HYDROLOGY, CARBON DYNAMICS AND HYDROCHEMICAL PROPERTIES OF PONDS IN AN EXTENSIVE LOW GRADIENT HIGH ARCTIC WETLAND, POLAR BEAR PASS, BATHURST ISLAND, NUNAVUT, CANADA

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A DISSERTATION SUBMITTED TO THE FACULTY OF GRADUATE STUDIES IN PARTIAL FULFILLMENT OF THE REQUIREMENTS FOR THE DEGREE OF DOCTOR OF PHILOSOPHY

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ABSTRACT

Ponds form the dominant feature of Polar Bear Pass (PBP), one of the largest wetlands in the Canadian High Arctic, and in order to understand the ramifications of climatic changes on PBP we must first understand the ponds' responses to seasonal changes in climatic, physical, chemical, and carbon components. Fieldwork (2007-2010) at PBP aimed (i) to determine water budgets of ponds with various hydrologic settings, (ii) to identify the processes controlling the changes in pond carbon and geochemistry on seasonal and inter-annual bases with a special focus on the snowmelt period, and (iii) to establish the baseline hydrochemistry and hydrology of ponds within the PBP wetland complex.

Pond systems at PBP have two hydrologic settings: (i) ones which are hydrologically connected to additional sources of water from their catchments beyond seasonal inputs of snowmelt and rainfall, or (ii) ponds which fail to form a link or only have a limited connection with their surrounding catchments. Intensive seasonal monitoring of water and carbon mass balance showed that elevated loads of dissolved organic carbon (DOC) in ponds were mostly of terrestrial origin and occurred in ponds receiving meltwater from snowbeds and/or discharge from hillslope creeks. The seasonal strength in the connectivity of a pond to its catchment from snowmelt to the post-snowmelt period was critical in controlling DOC loads and concentrations. This study provided the first estimates of DOC yields at Polar Bear Pass, and reported elevated DOC

loadings from wet meadow catchments into ponds. This highlights their importance as a source of carbon to pond ecosystems during snowmelt and heavy rainfall events.

The water chemistry and environmental data showed that waters at PBP were dominated by calcium and bicarbonate ions that fell on a common dilution line, however, they had distinct proportional major ionic variability due to the location, lithology, and level of water-bedrock interaction, and these dynamics were controlled by differences in climatic conditions and hydrologic connectivity. Results relating to pond-landscape linkages and their role in solute transport to ponds showed (i) elevated surface and subsurface water contribution to ponds in hydrologically connected catchments. The primary mechanism for solute and carbon transport was overland flow during snowmelt and surface/subsurface inflow during the post-snowmelt season. There was (ii) a potential for higher solute inflow during seasons with frequent or large precipitation events. Lastly, (iii) isolated ponds were subject to evapo-concentration resulting in solute enrichment in pond waters during warm, dry periods.

An analysis of carbon dioxide (CO_2) concentrations in surface waters during snowmelt was conducted to provide the first estimates of this greenhouse gas in ponds at PBP and to further support the interpretation of hydrologic and carbon dynamics in ponds during the snowmelt and early post-snowmelt season. Surface waters at PBP were strong sources of CO_2 to the atmosphere, with CO_2 emissions dramatically increasing at the beginning of snowmelt and then declining during peak snowmelt. The required inputs of carbon to support the estimated CO_2 emissions could be explained by surface or subsurface inflows of dissolved organic carbon and dissolved inorganic carbon, and possibly from mineralization of terrestrial organic carbon in the water column and sediments of ponds.

The findings of this study will aid in the future management of the PBP wetland, and may be applied to other arctic ponds situated in High Arctic wetland environments or in any area in the circumpolar Arctic that has similar geomorphologic features and climatic setting.

ACKNOWLEDGEMENTS

First I would like to thank my advisor and mentor, Professor Kathy L. Young, for giving me this valuable opportunity to conduct research in the Arctic. I have been always fascinated by the beauty and heterogeneity of arctic wetlands where ponds are the dominant component and which are now central in this thesis. This dissertation is a direct result of Kathy's generous financial support, unlimited guidance, and friendship when I needed it most.

I extend my deep appreciation to Professor Julia Boike (Alfred Wegener Institute for Polar and Marine Research) and her entire SPARC group for the enormous platform of knowledge and research information I have gained while visiting AWI in 2010. It has been an unforgettable year as a part of SPARC team! I have greatly benefited from numerous conversations with the researchers at AWI over a strong cup of German coffee.

I am thankful to Dr. Richard Petrone (Wilfred Laurier University), Dr. Scott Lamoureux (Queens University), and Dr. Andre Robert (York University) for serving as my committee members on this thesis. This project was funded by a National Science and Engineering Research Council held by Kathy Young; National Science and Engineering Research Council GSD award, Northern Scientific Training support, the W. Garfield Weston PhD Award for Northern Research, German Academic Exchange Program (DAAD) research grant, Susan Mann PhD scholarship, and additional York University funding for graduate student research.

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Throughout this thesis I have benefited from the friendship and assistance of Elizabeth Miller. I thank Liz for her friendship, scientific dedication, and long hours of work in the field and sleepless nights in the 203A Curtis Lecture Hall lab, and the final push that helped me to finish this thesis.

I would also like to thank John Siferd for proofreading this manuscript.

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Authors: Abnizova, A., K. L. Young and M. J. Lafrenière

This study, published in *Ecohydrology*, October 2012, describes seasonal hydrology, and carbon fluxes in ponds at Polar Bear Pass, Bathurst Island, Nunavut, Canada from two intensive summer seasons (2008, 2009). Study design was formulated by Abnizova and Young while data collection and analysis was carried out primarily by Abnizova. The writing of the manuscript was by Abnizova and Young, while Lafrenière contributed to editorial changes of the manuscript. Dissolved organic carbon and fluorescence analysis was conducted by Abnizova with support from Lafrenière's biogeochemistry laboratory at Queen's U.

SNOWMELT TRENDS AND VARIABILITY IN POND HYDROLOGY AND DISSOLVED CARBON DYNAMICS IN A HIGH ARCTIC WETLAND

Authors: Abnizova, A., K. L. Young and J. Boike

This study submitted to *Ecosystems,* focuses on snowmelt hydrology and carbon flux dynamics at PBP in early spring-summer 2010. Study design was formulated by Abnizova and Young with advice from Dr. Julia Boike, Alfred Wegner Institute (AWI), Potsdam, Germany. The approach utilized follows after Abnizova et al. 2012, Journal of

Geophysical Letters, see below. Data collection, analysis and the write-up of the manuscript was carried out primarily by Abnizova. Young assisted in editing the manuscript. Water chemistry analysis was conducted at AWI with support from Boike.

SEASONAL VARIABILITY IN HYDROLOGICAL AND PHYSICO-CHEMICAL CHARACTERISTICS OF SMALL WATER BODIES ACROSS A HIGH ARCTIC WETLAND, BATHURST ISLAND, NUNAVUT, CANADA

Authors: Abnizova, A., E. Miller, K.L. Young and S. Shakil

This study submitted to *Arctic, Antarctic and Alpine,* outlines the effects of local climatic and hydrologic differences on the water chemistry of PBP pond sites and establishes baseline conditions for this previously unexplored limnological area. Study design was formulated by Abnizova and Young. Data collection and analysis was carried out by Abnizova and Miller. Principal component analysis was done by Miller, while data compilation and table work was completed by S. Shakil. Manuscript writing was done by Abnizova with editorial assistance from Young.

SMALL PONDS WITH MAJOR IMPACT: THE RELEVANCE OF PONDS AND LAKES IN PERMAFROST LANDSCAPES TO CARBON DIOXIDE EMISSIONS

Authors: Abnizova A., J. Siemens, M. Langer and J. Boike

This study, published in *Geophysical Research Letters,* June 2012, investigated the inflows and outflows of dissolved organic and inorganic carbon along with carbon dioxide emissions from lakes, ponds, and outlets on Samoylov Island in the Lena Delta of north-eastern Siberia in September 2008. Abnizova and Boike designed the study, and Abnizova, Langer and Siemens contributed to data collection and processing. All authors participated in interpreting the results and writing the manuscript. This paper provides the methodology which was followed in the paper by Abnizova and Young, submitted (see above). Hence, it is included in this thesis for reference (see Appendix A). This research investigation also places the work at PBP in relation to a circumpolar wetland site experiencing a much different hydrologic and climatological setting.

I wish to indicate that the above work also resulted during the tenure of two DAAD scholarships to the AWI. In 2008, I participated in an expedition to Siberia where much of these data were collected, and then in 2010, I ran the analysis on these water samples and spent about six months in Germany writing up these data for the above paper.

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CHAPTER 1: Introducing the research – present knowledge of arctic wetland ponds

Arctic wetlands

The geographic setting of this dissertation is High Arctic. The Arctic is situated in the northern polar region of the planet spatially defined as the area above the Arctic Circle (Figure 1). Despite its spatial extent, the Arctic is generally characterised by harsh climatic conditions described as cold winters with no daylight and short cool summers with 24 hours of daylight. The Canadian High Arctic is largely classified as a polar desert because of low precipitation (600 to less than 100 mm), cold temperatures reaching a mean of -35 °C in January and 8 °C in summer, and low net radiation receipt (Muc and Bliss 1977). Permafrost which underlies arctic landscapes is defined as bedrock, organic and earth material with temperatures remaining below 0 °C for at least two consecutive years (Winter and Woo 1990). In the continuous permafrost environments (>500 meters of permafrost depth) most hydrological processes take place in and above the active layer, which is defined as the seasonally thawed zone typically not exceeding 1 meter in depth (Woo 1983). The presence of permafrost creates an impermeable soil layer with no vertical water drainage which encourages saturated ground conditions. These conditions promote development of wetlands in the areas where sources of water are higher than losses. Arctic wetlands represent complex hydrological systems varying in type and structure and can be differentiated based on their spatial extent: patchy, meso-wetlands to extensive wetland systems.

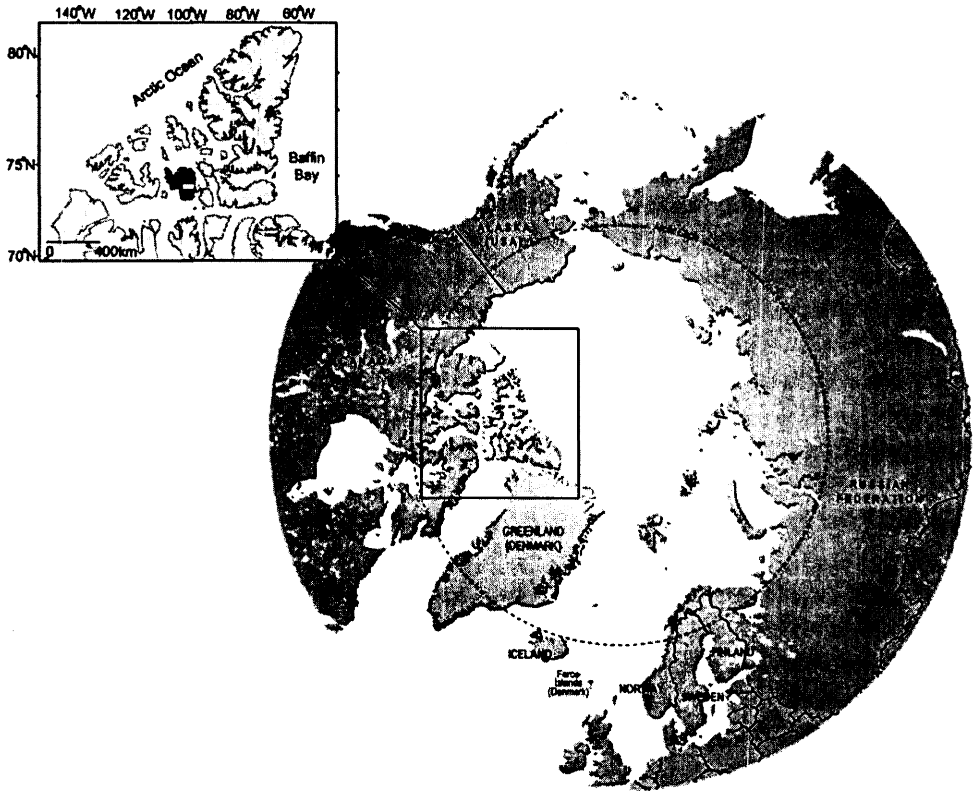

Figure 1. Arctic map showing location of the Canadian High Arctic. The inset shows location of Bathurst Island and Polar Bear Pass outlined with white rectangle. Source: Hugo Ahlenius, UNEP/GRID-Arendal).

Patchy wetlands are small catchments with distinctly wet and vegetated patches found in polar desert arid landscapes. A large number of studies have characterized the formation and development of patchy wetlands (Woo and Young 2003, Young and Woo 2003) and concluded that their formation is governed by local topography, vegetation and active layer conditions all of which constitute a set of unique hydrologic and geomorphic characteristics. According to a comprehensive summary of patchy wetland development by Woo and Young (1997), firstly, the condition of a shallow active layer reduces the amount of water required to saturate the thaw zone, allowing water tables to remain close to the ground; secondly, because this excess of water becomes readily accessible to plants it favours vegetation growth and results in peat accumulation which then favours ground ice formation due to insulating properties of peat. Water supply is essential in sustaining these wetlands and has many different sources, which may include ground water, snowmelt from late-lying snowbeds, water from the floodplains of lakes, ponds and streams, coastal flooding and water from hillslope channels to downslope fen wetlands (Figure 2; Woo and Young 2006).

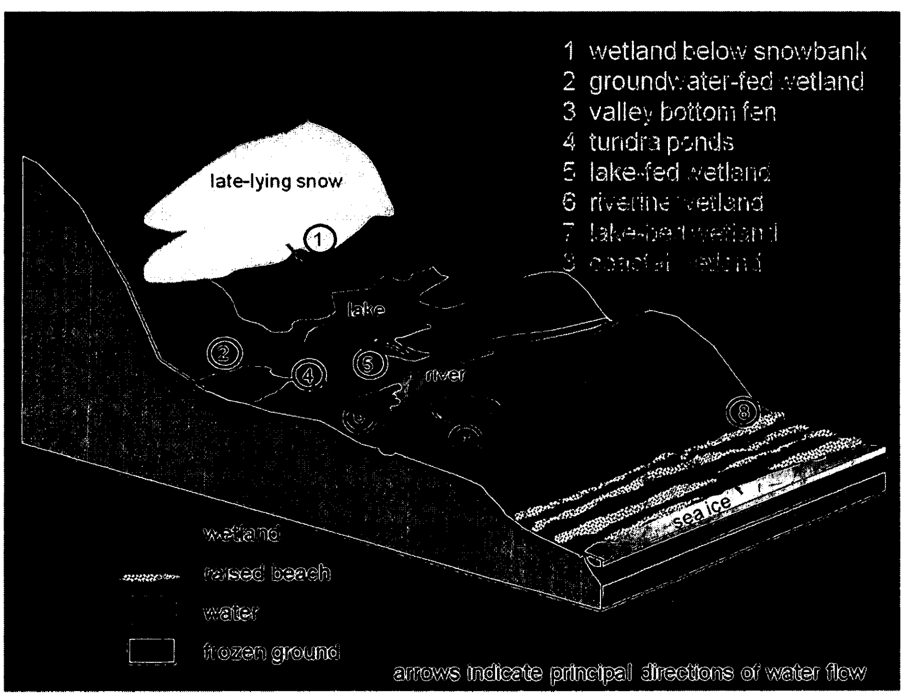

Figure 2. Conceptualization of occurrence of patchy wetlands in an Arctic environment (Source: Woo 2012, Figure 8. 33).

Meso-wetlands in the High Arctic consist of combination of wetland patches, dry and mesic grounds, and may have presence of ponds, lakes, streams, where all of these components may or may not be hydrologically connected (Woo and Young 2006). Woo and Young (2006) distinguished three types of meso-wetlands: ice-wedge polygonal ground, glacial till terrain and coastal wetlands. Reliable water supply is essential for sustainability of these large wetlands during the thawed season. A large number of studies have addressed hydrology of High Arctic meso-wetlands and shown that these wetlands may receive their inputs through a variety of hydrologic linkages to late-lying snowbanks, ground water discharge, and streamflow, flooding by lakes and the sea, and melt from ground ice if present (Woo and Young 2006). Abnizova and Young (2010) and Young and Abnizova (2011) examined a meso-wetland representing a glacial till terrain type near Creswell Bay, Somerset Island and showed a variety of hydrologic factors and connectivity mechanisms. This meso-wetland can be described by pond-wetland complexes in three various geomorphic terrain (moraine, coastal and bedrock) where pond water balances are governed by pond catchment characteristic (topography, soils, substrate material, vegetation conditions), presence of hydrologic linkage and climate variability. Various types of hydrologic connectivity were identified for wetland ponds receiving inflows from late-lying snowbeds, hillslope creeks, and undergoing losses as a result of water seepage controlled by substrate type and subsurface inflows and losses which may result from pond hydrologic connectivity to a frost crack (Figure 3a). According to Abnizova and Young (2010), the hydrologic sustainability of ponds in this wetland is defined by the degree of pond connectivity to other water sources and sets the thresholds for pond sustainability and demise (Figure 3b). Similarly, Woo and Guan (2006) examined hydrologic connectivity of tundra ponds in a meso-wetland system situated near Eastwind Lake, Ellesmere Island and identified strong surface connectivity during the snowmelt and subsurface flows during the post-snowmelt season. Losses to evaporation in this system resulted in a decline in pond water levels and surface areas with the possibility of pond recharge during major rain events.

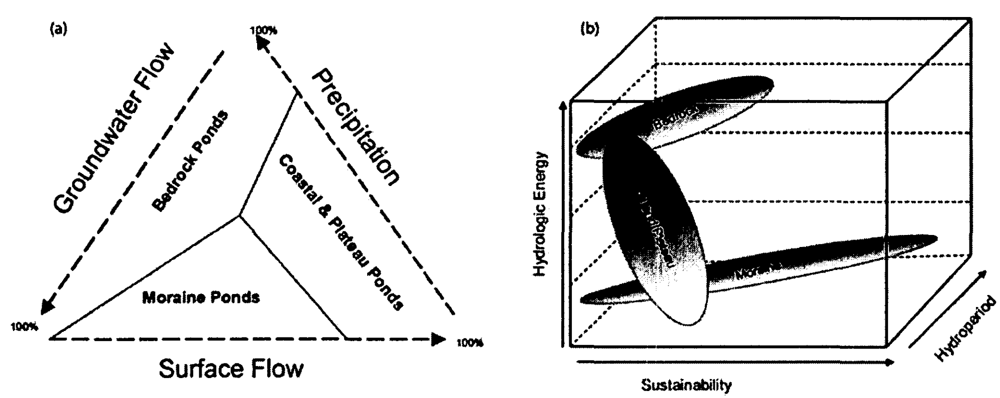

Figure 3. The relative contribution of precipitation, groundwater flow and surface inflow to a pond's catchment (a); a wetland system's functional integrity (b) (Source: Abnizova 2007, Figures 6.2 and 6.2).

While a good understanding of the hydrology of patchy wetlands in arctic environments has been attained, less is known about the hydrology of **extensive wetland** complexes encompassing areas larger than 100 km^2 (Woo 2012). Examples of extensive wetlands in the Canadian High Arctic can be found at Polar Bear Pass, Bathurst Island which represents an ice-wedge polygonal ground wetland and Truelove Lowland on Devon Island which represents an example of an extensive coastal wetland (Woo and Young 2006). Extensive wetlands like Polar Bear Pass and Truelove Lowland have been recognized as biological oases due to the richness of flora and fauna inhabiting these areas. Because of their unique presence in polar desert environments of the High Arctic, numerous ecological studies have been done on these biological oases with work including terrestrial faunal surveys (Nettleship and Smith 1975, Bliss 1977) and botanical surveys (Muc and Bliss 1977, Miller and Ireland 1978, Sheard and Geale 1983ab, Edlund et al. 1989, Edlund and Alt 1989). However, there is limited information to quantify the hydrological processes operating in these extensive wetlands. With the exception of one study by Rydén (1977) who studied a wet meadow in Truelove Lowland in the early seventies, there are no hydrologic investigations of the hydrology in extensive wetlands such as Polar Bear Pass wetland which comprises approximately 100 km^2 of lowland wetland and entire catchment area occupying 400 km^2 (Figure 4).

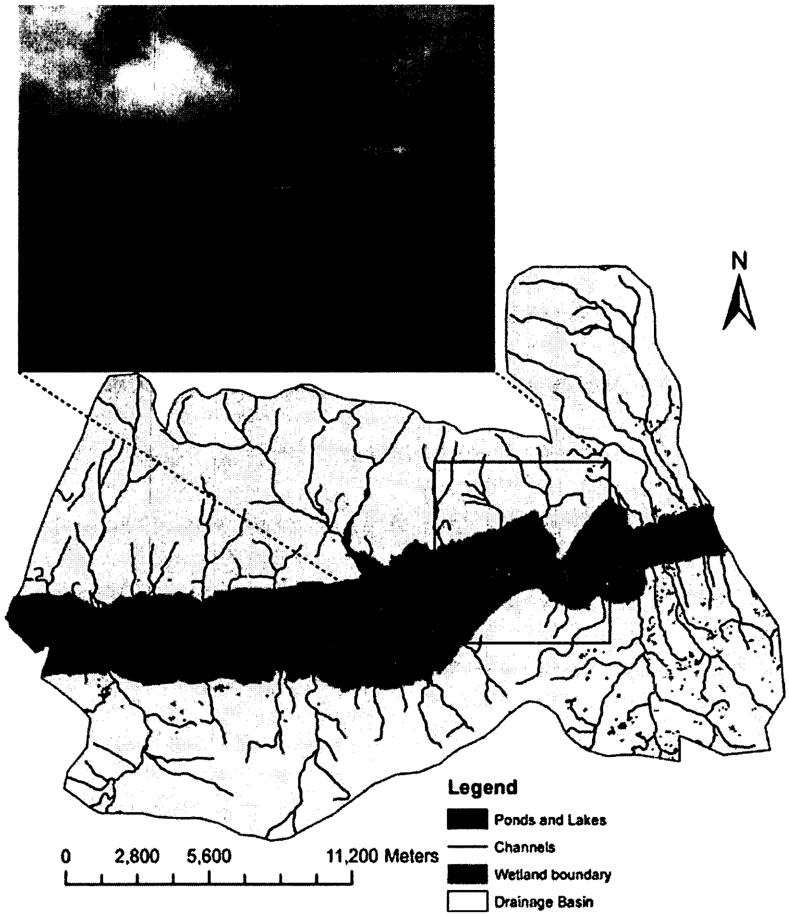

Figure 4. Polar Bear Pass catchment and lowland area with an air photo inset showing the wetland with numerous ponds. Image was taken on July 10, 2009.

Ponds are the most dominant component of extensive wetlands representing 95% of the total water bodies in some circumpolar areas (Muster et al. 2013). Due to their small areas these pond ecosystems are often underrepresented in climate models. For example, the Global Lakes and Wetland Database by Lehner and Döll (2004) only includes lakes larger than 10^5 m^2. Ponds are unique hydrogeomorphic features, forming during ground thaw and subsidence in ice-rich permafrost. They also exist in topographic depressions and intersections of polygon troughs (Hopkins 1949, Woo 2012). While the broad range of water depths and surface areas of arctic inland waters makes it difficult to establish a definitive distinction between ponds and smaller lakes, ponds typically have maximum water depth of less than 2 meters during the open water season (Woo 2012, National Wetlands Working Group 1997) and freeze to the bottom in the winter. Hydrologically, the abundance of ponds in arctic wetlands plays an important role by providing large capacities for surface water storage, especially at the time of seasonal snowmelt (Woo 2012).

At Polar Bear Pass (PBP) the total surface area occupied by water bodies comprises 20 % of the wetland, 60 % of which is contributed by ponds representing a significant portion of the area (Figure 4). PBP is considered a critical wetland area for migratory birds, caribou and muskox provide food and feeding grounds for northern wildlife and migratory birds, and it provides invaluable resources for local indigenous communities. While the ecology of PBP is relatively well studied (Young and Labine 2010), little is known about the climatology and hydrology here.

Young and Labine (2010) examined the short- and long-term summer climatic records for this wetland and identified that the climate at PBP is quite similar to Resolute Bay, a government weather station on Cornwallis Island (~ 90 km to southwest).They suggested that PBP experiences a polar desert climatic designation (Woo and Young 2003).

Recently some work has been conducted on the snowcover and snowmelt at PBP. Assini and Young (2012) and Young et al. (2013) measured end-of-winter snow storage, its distribution and monitored snowmelt patterns. Their work showed that topography and wind were the main drivers in snow distribution across the wetland areas which resulted in spatial variability in snowmelt patterns also affected by local microclimatic factors.

In 2008 Young et al. (2010) quantified the amount of meltwater coming into the wetland from hillslope streams (1 km^2 and 0.3 km^2). They showed that transport of water from upland areas was important in contributing meltwater to the low-lying wetland. These hydrologic inputs are sensitive to summer precipitation and are defined by basin characteristics such as type of soil, vegetation and ground ice and location within stream catchment. Even though recent studies document some climatic and hydrologic regimes at PBP, no detailed integration of wetland hydrology and the water balance of PBP have yet been attained.

The large numbers of ponds (~4000) here contribute significantly to the surface water storage of the PBP wetland, yet no research has been attempted to understand how these small waters function seasonally and between years. There is still a need to address the different types of connectivity of a pond to its catchment, upland areas, and to better understand the controlling factors shaping these linkages and the transfer of water and nutrients to downslope aquatic ecosystems.

Climate change and variability

Low amounts of summer precipitation in High Arctic areas and losses of water to evaporation predominantly define the summer water balance of arctic ponds making their sustainability sensitive to variable water inputs from their catchments. This is especially critical during warm, dry years (Abnizova and Young 2010). Examples of this hydrologic sensitivity (episodic shrinkage and drying) to dry episodes during the summer have been shown in ponds situated in Eastwind Lake wetland (Woo and Guan 2006) and Creswell Bay (Young and Abnizova 2011). In addition to the sensitivity of pond water storage to temperature and precipitation fluctuations their water balance also depends on the drainage area, basin geometry and bathymetry (Plug et al. 2008). Recent research has indicated that these aquatic systems play an important role as integrators of local geomorphology, limnology, hydrology and permafrost soil dynamics (Kirpotin et al. 2008), and that they vigorously respond to hydro-climatic variability (Vincent et al. 2007).

During the past decade arctic environments have been subjected to the highest summer temperatures in comparison to the last 2000 years, warming quicker than climate models have predicted (MacDonald 2010, Walsh et al. 2011). Because of this warming, scientific interest in this region has been growing, but there is still much uncertainty about how freshwater ecosystems in these polar regions are going to respond to these rapid changes in climate (Wrona et al. 2006).

Arctic ponds' heightened sensitivity to climate has signalled to researchers to suggest that they be regarded as 'sentinel' ecosystems through their ability to provide early warnings of change (Woodward et al. 2010). Woodward et al. (2010) suggest that by studying these changes now, a baseline of responses can be formulated that might be later applied to ponds and lakes at warmer latitudes, where the complexity in the environment is much higher (Woodward et al. 2009). Although scientific uncertainty remains about the freshwaters' responses to climatic changes, continuing climatic research has reduced some knowledge gaps by showing evidence of hydrologic alterations in Arctic ecosystems (Rowland et al. 2010). For instance, changes in the aerial extent and expansion rates of lakes, ponds and wetlands in permafrost environments is one of the documented effects of current high latitude climate warming (Plug et al. 2008). Climatic responses may also include changes in the lake water balance, increased groundwater outflow due to permafrost melt and thermokarst induced erosion of the shoreline that may possibly result in the drainage and disappearance of ponds and lakes (Marsh et al. 2009). A range of alterations in pond and lake size and number has recently been reported in permafrost regions (Yoshikawa and Hinzman 2003, Smith et al. 2005, Riordan et al. 2006, Hinkel et al. 2007, Smol and Douglas 2007). These studies demonstrate that arctic ponds are very vulnerable to climatic changes, yet no one has addressed the mechanisms responsible for their disappearance and drying.

Dissolved carbon

Northern wetlands store vast amounts of carbon and arctic ponds are active processors channelling terrestrial carbon to the atmosphere or storing it in their sediments and water column (Carroll et al. 2011). The net carbon budget of each aquatic system is dependent on a complex set of characteristics including water chemistry composition substrate, organic carbon quality, climate conditions and drainage pattern (Jones et al. 2011). Detailed studies of northern pond carbon budgets have proven the role of ponds in linking aquatic and terrestrial carbon cycles (Karlsson et al. 2012). Numerous studies have identified that northern lakes and ponds are emitters of carbon dioxide (CO_2) to the atmosphere which results from decomposition of terrestrial dissolved organic carbon (DOC) lost from soil organic carbon (SOC) through leaching by water (Sobek et al. 2007). Dissolved organic carbon represents one of the largest actively cycled reservoirs of organic matter (Bushaw et al. 1996) and is the most abundant hydrologically transported component of carbon cycle (Shindler et al. 1997). Dissolved inorganic carbon occurs in ionic form as HCO_3^-, CO_3^{2-}, H_2CO_3, or as dissolved or free gaseous CO_2 and generally forms from decomposition of terrestrial DOC and the weathering process in the mineral layers (Lyon et al. 2010). Together amounts of DOC and DIC in arctic ponds represent important signatures of a system's productivity responding to internal mechanisms of ecosystem processes such as photosynthesis and respiration and are sensitive to external inputs and losses of dissolved carbon.

Much limnological research (Ruhland and Smol 1998, Lim et al. 2001, Michelutti et al. 2002ab, Lim and Douglas 2003, Lim et al. 2005, Keatly 2007, Kumke et al. 2007) has documented variability in DOC in northern ponds and lakes and related the differences in concentrations to the presence of wetlands in the drainage basins. They have suggested that these sites serve as the main source of organic matter. The principal control on DOC yield from these basins is the proportion of the catchment that is constituted of wetlands along with inter-site variability (Tranvik and Jansson 2002).

Breton et al. (2009) and Laurion et al. (2010) conducted physico-chemical sampling in a series of arctic and subarctic ponds and suggested that pond dissolved organic matter (DOM) had significant control on the balance between heterotrophy and autotrophy in ponds, which originated from the thawing of organic soils and catchment vegetation.

Variability in DOC concentrations in arctic wetlands has been associated with changes in flow path due to seasonal snowmelt and active layer thaw (Clark et al. 2007). Snowmelt represents a period of meltwater runoff often resulting in extensive flooding and ponding of water and serves as an effective recharging mechanism for aquatic ecosystems and wetland soils (Woo 2012). This intensive hydrological event is known to transport large amounts of carbon and nutrients to the downstream water systems. For example, numerous studies have addressed changes in DOC concentrations in Arctic rivers during the spring freshet (Brooks et al. 1999, Rember and Trefry 2004, Schuster et al. 2004, Finlay et al. 2006, Holmes et al. 2008, and others), but a small number of studies have addressed the carbon dynamics in arctic lakes and ponds during snowmelt. Harsh climatic conditions and expensive logistics in the Arctic limit many limnological studies here to the open water season. Often times, sampling is limited to once or twice per season, a methodology referred to as a "snapshot" approach (Duff et al. 1999) or hydro-mapping. Only a few studies have focused on seasonal budgets of carbon stores in large arctic lakes (Char Lake, de March 1978; Toolik Lake, Whalen and Cornwell 1985, Crump et al. 2003; Lake 18, Tuktoyaktuk Peninsula, Ramlal et al. 1994), yet, seasonal dynamics of carbon in small ponds is limited to the results of the Tundra Biome project in Alaskan ponds (Hobbie 1980) which was conducted 40 years ago (Figure 5).

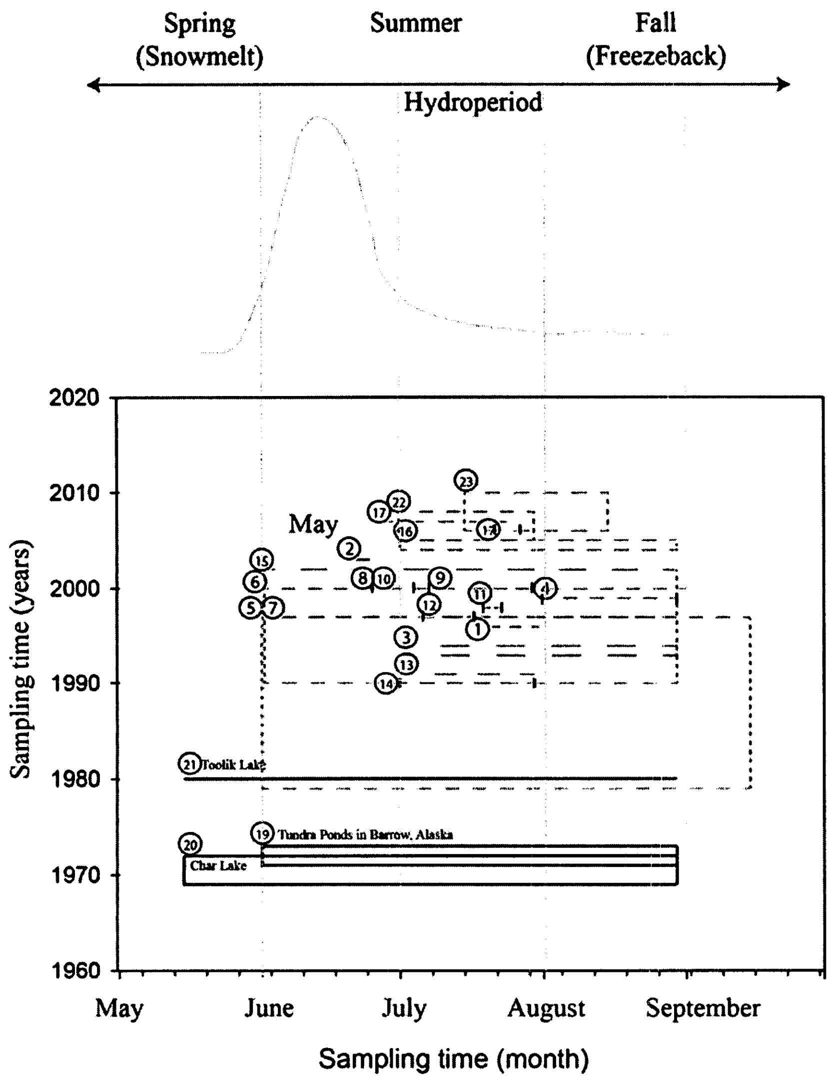

Figure 5. Graphical representation of sampling periods from selected publications reporting DOC composition of lakes and ponds in the Arctic. Dashed lines (one year) and dashed boxes (many years) indicate studies using the "snapshot" approach or short-term studies when samples have been taken two to three times during the sampling time. Solid lines (one year) and solid boxes (many years) represent complete seasonal studies with sampling conducted from snowmelt to

freezeback. Enumeration of studies is listed in Table 1, Chapter 2. Number 22 and 23 represent Lehnherr et al. (2012) and Medeiros et al. (2012), respectively.

While DOC represents a major source of carbon for freshwater heterotrophs fueling emissions of CO_2, studies monitoring CO_2 flux from wetland ponds and lakes are limited in northern and boreal regions (Zhu et al. 2012). Because of the small size of ponds and the lack of information regarding their location, extent and dynamics in the Arctic there is still a degree of uncertainty about their impact on regional and global carbon emissions (Carroll et al. 2011, Muster et al. 2013). Different emission trends are found in wetland ponds across the circumpolar arctic with studies reporting that some ponds are strong emitters during summer (Kling et al. 1991, Cole et al. 1994) and some studies report on ponds acting as sinks of carbon (Breton et al. 2009, Tank et al. 2009, Laurion et al. 2010). Many of these studies report on CO_2 emissions during summer, yet very limited research has been done on CO_2 dynamics pond arctic ponds during snowmelt and freezeback seasons. Abnizova et al. (2012) examined CO_2 emissions from arctic ponds in Siberia and showed much stronger CO_2 emissions during freezeback which were not reported for summer time effluxes. Similarly, Karrlson et al. (2013) quantified spring emissions in ponds in subarctic northern Sweden and showed that these ponds accumulate large amounts of dissolved CO_2 during winter resulting in a high flux during ice thaw.

Overall, a review of the literature suggests that there is a lack of studies on carbon dynamics during the entire snowmelt, summer open water and freeze-back period. The lack of an intensive monitoring strategy of freshwater carbon fluxes in these arctic water bodies not only during spring freshet but also for the remaining open water season limits our understanding of potential climate change effects on terrestrial DOC transport to these aquatic ecosystems. Considering the importance of aquatic ecosystems in landscape scale carbon fluxes, a better understanding of both short- and long-term carbon dynamics is warranted especially when considering their incorporation into terrestrial gas emission scenarios (Jones et al. 2011).

Hydrology and carbon dynamics in ponds at PBP

The aim of my dissertation is to provide a comprehensive understanding of hydrology of Polar Bear Pass wetland, particularly ponds. Polar Bear Pass is situated in a topographic depression bound by hillslope running south and north. This distinctive physiographic setting promotes a unique set of hydrological linkages often forming either permanent or episodic links between wetland ponds and hillslopes. These linkages not only help to transfer water but they also serve to exchange nutrients and carbon. Wetland ponds at PBP can be considered to play a dual role by (1) acting as storage units in topographic depressions by allowing for the pooling of water and (2) acting as transit hydro-systems through the transportation of water downslope when their pond storage capacity is exceeded or breeched (Figure 6). This latter hydrologic connectivity of a pond to various elements in its surrounding wetland landscapes can be multi-faceted as it supports a tight balance between hydrology and carbon dynamics in ponds at PBP which is essential for the sustainability of this biological oasis. To better understand how seasonal hydrology controls dissolved carbon movement and storage in downslope ponds, this research study will provide a complete seasonal (spring, summer and fall) picture of carbon variability in arctic ponds by focusing on (1) a quantification of carbon through its link with a pond's seasonal water budget, and (2) how the stores of carbon react to changes in a ponds' shifting water regime in response to variable hydro-climatic conditions and landscape linkages, from both the snowmelt to the freeze-back period. Previous research has highlighted local variability in inter-annual and inter-seasonal hydro-climatic conditions at PBP and research across the Arctic showed that arctic ponds are undergoing dramatic alterations in their ecohydrology. This study focuses on arctic wetland ponds and their water and carbon dynamics as a proxy for better understanding the current ecohydrology of Polar Bear Pass and its resilience to ongoing climate variability and climate change.

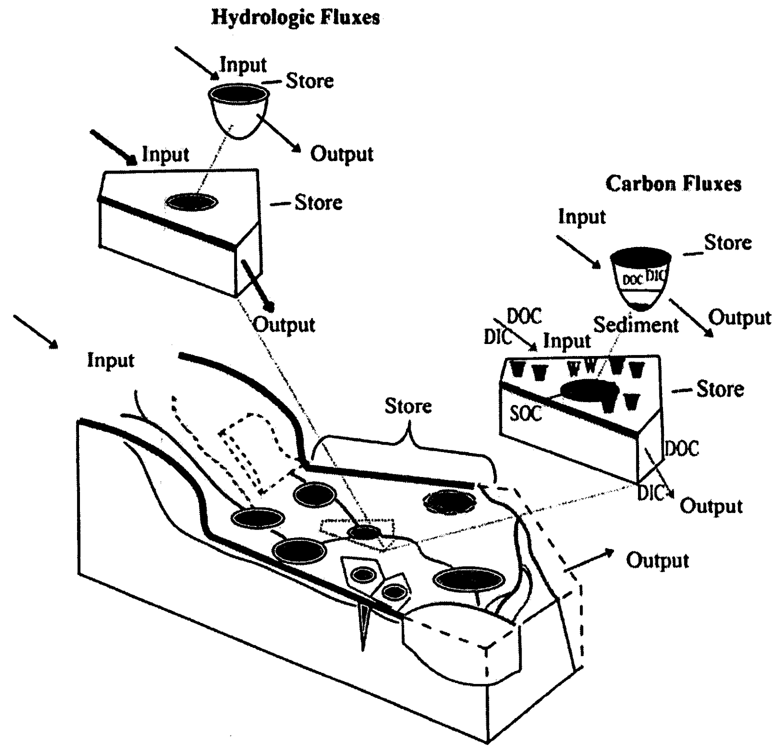

Figure 6. Schematic diagram of hydrologic and carbon fluxes in ponds situated in a hillslope-wetland system such as Polar Bear Pass. SOC, DOC and DIC stand for soil, dissolved and organic carbon.

Dissertation objectives

Considering the above observations, the objectives of this study are:

1) To identify and understand the key processes determining seasonal water budgets and carbon dynamics, and the factors that affect these processes in small ponds in a low-gradient High Arctic wetland during two field seasons (2008, 2009). This includes: a) examining hydrologic linkages in the sustainability of wetland ponds on a seasonal and inter-seasonal basis; b)

examining the seasonality of dissolved organic and inorganic carbon (DOC and DIC) concentrations and identifying the hydrologic pathways responsible for moving carbon into ponds and their adjacent wet meadows; and c) identifying the importance of wetland catchments (lowland wet meadows and adjoining hillslope areas) as sources of DOC to ponds in permafrost dominated landscapes

2) To elucidate the relationships between hydrology and carbon balance dynamics in ponds during snowmelt. This includes: a) documenting seasonal variability in the pond aquatic carbon (DOC, DIC) at Polar Bear Pass (PBP) during the snowmelt season (2010); and b) identifying the mechanisms regulating pond carbon stores during this season.

3) To provide initial estimates of CO_2 concentrations and emissions from ponds across the entire wetland at Polar Bear Pass and to understand its status in the context of these GHG contributions to the atmosphere in comparison to other circumpolar wetland sites.

4) To elucidate how hydrologic settings and variability in seasonal climatic conditions affect physico-chemical properties and inter-seasonal trends in ponds across the PBP wetland.

5) Given that this intensive fieldwork and lab-based study yielded a large physico-chemical dataset from different ponds across the PBP wetland over several years (2007-2010), this study aims to upscale the findings from intensively monitored ponds to similar ones found across PBP (94 km^2). This provides an

opportunity to explore and better understand the spatial and temporal variability (seasonal, annual) of the water quality and physico-chemical trends in all ponds (e.g. major cations, GHG and nutrients) across Polar Bear Pass.

To achieve these objectives the dissertation has been broken down into a series of papers that form the chapters with both an Introductory and Conclusions chapter bounding them. It should be noted that a literature review, description of methods pertaining to each aspect of the dissertation, and information about the study site are contained within each chapter (Two to Four). The organization of the chapters in this dissertation mimics the theme of scales, where the findings from detailed, intensive but small-scale studies are considered in Chapters Two and Three, while Chapter Four upscales this information to the regional level, across the entire PBP wetland area. In this final research chapter, a timeframe of three seasons (spring, summer, and fall) is also explored.

Specifically, Chapter Two, titled *Pond hydrology and dissolved carbon dynamics at Polar Bear Pass wetland, Bathurst Island, Nunavut, Canada*, provides information on the study area and a description of the field measurements used in 2008 and 2009. The chapter provides baseline information on seasonal hydrology, carbon fluxes in wetland ponds, and the transport of carbon both spatially and temporally between water pathways and pond sites. This chapter also provides information on how pond hydrology and connectivity to other water sources beside seasonal snowmelt and rainfall contribute to carbon stores in ponds.

Chapter Three, named *Snowmelt trends and variability in pond hydrology and dissolved carbon dynamics in a High Arctic wetland*, provides information on the study area and a description of the field measurements used in 2010. The chapter focuses on snowmelt hydrology (the timing and amount of snowmelt fluxes) and carbon flux dynamics at PBP in early spring/summer 2010. The study shows that variability in climate, topography, and surficial conditions (soils, vegetation) jointly affect the water sources, their distribution and ability to transport carbon. A mass balance approach was used to quantify the snowmelt and early post-snowmelt carbon retention efficiency in the intensively monitored pond and was used to test the hypothesis that these freshwater wetlands act as carbon sinks or sources. This chapter demonstrates that snowmelt is an important event in arctic hydrology and appears to be one of the main driving mechanisms in carbon transport, making hydrological transport the dominant processes in controlling the temporal and spatial variations of carbon.

Chapter Four, designated *Seasonal variability in hydrological and physicochemical characteristics of small water bodies across Polar Bear Pass, a High Arctic wetland, Bathurst Island, Nunavut, Canada*, provides information on the study area—both intensively studied ponds and those visited less frequently (i.e. *satellite ponds*). A detailed description of the field measurements pertaining to the hydro-chemical mapping techniques used from 2007-2010 is also found. The findings from previous studies (see Chapters 2 and 3) are utilized to address and quantify the physical and chemical limnological properties of 51 ponds located across the entire reach of Polar Bear Pass. It serves to establish baseline conditions for this previously unexplored limnological area. Second, the study explores the effects of climate variability (cool/wet years vs. warm/dry years) on the ponds' physico-chemical trends and patterns.

The main findings from chapters Two to Four are synthesized in Chapter 5-Conclusions. Aspects of future work are also considered in this final chapter.

This thesis represents a summary of research results obtained during the period from 2007 to 2010 at Polar Bear Pass, Bathurst Island, Nunavut, Canada (75.72'N 98.67'W). Additionally, I was a visiting research scholar at AWI, Potsdam in 2008 and 2010, which permitted my participation in a study examining GHG emissions from inland water bodies during freeze-back at a Siberian site in Fall 2008 (Samoylov Island, Lena Delta, Russia, N 72°22, E 126°28). Since 2009, I have had the opportunity to investigate a set of similar research techniques at the Polar Bear Pass site.

Appendix A, entitled *Small ponds with major impact: The relevance of ponds and lakes in permafrost landscapes to carbon dioxide emissions*, provides information on water and carbon fluxes in inland waters, and compares the water and carbon processes during the fall season at another arctic wetland on Samoylov Island, Siberia. We used a replicate approach in methodology by implementing the same sampling protocols and analogous data analyses and interpretation in the study reported in Chapter 3.

References

Abnizova A. 2007. Sustainability of High Arctic Ponds: Importance of Hydrologic Linkages. Unpublished master's thesis for master's degree, York University, Toronto, Canada.

Abnizova A, Young KL. 2010. Sustainability of arctic ponds in a polar desert environment. Arctic 63: 67-84.

Abnizova A, Siemens J, Langer M, Boike J. 2012. Small ponds with major impact: The relevance of ponds and lakes in permafrost landscapes to carbon dioxide emissions. Global Biogeochem Cycles 26: GB2041

Assini J, Young KL. 2012. Snow cover and snowmelt of an extensive High Arctic wetland: spatial and temporal seasonal patterns. Hydrological Sciences Journal 57(4):738-755.

Bliss LC. 1977. Truelove Lowland, Devon Island, Canada: A High Arctic ecosystem. Edmonton: University of Alberta Press. p 714.

Breton J, Vallières C, Laurion I. 2009. Limnological properties of permafrost thaw ponds in northeastern Canada. Canadian Journal of Fisheries and Aquatic Sciences 66 (10): 1635-1648.

Brooks PD, McKnight DM, Bencala KE. 1999. The relationship between soil heterotrophic activity, soil dissolved organic carbon (DOC) leachate, and catchment-scale DOC export in headwater catchments. Water Resources Research 35: 1895–1902.

Bushaw KL, Zepp RG, Tarr MA, Schulz-Janders D, Bourbonniere RA, Hodson RE, Miller WL, Bronk DA, Moran MA. 1996. Photochemical release of biologically available nitrogen from aquatic dissolved organic matter. Nature 381:404-407.

Carroll ML, Townshend JRG, DiMiceli CM, Loboda T, Sohlberg RA. 2011. Shrinking lakes of the Arctic: Spatial relationships and trajectory of change, Geophysical Research Letters 38, L20406, doi:10.1029/2011GL049427.

Clark JM, Lane SN, Chapman PJ, Adamson JK. 2007. Export of dissolved organic carbon from an upland peatland during storm events: Implications for flux estimates. Journal of Hydrology 347: 438– 447.

Crump BC, Kling GW, Bahr M, Hobbie JE. 2003. Bacterioplankton community shifts in an arctic lake correlate with seasonal changes in organic matter source. Applied and Environmental Microbiology 69: 2253–2268.

Cole JJ, Caraco NF, Kling GW, Kratz TK. 1994. Carbon dioxide supersaturation in the surface waters of lakes. Science 265: 1568-1570

de March L. 1975. Nutrient budgets for a high arctic lake (Char Lake, N.W.T.). Verhandlungen des Internationalen Verein Limnologie 19: 496-503.

