed) for \(^{14}\text{C}\) dating (see below), and into two separate 6 mL glass vials (helium flushed and vacuumed Exetainers) for stable isotope (see below) and gas chromatography analysis (Varian 3800, COMBI PAL head space injection system, CP PoraPLOT Q 0.53 mm ID \(\times\) 25 m, flame ionization detector). Ebullition flux (\(F_e\), in \(\text{mmol m}^{-2}\text{d}^{-1}\)) was calculated as

\[
F_e = \frac{(p_{\text{Gas}} \times V)}{(A \times MV \times t)},
\]

where \(p_{\text{Gas}}\) is the partial pressure of \(\text{CO}_2\) or \(\text{CH}_4\), \(V\) is the collected gas volume, \(A\) is the funnel area, \(MV\) is the gas molecular volume at ambient air temperature, and \(t\) is the collecting time.

### 3.4 Diffusive flux of greenhouse gases

Surface water dissolved GHG concentrations were obtained by equilibrating 2 L of lake or pond water with 20 mL of ambient air during 3 min (Hesslein et al., 1991). The resulting gaseous headspace was transferred into 6 mL glass vials and analyzed as above by gas chromatography. Dissolved GHG concentration at the surface (\(C_{\text{sur}}\)) was calculated using Henry’s law, and departure from saturation (sink vs. source) was calculated subtracting the gas concentration in the water at equilibrium with the atmosphere (\(C_{\text{eq}}\), global values of atmospheric partial pressures from IPCC, 2007, were used). To estimate diffusive flux (\(\text{Flux}_d\)), first the gas transfer coefficient (\(k_{600}\)) standardized to a Schmidt number (\(Sc\)) of 600 (Wanninkhof, 1992) was calculated with the wind-based model of Cole and Caraco (1998):

\[
k_{600} = 2.07 + 0.215u_{10}^{0.7},
\]

where \(u_{10}\) is the wind speed at 10 m above the ground, and then applying the equation:

\[
\text{Flux}_d = k(C_{\text{sur}} - C_{\text{eq}}),
\]

where \(k\) is the gas transfer coefficient for a given gas calculated as

\[
k = k_{600}(Sc/600)^{-0.5}.
\]

### 3.5 Radiocarbon analysis

Ebullition gas samples were analyzed at the Keck Carbon Cycle AMS facility at the University of California, Irvine. First, \(\text{CH}_4\) and \(\text{CO}_2\) were separated and purified by a zero air carrier gas flow-through line (Pack et al., 2015), and graphitized by the sealed tube \(\text{Zn}\) reduction method (Xu et al., 2007), then measured for radiocarbon (\(\text{^{14}C}\)) on a compact accelerator mass spectrometer (AMS) (Southon and Santos, 2007). Data presented here are expressed as \(\Delta^{14}\text{C} (\%e)\), which is normalized to radiocarbon activity of an oxalic acid standard OX1 (decay corrected to 1950) and corrected for isotopic fractionation (Reimer et al., 2004). \(\Delta^{14}\text{C} (\%e) > 0\) was further used to indicate “modern” carbon (1950 to present), and \(\Delta^{14}\text{C} (\%e) < 0\) for “older” carbon (pre-1950). This was particularly helpful for polygonal and trough ponds, which provided modern or very young GHG. The \(\Delta^{14}\text{C}\) analytical error was \(\approx 2\%\) for modern sample, based on long-term measurements of secondary standards. \(^{14}\text{C}\) age (yr BP) is as defined by Stuiver and Polach (1977).

### 3.6 Stable isotope analysis

Stable carbon and hydrogen isotopic compositions of GHG, \(\delta^{13}\text{CO}_2\), \(\delta^{13}\text{CH}_4\), and \(\delta\text{DCH}_4\), were analyzed at the Biogeochemistry Facility School of Earth and Ocean Sciences (BFSEOS, University of Victoria). Ebullition gas samples were analyzed for \(\delta^{13}\text{CH}_4\) by introducing the gas onto a GSQ plot column (0.32 mm ID, 30 m) using a Valco 6-port valve and sample loop. After chromatographic separation, the \(\text{CH}_4\) passes through an oxidation oven (1030°C), a Nafion water trap, and open-split interface to a continuous flow-isotope ratio mass spectrometer (CF-IRMS). The \(\delta^{12}\text{CO}_2\) was measured similarly by CF-IRMS, but bypassing the combustion oven. Precision for the \(\delta^{13}\text{CH}_4\) and \(\delta^{13}\text{CO}_2\) analyses was \(\pm 0.2\%\), relative to Vienna PeeDee Belemnite (VPDB). Hydrogen isotope ratios of \(\text{CH}_4 (\delta\text{DCH}_4)\) were measured by a TC/EA pyrolysis unit (1450°C) interfaced to a CF-IRMS. Precision for the \(\delta\text{DCH}_4\) analyses was \(\pm 3\%\), relative to Vienna Standard Mean Ocean Water (VSMOW). Carbon and hydrogen isotope ratios are expressed using standard delta (\(\delta\)) notation as described by deviations from a standard such that

\[
\delta_{\text{sample}}(\%e) = [(R_{\text{sample}} / R_{\text{standard}}) - 1] \times 1000,
\]

where \(R\) is the \(^{13}\text{C} / ^{12}\text{C}\) or \(^2\text{H} / ^1\text{H}\) ratio in the sample or standard. For isotope calibration, methane carbon and hydrogen standards from Isometric Instruments were used. These are traceable back to VPDB for carbon isotope ratios and VSMOW for hydrogen isotope ratios.