Duff KE, Laing TE, Smol JP, Lean DRS. 1999. Limnological characteristics of lakes located across arctic treeline in northern Russia. Hydrobiologia 391: 205-222

Edlund SA, Alt BT, Young KL. 1989. Interaction of climate, vegetation, and soil

hydrology at Hot Weather Creek, Forsheim Peninsula, Ellesmere Island, Northwest Territories. Current Research, Part D, Geological Survey of Canada Paper 89-1D, 125-133.
Edlund SA, Alt BT. 1989. Regional congruence of vegetation and summer climate patterns in the Queen Elizabeth Islands, Northwest Territories Canada. Arctic 42: 3-23.
Finlay J, Neff J, Zimov S, Davydova A, Davydov S. 2006. Snowmelt dominance of dissolved organic carbon in high-latitude watersheds: Implications for characterization of flux of river DOC. Geophysical Research Letters 33: L10401. DOI:10.1029/2006GL025754.
Jones BM, Grosse G, Arp CD, Jones MC, Walter Anthony KM, Romanovsky VE. 2011. Modern thermokarst lake dynamics in the continuous permafrost zone, northern Seward Peninsula, Alaska, Journal of Geophysical Research 116, G00M03, doi:10.1029/2011JG001666.
Hinkel KM, Jones BM, Eisner WR, Cuomo CJ, Beck RA, Frohn R. 2007. Methods to assess natural and anthropogenic thaw lake drainage on the western Arctic coastal plain of northern Alaska, Journal of Geophysical Research 112 , F02S16, doi:10.1029/2006JF000584
Hobbie JE. 1980. Limnology of tundra ponds, Barrow, Alaska. Stroudsburg, PA: Dowden, Hutchinson and Ross.
Holmes RM, McClelland JW, Raymond PA, Frazer BB, Peterson BJ, Stieglitz M. 2008. Lability of DOC transported by Alaskan rivers to the Arctic Ocean. Geophysical Research Letters 35, L03402, doi:10.1029/2007GL032837.
Hopkins, D. M. 1949. Thaw Lakes and Thaw Sinks in the Imuruk Lake Area, Seward Peninsula, Alaska. Journal of Geology 57 (2):119-131.
Karlsson JM, Lyon SW, Destouni G. 2012. Thermokarst lake, hydrological flow and water balance indicators of permafrost change in Western Siberia. Journal of Hydrology 464–465: 459–466.
Karlsson J, Giesler R, Persson J, Lundin E. 2013. High emission of carbon dioxide and methane during ice thaw in high latitude lakes. Geophysical Research Letters 40, doi:10.1002/grl.50152.
Keatley BE, Douglas MSV, Smol JP. 2007. Limnological Characteristics of a High Arctic Oasis and Comparisons across Northern Ellesmere Island. Arctic 60 (3): 294 – 308.
Kirpotin S, Polishchuk Y, Zakharova E, Shirokova L, Pokrovsky O, Kolmakova M, Durpe B. 2008. One of possible mechanisms of thermokarst lakes drainage in West-Siberian North. International Journal of Environmental Studies 65(5): 631–635.
Kling GW, Kipphut GW, Miller MC. 1991. Arctic lakes and streams as gas conduits to the atmosphere: implications for tundra carbon budgets. Science 251: 298-301.
Kumke T, Ksenofontova M, Pestryakova L, Nazarova L, Hubberten HW. 2007. Limnological characteristics of lakes in the lowlands of Central Yakutia, Russia. Journal of Limnology 66: 40-53.

Laurion I, Vincent WF, Macintyre S, Retamal L, Dupont C, Francus P, Pienitz R. 2010. Variability in greenhouse gas emissions from permafrost thaw ponds. Limnology and Oceanography 55: 115-133.

Lehner B, Doll P. 2004. Development and validation of a global database of lakes, reservoirs and wetlands. Journal of Hydrology 296: 1-22.

Lehnherr I, St. Louis VL, Emmerton CA, Barker JD, Kirk JL. 2012. Methylmercury cycling in High Arctic wetland ponds: sources and sinks. Environmental Science and Technology 46: 10514-10522.

Lim DSS, Douglas MSV, Smol JP, Lean DRS. 2001. Physical and chemical limnological characteristics of 38 lakes and ponds on Bathurst Island, Nunavut, Canadian High Arctic. International Review of Hydrobiology 86: 1-22.

Lim DSS, Douglas MSV. 2003. Limnological characteristics of 22 lakes and ponds in the Haughton Crater region of Devon Island, Nunavut, Canadian High Arctic. Arctic, Antarctic, and Alpine Research 35:509- 519.

Lim DSS, Douglas MSV, Smol JP. 2005. Limnology of 46 lakes and ponds on Banks Island, NWT, Canadian Arctic Archipelago. Hydrobiologia 545: 11-32.

Lyon SW, Destouni G, Giesler R, Humborg C, Mörth M. 2010. Investigating the relationship between subsurface hydrology and dissolved carbon fluxes for a sub-arctic catchment, Hydrology and Earth System Sciences 14, 941-950, doi:10.5194/hess-14-941-2010.

MacDonald GM. 2010. Global warming and the Arctic: a mew world beyond the reach of the Grinnellian niche? The Journal of Experimental Biology 213: 855-861.

Marsh P, Russel M, Pohl S, Haywood H, Onclin C. 2009. Changes in thaw lake drainage in the Western Canadian Arctic from 1950 to 2000. Hydrological Processes 23, 145–158.

Medeiros AS, Biastoch RG, Luszczek CE, Wang XA, Muir DGC, Quinlan R. 2012. Patterns in the limnology of lakes and ponds across multiple local and regional environmental gradients in the eastern Canadian Arctic. Inland Waters. Journal of the International Society of Limnology 21: 59-76.

Michelutti N, Douglas MSV, Muir DCG, Wang XW, Smol JP. 2002a. Limnological characteristics of 38 lakes and ponds on Axel Heiberg Island, High Arctic Canada. International Review of Hydrobiology 87: 385-399.

Michelutti N, Douglas MSV, Lean DRS, Smol JP. 2002b. Physical and chemical limnology of 34 ultraoligotrophic lakes and ponds near Wyniatt Bay, Victoria Island, Arctic Canada. Hydrobiologia 482: 1-13.

Miller NG, Ireland RR. 1978. A floristic account of the bryophytes of Bathurst Island Arctic Canada. Occasional Papers of the Farlow Herbarium of Cryptogamic Botany Harvard University, 1-38.

Muc M, Bliss LC. 1977. Plant communities of Truelove Lowland. In: Bliss, L.C., Ed. Truelove Lowland, Devon Island, Canada: A High Arctic ecosystem. Edmonton: University of Alberta Press. p 143 – 154.

Muster S, Heim B, Abnizova A, Boike J. Water Body Distributions Across Scales: A

Remote Sensing Based Comparison of Three Arctic Tundra Wetlands. Remote Sensing 5(4):1498-1523.
National Wetlands Working Group. 1997. The Canadian Wetland Classification System, 2nd Edition. In: Warner BG, Rubec CDA, Eds. Wetlands Research Centre, University of Waterloo, Waterloo, ON, Canada. p 68.
Nettleship DN, Smith PA. 1975. Ecological Sites in Northern Canada. Canadian Committee for the International Biological Program Conservation Terrestrial Panel 9, Canadian Wildlife Service, Ottawa, Canada.
Plug LJ, Walls C, Scott BM. 2008. Tundra lake changes from 1978 to 2001 on the Tuktoyaktuk Peninsula, western Canadian Arctic. Geophysical Research Letters 35, L03502, doi:10.1029/2007GL032303.
Ramlal P, Hesslein SRH, Hecky RE, Fee EJ, Rudd JWM, Guildford SJ. 1994. The organic carbon budget of a shallow arctic tundra lake on the Tuktoyaktuk Peninsula, N.W.T., Canada: arctic lake carbon budget. Biogeochemistry 24: 145-172.
Rember RD, Trefry JH. 2004. Increased concentrations of dissolved trace metals and organic carbon during snowmelt in rivers of the Alaskan arctic. Geochimica et Cosmochimica Acta 68(3): 477–489.
Riordan B, Verbyla D, McGuire AD. 2006. Shrinking ponds in subarctic Alaska based on 1950-2002 remotely sensed images. Journal of Geophysical Research 111: G04002 doi:10.1029/2005JG000150.
Rowland JC, Jones CE, Altmann G, Bryan R, Crosby BT, Geernaert GL, Hinzman LD, Kane DL, Lawrence DM, Mancino A, Marsh P, McNamara JP, Romanovsky VE, Toniolo H, Travis BJ, Trochim E, Wilson CJ. 2010. Arctic landscapes in transition: responses to thawing permafrost. EOS Journal 91: 229–30.
Ruhland K, Smol JP. 1998. Limnological Characteristics of 70 Lakes Spanning Arctic Treeline from Coronation Gulf to Great Slave Lake in the Central Northwest Territories, Canada. International Review of Hydrobiology 83(3):183-203.
Rydén BE. 1977. Hydrology of Truelove Lowland. In Truelove Lowland, Devon Island, Canada. In: Bliss LC, Ed. A High Arctic Ecosystem. Edmonton: University of Alberta Press. p 107-136.
Schuster PF, Reddy MM, Aiken GR, Shanley JB. 2004. What effect does permafrost have on dissolved organic carbon transport to streams during snowmelt. In: Wanty RB, Seal IIRR, Eds. Water Rock Interaction Volume 2, Proceedings of the Eleventh International Symposium on Water. New York: Saratoga Springs. p 1385-1389.
Sheard JW, Geale DW. 1983ab. Vegetation studies at Polar Bear Pass, Bathurst Island, N.W.T. 1) Classification of Plant Communities. Canadian Journal of Botany 61(6):1618-1636. N.W.T. 2) Vegetation-environment relationship. Canadian Journal of Botany 61(6):1637-1646
Schindler DW, Curtis PJ, Bayley SE, Parker BR, Beaty KG, Stainton MP. 1997. Climate-induced changes in the dissolved organic carbon budgets of boreal lakes. Biogeochemistry 36: 9–28.

Smith LC, Sheng Y, MacDonald GM, Hinzman LD. 2005. Disappearing Arctic lakes. Science 308: 1429–1429.
Smol JP, Douglas MSV. 2007. Crossing the final ecological threshold in High Arctic ponds. Proceedings of the National Academy of Sciences USA, 104: 12395–12397.
Sobek S, Tranvik LJ, Prairie YT, Kortelainen P, Cole JJ. 2007. Patterns and regulation of dissolved organic carbon: An analysis of 7,500 widely distributed lakes. Limnology and Oceanography 52(3), 2007, 1208–1219
Tank SE, Lesack LFW, McQueen DJ. 2009. Elevated pH regulates bacterial carbon cycling in lakes with high photosynthetic activity. Ecology 90: 1910–1922
Tranvik LJ, Jansson M. 2002. Terrestrial export of organic carbon. Nature 21.
Vincent WF, Rautio M, Pienitz R. 2007. Climate control of underwater UV exposure in polar and alpine aquatic ecosystems. In: Orbaek JB, Kallenborn R, Tombre I, Hegseth E, Falk-Petersen A, Hoel AH, Eds. Arctic Alpine Ecosystems and People in a Changing Environment. Berlin/Heidelberg: Springer, p 227-249.
Walsh JE, Overland JE, Groisman PY, Rudolf B. 2011. Ongoing climate change in the Arctic. In: Callaghan TV, Johansson M, Prowse TD, Eds. Arctic cryosphere—Changes and impacts. Ambio 40(S1). doi:10.1007/s13280-011-0211-z.
Whalen SC, Cornwell JC. 1985. Nitrogen, phosphorus and organic carbon cycling in an arctic lake. Canadian Journal of Fisheries and Aquatic Sciences 42, 797.
Winter TC, Woo MK.1990. Hydrology of lakes and wetlands. In: Wolman MG, Riggs HC, Eds. Surface water hydrology. Boulder: Geological Society of America. p 159–187.
Woodward G, Christensen JB, Olafsson JS, Gislason GM, Hannesdottir ER, Friberg N. 2009. Sentinel systems on the razor's edge: effects of warming on Arctic stream ecosystems. Global Change Biology. doi:10.1111/j.1365-2486.2009.02052.x
Woodward G, Perkins D, Brown LE. 2010. Climate change and freshwater ecosystems: impacts across multiple levels of organization. Philosophical Transactions of the Royal Society B: Biological Sciences: 365, doi: 10.1098/rstb.2010.0055
Woo MK. 1983. Hydrology of a drainage basin in the Canadian High Arctic. Annals of the Association of American Geographers 73:577–596.
Woo MK, Young KL. 1997. Hydrology of a small drainage basin with polar oasis environment, Forsheim Peninsula, Ellesmere Canada. Permafrost and Periglacial Processes 8 (3): 257-277.
Woo MK, Young KL. 2003. Hydrogeomorphology of patchy wetlands in the High Arctic, polar desert environment. Wetlands 23:291–309
Woo MK, Young KL. 2006. High arctic wetlands: Their occurrence, hydrological characteristics and sustainability. Journal of Hydrology 320: 432- 450.
Woo MK, Guan XJ. 2006. Hydrological connectivity and seasonal storage change of tundra ponds in a polar oasis environment, Canadian High Arctic. Permafrost and Periglacial Processes 17: 309-323.
Woo MK. 2012. Permafrost Hydrology. Springer, Heidelberg, 563.

Wrona FJ, Prowse TD, Reist JD, Hobbie JE, Lévesque LMJ, Vincent WF. 2006. Climate Impacts on Arctic Freshwater Ecosystems and Fisheries: Background, Rationale and Approach of the Arctic Climate Impact Assessment (ACIA). AMBIO: A Journal of the Human Environment 35 (7): 326-329.

Young KL, Labine C. 2010. Summer hydroclimatology of an extensive low-gradient wetland: Polar Bear Pass, Bathurst Island, Nunavut, Canada. Hydrology Research 41:492–502.

Young KL, Woo MK. 2003. Thermo-hydrological responses to an exceptionally warm, dry summer in a High Arctic environment. Hydrology Research 34:51–70.

Young KL, Assini J, Abnizova A, De Miranda N. 2010. Hydrology of hillslope-wetland streams, Polar Bear Pass, Nunavut, Canada. Hydrological Processes 24: 3345-3358

Young KL, Abnizova A. 2011. Hydrologic thresholds of ponds in a polar desert wetland environment, Somerset Island, Nunavut, Canada. Wetlands 31: 535-549.

Young KL, Assini J, Abnizova A, Miller EA. 2013. Snowcover and melt characteristics of upland/lowland terrain: Polar Bear Pass, Bathurst Island, Nunavut, Canada. Hydrology Research 44 (1): 2-20

Yoshikawa K, Hinzman LD. 2003. Shrinking thermokarst ponds and groundwater dynamics in discontinuous permafrost near Council, Alaska. Permafrost and Periglacial Processes 14: 151–160.

Zhu D, Chen H, Zhu Q, Wu Y, Wu, N. 2012. High carbon dioxide evasion from an alpine peatland lake: The central role of terrestrial dissolved organic carbon input. Water, Air, and Soil Pollution 223:2563–2569.

CHAPTER 2: Pond hydrology and dissolved carbon dynamics at Polar Bear Pass wetland, Bathurst Island, Nunavut, Canada

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Abstract

A large number of wetlands, lakes, and ponds exist in northern Canada, Alaska, and Siberia, and the hydrologic and ecological processes in these water bodies are responding to a changing climate. A large wetland, Polar Bear Pass (PBP), situated in the middle of Bathurst Island is considered to be one of the most important ecological sites in the region. Numerous ponds exist at PBP that are connected to their surrounding watersheds by streams and groundwater inflow, receiving varying amounts of water and nutrients. In 2008 and 2009 the representative hydrology of typical ponds at PBP along with their quantity of dissolved organic and inorganic carbon (DOC, DIC) was evaluated. Pond DOC and DIC loads and composition differ depending on the presence or absence of one or more hydrologic linkages that a pond has with its catchment. Elevated DOC loads were mostly of terrestrial origin and occurred in ponds receiving meltwater from snowbeds and discharge from hillslope creeks. The seasonal shift in connectivity of a pond to its catchment was critical in controlling DOC loads and concentrations. The frequency and duration of summer precipitation had a strong control on pond hydrologic connectivity and elevated the contribution of terrestrial DOC from wetland to ponds, especially ones that were hydrologically connected. The estimated DOC yields from wet meadow catchments highlight their importance as a source of carbon to pond ecosystems downstream. These wetland areas and ponds are potentially significant pools of carbon and are sensitive to future climate changes in permafrost-dominated environments.

Key Words: arctic ponds; arctic wetlands; carbon sources; dissolved organic carbon; hydrologic linkages; wetlands; precipitation

Introduction

Numerous small ponds and lakes form the dominant land-type of northern wetland environments. Today researchers are documenting a decline in pond and lake abundance, especially in Alaska, Northern Siberia, and Canada's Eastern Arctic (Yoshikawa and Hinzman, 2003; Smith *et al.*, 2005; Smol and Douglas, 2007). These aquatic systems are important reservoirs of carbon (Tranvik *et al.,* 2009) yet they have been largely ignored despite their major role in global cycles (Downing, 2010). Such water bodies become net sources of CO_2 to the atmosphere due to the mineralization of organic carbon imported from terrestrial systems and the degassing of resultant inorganic carbon (Cole *et al.*, 1994). Dissolved inorganic carbon (DIC) is an important part of the aquatic carbonate system and can be found not only as dissolved CO_2, but as HCO_3^- and CO_3^{2-}. Cole *et al.* (1994) calculated partial pressures of CO_2 in 1,837 global lakes and found 84 % of them supersaturated in CO_2. Supersaturation of CO_2 in lakes and ponds was also documented in other environments spanning from temperate to Arctic regions (Engstrom, 1987; Kling *et al.*, 1991; Hope *et al.*, 1996; Pienitz *et al.,* 1997a; Pienitz *et al.,* 1997b; Breton *et al.*, 2009).

Hydrologic processes drive the transport of organic carbon in wetlands through a variety of pathways such as snowmelt runoff, streamflow, meltwater draining from late-lying snowbeds, and groundwater discharge. Dissolved organic carbon (DOC) is an important constituent of aquatic organic carbon and is largely derived from the breakdown and leaching of organic matter in terrestrial areas of a catchment (allochtonous). It is then supplemented by autochthonous DOC generated by aquatic biota (Bradley *et al.*, 2007). Identification of DOC sources in hydrologic pathways is important for understanding of carbon fluxes in wetland environments. Allochtonous and autochtonous DOC sources can be qualitatively differentiated using DOC spectrofluorometric properties based on the general source of the fulvic acid fraction of DOC (McKnight *et al.*, 2001; Lafrenière and Sharp, 2004). Typically, snowmelt periods in wetlands are associated with high water tables contributing to dominantly allochtonous sources, while autochtonous sources are associated with late summer algae die-off (Peterson *et al.*, 2002). In northern environments seasonal hydrologic fluxes of DOC are dominated by spring snowmelt, which can carry high concentrations of DOC (Finlay *et al.*, 2006; Lewis *et al.*, 2011). Despite its importance, this season is not always well represented in sampling efforts to identify fluxes and composition of dissolved organic matter in ponds and lakes (Lyons and Finlay, 2008).

Numerous studies have focused on carbon dynamics and hydrology in alpine and subalpine environments (Baron *et al.*, 1991; Boyer *et al.*, 1997; Hood *et al.*, 2003; Lafrenière and Sharp, 2004; Hood *et al.*, 2005; Lafrenière and Sharp, 2005), the subarctic (Carey, 2003), and in Arctic Alaska (Kling *et al.*, 1991; Michaelson *et al.*, 1998), but presently there are only a few studies describing the flow of carbon in the Canadian Arctic (Guo *et al.*, 2007, Raymond *et al.*, 2007; Lewis *et al.*, 2011). For instance, to understand the transport pathways of soil organic carbon from permafrost to arctic rivers, Guo *et al.* (2007) measured dissolved and particulate organic carbon in the Mackenzie River basin and showed that radiocarbon composition of DOC in the river was largely derived from modern terrestrial biomass. They also attributed changes to altered plant ecology in response to regional climate change. Similarly, Raymond *et al.* (2007) reported that during snowmelt the major Arctic rivers export 60 % of their annual DOC flux, half of which is 1.5 years old.

Despite growing interest in carbon cycling in high latitudes, little research has focused on the seasonal carbon dynamics of High Arctic ponds and lakes (Table 1, Figure 1). This is of particular significance in the Arctic given the attention to greenhouse gas (GHG) emissions from arctic wetland ponds and lakes, which are fueled by the respiration or, photolytic degradation of DOC (Kling *et al.,* 1991). While mineralization of DOC may be an important source of DIC in aquatic ecosystems, a lack of work on processes governing DIC fluxes and trends remains (Hope *et al.,* 1994). Bradley *et al.* (2007) in a study of dissolved carbon trends in a headwater wetland showed that concentrations of DOC and DIC differed in groundwater and surface water samples. Their findings indicated that catchment carbon flux would be underestimated when the DIC and DOC fluxes are studied independently.

The high spatial and seasonal variability of GHG fluxes reported in numerous studies (*e.g.,* Huttunen *et al.,* 2002; Breton *et al.,* 2009; Walter Anthony *et al.,* 2010) challenges current estimates of the freshwater carbon balance in the Arctic, and points to an increasing need for improved quantification of carbon fluxes and sources to better understand the global carbon cycle and climate system (Tranvik *et al.,* 2009). While numerous studies have reported that northern lakes contain terrestrial inputs of carbon, complete estimates of carbon storage and its seasonal variability in these northern water bodies are currently limited to Alaska (*i.e.,* Prentki *et al.*, 1980; Whalen and Cornwell, 1985). We are currently aware of only four studies documenting the seasonal dynamics of carbon in Canadian High Arctic lakes (Kalff and Welch, 1974; Schindler *et al.,* 1974; de March, 1975; Ramlal *et al.*, 1994). The remoteness of the sites, extreme weather conditions, combined with high logistical costs of accessing field sites limit the length of a sampling season and often result in low data accuracy and precision. Scientists are often forced into a 'snap-shot' approach, reporting concentrations of DOC from only one sample per season (Duff *et al.,* 1999; Table 1). This present study aimed to address this research gap by examining seasonal pond water balance and the dynamics of pond DOC from snowmelt until freeze-back in relation to wetland water sources.

Arctic wetlands are important freshwater environments which provide feeding grounds for migratory birds, caribou, and muskox (Nettleship and Smith, 1975). Polar Bear Pass (PBP) wetland is a designated wildlife sanctuary, however no information exists on the intra-or inter-seasonal patterns of water flow in this wetland or on the quantity of carbon moving into the wetland. Questions surrounding the degree of hydrologic connectivity or isolation of ponds and how this influences organic carbon loads have not yet been addressed here. The objectives of this research are: (1) to examine the importance of hydrologic linkages in pond sustainability on a seasonal and inter-seasonal basis; (2) to examine the seasonality of DOC concentrations and identify the hydrologic pathways responsible for moving carbon into ponds and their adjacent wet meadows; and (3) to identify the importance of wetland catchments (lowland wet meadows and hillslope areas) as a source of DOC to ponds in a permafrost dominated catchment.

Study area

Fieldwork was carried out in one of the largest wetlands in the Queen Elizabeth Islands: Polar Bear Pass, Bathurst Island, Nunavut. The site (75.72′N 98.67′W) is located in a broad valley and the lowland covers an area of approximately 87 km^2 (Muster, 2011, per. comm.). The terrain is occupied by two lakes and over 4200 tundra ponds ranging from 8 m^2 to 0.8 km^2, separated by the raised rims of ice-wedge polygons or by limited extents of sedge meadows (Woo and Young, 2006). Water bodies occupy 22 % of the total wetland area with ponds (< 8000 m^2) contributing 5 %. Wet meadow terrain composes 74 % of the Pass. The Pass is bordered by poorly vegetated hillslopes which rise to about 200 m a.s.l. They are incised by numerous valleys draining the upland areas and seasonally recharge the wetland (Young *et al.,* 2010). Although PBP wetland is considered to be a biological oasis (Edlund and Alt, 1989) it experiences a polar desert climatic regime, much like Resolute Bay on Cornwallis Island located approximately 146 km to the south (Young and Labine, 2010). The Resolute Bay weather station reports a 1971-2000 mean annual air temperature of -16.4 °C, but its mean May, June, July, and August temperatures are -10.9, -0.1, 4.3 and 1.5 °C, respectively.

Previous wetland studies near Creswell Bay, Somerset Island, Nunavut, Canada suggested that a pond's water budget is often dependent on the type and duration of its hydrologic linkage to its surrounding catchment, which becomes more important during warm, dry years (Abnizova and Young, 2010; Young and Abnizova, 2011). These studies further indicated that the pond substrate, especially if it is coarse (*i.e.*, rocky terrain), serves to enhance thaw and water seepage, a process often detrimental to smaller ponds which can dry up quickly or transition into wet meadow zones. Since ponds are the dominant water body at Polar Bear Pass, five ponds representative of different terrain types (fine-grained and ice-rich versus coarse-textured and ice-poor) were selected for this study (Figure 2b, Table 2).

Two ponds with well-defined hydrologic linkages were located in fine grained and ice-rich terrain (Figure 2b). One of them is supplied by meltwater from a late-lying snowbed and is referred to as the *Snowbed pond*. The other receives runoff from an ephemeral hillslope creek (*Creek pond*). Two other selected ponds had no well-defined connections to their landscape apart from seasonal snowmelt inputs and rainfall. One was an isolated large pond in fine grained and ice-rich terrain (*Isolated Large*, or *IL pond)*, and the other was a smaller pond in similar terrain (*Isolated Small*, or *IS pond*, Figure 2b, Table 2). To understand the importance of a limited hydrologic connectivity in pond hydrology, one pond was selected in an isolated setting with an intermittent link to an ephemeral shallow pond located upslope (*Temporarily Connected*, or *TC pond*) (Figure 2b, Table 2).

Methodology

Field methods

To measure seasonal changes in pond water storage (ΔS), a 1- dimensional water balance framework after Woo *et al.* (1981) was utilized. Here,

$$\Delta S = P_{(Sn+R)} - E \pm Q_{(Qsur+Qsub)} \qquad (1)$$

where ΔS is the storage term used to describe the change in depth of water in mm in the ponds, Sn and R are precipitation inputs (P) from snow and rain, E is evaporation output, and Q is lateral flow, where Q_{sur} and Q_{sub} are surface and subsurface inflow or outflow to a pond, respectively. Snow surveys, following the approach by Woo (1997) were conducted at each study pond and adjoining catchment prior to snowmelt. Transects were laid out so as to encompass both the pond and its catchment, and depth measurements at 1 to 2 m intervals depending on the size of the pond and its catchment were made. Snow depth measurements were obtained by inserting a metric ruler (± 5 mm) and/or a long snow rod into the snowbank until the pond ice surface was reached. Depending on length of the transect (ranging from 30 to 137 m), 1 to 3 snow density cores were obtained (beginning, middle and transect end) with a Meteorological Service of Canada (MSC) snow corer. Daily snowmelt (M) was estimated with direct measurements (*cf.* Heron and Woo, 1978). Snow ablation poles with a 2.5 m marked line (every 10 cm) strung between them was established in the snowcover at different sites. Surface snowmelt (M) was then determined as

$$M\,(\text{mm}) = (z_{t-1}-z_t)(\rho_s/\rho_w) \tag{2}$$

where z_{t-1} - z_t is the difference based on the average of 10 measurements, between the vertical height value from the surface to the marked line from the previous day and the present day using a metric measuring tape (± 1 mm). Daily surface snow density (ρ_s) was based on the average weight of 5 surface measurements (each with a volume of 200 cm^3).

After the snow had ablated, the depth of ground thaw was measured twice a week near each well early in the season, and then weekly once thaw slowed. Thaw depth was determined as the depth to which a metal rod could be inserted into the ground until impeded by the frozen subsurface. Evaporation (E) at the study ponds and the wet meadow was calculated using the Priestley-Taylor (PT) approach (Priestley and Taylor, 1972). This method ranks high in accuracy in relation to other approaches (Rosenberry *et al.*, 2004). The α term was set to 1.26, an estimate generally considered appropriate for saturated surfaces (*e.g.*, Steward and Rouse, 1977; Bello and Smith, 1990).

An automatic weather station (AWS) was set up in the wet meadow to allow for estimates of snowmelt and evaporation. Air temperature (°C) and relative humidity (%) were measured with a CSI HC2-S3-L temperature and relative humidity probe (± 2 °C). Incoming and outgoing shortwave radiation (W m^{-2}) was measured with Eppley pyranometers (± 1 %). Net radiation (Q^*, W m^{-2}) was measured with an NR lite (±1 %), ground heat flux (W m^{-2}) was measured with a ground heat flux plate (± 3 %), wind speed and direction (m s^{-1}, degrees) were measured with a Davis anemometer (± 5 %), and precipitation (*P*, mm) was measured with a HOBO tipping bucket raingauge (± 0.25 mm). Net radiation was also measured over open water in the study pond. The water heat flux to the pond was calculated by multiplying the daily change in pond water depth with pond water temperature and the volumetric heat capacity of water (Woo and Guan, 2006). Water table in ponds was continuously measured with an Ecotone water level recorder (Remote Data Systems Inc., ± 2.54 mm) or a HOBO pressure transducer (HOBO U20 Water Level Data Logger, ± 0.5 cm) on an hourly basis and confirmed by manual measurements twice weekly at the center wells. The water temperature was routinely monitored by HOBO temperature probes (± 0.2 °C).

In 2007, 15 perforated and screened 5 cm diameter water table wells (3 transects, Figure 2b) were installed down to the permafrost table in the wet meadow catchment, downslope of the late-lying snowbed and upslope of the *Snowbed pond*. Here, manual water table and frost table were obtained twice weekly from these wells to assess surface and groundwater inflow into the wetland. Groundwater flow was determined using Darcy's law (Young *et al.*, 1997). The hydraulic conductivity of saturated soils was determined using bailing tests once or twice each season (Luthin, 1966). Near surface soil moisture in the unsaturated zone was monitored with a Theta probe soil moisture sensor-Ml2X (Delta-T Devices, ± 1 %) near each water well (average of three samples; *cf.* Young and Abnizova, 2011).

Stream discharge (L s^{-1}) flowing from the plateau area into the low-lying wetland area (*LSC*, Figure 2b) was estimated in both 2008 and 2009 using the mid-section velocity approach (Young *et al.*, 2010). A gauging station was established at the entrance to the wetland zone and an Ecotone water level recorder and HOBO pressure transducer provided continuous stage measurements (cm) every 30 minutes. Direct streamflow measurements were conducted up to four times daily during the main snowmelt period and then less frequently (once or twice daily) in the post-snowmelt period using an Ott C2 current meter with a calibrated accuracy of < 3 %. Reliable stage-discharge curves were established in 2008: $Q = 0.129H^{1.654}$, $R^2 = 0.76$, $n = 11$, for the snow covered channel, and $Q = 0.447H^{1.756}$, $R^2 = 0.94$, $n = 10$ for the snow free channel. Similarly, in 2009 $Q = 3.34 H^{0.562}$, $R^2 = 0.80$, $n = 16$, for the snow covered channel, and $Q = 0.401H^{1.750}$, $R^2 = 0.86$, $n = 15$, for the snow free channel.

All pond sites and their adjacent catchments were surveyed in 2007 and 2008 using a Total Survey Station (Leica ± 0.5 mm). ArcGIS 9.3 and Golden Software Surfer 7 allowed pond bathymetry and topographic maps of the catchments to be produced and provided estimates of pond coverage, surface area, volume, catchment size, and water linkages (Table 2).

Dissolved organic carbon (DOC) and dissolved inorganic carbon (DIC)

Snowmelt water and pond water samples were collected from each pond once a week from June to early September (2008, 2009). To determine concentrations of DOC entering ponds as a result of their hydrologic connectivity, water samples were also taken from the creek gauging station, at the edge of the late-lying snowbed and at the entry point of catchment waters to the connected ponds (Figure 2b). All samples were filtered in the field using pre-combusted GF/F filters (0.7 μm pore diameter) and a polyethylene syringe and stored in 40 ml glass acid-rinsed vials. Prior to the time of collection the glass sample bottles were rinsed three times with filtered sample water. Each sample was then acidified to pH 2 by adding 2M HCl and stored in the dark at 4 °C prior to analysis. Alkalinity was measured using a Hach Alkalinity Test Kit. Field measurements of pH (± 0.2), conductivity (± 0.001 mS cm^{-1}), dissolved oxygen (± 0.01 mg L^{-1}), and water temperature (± 0.15 °C) were made on a weekly basis with a YSI 600R Multiparameter Sonde.

DOC analysis was conducted at the Facility for Biogeochemical Research on Environmental Change and the Cryosphere (FaBRECC) at Queen's University. DOC levels were measured as non-purgeable organic carbon (NPOC) by high temperature combustion (680 °C) with a Shimadzu TOC-VCPH analyzer equipped with a high sensitivity platinum catalyst. The method detection limit for the analysis of water samples was 0.47 ppm based on the analysis of field blanks (n = 10). The variability in replicate DOC samples was ± 0.16 ppm (n = 27, 2008 and n = 29, 2009).

Fluorescence spectroscopy was used to quantitatively differentiate DOC resulting from terrestrial material such as soil and plant organic matter from DOC of microbial origin such as algal and bacterial products (McKnight *et al.*, 2001). Since fluorescent properties of DOC vary with the aromatic content of fulvic acids, and the aromaticity of fulvic acids from plants differ from those of microbial cells, the DOC resulting from terrestrial sources (allochthonous) has a different fluorescent signature than DOC produced by algae or bacteria within pond ecosystems (autochthonous) (*e.g.* Hood *et al.*, 2003; Lafrenière and Sharp, 2004). The fluorescence of the DOC was measured using a Shimadzu RF-1501 scanning spectrofluorometer with a xenon lamp. The fluorescence index (*FI* - the ratio of emission intensity at 450 nm to 500 nm for an excitation of 370 nm) was obtained by scanning the samples in an optically clear quartz cuvette at an excitation wavelength of 370 nm, for emission wavelengths between 370 and 700 nm at 1 nm increments (McKnight *et al.*, 2001). According to McKnight *et al.* (2001), the *FI* has values of *ca.* 1.9 – 2.0 for predominantly microbial DOC and values of *ca.* 1.4 for DOC with predominantly terrestrial sources. The Suwanee River fulvic acid (allochtonous, SRFA), Suwanee River Natural Organic Matter (allochtonous, SRNOM) and the Pony Lake fulvic acid (autochtonous, PLFA) were used as reference materials. The measured *FI* values of the reference materials (PLFA, $FI = 2.00$, SRFA, $FI = 1.54$, and SRNOM, $FI = 1.86$) indicated that signatures for microbial DOC would be expected to be above 2.00 and terrestrial DOC signatures should have index values around 1.86.

Dissolved inorganic carbon (DIC) was estimated using standard carbonate equilibrium relationships and field pH, alkalinity and temperature measurements (Stumm and Morgan, 1996).

At a given time, DOC and DIC stocks (S_c) (g C m^{-2}) in pond water were estimated as:

$$S_{c_i} = C_i V_i / A_i \tag{3}$$

where C_i is concentration at the time of sampling (i) (mg L^{-1}), V_i is pond volume (m^3) and

A_i is pond area (m^2). Shrinkage and expansion of pond surface area were monitored on a weekly basis from surveyed points around pond edges.

Synchrony between DOC concentrations in ponds was measured by calculating the Pearson correlation coefficient (R) amongst all study pond pairs (Pace and Cole, 2002). To establish the strength of agreement between two compared concentrations, Lin's concordance correlation coefficient (R_c) was calculated (Lin, 1989).

Results

Summer climatology and energy balance (2008-2009)

Climatic conditions in 2008 and 2009 at PBP are shown in Figure 3 and summarized in Table 3. In 2008 air temperature at PBP fluctuated from -2 to 0.5 °C prior to snowmelt (June 3), whereas in 2009 there was a continuous rise in air temperature with temperatures reaching above 0 °C by June 7 (Figure 3). Average daily air temperatures in 2008 were influenced by the net radiation regime and levels increased during the early 2008 post-snowmelt season coinciding with a period of sunny and clear conditions. The seasonal average in 2008 was 3.8 °C (June-August). In 2009 average daily air temperatures declined during the early post-snowmelt season as a result of low radiation inputs but the seasonal average was similar to 2008 (3.9 °C). Likewise, relative humidity remained > 80 % in both years reflecting cool, cloudy conditions (Table 3). In summers, monthly average air temperature and total precipitation values were not significantly different ($p<0.05$) from long-term records at Resolute Bay (1971-2000).

Total precipitation was similar in both years (94.5 mm vs. 94.9 mm), but the frequency at which the rainfall occurred varied (Table 3). Figure 3 indicates that precipitation events occurred less in 2008 but they were of longer duration and appeared mainly later in the season. In contrast, 2009 events were more frequent (generally < 3 mm per event) and of a shorter duration.

Over the course of the study season (June-August) total evaporation from the wet meadow was similar in both years (108 mm *vs.* 110 mm) and was the largest energy sink, consuming 43-47 % of Q^*. Total evaporation from the pond was slightly higher in 2008 than in 2009 (169 mm *vs.* 149 mm). Pond evaporation expended 58-61 % of Q^* (Figure 3).

Pond hydrology and water balance: hydrologically connected ponds

Despite variable climatic conditions, the water table of the *Snowbed pond* remained steady and relatively unchanged in both years (Figure 4). Maximum ground thaw depth was also similar in 2008 and 2009 (689 mm and 731 mm). End of winter snow accumulation of 29 mm in 2008 and 37 mm in 2009 at PBP was much lower than at other wetland sites (Table 2, Young and Abnizova, 2011) and summer rains did not increase pond water levels. Owing to cool and cloudy conditions in both years, daily evaporation from the *Snowbed pond* (open water) averaged only 2.0 mm d^{-1} in 2008, and 1.7 mm d^{-1} in 2009. In 2007 (a warm, dry summer), pond evaporation in July was 4.2 mm d^{-1} (0.5 to 5.9 mm d^{-1}) (Young and Labine, 2010). Like the *Snowbed pond*, the seasonal linkage of the *Creek pond* to a hillslope creek (Landing Strip Creek-*LSC*, Figure 4) served to maintain elevated water levels after spring snowmelt in both 2008 and 2009. Once snowmelt waters drained, this study pond maintained an ephemeral-type linkage with *LSC*. Pond water levels dropped to their seasonal minimum in 2008 after the creek dried up (July 24 - August 8, Figure 4), but they remained elevated in 2009 due to the higher frequency of rain events which ensured that the hillslope creek kept flowing.

In both seasons measured water storage ΔS_{meas} showed an increasing trend in both hydrologically connected study ponds during snowmelt (Figure 5, Table 4). The estimated water storage ($\Delta S_{est} = ((Sn+R) - E)$ is also shown in Figure 5 for comparison. Since ΔS_{est} shows changes in water budget of a pond as a result of vertical fluxes, its comparison to changes in ΔS_{meas} allows for identification of presence or absence of hydrologic lateral inflows (Q). Our results show that changes in ΔS_{est} and ΔS_{meas} varied during the post-snowmelt period in 2008 reaching a difference of 87 mm and 49 mm in the *Snowbed* and *Creek ponds*, respectively. After July 13, 2008, the changes in ΔS_{est} and ΔS_{meas} were similar in the linked ponds, the difference between the two increasing slowly to 113 mm in the *Creek pond* and then dropping in both ponds until August 31st. Pond water storage in 2009 was characterized by frequent rain events. The changes between ΔS_{est} and ΔS_{meas} were markedly different in the linked ponds as a result of net lateral inflows during this wet season. The differences between ΔS_{meas} and ΔS_{est} for *Snowbed* and *Creek ponds* (median = 133 and 137 mm, respectively) remained until July 25, 2009, well after the end of snowmelt. This finding demonstrates the critical role played by the timing of rain events coupled with the degree of connectivity that a pond has with its catchment. Only during the month of August, did ΔS_{meas} decrease due to evaporation losses. Here the difference between ΔS_{est} and ΔS_{meas} on August 31^{st} was 101 mm and 62 mm in the *Snowbed* and *Creek ponds*, respectively. Overall, the total water storage gain was similar in 2008 and 2009 for the *Snowbed pond* but two times smaller for the *Creek pond* (Table 4).

Hydrologically isolated ponds

The *Isolated Large (IL) pond* located about 2 km from base camp (see Figure 2b) is found in a wetland area which traps more snow than other areas (Figure 2, Table 2). While the water table here is relatively shallow (Table 2) and rapidly falls to 70 mm after snowmelt, the rate of ground thaw is slow in comparison to the other ponds and only reaches a maximum depth of 529 mm in 2008 and 659 mm in 2009. Likewise, due to its large surface area (Table 2), the pond requires substantial amounts of rain and lateral inflow to sustain stable water tables. The rainy conditions in both summers were sufficient to maintain a stable hydrologic regime here.

Of the five ponds studied, the *Isolated Small* (*IS*) *pond* can be considered the most vulnerable to desiccation. The small surface area and low snow accumulation led to rapid snowmelt and early drop in water table. Evaporation rates were higher in *IS* than in other ponds (1.9 mm d^{-1} in 2008, and 1.9 mm d^{-1} in 2009) owing to higher water temperatures, a small surface area, and shallow depth (Table 2). The pond became completely desiccated in 2008 with the water table falling below the ground (-109 mm), but late season rains managed to revive it back to freshet levels in 2008 (Figure 4a). The ability of late season rains to revive desiccated ponds has been reported elsewhere (Woo and Guan, 2006; Young and Abnizova, 2011). Frequent rainfall in 2009 maintained stable and relatively high water tables preventing this pond from desiccation.

Unlike the ponds with no hydrologic linkages, the *Temporarily Connected* (*TC*) *pond* provides an example of how ponds respond when subjected to intermittent or limited hydrologic linkages with its surrounding catchment (Figures 4 and 5). Maximum ground thaw here reached > 825 mm in both years, much deeper than the other sites owing to its coarse substrate. However, evaporation losses were comparable to the *Snowbed* and *Creek ponds* (1.9 mm d^{-1} in 2008, and 1.7 mm d^{-1} in 2009). Water tables were lower in 2008 and fell to 83 mm but rose again after a series of late season rainfalls. In 2009, water tables fell to 210 mm after snowmelt but later remained stable in response to steady rains, dropping only to 195 mm on August 17th.

The water storage in the hydrologically isolated ponds was characterized by the frequency and duration of rain events in 2008 and 2009. Measured change in water storage (ΔS_{meas}) showed a decreasing trend after the snowmelt season in 2008 and 2009 and was only markedly interrupted by precipitation events in both years (Figure 5). Changes in ΔS_{est} and ΔS_{meas} were comparable until August 13, 2009 when a series of end-of-season rain events produced marked differences. An end-of-season 50 mm difference indicated a net lateral gain of water to the *IS pond.* Overall, in both years the isolated ponds had negative end-of-season storage values (Table 4).

In comparison to the *IS* and *IL ponds*, ΔS_{meas} showed an increase towards the end of the season in the *TC pond* in both 2008 and 2009, and a positive end-of-season water storage. One of the main factors for this stability in 2008 and 2009 can be linked to this pond's geomorphic setting. Early in the season, the pond receives both snowmelt runoff and rainfall from an upslope ephemeral pond and its surrounding saturated catchment (Figure 2b). Later on in the summer, deep ground thaw together with a dry substrate ensures that most rainfall (< 3.2 mm d^{-1}) infiltrates into storage. These rains are for the most part ineffective in raising water levels. It is only when significant rainfalls occur (*e.g.*, 22.3 mm on July 27, 2009) did pond water levels rise to snowmelt freshet levels.

Seasonal carbon dynamics and stocks

Dissolved carbon concentrations

DOC concentrations in the five study ponds (*Snowbed, Creek, TC, IL and IS ponds)* in 2008 ranged from 1.48 to 41.55 mg L^{-1}, with an average of 13.24 ± 8.18 mg L^{-1} (n = 63) and a median of 10.00 mg L^{-1}. In 2009, concentrations were lower and ranged from 3.41 to 23.01 mg L^{-1}, with an average of 12.22 ± 3.84 mg L^{-1} (n = 67) and a median of 11.45 mg L^{-1}. Pond DOC concentrations varied over time, with maximum levels generally occurring during snowmelt in 2008 (24.22 ± 7.68 mg L^{-1}) and during the early post-snowmelt season in June 2009 (16.67 ± 2.81 mg L^{-1}) and July 2009 (17.17 ± 3.68 mg L^{-1}) (see Figure 6).

The *IS pond* exhibited the highest individual and mean seasonal DOC concentrations in 2008 (41.55 mg L^{-1}) and 2009 (23.01 mg L^{-1}). Lowest concentrations were found in snow collected at the *TC pond* in 2008 (1.48 mg L^{-1}) and at the *IS pond* in 2009 (3.41 mg L^{-1}). In 2008 DOC concentrations were much higher in the smaller ponds (*TC, IS*), compared to the larger ponds (*Snowbed, IL*) which shared similar DOC concentrations. DOC concentrations for all ponds, with the exception of the *IS pond*, peaked around June 12, 2008, coinciding with the onset of snowmelt. The DOC concentrations in the *IS pond* peaked again much later, at 40 mg L^{-1} in the season, when water levels dropped dramatically in this pond (Figure 4a).

In 2009 the concentration and seasonal trends were similar across all five study ponds. During snowmelt in 2009 DOC concentrations only reached a maximum of 12.97 mg L^{-1} on June 16th, but peak concentrations occurred after snowmelt on June 24th (21.22 mg L^{-1}) and then again around July 8th (23.01 mg L^{-1}). The first peak on June 24th coincided with highly saturated terrain conditions and shallow ground thaw; the second peak resulted from a series of rain events (12.8 mm) lasting from July 1 to July 8.

Opposite to DOC, snowmelt dissolved inorganic carbon concentrations (DIC) were typically at a minimum and then increased through to the end of summer. In 2008 DIC in all ponds were very similar, except on July 2nd and July 9th, when there was a significant increase in DIC concentrations in the *Snowbed Pond* (42.61 mg L^{-1}) and *Creek Pond* (28.60 mg L^{-1}) as a result of their connectivity to the catchment during the early post-snowmelt season. At this time saturated conditions prevailed throughout the landscape. In general, DIC increased gradually over the 2008 season in all study ponds, reaching their mean end-of-season concentration of 15.8 ± 2.1 mg L^{-1}. In 2009 patterns of DIC are comparable to 2008 and by the end of the season all ponds exhibited a two-fold seasonal increase in relation to initial post-snowmelt concentrations (Figure 6).

Dissolved carbon stocks

Due to the differences in pond sizes and seasonal effects of evapo-concentration (Benoy *et al.,* 2007), the area specific mass of carbon (or stocks of C) (g C m^{-2} for DOC and DIC) were estimated. These values were based on measured fluctuations in DOC and DIC concentrations, pond area, and volume (Figure 7). The stocks provide a more appropriate means of comparing the carbon budgets between ponds, given that concentrations are affected by variability in pond size and evaporative losses.

There were clear differences in DOC and DIC stocks for ponds linked to their catchment during the snowmelt period, in comparison to isolated ponds in 2008 (Figure 7). Ponds connected to the catchment had moderately higher DOC and DIC stocks than the isolated ponds. This pattern was replicated again during 2009 when DOC and DIC pools in well-connected ponds were higher than in isolated ponds. In 2009, continued connectivity of ponds to their catchments as a result of inflow during the early post-snowmelt period and after a stretch of continuous rainfall, resulted in the highest stocks of DOC and DIC (Figure 7). In addition, strong intra-seasonal variability in ponds linked to reliable hydrologic linkages was observed during the early post-snowmelt season in 2008. For example, during an interval of warm, dry conditions (June 30^{th} to July 14^{th}) there was a dramatic increase in DIC storage in the *Snowbed* and *Creek ponds*, whereas DIC remained stable in the other ponds. This pattern illustrates DIC inputs originate from surface runoff.

Sources of organic carbon

Fluorescence index values of dissolved organic matter (DOM) in 2008 revealed that DOC in all study ponds originated primarily from terrestrial organic matter (Figure 8). Note that the only instance where the DOM fluorescence index values approached the PLFA (autochtonous) values was in late August 2008 following a warm dry July, and rainy August. Major peaks in carbon stocks in ponds correspond to the snowmelt period in 2008, and although DOC stocks increased during the snowmelt period in 2009 the peaks in carbon stocks that year were associated instead with rainfall inputs and runoff during the early post-snowmelt season. Seasonal fluctuations in dissolved carbon stocks suggest that ponds connected to various hydrologic linkages tend to have elevated carbon amounts.