### 4 Results

#### 4.1 Morpho-limnological properties of ponds and lakes

Ponds were generally shallow (\(\sim 0.6–1.0\) and 1.0–1.5 m deep for polygonal and trough ponds, respectively) and thus froze to the bottom during winter, whereas lakes were more variable in depth depending on their origin and at least a portion of them did not freeze to the bottom in winter. The thermokarst lake was a few meters deep (< 5 m), while the kettle lake was deeper (< 12 m). Polygonal ponds, including different developmental stages and coalesced ponds, generally had flat bottoms covered by cyanobacterial mats (up
Figure 3. Temperature (°C; upper x axes) and dissolved oxygen (%; lower x axes) profiles for polygonal ponds BYL30 (a) and BYL80 (b), trough ponds BYL24 (c) and BYL27 (d), and lakes BYL66 (e) and BYL36 (f). Some profiles (a–c) were taken in July 2013, whereas the others (d–f) were taken in July 2014. Note the different vertical scales (depth).

to 5 cm thick), and stable (non-eroding) shores (Fig. 2b, c). Their surface area varied substantially (from 21 to 3350 m²) with a median of around 160 m². Trough ponds were elongated water channels (median width ∼ 3 m; median length ∼ 10 m), and their shores were either actively eroding with collapsing decimetric peat blocks (Fig. 2f), or stable and colonized by brown mosses (Fig. 2g). The thermokarst lake had sharp edges near the shore, a shallow and gently sloping lake bottom and a deeper central basin. The kettle lake had steeper slopes along its margins, and showed a deep section that was not in the center of the lake (Bouchard et al., 2015).

Ponds and lakes showed contrasting physicochemical conditions during the 2 sampling years (Table 1). Trough ponds generally had the highest concentrations of DOC, nutrients and ions, followed by polygonal ponds, whereas lakes showed the lowest values. Trough pond BYL27, where shore
erosion was active during summer time, had near- or higher-than-average concentrations, whereas trough pond BYL24, with stable shores, showed lower-than-average values. Pond DOC, nutrient and ion concentrations were substantially higher in 2014, a particularly dry year (total precipitations from January to June = 27.0 mm in 2014, compared to...
50.7 mm in average; Table B1 in Appendix B), with resulting low pond water levels as observed in the field. When considering specific solute species separately, all of them except NO₃⁻ and SO₄²⁻ were statistically different (p < 0.0001) among aquatic system types in 2013. In contrast, in 2014 only DOC (p < 0.0001), total nitrogen (TN, p < 0.001) and soluble reactive phosphorus (SRP, p < 0.05) showed significant differences, and only between lakes and ponds (i.e., not between polygonal and trough ponds). Among all the water chemical properties and regardless of the sampling year, DOC showed the highest statistical contrasts between the different types of water bodies.

Polygonal ponds (BYL30, BYL80) had a thermally homogenous and well-oxygenated water column in July, whereas trough ponds (BYL24, BYL27) were notably stratified (Figs. 3 and 4). Thermokarst lake BYL66 was relatively well mixed over most of the water column, except near the sediment-water interface where dissolved oxygen decreased rapidly. Kettle lake BYL36, deeper than the other sampled water bodies, showed a steep gradient between the warmer, well-oxygenated epilimnion and the much colder, anoxic hypolimnion. The thermal profiles on Fig. 3 are representative of the conditions generally prevailing from July to mid-August in each type of water bodies.

4.2 Age and concentration of greenhouse gases released through ebullition

Radiocarbon age (Δ¹⁴C signature) and concentration of GHG (CO₂ and CH₄) emitted through ebullition showed strikingly different trends between the various types of aquatic systems (Fig. 5). Polygonal and trough ponds produced modern CH₄ and modern to a few hundred years old (< 550 yr BP) CO₂, whereas lakes generally released older GHG, ranging from 510 to 1425 yr BP for CO₂ and from 125 to 3405 yr BP for CH₄ (Table 2). Moreover, samples from lake edges had younger and less concentrated CH₄ than those coming from lake central area. No such trend was observed for CO₂ in lakes. Considering all ponds and lakes as a whole, CH₄ was generally 1 to 2 orders of magnitude more concentrated than CO₂ in emitted bubbles in July.