In general, DOC concentrations of inflow entering the creek-fed pond exhibited similar trends in seasonal concentrations than that measured in the pond or its outflow (Figure 9a), suggesting a strong connection between the carbon in the pond and its upslope catchment. The concentrations in the inflow and in the pond remained the same until mid July, and then declined possibly due to DOC being consumed by heterotrophic respiration, or a shift of flowpath as a result of active layer thaw. Likewise, in 2008 higher DOC concentrations were recorded at the point of inflow to the *Snowbed pond* versus both in-situ and outflow values (Figure 9b). A positive hydraulic gradient (0.01 m m^{-1}) and groundwater inflow (here hydraulic conductivity = 0.45 m d^{-1}) from an upslope wet meadow zone supports the importance of this pond's linkage to its upslope catchment (Figure 10). Moreover, DOC concentrations recorded at the source of the hillslope creek waters in 2008 (Figure 2b) were much lower on average (6.3 ± 4.2 mg L^{-1}) than DOC concentrations in waters entering the pond (19.8 ± 10.0 mg L^{-1}). This difference corresponds to the length of the hydrologic path (*ca.* 500 m) that the hillslope creek waters took from the source (upper plateau zone) to their sink (the *Creek pond*). Like *LSC*, concentrations of DOC for the late-lying snowbank waters in 2008 (4.3 ± 0.3 mg L^{-1}) were more dilute than samples taken at the *Snowbed pond* entrance (9.6 ± 3.8 mg L^{-1}). Here, a distance of 200 m separated the sampling points (see Figure 2b). In summary, these results suggest that hillslope creeks here pick up DOC along their pathway and deliver it to the adjacent low-lying ponds where it is either consumed or diluted by DOC limited waters.

Discussion

This High Arctic study highlights a number of conditions under which pond connectivity to a hydrologic source(s) represents a significant supply of dissolved carbon to its ecosystem.

Importance of hydrologic connectivity

Our hydrologic data suggests marked differences in pond water budgets for ponds connected to late-lying snowbeds or creeks versus ponds receiving water from intermittent single sources or ones receiving only seasonal meteoric inputs.

The data reveal that water levels in some wetland ponds can be maintained by steady hydrologic linkages even after snowmelt has ended. Similar findings were reported by Abnizova and Young (2010) and Young and Abnizova (2011) for a series of wet meadow ponds near Creswell Bay, Somerset, Nunavut. They found that hydrologically linked ponds had relatively steady water tables throughout the post-snowmelt season, since reliable water inputs negated losses from both evaporation and ground seepage. However, seasonal hydrologic gain for some connected ponds can still result in low water table values (*e.g. Creek pond)*, and this can be attributed to a pond's maximum storage capacity being filled prior to the start of the following spring season and hydrologic outflows. For instance, summer rains at the end of the 2008 season ensured that the pond water storage was at its full capacity prior to freeze-back. This small pond with limited storage at the start of the 2009 thaw season, together with persistent saturated conditions upslope, channelled inflow from creek waters to move through it quickly to the downslope wetland area, much like how meltwater might drain across a still frozen ice-covered lake. Thus, in 2009 the *Creek pond* did not play a water reservoir role in this landscape, but instead acted as a transmitting feature (*cf.* Spence and Woo, 2003).

While hydrologic linkages deliver water to connected ponds (*e.g.*, *Snowbed* and *Creek ponds*), these water sources are not always readily available for isolated ponds, especially during warm, dry summers. It is only during wetter years that they are linked to upslope catchments (*e.g., TC pond*). Elevated soil moisture levels in pond catchments typically increase hydrologic connectivity. Overall, high soil moisture conditions in pond catchments prevailed during both seasons ranging from 68 ± 23.9 to 92.0 ± 23.9 %. The highest values were recorded during the wet 2009 season. Occasionally, initial snowpack conditions are critical and can serve to offset pond water losses and desiccation. Higher accumulation of snow in the *IL pond* helped to delay snowmelt, ground thaw, and shorten the evaporation period. Abnizova and Young (2010) also observed this pattern for deeply incised rocky ponds on Somerset Island.

Pond water balance results illustrate that in the absence of precipitation during mid-season (2008) both small and large isolated ponds (*IS* and *IL*) experience large evaporative losses, with complete desiccation of smaller ponds. However, late rains can rejuvenate small ponds and allow positive water storages to occur, a phenomenon that has been found by others in arctic areas (e.g., Kane and Yang, 2004; Woo and Guan, 2006; Young and Abnizova, 2008). While the large isolated pond in this study showed a negative summer water balance, it did not completely dry out due to heavy and frequent summer precipitation. Similar findings on the importance of late summer or early fall rains for rejuvenating wetland ponds prior to freeze-back have been reported by Woo and Guan (2006) and Young and Abnizova (2011). We surmise that summer rains also play an important role in how a pond responds to snowmelt inputs the following season (*i.e.*, ponds serve as either a reservoir of water or a transmitting feature). Taken together, our study and those of others demonstrate the importance in the timing and duration of precipitation events in sustaining small and shallow ponds with limited hydrologic linkages in High Arctic wetland landscapes. These variations in hydrologic processes have considerable implications when considered in conjunction with dissolved carbon results.

Dissolved carbon dynamics

A review of the literature suggests that previous studies have focused on dissolved carbon concentrations in arctic ponds based on a single sample or a few samples taken over a season (Table 1). Our results suggest this sampling strategy may lead to significant underestimates in seasonal dissolved carbon stocks in ponds. Water samples taken over an entire study season (June to September) reveal that arctic wetland ponds receive spring snowmelt high in DOC concentrations, while meltwaters bring diluted DIC concentrations. Elevated DOC and DIC concentrations can also be triggered by frequent rain and connectivity to an external water source (stream, creek, late-lying snowbed) (Figure 6).

Pearson bivariate correlation analysis of pond DOC concentrations (Andersson *et al.*, 1991; Kling *et al.,* 2000; Pace and Cole, 2002) showed increased synchrony across all study ponds during 2009, a year with a higher frequency of rain events. This finding supports the importance of catchment connectivity and summer precipitation regimes in regulating pond hydrologic balances and dissolved carbon fluctuations. Strong agreement between DOC concentrations between study ponds (concordance coefficient > 0.7) was also found for the connected ponds (Table 5). The highest synchrony in concentrations between the connected ponds was during the drier season of 2008 (Table 5), and this can be linked to snowmelt transferring DOC into ponds. These study findings are supported by Baron *et al.* (1991) and Miller and McKnight (2010) who also indicate that DOC concentrations in alpine lakes increase during snowmelt. Minimal ground thaw during the snowmelt season and shortly thereafter ensures some interaction of surface waters with the uppermost organic-rich soil materials. Others have reported that a large amount of

DOC originates in the organic-rich upper soil horizons which can be delivered into streams and lakes during spring snowmelt or during high precipitation events (McKnight *et al.*, 1993; Hinton *et al.*, 1997; Schindler *et al.*, 1997; Lafrenière and Lamoureux, 2008).

The elevated pond DOC stocks in 2009 further highlight the importance of post-snowmelt season rain events as a mechanism for delivering DOC to wetland ponds. Immediately following snowmelt in 2009, shallow ground thaw combined with saturated soil conditions and frequent rainfall promoted flushing of DOC through the upper organic rich soil horizon. A similar pattern has been reported recently by Lewis *et al.* (2011) in a paired High Arctic watershed study at Cape Bounty, Melville Island, however our study further indicates that secondary mid-season peaks in DOC stocks emerge only after continuous rain events and only for connected ponds. Our estimates of water flux show that while DOC concentrations at the source (creek and edge of late-lying snowbed) were low, these increased in a relatively short distance, while the sink (pond) received high DOC concentrations as a result. While changes in seasonal DOC stocks in the ponds are mainly controlled by pond hydrologic connectivity and changes in seasonal climatic conditions (*e.g.,* frequent rain events in 2009), delivery of DIC to ponds is possibly controlled by the contribution of carbon-rich subsurface inflows. DIC fluxes into ponds increase towards the end of the season as a result of prolonged interaction of subsurface waters with thawed carbonate-rich mineral layers generally found in this calcareous bedrock environment.

Overall, when compared with other DOC studies, the DOC stocks at PBP are moderate. Although maximum active layer development in this wetland may approach 1 m, the organic layer is relatively shallow, averaging only 3 to 5 cm in most locations, potentially reaching a maximum of 10 to 15 cm in lush wet meadow zones. Therefore it is possible that this thinner organic horizon limits the delivery of DOC in runoff and the amount of DOC in ponds compared to ones located in subarctic terrain where peat layers would be much thicker (Carey, 2003).

Our *FI* data confirm that these ponds receive organic carbon of predominantly terrestrial origin. Soluble *OM* likely leached from soils and vegetation and was carried into ponds by seasonal snowmelt, meltwater from late-lying snowbeds, and water draining from nearby hillslope creek catchments. Miller and McKnight (2010) report similar *FI* values for a small, clear alpine lake which is strongly influenced by snowmelt processes early in the summer. However, in their study, as the summer progressed, the dominant DOM pool changed from one dominated by allochthonous sources to one dominated by autochthonous sources, expressed in higher *FI* values. McKnight *et al.* (2001) found high *FI* index for autochthonous DOC sources in permanently ice-covered and ice-free in summer time Antarctic lakes, where organic carbon originated from internal biotic process. The only occasion when our *FI* data suggested a strong autochthonous source was in the smallest pond (*Creek pond*) at the end of the summer.

DOC yields

Our study reveals that DOC contributions from adjacent wet meadow catchments to wetland ponds are linked to the hydrology of the site. We estimate, based on our limited surface inflow measurements, that a seasonal DOC yield from the wet meadow to the ponds was 2.9 g C m^{-2} y^{-1}, whereas inflow from a hillslope source ranged from 0.0 to 3.1 g C m^{-2} y^{-1} for the *LSC* and the late-lying snowbed respectively (Table 6). In 2009, the DOC inputs to the ponds were highest during the snowmelt period followed by the early post-snowmelt season with frequent rain events. The lowest values occurred during the post-snowmelt season. This sequence roughly corresponds to a seasonal pattern where snowmelt contributes the largest seasonal water fluxes followed by periods of intensive rain events which generate extensive overland flows enhanced by shallow ground thaw conditions, and, finally, a drier post-snowmelt season where reduced hydrologic flows occur. It is important to note that our estimate of the seasonal DOC yield is likely an underestimated value since it is based on only two samples of DOC during the snowmelt period, only one sample taken after a series of prolonged rain events, and an average value of samples taken once per week during the post-snowmelt period (Table 6). Nevertheless, our estimated seasonal DOC yields are similar to those cited by Moore (1987) and Koprivnjak and Moore (1992) working in subarctic wetlands. Understandably, they are lower than estimates from wetlands located in temperate environments (Elder *et al.*, 2000; Naiman, 1982; Fraser *et al.*, 2003).

Conclusions

The aim of this study was to understand the seasonal hydrology, degree of connectivity of wetland ponds to their surrounding catchment at Polar Bear Pass, and the control these processes have on dissolved carbon dynamics. It was hypothesized that catchment characteristics, especially type and seasonality of hydrologic inputs are important to in-situ pond carbon dynamics because they help to determine both the quality and quantity of carbon input in them. Water depth, surface area, and volume were also expected to have important implications for carbon delivery and storage in ponds.

This study revealed that pond connectivity to various water source(s) like late-lying snowbed, hillslope creek, or another pond upslope can be critical for a pond's sustainability, especially during dry spells with little rain. Our results show that while the snowmelt season continues to contribute the highest concentrations and stocks of carbon to ponds, the frequency and duration of rain events can have a strong control on pond hydrologic connectivity and increasing rainfall elevates DOC contributions from terrestrial sources to the ponds. While spring snowmelt may set initial levels of DOC and DIC in all ponds, additional connectivity to a hydrologic source affects subsequent variations in DOC and DIC. Rainfall events sometimes augment DOC inflows and help to determine the variations in concentrations amongst a range of tundra ponds with relatively high DOC yields entering the ponds from adjacent wet meadow catchments. Finally, inter-annual variability in pond water and carbon budgets points to an increasing need for extended field season investigations that capture both the snowmelt period and late season rainfall events. This seasonal overview provides for an enhanced understanding of the sources of carbon to pond ecosystems. This is of particular significance in the Arctic, given the growing attention to GHG gas emissions from arctic wetland ponds and lakes.

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References

Abnizova A, Young KL. 2010. Sustainability of High Arctic ponds in a polar desert environment. *Arctic* **63**: 64-68.

Andersson T, Nilsson A, Jansson M. 1991. Colored substances in Swedish lakes and rivers-Temporal variation and regulating factors. *Lecture Notes in Earth Sciences* **33**: 243-253.

Antoniades D, Douglas MSV, Smol JP. 2003. Comparative physical and chemical limnology of two Canadian high arctic regions: Alert (Ellesmere Island, NU) and Mould Bay (Prince Patrick Island, NWT). *Archives of Hydrobiology* **158**: 485-516.

Barley EM, Walker IR, Kurek J, Cwynar LC, Mathewes RW, Gajewski K, Finney BP. 2006. A northwest North American training set: distribution of freshwater midges in relation to air temperature and lake depth. *Journal of Paleolimnology* **36**: 295-314.

Baron J, McKnight D, Denning AS. 1991. Sources of dissolved and particulate organic material in Loch Valley Watershed, Rocky Mountain National Park, Colorado, USA. *Biogeochemistry* **15**: 89-110.

Bello R, Smith JD. 1990. The effect of weather variability on the energy balance of a lake in the Hudson Bay Lowlands, Canada. *Arctic and Alpine Research* **22**(1): 98-107.

Benoy GA, Cash K, McCauley E, Wrona F. 2007. Carbon dynamics in lakes of the boreal forest under a changing climate. *Environmental Reviews* **15**(1):175-189.

Bradley C, Baker A, Cumberland S, Boomer I, Morrissey, I. 2007. Dynamics of water movement and trends in dissolved carbon in a headwater wetland in a permeable catchment. *Wetlands* **27** (4): 1066-1080.

Boyer EW, Hornberger GM, Bencala KE, McKnight DM. 1997. Response characteristics of DOC flushing in an alpine catchment. *Hydrological Processes* **11:** 1635-1647.

Breton J, Vallières C, Laurion I. 2009. Limnological properties of permafrost thaw ponds in northeastern Canada. *Canadian Journal of Fisheries and Aquatic Sciences* **66** (10): 1635-1648.

Brutemark A, Rengefors K, Anderson NJ. 2006. An experimental investigation of phytoplankton nutrient limitation in two contrasting low arctic lakes. *Polar Biology.* **29**: 487-494.

Carey SK. 2003. Dissolved organic carbon fluxes in a discontinuous permafrost subarctic alpine catchment. *Permafrost and Periglacial Processes* **14**(2): 161-171.

Cole JJ, Caraco NF, Kling GW, Kratz TK. 1994. Carbon dioxide supersaturation in the surface waters of lakes. *Science* **265**: 1568-1570.

de March L. 1975. Nutrient budgets for a high arctic lake (Char Lake, N.W.T.). *Verhandlungen des Internationalen Verein Limnologie* **19**: 496-503.

Downing JA. 2010. Emerging global role of small lakes and ponds: little things mean a lot. *Limnetica* 29(1): 9-24.

Duff KE, Laing TE, Smol JP, Lean DRS. 1999. Limnological characteristics of lakes located across arctic treeline in northern Russia. *Hydrobiologia* **391**: 205-222.

Edlund SA, Alt BT. 1989. Regional congruence of vegetation and summer climate patterns in the Queen Elizabeth Islands, Northwest Territories Canada. *Arctic* **42**: 3-23.

Elder JF, Rybicki NB, Carter V, Weintraub V. 2000. Sources and yields of dissolved carbon in northern Wisconsin stream catchments with differing amounts of peatlands. *Wetlands* **20:** 113-125.

Engstrom DR. 1987. Influence of vegetation and hydrology on the humus budgets of Labrador lakes. *Canadian Journal of Fisheries and Aquatic Sciences* **44**: 1306-1314.

Finlay J, Neff J, Zimov S, Davydova A, Davydov S. 2006. Snowmelt dominance of dissolved organic carbon in high-latitude watersheds: Implications for characterization of flux of river DOC. *Geophysical Research Letters* **33**: L10401. DOI:10.1029/2006GL025754.

Fraser, CJD, Roulet BT, Moore TR. 2001. Hydrology and dissolved organic carbon biogeochemistry in an ombrotrophic bog. *Hydrological Processes* **15:** 3551-3166.

Graneli W, Bertilsson S, Philibert A. 2004. Phosphorus limitation of bacterial growth in High Arctic lakes and ponds. *Aquatic Sciences* **66**:430-439.

Guo L, Ping CL, Macdonald RW. 2007. Mobilization pathways of organic carbon from permafrost to arctic rivers in a changing climate. *Geophysical Research Letters* **34**:L13603. DOI:10.1029/2007GL030689.

Hamilton PB, Gajewski K, Atkinson DE, Lean DRS. 2001. Physical and chemical limnology of 204 lakes from the Canadian Arctic Archipelago. *Hydrobiologia* **457**: 133-148.

Heron R, Woo MK. 1978. Snowmelt computations for a high arctic site. *Proceedings of the 35th Eastern Snow Conference,* Hanover, New Hampshire, 162-172.

Hinton MJ, Schiff SL, English MC. 1997. The significance of storms for the concentrations and export of dissolved organic carbon from two Precambrian Shield catchments. *Biogeochemistry* **36:** 67-88.

Hood EW, McKnight DM, Williams MW. 2003. Sources and chemical character of dissolved organic carbon across an alpine/subalpine ecotone, Green Lakes Valley, Colorado Front Range, United States. *Water Resources Research* **39**: 1188. DOI:10.1029/2002WR001738.

Hood EW, Williams MW, McKnight DM. 2005. Sources of dissolved organic matter (DOM) in a Rocky Mountain stream using chemical fractionation and stable isotopes. *Biogeochemistry* **74**: 231-255. DOI:10.1007/s10533-004-4322-5.

Hope D, Billett MF, Cresser MS. 1994. A review of the export of carbon in river water: fluxes and processes. *Environmental Pollution* **84:** 301-24.

Huttunen JT, Väisänen TS, Heikkinen M, Hellsten S, Nykänen H, Nenonen O, Martikainen PJ. 2002. Exchange of CO_2, CH_4 and N_2O between the atmosphere and two northern boreal ponds with catchments dominated by peatlands or forests. *Plant Soil* **242**: 137-146.

Jonsson A, Karlsson J, Jansson M. 2003. Sources of carbon dioxide in clearwater and humic lakes in northern Sweden. *Ecosystems* **6:** 224–235.

Kalff J, Welch HE. 1974. Phytoplankton production in Char Lake, a natural polar lake, and in Meretta Lake, a polluted polar lake, Cornwallis Island, Northwest Territories. *Journal of the Fisheries Research Board of Canada* **31:** 621-636.

Kane DL, Yang D. (eds.) 2004. *Northern Research Basins Water Balance*. International Association of Hydrological Sciences 290: Wallingford, United Kingdom.

Kling GW, Kipphut GW, Miller MC. 1991. Arctic lakes and streams as gas conduits to the atmosphere: implications for tundra carbon budgets. *Science* **251:** 298-301.

Kling GW, Kipphut GW, Miller MC, O'Brien WJ. 2000. Integration of lakes and streams in a landscape perspective: The importance of material processing on spatial patterns and temporal coherence. *Freshwater Biology* **43:** 477-497.

Koprivnjak J-F, Moore TR. 1992. Sources, sinks and fluxes of dissolved organic carbon in subarctic fen catchments. *Arctic Alpine Research* **24:** 204-210.

Lafrenière M, Sharp M. 2004. The concentration and fluorescence of Dissolved Organic Carbon (DOC) in glacial and non-glacial catchments: interpreting hydrological flow routing and DOC sources. *Arctic, Antarctic, and Alpine Research* **36**(2): 156-165.

Lafrenière M, Sharp M. 2005. A comparison of solute fluxes and sources from glacial and non-glacial catchments over contrasting melt seasons. *Hydrological Processes* **19**(15): 2991-3012.

Lafrenière M, Lamoureux S. 2008. Seasonal dynamics of dissolved nitrogen exports from two High Arctic watersheds, Melville Island, Canada. *Hydrology Research* **39:** 323-335.

Laurion I, Vincent WF, MacIntyre S, Retamal L, Dupont C, Francus P, Pienitz R. 2010. Variability in greenhouse gas emissions from permafrost thaw ponds. *Limnology and Oceanography* **55:** 115-133.

Lewis T, Lafrenière M, Lamoureux S. 2011. Hydrochemical and sedimentary responses of paired High Arctic watersheds to unusual climate and permafrost disturbance, Cape Bounty, Melville Island, Canada. *Hydrological Processes* DOI: 10.1002/hyp.8335.

Lim DSS, Douglas MSV, Smol JP, Lean DRS. 2001. Physical and chemical limnological characteristics of 38 lakes and ponds on Bathurst Island, Nunavut, Canadian High Arctic. *International Review of Hydrobiology* **86:** 1-22.

Lim DSS, Douglas MSV. 2003. Limnological characteristics of 22 lakes and ponds in the Haughton Crater region of Devon Island, Nunavut, Canadian High Arctic. *Arctic, Antarctic, and Alpine Research* **35:**509- 519.

Lim DSS, Douglas MSV, Smol JP. 2005. Limnology of 46 lakes and ponds on Banks Island, NWT, Canadian Arctic Archipelago. *Hydrobiologia* **545**: 11-32.

Lin LI-K. 1989. A concordance correlation coefficient to evaluate reproducibility. *Biometrics* **45:**255-268.

Luthin JN. 1966. *Drainage Engineering*. Wiley: New York.

Lyons WB, Finlay JC. 2008. Biogeochemical processes in high-latitudes lakes and rivers. In *Polar Lakes and Rivers — Limnology of Arctic and Antarctic Aquatic Ecosystems.* Vincent WF, Laybourn-Parry J (eds). 2008. Oxford University Press: UK.

McKnight DM, Smith RL, Harnish RA, Miller CL, Bencala KE. 1993. Seasonal relationships between planktonic microorganisms and dissolved organic material in an alpine stream. *Biogeochemistry* **21:** 39-59.

McKnight DM, Boyer EW, Westerhoff PK, Doran PT, Kulbe T, Andersen DT. 2001. Spectrofluorometric characterization of dissolved organic matter for indication of precursor organic material and aromaticity. *Limnology and Oceanography* **46:** 38-48.

Michaelson GJ, Ping CL, Kling GW, Hobbie JE. 1998. The character and bioactivity of dissolved organic matter at thaw and in the spring runoff waters of the arctic tundra north slope, Alaska. *Journal of Geophysical Research* **103:** 28,939-28,946.

Michelutti N, Douglas MSV, Muir DCG, Wang XW, Smol JP. 2002a. Limnological characteristics of 38 lakes and ponds on Axel Heiberg Island, High Arctic Canada. *International Review of Hydrobiology* **87:** 385-399.

Michelutti N, Douglas MSV, Lean DRS, Smol JP. 2002b. Physical and chemical limnology of 34 ultraoligotrophic lakes and ponds near Wyniatt Bay, Victoria Island, Arctic Canada. *Hydrobiologia* **482:** 1-13.

Miller MP, McKnight DM. 2010. Comparison of seasonal changes in fluorescent dissolved organic matter among aquatic lake and stream sites in the Green Lakes Valley. *Journal of Geophysical Research-Biogeosciences* **115**: GOOF12. DOI:10.1029/2009JG000985.

Moore TR. 1987. Patterns of dissolved organic matter in subarctic peatlands. *Earth Surface Processes and Landforms* **12:** 387–397.

Muster S. November 18, 2011, Personal Communication.

Naiman RJ. 1982. Characteristics of sediment and organic carbon export from pristine boreal forest watersheds. *Canadian Journal of Fisheries and Aquatic Science* **39:** 1699-1718.

Nettleship DN, Smith PA. 1975. *Ecological Sites in Northern Canada*. Canadian Committee for the International Biological Program Conservation Terrestrial Panel 9, Canadian Wildlife Service, Ottawa, Canada.

Pace ML, Cole JJ. 2002. Synchronous variation of dissolved organic carbon and color in lakes. *Limnology and Oceanography* **47:** 333-342.

Penman HL. 1948. Natural evaporation from open water, bare soil, and grass. *Proc. Roy. Soc. London* **A193:**120-146.

Peterson GS, Johnson LB, Axler RP, Diamond SA. 2002. Assessment of the risk of solar ultraviolet radiation to amphibians. II. In situ characterization of exposure in amphibian habitats. *Environmental Science and Technology* **36:** 2859-2865.

Pienitz R, Smol JP, Lean DRS. 1997a. Physical and chemical limnology of 24 lakes located between Yellowknife and Contwoyto Lake, Northwest Territories (Canada). *Canadian Journal of Fisheries and Aquatic Sciences* **54:** 347-358.

Pienitz R, Smol JP, Lean DRS. 1997b. Physical and chemical limnology of 59 lakes located between the southern Yukon and Tuktoyaktuk Peninsula, Northwest Territories (Canada). *Canadian Journal of Fisheries and Aquatic Sciences* **54:** 330-346.

Prentki RT, Miller MC, Barsdatve RJ, Elley A, Coyne P. 1980. Chemistry. In *Limnology of tundra ponds*. Hobbie JE (ed.). Dowden, Hutchinson & Ross, Stroudsburg, PA.

Priestley CHB, Taylor RJ. 1972. On the assessment of surface heat flux and evaporation using large-scale parameters. *Monthly Weather Review* **100:** 81-82.

Ramlal P, Hesslein SRH, Hecky RE, Fee EJ, Rudd JWM, Guildford SJ. 1994. The organic carbon budget of a shallow arctic tundra lake on the Tuktoyaktuk Peninsula, N.W.T., Canada: arctic lake carbon budget. *Biogeochemistry* **24:** 145-172.

Raymond PA, McClelland JW, Holmes RM, Zhulidov AV, Mull K, Peterson BJ, Striegl RG, Aiken GR, Gurtovaya TY. 2007. Flux and age of dissolved organic carbon exported to the Arctic Ocean: A carbon isotopic study of the five largest arctic rivers. *Global Biogeochemical Cycles* **21:** GB4011 DOI:10.1029/2007GB002934.

Rosenberry DO, Stannard DI, Winter TC, Martinez MJ. 2004. Comparison of 13 equations for determining evapotranspiration from a prairie wetland, Cottonwood Lake Area, North Dakota, USA. *Wetlands* **24:**483-497.

Schindler DW, Welch HE, Kalff J, Brunskill GH Kritsch N. 1974. Physical and chemical limnology of Char Lake, Cornwallis Island (75 N lat.). *Journal of the Fisheries Research Board of Canada* **31:** 585-607.

Schindler DW, Curtis PJ, Bayley SE, Parker BR, Beaty KG, Stainton MP. 1997. Climate-induced changes in the dissolved organic carbon budgets of boreal lakes. *Biogeochemistry* **36:** 9–28.

Smith LC, Sheng Y, Macdonald GM, Hinzman LD. 2005. Disappearing Arctic lakes. *Science* **308:**1429.

Smol JP, Douglas MSV. 2007. Crossing the final ecological threshold in High Arctic ponds. *Proceedings of the National Academy of Sciences* **104:**12395-12397.

Spence C, Woo MK. 2003. Hydrology of a subarctic Canadian shield: soil-filled valleys. *Journal of Hydrology* **278:** 111-127.

Stewart RB, Rouse WR. 1977. Substantiation of the Priestley and Taylor parameter α = 1.26 for potential evaporation in high latitudes. *Journal of Applied Meteorology* **16**(6): 649-650.

Stumm W, Morgan JJ. 1996. *Aquatic Chemistry: Chemical Equilibria and Rates in Natural Waters,* 3rd ed. John Wiley: Hoboken, N. J.

Tranvik LJ, Downing JA, Cotner JB, Loiselle SA, Striegl RG, Ballatore TJ, Dillon P, Finlay K, Fortino K, Knoll LB, Kortelainen PL, Kutser T, Larsen S, Laurion I, Leech DM,. McCallister SL, McKnight DM, Melack JM, Overholt E, Porter JA, Prairie YT, Renwick WH, Roland F, Sherman BS, Schindler DW, Sobek S, Tremblay A, Vanni MJ, Verschoor AM, Von Wachenfeldt E, Weyhenmeyer GA. 2009. Lakes and reservoirs as regulators of carbon cycling and climate. *Limnology and Oceanography* **54**: 2298-2314.

Walter Anthony K, Dragos M, Vas A, Brosius L, Chapin III FS, Zimov SA, Zhuang Q. 2010. Estimating methane emissions from northern lakes using ice bubble surveys. *Limnology and Oceanography Methods* **8**: 592-609.

Whalen SC, Cornwell C. 1985. Nitrogen, phosphorus, and organic carbon cycling in an arctic lake. *Canadian Journal of Fisheries and Aquatic Sciences* **42:** 797-808.

Woo MK. 1997. *A guide for ground based measurement of the arctic snow cover.* Canadian Snow Data CD, Meteorological Service of Canada, Downsview, Ontario. http: www.ccin.ca/datasets/snowcd/docs/1999/DOCUMENTS/WOO.PDF [31 October 2006].

Woo MK, Heron R, Steer P. 1981. Catchment hydrology of a High Arctic lake. *Cold Regions Science and Technology* **5:** 29-41.

Woo MK, Guan XJ. 2006. Hydrological connectivity and seasonal storage change of tundra ponds in a polar oasis environment, Canadian High Arctic. *Permafrost and Periglacial Processes* **17:** 309-323.

Woo MK, Young KL. 2006. High Arctic wetlands: Their occurrence, hydrological characteristics and sustainability. *Journal of Hydrology* **320:** 432-450.

Yoshikawa K, Hinzman LD. 2003. Shrinking thermokarst ponds and groundwater dynamics in discontinuous permafrost near Council, Alaska. *Permafrost and Periglacial Processes* **14:** 151-160.

Young KL, Woo MK, Edlund SA. 1997. Influence of local topography, soils, and vegetation on microclimate and hydrology at a High Arctic site, Ellesmere Island, Canada. *Arctic and Alpine Research* **29:**270-284.

Young KL, Abnizova A. 2008. Hydrologic status of High Arctic ponds in a continuous permafrost environment, Somerset Island, Nunavut. *9th International Permafrost Conference,* Fairbanks, Alaska, p. 2003-2008.

Young KL, Assini J, Abnizova A, DeMiranda N. 2010. Hydrology of hillslope-wetland streams, Polar Bear Pass, Nunavut, Canada. *Hydrologic Processes* **24**: 3345-3358.
Young KL, Labine C. 2010. Summer hydroclimatology of an extensive low gradient wetland: Polar Bear Pass, Bathurst Island, Nunavut, Canada. *Hydrology Research* **41:** 492-502.
Young KL, Abnizova A. 2011. Hydrological thresholds of ponds in a polar desert wetland environment, Somerset Island, Nunavut, Canada. *Wetlands* **31:** 535-549.

Tables

Table 1. DOC composition of lakes and ponds for selected Arctic regions (adapted and modified from Table 8.2. Lyons and Finlay, 2008). Snapshot studies refer to samples taken once a year (*e.g.*, helicopter survey); short-term studies indicate when samples have been taken several times over a period of two months; and seasonal implies that samples have been regularly obtained from spring to fall

Study	Land Cover	Mean DOC (mg L^{-1})	Range	Period of sampling	Category
1. Antoniades *et al.* (2003)	polar desert	2.1	0.9-46	July 21-31, 1996	snapshot
2. Brutemark *et al.* (2006)	tundra	65.5	41-90	June 20-26 2003	snapshot
3. Duff *et al.* (1999)	tundra	4.05	1.7-12.3	July and Aug 1993, 1994	snapshot
4. Granelli *et al.* (2004)	tundra	1.1	0.6-2.6 (lakes)	August 1999	snapshot
		7.6	4.4-14.7 (ponds)		
5. Hamilton *et al.* (2001)	variable	3.93	0.1-31.9	18 years compilation	snapshot
6. Jonsson *et al.* (2003)	tundra	4.18	0.7-9.9	June- September 2000 (3 times)	snapshot
7. Kling *et al.*(2000)	tundra	3.48	2.0-13.5	1-4 times in 1990-97	snapshot
8. Lim *et al.* (2001)	polar desert	3.55	1.2-17.4	June 25-July 4 2000	snapshot
9. Lim and Douglas (2003)	polar desert	1.95	1.2-7.4	July 8-29 2000	snapshot
10. Lim *et al.* (2005)	polar desert	6.11	1.2-17.4	June 25th–July 4th, 2000,	snapshot
11. Michelutti *et al.* (2002a)	polar desert	3.51	1.0-18.5	July 19-23 1998	snapshot
12. Michelutti *et al.* (2002b)	polar desert	1.19	0.3-3.7	July 6-17 1997	snapshot
13. Pienitz *et al.* (1997a)	boreal forest, forest tundra, arctic tundra	4.3	1.6-0.1	July1991	snapshot
14. Pienitz *et al.* (1997b)	boreal forest, forest tundra, arctic tundra, alpine tundra	12.3	3.1-35.1	July 1990	snapshot
15. Barley *et al.* (2006)	variable	9.4	0.5-31.3	1992-2002	short-term, snapshot
16. Breton *et al.* (2011)	tundra	9.9	1.3-26.0	July and Aug 2004 and 2005	short-term
17. Laurion *et al.* (2010)	tundra	9.4	1.5-20.8	July 21-27 2006, June 28-July 24	short-term

				2007	
18. Abnizova *et al.* (this study)	tundra	12.8	1.5-41.6	June- August 2008-2009	seasonal
19. Prentki *et al.* (1980)	tundra	n/a	4.5-15.5	June-August 1971-73	seasonal
20. Schindler *et al.* (1974)	tundra	<0.09	n/a	1969-72	seasonal
21. Whalen and Cornwell (1985)	tundra	6.9	3.2-9.4	May-August 1980	seasonal

Table 2. Detailed limnological characteristics and snow cover of the study ponds, June to August in 2008 and 2009.

Type and Name	Snowbed Pond		Creek Pond		Temporarily Connected Pond		Isolated Large Pond		Isolated Small Pond	
Terrain type	fine-grained/ice-rich		fine-grained/ice-rich		coarse-textured/ice poor		fine-grained/ice-rich		fine-grained/ice-rich	
Maximum ground thaw depth (m)	0.40		0.29		0.28		0.18		0.20	
Surface area (m^2)	7242		918		380		6554		818	
Volume (m^3)	1452		222		36		941		72	
Area/Volume	4.99		4.14		10.66		6.96		11.39	
	2008	2009	2008	2009	2008	2009	2008	2009	2008	2009
Mean snow depth (mm)	147	131	255	161	186	176	307	250	185	185
Mean snow density (kg m^{-3})	195	285	229	294	212	301	270	304	260	298
Mean SWE (mm)	29	37	59	48	39	53	83	76	48	55
Mean pH	8.1	8.2	8.1	8.2	8.3	8.4	8.7	8.2	8.2	8.3

Table 3. Monthly air temperature, Penman potential evaporation and total precipitation at Polar Bear Pass, June to August, 2008 and 2009.

Month		T_{air} (°C)		E_0 (mm d^{-1})	P (mm)
	Mean	Max	Min	Mean	Total amount
2008					
June	2.7	8.8	-2.5	3.5	19.5
July	7.0	13.0	1.3	3.5	42.6
August	2.0	8.2	-3.4	1.5	32.4
2009					
June	1.9	9.8	-3.7	3.3	2.7
July	5	12.1	1.6	2.7	61.7
August	4.5	8.6	0.7	1.7	31.1

Table 4. Pond water balance in 2008 and 2009[a] .Values are given as totals in mm for 2008 from 2 June to August 31 and for 2009 from 3 June to August 31.

Snowbed pond	P	E	ΔS_{meas}	$\Delta S_{est} = (P-E)$
2008	82	170	55	-87
2009	95	148	47	-54
Creek pond				
2008	82	154	51	-72
2009	95	138	20	-43
TC pond				
2008	82	155	31	-73
2009	95	138	77	-43
IL pond				
2008	82	154	-47	-72
2009	95	139	-11	-45
IS pond				
2008	82	155	-7	-73
2009	95	139	-6	-44

Table 5. Synchrony values determined by Pearson correlation (*R*) with *s* indicating the slope of the regression line and strength of agreement determined by Lin's (Lin, 1989) concordance correlation coefficient (*Rc*) for DOC concentrations across all pond pairs. In bold are significant coefficient values ($p < 0.05$).

	Snowbed pond		*Creek pond*		*TC pond*		*IL pond*	
	2008	2009	2008	2009	2008	2009	2008	2009
IS pond	*R*=0.30	***R*=0.65**	*R*=0.28	***R*=0.73**	*R*=0.56	***R*=0.88**	*R*=0.32	***R*=0.79**
	s=0.11	*s*=0.46	*s*=0.8	*s*=0.47	*s*=0.31	*s*=0.59	*s*=0.06	*s*=0.57
	Rc=0.07	*Rc*=0.10	*Rc*=0.06	***Rc*=0.15**	***Rc*=0.46**	***Rc*=0.37**	*Rc*=0.05	***Rc*=0.18**
	R=0.56	***R*=0.87**	*R*=0.87	***R*=0.90**	*R*=0.50	***R*=0.93**		
	s=0.35	*s*=0.85	*s*=-0.07	*s*=0.81	*s*=0.82	*s*=0.86		
IL pond	*Rc*=0.21	***Rc*=0.66**	*Rc*=-0.07	***Rc*=0.89**	*Rc*=0.06	***Rc*=0.54**		
	***R*=0.88**	***R*=0.90**	***R*=0.85**	***R*=0.93**				
	s=0.58	*s*=0.99	*s*=0.45	*s*=0.90				
TC pond	**Rc=0.17**	**Rc=0.28**	**Rc=0.15**	**Rc=0.50**				
	***R*=0.92**	***R*=0.95**						
	s=1.15	*s*=1.03						
Creek pond	**Rc=0.86**	**Rc=0.72**						

Table 6. Yields of dissolved organic carbon (DOC) in g C m^{-2} day^{-1} during 2009 estimated from inflow measurements into a lowland pond, a pond fed by a late-lying snowbed and inflow from Landing Strip Creek (*LSC*) into wetland area.

Terrain type	Source area	End point	Watershed area (m^2)	Average snowmelt yield (Y_s)	Rain event yield (Y_r)	Average post-snowmelt yield (Y_p)	Annual yield [a] $(Y_s + Y_r)+ 60Y_p$
Lowland	wet meadow wetland catchment	a lowland pond	28620	0.85 (n=2)	0.27 (n=1)	0.016 (n=5)	(1.96) + 0.96 = 2.92
Hillslope	*late-lying snowbed* catchment	*Snowbed pond*	78280	0.43 (n =1)	0.17 (n =1)	0.04 (n=6)	(0.60) + 2.54 = 3.14
Hillslope	*LSC* catchment	lowland wet meadow	287508	0.02 (n =2)	3.47×10^{-6} (n=1)	1.84×10^{-4} (n=7)	(0.02) + 0.01 =0.03

[a]Calculation of annual yield is based on multiplication of average post-snowmelt yield by 60 days. This estimate does not include snowmelt and rain event yields as indicated in () brackets.

Figures

Figure 1. A circumpolar map representing locations of study sites listed in Table 1. Locations of the studies are numbered using the order they are shown in Table 1. Snapshot studies are shown with circles. Here, shaded circles represent the studies where study ponds are nearby each other. Open circles represent studies where study ponds are sampled along long transects and show the centres of these sampling transects. Short-term studies are shown with triangles and long-term studies are indicated with stars. Source: a collection of assorted illustrations for the Arctic Council by Hugo Ahlenius, UNEP/GRID-Arendal.

Figure 2. Radar image (August 13, 2009) showing location of study area (75° 40'N, 98° 30'W) at Polar Bear Pass, Bathurst Island, Nunavut (a) (source: S. Muster, Alfred Wegener Institute for Polar and Marine Research). Inset shows location of Bathurst Island in black and Polar Bear Pass is highlighted with a white rectangle within the Queen Elizabeth Islands. Main study ponds, groundwater well transects, a late-lying snowbed, and a hillslope creek (LSC) are indicated in (b). The Automatic Weather Station (AWS) in the central part of the wetland is indicated by a circle and sampling points for carbon sources are labelled with stars. A pond located upslope from the *TC pond* is shown with a rectangle. Well transect lines downslope and adjacent to the late-lying snowbed are indicated with white dashed lines.

Figure 3. Maximum and minimum daily air temperature, daily precipitation, daily net radiation, and daily evaporation estimated for open water and the wet meadow in 2008 and 2009. Data are plotted using information from the wetland automatic weather station (AWS).

Figure 4. Seasonal water table (solid line), frost table (black cross), and daily evaporation (vertical bar) of the study ponds in 2008 (a) and 2009 (b). Dashed line indicates the ground surface. Start and end of ground thaw is indicated with arrows. Note that 2008 was a leap year.

Figure 5. Water balance of study ponds during the snowfree season in 2008 and 2009. Cumulative calculated storage change ($\Delta S_{est} = P{-}E$, Precipitation minus Evaporation) is shown for comparison. Note that 2008 was a leap year.

Figure 6. DOC and DIC concentrations of water samples collected from the study ponds in 2008 and 2009. Gray symbols indicate large ponds (> 6000 m^2) and open symbols indicate small ponds (< 1000 m^2). Circles indicate hydrologically connected ponds, triangles indicate ponds with intermittent hydrologic linkage, and squares indicate hydrologically isolated ponds. Snowmelt onset (black solid line) and end (dashed line) are indicated.

Figure 7. DOC and DIC stocks (g C m^{-2}) in the study ponds in 2008 and 2009. Total daily precipitation values are provided for comparison. Snowmelt onset (black solid line) and end (dashed line) are provided. Disappearance of the *late-lying snowbed* is indicated with a dashed gray line. Circles indicate hydrologically connected ponds, triangles indicate ponds with intermittent hydrologic linkage, and squares indicate hydrologically isolated ponds. Discharge and DOC from Landing Strip Creek (*LSC)* are also shown.

Figure 8. Fluorescent index values in the *Snowbed* and *Creek ponds* (a) and *IS, IL* and *TC ponds* (b) in 2008. Values of the terrestrial (Suwanee River, SWFA = 1.54 and Suwanee River Natural Organic Matter, SRNOM = 1.86) and microbial (Pony Lake, PLFA = 2.00) end-member fulvic acid values are shown as dashed lines. *FI* in all ponds fluctuated around the value of the SRNOM terrestrial reference throughout the season. Circles indicate hydrologically connected ponds, triangles indicate ponds with intermittent hydrologic linkage, and squares indicate hydrologically isolated ponds.

Figure 9. DOC concentrations, in inflow, in outflow and in the ponds connected to a hillslope creek (*Creek pond*) (a), and a late-lying snowbed (*Snowbed pond*) (b) in 2009. Water movement from the catchment into the *Snowbed pond* (c) and from (d) the *Snowbed pond* during snowmelt on June 16, 2010 (d).

Figure 10. Estimated ground water flux (mm d^{-1}) from the late-lying snowbed in 2008 (a) and 2009 (b). Also provided are DOC concentrations from the snowbed in 2008.