4.3 Dissolved and ebullition fluxes of greenhouse gases

Polygonal ponds were generally CO₂ sinks, but they were CH₄ sources with a relatively broad range of saturation levels (~0–2.4 µM) (Fig. 6). Lakes were near the equilibrium with the atmosphere (all samples clustered near 0 for both gases), being small sinks or sources of CO₂, and small sources of CH₄. Trough ponds were in general supersaturated in both gases, especially when their margins were actively eroding (highest GHG saturation values) (Fig. C1). Trough ponds showed the highest diffusive flux, especially of CO₂ (65.5 mmol m⁻² d⁻¹; Table 3) with a median diffusive CO₂ flux (21.8 mmol m⁻² d⁻¹) more than 12 times higher than the median value of all sampled water bodies (1.7 mmol m⁻² d⁻¹). Polygonal ponds, on the other hand, showed the highest ebullition flux for both CO₂ (16.3 mmol m⁻² d⁻¹) and CH₄ (534.5 mmol m⁻² d⁻¹), with a median ebullition CH₄ flux that, although relatively low (~1.0 mmol m⁻² d⁻¹), was ~5 times higher than the median value for all ponds and lakes (~0.2 mmol m⁻² d⁻¹). Lakes generally showed the lowest fluxes (both diffusion and ebullition). Globally, diffusion appeared as the dominant mechanism for CO₂ emission, whereas CH₄ was mainly emitted through ebullition. Statistical tests ran on the GHG data showed that trough ponds (BYL24, BYL27) were significantly different (p < 0.001) from the other two types of water bodies (polygonal ponds and lakes), but also from each other. Furthermore, dissolved CO₂ and CH₄ fluxes were significantly correlated (p < 0.006) with CDOM (r = 0.79 and 0.78, respectively; N = 22; a320 only available in 2013), but only CH₄ fluxes were correlated (p < 0.003) with DOC (R = 0.61, N = 28; data available in both years).

4.4 Carbon and hydrogen stable isotope ratios in ebullition gas samples

The stable isotope ratios of methane (δ¹³CH₄, δDCH₄) and carbon dioxide (δ¹³CO₂) were measured on 18 ebullition samples collected in 2013 and 2014 (Table 2; Figs. 7 and 8). The δ¹³CH₄ average values were −60.5‰ and ranged from −52.1‰ to the most ¹³C-depleted value of −67.6‰, both from polygonal ponds. The δDCH₄ values, which averaged −376.8‰, were relatively ²H-depleted for naturally
occuring methane. The $\delta^{13}{\text{CH}}_4$ with the most $^2$H-enriched value came from the thermokarst lake sample collected at its center ($−319.56\permil$; BYL66; Fig. 7). In contrast, the $\delta^{13}{\text{CH}}_4$ values from trough ponds (BYL24 and BYL27) were consistently and extremely $^2$H-depleted, with values from $−397.7\permil$ to a very low value of $−448.1\permil$. There was no apparent correspondence between the methane concentration and $\delta^{13}{\text{CH}}_4$ or $\delta^{13}{\text{DCH}}_4$. The CO$_2$ contents of ebullition samples were sometimes insufficient for carbon isotope measurements. For those with more CO$_2$, the average $\delta^{13}\text{CO}_2$ was $−14.3\permil$ and varied from +0.3 (polygonal pond BYL80) to $−21.8\permil$ (trough pond BYL24). There was also no apparent correspondence between the CO$_2$ concentration and $\delta^{13}\text{CO}_2$.

However, it is worth noting that the sample with the most $^{13}$C-enriched CO$_2$ also corresponded to the one with the most $^{13}$C-depleted CH$_4$ (polygonal pond BYL80; Fig. 8).

### 5 Discussion

#### 5.1 The strong heterogeneity in greenhouse gas age and concentration

We observed large variability in the age, composition and emission rate of GHG released by the studied aquatic systems. The GHG escaping through ebullition ranged from modern to a few centuries old for polygonal and trough ponds.
Figure 8. Carbon isotope composition ($\delta^{13}$C) of CH$_4$ (x axis) and CO$_2$ (y axis) emitted by the sampled ponds and lakes. HM = hydrogenotrophic methanogenesis; AM = acetoclastic methanogenesis; MO = methane oxidation.

and from a few centuries to a few millennia old for lakes (Fig. 5). Trough ponds emitted slightly but significantly older CH$_4$ than polygonal ponds ($\Delta^{14}$C = 10$\pm$18 $\sigma$ vs. 43$\pm$28 $\sigma$, respectively; $p < 0.05$), as observed earlier at the same site (Negandhi et al., 2013), although still classified as modern carbon, suggesting a small contribution of peat-derived carbon pool to microbial activity in trough ponds. Surprisingly, trough ponds did not emit millennium-old CH$_4$, at least in July, despite the fact that they were exposed to eroding peat from down to the base of the active layer in the surroundings (14C dates ranging from $\sim$ 2.2 to 2.5 kyr BP; Table 2) and even older peat strata up-thrust along ice wedges by cryo-turbation and now in contact with surface waters (Fortier and Allard, 2004). Eroding peat was likely leaching old carbon into the water column, but bottom sediment interstitial water, where CH$_4$ is mostly produced, did not predominantly emit carbon of this age. Permafrost disturbance was indeed shown to deliver millennia-old particulate organic carbon and DOC to arctic streams and rivers (Lamoureux and Lafrenière, 2014; Guo et al., 2007; Vonk et al., 2013), acting as a significant source of bioavailable carbon in Arctic freshwaters (Mann et al., 2015). We speculate that microbes were preferably using young carbon, putatively more labile and more abundant at this time of the year, and may use older carbon stocks later when primary producers are less active. If the CH$_4$ released from trough ponds is indeed older during the autumn and spring, this could represent a positive climate feedback, but our results now indicate a limited role.

On the other hand, CH$_4$ ebullition samples collected from lakes provided older dates, up to nearly 3500 yr BP (thermokarst lake BYL66), which is very close to the maximum known age of the permafrost peat layers in the valley (3670 $\pm$ 110 yr BP; Fortier and Allard, 2004). It may suggest that permafrost thaw underneath this lake has proceeded through the organic layers at this site, which could result in decreased emissions in the future after the microbial exhaustion of the labile fraction of the organic matter pool (Walter et al., 2007). However the timing of this reduction is unknown. We observed a spatial gradient in the age and concentration of CH$_4$ in bubbles emitted from the thermokarst lake, with younger and less concentrated CH$_4$ from the lake edge ($\sim$ 3%), and older and more concentrated CH$_4$ from the center (up to 57%). The development of a talik (unfrozen soil under lake) explains the mobilization of deeper and older CH$_4$ at the lake center where water remains unfrozen under the ice cover in winter (maximum lake depth $> 4$ m, ice cover thickness $\sim 2$ m). Methane emitted from a given location would thus be composed of a mixture of young CH$_4$ from the edge with older CH$_4$ from the center (Fig. D1). To our knowledge, the only other studies of thermokarst lakes presenting 14C dates on GHG are in yedoma deposits (Alaska, Siberia), which have very different ground ice, sediment and organic carbon contents, and chronostratigraphic history. For these lakes, the release of very old (> 40 kyr BP) and highly concentrated (up to 90%) CH$_4$ from deep unfrozen lake sediments has been found (Walter et al., 2008). However, this study also reported younger ages for ebullition samples emitted from different parts of the lakes, and generally younger towards the lake center (when from background ebullition). At our study site, even though older GHG were emitted from lakes compared to ponds, ebullition fluxes remained low during the study period (July). Walter-Anthony and Anthony (2013) concluded that the classic randomized bubble-trap method for estimating mean lake ebullition is highly median-biased toward underestimation of fluxes, and this was possibly also occurring for our data set, although no systematic point source studies have been conducted so far at our study site.

We also observed strong differences in dissolved GHG flux depending on pond and lake types (Fig. 6; Table 3): polygonal ponds were CO$_2$ sinks but CH$_4$ sources while trough ponds were significant sources of both GHG, as previously reported in the valley (Laурion et al., 2010; Negandhi et al., 2013), and lakes were small sources of GHG. This pattern can be explained by the morpho-limnological properties of the water bodies. Polygonal ponds had stabilized shores (no apparent slumping) and more transparent waters compared to other systems, as shown by their lower CDOM content (Laурion et al., 2010). Moreover, they had flat and shallow bottoms covered by abundant cyanobacterial mats actively photosynthesizing and acting as a relatively efficient CO$_2$ sink (flux reaching $-11.8$ mmol m$^{-2}$ d$^{-1}$). This is however 1 order of magnitude lower than the net ecosystem CO$_2$ uptake measured over the summer from a wet polygonal tundra site in Siberia (flux reaching $-104.7$ mmol m$^{-2}$ d$^{-1}$; Kutzbach et al., 2007). Bottom sediments of the studied polygonal ponds were also colonized by methanotrophic bacteria (Negandhi et al., 2014), which can be a significant control mechanism on CH$_4$ emissions as shown in polygonal ponds of the Lena region (Liebner et al., 2011).
Lakes were larger and deeper, thus they were exposed to wind-induced mixing of their epilimnetic waters promoting venting of the GHG from this layer. When the water column is seasonally stratified (like in BYL36), the hypolimnion likely stores a large fraction of the GHG produced by the lake until the autumnal overturn period (Bastviken et al., 2004), allowing more space and time for the oxidation of dissolved CH₄, and for the dissolution of a fraction of ebullition CH₄ (Bastviken et al., 2008). Therefore, it is possible that higher flux of old carbon would be observed later in the season. To fully account GHG emissions from lakes and compare them to other aquatic systems, summer and winter storage fluxes will need to be estimated (Boereboom et al., 2012; Langer et al., 2015; Walter-Anthony et al., 2010; Wik et al., 2011).