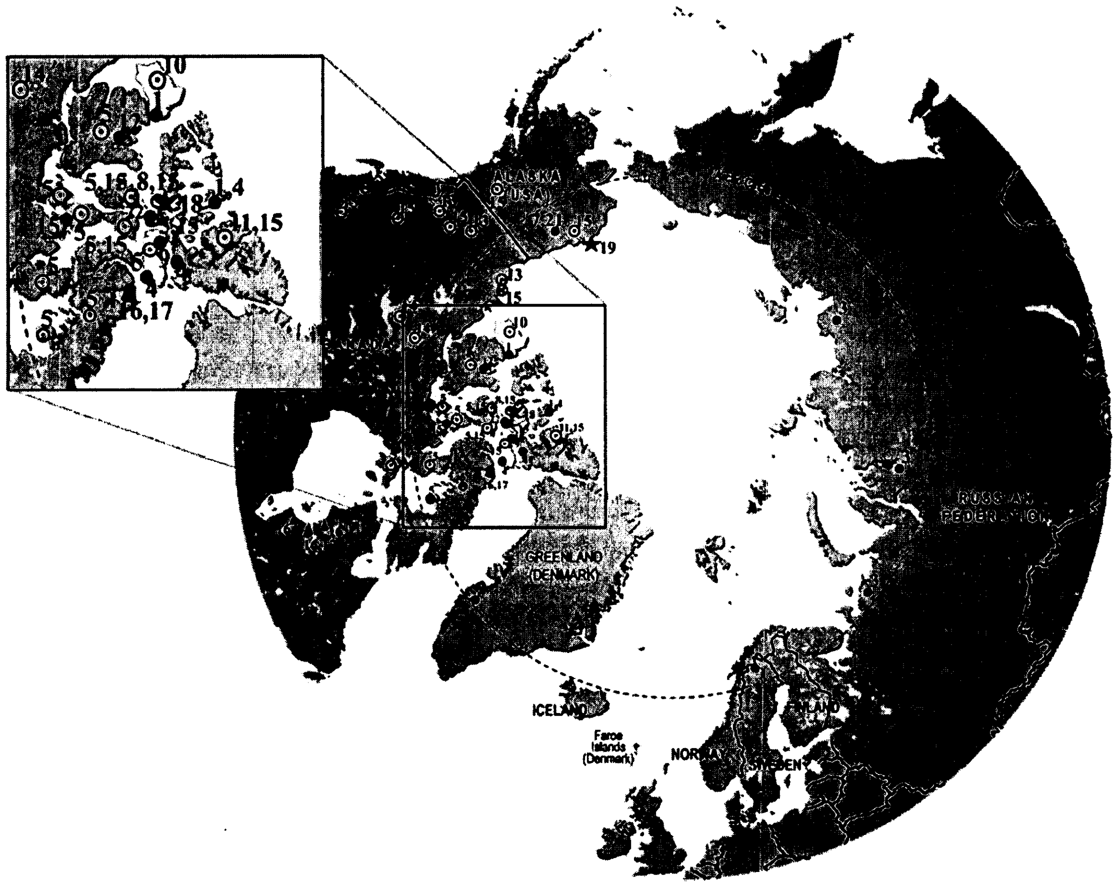
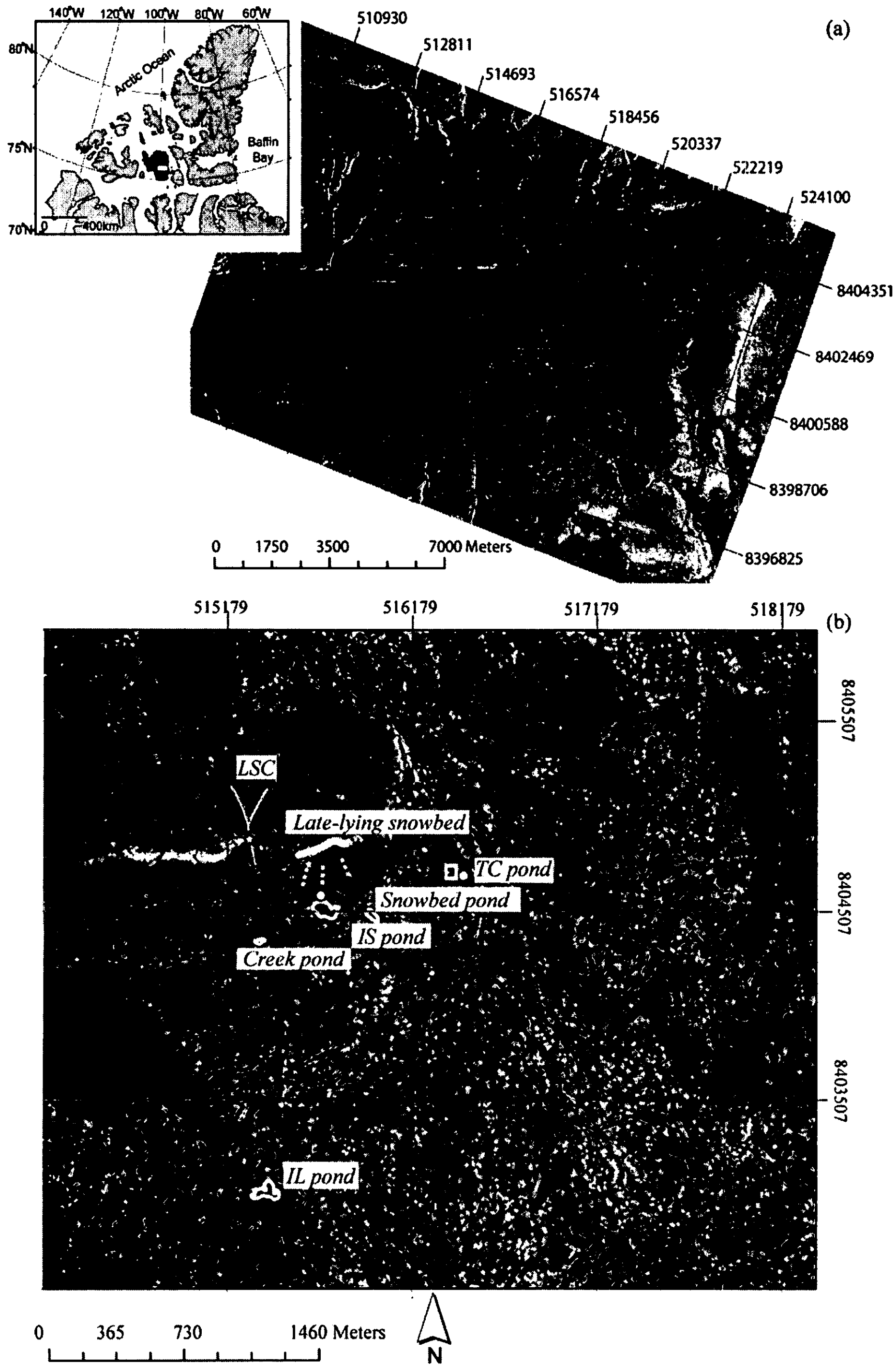
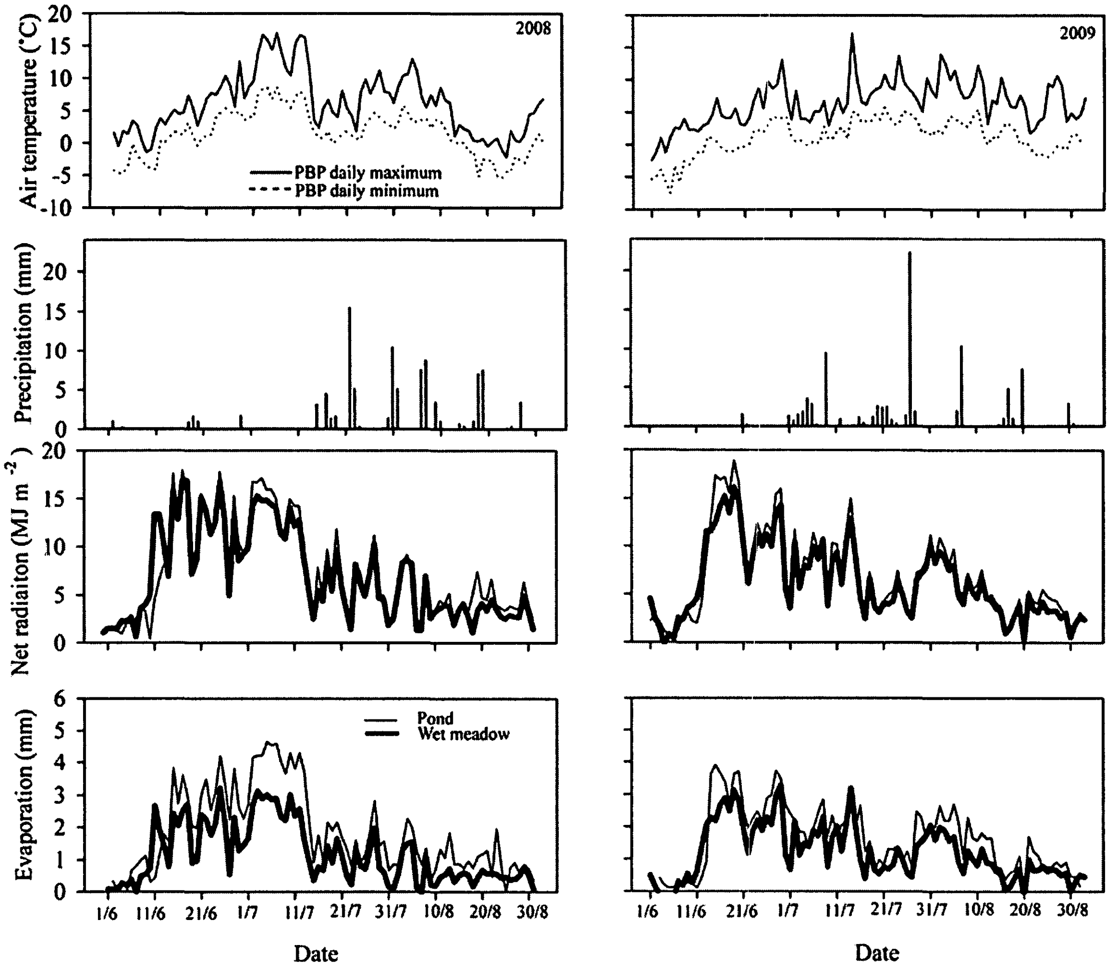


Figure 1



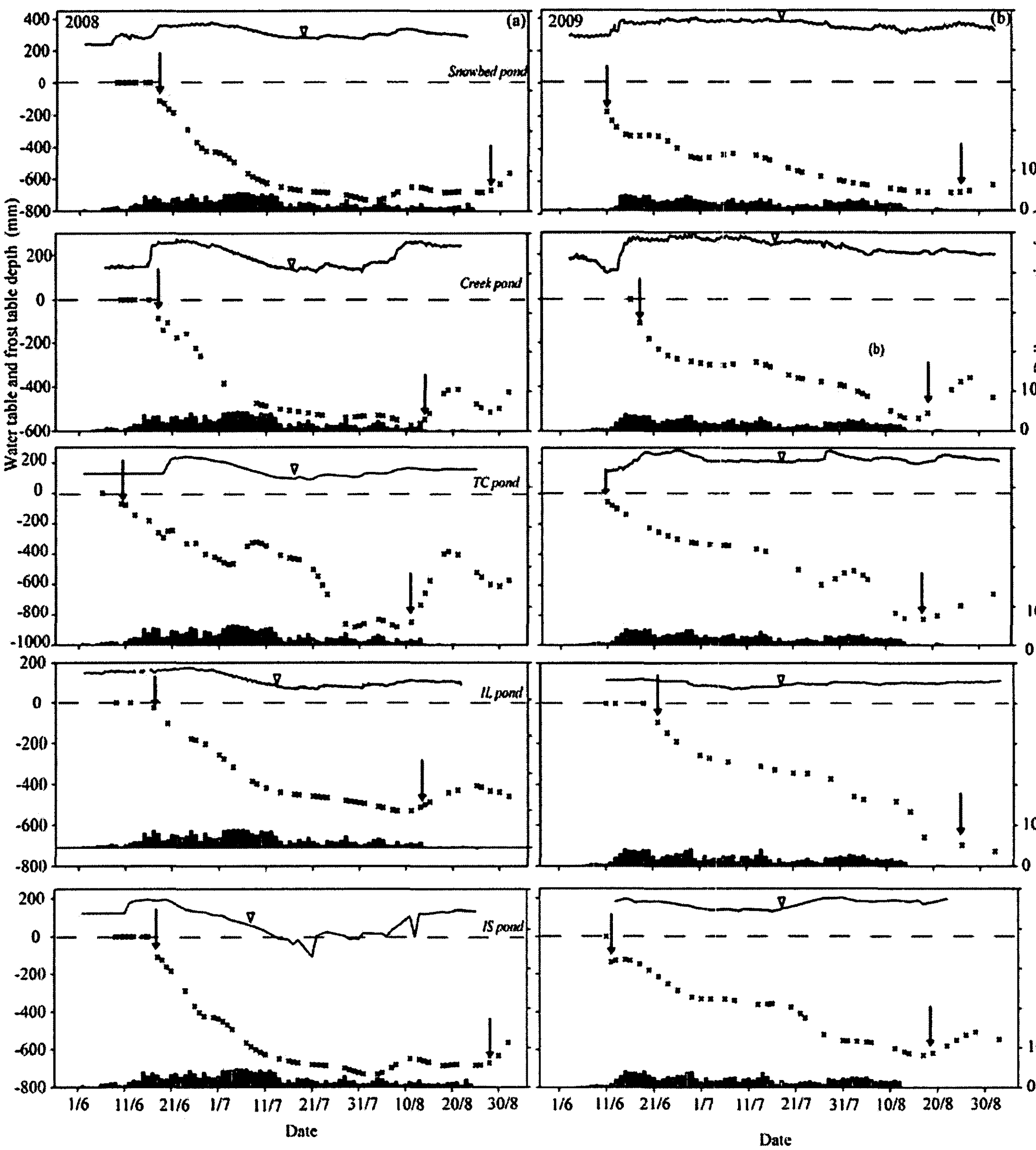
140°W
120°W
100°W
80°W
60°W
80°N
75°N
70°N
Arctic Ocean
Baffin Bay
0
400km
510930
512811
514693
516574
518456
520337
522219
524100
(a)
8404351
8402469
8400588
8398706
8396825
0
1750
3500
7000 Meters
515179
516179
517179
518179
(b)
8405507
8404507
8403507
LSC
Late-lying snowbed
TC pond
Snowbed pond
IS pond
Creek pond
IL pond
0
365
730
1460 Meters
N

Figure 2



Air temperature (°C)
20
15
10
5
0
-5
-10
2008
2009
PBP daily maximum
PBP daily minimum
Precipitation (mm)
20
15
10
5
0
Net radiaiton (MJ m -2)
20
15
10
5
0
Evaporation (mm)
6
5
4
3
2
1
0
Pond
Wet meadow
1/6
11/6
21/6
1/7
11/7
21/7
31/7
10/8
20/8
30/8
Date

Figure 3



2008
(a)
2009
(b)
Snowbed pond
Creek pond
TC pond
IL pond
IS pond
Water table and frost table depth (mm)
Daily evaporation (mm)
Date
Date

Figure 4

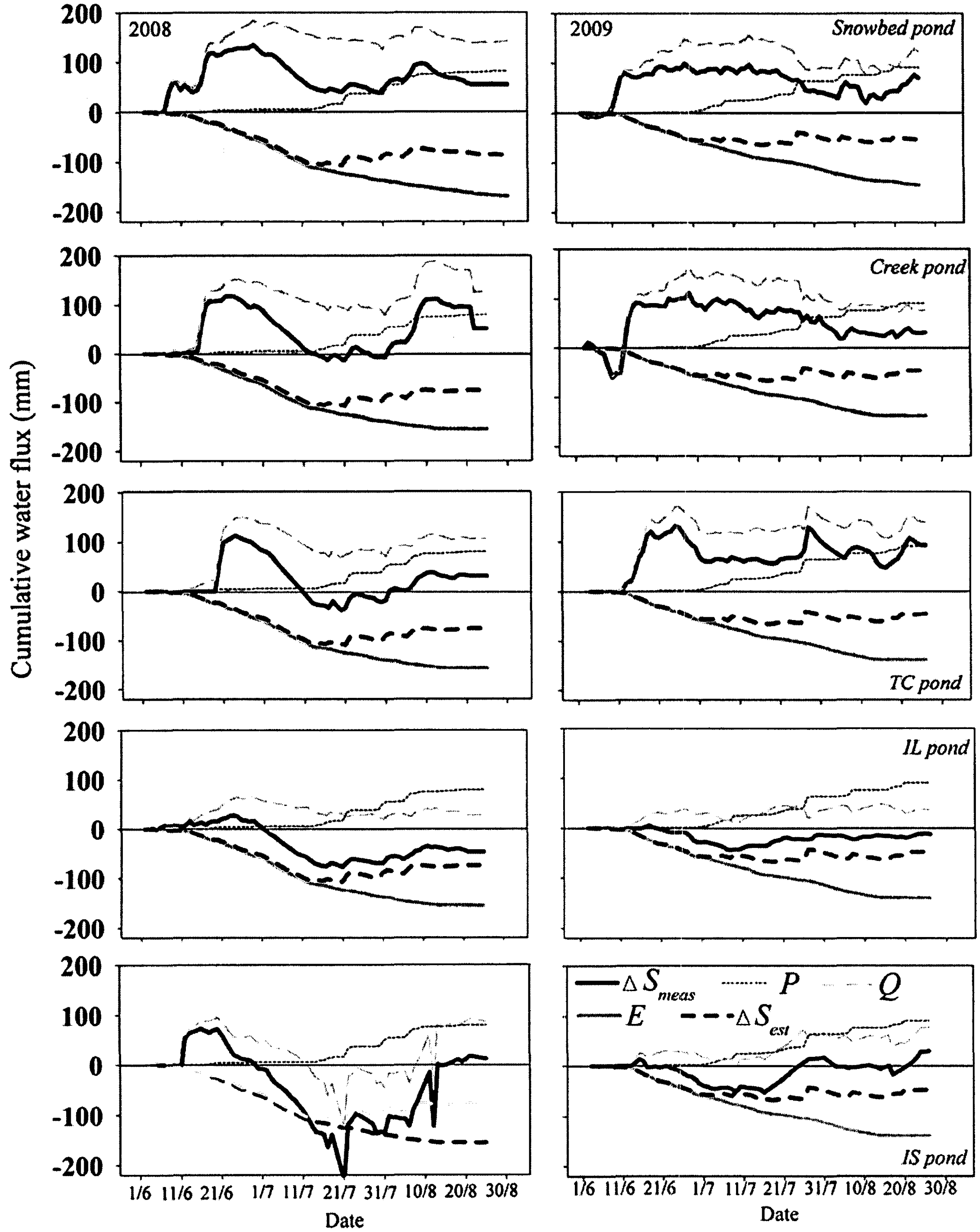
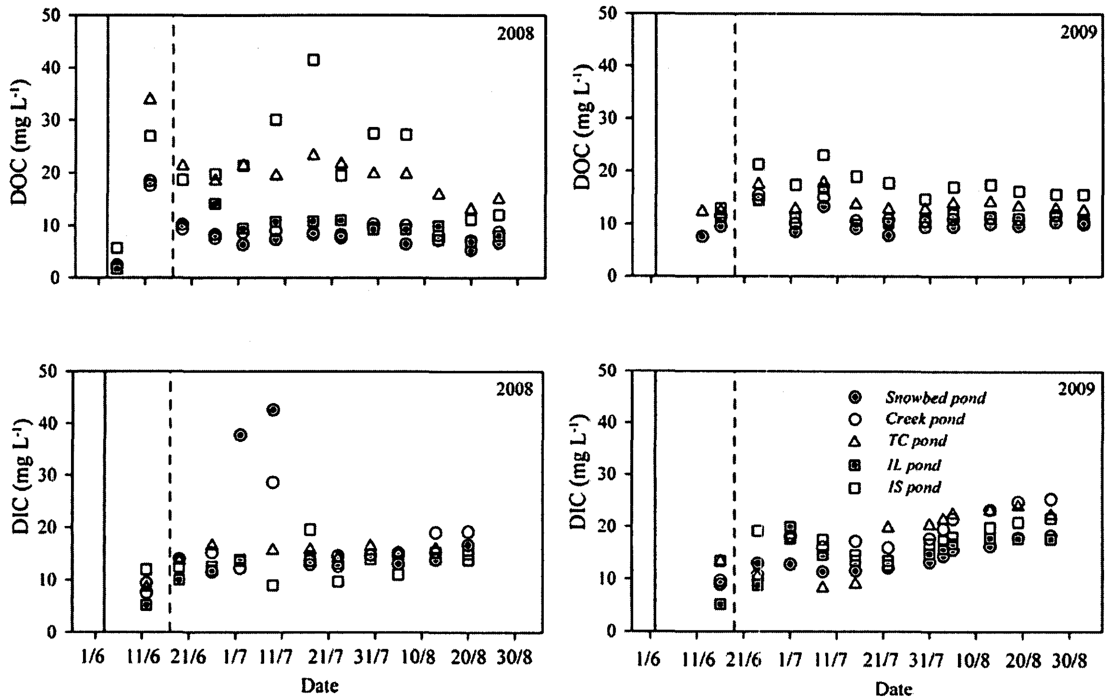
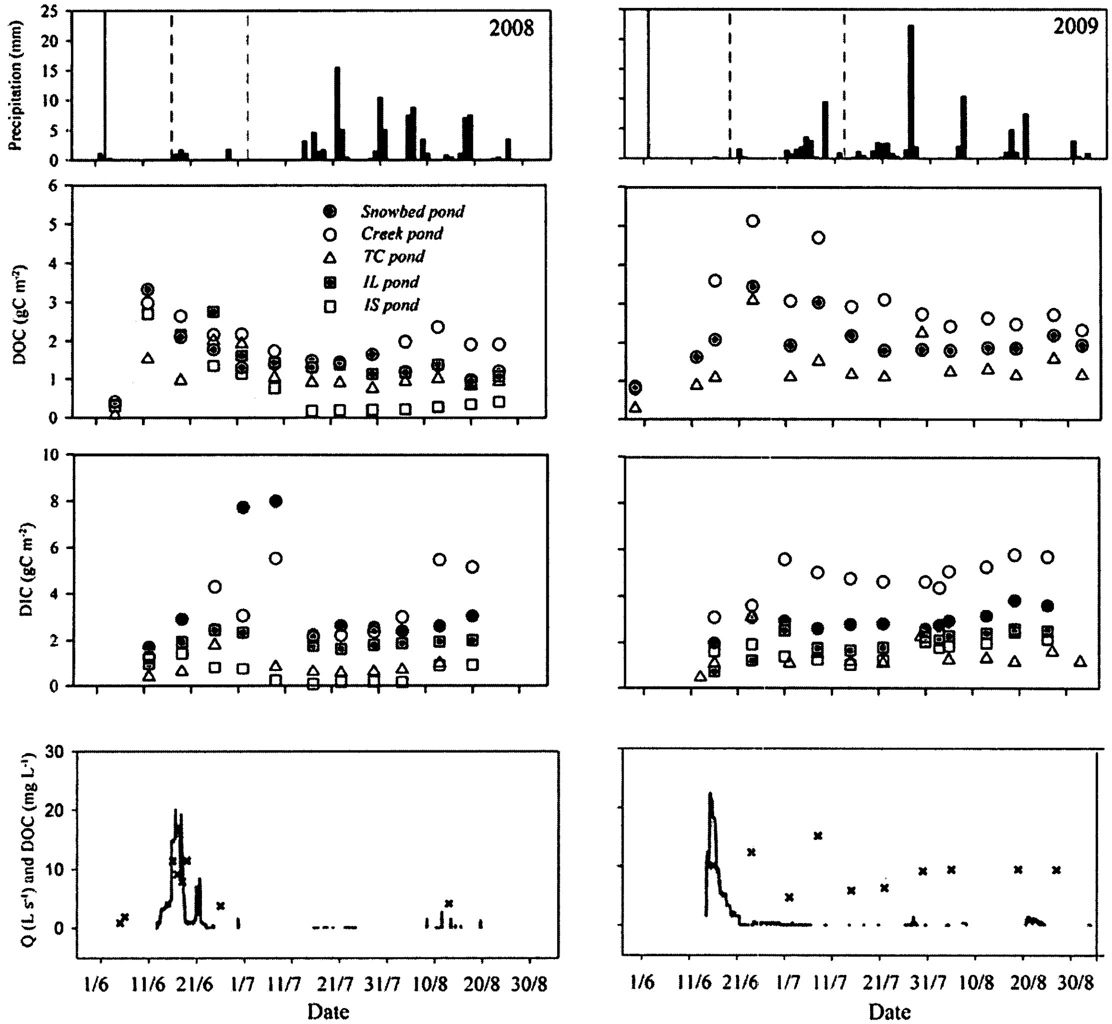

2008
2009
Snowbed pond
Creek pond
TC pond
IL pond
IS pond
200
100
0
-100
-200
Cumulative water flux (mm)
1/6 11/6 21/6 1/7 11/7 21/7 31/7 10/8 20/8 30/8
Date
ΔS_{meas}
P
Q
E
ΔS_{est}

Figure 5



2008
DOC (mg L-1)
2009
DOC (mg L-1)
2008
DIC (mg L-1)
2009
DIC (mg L-1)
Snowbed pond
Creek pond
TC pond
IL pond
IS pond
1/6 11/6 21/6 1/7 11/7 21/7 31/7 10/8 20/8 30/8
Date

Figure 6



2008
2009
Precipitation (mm)
DOC (gC m-2)
DIC (gC m-2)
Q (L s-1) and DOC (mg L-1)
Snowbed pond
Creek pond
TC pond
IL pond
IS pond
1/6 11/6 21/6 1/7 11/7 21/7 31/7 10/8 20/8 30/8
Date

Figure 7

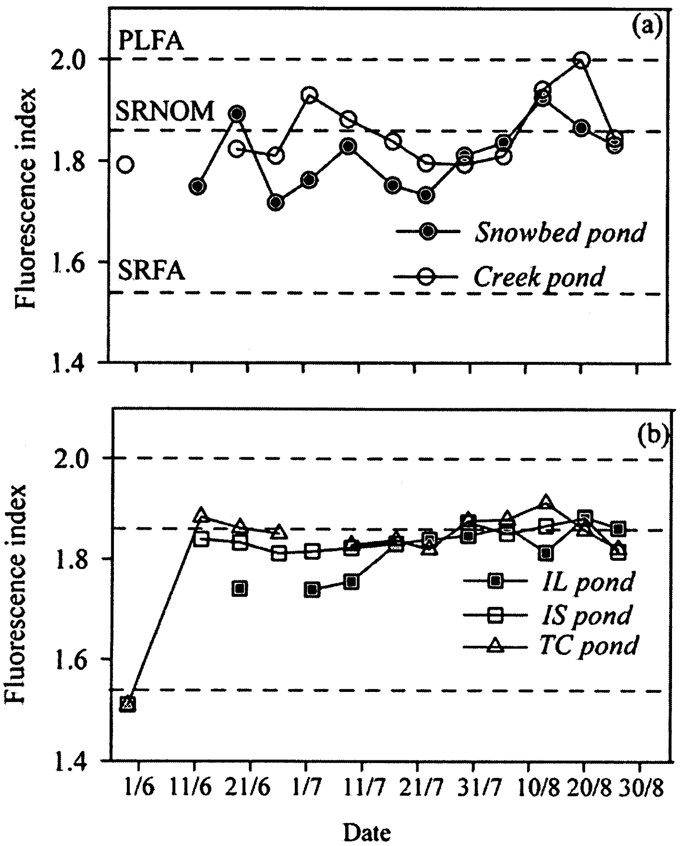

(a)
PLFA
SRNOM
SRFA
Fluorescence index
2.0
1.8
1.6
1.4
Snowbed pond
Creek pond
(b)
IL pond
IS pond
TC pond
1/6 11/6 21/6 1/7 11/7 21/7 31/7 10/8 20/8 30/8
Date

Figure 8

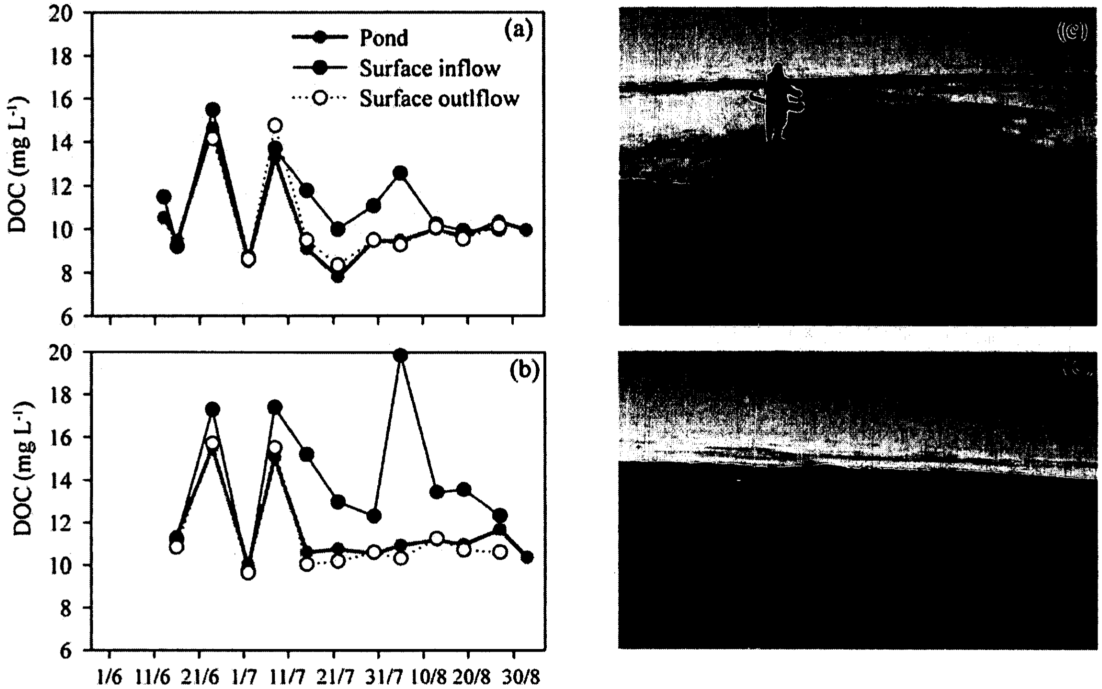
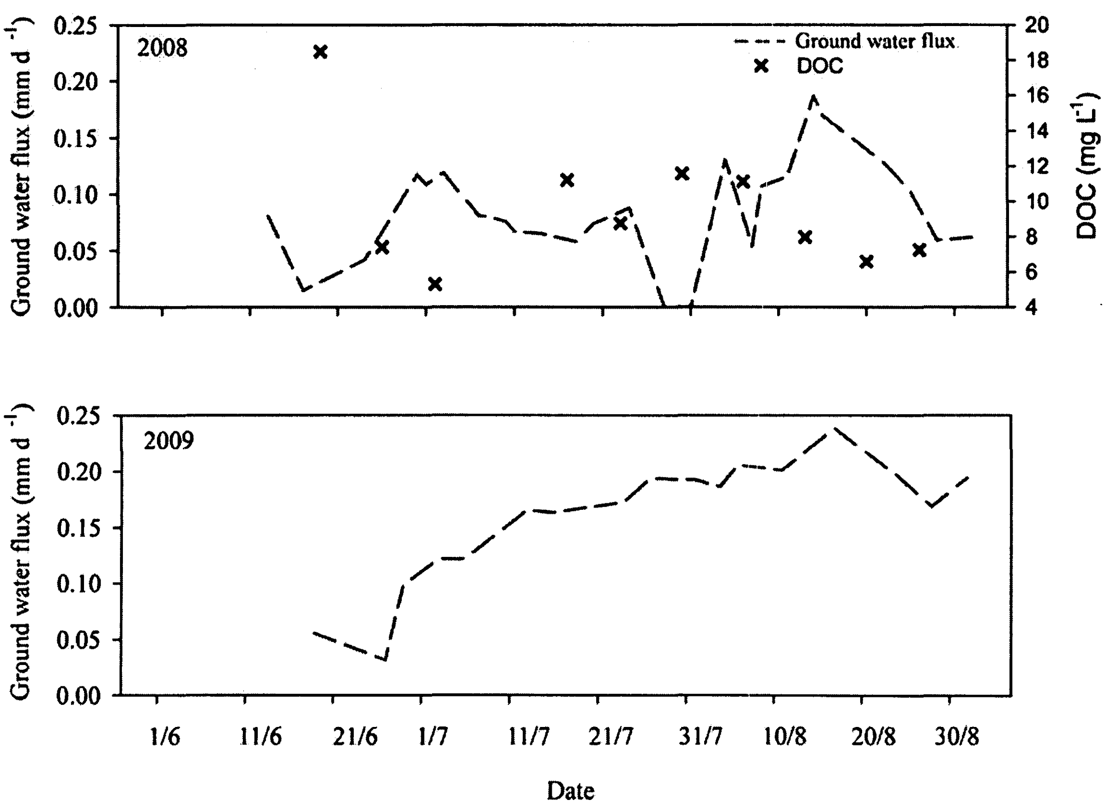

(a)
(b)
(c)
Pond
Surface inflow
Surface outlflow
DOC (mg L-1)
20
18
16
14
12
10
8
6
1/6 11/6 21/6 1/7 11/7 21/7 31/7 10/8 20/8 30/8

Figure 9



2008
Ground water flux
DOC
Ground water flux (mm d -1)
DOC (mg L-1)
0.25
0.20
0.15
0.10
0.05
0.00
20
18
16
14
12
10
8
6
4
2009
1/6
11/6
21/6
1/7
11/7
21/7
31/7
10/8
20/8
30/8
Date

Figure 10

Additional comments:

Page 42: It is important to note that the certainty of the DIC estimates and associated calculations in the current study can be possibly affected by 1) addition of CO_2 resulted from degradation and oxidation of organic carbon in water samples during transportation, and 2) the precision of in-field pH values applied when estimating DIC using standard carbonate equilibrium relationships following approach by Stumm and Morgan (1996).

Page 50: Examination of seasonal changes in the extent of surface area and volume of the study ponds at PBP showed that hydrologically connected ponds experienced 56-57 % in volume and 7-31 % surface area reduction. Hydrologically isolated ponds experienced 18-100 % reduction in surface area extent and 49-100 % in the extent of their volumetric change.

CHAPTER 3: Snowmelt trends and variability in pond hydrology and dissolved carbon dynamics in a High Arctic wetland

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Author Contributions: A. Abnizova, K. Young - formulated study design with advise from J. Boike; A Abnizova - performed data collection, analysis and the write-up of the manuscript; K. Young – edited the manuscript; J. Boike – contributed to water chemistry analyses.

Abstract

The snowmelt period is recognized as an important event transporting large fluxes of water, carbon and nutrients in High Arctic watersheds, yet studies seldom include intensive measurements and estimates of aquatic-atmospheric carbon exchange during this period. Ponds are the dominant water bodies in arctic wetland environments and they represent substantial reservoirs of carbon by acting as hydrological conduits and sources of carbon to the atmosphere. In this study we quantified snowmelt water budgets and carbon stores of dissolved organic and inorganic carbon in small shallow ponds with various degrees of hydrologic connectivity to the catchment at an extensive low-gradient Arctic wetland Polar Bear Pass (75°40'N, 98°30'W). Water budgets of ponds hydrologically connected to a source (e.g. a late-lying snowbed or a hillslope creek) remained positive even until the early post-snowmelt period when the inputs from their sources diminished. Opposite to ponds with stable hydrologic linkages, negative water budgets following the snowmelt freshet were observed in a pond temporarily connected to an upslope neighbour pond and an isolated pond receiving water from precipitation only. Dissolved carbon stores in hydrologically connected ponds were higher than those in the isolated ponds and remained stable after the snowmelt period ended.

We also estimated the flux of carbon dioxide (CO_2) over the snowmelt period from 42 ponds across the wetland. Most ponds were strong net sources of CO_2 to the atmosphere during snowmelt. CO_2 evasion from wetland ponds was different across the Pass and was controlled by spatially variable melt patterns in the wetland. A tentative dissolved carbon balance estimate of a wetland pond hydrologically connected to a late-lying snowbed showed that snowmelt runoff is important to gas emissions by providing inflows rich in DOC and DIC. A decline in melt runoff from the snowbed reduced inputs of dissolved carbon reversing the CO_2 flux of this pond from a source to a sink. Our results demonstrate that variability in pond carbon pools is strongly controlled by the presence of hydrologic linkages to the catchment. Hydrologic connectivity during snowmelt augments the delivery of DIC and DOC to the ponds contributing to pond CO_2 emissions. The release of CO_2 from wetland ponds averaged 0.37 g C $m^{-2}d^{-1}$ demonstrating that snowmelt is an important period in the seasonal estimates of CO_2 emissions.

Key words: wetlands; arctic ponds; snowmelt hydrology; carbon; DOC and DIC

Introduction

The vulnerability of arctic wetlands to climate change has been widely recognized because of the potential for higher temperatures to significantly alter the hydrology and hence the ecology of lakes and ponds which typically dominate these landscapes (Rouse and others 1997). In contrast to lakes, ponds are the most abundant water bodies in the Arctic measuring less than 0.0001 km^2 on an aerial scale (Downing and others 2006). Some scientists classify them as being shallow water bodies smaller than 1 ha and 3 m in depth (Rautio and others 2011), though the Canadian Wetland Classification scheme suggests that they can only reach 2 m in depth (National Wetlands Working Group 1997).

The hydrological balance of these shallow arctic water bodies is highly sensitive to hydro-climatic changes such as seasonal snowmelt, rainfall and evaporation losses. Occasionally, depending on a pond's linkage with its surrounding catchment, additional water can be gained from meltwaters draining from an upslope late-lying snowbed, a nearby creek or an runoff from a nearby pond (Woo and Guan 2006; Boike and others 2008; Abnizova and Young 2010; Young and Abnizova 2011; Abnizova and others 2012a).

Recently, numerous studies have reported on the disappearance or drying of arctic shallow lakes and ponds and associated these hydrologic responses with climate change, in particular warmer temperatures (Yoshikawa and Hinzman 2003; Smith and others 2005; Prowse and others 2006; Riordan and others 2006; Smol and Douglas 2007). It is expected that such climate-induced shifts in hydrology will eventually trigger changes in the overall functioning of these sensitive ecosystems (Helbig and others 2013) altering biochemical composition and seasonal variability in nutrients and carbon dynamics, particularly in the ponds and lakes.

Northern wetlands are considered long-term sinks of atmospheric carbon (Heikkinen and others 2002) storing approximately 455 Pg of carbon (Gorham 1991), and Arctic ponds and lakes are active carbon processors and conduits of greenhouse gases (carbon dioxide and methane) to the atmosphere. Their supersaturation in carbon dioxide (Kling and others 1992; Hamilton and others 1994; Cole and others 2007; Blodau and others 2008; Abnizova and others 2012b) and methane (Walter and others 2006, 2007) is directly associated with their role in the hydrologic and carbon cycles in northern wetlands which links the terrestrial with aquatic environments (Cole and others 1994; Koretailainen and others 2006; Smith and others 2007; Dinsmore and others 2010). The sources for CO_2 supersaturation in ponds and lakes include terrestrial hydrologic inputs of CO_2-rich hydrologic inputs resulted from DOC mineralization and soil respiration processes, dissolved inorganic carbon (DIC) that results from dissolution of carbonates (Hope and others 2004; Tranvik and others 2009), and inputs of dissolved organic carbon (DOC), which is mineralized in water bodies (Striegl and others 2001). Recently, a number of studies reported that arctic ponds may experience summertime undersaturation in CO_2 associated with macrophyte productivity (Tank and others 2009) and photosynthetic process in benthic cyanobacteria mats (Laurion and others 2010).

A global survey of inland waters by Cole and others (2007) showed that 80% of terrestrial carbon inputs in ponds and lakes are returned to the atmosphere through gas exchange making these environments important in regional carbon balances. Despite their high spatial abundance (2.0 % of the global lake coverage of land surface area - Lehner and Doll 2004), key questions associated with hydrological processes and quantification of GHG emissions in these aquatic ecosystems still remain unanswered (Zhang and others 2009), challenging our ability to predict how northern wetlands will sequester and store carbon in the near future. Given the importance of wetland waters for facilitating the transport of terrestrial carbon to its hydrologic and atmospheric forms, a thorough understanding of the interactions between hydrological and carbon fluxes in shallow water bodies is necessary to help identify the possibility of a shift in these wetlands from being sinks to sources of carbon to the atmosphere as a response to climate change.

An important control on aquatic carbon fluxes in northern wetlands is the degree of hydrologic connectivity between a pond/lake and its catchment during snowmelt - the time period which contributes to the bulk of total runoff in Arctic watersheds (Woo 1976). Snowmelt is also considered to be a dynamic season for soil biogeochemistry triggered by this rapid melt, ground thaw and warmer air temperatures above 0°C (Buckeridge and others 2010). It has been documented that a significant portion or all of the nutrients and carbon in spring runoff may be exported to aquatic systems especially when overland flow prevails as a result of the melting snowpack, and poor drainage from a thawing active layer (Michaelson and others 1998; Buckeridge and Grogan, 2010; Townsend-Small and others 2011). In particular, the spring freshet transports a large amount of dissolved organic matter which becomes available as a result of freeze-thaw processes which release carbon from near surface soil organic layers (Michaelson and others 1998).

To date, only a limited number of studies document carbon input into arctic inland waters during the spring melt. Kling (1995) showed that the snowmelt influx of dissolved organic matter to Toolik Lake made for a readily available carbon source for bacterial communities' metabolic processes. Michaelson and others (1998) measured dissolved organic carbon concentrations in inlet waters to Toolik Lake and showed that the highest DOC concentrations were found in the soil water during early thaw which was consequently reduced during transport in hydrologic pathways to streams and lakes mainly as a result of dilution, but also from selective sorption, precipitation, and microbial decomposition (Michaelson and others 1998). While these studies have focused on the quality and quantity of transported terrestrial carbon during spring thaw, a complete temporal pattern of carbon, including its type and fate remains poorly investigated in the Arctic, especially during the short snowmelt season.

Recent research in temperate and boreal environments has revealed that temporal snowmelt DOC concentrations of lakes are well correlated with the proportion of wetlands in the watershed (McEachern and others 2010). This suggests that in-lake DOC concentrations are a useful index of watershed influence (Gergel and others 1999). However, while Medeiros and others (2012) and Lehnherr and others (2012) and many others (Abnizova and others 2012a and references herein) measured DOC concentrations temporally in arctic ponds during summer periods and alluded to the fact that some ponds had signatures of systems with high inputs of terrestrial carbon likely due to water inputs from the wetland catchment, the spring variability of DOC concentrations in High Arctic ponds remains unexplored. Hence, a better understanding of the temporal variability in pond carbon fluxes utilizing an intensive monitoring scheme is needed, specifically during the short snowmelt season.

In this study, we examine hydrology and carbon dynamics: carbon concentrations and storages in ponds situated in an extensive, low-gradient wetland at Polar Bear Pass, Bathurst Island (75°40'N, 98°30'W). Previously, Abnizova and others (2012a) examined seasonal dynamics of in-pond DOC and DIC pools at this location and concluded that differences in hydrologic connectivity from snowmelt to freeze-up affected the delivery and stores of carbon in ponds. Their research was based on a weekly monitoring strategy. Instead, we report on the effects of hydrologic connectivity on a series of ponds' water balance and DOC and DIC mass balances during the entire snowmelt period. This research is grounded on a daily monitoring interval. We also synthesise existing study findings to determine the importance of hydrologic controls on DOC and DIC stores in arctic ponds.

The main objectives of this study were 1) to intensively measure water budgets and carbon stores in arctic wetland ponds during the period of snowmelt; 2) to evaluate how differences in hydrologic connectivity of a range of pond types (isolated/linked) with their catchment affect DOC and DIC pools; and 3) to identify spatial snowmelt patterns in CO_2 evasion in the larger wetland region. The study shows in detail pond water budgets in response to hydrologic connectivity to their catchments and the transition in carbon mass balance components during snowmelt.

Study description and methods

The study area was located at Polar Bear Pass, Bathurst Island Nunavut (Figure 1). It is considered one of the most important wetlands in the Queen Elizabeth Islands, and besides being a haven for migratory birds, polar bears, along with muskox and caribou, local Inuit often frequent the area to hunt. This wetland is approximately 94 km^2 in area and is situated in a broad valley, which is occupied by two large lakes and a multitude of tundra ponds separated by the raised rims of ice-wedge polygons and sedge meadows (Woo and Young 2006). The wetland which runs east to west across the central part of the island is bounded by hillslopes rising 100 to 200 meters above sea level (m. a.s.l.). They are incised by numerous v-shaped stream valleys which along with capturing much seasonal snow drain the upland areas of its snowmelt, recharging the low-lying wetland on a seasonal basis (Young and others 2010).

To justify the selection of the study pond sites at PBP, field reconnaissance in 2007 allowed an initial assessment of their variability including pond size/area, water depth and ground thaw. Several water quality parameters (pH, conductivity) were also measured. Polar bears frequently move through the area, so for safety and other logistical concerns only ponds that were readily accessible (within 2 km from base camp) and representative of other ponds in the region were considered for detailed monitoring.

We selected and examined one large and one small pond both having well defined hydrologic linkages and which were located in fine grained/ice-rich terrain (Figure 1, Table 1). The large pond which was supplied by meltwater from an upslope late-lying snowbed is referred to as *Snowbed Pond* and the small pond receiving runoff from an ephemeral hillslope creek is called *Creek Pond*. We also selected one small pond which had no well-defined connections to its surrounding landscape aside from seasonal snowmelt inputs and rainfall and refer to it as *Isolated Pond*. It is located in fine grained/ice-rich terrain (see Figure 1 and Table 2 in Abnizova and others 2012a). To highlight the importance of limited hydrologic connectivity to a pond's hydrologic regime one small pond was selected having an intermittent link to an ephemeral shallow pond located upslope from it. It is here referred to as *Temporarily Connected Pond*, and is found in coarse-textured/ice poor terrain (see Figure 1 and Table 2 in Abnizova and others 2012a). These ponds will also be referred to as *Control Ponds* due to the intensive monitoring of them in this study.

To better understand the hydrology and physico-chemical characteristics of other ponds at PBP we also selected a range of ponds located within a 2 km radius of these *Control Ponds* and refer to them as *North Control Ponds* and *Central Control Ponds, or 'Other Control Ponds'*. These ponds were only sampled 2-3 times over the melt season. Finally, to ensure that our results were applicable to ponds situated across the entire Pass, a helicopter survey was conducted once during the season. Here, ponds in the east, south-east, south-central, south-west, and western areas of the wetland were sampled. They are referred to as *North East, South Central, South West,* and *Central West Ponds, or Satellite Ponds*.

Field methods and analysis

Hydroclimotology

Evaluating the water budget of arctic ponds often provides a good understanding of how water is moving into a pond from its catchment and whether it will be stored in the pond or flushed out of it. It is also possible to determine if the pond is losing water to evaporation or losses are seeping instead into a thawing substrate. These seasonal changes in water delivery and volume for a pond will of course impact nutrient and carbon stores.

The water balance for each study pond was assessed from June 10 to July 31 2010 and is defined as:

$$\Delta S = P_{(Sn + R)} - E \pm Q_{(Qsur+Qsub)} \tag{1}$$

where ΔS is the change in the pond's volume of water, Sn and R is precipitation input of snowfall (Sn) and rainfall (R), E is evaporation output with negative values of E implying condensation. Q is inflow/outflow to (from) the pond including both surface (Q_{sur}) and subsurface waters (Q_{sub}).

Detailed snow surveys (Sn) which includes the measurements of snow depth and density were conducted at each study pond and adjoining catchment following the approach by Woo (1997). Nearby late-lying snowbeds and several upland stream catchments were also surveyed. These end-of the year snow surveys, along with direct and modelled snowmelt measurements are described in fully in Abnizova and others (2012a) and will not be reported further here. Rainfall (R) was recorded with a HOBO tipping bucket raingauge (± 0.25 mm) at the centrally located wet meadow automatic weather station (see below). To monitor changes in pond water tables in ponds and in their catchments, perforated and screened 5 cm diameter water table wells were installed down to the permafrost table in August 2007. In 2010, pond water tables and temperature were monitored continuously with HOBO pressure transducers (HOBO U20 Water Level Data Logger, ± 0.5 cm), and manually twice weekly at the centre wells with water level sensors (± 0.2 cm). Water levels in catchment wells were manually monitored on a daily basis during the snowmelt period and less intensively during the early post-snowmelt period. Likewise, ground thaw (± 1 cm) was measured daily here using a metal rod inserted into the ground until impeded by frozen ground. Towards the end of the snowmelt season, measurements were made every few days.

Evaporation (E) at the study ponds was calculated using the Priestley and Taylor approach (1972) following the approach by Woo and Guan (2006). The α variable was equal to 1.26 which is appropriate for saturated surfaces in these environments (Stewart and Rouse 1977). The water heat flux to the pond required for evaporation estimates was calculated following the approach by Woo and Guan (2006), with instrumentation fully described in Abnizova and others (2012a). Volumes of ponds were calculated using established relationships between water depth and surface area, which were intensively measured during the 2008 and 2009 field seasons (see Abnizova and others 2012a). To estimate magnitude of lateral inflows and outflows, changes in pond water storage (ΔS_{meas}) were calculated as daily changes in pond volumes and computed to estimated water storage of (ΔS_{est} = ((Sn+R) – E). A similar approach was previsouly used by Woo and Guan (2006) and Abnizova and others (2012a).

Components required in water balance calculations, namely snowmelt and evaporation were obtained from an automated weather system (AWS) established in the wet meadow area of the wetland. Here, hourly values of air temperature (°C) and relative humidity (%) using a CSI-HC2-S3-L temperature and relative humidity probe (± 0.2 °C) were made and logged with a Campbell Scientific CR21X. Measurements were scanned every 60 sec and averaged over the hour. Net radiation (W m^{-2}) was measured using a NR lite radiometer (± 1 %), and incoming and outgoing shortwave radiation (W m^{-2}) was determined with Eppley pyranometers (± 1 %). A Davis anemometer (± 1 %) monitored wind speed (m s^{-1}), and precipitation (mm) was recorded with a HOBO tipping bucket raingauge (± 0.25 mm). Net radiation over the pond was not measured in 2010, so the relationship between wet meadow and pond net radiation from 2009 was used to estimate 2010 pond net radiation values.

Stream discharge (L s^{-1}) from the Landing Strip Creek (LSCR) was monitored following the approach by (Young and others 2010) using a HOBO pressure transducer and Ecotone (Remote Data Systems Inc., ± 2.54 mm) for continuous hourly water level recording and an Ott C2 current meter (± 1 %) to measure stream velocity four times daily at the gauging station established at the beginning of the season. Two stage-discharge curves for LSCR were developed as a result of the different snowcover conditions in the channel, and are as follows: $Q = 0.0129H^{2.3894}$until June 20 and $Q = 1.310H^{1.7551}$, after June 20.

Water chemistry

Water samples (60-100 ml) were taken from the control ponds, inflow into the ponds from the upslope melting late-lying snowbed and the hillslope creek (LSCR) every day or every second day during the study period to assess concentrations of dissolved inorganic carbon (DIC), organic carbon (DOC), cations, and anions. Estimates of pH (± 0.2), conductivity (± 0.001 mS cm^{-1}), water temperature (± 0.15 °C), and dissolved oxygen (± 0.01 mg L^{-1}) were also obtained with a handheld portable YSI unit (YSI 600QS, YSI) at the time of sampling. During the entire snowmelt period which lasted from 17 June to 29 June, the 12 *Other Control Ponds,* specifically ones located 2 km from our intensively studied ponds were sampled two to three times. Additionally, on June 19, 2010, a helicopter survey of the PBP wetland region allowed us to sample 24 *Satellite Ponds*. Similar to our 'intensive' pond sites pond water tables, substrate thaw, and various water quality constituents (pH, temperature, conductivity, and dissolved oxygen) were monitored. Water samples for DOC, DIC and anion-cation analyses were also obtained from these sites following the same sampling protocol mentioned above (Figure 1).

At the time of water collection in the field, all sampling bottles were rinsed 3 times with filtered sample water. Water samples for cation and anion analysis were then filtered using cellulose-acetate filters (0.45 μm pore diameter) and pre-cleaned poly-ethylene-syringes, and then stored in 15 ml HDPE plastic containers. Samples retained for analysis of cation concentrations were acidified with 100-200 μL (12 N) HNO_3. DOC samples were filtered using GF filters (0.7 μm pore diameter) and a PE-syringe, and were collected in 30 ml high-density polyethylene plastic containers. The sample was then acidified to pH 2 by adding 2M HCL and stored in darkness at 4°C until analysis in laboratory environment. Unfiltered water samples for analysis of DIC concentrations were collected in headspace-free, sealed glass vials and kept in a cold, dark environment until laboratory analysis (Abnizova and others 2012b). The preservation method was further tested in the laboratory environment to identify the extent of alteration in DIC concentration as a result of degradation and oxidation of organic carbon in water samples during transportation period. The experiment showed small changes in DIC (8.53 %) in the tested samples compared to their initial concentrations at the time of sampling.

Anion analysis was performed on an IC Dionex DX 320, and cation analysis was made on an ICP-OES Perkin-Elmer Optima 3000xl. DOC concentrations were determined in the filtrate by high temperature catalytic oxidation with a Shimadzu TOC-VCPN analyser. Concentrations of DIC were analyzed using a gas chromatograph for simultaneous detection of CO_2. The Shimadzu GC-2014AF gas chromatograph was equipped with an AOC-5000 autosampler, a 1 m × 1/8" HayeSep Q 80/100-mesh column, an electron capture detector, and two flame ionization detectors. Gas concentrations and total pressure were analyzed in the headspace after shaking solutions at 90°C for 20 minutes on a rotary shaker. The dissolved concentrations of CO_2 (+ H_2CO_3) required for calculating the gaseous emissions were derived from the headspace DIC concentrations and temperature-corrected pH values (Ben-Yaakov 1970), determined from the gas-tight vials in the laboratory, using temperature-adjusted dissociation constants (Plummer and Busenberg 1982).

Stable isotopes of oxygen and hydrogen (δD, $\delta^{18}O$) in pond surface waters were also determined. Here, an equilibration technique using a mass-spectrometer (FinniganMAT Delta-S) was used (Meyer and others 2000).

Carbon mass balance calculation

To better understand carbon dynamics during snowmelt, the mass balance of organic and inorganic carbon, or the total carbon pool based on water budget estimates in the *Snowbed Pond* was quantified (Åberg and others 2004). Specifically, DOC and DIC stocks in this pond were estimated from concentrations of carbon (mg m^{-3}) and pond volume in m^3. Estimates of mass balance and hydrological retention of these carbon constituents in this particular pond was based on the high frequency of sampling here in comparison to other ponds.