Trough ponds presented the highest combined (CO₂ + CH₄; diffusion + ebullition) GHG fluxes at the time of sampling (Fig. 9). Considering a global warming potential (GWP) of 34 for CH₄ on a 100-year horizon (Myhre et al., 2013), trough ponds presented the highest net carbon efflux (1.5 g CO₂-equivalent m⁻² d⁻¹, compared to 0.7 and 0.2 g CO₂-equivalent respectively for polygon ponds and lakes). Despite their shallow depths, trough ponds were strongly stratified with oxygen-depleted and cold bottom waters. The bottom temperature in these ponds was indeed near 0°C (Fig. 3) because this layer of water is lying just above the melting ice wedge (as part of the active layer), it does not mix with surface waters and it is cooled down through sensible heat transfer. Moreover, trough ponds were not colonized by photosynthesizing (CO₂ sink) and methanotrophic (CH₄ sink) bacteria such as in polygonal ponds (Negandhi et al., 2014). Stronger water column hypoxia generated anoxia more rapidly in the sediments, and the organic material inputs caused by active erosion likely led to higher CH₄ production, although the young carbon signature of emitted CH₄ is still puzzling (see below). Meanwhile, the eroding conditions and reduced light availability (higher CDOM, TP and turbidity; Table 1) in trough ponds favored net heterotrophy and net CO₂ emissions, as found in subarctic thermokarst lakes (Roiha et al., 2015). Similar to polygonal ponds, the shallow depth of trough ponds reduces the chances for dissolution of CH₄ bubbles into the water column and its subsequent oxidation before reaching the atmosphere. Moreover, the thermal structure of trough ponds (low transparency, microtopography) can impede mixing for several weeks (Fig. 4), thus favoring GHG summer storage in bottom waters, and likely generating stronger diffusive flux later at the autumnal overturn period. Thermal structure might become even stronger in years of low precipitations such as in 2014, when concentrations of solutes (DOC, ions) increase through evaporation, intensifying density gradients thus GHG storage. Diffusive CH₄ fluxes were indeed statistically higher (p < 0.01) in 2014 compared to 2013, although no such trend was observed for CO₂.

The highest GHG saturation levels observed over the sampling period were measured in a trough pond the day following a major erosion event (peat block collapsing in pond BYL27; Fig. C1). This might result from the disturbance of the thermal structure and transfer of stored GHG to the surface, or from the causal effect of a new input of organic matter to microbial activity. Active shore erosion around tundra ponds, potentially increasing CH₄ production by 2 to 3 orders of magnitude, has been reported from similar systems in Siberia (Langer et al., 2015), suggesting a direct impact of permafrost slumping on GHG emissions. The effect of erosion events on GHG flux must be further evaluated as other factors, such as fluctuating wind and air temperature, can also influence mixing and surface GHG concentrations (Tedford et al., 2014).

Interestingly, we also observed substantial differences in GHG concentrations among trough ponds, some presenting much lower values. Trough ponds such as BYL24 (Fig. 2g) had relatively stable (non eroding) shores, and were colonized by abundant vegetation dominated by brown mosses. Methane oxidation by bacteria associated with submerged brown mosses has been reported in Siberian ponds, contributing to smaller CH₄ concentrations in these ecosystems (Liebner et al., 2011). Therefore, there might be cases where
the methanotrophic community is also efficient in limiting CH$_4$ emissions from trough ponds (Negandhi et al., 2014).

5.2 Production pathways of CO$_2$ and CH$_4$

We obtained different radiocarbon ages for CO$_2$ and CH$_4$ within the same ebullition samples, as collected from funnels placed at the water surface (Table 2, Fig. 5), suggesting that GHG production was derived from different carbon sources. This divergence in carbon age was even more pronounced for the lakes, where it could reach almost 3000 years. The presence of unfrozen sediment layers (talik) underneath the lakes would explain the older bubbling CH$_4$ emitted from deeper and older sediments exposed to microbrial degradation, as found in thermokarst lakes of Siberia and Alaska (Walter et al., 2007). Younger CO$_2$ could then be explained by a larger contribution of younger and shallower surface sediments to bacterial production and respiration. It could also result from lateral inputs of CO$_2$ produced within younger organic material or from exchanges with atmospheric CO$_2$.

On the other hand, century-old CO$_2$ collected from ponds in parallel to modern CH$_4$ is more difficult to explain. As stated above, emission of young CH$_4$ suggests the preferential use of modern carbon by methanogens, and also a dominance of background ebullition mode (from surface sediments) in thaw ponds. Meanwhile, emission of older CO$_2$ could be related to anaerobic CO$_2$ production in water-saturated and reductive soils and its subsequent lateral transport, as observed in a flooded tundra site in Alaska (Zona et al., 2012). Characterizing organic matter properties and oxidation versus reduction (redox) potential of pond and lake sediments at our study sites are required to confirm if such a mechanism can contribute to modern CH$_4$ emissions from surface layers and, at the same time, older CO$_2$ emissions from deeper layers. Moreover, a quantification of lateral fluxes of carbon within the active layer (groundwater and streams), an important yet rarely mentioned process driven by the coupling between carbon and water cycles (vonk and Gustafsson, 2013; Paytan et al., 2015), could help to better understand these results.