Carbon retention was calculated from the difference between carbon loadings in hydrologic inflows and outflows. Using a previously developed relationship in pond volumetric change in relation to hydrologic inflow from the nearby late-lying snowbed, maximum storage capacity of the pond was estimated. Changes in surface area and volume remain linear (Area = 3.1xVolume + 2943.7, R^2=0.996, $p<0.05$) until reaching a threshold of 1600 m^3 which results in overflow. Since no direct measurements of surface inflow and outflow were obtained in 2010, maximum storage capacity was used to estimate inflows and outflows. No precipitation was recorded during the study period in 2010 and retention (R) in 2010 was calculated using the following equation:

$$R = C_{inflow}V_{inflow} + C_{outflow}V_{outfow} - \text{gaseous emission} \quad (2),$$

Gaseous emissions were calculated from the local wind speed from the AWS and surface water gas concentrations using the two-layer model of transfer across an air-water interface (Liss and Slater 1974) following the approach by Repo and others (2007):

$$\text{Flux} = k_{gas}\,(c_{sur} - c_{atm}) \quad (3),$$

where c_{sur} is surface water gas concentration (mol L^{-1}), c_{atm} is gas concentration in equilibrium with the atmosphere (mol L^{-1}), and k_{gas} is the gas exchange constant (cm h^{-1}). To calculate k_{gas} the method used by Jahne and others (1987) was followed:

$$k_{gas} = k_{600}\,(Sc/600)^{x} \qquad (4),$$

where k_{600} is the transfer velocity adjusted to Sc= 600 (the ratio of kinematic viscosity of water and the diffusion coefficient), x = -0.67 when $u_1 < 3$ m s^{-1}, or x = -0.5 when $u_1 > 3$ m s^{-1} (Crisius and Wanninkhof 2003) and Sc was calculated for CO_2 using the method by Wanninkhof (1992) for pond temperatures at the time of sampling.

To calculate k_{600} the power relationship from Cole and Caraco (1998) was used:

$$k_{600} = 2.07 + 0.215 \times u_{10}^{1.7} \qquad (5),$$

The average wind speed 10 m above the water surface (u_{10}) on the sampling day was calculated using the formula by Crisius and Wanninkhof (2003) developed for lake surfaces:

$$u_{10} = u_1\,(1+C_{d10}^{0.5}/k \times \ln(10/1)) \qquad (6),$$

where u_{10} is the wind speed at 10 m height, u_1 is the wind speed at 1 m height, C_{d10} is the drag coefficient at 10 m and k is the von Karman's constant under assumption of a neutrally stable boundary layer. This approach was successfully applied in other arctic and subarctic ponds (Repo and others 2007). Cole and others (2010) compared multiple approaches estimating air–water gas exchange in small lakes and showed that this method showed reliable results when direct measurements are not possible.

Results

Climate

Climatic conditions during 2010 at PBP are shown in Figure 2. In 2010, air temperature at PBP fluctuated from -5 to -2 °C prior to snowmelt (June 1). A continuous rise in air temperature was recorded and eventually temperatures reached above 0°C by June 9 (see Figure 2). Average daily air temperatures in June were influenced by the net radiation regime and levels increased, remaining relatively high throughout the snowmelt season (June 1 – 23), as a result of clear conditions (Figure 2). June average air temperature was only 2.7 °C though. Relative humidity declined to 60 % at the end of snowmelt, reflecting warmer air temperatures, and dry, clear conditions (Figure 2). Wind speeds ranged from 1.6 to 6.6 m s^{-1}, but occasionally they gusted to > 8.1 m s^{-1}. As previously mentioned, average wind speed during the melt season was 3 m s^{-1} at the height of 2 m.

Snowcover and melt

Snowmelt remains a dominant hydrologic input of water into arctic ponds. Table 1 outlines the range in snow depth, snow density, and snow water equivalent of the study ponds and provides an indication of the variability of SWE (mm) between pond catchments. SWE values differed amongst the ponds, with variable snow depths accounting for most of this difference. *Snowbed Pond* and *Isolated Pond* captured high amounts of snow (69 and 73 mm, respectively), whereas the *Temporarily Connected Pond* had the least SWE (33 mm). Creek Pond had a SWE intermediate to these two extremes (56 mm). The former sites were located in a sheltered locale, which captured more snow and delayed melt (Table 1).

At PBP, pond snowmelt, in terms of its timing, rate of melt and duration is modified by catchment melt water inputs subsequent to initial snow coverage and variable climatic conditions. An example of this spatial melt pattern (early, mid, and late period) is provided in Figure 3. In 2010, snowmelt commenced on 1 June. While all the *Study Ponds* were mostly snow-covered before 15 June (Figure 3), a dramatic inflow of meltwaters from the surrounding catchment led to a reduced snow cover on them and they were generally snowfree by 20 June (Figure 3).

The spatial pattern of snowmelt helps define the release of water for runoff, and initiation of ground thaw, and evaporation loss. It is important to note that while the ponds' snowcover was completely reduced by the inflow of warmer catchment waters, ice lenses were observed and remained in the ponds until 22 June. Overall, the melt period lasted for about 16 days at all sites except for *Snowbed Pond,* which received meltwaters from the lingering late-lying snowbed until 29 June.

Hydrology

The ponds with strong hydrologic connectivity (e.g. late-lying snowbed or hillslope creek) demonstrated good seasonal stability in their water levels during the melt season (Figure 4). Observations of water table dynamics at the *Snowbed and Creek Ponds* showed generally steady levels throughout the 2010 season. For instance, water tables in the *Snowbed Pond* remained elevated and declined only when the residual snowbed shrunk in size (June 29, Figure 3, 4) clearly demonstrating its dependence on this particular hydrological linkage. Water level stability at the *Creek Pond* was attributed to its linkage to the hillslope stream during both snowmelt and post-snowmelt seasons (Figure 5d, 6b). Initially, the pond captured much snow due to its basin area (snow depth = 211 mm) but after snowmelt, pond water tables remained steady owing to continuous inputs of water from the creek (Figure 4d). This differed from the *Isolated* and *Temporarily Connected Ponds* where water levels declined sharply following the main melt season. They fell below 100 mm by July 31 during the post-snowmelt period in spite of much rain (25.3 mm).

In terms of the pond water budgets, during the melt period, measured water storage ΔS_{meas} showed an increasing trend in both hydrologically connected study ponds reaching 90 mm in the *Snowbed Pond* and 68 mm in the *Creek Pond* (Figure 5). This differed from the other ponds which showed losses: -91 mm in the *Temporarily Connected Pond* and -106 mm in the *Isolated Pond* (see Figure 5).

Estimation of spring water budgets for 2010 showed that losses through evaporation were dominant at all ponds. The *Snowbed* and *Creek Ponds* lost on average 3.8 mm d^{-1}, amounting to 194 and 192 mm from June 10 to July 31. Similar evaporation rates (3.7 mm d^{-1}) were observed at the *Temporarily Connected Pond* and the *Isolated Pond* (3.8 mm d^{-1}). Seasonal totals for the isolated-type ponds similarly reached about 194 mm.

The estimated water storage (ΔS_{est} = ((Sn+R) – E) shows changes in the water budget of a pond as a result of precipitation inputs and evaporation losses and can be used to help identify the presence or absence of hydrologic lateral inflows (Q) when compared to changes in ΔS_{meas}. Our results illustrate that changes in ΔS_{est} and ΔS_{meas} were similar in 2010, reaching a difference of 222 mm and 240 mm in the *Snowbed* and *Creek Ponds* respectively. These sizeable values suggest substantial lateral inflows into both of these ponds. The differences between ΔS_{meas} and ΔS_{est} for the *Temporarily Connected* and *Isolated Ponds* were much smaller (68 and 56 mm respectively), demonstrating that ΔS_{meas} decreased in these ponds primarily in response to evaporation losses (Figure 5).

The isotopic composition of the *Control Ponds* during the snowmelt period in 2010 strongly reflected differences in water balances between hydrologically connected and isolated ponds (Figure 8). The *Snowbed* and *Creek Ponds* showed isotopic depletion and plotted close to the Global Meteorological Water Line (GMWL) and along the isotopic signatures of snow, LSCR, and meltwaters from the late-lying snowbed (Figure 8a). The *Isolated Pond* and *Temporarily Connected Pond* showed values that were more isotopically enriched and plotted away from the GMWL. This separation was also evident in isotopic signatures of the *Other Control Ponds* dividing them into ponds that are more isotopically depleted (hydrologically connected) and more isotopically enriched (isolated) (Figure 8b).

Snowmelt carbon dynamics and stocks

In 2010, DOC concentrations in the four *Control Ponds* (*Snowbed, Creek, Temporarily Connected*, and *Isolated*) during snowmelt ranged from 5.02 to 15.4 mg L^{-1}, with an average ± standard deviation equal to 7.76 ± 2.32 mg L^{-1} (n = 30) and a median of 6.80 mg L^{-1}. Pond DOC concentrations varied over time, with maximum levels generally occurring towards the end of snowmelt. The isolated ponds (*Isolated Pond* and *Temporarily Connected Pond*) exhibited the highest individual and mean seasonal DOC concentrations (15.40 mg L^{-1} and 10.29 mg L^{-1} respectively). The lowest concentrations were found in the *Snowbed Pond* on 20 June (5.02 mg L^{-1}). DOC concentrations in snow sampled prior to the snowmelt on 31 May 31 and 5 June were low with an average of 0.98 ± 0.34 mg L^{-1} (n = 10). DOC concentrations differed for all ponds including *Other Control Ponds* and *Satellite Ponds* with the isolated ponds peaking towards the end of the snowmelt season while peaks in the connected ponds coincided with the onset of snowmelt (Figure 7).

Significant positive correlations were found between DOC concentrations in water draining soils immediately below the snowbed and the *Snowbed Pond* ($R^2 = 0.62$, $P < 0.05$), and in LSCR's outflowing waters and the nearby *Creek Pond* ($R^2 = 0.36$, $P < 0.05$). This hydrologic link was also tested in a neighbor pond located about 100 m from the *Snowbed Pond* site and resulted in a significant positive correlation as well ($R^2 = 0.73$, $P < 0.05$). It too received meltwater from the same large snowbed. Analysis of stable water isotopes confirmed the snowmelt connectivity of the *Snowbed* and *Creek Ponds* to their respectful sources, in contrast to the isolated ponds (*Isolated* and *Temporarily Connected Ponds*) see Figure 8.

Dissolved inorganic carbon concentrations (DIC) were typically at a minimum at the time of snowmelt and then increased through to the end of the observation period (Figure 7). At this time, saturated conditions prevailed throughout the landscape (Figure 3a,b,c). In 2010, DIC concentrations in the four *Control Ponds* (*Snowbed, Creek, Temporarily Connected,* and *Isolated Ponds*) ranged from 9.18 to 45.06 mg L^{-1}, with an average of 23.44 ± 10.49 mg L^{-1} (n = 27) and a median of 21.57 mg L^{-1}. Overall, in all *Control* and *Satellite Ponds,* DIC concentrations varied over time, with maximum levels generally occurring towards the end of snowmelt.

An analysis of the specific mass of carbon within the *Control Ponds* showed differences in DOC and DIC stocks for ones linked to their catchment in comparison to isolated-types (Figure 7). The *Creek Pond* and the *Temporarily Connected Pond* had higher DOC stocks at the onset of snowmelt with a dramatic increase in DOC and DIC stocks in *Creek Pond* during the period of highest discharge in LSCR (Figure 4d). DOC stocks in *Snowbed Pond* were stable during snowmelt. Calculation of carbon mass balance in this pond because of more intensive sampling in comparison to other ponds showed positive stores of DOC and DIC at the end of snowmelt (1.23 g C m^{-2} and 3.34 g C m^{-2}, Figure 10). As a result of hydrologic inputs during snowmelt DOC retention calculated as a mass balance remained positive, and equal to 1.19 g C m^{-2}. Strong CO_2 losses from the pond (3.77 g C m^{-2}) contributed to removal of all available DIC from the inflow resulting in retention values equal to 0.10 g C m^{-2}. Diverging patterns emerged amongst the ponds at the end of snowmelt when hydrologically connected ponds (*Snowbed* and *Creek Ponds*) continued to increase in carbon and isolated ponds showed a decline in both DOC and DIC stocks.

Greenhouse gas fluxes

Most of the *Control Ponds* acted as sources of CO_2 to the atmosphere during snowmelt (Figure 9). Concentrations of CO_2 were only significantly correlated with DOC in the *Snowbed Pond* ($R^2 = 0.42$, $P < 0.06$) with a similar yet less significant relationship for the *Creek Pond* ($R^2 = 0.38$, $P < 0.14$). In the entire data set, net CO_2 influx was found only in the *Snowbed Pond* on June 22, when the surface water CO_2 concentrations in the pond were close to equilibrium with concentrations in the atmosphere and the CO_2 flux was slightly negative (-0.002 g C m^{-2} d^{-1}). The mean CO_2 emissions from the *Isolated* (0.50 ± 0.14 g C m^{-2} d^{-1}) and *Temporarily Connected Ponds* were higher (0.47 ± 0.54 g C m^{-2} d^{-1}) than in the hydrologically connected *Snowbed Pond* (0.30 ± 0.18 g C m^{-2} d^{-1}) and *Creek Pond* (0.25 ± 0.17 g C m^{-2} d^{-1}).

Examination of CO_2 fluxes in the *Other Control Ponds* showed an average value of 0.30 g C m^{-2} d^{-1} and revealed two distinct increases in the emissions; one observed before 19 June and the other after 22 June. This latter 'peak' coincided with end of the snowmelt or the early post-snowmelt period (Figure 9). Snowmelt CO_2 fluxes in the *North Control Ponds* and *Central Control Ponds* shared a similar trend in emissions

(Figure 9) with an average value estimated at 0.30 g C m^{-2} d^{-1} (range from 0.10 to 1.00 g C m^{-2} d^{-1}).

On 19 June 2010, in the middle of the main snowmelt period, the mean CO_2 flux in the aerially surveyed *Satellite Ponds* varied from 0.00-1.00 g C m^{-2} d^{-1} with a mean of 0.40 g C m^{-2} d^{-1}. When these data are plotted on a δD versus $\delta^{18}O$ scatter plot, ponds in the initial stage of snowmelt and/or receiving meltwaters like *the Snowbed* and *Creek Ponds* were located closer to the Global Meteorological Water Line and ponds in a progressed stage of thaw are located away from it (Figure 8c).

Discussion

Hydrology

The differences in hydrologic settings of PBP ponds and hydro-climatic conditions during snowmelt resulted in variable responses in the water balances of the studied ponds. In 2010, low precipitation inputs and clear sky conditions led to a rapid onset of snowmelt. This large influx of snowmelt waters in combination with frozen ground resulted in overland flow and was a primary source of water to the ponds. This was clearly evident in elevated pond water tables and storages. This recharging of ponds during the short snowmelt freshet has been previously reported in this wetland landscape (Abnizova and others 2012a), as well as others (e.g. Woo and Guan 2006). Most snow in the pond basins at PBP were depleted by 18 June, 2010 with only late-lying snowbeds remaining in the lee of hillslopes, marking the end of the snowmelt season.

The latter part of the melt period was when clear differences in the extent of hydrological connectivity of the ponds emerged. Meltwaters from a late-lying snowbed sustained the *Snowbed Pond* with additional waters helping to maintain elevated water levels at a time when other pond water levels were declining (see Figure 3g). Likewise, in the *Creek Pond* water levels remained stable in response to steady water inflow from LSCR, a hillslope creek which drained the upland area (Figure 6b). On the other hand, while the *Temporarily Connected* and *Isolated Pond* both received meltwater inflow at the time of the freshet, (Figure 6 a,c), ultimately, they experienced the lowest water tables, deepest frost tables (*Temporarily Connected*), and warmest water temperatures at the end of this period. In the absence of rain and lateral inflow into these ponds, losses to both ground seepage and evaporation prevailed and led to negative water balances.

Storage decline in the *Temporarily Connected Pond* can be attributed to its rocky substrate (Figure 6f), which when compared to the silty substrate of *Snowbed* and *Creek Ponds* (Figure 6e) would experience a higher thermal and hydraulic conductivity than other pond substrates. These rocky substrates would have enhanced seepage and subsurface flow. In fact, field observations revealed that the *Temporarily Connected Pond* drained into the low-lying Goodsir River channel through a rocky soil layer which would exhibit a high hydraulic conductivity (Figure 6d). Such subsurface hydrologic connections may also act as corridor for carbon loss from a pond. Similar high seepage rates into underlying rocky substrates have been previously documented for similar ponds near Creswell Bay, Somerset Island (Abnizova and Young 2010).

Carbon dynamics during the snowmelt and early post-snowmelt season

Our results show that wetland ponds at PBP have variable DOC and DIC concentrations and stocks during snowmelt. High DOC concentrations in *Control Ponds* (5.0-15.4 mg L^{-1}) during snowmelt are typically found in ponds situated in wetland environments which receive runoff from a catchment rich in organic matter. Abnizova and others (2012a) summarized previously reported data on DOC concentrations in ponds from selected studies in arctic and subarctic regions of Canada and Alaska showing a high variability in DOC concentrations ranging from 0.1–90 mg L^{-1}. Similar findings were reported in a number of recent studies documenting high DOC concentrations in arctic ponds and lakes across the Canadian Arctic. They linked elevated organic carbon values to high inputs of terrestrial carbon, and attributed it to inputs from the surrounding wetland catchment (0.6-21.2 mg L^{-1}, Medeiros and others 2012; 4.7-40.6 mg L^{-1}, Lehnherr and others 2012; 4.8-16.4 mg L^{-1}, Bos and Pellatt 2012; 10-35 mg L^{-1}, McEachern and others 2010). Gergel and others (1999) conducted a survey of boreal and temperate lakes and also suggested that lake DOC concentrations relate directly to wetland extent in the watershed with its variability related to export from surrounding wetlands mediated by the climate, temperature, or hydrologic changes occurring within (Hansen and others 2006).

High pond DIC concentrations in our study suggest that hydrologic inflows into ponds during snowmelt may also transport large amounts of waterborne DIC. The DIC concentrations in ponds reported in this study show elevated values (9.2 to 45.1 mg L^{-1}) in comparison to others (McEachern and others 2010; Bos and Pellatt, 2012; Lehnherr and others 2012; Medeiros and others 2012). This pattern might possibly be attributed to the interaction of surface and subsurface waters with thawed carbonate-rich mineral layers, typical of this calcareous bedrock environment (Abnizova and others 2012a). Field observations showed that the active layer reached 40 cm in the LSCR and late-lying snowbed catchments. This is deeper than the organic layer, depth of which typically reaches only 15 cm from the surface.

End-of-snowmelt elevated DOC and DIC stocks in hydrologically linked ponds highlight the importance of connectivity of a pond to its catchment particularly during periods of little to no precipitation. Towards the end of snowmelt, the effect of evapo-concentration (Benoy and others 2007) is clearly evident in the declining DOC and DIC stocks in the hydrologically isolated ponds. Similar results were found in 2008 and 2009 (Abnizova and others 2012a). Finally, it is possible that increased DOC concentrations in the hydrologically isolated ponds, especially in ones with small volumes could also result from terrestrial DOC-rich inflows at the onset of snowmelt, coupled with reduced losses of waterborne DOC to microbial respiration (Gergel and others 1999).

A tentative carbon budget of the *Snowbed Pond* revealed that CO_2 emissions to the atmosphere exceeded the retention of DOC and was primarily derived from DIC retention from hydrologic inflows into the pond (Figure 10). Therefore most CO_2 emitted to the atmosphere possibly came from DIC accumulated in the pond as a result of hydrologic inflow or mineralized from existing DOC stores available from retention and water column storage. Similar findings were reported by Kratz and others (1997) in northern Wisconsin lakes receiving substantial bicarbonate loadings. Kling and others

(1992) also showed that CO_2 in arctic lakes may originate from DIC inflows because subsurface waters are confined by permafrost to shallow organic rich soils and may accumulate CO_2 produced by plant and microbial respiration. In this area, the organic layer only reaches about 15 cm.

The CO_2 values reported in this study are similar to the range of CO_2 concentrations reported for Bylot Island, located ca.700 km south-east of PBP. Polygon ponds sampled during the summer of 2006 and 2007 ranged in DOC (20.5–172.9 μm L^{-1}) and DIC (20.1–238.2 μm L^{-1}) (Laurion and others 2010). Comparable values were also reported in Western Siberian ponds (36-115 μm L^{-1}) by Repo and others (2007)

Most of the PBP ponds studied acted as sources of CO_2 to the atmosphere during the snowmelt season, with the exception of *Snowbed Pond*, which acted temporarily as a sink of carbon during peak snowmelt. The decline in CO_2 evasion could also be explained by decline in inflow as a result of the snowbed's demise (Figure 3b) which would have depleted the influx of waterborne DIC. While an increase in CO_2 in inland waters has been reportedly linked to mineralization of waterborne DOC transported from terrestrial environments (Cole and others 2007) and the season of decomposition in aquatic habitats which is apparently longer than that of terrestrial habitats even during colder seasons (Coyne and Kelley 1974), the decline in CO_2 evasion during summer time has also been found in northern ponds. Tank and others (2009) explained the decline in CO_2 concentrations in Arctic delta lakes as a result of macrophyte productivity. Laurion and others (2010) measured negative CO_2 fluxes in arctic ponds colonized with benthic microbial mats and explained the undersaturation in CO_2 as a result of benthic cyanobacteria developing active photosynthetic mats. Similar findings were reported by Coyne and Kelly (1974) in Alaskan ponds suggesting the possibility of a phytoplankton bloom. A brief interval of CO_2 undersaturation observed in the *Snowbed Pond* may also have been caused by photosynthetic activities, but cannot yet be proved here due to insufficient evidence.

The highest surface water CO_2 evasion in the study ponds was measured after the onset of snowmelt. When snowmelt occurs, sufficient amounts of nutrients and dissolved carbon are carried into ponds via overland flow and varying hydrologic linkages, influencing ecosystem dynamics (Townsend-Small and others 2011). The labile fraction of DOC originating from the catchment is metabolised in ponds and lakes and becomes a source for high CO_2 emissions (Rautio and others 2011). The connectivity of ponds to their catchments during peak snowmelt (Figure 3) suggests that DOC was available in all ponds and possibly fuelled CO_2 emissions. In a recent study of arctic ponds Karlsson and others (2013) measured a significant amount of CO_2 being stored over the winter in subarctic northern lakes in Sweden, which was then released to the atmosphere at the time of spring ice-thaw. Hanson and others (2006) suggests that the spring ice-thaw in temperate lakes results in rapid degassing of CO_2 and influx of DO as lakes adjust to temperature-dependent equilibrium conditions. This process, however, is not likely for the PBP ponds during peak snowmelt. First, the tundra ponds here freeze completely to the bottom in winter. In the spring, snowmelt runoff delivers warm catchment waters to most ponds. According to our field observations, while warmer terrestrial runoff melts a pond's snowcover, ice-lenses pans are still present during snowmelt onset (Figure 3g).

This physical barrier between the pond substrate and the atmosphere might possibly explain lower than reported CO_2 concentrations at the time of ice-thaw in comparison to other northern ponds that do not freeze to the bottom (Striegl and Michmerhuizen 1998; Karlsson and others 2010). In our study, CO_2 concentrations ranged from 17-173 μM which is smaller than spring concentrations (470-2344 μM) but higher than CO_2 concentrations (25-56 μM) during the ice-free period in subarctic lakes in Sweden (Karlsson and others 2013).

Potential implications for ponds across PBP

Estimated CO_2 emissions from wetland ponds at PBP showed that 39 out of 40 measured ponds were strong sources of carbon to the atmosphere during snowmelt. Temporal variation of CO_2 evasion from *Control Ponds* showed two distinct peaks with the first increase in emissions occurring during the snowmelt (*Stage 1*, Figure 9) and a second peak during the early post-snowmelt period (*Stage 2*, Figure 9). An aerial survey conducted during peak snowmelt revealed that the *Satellite Ponds* exhibited spatial differences in CO_2 emissions across the Pass. We suggest that this pattern might be linked to differences in the timing and duration of snowmelt between the northern and southern sectors of PBP (Assini and Young 2012). According to Assini and Young (2012), differences in snowmelt timing are driven by spatial snow distribution, which in turn is initially controlled by interactions between topography and wind. In the spring of 2009 Assini and Young (2012) reported a delay in snowmelt onset in the southern part of the Pass of 7 days versus northern areas. They also observed that *North-eastern Ponds* also experienced earlier snowmelt onset when compared to other *Northern or Western Ponds* because of a shallower snowpack, again relating this to local wind patterns. Our visual observations along with available pond temperature data in 2010 also showed that ponds situated in the northern sector (*Control Pond*) melted-out earlier than *Southern Satellite Pond(s)* by at least 5 to 6 days (Figure 4e). In addition, the isotopic composition of the *Satellite Ponds* sampled during peak snowmelt showed that most *Satellite Ponds* were isotopically depleted and plotted close to GMWL similar to the linked and isolated *Control Ponds*. This reflects their snow-covered and early-snowmelt conditions. The *North-eastern Ponds* which experienced earlier melt than other *Satellite Ponds* based on our field observations showed signatures similar to the *Snowbed Pond.* This suggests a hydrological connectivity to their catchment, which in this case was also lingering snowbeds in the lee of slopes.

The delay and non-uniformity of melt across other sectors of the PBP wetland in 2010 resulted in two distinct groups of ponds with different CO_2 evasion rates. Separation of emission values based on pond location (northern and southern) and assigning the groups of ponds to the state of snowmelt based on the study by Assini and Young (2012) confirmed that the range of emissions estimated from *Satellite Ponds* fit the temporal trends in emissions from *Control Ponds* in the order of their melting stage (Figure 9). *North-eastern Ponds* melted earlier than the *Control Ponds* and this group was assigned to *Stage 2* of the snowmelt evasion pattern. *Southern Ponds* were still partly snow-covered placing their CO_2 fluxes under *Stage 1* of the snowmelt evasion model. The reported differences in the evasion rates of CO_2 from wetland ponds during this freshet season demonstrates the unique inter-play of topography (snowcover) and microclimatic variability (wind, radiation) on hydrologic processes of melt and runoff and ultimately carbon dynamics in PBP ponds.

To the best of our knowledge, no one has yet identified these stages of carbon emissions in High Arctic ponds during the snowmelt season. While additional years of sampling might be important in replicating these findings, and show the range of emissions that can occur under a range of melt situations (e.g. short vs. prolonged melt), our findings do serve to caution other scientists about how to interpret carbon water samples, especially if they are taken at the time of the spring freshet, and especially from ponds similar to the ones at PBP. Our study also confirms that CO_2 fluxes produced in the ponds at PBP draw their carbon from different sources but predominantly from catchment inputs. The snowmelt emissions estimated in our study (0-1.57 g C m^{-2} d^{-1}) with mean emissions equaling 0.37 g C m^{-2} d^{-1} were slightly lower than the daily summer fluxes reported in arctic ponds in West Siberian lowlands (0.5-2.6 g C m^{-2} d^{-1}) (Repo and others 2007) and similar to the mean seasonal emission value for 25 arctic lakes in Alaska (0.25 ± 0.04 g C m^{-2} d^{-1}, Kling and others 1992).

To compare the CO_2 fluxes obtained in this study with literature values from other circumpolar environments we extrapolated our instantaneous flux values over 60 days in order to provide an estimate of the active season rate. Assuming that ponds in these environments are frozen to the bottom and remain ice covered from September to June, this estimate should represent an annual CO_2 flux from PBP wetland ponds. Our calculations suggest that annual CO_2 fluxes can range from 9.3 to 37.8 g C m^{-2} in these arctic ponds. These estimates are lower than those reported for small boreal lakes (102 g C m^{-2} per year for lakes smaller than 0.1 km^2, Kortelainen and others 2006) and wetland ponds in the Hudson Bay lowland (1350–4015 g C m^{-2} per year, Hamilton and others 1994). However, emissions from our study ponds were higher than those estimated from a large series of boreal lakes in Sweden (0.6-5.1 to g C m^{-2} per year, Algesten and others 2004) or from arctic lakes and rivers in Alaska (24 g C m^{-2} per year, Kling and others 1992). Based on total area of pond coverage at PBP using Terra-SARX imagery (11.38 km^2, S. Muster, pers. Communication, March 15, 2013) our CO_2 flux estimates from PBP ponds can range from -0.14 to 65.69 T of CO_2 per day at the time of snowmelt.

Conclusions

In this study, we examined the interactions between snowmelt variability and carbon dynamics in tundra ponds experiencing a varying range of hydrologic linkages with their surrounding catchment, some strong to weak, or non-existent. In 2010, carbon dynamics in these wetland ponds was controlled by three major factors:

1) ponds with strong hydrological linkages to their catchment show higher DOC and DIC stores than ponds connected to their catchment only during snowmelt runoff because they do not receive water carbon inputs after being de-coupled from the catchment. Hydrologically connected ponds continue receiving waterborne DOC and DIC which elevates their carbon stocks toward the end of the snowmelt;

2) Pond CO_2 emissions showed two distinct peaks during snowmelt and these were strongly controlled by hydrologic inputs of terrestrial DOC and DIC. Early

snowmelt emissions were likely driven by DOC inputs, and the later emissions fuelled by increased DIC inputs as a result of surface and subsurface water interacting with thawed carbonate-rich catchment soils;

3) an aerial survey of ponds across PBP during peak snowmelt illustrated that while most ponds were strong sources of CO_2 to the atmosphere, the magnitude of the carbon flux was strongly dependent on the spatial pattern of melt in the wetland with southern ponds melting later and therefore showing higher CO_2 emissions than northern ponds.

The high carbon emissions from ponds at PBP demonstrate the importance of ponds as active conduits of terrestrial carbon to the atmosphere. This influences carbon balances of northern wetlands, which typically possess a high number of small and shallow water bodies. Evaluation of seasonal dynamics in DOC and DIC components in arctic ponds, particularly during snowmelt dramatically improves our knowledge of the processes driving carbon balance variability in aquatic ecosystems at this time of the year.

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References

Åberg J, Bergström AK, Algesten G, Söderback K, Jansson, M. 2004. A comparison of the carbon balances of a natural lake (L. Örträsket) and a hydroelectric reservoir (L. Skinnmuddselet) in northern Sweden. Water Res 38: 531-538, doi:10.1016/j.watres.2003.10.035.

Abnizova A, Young KL. 2010. Sustainability of arctic ponds in a polar desert environment. Arctic 63: 67-84.

Abnizova A, Young KL, Lafrenière MJ. 2012a. Pond hydrology and dissolved carbon dynamics at Polar Bear Pass wetland, Bathurst Island, Nunavut. Ecohydrol doi: 10.1002/eco.1323

Abnizova A, Siemens J, Langer M, Boike J. 2012b. Small ponds with major impact: The relevance of ponds and lakes in permafrost landscapes to carbon dioxide emissions. Global Biogeochem Cycles 26: GB2041.

Algesten G, Sobek S, Bergström AK, Ågren A, Tranvik LJ, Jansson M. 2004. Role of lakes for organic carbon cycling in the boreal zone. Glob Change Biol 10:141-147.

Assini J. Young KL. 2012. Snow cover and snowmelt of an extensive High Arctic wetland: spatial and temporal seasonal patterns. Hydrolog Sci J 57(4):738-755.

Benoy G, Cash K, McCauley E, Wrona F. 2007. Carbon dynamics in lakes of the boreal forest under a changing climate. Environ. Rev. 15: 175–189.

Ben-Yaakov S. 1970. A method for calculating the in situ pH of seawater. Limnol. Oceanogr.15: 326–328.

Blodau C, Rees R, Flessa H, Rodionov A, Guggenberger G, Knorr K-H, Shibistova O, Zrazhevskaya G, Mikheeva N, Kasansky OA. 2008. A snapshot of CO_2 and CH_4 evolution in a thermokarst pond near Igarka, northern Siberia. J Geophys Res 113, doi:10.1029/2007JG000652.C

Boike J, Wille C, Abnizova A. 2008. Climatology and summer energy and water balance of polygonal tundra in the Lena River Delta, Siberia. J Geophys Res 113: G03025 doi:10.1029/2007JG000540

Bos DG, Pellatt MG. 2012. The water chemistry of shallow ponds around Wapusk National Park of Canada, Hudson Bay Lowlands. Can Water Resources J, 37:163-175.

Buckeridge KM, Cen YP, Layzell DB, Grogan P. 2010. Soil biogeochemistry during the early spring in low arctic mesic tundra and the impacts of deepened snow and enhanced nitrogen availability. Biogeochemistry 99: 127–141.

Buckeridge KM, Grogan P. 2010. Deepened snow increases late thaw biogeochemical pulses in mesic low arctic tundra. Biogeochemistry 101:105–121.

Cole JJ, Caraco NF, Kling GW, Kratz TK. 1994. Carbon dioxide supersaturation in the surface waters of lakes. Science 265: 1568-1570.

Cole JJ, N. F. Caraco NF. 1998. Atmospheric exchange of carbon dioxide in a low-wind oligotrophic lake measured by the addition of SF6. Limnol Oceanogr 43: 647–656.

Cole JJ, Prairie, Caraco NF, McDowell WH, Tranvik LJ, Striegl RG, Duarte CM, Kortelainen P, Downing JA, Middleburg JJ, Melack J. 2007. Plumbing the global carbon cycle: Integrating inland waters into the terrestrial carbon budget. Ecosystems 10: 171–184.

Cole JJ, Bade DL, Bastviken D, Pace ML, Van de Bogert M. 2010. Multiple approaches to estimating air-water gas exchange in small lakes. Limnol Oceanogr Methods 8:285-293.

Coyne PI, Kelley JJ. 1974. Carbon dioxide partial pressures in arctic surface waters. Limnol Oceanogr 19(6): 928-938.

Crusius J, Wanninkhof R. 2003. Gas transfer velocities measured at low wind speed over a lake. Limnol Oceanogr 48: 1010-1017.

Dinsmore KJ, Billett MF, Skiba UM, Rees RM, Drewer J, Helfter C. 2010. Role of the aquatic pathway in the carbon and greenhouse gas budgets of a peatland catchment. Glob Change Biol 16: 2750-2762.

Downing JA, Prairie YT, Cole JJ, Duarte CM, Tranvik LJ, Striegl RG, McDowell WH, Kortelainen P, Caraco NF, Melack JM, Middelburg JJ. 2006. The global abundance and size distribution of lakes, ponds, and impoundments. Limnol Oceanogr 51: 2388-2397.

Gergel SE, Turner MG, Kratz TK. 1999. Scale-dependent landscape effects on north temperate lakes and rivers. Ecological Applications 9: 1377–1390.

Gorham E. 1991. Northern peatlands: Role in the carbon cycle and probable responses to climatic warming. Ecol Appl 1: 182–195.

Hamilton JD, Kelly CA, Rudd JWM, Hesslein RH, Roulet NT. 1994. Flux to the atmosphere of CH_4 and CO_2 from wetland ponds on the Hudson-Bay lowlands (HBLs). J Geophys Res 99: 1495–1510.
Hanson P, Carpenter S, Armstrong DE, Stanley EH, Kratz TK. 2006. Lake dissolved inorganic carbon and dissolved oxygen: changing drivers from days to decades. Ecological Monographs 76:343-363.
Helbig M, Boike J, Langer M, Schreiber P, Runkle BRK, Kutzbach L. 2013. Spatial and seasonal variability of polygonal tundra water balance: Lena River Delta, northern Siberia (Russia). Hydrogeol J 21: 133-147.
Heikkinen JEP, Maljanen M, Aurela M, Hargreaves KJ, Martikainen PJ. 2002. Carbon dioxide and methane dynamics in a sub-Arctic peatland in northern Finland. Polar Res 21(1): 49-62.
Hope D. Palmer SM. Billett MF. Dawson JJC. 2004. Variations in dissolved CO_2 and CH_4 in a first-order stream and catchment: an investigation of soil–stream linkages. Hydrol Process 18: 3255–3275. doi: 10.1002/hyp.5657
Jähne B, Münnich KO, Bosinger R, Dutzi A, Huber W, Libner P. 1987b. On the parameters influencing air-water gas exchange. J of Geophys Res 92: 1937-1949.
Karlsson J. Giesler R. Persson J. Lundin E. 2013. High emission of carbon dioxide and methane during ice thaw in high latitude lakes. Geophys Res Lett 40, doi:10.1002/grl.50152.
Kling GW, Kipphut GW, Miller MC. 1992. The flux of CO_2 and CH_4 from lakes and rivers in arctic Alaska. Hydrobiologia 240: 23-36. doi:10.1007/BF00013449.
Kling GW. Land-water linkages: the influence of terrestrial diversity on aquatic systems. In: Korner C The role of biodiversity in arctic and alpine tundra ecosystems. The role of biodiversity in arctic and alpine tundra ecosystems. Berlin; 1995. p. 297-310.
Kortelainen P, Rantakari M, Huttunen JT, Mattsson T, Alm J, Juutinen S, Larmola T, Silvola J, Martikainen PJ. 2006. Sediment respiration and lake trophic state are important predictors of large CO_2 evasion from small boreal lakes. Glob Change Biol 12: 1554-1567, doi:10.1111/j.1365-2486.2006.01167.x.
Kratz T, Webster K, Bowse C, Maguson J, Benson B. 1997. The influence of landscape position on lakes in northern Wisconsin. Freshwater Biol 37: 209–217. doi: 10.1046/j.1365-2427.1997.00149.x
Laurion I, Vincent WF, Macintyre S, Retamal L, Dupont C, Francus P, Pienitz R. 2010. Variability in greenhouse gas emissions from permafrost thaw ponds. Limnol Oceanogr 55: 115-133.
Lehnherr I, St. Louis VL, Emmerton CA, Barker JD, Kirk JL. 2012. Methylmercury cycling in High Arctic wetland ponds: sources and sinks. Environ Sci Technol 46: 10514-10522.
Lehner B, Doll P. 2004. Development and validation of a global database of lakes, reservoirs and wetlands. J Hydrol 296: 1-22.
Liss PS, Slater PG. 1974. Flux of gases across the air-sea interface. Nature 247: 181-184.
McEachern P. Prepas EE. Gibson JJ. Dinsmore WP. 2010. Forest fire induced impacts on

phosphorus, nitrogen, and chlorophyll a concentrations in boreal subarctic lakes of northern Alberta. Can J Fish Aquat Sci 57(Suppl. 2): 73–81.

Medeiros AS, Biastoch RG, Luszczek CE, Wang XA, Muir DGC, Quinlan R. 2012. Patterns in the limnology of lakes and ponds across multiple local and regional environmental gradients in the eastern Canadian Arctic. Inland Waters. J Int Soc Limnol 21: 59-76.

Meyer, H. L. Schonicke, U. Wand, H. W. Hubberten, H. Friedrichsen, Isotope studies of hydrogen and oxygen in ground ice—experiences with the equilibration technique. Isot Environ Health Stud 36, 133–149 (2000).

Michaelson GJ, Ping CL, Kling GW, Hobbie JE. 1998. The character and bioavailability of dissolved organic matter at thaw and in spring runoff waters of the arctic tundra north slope. J Geophys Res 103: 28939–28946.

National Wetlands Working Group. 1997. The Canadian Wetland Classification System, 2nd Edition. In: Warner BG, Rubec CDA, Eds. Wetlands Research Centre, University of Waterloo, Waterloo, ON, Canada. p 68.

Plummer LN, Busenberg E. 1982. The solubilities of calcite, aragonite and vaterite in CO_2-H_2O solutions between 0 and 90°C, and an evaluation of the aqueous model for the system $CaCO_3$-CO_2-H_2O. Geochimica et Cosmochimica Acta. 46. 1011-1040.

Prowse TD, Wrona FJ, Reist JD, Gibson JJ, Hobbie JE, Lévesque LMJ, Vincent WF. 2006. Climate change effects on hydroecology of arctic freshwater ecosystems. Ambio 35: 347-358.

Rautio M, Dufresne F, Laurion I, Bonilla S, Vincent WF, Christoffersen KS. 2011. Shallow freshwater ecosystems of the circumpolar Arctic. Ecoscience 18: 204-222.

Repo ME, Huttunen JT, Naumov AV, Chichulin AV, Lapshina ED, Bleuten W, Martikainen PJ. 2007. Release of CO_2 and CH_4 from small wetland lakes in western Siberia. Tellus B 59: 788-796, doi:10.1111/j.1600-0889.2007.00301.x.

Riordan B, Verbyla D, McGuire AD. 2006. Shrinking ponds in subarctic Alaska based on 1950-2002 remotely sensed images. JGeophys Res 111: G04002 doi:10.1029/2005JG000150.

Rouse WR, Douglas MSV, Hecky RE, Hershey AE, Kling GW, Lesack L, Marsh P, McDonald M, Nicholson BJ, Roulet NT, Smol JP. 1997. Effects of climate change on the freshwaters of arctic and subarctic North America. Hydrol Process 11: 873-902.

Smith LC, Sheng Y, MacDonald GM, Hinzman LD. 2005. Disappearing Arctic lakes. Science 308: 1429–1429.

Smith LC, Sheng Y, MacDonald GM. 2007. A first pan-Arctic assessment of the influence of glaciation, permafrost, topography and peatlands on northern hemisphere lake distribution. Permafrost Periglac 18: 201-208.

Smol JP, Douglas MSV. 2007. Crossing the final ecological threshold in High Arctic ponds. Proc Natl Acad Sci USA 104: 12395–12397.

Stewart RB, Rouse WR. 1977. Substantiation of the Priestley and Taylor parameter α =

1.26 for potential evaporation in high latitudes. J Appl Meteorol 16(6): 649-650.
Striegl RG, Michmerhuizen CM. 1998. Hydrologic influence on methane and carbon dioxide dynamics at two north-central Minnesota lakes. Limnol Oceanogr 1519-1529.
Striegl RG. Kortelainen P. Chanton JP. Wickland KP. Bugna GC. Rantakari M. 2001. Carbon dioxide partial pressure and 13C content of north temperate and boreal lakes at spring ice melt. Limnol Oceanogr 46(4): 941-945.
Tank SE, Lesack LFW, McQueen DJ. 2009. Elevated pH regulates bacterial carbon cycling in lakes with high photosynthetic activity. Ecology 90: 1910–1922.
Townsend-Small A, McClelland J, Max Holmes R, Bruce P. 2011. Seasonal and hydrologic drivers of dissolved organic matter and nutrients in the upper Kuparuk River, Alaskan Arctic. Biogeochem 103(1):109–124.
Tranvik LJ, Downing JA, Cotner JB, Loiselle SA, Striegl RG, Ballatore TJ, Dillon P, Finlay K, Fortino K, Knoll LB, Kortelainen PL, Kutser T, Larsen S, Laurion I, Leech DM, Mccallister SL, Mcknight DM, Melack JM, Overholt E, Porter JA, Prairie YT, Renwick WH, Roland F, Bradford SS, Schindler DW, Sobek S, Tremblay A, Vanni MJ, Verschoor AM, Wachenfeldt EV, Weyhenmeyer GA. 2009. Lakes and reservoirs as regulators of carbon cycling and climate. Limnol Oceanogr 54 : 2298-2314.
Walter KM, Zimov SA, Chanton JP, Verbyla D, Chapin FS. 2006. Methane bubbling from Siberian thaw lakes as a positive feedback to climate warming. Nature 443: 71–75.
Walter KM, Smith LC, Chapin FS. 2007. Methane bubbling from northern lakes: Present and future contributions to the global methane budget. Phil. Trans. R. Soc. A 365: 1657–1676.
Wanninkhof R. 1992. Relationship between wind speed and gas exchange over the ocean. J Geophys Res 97(C5): 7373–7382 doi:10.1029/92JC00188.
Woo MK. 1976. Hydrology of a small Canadian High Arctic basin during the snowmelt period. Catena 3: 155-168.
Woo MK. 1997. A guide for ground based measurement of arctic snow cover, Canadian Snow Data CD, Meteorological Service of Canada, Downsview, Ontario, 30 pp., 1997.
Woo MK, Guan XJ. 2006. Hydrological connectivity and seasonal storage change of tundra ponds in a polar oasis environment, Canadian High Arctic. Permafrost Periglac 17: 309-323.
Woo MK, Young KL. 2006. High arctic wetlands: Their occurrence, hydrological characteristics and sustainability. J Hydrol 320: 432- 450.
Young KL, Assini J, Abnizova A, De Miranda N. 2010. Hydrology of hillslope-wetland streams, Polar Bear Pass, Nunavut, Canada. Hydrol Process 24: 3345-3358.
Young KL, Abnizova A. 2011. Hydrologic thresholds of ponds in a polar desert wetland environment, Somerset Island, Nunavut, Canada. Wetlands 31: 535-549.

Yoshikawa K, Hinzman LD. 2003. Shrinking thermokarst ponds and groundwater dynamics in discontinuous permafrost near Council, Alaska. Permafrost Periglac 14: 151–160.
Zhuang Q, Melack JM, Zimov SA, Walter KM, Butenhoff CL, Khalil MAK. 2009. Global methane emissions from wetlands, rice paddies, and lakes, Eos Trans. AGU 90: 37 – 38, doi:10.1029/2009EO050001.

Tables

Table 1. Limnological and chemical characteristics as well as concentrations of carbon together with isotopic compositions, for ponds at Polar Bear Pass during the study period (mean values, with ranges and sample size in parentheses). Snow depth, snow density and snow water equivalent (SWE) estimates are based on 13 measurements across each pond.

	Snowbed Pond	Creek Pond	Temporarily Connected Pond	Isolated Pond
Trophic state	Oligotrophic	Oligotrophic	Oligotrophic	Oligotrophic
Terrain type	fine-grained/ice-rich	fine-grained/ice-rich	coarse-textured/ice poor	fine-grained/ice-rich
Mean snow depth (mm)	255.9	211.2	123	271
Mean snow density (g cm^{-3})	0.40	0.30	0.40	0.35
Mean SWE (mm)	69	56	33	73
Volume (m^3)	1452	222	36	72
Surface area (m^2)	7242	918	380	818
Area/Volume	4.99	4.14	10.66	11.39
Maximum seasonal depth (m)	0.38	0.38	0.31	0.22
Maximum seasonal thaw depth (m)	0.65	0.50	1.07	0.56
Mean seasonal water	8.48 (-3.32-18.90)	4.19 (-9.80-24.06)	5.66 (-10.21-20.52, 3)	11.14 (0.00-25.84,

temperature (°C)				3)
Ca (mg L^{-1})	30.8 (26.2–35.7, 3)	18.0 (41.6–55.4, 3)	64.0 (59.0–68.0, 3)	63.3 (52.9–68.3, 3)
Mg (mg L^{-1})	9.6 (8.5–11.0, 3)	9.5 (8.0–10.7, 3)	13.0 (11.2–14.6, 3)	13.5 (12.0–15.6, 3)
Na (mg L^{-1})	3.1 (2.6–3.4, 3)	4.5 (3.7–5.3, 3)	12.7 (10.3–14.9, 3)	8.0 (6.6–10.5, 3)
K (mg L^{-1})	0.4 (0.1–0.9, 3)	0.4 (0.1–1.0, 3)	1.4 (1.3–1.6, 3)	1.9 (1.5–2.2, 3)
Fe (μg L^{-1})	<50 (<50-<50, 3)	38.9 (25.0–66.7, 3)	62.0 (25.0–136.0, 3)	<50 (<50-<50, 3)
Ba (μg L^{-1})	145.0 (125.0-166.0, 3)	197 (148.0-226.0, 3)	210.7 (181.0-242.0, 3)	339.9 (323.0-354.5, 3)
Si (mg L^{-1})	0.6 (0.5–0.8, 3)	1.1 (0.9–1.4, 3)	0.9 (0.4–1.4, 3)	2.5 (2.3–3.0, 3)
Sr (μg L^{-1})	40.9 (35.3–47.0, 3)	83.9 (73.4–99.5, 3)	134.0 (117.0–147.0, 3)	178.4 (160.0–190.0, 3)
Cl (mg L^{-1})	7.6 (7.3–8.2, 3)	10.4 (9.5–11.7, 3)	32.0 (24.9–38.0, 3)	20.4 (16.5–24.7, 3)
SO_4 (mg L^{-1})	18.4 (12.3–22.3, 3)	7.2 (4.9–8.9, 3)	14.6 (13.1–16.1, 3)	2.4 (1.7–3.1, 3)
Alkalinity (mmol L^{-1})	1.7 (1.1–2.0,10)	2.1 (0.9–3.0, 8)	2.9 (1.6–3.8, 8)	3.9 (3.4–4.4, 3)
Conductivity (μS cm^{-1})	133.4 (57.0–214.0, 8)	158.3 (30.0–261.0, 7)	272.9 (158.0–388.0, 7)	221.1 (81.0–588.0, 8)
DO (mg L^{-1})	12.1 (6.0–15.2, 8)	12.2 (5.9–16.0, 8)	11.5 (9.1 –13.5, 8)	11.7 (6.1–16.4, 9)
DOC (mg L^{-1})	6.0 (5.0–6.9, 11)	6.9 (6.1–8.1, 8)	9.4 (8.9 –10.3, 8)	11.9 (7.6–15.4, 3)
DIC(mg L^{-1})	16.6 (10.0–22.8, 9)	22.6 (9.2–31.2, 8)	27.0 (14.0 –45.1, 7)	37.8 (27.3–44.5, 3)
pH	7.85 (7.48–8.15, 9)	8.00 (7.67–8.25, 8)	8.08 (7.56 –8.5, 7)	8.34 (8.28–8.38, 3)
CO_2(mg L^{-1})	0.78 (0.27–1.21, 9)	0.71 (0.46–1.13, 8)	0.76 (0.25 –1.66, 7)	0.45 (0.34–0.53, 3)
$\delta^{18}O$ (‰)	-22.6 (-23.2– -21.9, 2)	-21.3 (-22.0– -19.7, 5)	-18.2 (-19.7– -16.4, 3)	-15.8 (-17.0– -14.6, 2)
δ D(‰)	-177.9 (-181.7– -174.2, 2)	-168.3 (-173.8– -159.1, 5)	-149.3 (-156.4– -140.4, 3)	-136.1 (-142.0– -130.3, 2)

Figures

Figure 1. Topographic map of the study area showing the wetland and wetland basin boundary at Polar Bear Pass, Bathurst Island, Nunavut (75° 40'N, 98° 30'W). *Control Ponds* (C) are shown with white circles, *Other Control Ponds* (C) and *Satellite Ponds* (S) are shown with black circles. Contour lines are spaced at 30-m intervals in the drainage basin area and at 10-m interval in the wetland area.

Figure 2. Incoming shortwave radiation, net radiation, mean daily air temperature, relative humidity, wind speed and daily total precipitation in 2010. Data are plotted using information from the wetland AWS. Snowmelt onset (black solid line), end (grey line) and end of snowbed melt (dashed grey) are indicated.

Figure 3. S*nowbed pond* SWE distribution (a), duration of melt at the pond and the snowbed (b), and process of spatial melt at *Snowbed Pond* over time (c-h).

Figure 4. Seasonal water table (a), sediment thaw (b), and pond water temperatures (d) of the study ponds in 2010. Discharge from LSC is also plotted (c). Pond water temperature from *South Satellite Pond* is shown in comparison to *North Control Pond* and Air Temperature (e). Snowmelt end (grey line) and end of the late-lying snowbed melt (dashed grey) are indicated. Daily values of pond sediment thaw were interpolated using MATLAB 7.1. No frost table data are available for the T*emporarily Isolated Pond.* Rapid frost table descent which prevented accessibility with the 1.5 m long frost rod.

Figure 5. Water balance of study ponds during the snowmelt and early post-snowmelt season in 2010. Cumulative calculated storage change (ΔSest = P–E, Precipitation minus

Evaporation) is shown for comparison. End of snowmelt (grey line) and end of the late-lying snowbed melt (dashed grey) are also indicated.

Figure 6. Field photographs displaying overland flow during peak snowmelt and *Isolated Pond* (a), a hillslope creek and creek waters inflow into *Creek Pond* (b), surface inflow into *Temporarily Isolated Pond* from the upslope pond (c), subsurface outflow from *Temporarily Isolated Pond* to the channel Goodsir River (d), substrate of *Snowbed Pond* (e), substrate of *Temporarily Isolated Pond* (f).

Figure 7. DOC and DIC concentrations of water samples (left panel) collected from the study ponds in 2010. DOC and DIC stocks (g C m^{-2}) (right panel) in the study ponds in 2010. End of snowmelt (gray line) and disappearance of the late-lying snowbed is indicated with a dashed gray line. Based on field observations, the ice-off date was on June 20, 2010.

Figure 8. Isotopic composition ($\delta^{18}O$ and δD) of *Control Ponds,* snow, outflow from a late-lying snowbed, and water samples from LSCR creek during snowmelt in 2010 (a) , *Control Ponds* in association with *Other Control Ponds* (b), and *Satellite Ponds* (c).

Figure 9. CO_2 flux in *Control Ponds* using k derived according to Cole and Caraco (1998) into atmosphere (a), CO_2 flux from the *North Control, Central Control Ponds* presented with single points and the box of emission values from *Satellite Ponds* with frames values between the 25th and 75th percentiles, the horizontal line represents the median, and whiskers show the 10th and 90th percentiles (b). Extreme values are shown as dots. Temporal trend in CO_2 emissions from *Snowbed Pond* and *Temporarily*

Connected Pond are shown for comparison in grey. Distribution of CO_2 emissions from *Satellite Ponds* based on temporal differences in spatial melt at PBP (c).

Figure 10. Schematic of the carbon mass balance of the *Snowbed Pond* from June 16 to June 29, 2010, in g C m^{-2}. The diagram indicates the average stock (average amount during the study period) of dissolved carbon during the study period in g C m^{-2}. Carbon emissions are estimated with the gas exchange coefficient derived according to Cole and Caraco (1998).

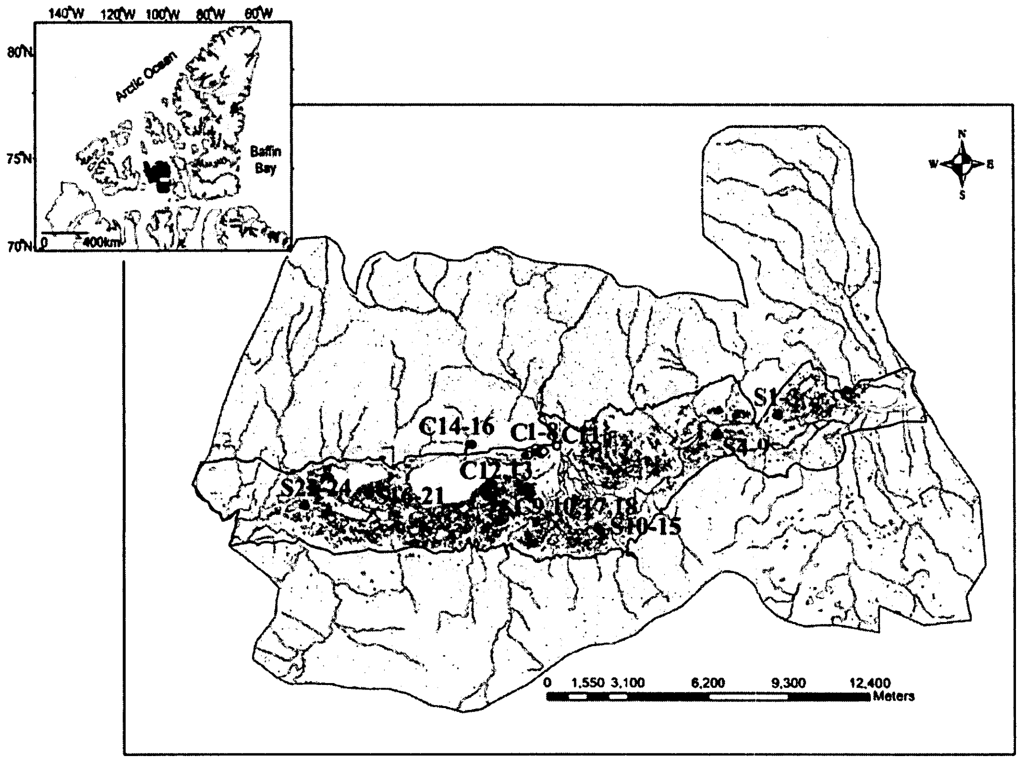

Arctic Ocean
Baffin Bay
400km
C14-16
C12-13
S10-15
0 1,550 3,100 6,200 9,300 12,400
Meters

Figure 1

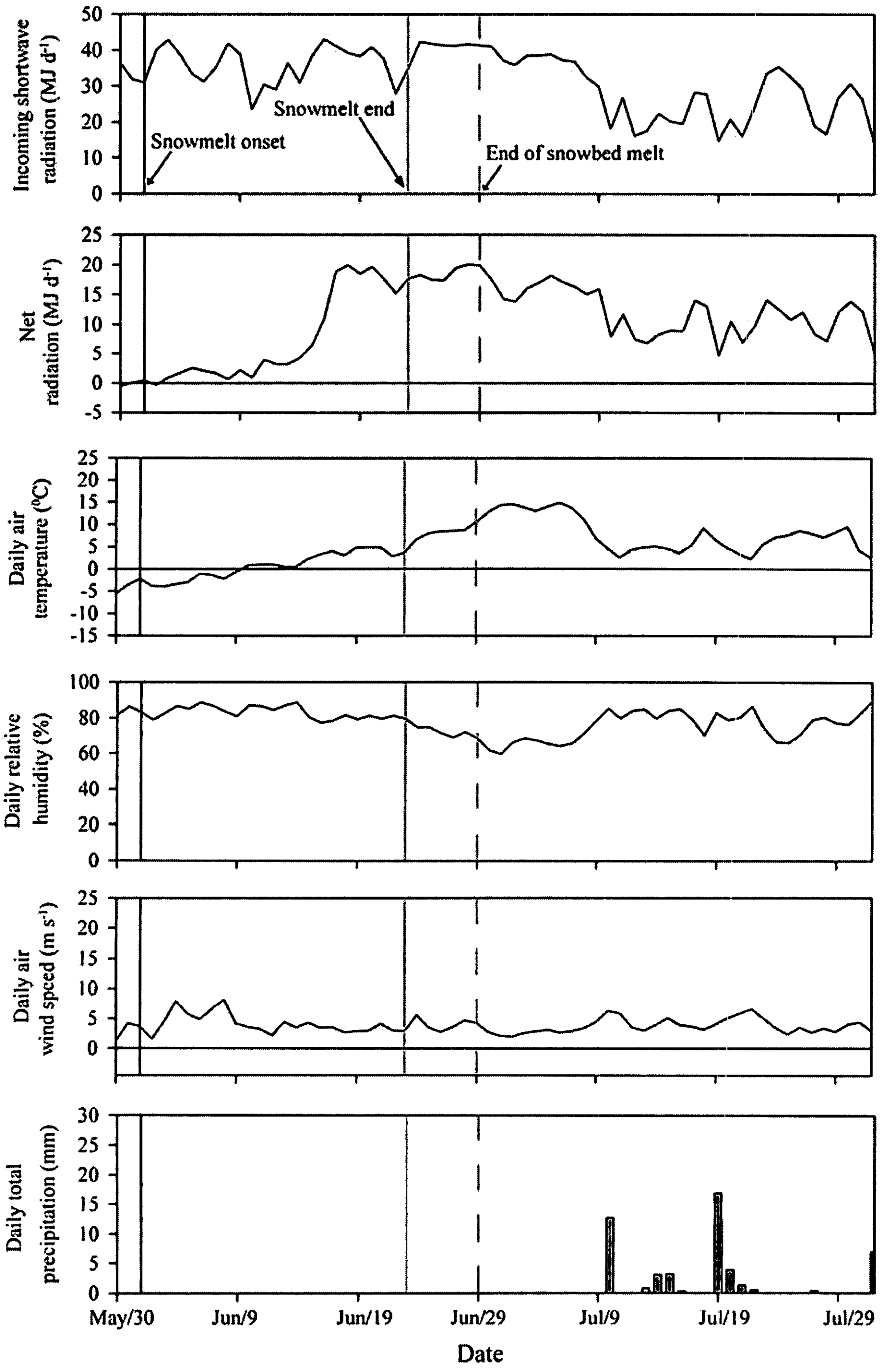

Incoming shortwave radiation (MJ d-1)
Snowmelt onset
Snowmelt end
End of snowbed melt
Net radiation (MJ d-1)
Daily air temperature (°C)
Daily relative humidity (%)
Daily air wind speed (m s-1)
Daily total precipitation (mm)
May/30
Jun/9
Jun/19
Jun/29
Jul/9
Jul/19
Jul/29
Date

Figure 2

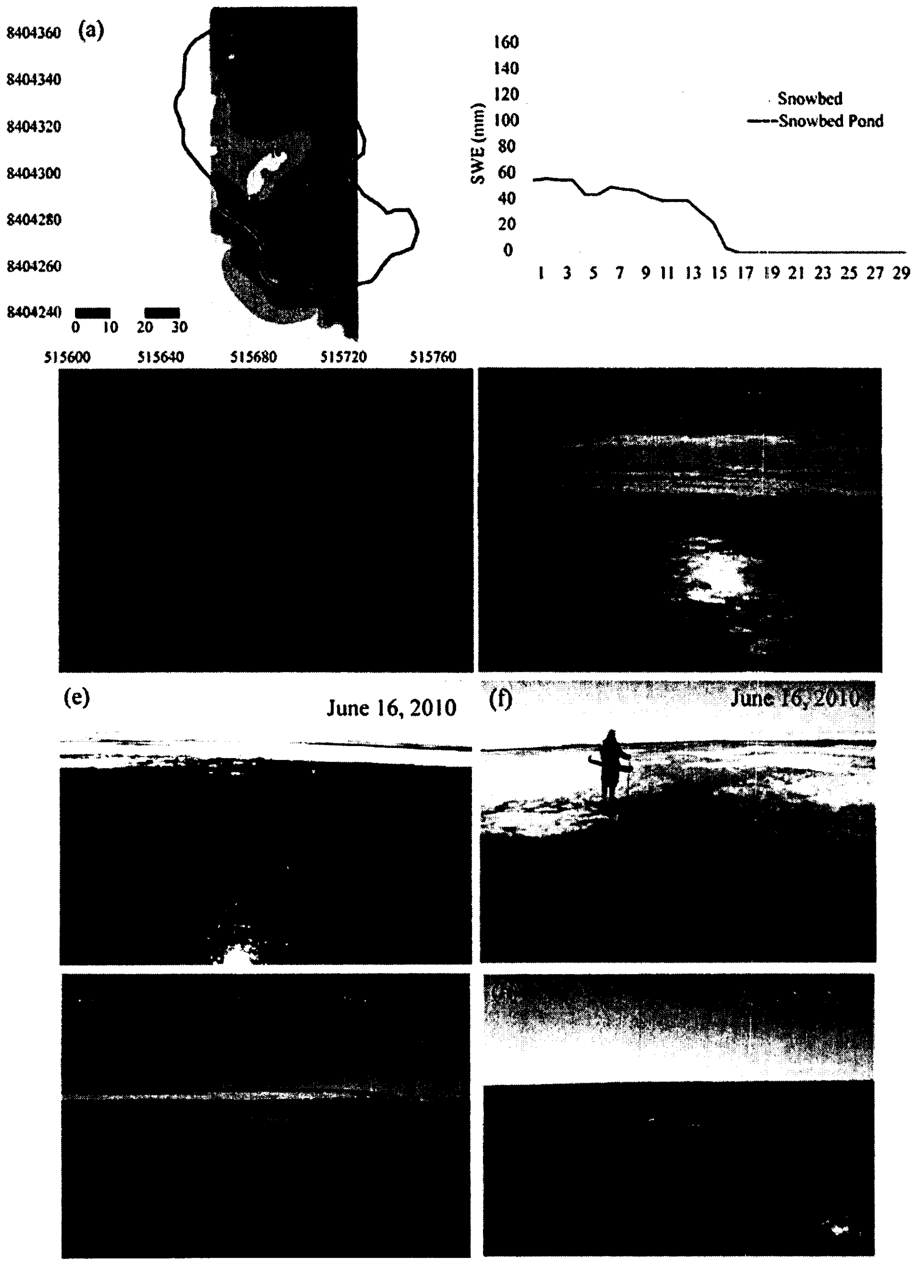
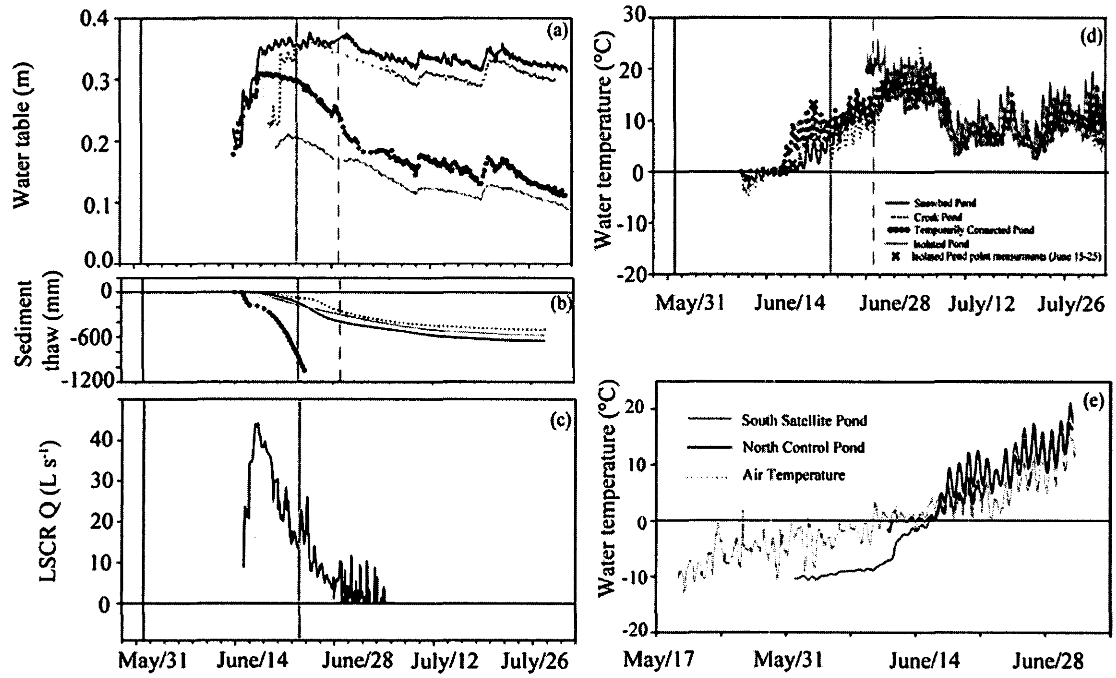
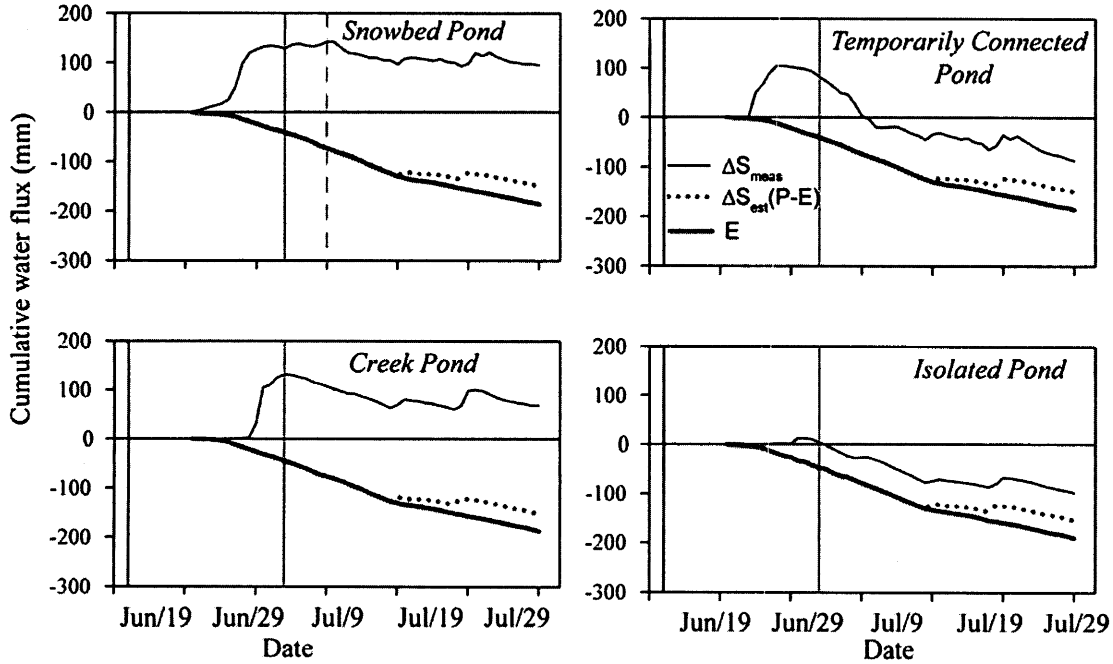

(a)
8404360
8404340
8404320
8404300
8404280
8404260
8404240
0 10 20 30
515600
515640
515680
515720
515760
160
140
120
100
80
60
40
20
0
SWE (mm)
Snowbed
Snowbed Pond
1 3 5 7 9 11 13 15 17 19 21 23 25 27 29
(e)
June 16, 2010
(f)
June 16, 2010

Figure 3



(a)
(b)
(c)
(d)
(e)
Water table (m)
Sediment thaw (mm)
LSCR Q (L s⁻¹)
Water temperature (°C)
May/31
June/14
June/28
July/12
July/26
May/17
South Satellite Pond
North Control Pond
Air Temperature

Figure 4



Snowbed Pond
Temporarily Connected Pond
Creek Pond
Isolated Pond
Cumulative water flux (mm)
200
100
0
-100
-200
-300
ΔS_{meas}
ΔS_{est}(P-E)
E
Jun/19
Jun/29
Jul/9
Jul/19
Jul/29
Date

Figure 5

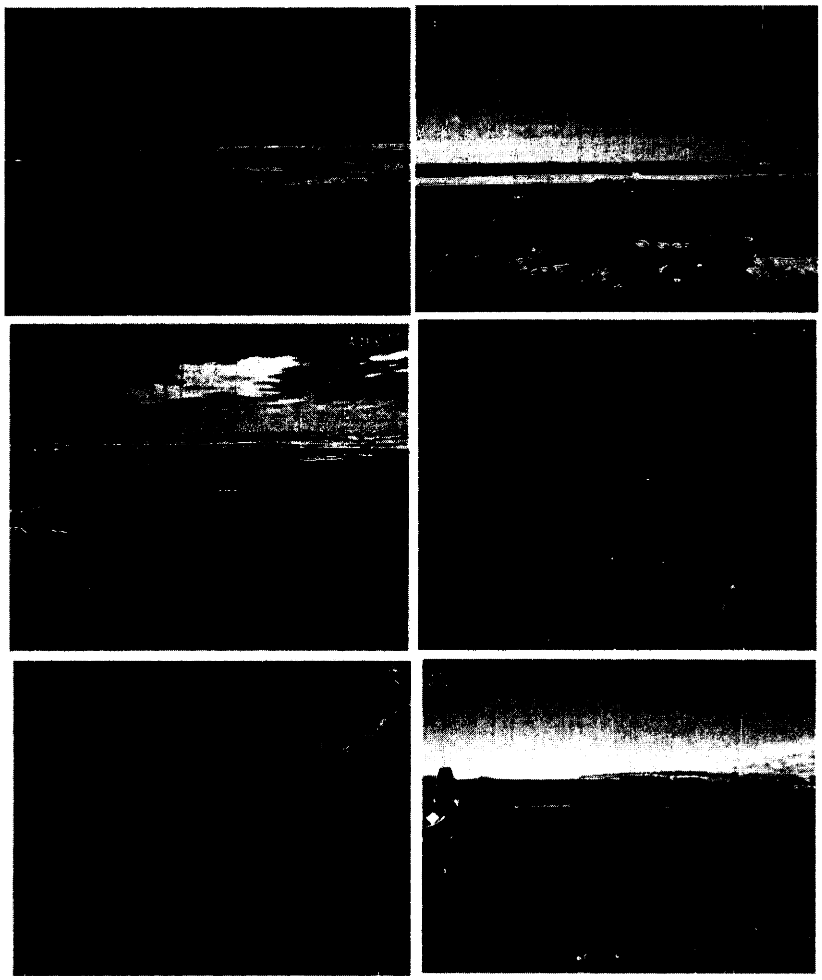
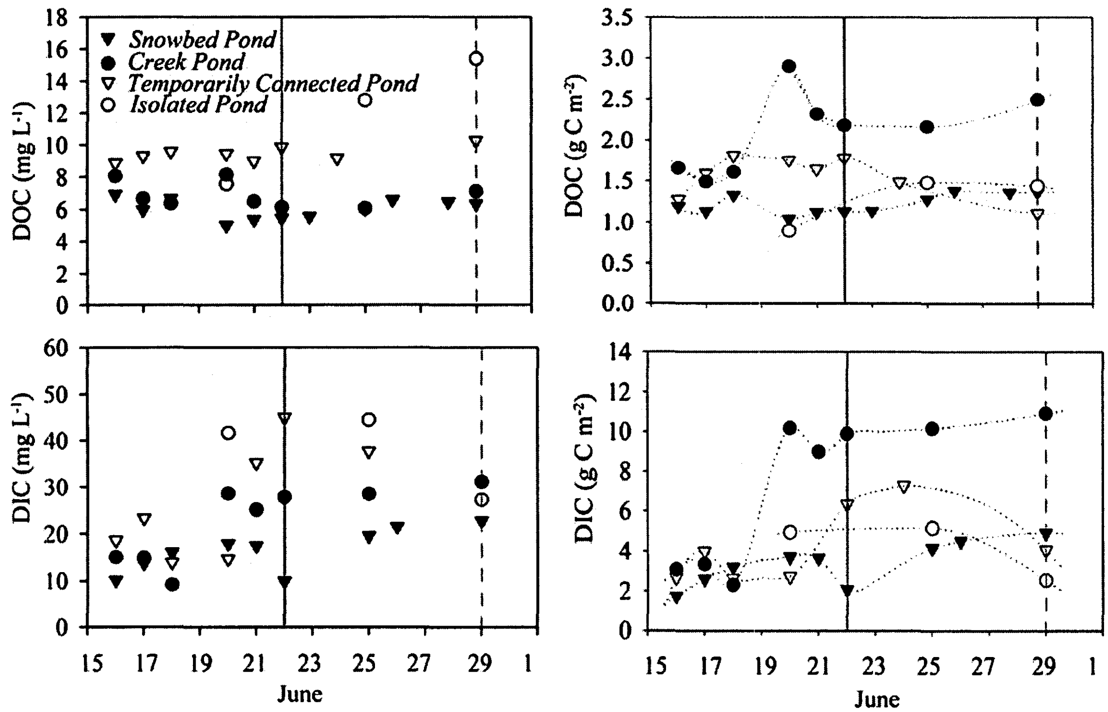


Figure 6



Snowbed Pond
Creek Pond
Temporarily Connected Pond
Isolated Pond
DOC (mg L-1)
DOC (g C m-2)
DIC (mg L-1)
DIC (g C m-2)
June
June

Figure 7

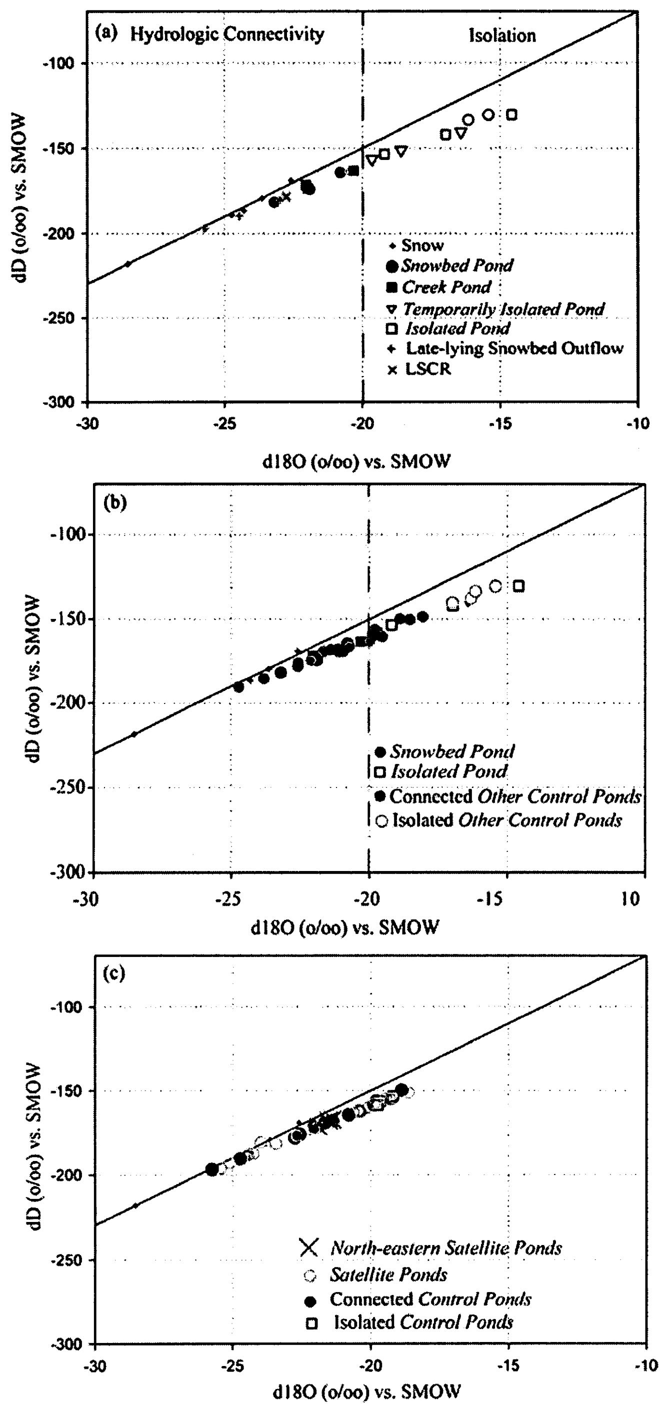
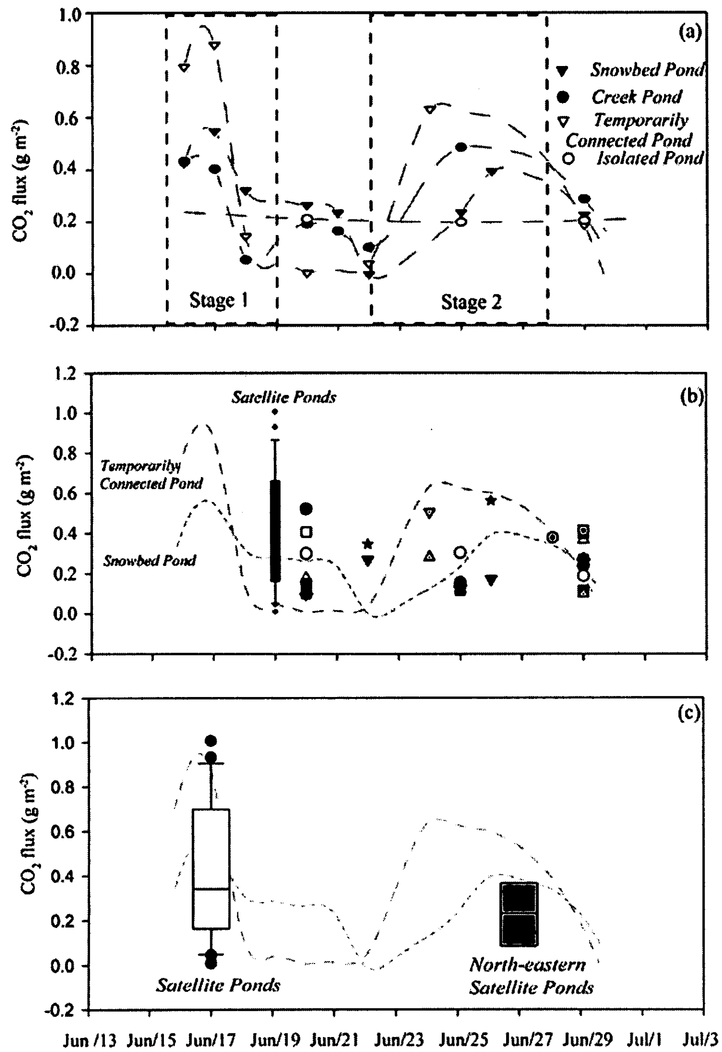
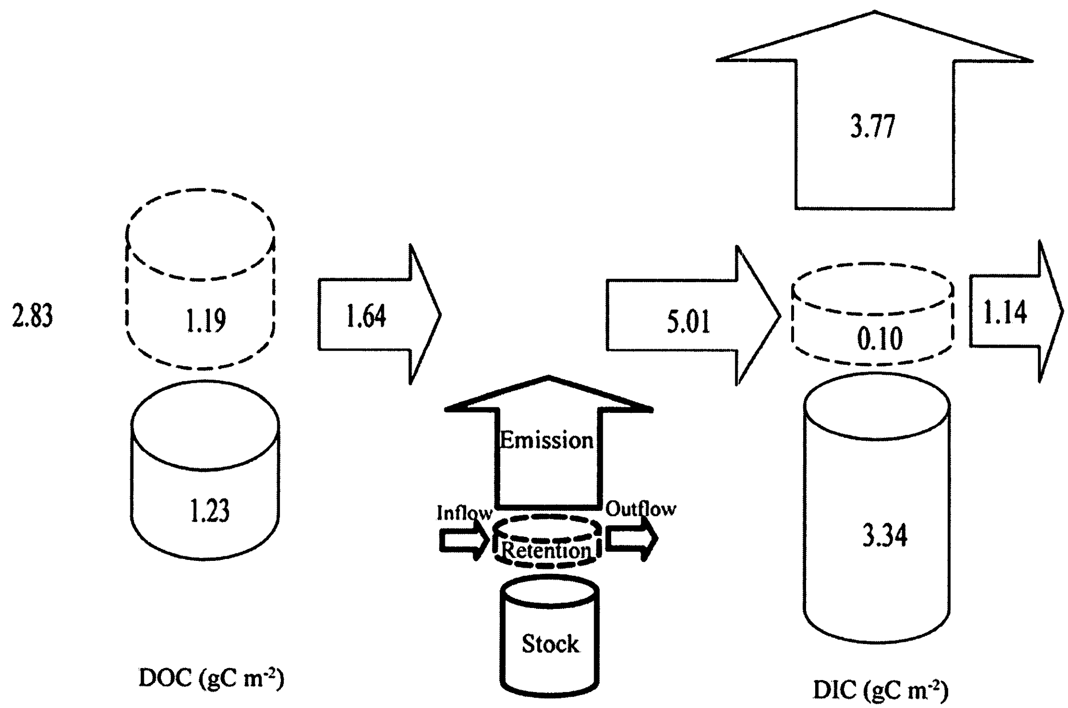

(a) Hydrologic Connectivity
Isolation
Snow
Snowbed Pond
Creek Pond
Temporarily Isolated Pond
Isolated Pond
Late-lying Snowbed Outflow
LSCR
dD (o/oo) vs. SMOW
d18O (o/oo) vs. SMOW
(b)
Snowbed Pond
Isolated Pond
Connected Other Control Ponds
Isolated Other Control Ponds
dD (o/oo) vs. SMOW
d18O (o/oo) vs. SMOW
(c)
North-eastern Satellite Ponds
Satellite Ponds
Connected Control Ponds
Isolated Control Ponds
dD (o/oo) vs. SMOW
d18O (o/oo) vs. SMOW

Figure 8



(a)
Snowbed Pond
Creek Pond
Temporarily Connected Pond
Isolated Pond
Stage 1
Stage 2
CO2 flux (g m-2)
(b)
Satellite Ponds
Temporarily Connected Pond
Snowbed Pond
(c)
Satellite Ponds
North-eastern Satellite Ponds
Jun /13 Jun/15 Jun/17 Jun/19 Jun/21 Jun/23 Jun/25 Jun/27 Jun/29 Jul/1 Jul/3

Figure 9



3.77
2.83
1.19
1.64
5.01
0.10
1.14
Emission
1.23
Inflow
Outflow
Retention
3.34
Stock
DOC (gC m-2)
DIC (gC m-2)

Figure 10

CHAPTER 4: Seasonal variability in hydrological and physico-chemical characteristics of small water bodies across a High Arctic wetland, Nunavut, Canada

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Abstract

Shallow arctic ponds represent the dominant components of arctic wetlands and play an important role as hydrological storage units and transit links by transporting water, nutrient and carbon to downslope ecosystems. Numerous studies from Alaska, Siberia and Arctic Canada show that arctic ponds are sensitive to the ongoing climatic variability and report alterations in ponds' surface areas and volumes as a result of climate change. The 94 km^2 extensive high arctic wetland at Polar Bear Pass (PBP) is composed of two large lakes and about 4000 small to medium-sized shallow ponds which together with the lakes occupy 60 % of the wetland area covered by water. The ponds contribute significantly to the surface water storage of this wetland, but to date no comprehensive information is available on the hydrology, limnology or the physico-chemical properties of the ponds across PBP, and how this ecosystem functions both seasonally and across years. We conducted a detailed hydrological and hydro-chemical monitoring program of 18 ponds (*Control Ponds*) and 3 frost cracks (linear water bodies) and sampled 30 ponds (*Satellite Ponds*) across PBP two to three times a season from 2007 - 2010. Using the measured water isotope tracers in 2007, 2009 and 2010, hydrological analysis, and our field observations we identified two hydrologic settings for PBP ponds: (i) ponds which are hydrologically connected to additional sources of water from their catchments beyond seasonal inputs of snowmelt and rainfall, or (ii) ponds which fail to form a steady link or only have a limited connection with their surrounding

catchments. The presence or absence of a hydrologic linkage (i.e. an external water source) to a pond was used to explain the variability in properties and gradients of the physico-chemical variables. The results indicate that pond physico-chemical characteristics are mainly influenced by seasonal changes in climatic conditions and hydrologic settings. Our classification of ponds at PBP identifies water bodies that have strong connectivity to their catchment and therefore are more resilient to ongoing climate change when compared to the isolated ponds that have lower hydrologic stability and show more variability in their seasonal hydro-chemical signatures.

Introduction

Arctic ponds represent the main components of extensive arctic wetlands and may occupy up to 95 % of the total water area in some circumpolar wetlands (Muster et al., 2013), yet the mechanisms responsible for their hydrologic sustainability and their relevance to nutrients and carbon fluxes on a landscape scale has, to date, remained largely unknown (Downing, 2009). Arctic ponds form in depressions resulting from ground thaw and subsidence of rich in ground ice permafrost, in topographic depressions and in the areas of polygon troughs connections (Woo 2012). These shallow water bodies have maximum water depths reaching 2 m during the open water season which lasts from the end of June until late August or early September and they are completely frozen during the remaining 8-9 months of the year. This short open water season and the harsh climatic conditions with low amounts of precipitation and low average air temperatures place arctic ponds among the most sensitive ecosystems in polar environments. Their sensitivity to the ongoing climatic changes combined with seasonal variations in precipitation, evaporation, and drainage alters their hydrologic balance and has a major effect on the surface areas and volumes of these water bodies. Today, researchers are already documenting a decline in pond and lake abundance, especially in Alaska, Northern Siberia, and Canada's Eastern Arctic (Yoshikawa and Hinzman, 2003; Smith et al., 2005; Smol and Douglas, 2007a; Keatley et al. 2007). These studies demonstrate that arctic ponds are very vulnerable to climatic changes highlighting the urgent need for research to address the mechanisms responsible for arctic ponds' hydrologic sustainability and resilience to disappearance and drying.

Predicted changes in climate are expected to result in alterations of not only hydrological, but also hydro-chemical and biogeochemical characteristics of arctic freshwater ecosystems. For example, the reduction in ice cover predicted by numerous climate change scenarios will elevate solar radiation receipt causing warmer temperatures and a longer growing season which will elevate primary productivity and nutrient cycling (Flanagan et al., 2003; Perren et al., 2003). Current research on the ponds in Barrow, Alaska has revealed significant changes in the physical, chemical, and biological characteristics of these ponds over time (Lougheed et al., 2011). By comparing data collected in the early 1970s with data from 2007-2010, Lougheed et al. (2011) observed a change in the thermal regime, with the greatest differences occurring early and late in the season, indicating an extended growing season and an overall increase in nutrients. While numerous studies have described arctic ponds as oligotrophic environments due to their low nutrient and solutes content in the water column, recent studies indentified high nutrient concentrations and carbon stores in sediments as being important (Rautio and Vincent, 2006; Carroll et al., 2011; Rautio et al., 2011). High carbon storage in pond sediments contributes substantial quantities of greenhouse gases (GHG) CO_2 and CH_4 to the atmosphere as a result of the mineralization processes of predominantly terrestrial dissolved organic carbon (DOC) and methanotrophic bacterial activity. Thus, today these small arctic aquatic systems are considered important reservoirs of carbon (Tranvik et al., 2009) and alterations in their water balance may potentially result in the production of considerable GHG emissions over large arctic regions.

A detailed monitoring of the physical, chemical, and biological characteristics of high arctic water bodies allows for better understanding of their response to inter-seasonal and intra-seasonal variability in climate (e.g. precipitation, air temperature). In the early seventies, intensive limnological studies were conducted in Barrow, Alaska and focused on detailed monitoring of hydro-chemistry, biomass, and biological productivity (Hobbie, 1980). The aquatic research in Alaska was followed by the formation of the Long Term Ecological Research (LTER) project in 1975 focusing on Toolik Lake to compare this arctic lake with the small shallow ponds in the area previously studied by Hobbie (1980). This long-term research project investigated biogeochemistry, nutrient and carbon cycles, biology of the lake and surrounding lakes and ponds and showed the importance of land-water interactions and hydrologic inputs of carbon to the otherwise nutrient-limited oligotrophic aquatic ecosystems (Whalen and Cornwell, 1985; Miller et al., 1986; Peterson et al., 1986; Hobbie et al., 1995). Similar detailed research was also conducted in the Canadian High Arctic during this time period but was limited to only a few studies

(Danks et al., 1974; Schindler et al., 1974ab). Beginning in the 1980s, numerous research studies were conducted in lakes and ponds across the Canadian High Arctic describing present-day water chemistry and other limnological characteristics (Prentiki et al., 1980; Hamilton et al., 1994; Duff et al., 1999; Hamilton et al., 2001; Jonsson et al., 2003; Smol, 2005; Smol et al., 2007b; Laurion et al., 2010; Breton et al., 2011). The main focus of these latter studies was the development of limnological and paleolimnological techniques (e.g. diatom algae; Peterson et al., 1985; Douglas and Smol, 2000; Lim et al., 2001) that could be used to track a variety of environmental trends throughout the Arctic (Smol, 2005). Eventually, a large number of survey studies were conducted in the Canadian High Arctic Archipelago encompassing small and shallow ponds and lakes (Pienitz et al., 1997ab; Hamilton et al., 2001; Lim et al., 2001; Michelutti et al., 2002ab; Antoniades et al., 2003; Lim and Douglas, 2003; Lim et al., 2005). However the harsh climate of the High Arctic along with financial and logistical constraints often impedes frequent and detailed monitoring of small arctic water bodies (Smol and Douglas, 1996), which limits the field work of many studies to a few sampling times during the open water season without considering seasonal dynamics. To date only a limited number of studies have considered the influence of seasonal hydrology and water movement on either lakes' or ponds' physical, chemical, and biological characteristics in the Canadian High Arctic (Kalff and Welch, 1974; Schindler et al., 1974 ab; Whalen and Cornwell, 1985; Dickman and Oullet, 1987; Breton et al., 2009; Laurion et al., 2010; Abnizova et al., 2012a).

A large numbers of ponds (~4000) exist in one of the most extensive wetlands in the Canadian Archipelago Polar Bear Pass (PBP), Bathurst Island (75° 40' N, 98° 30' W). These ponds contribute significantly to the surface water storage of PBP wetland, yet no research has been attempted to understand how these small waters function seasonally and between years. Past limnological data from Bathurst Island are very sparse (Lim et al., 2001). Lim et al. (2001) only sampled one pond in PBP during their water quality survey across Bathurst Island. No studies as yet have provided a summary of the pond limnology across this extensive wetland or illustrated how hydrology and inter-seasonal climatic conditions modify physico-chemical trends in the ponds here. Numerous biological studies have been conducted here to address vegetation (Miller and Ireland, 1978; Edlund and Alt, 1989) and wildlife population dynamics (Ferguson, 1987; Miller, 1991). The climatology and hydrology of the site has recently been addressed in research studies by Woo and Young (2006), Young et al. (2010), Young and Labine (2010), Abnizova et al. (2012a), and Abnizova et al. (submitted). These studies highlighted local variability in inter-annual and inter-seasonal hydro-climatic conditions at PBP.

Hydro-chemical properties of arctic pond waters are strongly modified by seasonal changes in their hydrology, causing variations in the proportions of biogenic and chemical components. Many existing limnological summaries of Canadian High Arctic lakes and ponds provide important information for analysis of the current and past environmental and paleolimnological changes (Douglas and Smol 1994; Antoniades et al., 2000; Gregory-Eaves et al., 2000; Hamilton et al., 2001; Antoniades et al., 2003 and references therein). Clearly, given the threat of climate change and recent research findings across the Arctic demonstrating that arctic ponds are already undergoing dramatic alterations in their ecohydrology, there is a real need to establish baseline conditions for sustainability of ponds which are often dominant components in arctic wetlands like PBP. This wetland represents an area of both local and international importance and has been designated a Ramsar site since 1982.

We examined the variability and trends in physical and chemical characteristics of arctic ponds at PBP with respect to their hydrological status (linked versus isolated) through the analysis of a comprehensive limnological dataset collected from 2007-2010. The study comprised 51 small freshwater bodies spanning the extensive area of PBP wetland (Fig. 1). This wetland is surrounded by hillslopes, and the location of some ponds close to the hillslopes provides them with an important hydrological connection to hillslope creeks and snowbeds in comparison to isolated ponds, ones, for instance, located centrally in the wetland. These latter ponds can be considered to have no hydrological connections. Consequently, the presence or absence of hydrological connectivity should elucidate different physico-chemical trends in the ponds across the wetland and help to identify the relative importance of hydrologic sources. Ultimately, pond response to inter-seasonal variability in climate (duration of flooding, shrinkage) should influence carbon components, and associated differences in physico-chemical characteristics of hydrologically connected versus isolated ponds.

Accordingly, this study aims:

1) to evaluate the differences in hydrologic connectivity with the catchment in a range of ponds across the wetland based on analysis of comprehensive hydrologic datasets, and seasonal evolution of stable water isotope signatures in the study ponds; and

2) to assess whether differences in hydrological settings and climatic conditions between observation periods impact physico-chemical compositions and trends in ponds across the wetland.

Site Description

The study took place at an extensive wetland located at Polar Bear Pass (PBP), Bathurst Island, Nunavut (98° 30' W, 75° 40' N). It is located in the centre of the Canadian Arctic Archipelago, northwest of Cornwallis Island and east of Melville Island (Fig. 1). This area experiences long, cold winters and short, cool, moist summers (Young and Labine, 2010), yet the wetland provides grounds for a rich floral and faunal diversity making it a biological oasis within a polar desert climatic setting. PBP has been identified as the most important wetland on Bathurst Island based on extent, productivity, and ecological significance (Babb and Bliss, 1974). The wetland is a designated national wildlife area due to its rich wildlife population and provides both feeding and breeding grounds for numerous migratory bird species, muskox, polar bear, wolves, foxes, lemmings, and Peary caribou. PBP is situated in a low relief valley that spans from the east to west coasts of Bathurst Island, and is approximately 94 km^2 in area. Low rolling hills (100 to 200 m a.s.l.) border it. A large number of thaw ponds with surface areas smaller than 1 ha (n = 4015, Muster, personal communications) represent the dominant feature of the wetland. There are a number of larger water bodies ranging from 0.1-0.9

km^2 (n = 11, Muster, personal communications) as well as two large lakes with areas of 5.3 km^2 and 2.1 km^2 that are centrally located in the wetland (Fig. 1). Numerous frost cracks are also found in areas with more prominent patterned ground (Fig. 2). The initiation and formation of frost cracks in arctic environments has been well described by Mackay (1964). Frost cracks can be sources of water to ponds but they can also drain water away from ponds, especially when they are at a lower elevation than a nearby pond (Abnizova and Young, 2010; Young and Abnizova, 2011).

The geological characteristics of the area are described as Devonian aged sedimentary rocks such as dolomitic sandstone, dolomitic siltstone, limestone, chert boulder conglomerate, dolomite, and remnants of patch reefs (Kerr, 1974). Glacial sands, gravels, or felsenmeer represent most of the surficial materials (Blake, 1964; Kerr, 1974). The remnants of reefs that are composed of algae, stromatoporoids, and corals are found as 2-m high rock formations seen on northern hillslopes (Kerr, 1974). Bathurst Island has been ice-free for approximately 9000 years. it was not covered by the large Laurentide ice sheet but was glaciated during the Wisconsin era (Blake, 1964). Evidence of this latter glaciation can be found in the area due to the presence of till, erratics, and v-shaped meltwater channels that are common across the region (Blake, 1964; Kerr, 1974).