Notwithstanding the above-mentioned differences, the concentrations of CO$_2$ and CH$_4$ emitted through ebullition also need to be taken into account when evaluating the climate feedback potential of these emissions. Even though the age of CO$_2$ could reach several centuries (>1000 yr BP for one sample; Fig. 5), it was 1 to 2 orders of magnitude less concentrated in the emitted bubbles than CH$_4$. Hence, such emissions have a much lower potential to generate a positive feedback effect, at least during the ice-free season and under current climate conditions. Similar observations were reported from Siberian lakes, despite notably different geomorphological, geocryological and limnological conditions (Walter et al., 2007).

Methanogenesis in cold wetland systems typically proceeds via the anaerobic fermentation pathways of acetoclastic methanogenesis (AM) and/or hydrogenotrophic carbonate reduction methanogenesis (HM) (e.g., Kotsyurbenko et al., 2004; Alstad and Whiticar, 2011). AM utilizes the transfer of a CH$_3^-$ group from preformed organic substrates (i.e., acetate, methanol, methylated substrates, etc.), whereas HM utilizes H$_2$ and CO$_2$. Numerous studies have demonstrated the ability of using methane C and H isotope signatures to discriminate AM from HM pathways, and to characterize secondarily altered methane (oxidation, mixing, etc.). Polygonal ponds and lakes had combined methane C and H stable isotope signatures that were typical for methanogenesis dominated by AM, as clearly illustrated in the plot of $\delta^{13}$CH$_4$ versus $\delta$DCH$_4$ (Fig. 7). Trough ponds shared similar $\delta^{13}$CH$_4$ values with the other water bodies, but had substantially more $^2$H-depleted values ($\delta$DCH$_4$ from −398 to −448‰; Table 2, Fig. 7). These values are among the most $^2$H-depleted values known for naturally occurring methane (e.g., Whiticar, 1999). Although there was some variation between sites, the isotope signatures designate that all CH$_4$ emitted by ebullition in July is produced by AM, consistent with an earlier study at the same site (Negandhi et al., 2013). There is no indication of HM, which has a very different isotope signature, although the signature of samples collected from the center of lakes tend to lie towards the HM region, suggesting that a small proportion of the CH$_4$ produced could be through this pathway. This finding of AM dominance is consistent with ombrotrophic bogs with higher pH (ranging from $\sim$6.7 to 10.0 in 2014) compared with more acidic minerotrophic wetlands, which can be HM dominated (e.g., Bowes and Hornibrook, 2006; Prater et al., 2007). The dominance of AM is likely related to the carbon precursors; our sites may have more labile organic material present (e.g., organic acids) supporting acetoclastic methanogenesis and recently made available to methanogens. As this labile carbon pool is exhausted, the methanogenic pathway shifts from acetoclastic to more recalcitrant compounds and hence hydrogenotrophic methanogenesis (e.g., Alstad and Whiticar, 2011). It is therefore possible that other periods of the year would show a stronger HM signature, which would also be consistent with the presence of a large fraction of microbes able to perform HM in thaw ponds from this site (Negandhi et al., 2013).

Previous work in this valley indicated a significant relationship between water oxygen concentration and dissolved CH$_4$ oxidation level (Negandhi et al., 2013). This work also showed evidence that diffusive CH$_4$ was more susceptible to oxidation in polygonal ponds where a methanotrophic community was favored (Negandhi et al., 2014). This conclusion was supported by the strong shift in $\delta^{13}$CH$_4$ and $\delta$DCH$_4$ to the heavier isotopes, as expected (Whiticar et al., 1986). In the present study, there was no evidence of methane oxidation in any of the collected ebullition samples (Fig. 8), indicating that the conditions did not favor oxidation at the
5.3 Bylot ponds and lakes within the circumpolar North