The distinct geologic structures shape the physiography of Bathurst Island. The wetland valley, spanning from Goodsir Inlet in the east to Bracebridge Inlet in the west, was formed when the PBP anticline was preferentially eroded (Kerr, 1974). Surrounding the valley are large ridges that run east to west on the north and south side of PBP and serve as common features of Bathurst Island. The influence of isostatic rebound is seen in the raised beaches along these ridges which mark the vertical marine limit of approximately 90 m above present day sea level (Blake, 1964; Kerr, 1974). Numerous (> 50) v-shaped valley streams of various sizes (1^{st} to 4^{th} order) dissect the northern and southern ridges, reaching an elevation of 180 m a.s.l and 120 m a.s.l respectively. Since the wetland (24 – 26 m a.s.l.) has a very small vertical gradient (0.002) and poor vertical drainage as a result of continuous permafrost and shallow active layer, the water flow out of the wetland is thought to be greatly reduced. Fieldwork in 2012/13 is now evaluating this premise.

Vegetation cover typically found in the upland areas is described by Sheard and Geale (1982ab) as sparsely vegetated with small clusters of herbaceous perennials: *Saxifraga oppositifolia, Saxifraga caespitose, Papaver radicatum, Cerastium alpinum, Draba, Puccinellia bruggemannii, Braya purpurascens*, mosses (*Ditrichum flexicaule)*, and lichens (*Collema* sp, *Lecidea ramulosa*) (Fig. 2). The lowland areas contain vascular vegetation such as grasses and sedges (*Carex stans, Salix arctica,* and *Dupontia fisheri*) interspersed with moss mounds (*Orthothecium chryseum, Tomenthypnum nitens)* that make up the majority of the valley floor (Sheard and Geale, 1982ab) (Fig. 2).

Methods

Hydrologic and physico-limnological variables

Fifty-one small water bodies were sampled at PBP during the field seasons of 2007-2010. The sampling was conducted from June 21-August 1 in 2007, June 1-September 1 in 2008, June 1- September 2 in 2009, June 1-August 1 in 2010. In 2007 we surveyed fifteen ponds with the remaining thirty-six ponds being added in early 2008 (Fig. 1). Sites were selected to represent as wide an environmental gradient as possible, and were sampled within a short period of time each year to minimize variability due to seasonal fluctuations in climate. Manual (once a week or at the time of every visit), and, if possible, continuous measurements (using HOBO water level pressure transducers) of water tables were made at every monitored site. We selected and monitored in greater detail (once or twice a week) 5 ponds that are representative of the wetland and will refer to these ponds as *Control Ponds* (C1, C8, C10, C11, C14, see Table 2). The *Control Ponds* differed based on the various degree of hydrologic connectivity to their catchments varying from a link to a late-lying snowbed (C1), a hillslope creek (C14) or a neighbouring pond (C11). The ponds situated in isolated locales and receiving water inputs from precipitation only are listed as C8 and C10. A detailed description of the *Control Ponds* and their seasonal hydrology can be found in Abnizova et al. (2012a) and Abnizova et al. (submitted). We also selected a set of 13 ponds with varying degrees of hydrological connectivity to their catchment and monitored these ponds at least once a week during each study season, and refer to them as *Other Control Ponds* (C2-7, C9, 12, 13, 15-17, 21, see Table 2 and Fig. 1).

Two frost cracks were also studied from 2008-2010 to assess their role in delivering water to ponds. Frost crack CF18 (max depth = 346 mm) was located in the northern upland part of the wetland and was connected to pond C16. The second frost crack CF19-20, located in the central part of the wetland, was shallower (max depth = 136 mm) and was connected to the pond C21. During periods of warm temperatures and no precipitation this second frost crack separated into two sections and was measured separately as CF19 and CF20. In the context of this study, this frost crack is reported as a single system and averages from both sections are shown.

To better understand the hydrology and physico-chemical characteristics of other ponds at PBP, we selected *Satellite* ponds located in the east (S22 - S24), southeast (S25 - S30), south (S31 - S36), west (S37 - S39), west central (S40 - S45), and southwest (S46 - S51) sectors of the wetland (Fig. 1). In order to monitor these satellite ponds we followed the sampling approach of Lim et al. (2001). They used a hydro-mapping approach to identify the limnological conditions of ponds over a short open water season (~2 months). To compensate for changes in climatic conditions between seasons the same ponds were sampled during warm/dry (2007, 2010) and cool/wet seasons (2008, 2009). Limnological changes during the season were also considered by sampling the ponds during the snowmelt and pre-freezeback periods, and occasionally during peak summer. In all years, all samples were collected within a three day window, a time-frame similar to other limnological surveys (e.g., Douglas and Smol 1994; Antoniades et al., 2000; Lim et al., 2001). Surface area estimates in all investigated ponds varied with 52% ponds having surface areas less than 1,000 m^2 (small) and the remaining 48% ranging from 1,150 - 31,400 m^2 (medium to large). The ponds ranged in water depth from 23 mm to 636 mm (Table 2 and 3) and no thermal stratification was observed during all the study seasons.

Various physical and chemical variables were collected from 2007 - 2010. In the summer of 2007, fall of 2008, and spring of 2009 water table (WT, mm), frost table (FT, mm), water temperature (TEMP, °C), conductivity (CON, $\mu S\ cm^{-1}$), and pH were measured in the field. Water samples were collected from central locations within the ponds and frost cracks at 10 cm below the surface water and were later analyzed for major cation concentrations including: calcium (Ca, mg L^{-1}), magnesium (Mg, mg L^{-1}), potassium (K, mg L^{-1}), and sodium (Na, mg L^{-1}). In 2009, two additional variables were analyzed: dissolved oxygen (DO, mg L^{-1}) and chloride (Cl, mg L^{-1}). A more detailed chemical analysis was performed in 2010. In addition to the ions sampled in the previous years, the following ions were included: aluminum (Al, μg L^{-1}), barium (Ba, μg L^{-1}), iron (Fe, μg L^{-1}), manganese (Mn, μg L^{-1}), total dissolved phosphorus (P, mg L^{-1}), silicon (Si, mg L^{-1}), strontium (Sr, μg L^{-1}), fluoride (F, mg L^{-1}), sulphate (SO_4, mg L^{-1}), bromide (Br, mg L^{-1}), nitrate (NO_3, mg L^{-1}), phosphate (PO_4, mg L^{-1}), hydrogen carbonate (HCO_3, mg L^{-1}), dissolved organic carbon (DOC, mg L^{-1}), and dissolved inorganic carbon (DIC, mg L^{-1}). Samples were kept cool and dark until they had returned from the field, at which time they were immediately sent to laboratories where quality assurance and quality control procedures were followed for storage and analysis. Chemical analyses were carried out at York University Polar Biogeochemistry Laboratory, at the Facility for Biogeochemical Research on Environmental Change and the Cryosphere at Queen's University, and at the Geosciences and Periglacial Research Laboratory at Alfred Wegener Institute. Protocols for bottling and filtering, and methodology for chemical analyses followed after Abnizova et al. (2012a), Abnizova et al. (2012b), and Thompson and Woo (2009).

Water sampling in the local ponds was carried out once or twice every week during the study periods. Values of pH, conductivity, water temperature, and dissolved oxygen were obtained with handheld portable YSI units at the time of sampling. DOC samples were filtered using GF/F filters (0.7 μm pore diameter) and a Polyethylene (PE) syringe, and collected in 30 mL High Density Polyethylene (HDPE) plastic containers. At the time of collection the sampling bottles (30-60 mL) were rinsed 3 times with filtered sample water. The samples were acidified to pH 2 by adding 2M HCl. Water samples were then stored in the dark at 4°C. Unfiltered water samples for analysis of DIC concentrations were collected in headspace-free, sealed glass vials and kept in a cold, dark environment until laboratory analysis (Abnizova and others 2012b). The preservation method was further tested in the laboratory environment to identify the extent of alteration in DIC concentration as a result of degradation and oxidation of organic carbon in water samples during transportation time. The experiment showed small changes in DIC (8.53 %) in the tested samples compared to their initial concentrations at the time of sampling.

Water samples for cation/anion analysis were filtered using cellulose-acetate filters (0.45 μm pore diameter) and pre-cleaned PE syringes and stored in 15 mL HDPE plastic containers. At the time of sample collection, the sampling bottles were rinsed 3 times with filtered sample water. The samples for the analysis of cation concentrations were acidified with 100-200 μL (12 N) HNO_3. Anion analysis was performed on an IC Dionex DX 320 and cation analysis was evaluated with an ICP-OES Perkin-Elmer Optima 3000xl. Total dissolved nitrogen (TDN) and dissolved organic carbon (DOC) concentrations were determined simultaneously in the filtrate by high temperature catalytic oxidation with a Shimadzu TOC-VCPN analyser. Concentrations of DIC were analysed using a gas chromatograph for simultaneous detection of CO_2. The Shimadzu GC-2014AF gas chromatograph was equipped with an AOC-5000 autosampler, a 1 m × 1/8" HayeSep Q 80/100-mesh column, an electron capture detector, and two flame ionization detectors.

Stable isotopes of oxygen and hydrogen ($\delta^{18}O$ and δD)) in pond surface waters were sampled at the same time as hydro-chemical samples in the summers of 2007, 2009 and 2010 and stored collected in 30 mL High Density Polyethylene (HDPE) plastic containers. An equilibration technique using a mass-spectrometer (FinniganMAT Delta-S) was used (Meyer et al., 2000).

Statistical analyses and data screening

Principal component analysis is a multivariate statistical technique that reduces dimensionality by using ordination to identify the key variables that are responsible for the majority of variation within a dataset (Leps and Šmilauer, 2003). In total, five Principal Component Analyses (PCA) were run using the CANOCO program V. 4.5 (ter Braak and Šmilauer, 2002) to examine the variation of physico-chemical components between sites for summer 2007 (July 26th- Aug 2nd), fall 2008 (Aug 19th – Aug 31st), summer 2009 (Aug 3rd – Aug 5th), fall 2009 (Aug 26th – Aug 30th), and spring 2010 (Jun 19th – Jun 24th). These datasets were chosen to provide both seasonal as well as inter-annual comparisons. The number of principal components produced by a PCA is equal to the number of variables, however the majority of the variation within a dataset can usually be explained by the first and second principal components (axis 1 and axis 2), and those are the only ones that will be reported in this paper. The amount of variation explained by a principal component is indicated by its Eigen value that is often expressed as a percentage.

All variables that did not have normal distributions were transformed using either log x, log (x + 1), or square root following Lim et al. (2001). Only variables that displayed normal distributions initially or after transformation were included in the analysis and they are represented in the PCA diagrams by solid lines. Non-normal data were added back into the PCA diagrams post analysis as passive variables and are indicated in figures by dashed lines. In 2007 three variables were transformed (WT, CON, and K); in 2009 summer two variables were transformed (CON and Na); in 2009 fall four variables were transformed (FT, Cl, K and SA) in 2010 seven variables were transformed (CON, TEMP, Cl, Ba, Mg, Sr and Si,). No data from 2008 fall was transformed. Initial screening of the data showed the frost cracks to be outliers and they were removed from the PCA analyses.

The linear discriminant analysis (LDA) or canonical correlation analysis was applied to find the linear combinations of the hydro-chemical variables used in the PCA analysis. The LDA analysis was used to confirm the best possible separation between the pond groups (hydrologically linked vs. isolated ponds) in the data sets that resulted from the analysis of stable water isotope tracers. Non-normalized datasets used in the PCA screening were further analyzed using the function lda {MASS} with software R version 2.15.3 (R Development Core Team 2008).

Four variables (NO_3, PO_3, Al, TDP) were removed from the data set because they occurred below the detection limits in greater than 50% of the sites. Variables that were below the detection limits in less than 50% of the sites were not removed from the data-set, but were replaced with the measurement of detection for that variable. Since many variables had a skewed distribution, a non-parametric Spearman correlation matrix was performed following Kumke et al. (2007).

Results

Climate 2007-2010

The climate conditions varied among the four field seasons: 2007 experienced a warm and dry July, the summers of 2008 and 2009 were cool but received elevated levels of precipitation (96 mm - 2008; 95 mm - 2009), while the summer of 2010 was warm (July) and experienced a number of large rainfall events (Fig. 3, Table 1). For comparison, climatic data from the nearest Government Weather station at Resolute Bay, Cornwallis Island is also presented. Evaluation of the climate record of air temperature and precipitation indicates that weather conditions at PBP in 2007-2010 were similar to Resolute Bay, confirming a polar desert climatic designation (Woo and Young, 2003; Young and Labine, 2010).

Hydrology

Most of the *Control Ponds* were shallow with a mean water depth of 239 mm (Table 2). There were some exceptions though, such as pond C2 and C16, with measured water depths reaching 636 mm and 568 mm, respectively. Apart from these ponds all *Other Control Ponds* varied from 26 mm to 446 mm and the *Satellite Ponds* had similar water depths ranging from 52 mm to 505 mm. All study sites represented well-mixed water bodies and were not stratified at any time during the observation period and sampling.

Previous findings from more intensive hydrologic investigations at this site which focused primarily on the *Control Ponds* showed that the hydrology of ponds at PBP is controlled by the presence or absence of a hydrologic linkage (e.g. snowbed, hillslope creek, upslope pond) (see Abnizova et al., 2012b; Abnizova et al., submitted). To visually display these differences and to better understand differences in the physico-chemical properties of the study ponds, selected hydrological plots are presented in Fig. 4. This diagram shows two general groups that represent the variability in pond water tables at PBP: hydrologically linked small and large ponds (referred to as linked ponds), and hydrologically isolated small (<1,000 m^2) and larger sized ponds (referred to as isolated small or large ponds, respectively). In addition, changes in water levels of a frost crack (CF18) are also displayed. These hydrological groups were used when examining the trends indentified in the stable water isotopes and in the measured physico-chemical variables.

The observed linked ponds at PBP demonstrated differences in the seasonal stability of water tables over the 2007-2010 study seasons. These water tables ranged from 144 mm to 365 mm (Table 2, 3 and Fig. 4). The isolated ponds exhibited even greater seasonal variability in water levels when compared to the linked ponds (-280 mm to 568 mm). Like the isolated ponds, water table changes in frost cracks were considerable over the 2008 to 2010 study seasons. One frost crack's water table fluctuated greatly during the dry periods in 2010 (149-280 mm) and again in the early season of 2008 (CF18, Fig. 4e), but its water levels remained relatively stable during the wet study season of 2009.

Examination of the seasonal changes in the selected *Satellite Ponds* (east, south, and west ponds, Table 3) showed S23 (a medium-sized pond) in the east sector as being a hydrologically connected pond in 2009 (Table 3, Fig. 2d). Pond water levels increased even after the snowmelt season suggesting an additional source of water draining into it. This differed from S33 (south small pond) and S38 (a west small pond) whose water levels were more typical of isolated ponds (Fig. 4 c, d, f). Water levels dropped quickly after snowmelt implying that they were dependent on snowmelt and seasonal precipitation for their water supply.

This pattern is supported by stable water isotope analysis. Fig. 5 represents a δD versus $\delta^{18}O$ scatter plot of seasonal isotopic evolution from the hydrologically connected and isolated *Control Ponds* in 2007, 2009 and 2010. Each data set plots along a line with a slope that intersects the global meteoric water line (GMWL). This isotopic composition of pond waters represents the weighted average of regional precipitation that has undergone varying degrees of fractionation due to evaporation (Criss and Davisson, 1996). Linear regression analysis demonstrates variability in the slopes between the linked and isolated ponds (Fig. 5b). While the slopes of regression are different between the linked (estimate=7.18) and the isolated ponds (estimate=5.01) in 2010, regression slopes were similar for the linked and isolated ponds in 2007 (slope estimates are 5.66 and 5.28, respectively). The data matrix in Fig. 5 shows strong grouping in the plot with a different segregation between the water isotopic compositions from the linked and isolated ponds and demonstrates a seasonal separation between these ponds' isotopic signatures. The hydrologically connected ponds are located closer to the GMWL during the snowmelt and retain this separation that ends only in the early post-snowmelt season. The snowmelt signatures of the isolated ponds plot clearly away from GWML and are found together with the hydrologically connected ponds' signatures in the post-snowmelt season. The third type of segregation in isotopic compositions is observed in the isolated ponds plotting in the area of the most isotopically-enriched values and begins right after the snowmelt.

Fig. 6 shows the application of the isotopic framework from Fig. 5 to evaluate hydrological status of ponds at PBP that were studied with less intensity or only sampled a few times during the season. The inter-seasonal isotopic separation between the hydrologically connected and isolated *Control Ponds* is clearly seen in 2007, 2009 and the early part of 2010 season (Fig. 6a-c). Similarly, in 2009 and 2010, the seasonal evolution of isotopic compositions in the *Other Control Ponds* shows segregation between the hydrologically connected and isolated ponds (Fig. 6d).

In 2009, the *Satellite ponds* plot in the area of isotopic composition which is similar to the isotopic composition of the isolated *Control Ponds* during the post-snowmelt season. In 2010, the *Satellite ponds* sampled only once during peak snowmelt on June 19 (Fig. 6f) are located close to GWML and have a similar range of isotopic signatures as hydrologically linked and isolated *Control Ponds* at this time of year.

Physico-chemical characteristics

All the investigated wetland ponds showed varying dissolved salt concentrations and ion compositions (Table 2 and Table 3). The mean pH of the control sites was pH 8.2 with a relatively wide range of 5.7-9.1 during the four seasons. The mean pH of the satellite ponds was also 8.2 indicating that the pH of ponds across PBP is similar. The range of pH values for the satellite ponds was slightly narrower (6.6 - 9.0). Conductivity values had a mean value of 273 $\mu S\ cm^{-1}$ with a range of 4 $\mu S\ cm^{-1}$ to 740 $\mu S\ cm^{-1}$. The four highest values were recorded in C8, C11, CF18, CF19-20 sites. The satellite ponds had a similar average and range in conductivity (271, 47 - 752 $\mu S\ cm^{-1}$) compared to the centrally-located ponds. Ponds in the eastern sector (S22-S24) generally had the highest conductivity values in a given season, with the largest variation in conductivity between ponds with small and large surface areas.

Low conductivity values (between 4 and 444 $\mu S\ cm^{-1}$, Table 2) were observed during the snowmelt seasons in all four years in most of the investigated ponds. The lowest post-snowmelt seasonal value (mean = 223 $\mu S\ cm^{-1}$) was found in pond C16 in 2009. In 2009, C11, an isolated type of pond, showed the highest mean conductivity level (mean = 440 $\mu S\ cm^{-1}$). All study sites were well oxygenated with median DO of *ca* 12.0 mg L^{-1} and a variability of CV = 18%.

The more intensive hydro-chemical mapping in 2010 of ponds and frost cracks revealed significant differences in the proportions of HCO_3, SO_4 and Ca ions. A significant relationship was found between conductivity and the concentrations of HCO_3, Ca and Mg cations ($P < 0.05$). Conductivity was also a function of the chloride ion content and exhibited a correlation coefficient of 0.707 ($P < 0.05$).

When plotted in a Piper diagram (Fig. 7; Piper, 1944) all water samples of ponds plot within the following hydro-chemical categories: HCO_3-Ca, HCO_3-Ca-Mg, HCO_3-Cl-SO_4, listed in order of dominance. Based on the measurements taken in 2010, hydrologically connected ponds only fall into the HCO_3-Ca-Mg category and plot along a straight line in each of the fields of the diagram as the result of mixing Ca and Mg ions in the water. Isolated ponds' and satellite ponds' chemical compositions of the water samples do not plot along the straight line on the Piper diagram, and cannot be related by a simple mixing between two end members. Here chloride ions tend to occupy up to 24% of the chemical compositions. Overall, regardless of their hydrological group and location, the ion content of all water bodies across the wetland was dominated by calcium cations and bicarbonate anions. Bicarbonate ions accounted for 79-100% of total anions, whereas calcium ions accounted for 56-76% of total cations in 2010. The (Ca+Mg)/ HCO_3 ratio was never greater than one (median = 0.33), indicating that the total water hardness in the ponds was not caused by the dissolution of calcium carbonate alone.

The positive relationship between Na and Cl is apparent, as suggested by the high correlation coefficient ($R = 0.877$, $P < 0.05$). This trend supports the hypothesis that these chemical elements have the same source. A weak but positive relationship between CON and SO_4 is also observed (R = 0.377, $P < 0.01$). Sulphates are the second most important anion in pond waters and are responsible for 10-25% of the total ion content. A strong positive correlation was found between the Na and K ions (R = 0.755, $P < 0.05$).

Generally, the concentrations of most chemical components slowly increased from the beginning of the snowmelt season to the end of summer in linked ponds. A more pronounced response was found in the isolated ponds, with dramatic increases in solute concentrations during the dry warm periods in small isolated ponds (2008, 2010). The highest specific conductivity values were observed when the air temperatures started to decrease and an ice cover formed in 2008. During this period the conductivity level rose from 354 to 373 $\mu S\ cm^{-1}$ in a small hydrologically connected pond (Fig. 3) and to a lower extent in other ponds. An increase in air temperature then led to the melting of this initial skim ice and the conductivity values leveled or decreased in the studied ponds (Fig. 3).

The major ion concentrations were relatively similar among all sites; however deeper ponds had lower concentrations than shallower ponds with respect to ionic concentration. In 2007-2009, the relative concentration of cations among sampled sites was Ca > Na > Mg > K with means of 39.1 mg L^{-1}, 14.7 mg L^{-1}, 9.4 mg L^{-1}, and 1.3 mg L^{-1} (1.95 , 0.64, 0.73 and 0.36 in meq L^{-1}) respectively for the centrally located ponds, and means of 40.5 mg L^{-1}, 17.9 mg L^{-1}, 7.7 mg L^{-1}, and 1.5 mg L^{-1} (2.02 > 0.78 > 0.64 > 0.04 in meq L^{-1}) respectively for the satellite ponds. In 2010, the relative concentrations of cations was different (Ca > Mg > Na > K), with means of 43.6 mg L^{-1}, 10.2 mg L^{-1}, 4.6 mg L^{-1}, and 0.9 mg L^{-1} (2.18 > 0.85 > 0.20 > 0.02 in meq L^{-1}) respectively for the centrally located ponds, and means of 17.9 mg L^{-1}, 2.9 mg L^{-1} 3.0 mg L^{-1}, and 0.9 mg L^{-1}

(0.89 > 0.24 > 0.13 > 0.02 in meq L^{-1}) respectively for the satellite ponds (Table 2, Table 3). High concentrations of K were found in some ponds at PBP and concentrations ranged from 0.0-5.9 mg L^{-1}. The K concentrations in the satellite ponds ranged from 0.2-3.8 mg L^{-1}.

Major anions showed the following trend in concentration levels: HCO_3 > Cl > SO_4 (see Table 2). High concentrations of Na and Cl were largely a function of proximity to the arctic coastal sea-waters, a distance of less than 2 kilometres from the east ponds (S22, S23, and S24, Table 3). SO_4 concentrations were usually high, ranging from <0.1 mg L^{-1} (C10) to 23.7 mg L^{-1} (C2), with a mean concentration of 5.6 mg L^{-1} (Table 2). *Satellite ponds* had a much lower mean SO_4 concentration of 0.9 mg/L and a narrower range of 0.05 mg L^{-1} (C32) to 5.0 mg L^{-1} (C23, Table 3). Eastern sector ponds (C22-C24) exhibit SO_4 concentrations higher than the other satellite ponds, ranging from 2.4-5.0 mg L^{-1}.

DOC values averaged 12.0 mg L^{-1} and ranged from 0.4 - 41.6 mg L^{-1}. The highest DOC values were typically found in water bodies with small surface areas. The mean alkalinity value was 124 mg L^{-1} for the ponds/frost cracks and ranged from 30-249 mg L^{-1}.

The Spearman correlation matrix revealed that conductivity in 2009 was negatively correlated with surface area (SA) and water table (WT) but positively with TEMP, Ca, and Mg (Table 4). The correlation matrix revealed significant positive correlations in 2010 hydrological mapping between CON, DO, TEMP, Ca, Mg, Na, K, Ba, Si, Sr, HCO3, Cl, SO4, DOC, and DIC (Table 5).

Concentrations of Al, NO_3, and PO_3 in the investigated ponds were low and therefore were not used in the statistical analysis.

Ordination analysis

The PCA plot for the spring of 2010 which analyzed twenty one variables from 40 of the pond sites is found in Figure 8. In total, axis 1 and axis 2 explained 70.4% of the variation (60.6% and 9.7% respectively). The species scores of major ions, including Ca (0.952), HCO_3 (0.945), Mg (0.935), Ba (0.910), CON (0.902) were most strongly associated with axis 1 (Fig. 11) while WT (-0.778) was associated with axis 2 suggesting the ponds can be differentiated by a primary chemical gradient and a secondary physical gradient. In the ordination diagram (Fig. 8) almost all control ponds (except C16) are located in the right quadrants while the majority of the satellite ponds are located in the left quadrants away from the variables associated with axis 1. Some ponds from the east site (S22 and S23) are clustered with the hydrologically connected control ponds. Isolated control ponds C8 and C11 are located relatively far from the other control ponds while C9 and C10 are near the control ponds.

PCA analyses were also conducted for the summer of 2007, fall of 2008, summer of 2009 and fall of 2009. The number of sampled ponds was large (16 to 45 ponds) however only eight to eleven environmental variables were collected during those surveys so these analyses are only briefly mentioned here and the plots are not shown. In general conductivity, pH, calcium and magnesium were the factors that dominated the first axes and were responsible for most of the variability in the datasets. Similar to the spring 2010

PCA, results from the PCAs in the summer of 2007 and the fall of 2009 also displayed axes divided along physical and chemical gradients where chemical variables were associated with axis 1 and a physical variable, specifically frost table, was most strongly associated with the second axes.

Discussion

Hydrology and hydrologic connectivity

The hydrological properties of wetland ponds (e.g. water table, surface area) on a catchment scale are defined by water balance, which in turn is dependent on vertical fluxes (precipitation and evaporation) and lateral fluxes (surface and subsurface inflows and losses). These hydrologic fluxes are determined by climate and local geomorphologic settings, making arctic ponds' water balance highly sensitive to climatic variability. Seasonal climatic conditions determine the amount of available snow prior to snowmelt, the rate and duration of melt, the extent of hydrologic connectivity, the rate of evaporation, the degree of active layer thawing and thus exert controls over the water balance of the wetland ponds. In dry years (e.g. 2007 and 2010) the rate of evaporation is typically high for ponds at PBP, often resulting in negative water balances, especially during extended periods of warm air temperatures and absence of precipitation (see Abnizova et al., submitted). During these periods, isolated ponds typically experience large drops in water levels, which sometimes led to their complete desiccation or extreme surface area shrinkage (Abnizova et al., 2012b).

The ponds at PBP showed a high level of variability in water depths during the four observation periods. Large isolated ponds showed considerable variability (e.g. C10) in their water tables but did not undergo desiccation during dry periods as a result of delayed and shallow snowmelt. This is quite typical for the central and southern sectors of the wetland. According to Assini and Young (2012) the spatial snowmelt pattern evident at PBP can be attributed to a non-uniform snowcover distribution and local microclimate conditions.

Rapid snowmelt and a lack of lateral hydrological linkages led to a rapid decline in water levels in small isolated ponds (e.g. C8) during all seasons and led to the desiccation of surface waters in 2007 and 2008. Similar findings were reported by Abnizova et al. (2012b) at this site, and for ponds located in a small wetland near Creswell Bay, Somerset Island (Abnizova and Young, 2010). Smol and Douglas (2007) similarly surmised that high rates of evaporation and low amounts of summer precipitation were the cause of several small lakes drying up on the NE coast of Ellesmere Island. This phenomenon was also observed in small lakes located in Spitzbergen, Norway (Mazurek et al., 2012).

Changes in the water tables of frost cracks at PBP showed that these small water bodies typically experienced large fluctuations in water levels as a result of evaporation (e.g. CF19-20). Initial ordination analyses often showed frost cracks as outliers compared to the rest of the pond groups due to large differences in some chemical and physical characteristics that caused them to be situated at the extreme end of certain gradients (e.g. very high DOC concentrations in the spring of 2010) and thus were eventually removed from the analyses. The small surface area and shallow depth of these frost cracks likely created conditions in which unique chemical and physical characteristics developed. Similar findings were reported by Thompson and Woo (2009) for a site on the Fosheim Peninsula, Ellesmere Island.

Ponds with good seasonal water level stability during all four study seasons were typically located in the lee of a slope and received more snow than exposed ponds. These ponds received meltwater from a lingering snowbed adjacent to the ponds or due to their connection to a hillslope stream draining into the sector of the wetland which was found within 500 m from the hillslope where they were situated (Fig. 1). These waters helped to keep pond water tables elevated. These ponds experienced a decline in water tables later in the season with the gradual cessation of this linkage and its water supply. For example, water levels in linked ponds declined as the snowbank shrunk and the creek dried up (C1 and C14 respectively), highlighting the critical role that these types of hydrological linkages play in these wetland ecosystems.

A difference in size (large vs. small) did not affect water table stability in these ponds between the seasons. Higher amounts of precipitation in 2008 and frequent episodes of rainfall in 2009 raised water tables in both years. These ponds also received continual water inputs from the snowbed and stream throughout the summer, offsetting any evaporation losses that might have occurred in 2008 and 2009. Similar results have been reported by others elsewhere in the High Arctic (Woo and Guan 2006; Abnizova and Young, 2010; Young and Abnizova, 2011).

Differences in pond hydrology are also reflected in stable water isotopic composition and the seasonal evolution of isotopic values in the *Control Ponds*. The isolated ponds tend to have a more enriched isotopic composition demonstrating the negative water balance conditions resulting from evaporation losses. Predominantly depleted isotopic compositions in the ponds that are hydrologically connected to a source (e.g. late-lying snowbed, a hillslope creek) characterise the conditions of positive water balances. In 2007, 2009 and 2010 study seasons the isolated *Control Ponds* have enriched isotopic compositions even during snowmelt which is opposite to the hydrologically connected *Control Ponds* plotting close to the GMWL. This unique separation of ponds with strong hydrologic linkages to their catchment from the isolated ponds lasting until the early post-snowmelt season was observed during both the dry 2007 and 2010 seasons and the wet season of 2009. Similarities in slopes of regressions between δD and $\delta^{18}O$ of hydrologically connected and isolated ponds in 2007 point to evaporation as the main loss of pond water during this dry season. Abnizova et al. (2012b) also analysed stable water isotope signatures in the isolated polygonized and thermokarst ponds in Lena Delta, Russia and found similar signatures in the hydrologically isolated ponds. This distinction is lost during the snowmelt season in 2010 which shows dramatic differences in the slopes of regression between the hydrologically linked and isolated ponds, and demonstrates the dominant role of inflow in the water balance of the linked pond. Slightly different slopes of regression between δD and $\delta^{18}O$ in the linked and isolated ponds (estimates equal to 6.88 and 7.53) in 2009 point to precipitation as the main source of water to these ponds during this season.

Similar patterns of separation were observed in the isotopic compositions in the *Other Control Ponds* that were indentified as hydrologically linked or isolated based on our field observations and the intensive hydrological monitoring.

Isotope analysis suggests that most *Satellite Ponds* have predominantly isotopically enriched values similar to the evaporative signatures of the isolated *Control Ponds*. During snowmelt these ponds had similar values to those from the hydrologically linked and isolated *Control Ponds* which tend to overlap due to the meltwater inflows into all ponds at PBP.

The linear discriminant analysis confirmed that our classification of ponds across PBP was correct with most classification count values reflecting the correct group of ponds (linked or isolated). The LDA analysis of pond grouping from fall 2008 and 2009 showed slightly lower predicted values for hydrologically connected ponds (92% and 75%, respectively) and 95% for isolated ponds in fall 2009. This indicates that some hydrologically linked pond may have reduced connectivity to their catchment late in the season and therefore experience negative water balances similar to the isolated ponds. Based on the isotopic separation from the dry study season in 2007, hydrologically connected ponds can experience isotopic signatures like the isolated ponds, and this is associated with a cessation of hydrologic inflows from linkages.

Overall, the results from the intensive hydrologic monitoring of the *Control Ponds* demonstrate that for the most part ponds at PBP are distributed into two distinct hydrological groupings. These are predominantly identified by the presence or absence of an external hydrologic linkage to pond catchment. The extent of this hydrological connectivity is defined by the topographic location of the ponds, and in relation to their proximity to hillslopes. The strength of hydrologic connectivity varies between the seasons and inter-annually and is controlled by a set of hydro-climatic factors including the end-of-winter snow accumulation, frequency and intensity of rain events during the open water season, and hydro-climatic conditions from the preceding season, and the previous year. This suggests that while the ponds show the hydrologic characteristics of the linked or isolated types based on their topographic location in the wetland, their hydrologic responses to variability in hydro-climatic conditions can place them in an "overlapping" group identified in our water isotopic framework. This pattern can be seen in some hydrologically connected *Control Ponds* that show similar hydrologic dynamics to the isolated ponds which becomes pronounced during years with warmer than average air temperatures, lower inputs of summer precipitation (e.g. 2007).

Turner et al. (2010) also developed an isotopic framework for lakes in the Old Crow Flats landscape in northern Yukon Territory, Canada and identified 5 categories of lakes based on the dominant hydrologic process controlling their water balance: snowmelt-dominated, rainfall-dominated, groundwater-influenced, evaporation-dominated and drained lakes. Similar to our study, Turner et al. (2010) found that the isolated water bodies receiving inputs from precipitation only (rainfall-dominated) show the highest fluctuations in their water levels and will be subject to evaporation (evaporation-dominated) during dry open water seasons. Similar findings were reported in other arctic ponds situated in a smaller high arctic wetland near Creswell Bay, Somerset Island, Nunavut (Abnizova and Young, 2010; Young and Abnizova, 2011). Seasonal changes in a pond's surface area and depth depend on a pond's hydrological sources (types, duration), along with rates of evaporation and ground thaw.

Geochemistry of surface waters at PBP from 2007-2010

Seasonal changes in pond water balance define the dynamics of physico-chemical parameters in wetland ponds. The year-to-year and seasonal variability in precipitation and temperature affects the quantities of lateral inflow into the ponds, variation in surface area and depth, and their hydro-chemical properties.

General geochemical properties

Chemical analysis of water samples indicates that the water composition of ponds across PBP is predominantly a calcium bicarbonate type and is classified as a Ca+Mg, Na+K and HCO_3, Cl+SO_4 hydrogeochemical facie based on the classification by Kehew (2011). Analysis of stable water isotope signatures in all pond water samples from the 2009 observation period showed their typical chemical composition of snow and rainfall have undergone alteration due to evaporation. Here, linked ponds typically have lower evaporative signatures than isolated ponds. Chemical analysis shows that among major ions, Ca and HCO_3 ions were dominant. Examination of major cation concentrations in one of the large lakes also showed the dominance of calcium cations with average values reaching 35.1 mg L^{-1} in the 2008 observation period. The dominance of Ca and HCO_3 ions in all ponds and frost cracks reflects the presence of calcium-carbonate rich limestone bedrock or dolostone in this region. The chemical reactions, such as the carbonation of carbonates, typically release calcium ions, which are a product of the weathering of carbonate rocks and gypsum in bedrock (Mazurek et al., 2012). High concentrations of these dissolved solids could also be attributed to the argillaceous limestone formations draining large amounts of $CaCO_3$ into these particular sites (Edlund, 1990). Similar findings were reported for Canadian arctic ponds on Ellesmere Island (Douglas and Smol, 1994) and elsewhere on Bathurst Island (Lim et al., 2001). Results of the spring 2010 PCA showed that Ca, HCO3, Mg, Ba and conductivity were responsible for much of the variability between ponds. This corresponds well with Lim et al. (2001) who also discovered that major ion content, particularly Ca and Mg, in addition to dissolved organic carbon and pH, were the main drivers of variation among their survey of 38 ponds on Bathurst Island, late July 1994. Measured values of conductivity in water are affected by the presence of inorganic dissolved solids such as Cl, NO_3, SO_4, and PO_3 anions or Na, Mg, Ca, Fe, and Al cations (Boman et al., 2011).

Ca and HCO_3 are the predominant ions in all wetland waters at PBP. The concentration values of other major cations and anions differ and are associated with pond hydrological settings and seasonal differences in climate. The chemical composition of hydrologically linked ponds shows the results of the sole mixing of Ca and Mg, whereas isolated ponds' waters cannot be characterised by such simple mixing. Proportions of Mg and Na cations and Cl and SO_4 anions in these ponds are different and vary among isolated ponds.

Further evaluation of seasonal differences in hydro-chemical concentrations identified two hydro-chemical trends in meq L^{-1}: 1) Ca > Na > Mg > K (*Satellite* Ponds) and Ca > Mg ~ Na > K (*Control* Ponds) derived from two complete season averages (May-September, 2008 and 2009) and summer 2007; and 2) Ca > Mg > Na > K derived from a shorter 2010 observation period (May-June). The Ca > Na > Mg > K signature is very similar to the relative concentrations of major cations found elsewhere on this island (Lim et al., 2005), in the Canadian High Arctic (Michellutti et al., 2002; Antoniades et al., 2003a), Subarctic (Pietitz et al., 1997ab; Ruland et al., 2003), and in Arctic Russia (Duff et al., 1999).

Geochemical trends during wet and cool seasons

The observed differences in the hydro-geochemical properties of study ponds may be shaped by different characteristics and are strongly affected by atmospheric deposition, chemical weathering, surficial processes, periglacial activity, and biological inputs (Mazurek et al., 2012). The investigated ponds were located at elevations of less than 25 m a.s.l., however at various distances from the seashore varying from less than 3 to 10 km for centrally located ponds (C1-C21). Therefore, the chemistry in these ponds might reflect the effects of marine aerosols and sea sprays. Ions of marine origin, such as Cl and Na ions, are delivered to the coastal wetlands typically through precipitation (Mazurek et al., 2012). With the exception of three ponds (C10, S28 and S36 with significant difference), water samples from all of the ponds that were investigated in 2010 showed that the ratio of Cl to Na exceeded the widespread sea water level of 1.14 (Mazurek et al., 2012). High amounts of precipitation in 2008 and 2009 may have contributed to NaCl enrichment, explaining the elevated concentrations of Na. For example, measured rain concentrations in 2009 samples ranged from 1.56 – 41.63 mg L^{-1} (Na) and 0–38.22 (mg L^{-1}) (Cl). Similarly, Lim et al. (2005) reported high Cl concentration in ponds on Banks Island and explained it by the proximity of the study site to the coast, and receipt of marine aerosols. A similar finding was reported from Axel Heiberg Island (Michelutti et al., 2002a). Interestingly, Antoniades et al. (2003) also found this trend in ponds at Isachsen and Ellef Ringnes Island but suggested that it was driven by local soil characteristics. Similarly, the mineral substrate at PBP contains evaporitic gypsum deposits, which could have contributed to elevated levels of Cl. This region of the High Arctic tends to experience very poor weather (fog, rain). Its proximity to coastal sea-waters, suggests that rainwater falling on the ground would be enriched with marine salts (Cl and Na).

Geochemical trends during dry and warm seasons

The Ca>Mg>Na>K signature was observed in PBP ponds during the 2010 observation season. Evaluation of the near-surface water samples revealed that hydrologic inflows in ponds connected to their catchment can transport a significant pool of solutes, which will then be released into the active layer during periods of active ground thaw. To highlight the magnitude of solutes contained in surface waters, we monitored concentrations of major cations at a depth of approximately 75 cm in a well near pond C1 in 2009. The water exhibited high concentrations of major cations including Ca, which ranged from 56.8 to 126.4 mg L^{-1}, and Mg, which ranged from 11.5 to 26.9 mg L^{-1}. In addition, relatively high concentrations of Na and K cations in subsurface waters (Na, 6.5-9.2 mg L^{-1} and K, 0.8-5.2 mg L^{-1}) indicated that ponds connected to their catchments probably receive elevated concentrations of these solutes. Therefore, the 2010 Ca>Mg>K>Na chemical trend can be explained by pond hydrologic connectivity to its catchment which provided higher Mg concentrations than Na in solute enriched lateral inflows. Similar findings to ours were reported for ponds in Spitzbergen, where high concentrations of Ca, Mg, HCO_3, and SO_4 were found in suprapermafrost water (Mazurek et al., 2012).

The Ca>Mg>K>Na trend was not apparent in the warm and dry 2007 observation period. Hydro-chemical mapping and pond monitoring was limited to only the month of July and only 15 ponds, of which 6 satellite ponds were visited twice. Warm conditions and the complete absence of precipitation contributed to evapo-concentration and therefore enriched solute concentrations. In comparison to Na, the lower average concentration values of Mg can be attributed to the cessation of hydrologic connectivity to the catchment and limited contribution of weathered Mg cations typically arriving from suprapermafrost inflows.

The average seasonal K concentrations were lower in the linked ponds than in the isolated ponds, which are typically prone to frequent desiccation, and have negative water balances that are intensified during drier and warmer seasons. Therefore, high average K concentrations in the isolated ponds result from the evapo-concentration process (Benoy et al., 2007). Previous studies have related variability in K concentrations to the vegetation characteristics of a catchment (Douglas and Smol, 1994; Lim et al., 2001) since leaching from vascular plants can affect the K concentrations (Prentiki et al., 1980).

The high concentrations and wide range (0.0-5.9 mg L^{-1}) of K in PBP ponds indicate that catchment vegetation and connectivity play a major role in affecting K concentrations. This is especially evident for the ponds with well known hydrologic links (C1-C6), as well as two ponds (e.g. C6, C8) and one frost crack (CF18) which contained hydrophilic vegetation (Fig. 2). An additional source of K could be the enhanced plant growth observed in isolated ponds, as this was typically seen during warm and dry periods.

Physico-chemical signatures of hydrologic connectivity

SO_4 ions in ponds at PBP probably originate from the interaction of suprapermafrost waters coming in contact with anhydrite (calcium sulphate) naturally found in gypsum (Hodson, 1989) which then undergoes dissolution. The occurrence of SO_4 was indicated by the increase in SO_4 levels in ponds over time and further supported by representation of the Ca-Mg-HCO_3-SO_4 hydro-chemical group in the Piper diagram (Fig. 7). No estimates of SO_4 anions were made in 2007-2009. Analysis of 2010 hydro-chemical mapping revealed that prevalence of SO_4 over Cl in pond waters was dependent upon the presence or absence of a hydrological linkage in the investigated ponds. Ponds that were hydrologically connected had typically higher average SO_4 concentrations than Cl. Analysis of hydro-chemical data in runoff waters from a late-lying snowbed and hillslope creek showed high SO_4 concentrations (23.9 mg L^{-1} and 16.0 mg L^{-1}, respectively). Our records further indicate that hydrologic linkages typically transfer higher SO_4 amounts later in the season (e.g. 32.2 mg L^{-1}) than snowmelt levels (e.g. 5.1 mg L^{-1}), possibly due to bedrock weathering (Pienitz et al., 1997a). Although the satellite ponds were not continually monitored throughout the season, their low SO_4 concentrations (< 1.4 mg L^{-1} excluding the east pond sites) could be attributed to their hydrologic isolation or patchy distribution of gypsum in limestone deposits.

Analyses of stable water isotopes showed that all *Satellite Ponds*, with exception of the eastern sector, had strong signatures of evaporation and have an isotope composition that is similar to the isolated control ponds. A possibility of hydrological connectivity was observed in the water levels of an eastern pond (S23) when compared to western (S38) and southern ponds (S33), showing an increasing trend even after spring meltwaters ceased. High SO_4 concentrations ranging from 2.43–5.02 mg L^{-1} indicates that these ponds might have been receiving water in the post-snowmelt season. These three eastern ponds are located near the base of a hillslope where a large late-lying snowbed was located (Fig. 2). It is reasonable to expect that meltwaters from this snowbed were the source for these waters. Results from the 2010 PCA also suggest that the east ponds (S22 and S23) may be connected as they were plotted closer to the *Control ponds* than the *Satellite ponds*. Other studies also found high SO_4 concentrations in lakes which were located in a landscape with prevalent limestone-dolomite, sandstone, siltstone, and conglomerate rock (Pienitz et al., 1997b), much like the one at PBP.

Dissolved organic carbon

Dense vegetation cover in pond catchments provides a substantial source of biogenic components. Abnizova et al. (2012) and Abnizova et al. (submitted) found higher amounts of DOC in ponds that are hydrologically connected to their vegetated catchment long after the snowmelt connectivity is over. Similar findings were reported by Lim et al. (2001), reporting higher DOC concentrations in Canadian High Arctic ponds with vegetated catchments. In comparison to other studies, average DOC concentrations reported for unvegetated arctic tundra sites by Pienitz et al. (1997a) were the lowest (1.7–3.8 mg L^{-1}), whereas they were the highest (3.8–29.9 mg L^{-1}) in boreal forest and forest-tundra sites. Lim (1999) similarly illustrated that all ponds with high DOC values from the arctic tundra sites had waters which drained through relatively vegetated catchments. In this study the highest DOC value was reported in a hydrologically dynamic pond, one where the water level fluctuated in response to precipitation inputs and water losses (max 41.6 mg L^{-1}). Similar findings were recently reported by Abnizova et al. (2012a), Barley et al. (2006) (maximum value 31.3 mg L^{-1}), and Duff et al. (1999) (max = 47.0 mg L^{-1}). Lim et al. (2001) reported low to moderate DOC concentrations on Bathurst Island (1.1-11.2 mg L^{-1}), however their sampling scheme allowed for only one pond at PBP to be tested. Moreover, their cross-island survey contained catchments with limited vegetation. This differs from Abnizova et al. (2012a) who reported that most ponds at PBP are all typically connected to their vegetated catchments during the snowmelt period. Other ponds may be located so as to receive additional meltwaters from late-lying snowbeds or hillslope creeks that drain waters from the uplands (Young et al., 2010). Abnizova et al. (2012a) also verified that ponds at PBP typically contain terrestrial DOC based on fluorescence index analysis. Pienitz et al. (1997a) also found a significant correlation between watershed vegetation and DOC levels.

Conclusions

We conducted hydrologic and hydro-chemical surveys of 51 small water bodies located in Polar Bear Pass, a low-gradient extensive High Arctic wetland from 2007-2010. The methodology of this study included 1) analysis of seasonal and inter-seasonal pond hydrology and development of a pond water isotopic framework based on the research findings from the *Control Ponds* at PBP, and 2) analysis of the isotopic compositions from the *Satellite Ponds* superimposed on the developed framework to categorize these pond into two hydrological groupings: hydrologically linked or isolated ponds. The hydro-chemical surveys showed that the PBP ponds have large seasonal variability in physico-chemical parameters which is linked to the unique hydrological settings of ponds in this wetland. Presence of hillslopes in this area creates a distinctive physiographic setting which promotes a unique set of hydrological linkages between the hillslopes and wetland ponds extending to 500 meters in distance creating a *hydrologic buffer zone*. In this *hydrologic buffer zone* PBP ponds often form either permanent or episodic links with various types of hillslope sources like late-lying snowbeds, creeks and upslope ponds. The resulting differences in hydrological inputs to the ponds establishes an effective mechanism that ensures positive pond water balances, stable and elevated water levels all of which creates a distinct set of hydrological characteristics unique to these connected ponds. Based on the results of this study we have developed a conceptual diagram of PBP wetland ponds which with the help of the topographic map (Fig. 1) identifies the ponds that are hydrologically connected to their catchment and the isolated ponds located centrally in the wetland (Fig. 9).

The results of this intensive hydro-chemical study expand the seasonal range of limnological data sets from the High Arctic. This study reveals the importance of understanding the hydrology of wetland ponds between seasons and various years. It also demonstrates the control that hydrological settings may exert on physico-chemical properties of pond waters. The trends in major ion concentration, pH, and conductivity values for the wetland ponds are based on four consecutive observation periods, a comprehensive seasonal monitoring scheme which is unusual for most scientific studies in arctic regions. An analysis of the geochemical properties of the pond waters across PBP wetland indicated:

1) the main differences in proportions of major ions reflect the lithology of bedrock material found on Bathurst Island. However, the trends in geochemistry of the investigated ponds varied depending on the hydrological status and seasonal climatic conditions;

2) the high estimates of major ions, pH, and conductivity in hydrologically connected ponds were typically evident towards the end of arctic summer (beginning of August). These patterns resulted from continuous lateral inflows of surface and subsurface waters enriched in major ions from bedrock weathering processes and organic matter from densely vegetated catchments. These patterns illustrated the importance of hydrologic linkages to ponds and frost cracks on a temporal scale;

3) a more distinct pattern to (2) in chemical properties was observed for isolated ponds indicating their hydrologic sensitivity to changes in temperature and precipitation regimes. Here, maximum concentrations typically occurred during drought episodes, corresponding with warm air temperatures.

Finally, this study strongly suggests that both seasonality and climate conditions must be considered when interpreting the physico-chemical characteristics of arctic ponds. Wetland researchers monitoring water quality are also cautioned that their sampling protocol should take into account the spatial variability of the terrain, as well as the local climate.

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References cited

Abnizova, A. K. L. Young and M. J. Lafrenière. 2012a: Pond hydrology and dissolved carbon dynamics at Polar Bear Pass wetland, Bathurst Island, Nunavut. *Ecohydrology,*doi: 10.1002/eco.1323.