The general topography and geology of the southwest plain of Bylot, together with the distinct local conditions of the Qrollertukvik valley (e.g., glacier and outwash plain activity, valley orientation in relation to dominant winds, snow cover depth and density), have contributed to the development over thousands of years of what is arguably one of the richest ecosystems in the region. However, taken separately, most of the landscape features in the valley (e.g., tundra polygons, ice-wedges, thermokarst ponds and lakes) are widespread across the Arctic (e.g., Walter-Anthony et al., 2010; Abnizova et al., 2012; Langer et al., 2015). When compared with flux values reported in the literature, our results, representing a snapshot of mid-summer conditions, generally appear in the range of what has been observed in other ponds and lakes from northern regions (Table 4). For example, we measured total CO$_2$ fluxes (diffusion + ebullition) of up to $\sim 0.8$ g C m$^{-2}$ d$^{-1}$, which is in the range of those reported from Alaska (0.7–2.3 g C m$^{-2}$ d$^{-1}$; Kling et al., 1992; Sepulveda-Jauregui et al., 2015), Siberia (0.02–1.1 g C m$^{-2}$ d$^{-1}$; Abnizova et al. 2012; Blodau et al., 2008), and Scandinavia (0.9–1.6 g C m$^{-2}$ d$^{-1}$; Huttunen et al., 2003; Kankaala et al., 2013). Methane fluxes (diffusion + ebullition) at our study site varied substantially (0.0005–6.4 g C m$^{-2}$ d$^{-1}$), but could reach values 1 order of magnitude higher than those from lakes in Alaska (0.01–0.5 g C m$^{-2}$ d$^{-1}$; Kling et al., 1992; Sepulveda-Jauregui et al., 2015; Walter-Anthony and Anthony, 2013) and Scandinavia (0.01–0.1 g C m$^{-2}$ d$^{-1}$; Bastviken et al., 2004; Huttunen et al., 2003; Kankaala et al., 2013). However, median values for polygonal and trough ponds ($\sim 0.02$ and 0.01 g C m$^{-2}$ d$^{-1}$, respectively) were more similar to published ranges. Yet, these fluxes were lower than those reported from Siberian thermokarst lakes in yedoma deposits (nearly 20 g C m$^{-2}$ d$^{-1}$; Walter-Anthony et al., 2010), which however include discrete ebullition seeps and hotspots that were not observed in our study, and most likely do not exist in the case of ponds.

Table 4. Greenhouse gas fluxes of CO$_2$ and CH$_4$ from high-latitude sites across the circum-Arctic. D = diffusion; E = ebullition.

<table>
<thead>
<tr>
<th>Reference</th>
<th>Region</th>
<th>Type</th>
<th>Mode</th>
<th>CO$_2$ (mg C m$^{-2}$ d$^{-1}$)</th>
<th>CH$_4$ (mg C m$^{-2}$ d$^{-1}$)</th>
<th>Notes</th>
</tr>
</thead>
<tbody>
<tr>
<td>Bouchard et al. (this study)</td>
<td>NE Canada</td>
<td>Polygon ponds</td>
<td>D + E</td>
<td>$-141.4$ 741.1</td>
<td>0.5 6432.0</td>
<td>July measurements</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Troughs</td>
<td>D + E</td>
<td>$-84.6$ 61.6</td>
<td>0.7 74.5</td>
<td></td>
</tr>
<tr>
<td>Laurion et al. (2010)</td>
<td>NE Canada</td>
<td>Subarctic ponds</td>
<td>D</td>
<td>27.6 746.4</td>
<td>0.4 5.4</td>
<td>July measurements</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Arctic ponds</td>
<td>D</td>
<td>$-246.0$ 1372.8</td>
<td>0.4 67.4</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Arctic lakes</td>
<td>D</td>
<td>$-63.6$ 70.8</td>
<td>0.1 0.4</td>
<td></td>
</tr>
<tr>
<td>Buell (2015)</td>
<td>NW Canada</td>
<td>Ponds</td>
<td>D + E</td>
<td>$-3.5$ 120.0</td>
<td>1.0 12.2</td>
<td>Headspace, chamber and flux tower methods</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>0.6 155.7</td>
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</tr>
<tr>
<td>Kling et al. (1992)</td>
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<td>Lakes and rivers</td>
<td>D</td>
<td>$-66.0$ 717.6</td>
<td>3.0 455.4</td>
<td>From annual fluxes, considering the ice-free period = 180 days</td>
</tr>
<tr>
<td>Walter Anthony and Anthony 2013</td>
<td>Alaska</td>
<td>Thermokarst lakes</td>
<td>E</td>
<td>51.9 2276.9</td>
<td>0.0 18716.8</td>
<td>Background + seep ebullition</td>
</tr>
<tr>
<td>Sepulveda-Jauregui et al. (2015)</td>
<td>Alaska</td>
<td>Lakes</td>
<td>D + E</td>
<td>200.0 1100.0</td>
<td>8.2 127.2</td>
<td>September measurements, flux tower</td>
</tr>
<tr>
<td>Walter-Anthony et al. (2010)</td>
<td>Alaska, Siberia</td>
<td>Thermokarst lakes</td>
<td>E</td>
<td>140.0 1586.7</td>
<td>0.6 26.7</td>
<td>From annual fluxes, considering the ice-free period = 180 days</td>
</tr>
<tr>
<td>Abnizova et al. (2012)</td>
<td>Siberia</td>
<td>Whole landscape</td>
<td>D + E</td>
<td>$0.0$ 1100.0</td>
<td>0.0 18716.8</td>
<td></td>
</tr>
<tr>
<td>Blodau et al. (2008)</td>
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<td>Ponds</td>
<td>D</td>
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<td>0.2 6.7</td>
<td>From annual fluxes, considering the ice-free period = 180 days</td>
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<tr>
<td>Kankaala et al. (2013)</td>
<td>Finland</td>
<td>Lakes</td>
<td>E</td>
<td>200.0 1100.0</td>
<td>8.2 127.2</td>
<td></td>
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<tr>
<td>Huttunen et al. (2003)</td>
<td>Finland</td>
<td>Lakes and reservoirs</td>
<td>D + E</td>
<td>$-21.6$ 876.0</td>
<td>0.8 99.6</td>
<td>From annual fluxes, considering the ice-free period = 180 days</td>
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<td>Bastviken et al. (2004)</td>
<td>Sweden</td>
<td>Lakes</td>
<td>D + E</td>
<td>51.9 2276.9</td>
<td>0.6 11.0</td>
<td></td>
</tr>
</tbody>
</table>

Production site (likely in anoxic sediment but also potentially in the water column; Grossart et al., 2011), and that the exchange with a pool of oxidized methane during the transport of bubbles to surface waters was undetectable, possibly linked to the short residence time. This was expected for shallow waters where bubbles can rapidly escape, but it was also the case in larger and deeper stratified lakes such as BYL36.
6 Conclusions

Aquatic systems are widespread across permafrost landscapes and play a crucial role in large-scale biogeochemical cycles. Yet, there is still much uncertainty about whether or not the Arctic can globally be considered a carbon source or sink, and how this will change in the coming decades. One element of such uncertainty is the highly heterogeneous distribution of ponds and lakes at the local scale and their different geomorphological and limnological properties, which influence their biogeochemistry and result in highly variable fluxes, especially for trough ponds. Our study demonstrates that local geomorphology and shoreline erosion around permafrost ponds and lakes can have a strong impact on their GHG concentrations and fluxes. We also report substantially different GHG ages among ponds and lakes of contrasting sizes and depths, and unexpectedly the emission of mainly modern CH$_4$ from trough ponds despite their exposure to a stock of eroding old carbon. Such results underscore the importance of the combined effects of geomorphology (talik development level, chronostratigraphy), limnology (organic matter concentration, CH$_4$ production and storage in anoxic/hypoxic bottom waters) and hydrology (lateral runoff inputs of organic material or GHG) on GHG emissions by permafrost thaw ponds and lakes. Interestingly, the significant correlation between GHG flux and DOM once more suggests the key role of this limnological characteristic, and calls for a deeper investigation as it could be used as a proxy for upscaling and modeling. The dominance of acetoclastic methanogenesis indicates that the system is presently rich in labile precursor substrates (e.g., acetate, formate, methylated substrates). However, the oldest CH$_4$ ages (~3.5 kyr BP) obtained from a thermokarst lake corresponded to the maximal age of the frozen organic (peat) layers in the valley, suggesting that permafrost thaw might have (or will soon have) proceeded through the organic substrate at this site. The local differences in surface areas, emissions rates, carbon age and sources reported in this study need to be further characterized in other regions of the Arctic in order to properly upscale and model GHG emissions and carbon-climate feedbacks across permafrost lake-rich landscapes.
Appendix A

Figure A1. Picture of the homemade funnels deployed in ponds and lakes (photo taken in July 2014 just after their removal).

Appendix B

Table B1. Temperature and total precipitation data for the 6 months preceding the sampling period in July 2013 and 2014. The climate normal (1981–2010) is also indicated (Environment Canada, 2015).

<table>
<thead>
<tr>
<th>Month</th>
<th>Temperature (°C)</th>
<th>Precipitation (mm)</th>
<th>Normal</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>2013</td>
<td>2014</td>
<td>Normal</td>
</tr>
<tr>
<td>Jan</td>
<td>−28.7</td>
<td>−30.8</td>
<td>−33.4</td>
</tr>
<tr>
<td>Feb</td>
<td>−30.5</td>
<td>−32.7</td>
<td>−33.7</td>
</tr>
<tr>
<td>Mar</td>
<td>−22.2</td>
<td>−29.8</td>
<td>−30.0</td>
</tr>
<tr>
<td>Apr</td>
<td>−19.5</td>
<td>−19.8</td>
<td>−21.8</td>
</tr>
<tr>
<td>May</td>
<td>−12.4</td>
<td>−7.2</td>
<td>−9.3</td>
</tr>
<tr>
<td>Jun</td>
<td>3.4</td>
<td>2.5</td>
<td>2.4</td>
</tr>
<tr>
<td>Total (Jan–Jun)</td>
<td>−18.3</td>
<td>−19.6</td>
<td>−21.0</td>
</tr>
</tbody>
</table>

Appendix C

Figure C1. Picture of eroding shores (slumping peat) along trough pond BYL27 (photo taken in July 2014). The sampling funnel syringe can be seen just above the water surface.

Appendix D

Figure D1. Keeling plot of lake ebullition CH₄ sampled in 2014, showing a mixing of millennium-old and highly concentrated gas with near-modern and less concentrated gas. Concentration (x axis) is expressed as 1000/partial pressure (in ppmv, parts per million volumetric), whereas radiocarbon age is expressed as the normalized radiocarbon activity ($\Delta^{14}C$, in ‰; left y axis) and in thousands of years before present (kyr BP; right y axis).
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