Abnizova A., Young K.L., and Boike, J. Snowmelt hydrology and carbon dynamics in wetland ponds at Polar Bear Pass, Bathurst Island, Nunavut, Canada. *Ecosystems* (submitted).

Abnizova, A., Siemens, J., Langer, M., and Boike, J., 2012b: Small ponds with major impact: The relevance of ponds and lakes in permafrost landscapes to carbon dioxide emissions. *Global Biogeochemical Cycles*, 26, GB2041, doi:10.1029/2011GB004237.

Abnizova, A., and Young, K. L., 2010: Sustainability of High Arctic ponds in a polar desert environment. *Arctic,* 63: 67-84.

Antoniades, D., Douglas, M. S. V., and Smol, J.P., 2000: Limnology and autecology of freshwater diatoms from Alert, northern Ellesmere Island, Nunavut. *Proceeding of the 6th National Student Conference on Northern Studies*: 1–11.

Antoniades, D., Douglas, M. S. V., and Smol, J. P., 2003: The physical and chemical limnology of 24 ponds and one lake from Isachsen, Ellef Ringnes Island, Canadian High Arctic. *International Review of Hydrobiology*, 88: 519–538.

Assini, J., and Young, K. L., 2012: Snow cover and snowmelt of an extensive High Arctic wetland: spatial and temporal seasonal patterns. *Hydrological Sciences Journal*, 57: 738-755.

Babb, T. A., and Bliss, L. C., 1974. Susceptibility to environmental impact in the Queen Elizabeth Islands. *Arctic*, 24: 234-237.

Barley, E. M., Walker, I. R., and Kurek, J., 2006: A northwest North American training set: distribution of freshwater midges in relation to air temperature and lake depth. *Journal of Paleolimnology*, 36: 295–314.

Blake, W. Jr., 1964: Preliminary account of the glacial history of Bathurst Island, Arctic Archipelago. *Canada Geological Survey,* Paper 64-30.

ter Braak, C. J. F., and Šmilauer, P., 2002: *Canoco Reference Manual and CanoDraw for Windows User's Guide: Software for Canonical Community Ordination (Version 4.5)*. Ithaca, New York: Microcomputer Power.

Breton, J., Vallières, C., and Laurion, I., 2009: Limnological properties of permafrost thaw ponds in northeastern Canada. *Canadian Journal of Fisheries and Aquatic Sciences*, 66: 1635-1648.

Boman, J.B., Wilson, P.C., and Ontermaa, E.A., 2011. Understanding Water Quality Parameters for Citrus Irrigation and Drainage Systems. *Circular*, 1406: 1-11.

Carroll, M.L., Townshend, J.R.G., DiMiceli, C.M., Loboda, T., and Sohlberg, R.A. 2011: Shrinking lakes of the Arctic: Spatial relationships and trajectory of change. *Geophysical Research Letters*, 38: L20406, doi:10.1029/2011GL049427.

Criss, R. E., and Davisson, M. I., 1996: Isotopic imaging of surface water/groundwater interactions, Sacramento Valley, California. *Journal of Hydrology*, 178: 205-222.

Dickman, M., and Ouellet, M., 1987: Limnology of Garrow Lake, N.W.T., Canada. *Polar Record*, 23: 531-549.

Douglas, M. S. V., and Smol, J. P., 1994: Limnology of high arctic ponds (Cape Herschel, Ellesmere Island, N. W. T.) *Archiv für Hydrobiologie*, 131: 401–434.

Douglas, M. S. V., and Smol, J. P., 2000: Eutrophication and recovery in the High Arctic; Meretta Lake (Cornwallis Island, Nunavut, Canada) revisited. *Hydrobiologia* 431: 193–204.

Danks, H.V., 1974: Spring and early summer temperatures in a shallow arctic pond. *Arctic,* 24: 113-123.

Duff, K. E., Laing, T. E., Smol, J. P., and Lean, D. R. S., 1999: Limnological characteristics of lakes located across arctic treeline in northern Russia. *Hydrobiologia*, 391: 205-222.

Downing, J. A., 2009: Global limnology: up-scaling aquatic services and processes to the planet Earth. *Verhandlungen der Internationale Vereinigung für Theoretische und Angewandte Limnologie*, 30: 1149-1166.

Edlund, S. A., Alt, B. T., and Young, K.L., 1989: Interaction of climate, vegetation, and soil hydrology at Hot Weather Creek, Fosheim Peninsula, Ellesmere Island, Northwest Territories. *Current Research, Part D, Geological Survey of Canada Paper,* 89- 1D, 125-133.

Edlund, S. A., 1990: Bioclimatic zones in the Canadian Arctic Archipelago. *In* Harrington, C. R. (ed.), *Canada's missing dimension - science and history in the Canadian Arctic Islands.* Ottawa: Canadian Museum of Nature, 421-441.

Ferguson, M. A. D., 1987: Status of Peary caribou and muskox populations on Bathurst Island, N.W.T., August 1981. *Arctic*, 40:131–137.

Flanagan, K. M., McCauley, E., Wrona, F., and Prowse, T., 2003: Climate change: the potential for latitudinal effects on algal biomass in aquatic ecosystems. *Canadian Journal of Fisheries and Aquatic Sciences*, 60: 635–639.

Gregory-Eaves, I., Smol, J. P., Finney, B. P., Lean, D. R. S., and Edwards, M., 2000: Characteristics and variation in lakes along a north-south transect in Alaska. *Archiv für Hydrobiologie*, 147: 193–223.

Hamilton, J. D., Kelly, C. A., Rudd, J. W. M., Hesslein, R. H., and Roulet, N.T., 1994: Flux to the atmosphere of CH_4 and CO_2 from wetland ponds on the Hudson-Bay Lowlands (HBLs). *Journal of Geophysical Research - Atmospheres*, 99: 1495–1510.

Hamilton, P. B., Gajewski, K., Atkinson, D.E., and Lean, D.R.S., 2001: Physical and chemical limnology of 204 lakes from the Canadian Arctic Archipelago. *Hydrobiologia*, 457: 133–148.

Hobbie, J.E. (ed.), 1980: *Limnology of Tundra Ponds: Barrow, Alaska.* Stroudsburg, PA: Dowden, Hutchinson and Ross. 514 pp.

Hobbie, J.E., Deegan, L.A., Peterson, B.J., Rastetter, E.B., Shaver, G.R., Kling, G.W., O'Brien, W.J., Chapin, F.S. III, Miller, M.C., Kipphut, G.W., Bowden, W.B., Hershey, A.E., and McDonald, M.E., 1995: Long-term measurements at the Arctic LTER site. *In* Powell, T.M. and Steele, J.H. (eds.), *Ecological Time Series.* New York: Chapman and Hall, 391-409.

Jonsson, A., Karlsson, J., and Jansson, M., 2003: Sources of carbon dioxide supersaturation in clearwater and humic lakes in northern Sweden. *Ecosystems*, 6: 224-235.

Kalff, J., and Welch, H. E., 1974: Phytoplankton production in Char Lake, a natural polar lake, and in Meretta Lake, a polluted polar lake, Cornwallis Island, Northwest Territories. *Journal of the Fisheries Research Board of Canada*, 31: 621-636.

Keatley, B. E., Douglas, M. S. V., and Smol, J. P., 2007: Physical and chemical limnological characteristics of lakes and ponds across environmental gradients on Melville Island, Nunavut/N.W.T., High Arctic Canada. *Fundamental and Applied Limnology*, 168: 355 – 376.

Kerr, J. W., 1974: Geology of Bathurst Island Group and Byam Martin Island, Canada. Geological Survey of Canada, Dept. of Energy, Mines and Resources. Memoir 378, pp
152.

Kehew, A.E., *Applied Chemical Hydrogeology.* Prentice Hall, Upper Saddle River, New Jersey, 2001.

Kumke, T., Ksenofontova, M., Pestryakova, L., Nazarova, L., and Hubberten, H.-W., 2007: Limnological characteristics of lakes in the lowlands of Central Yakutia, Russia. *Journal of Limnology*, 66: 40-53.

Laurion. I, Vincent W.F., MacIntyre, S., Retamal, L., Dupont, C., Francus, P. and Pienitz, R., 2010: Variability in greenhouse gas emissions from permafrost thaw ponds. *Limnology and Oceanography*, 55: 115-133.

Leps, J., and Šmilauer, P., 2003. *Multivariate Analysis of Ecological Data using CANOCO.* Cambridge U.K.: Cambridge University Press. 284 pp.

Lim, D. S. S., Douglas, M. S. V., Smol, J. P., and Lean, D. R. S., 2001: Physical and chemical limnological characteristics of 38 lakes and ponds on Bathurst Island, Nunavut, Canadian High Arctic. *International Review of Hydrobiology*, 86: 1–22.

Lim, D. S. S., Douglas, M. S. V., and Smol, J. P., 2005: Limnology of 46 lakes and ponds on Banks Island, N.W.T., Canadian Arctic Archipelago. *Hydrobiologia*, 545:11-

32.
Lim, D. S. S., and Douglas, M. S. V., 2003: Limnological characteristics of 22 lakes and ponds in the Haughton Crater region of Devon Island, Nunavut, Canadian High Arctic. *Arctic, Antarctic, and Alpine Research*, 35: 509–519.
Lougheed, V.L., Butler, M.G., McEwen, D.C., and Hobbie, J.E. 2011: Changes in Tundra Pond Limnology: Re-sampling Alaskan Ponds After 40 Years. *Ambio,* 40: 589-599.
Mackay, J.R., 1964: Field observation of patterned ground. *Journal of Geology,* 72: 378-380.
Mazurek, M., Paluszkiewicz, R., Rachlewicz, G., and Zwoliński, Z., 2012: Variability of Water Chemistry in Tundra Lakes, Petuniabukta Coast, Central Spitsbergen, Svalbard. *The Scientific World Journal*, doi:10.1100/2012/596516.
Meyer, H., Schonicke, L., Wand, U., Hubberten, H. W., Friedrichsen H., 2000: Isotope studies of hydrogen and oxygen in ground ice—experiences with the equilibration technique. *Isotopes* in *Environmental* and *Health Studies,* 36: 133–149.
Miller, N.G., and Ireland, R. R., 1978: A floristic account of the bryophytes of Bathurst Island, Arctic Canada. *Occasional Papers of the Farlow Herbarium of Cryptogamic Botany,* No. 13: 1-38.
Miller, M.C., Hater, G.R., Spatt, P., Westlake, P., and Yeakel, D., 1986: Primary production and its control in Toolik Lake, Alaska. *Archivs fur Hydrobiologie Supplement*, 74(1): 97-131.
Miller, F. L., 1991: Estimating Bathurst Island Peary caribou and muskox populations. *Arctic*, 44: 57 – 62.
Michelutti, N., Douglas, M. S. V., Muir, D. C. G., and Smol, J. P., 2002a: Limnological characteristics of 38 lakes and ponds on Axel Heiberg Island, High Arctic Canada. *International Review of Hyrobiology*, 87: 385–399.
Michelutti, N., Douglas, M. S. V., Lean, D. R. S., and Smol, J. P., 2002b: Physical and chemical limnology of 34 ultra-oligotrophic lakes and ponds near Wynniatt Bay, Victoria Island, Arctic Canada. *Hydrobiologia*, 48: 1–13.
Muster, S., Langer, M., Heim, B., Westermann, S., and Boike, J., 2012: Subpixel heterogeneity of ice-wedge polygonal tundra: a multi-scale analysis of land cover and evapotranspiration in the Lena River Delta, Siberia. *Tellus B*, 64: 17301, doi: 10.3402/tellusb.v64i0.17301
Muster, S., Heim, B., Abnizova, A., and Boike, J., 2013: Water body distribution across scales: a comparison of three Arctic wetlands. *Remote Sensing*, 5(4):1498-1523.
Perren, B. B., Bradley, R. S., and Francus, P., 2003: Rapid lacustrine response to recent high Arctic warming: a diatom record from Sawtooth Lake, Ellesmere Island, Nunavut. *Arctic Antarctic and Alpine Research*, 35: 271–78.
Peterson, B. J., Hobbie, J. E., Hershey, A., Lock, M., Ford, T., Vestal, R., Hullar, M., Ventullo, R., and Volk, G., 1985: Transformation of a tundra river from heterotrophy to autotrophy by addition of phosphorus. *Science*, 229:1383-1386.
Peterson, B.J., Hobbie, J.E., and Corliss, T.L., 1986: Carbon flow in a tundra stream

ecosystem. *Canadian Journal of Fisheries and Aquatic Sciences*, 43(6): 1259-1270.

Prentiki, R. T., Miller, M. C., Barsdate, R. J., Alexander, V., Kelley, J., and Coyne, P., 1980. *Chemistry*, pp. 76-178. In: J. E. Hobbie (ed.), *Limnology of Tundra Ponds, Barrow, Alaska*. Stroudsburg, PA: Dowden, Hutchinson and Ross, 514 pp.

Pienitz, R., Smol, J. P., and Lean, D. R. S., 1997a: Physical and chemical limnology of 24 lakes located between Yellowknife and Contwoyoto Lake, Northwest Territories (Canada). *Canadian Journal of Fisheries and Aquatic Sciences*, 54: 347–358.

Pienitz, R., Smol, J. P., and Lean, D. R. S., 1997b: Physical and chemical limnology of 59 lakes located between the southern Yukon and the Tuktoyaktuk Peninsula, Northwest Territories (Canada). *Canadian Journal of Fisheries and Aquatic Sciences*, 54: 330–346.

R Development Core Team, 2008: R: A language and environment for statistical computing. R Foundation for Statistical Computing, Vienna, Austria. ISBN 3-900051-07-0, URL http://www.R-project.org.

Ramlal P, Hesslein SRH, Hecky RE, Fee EJ, Rudd JWM, Guildford SJ. 1994. The organic carbon budget of a shallow arctic tundra lake on the Tuktoyaktuk Peninsula, N.W.T., Canada: arctic lake carbon budget. *Biogeochemistry* 24: 145-172.

Rautio, M., Dufresne, F., Laurion, I., Bonilla, S., Vincent, W.F., and Christoffersen, K., 2011: Shallow freshwater ecosystems of the circumpolar Arctic. *Écoscience* 18: 204-222.

Rautio, M., Vincent, W.F., 2006: Benthic and pelagic food resources for zooplankton in shallow high-latitude lakes and ponds. *Freshwater Biology*, 51: 1038-1052.

Sheard, J.W. and D.W. Gale. 1983a. Vegetation studies at Polar Bear Pass, Bathurst Island, N.W.T. Classification of plant communities I. *Canadian Journal of Botany*, 61: 1618-1636.

Sheard, J.W. and D.W. Gale. 1983b. Vegetation studies at Polar Bear Pass, Bathurst Island, N.W.T. Vegetation-environment relationships II. *Canadian Journal of Botany*, 61: 1637-1646.

Schindler, D.W., Welch, H.E., Kalff, J., Brunskill, G.T., and Kritsch, N., 1974a: Physical and chemical limnology of Char Lake, Cornwallis Island (75°N Lat.). *Journal of theFisheries Research Board of Canada*, 31: 585-607.

Schindler, D. W., Kalff, J., Welch, H. E., Brunskill, G. J., Kling, H., and Kritsch, N., 1974b: Eutrophication in the high arctic-Meretta Lake, Cornwallis Island (75°N Lat). *Journal of the Fisheries Research Board of Canada*, 31: 647-662.

Smith, L. C., Sheng, Y., MacDonald, G. M., and Hinzman, L. D., 2005: Disappearing Arctic lakes. *Science*, 308: 1429.

Smol, J.P., Walker, I.R. and Leavitt, P.R. 1991. Paleolimnology and hindcasting climatic trends.
Verhandelungen der Internationalen Verein für Limnologie, 24:1240-1246.

Smol, J. P., and Douglas, M. S. V., 1996: Long-term environmental monitoring in Arctic lakes and ponds using diatoms and other biological indicators. *Geoscience*

Canada, 23: 225-230.
Smol, J.P., 2005. Tracking long-term environmental changes in arctic lakes and ponds: a paleolimnological perspective. *Arctic,* 58 (2): 227-229.
Smol, J. P., Wolfe, A. P., Birks, H. J. B., Douglas, M .S. V., Jones, V. J, Korhola, A., Pienitz, R., Rühland, K., Sorvari, S., Antoniades, D., Brooks, S. J., Fallu, M-A., Hughes, M., Keatley, B. E., Laing, T. E., Michelutti, N., Nazarova, L., Nyman, M., Paterson, A. M., Perren, B., Quinlan, R., Rautio, M., Saulnier-Talbot, É, Siitonen, S., Solovieva, N., and Weckström, J., 2005: Climate-driven regime shifts in the biological communities of arctic lakes. *Proceedings of the National Academy of Sciences,* 102: 4397-4402.
Smol, J. P., and Douglas, M. S. V., 2007a: Crossing the final ecological threshold in High Arctic ponds. *Proceedings of the National Academy of Sciences,* 104: 12395 – 12397.
Smol, J.P., and Douglas, M.S.V., 2007b: From controversy to consensus: making the case for recent climatic change in the Arctic using lake sediments. *Frontiers in Ecology and the Environment* 5: 466-474.
Thompson, D. K., and Woo, M. K., 2009: Seasonal hydrochemistry of a High Arctic wetland complex. *Hydrological Processes,* 23: 1397-1407.
Tranvik, L. J., Downing, J. A., Cotner, J. B., Loiselle, S. A., Striegl, R. G., Ballatore, T. J., Dillon, P., Knoll, L. B., Kutser, T., Larsen, S., Laurion, I., Leech, D. M., McAllister, S. L., McKnight, D. M., Melack, J., Overholt, E., Porter, J. A., Prairie, Y. T., Renwick, W. H., Roland, F., Sherman, B. S., Schindler, D. W., Sobek, S., Tremblay, A., Vanni, M. J., Verschoor, A. M., Wachenfeldt von, E., Weyhenmeyer, G., 2009: Lakes and reservoirs as regulators of carbon cycling and climate. *Limnology and Oceanography,* 54: 2298-2314.
Whalen, S. C., and Cornwell, J. C., 1985: Nitrogen, phosphorus, and organic carbon cycling in an Arctic Lake. *Canadian Journal of Fisheries and Aquatic Sciences,* 42: 797-808.
Woo, M. K., and Young, K. L., 2003: Hydrogeomorphology of patchy wetlands in the High Arctic, polar desert environment. *Wetlands,* 23(2): 291-309.
Woo, M. K. and Young, K. L., 2006: High arctic wetlands: Their occurrence, hydrological characteristics and sustainability. *Journal of Hydrology,* 320: 432-450.
Woo, M. K. 2012: *Permafrost Hydrology*. Springer, Heidelberg, 563.
Young, K. L., Assini, J., Abnizova, A., and De Miranda, N., 2010: Hydrology of hillslope-wetland streams, Polar Bear Pass, Nunavut, Canada. *Hydrological Processes,* 24: 3345-3358.
Young, K. L., and Labine, C., 2010: Summer hydroclimatology of an extensive low-gradient wetland: Polar Bear Pass, Bathurst Island, Nunavut. *Hydrology Research,* 41: 492-502.
Young, K.L., and Abnizova, A. 2011: Hydrologic thresholds of ponds in a polar desert wetland environment, Somerset Island, Nunavut, Canada. *Wetlands,* 31: 535- 549.
Yoshikawa, K., and Hinzman, L. D., 2003: Shrinking thermokarst ponds and

groundwater dynamics in discontinuous permafrost near Council, Alaska. *Permafrost and Periglacial Processes*, 14: 151–160.

Tables

TABLE 1. Average monthly air temperature and total monthly precipitation data from the main wetland site at PBP and from Environment Canada for Resolute Bay for the 2007 -2010 observation periods.

Year	Month	PBP Mean T_{air}(°C)	PBP Total PPT (mm)	Resolute Mean T_{air}(°C)	Resolute Total PPT (mm)
	June	1.4	5.4*	1.3	1.8
2007	July	8.7	3.2	7.4	10.2
	August	3.5	n/a	4.9	8.3
	June	2.6	19.5	2.2	21.6
2008	July	6.9	44.1	5.3	33.0
	August	2.0	32.4	2.3	26.0
	June	1.9	2.1	1.0	3.4
2009	July	5.0	61.7	5.3	64.3
	August	4.5	31.1	4.8	16.9
	June	2.6	trace	2.2	0.0
2010	July	7.7	35.8	5.3	33.9
	August	n/a	32.3*	4.3	57.0

Data do not include the entire month only June 25-30, 2007, August 1-4, 2010; n/a-not available.

TABLE 2. Mean seasonal values of physicochemical variables measured from 20 water bodies (18 ponds and 2 frost cracks) from Polar Bear Pass, Bathurst Island. A 'less than' symbol (<) indicates that sample was below the detection limit for that particular variable. Environmental variables at or below their detection limits across all sites are not shown. Frost table (FT) values report maximum seasonal thaw. Duration of each season is indicated in the Methods section. Sampling was conducted twice a week during each season.

	C1	C2	C3	C4	C5	C6	C7	C8	C9
Coordinates	75° 43' 30.50" 98° 25' 53.83"	75° 43' 30.82" 98° 25' 54.60"	75° 43' 26.67" 98° 25' 59.99"	75° 43' 28.71" 98° 25' 33.49"	75° 43' 27.24" 98° 25' 28.71"	75° 43' 27.24" 98° 25' 28.71"	75° 43' 23.72" 98° 25' 26.76"	75° 43' 27.24" 98° 25' 14.45"	75° 42' 42.74" 98° 26' 32.93"
Surface Area (m^2)	6375	275	7975	750	1850	1850	10950	1000	5200
2007									
WT (mm)	344	415	359	265	243	176	177	68	99
FT (mm)	635	410	645	485	310	710	650	595	851
pH	8.0	8.1	8.2	8.1	8.3	8.4	8.5	8.4	8.3
CON(μS m^{-1})	196	184	227	217	254	254	241	315	178
TEMP (°C)	8.3	9.8	9.1	10.0	10.6	10.2	10.2	12.1	10.2
Ca (mg L^{-1})	26.2	25.5	31.2	33.3	36.1	42.2	32.6	43.0	20.3
Mg (mg L^{-1})	10.0	9.9	11.5	10.6	10.2	11.0	10.8	11.9	5.3
Na (mg L^{-1})	13.8	13.6	9.9	15.4	14.8	15.0	12.5	15.6	12.1
K (mg L^{-1})	0.6	3.1	0.8	0.5	0.6	1.2	1.2	1.6	1.1
2008									
WT (mm)	302	382	347	215	247	107	162	80	-
FT (mm)	731	569	738	544	580	814	745	675	-
pH	8.1	7.9	7.9	8.0	8.1	8.1	8.3	8.2	-
CON (μS m^{-1})	229	226	244	256	264	260	261	313	-
TEMP (°C)	5.6	5.9	6.4	5.8	5.4	6.3	5.0	6.3	-
DO (mg L^{-1})	12.7	12.2	12.6	12.5	12.8	12.6	12.8	12.3	-
Ca (mg L^{-1})	41.0	37.2	40.9	48.3	43.3	52.8	39.2	46.6	-
Mg (mg L^{-1})	9.6	10.5	10.5	10.0	8.9	9.8	10.0	12.0	-
Na (mg L^{-1})	16.9	7.4	10.6	17.6	11.2	19.6	12.9	25.8	-
K (mg L^{-1})	0.8	0.3	0.5	0.5	0.5	0.4	1.1	1.5	-
Cl (mg L^{-1})	5.0	-	-	6.1	6.7	9.6	-	23.7	
DOC(mg L^{-1})	8.0	9.4	8.6	9.5	6.9	7.9	2.2	20.7	-
Alk (mg L^{-1})	106	-	-	103	-	110	-	101	-
2009									
WT (mm)	348	433	348	236	200	195	171	158	286
FT (mm)	689	489	644	534	524	707	680	631	408
pH	8.2	8.1	8.3	8.1	8.3	8.3	8.3	8.3	8.4
CON (μS m^{-1})	240	282	281	292	311	313	308	330	280
TEMP (°C)	6.7	6.8	7.0	7.0	7.5	6.7	6.8	8.2	8.3

DO (mg L^{-1})	12.6	11.5	12.4	12.4	11.7	12.8	12.6	12.6	12.0
Ca (mg L^{-1})	25.6	32.5	33.1	28.7	33.8	36.2	32.7	37.1	31.6
Mg (mg L^{-1})	8.1	10.0	8.9	9.0	8.4	8.9	9.9	10.3	6.7
Na (mg L^{-1})	11.4	9.6	11.1	9.9	12.4	14.5	16.1	19.0	8.8
K (mg L^{-1})	0.9	0.9	1.1	1.7	1.0	1.2	1.4	2.1	1.4
Cl (mg L^{-1})	3.9	6.4	4.9	5.0	8.4	9.1	10.8	18.3	7.4
DOC(mg L^{-1})	10.2	11.7	-	11.4	-	11.3	11.1	17.4	-
Alk (mg L^{-1})	112	-	-	133	-	144	-	144	-
d^{18}Owater	-225	-23.2	-20.1	-20.4	-19.4	-18.8	-17.7	-20.4	-16.3
dDwater	-117.4	-180.0	-160.3	-162.4	-156.4	-151.1	-144.1	-167.3	-135.0
2010									
WT (mm)	334	456	373	270	266	152	195	178	275
FT (mm)	649	502	632	492	512	702	678	529	334
pH	8.0	7.9	8.0	8.0	8.1	8.1	8.1	8.1	7.8
CON (µS m^{-1})	225	237	243	238	263	287	252	386	241
TEMP (°C)	8.2	9.6	8.1	9.2	9.2	8.7	6.8	9.7	13.2
DO (mg L^{-1})	11.3	10.1	11.3	11.2	11.2	11.3	11.9	11.4	9.4
Ca (mg L^{-1})	30.8	30.6	37.5	37.4	43.8	48.2	51.8	61.6	38.4
Mg (mg L^{-1})	9.6	10.0	10.8	9.4	9.9	10.5	11.4	14.0	7.0
Na (mg L^{-1})	3.1	3.1	3.1	3.3	4.1	4.5	5.5	8.5	2.9
K (mg L^{-1})	0.4	0.3	0.5	0.4	0.5	0.7	1.2	1.9	1.0
Ba (µg L^{-1})	145	96.6	160.7	235.7	273.0	291.7	296.3	345.5	304.0
Fe (µg L^{-1})	<50	48.1	<50	<50	<50	<50	<50	<50	87.1
Si (mg L^{-1})	0.6	0.4	0.7	0.6	0.7	0.9	1.6	2.5	0.5
Sr (µg L^{-1})	40.9	35.5	58.8	57.7	79.2	93.8	115.8	184.5	88.5
SO4 (mg L^{-1})	18.4	21.0	9.9	7.0	5.5	4.4	2.4	2.4	0.3
Cl (mg L^{-1})	7.6	7.2	7.6	7.4	10.2	11.5	9.9	20.4	6.7
DOC(mg L^{-1})	6.0	5.5	5.9	6.8	9.1	6.5	6.6	11.9	7.5
Alk (mg L^{-1})	110	108	145	139	160	177	155	230	143
d^{18}Owater	-22.6	-23.7	-21.0	-22.2	-21.1	-20.0	-18.5	-15.8	-16.7
dDwater	-177.9	-184.5	-167.3	-175.5	-168.8	-160.9	-149.6	-136.1	-138.0
2007-2010									
WT (mm)	191-397	205-636	232-469	145-364	144-311	47-246	86-271	-280-285	49-346
pH	6.0-9.0	6.0-8.9	6.0-9.1	6.3-9.0	5.9-9.1	6.9-9.0	6.9-9.0	6.9-9.0	7.5-8.7
CON (µS m^{-1})	57-354	103-403	4-347	94-366	105-388	85-363	100-355	155-740	77-313
TEMP (°C)	0.4-16.2	1.2-16.2	0.2-16.3	0.5-16.9	0.6-16.6	0.2-16.3	0.5-15.3	0.2-18.5	3.2-20.1
DO (mg L^{-1})	6.2-15.9	5.9-15.7	6.2-17.0	6.1-16.1	6.3-15.9	6.1-16.2	6.0-16.1	6.1-16.4	8.5-13.7
Ca (mg L^{-1})	1.9-62.3	1.7-60.6	1.5-52.7	3.2-82.3	9.9-62.7	1.9-76.6	0.2-68.3	12.6-71.9	0.8-48.0
Mg (mg L^{-1})	0.4-13.9	0.6-17.8	0.5-15.4	0.9-18.7	1.4-13.9	0.6-13.6	1.8-13.8	1.7-18.3	2.3-8.3
Na (mg L^{-1})	2.1-23.8	2.7-23.8	1.9-26.4	2.8-20.7	3.5-26.9	3.8-24.9	4.3-25.7	6.8-32.2	2.5-18.1

K (mg L^{-1})	0.1-5.9	0.1-4.6	0.2-3.4	0.0-4.9	0.1-3.7	0.1-4.0	0.3-2.6	0.3-4.8	0.6-2.1
Cl (mg L^{-1})	0.0-9.2	2.8-8.9	4.5-9.1	0.0-11.3	5.3-13.1	3.8-15.9	8.1-12.7	0.0-44.8	5.4-9.4
DOC(mg L^{-1})	2.1-18.5	2.4-16.7	0.8-16.5	2.3-23.3	1.4-15.8	1.0-16.3	1.2-15.7	5.7-41.6	6.7-8.4
Alk (mg L^{-1})	68-154	96-123	169-117	48-184	136-176	83-199	120-186	73-249	130-156

C10	C11	C12	C13	C14	C15	C16	C17	CF18	CF19-20	C21
		75°	75°	75°		75°				
75° 42'	75° 43'	43'	43'	43'	75° 43'	43'	75° 43'	*75° 43'*	*75° 42'*	*75° 42'*
40.50"	33.86"	26.91"	26.91"	23.08"	22.08"	34.92"	36.14"	*36.92"*	*43.02"*	*43.34"*
98° 26'	98° 24'	98°	98°	98°	98° 22'	98°	98° 31'	*98° 31'*	*98° 26'*	*98° 26'*
27.14"	8.66"	22'	22'	26'	50.65"	31'	29.93"	*16.43"*	*27.43"*	*26.90"*
		59.49"	59.49"	33.74"		2.89"				
6475	550	10475	1725	650	1250	10666	726	80	100	50
123	132*	-	-	-	-	-	-	-	-	-
432	555*	-	-	-	-	-	-	-	-	-
7.9	8.2*	-	-	-	-	-	-	-	-	-
236	308*	-	-	-	-	-	-	-	-	-
10.2	12.0*	-	-	-	-	-	-	-	-	-
34.2	43.1	-	-	-	-	-	-	-	-	-
7.7	11.5	-	-	-	-	-	-	-	-	-
15.0	14.6	-	-	-	-	-	-	-	-	-
1.2	1.7	-	-	-	-	-	-	-	-	-
58	141	168	236	202	196	273	149	211	-	-
529	863	951	819	549	698	388	852	675	-	-
8.7	8.3	8.6	8.8	8.1	8.0	8.5	8.4	8.3	-	-
247	380	230	272	277	291	213	249	415	-	-
6.6	6.8	6.4	6.3	5.5	5.9	8.0	8.2	7.6	-	-
13.3	12.6	13.2	13.4	12.4	12.5	12.7	12.7	12.8	-	-
47.4	54.2	41.8	44.3	59.3	55.0	33.7	55.1	70.5	-	-
8.3	8.2	6.3	7.8	10.3	13.1	6.8	9.4	14.9	-	-
18.6	30.9	10.9	12.4	19.4	18.9	18.2	20.5	16.3	-	-
0.6	1.7	2.1	1.9	0.5	0.4	1.2	1.7	3.4	-	-
9.1	39.8	-	-	8.6	12.1	11.3	11.7	-		
9.2	18.8	-	-	8.9	9.6	7.1	18.8	25.1	-	-
107	119	-	-	115	112	71	108	-	-	-
103	243	256	303	261	215	525	380	268	91	156
659	825	903	882	544	612	307	252	322	354	491
8.2	8.4	8.5	8.4	8.2	8.2	8.3	8.1	8.3	8.5	8.3
276	440	253	301	355	338	195	276	356	393	279
8.8	7.7	8.5	8.2	6.6	6.4	7.2	7.2	8.7	10.2	8.7
12.3	11.6	12.2	11.8	12.3	12.4	12.6	11.7	13.7	12.0	11.9
31.5	46.4	26.5	30.7	48.6	40.7	21.7	27.0	45.0	49.7	37.1
7.0	10.0	4.1	7.9	10.8	10.9	5.5	8.3	9.1	13.9	7.4
11.5	28.7	11.6	14.4	14.4	9.0	13.4	13.3	11.6	13.2	10.8
1.3	2.2	2.1	2.3	1.1	1.0	1.8	2.9	2.0	0.7	2.2
8.6	27.5	5.2	11.6	7.0	13.9	9.6	9.8	12.1	-	9.4
11.7	13.9	-	-	11.2	11.7	9.9	17.4	20.5	21.0	-
119	154	-	-	150	153	79.5	110	-	-	-
-15.0	-19.2	-23.3	-15.9	-21.1	-19.4	-17.7	-15.4	-15.6	-13.8	-15.8
-126.1	-155.7	-186.4	-133.3	-166.6	-156.3	-146.7	-131.8	-137.9	-116.1	-132.1
114	247	193	254	328	172	525	435	243	77	124
428	610	661	534	502	609	294	211	303	364	413
7.8	8.0	8.4	8.3	8.1	8.0	7.8	7.8	7.5	7.7	7.7
234	398	233	250	271	303	117	181	266	451	301
12.7	11.7	17.5	17.4	8.3	8.8	6.3	8.3	8.7	13.9	14.5

9.5	11.0	6.8	6.7	11.8	10.6	11.2	10.3	10.4	10.2	8.6
33.3	64.0	-	-	48.0	49.5	17.1	31.9	37.1	63.5	49.0
6.6	13.0	-	-	9.5	13.2	4.0	7.7	7.5	16.0	9.4
3.0	12.7	-	-	4.5	4.5	2.9	3.6	2.9	4.5	3.5
0.9	1.4	-	-	0.4	0.8	0.8	1.5	2.0	2.0	1.5
272.0	211.5	-	-	197.0	199.3	65.4	185.0	115.2	377.0	375.0
<50	62.0	-	-	38.9	42.7	<50	40.4	284.0	<50	290.5
0.4	0.9	-	-	1.1	1.1	0.5	1.1	1.3	0.8	0.7
76.3	134.0	-	-	83.9	72.4	37.0	70.7	79.2	138.0	103.5
0.1	14.6	-	-	8.3	10.6	4.6	2.4	7.3	<0.1	0.4
6.4	32.0	-	-	10.4	11.2	6.9	6.2	8.4	9.4	9.4
6.6	9.4	-	-	6.9	7.9	5.2	12.5	10.1	17.9	17.9
127	209	-	-	169	187	59	128	130	260	260
-15.8	-18.2	-	-	-22.0	-20.8	-19.3	-16.6	-19.4	-14.8	-16.8
-132.1	-149.3	-	-	-172.7	-165.4	-156.0	-138.8	-155.9	-129.4	-141.2
26-182	47-340	127-326	197-331	114-365	136-354	232-568	101-471	86-346	23-179	65-249
7.1-9.0	7.3-8.9	7.7-9.1	7.6-9.0	6.4-8.9	5.7-9.1	6.6-8.9	6.7-8.9	7.2-8.7	7.2-8.4	7.3-8.3
85-322	151-549	94-284	101-339	55-473	149-416	34-359	70-392	187-449	283-540	159-340
1.2-19.1	0.9-19.5	0.8-19.7	0.8-19.2	0.5-17.8	0.2-16.6	1.4-17.6	0.3-16.3	1.6-14.3	2.1-21.9	3.2-21.0
8.3-15.6	7.7-16.3	5.1-15.2	4.9-15.7	5.9-16.0	5.5-16.7	9.4-15.5	9.4-16.2	9.0-15.6	2.9-13.1	7.2-13.8
0.1-71.7	3.7-92.7	8.5-51.7	1.2-61.2	2.5-116.6	2.4-74.6	1.4-51.7	0.0-79.0	20.3-95.7	40.3-65.0	26.9-53.4
0.1-12.7	0.0-17.5	1.5-9.1	0.4-13.8	0.7-20.5	0.7-16.9	0.2-11.5	0.0-12.5	2.2-20.0	9.6-22.5	5.0-10.6
2.4-22.3	10.3-55.2	6.0-21.4	8.3-22.2	3.7-26.6	3.0-21.0	2.7-26.8	3.2-31.7	2.6-26.5	3.3-26.8	3.2-22.7
0.0-2.2	0.1-3.8	0.2-4.2	0.2-4.1	0.1-3.5	0.1-3.9	0.1-3.4	1.3-3.6	1.0-5.6	0.2-2.1	1.4-3.7
0.0-11.7	0.0-52.8	0.3-9.8	9.4-13.6	2.1-13.4	7.6-23.0	0.0-19.3	1.7-15.1	8.0-13.9	6.2-12.6	5.3-13.5
1.7-16.7	1.5-33.9	-	-	2.4-17.6	6.9-18.7	0.9-13.2	8.5-24.5	5.5-29.5	12.4-25.1	5.3-22.5
40-160	62-223	-	-	55-208	80-212	30-120	46-144	108-152	225-293	167-207

TABLE 3. Limnological variables measured from 36 satellite ponds at Polar Bear Pass, Bathurst Island. A 'less than' symbol (<) indicates that sample was below the detection limit for that particular variable. Environmental variables at or below their detection limits across all sites are not shown.

	S22	S23	S24	S25	S26	S27	S28	S29	S30	S31	S32	S33
Coordinates	75° 44' 6.01" 98° 5' 42.73"	75° 44' 4.80" 98° 5' 28.15"	75° 44' 10.01" 98° 5' 31.62"	75° 43' 42.89" 98° 10' 51.64"	75° 43' 45.00" 98° .10' 35.03"	75° 43' 44.56" 98° 10' 49.19"	75° 43' 46.17" 98° 10' 36.93"	75° 43' 19.45" 98° 10' 29.41"	75° 43' 43.97" 98° 10' 27.76"	75° 41' 51.00" 98° 20' 53.64"	75° 41' 48.73" 98° 20' 36.25"	75° 41' 50.93" 98° 20' 49.33"
Surface Area	100	1500	24500	400	200	4469	706	2025	4875	325	450	350
Summer 2007 (Jul 26 – Aug 2)										-	-	-
WT (mm)	70	110	125	-	-	-	-	-	-	-	-	-
FT (mm)	460	740	760	-	-	-	-	-	-	-	-	-
pH	6.6	6.9	7.1	-	-	-	-	-	-	-	-	-
CON (µS m^{-1})	439	331	228	-	-	-	-	-	-	-	-	-
TEMP (°C)	10.9	10.4	11.4	-	-	-	-	-	-	-	-	-
Ca (mg L^{-1})	58.8	48.3	39.6	-	-	-	-	-	-	-	-	-
Mg (mg L^{-1})	11.7	9.1	6.2	-	-	-	-	-	-	-	-	-
Na (mg L^{-1})	16.2	12.8	12.4	-	-	-	-	-	-	-	-	-
K (mg L^{-1})	3.6	2.1	2.1	-	-	-	-	-	-			
2008 Fall (Aug 26 – Sep 1)												
WT (mm)	135	222	275	370	183	532	314	280	541	75	66	180
FT (mm)	556	648	588	345	387	521	351	499	283	677	427	675
pH	8.75	8.85	8.79	8.94	9.01	9	9.04	9.01	9.02	8.92	8.94	9
CON (µS m^{-1})	726	561	413	368	331	299	307	352	287	241	276	227
TEMP (°C)	3.1	3.0	3.3	3.1	3.2	3.1	3.2	3.0	3.1	3.6	4.0	3.3
Ca (mg L^{-1})	128.1	82.4	93.1	25.5	42.5	54.7	42.5	52.1	38.4	49.9	40.6	20.1
Mg (mg L^{-1})	14.0	8.9	6.6	9.5	10.0	10.4	8.8	9.2	7.9	7.3	9.2	5.7
Na (mg L^{-1})	47.1	35.3	27.7	34.9	26.6	21.6	24.4	19.2	23.7	13.1	21.1	26.9
K (mg L^{-1})	1.6	1.1	0.9	1.7	1.5	2.5	1.6	2.3	1.4	0.8	1.0	1.3
2009 Summer (Aug 3 – 5)												
WT	270	290	245	-	-	-	-	-	-	250	95	410

(mm)												
FT (mm)	921	603	625	-	-	-	-	-	-	386	413	386
CON (μS m^{-1})	559	430	326	-	-	-	-	-	-	217	285	217
TEMP (°C)	11.1	10.8	11.3	-	-	-	-	-	-	15.9	17.8	14.9
Ca (mg L^{-1})	55.5	30.2	38.6	-	-	-	-	-	-	22.2	20.3	22.8
Mg (mg L^{-1})	6.0	3.8	4.7	-	-	-	-	-	-	4.6	10.6	4.3
Na (mg L^{-1})	-	-	-	-	-	-	-	-	-	10.1	9.8	5.9
K (mg L^{-1})	1.3	0.9	1.0	-	-	-	-	-	-	1.5	1.4	1.8
2009 Fall (Aug 26 – 30)										-	-	-
WT (mm)	260	294	223	407	297	494	392	401	505	-	-	-
FT (mm)	968	948	1020	407	506	322	330	602	359	-	-	-
pH	8.3	8.5	8.6	8.4	8.6	8.6	8.1	8.5	8.6	-	-	-
CON (μS m^{-1})	752	518	329	384	330	326	322	352	309	-	-	-
TEMP (°C)	4.7	4.2	3.9	4.0	3.9	4.3	4.0	4.2	4.2	-	-	-
DO (mg L^{-1})	11.5	12.1	12.5	10.7	11.7	11.8	10.9	12.2	12.0	-	-	-
Ca (mg L^{-1})	58.8	22.8	30.4	36.4	36.2	41.5	28.1	30.3	17.2	-	-	-
Mg (mg L^{-1})	5.4	5.7	5.7	8.8	7.1	8.4	8.1	7.6	3.6			
K (mg L^{-1})	2.0	1.8	1.3	1.3	1.7	2.9	2.9	2.6	1.1			
Cl (mg L^{-1})	15.2	77.8	66.7	25.4	17.6	20.9	18.0	13.5	24.2			
2010 Spring (Jun 19 – 24)												
WT (mm)	198	165	104	400	195	500	381	265	320	326	146	425
FT (mm)	15	23	20	10	21	0	20	40	25	24	22	0
pH	7.7	8.0	7.3	7.5	7.7	7.5	7.7	8.1	8.1	7.4	7.7	7.4
CON (μS m^{-1})	283	225	53	105	196	186	107	202	133	68	109	59
TEMP (°C)	5.1	9.6	6.2	3.7	7.4	3.1	5.0	7.2	2.5	5.1	10.2	3.5
DO (mg L^{-1})	11.6	10.9	14.6	12.8	11.3	11.7	12.9	11.3	12.6	12.6	11.7	12.5
Ca (mg L^{-1})	37.3	35.3	26.0	24.2	30.8	29.1	33.7	22.6	22.6	8.5	14.5	6.2
Mg (mg L^{-1})	5.5	4.8	3.3	3.6	5.1	3.7	4.1	3.3	3.6	1.7	5.0	1.4
Na (mg L^{-1})	3.4	11.9	4.3	3.4	2.8	3.2	9.6	1.8	2.4	1.7	1.4	1.6
K (mg L^{-1})	1.6	2.0	1.2	0.9	1.2	1.2	2.4	0.8	1.0	0.7	0.8	0.5

Coordinates	S34	S35	S36	S37	S38	S39	S40	S41	S42
Summer 2007 (Jul 26 – Aug 2)	75° 41' 49.00" 98° 20' 29.56"	75° 41' 51.36" 98° 20' 40.83"	75° 41' 47.19" 98° 20' 28.74"	75° 42' 20.53" 98° 45' 18.70"	75° 42' 25.24" 98° 45' 17.19"	75° 42' 21.47" 98° 45' 29.00"	75° 42' 39.33" 98° 39' 49.63"	75° 42' 36.65" 98° 39' 49.82"	75° 42' 38.67" 98° 40' 4.02"
Surface Area	300	2075	3575	2550	7050	31400	200	575	1225
WT (mm)	-	-	-	60	100	130			
FT (mm)	-	-	-	885	1088	1020	-	-	-
pH	-	-	-	7.2	7.5	7.5	-	-	-
Cond (µS m^{-1})	-	-	-	209	266	186	-	-	-
Temp (°C)	-	-	-	11.4	12.0	11.6	-	-	-
Ca (mg L^{-1})	-	-	-	34.4	37.4	27.7	-	-	-
Mg (mg L^{-1})	-	-	-	7.6	11.1	6.8	-	-	-
Na (mg L^{-1})	-	-	-	8.5	13.2	19.2	-	-	-
K (mg L^{-1})	-	-	-	3.8	3.4	3.2	-	-	-
2008 Fall (Aug 26 – Sep 1)									
WT (mm)	80	400	328	281	150	262	130	-	256
FT (mm)	522	653	443	527	852	893	544	-	479
pH	8.9	8.9	9.0	8.9	8.9	8.9	8.9	-	8.8
Cond (µS m^{-1})	326	283	283	279	345	275	304	-	265
Temp (C)	3.8	3.1	3.1	4.7	4.5	4.4	5.5	-	265
Ca (mg L^{-1})	61.3	29.2	38.9	58.4	72.1	78.2	64.0	-	5.2
Mg (mg L^{-1})	11.2	9.7	9.5	7.2	10.3	14.3	7.8	-	5.5
Na (mg L^{-1})	13.7	18.4	20.1	14.2	25.9	30.0	14.2	-	17.8
K (mg L^{-1})	1.1	1.2	0.8	0.2	0.8	1.4	1.4	-	0.7
2009 Summer (Aug 3 – 5)									
WT (mm)	175	390	381	340	268	350			
FT (mm)	513	323	492	677	713	876	-	-	-
Cond (µS m^{-1})	299	294	310	289	373	291	-	-	-
Temp (°C)	16.3	14.4	14.6	12.0	12.0	11.6	-	-	-
Ca (mg L^{-1})	28.9	27.5	28.9	18.0	35.9	25.0	-	-	-
Mg (mg L^{-1})	8.4	7.8	7.4	3.0	7.0	5.6	-	-	-
Na (mg L^{-1})	7.4	6.4	7.2	6.9	16.9	13.7			
K (mg L^{-1})	1.7	1.9	1.2	1.4	2.2	2.1	-	-	-
2009 Fall (Aug 26 – 30)							-	-	-
WT (mm)	-	-	-	406	222	446			
FT (mm)	-	-	-	618	959	798	-	-	-
pH	-	-	-	8.2	8.4	8.5	-	-	-
Cond (µS m^{-1})	-	-	-	184	363	271	-	-	-
Temp (°C)	-	-	-	4.2	4.2	4.5	-	-	-
DO (mg L^{-1})	-	-	-	11.5	11.4	11.4	-	-	-
Ca (mg L^{-1})	-	-	-	31.6	33.8	25.5	-	-	-
Mg (mg L^{-1})	-	-	-	5.5	8.8	5.0	-	-	-
K (mg L^{-1})	-	-	-	1.1	1.6	1.8	-	-	-
Cl (mg L^{-1})	-	-	-	11.0	30.0	16.4	-	-	-
2010 Spring (Jun 19 –							-	-	-

24)									
WT (mm)	242	455	284	228	188	195	80	52	252
FT (mm)	78	6	30	75	68	30	30	25	0
pH	7.7	7.4	7.3	7.7	7.7	7.6	6.9	7.4	7.2
Cond (μS/m)	90	81	76	113	111	107	53	63	100
Temp (°C)	6.9	4.8	5.7	3.3	1.7	1.9	4.6	5.3	4.6
DO (mg L^{-1})	12.5	13.2	13.4	14.1	14.2	15.3	12.8	13.1	12.2
Ca (mg L^{-1})	12.8	10.3	6.9	18.1	17.9	15.5	6.6	8.1	9.4
Mg (mg L^{-1})	3.9	2.9	2.1	2.3	2.6	2.3	1.1	1.2	1.4
Na (mg L^{-1})	1.8	1.0	1.8	2.5	2.5	2.5	1.6	2.0	2.4
K (mg L^{-1})	0.7	0.6	0.5	0.8	0.8	0.7	0.5	0.6	0.7

S43	S44	S45	S46	S47	S48	S48	S49	S50	Mean	Max	Min	St. Dev.
75° 42' 38.00" 98° 39' 46.00"	75° 42' 41.65" 98° 39' 51.66"	75° 43' 5.46" 98° 39' 51.77"	75° 41' 37.83" 98° 34' 26.27"	75° 41' 37.73" 98° 34' 21.62"	75° 41' 37.49" 98° 34' 15.23"	75° 41' 37.32" 98° 34' 10.15"	75° 41' 34.67" 98° 34' 9.44"	75° 41' 35.10" 98° 33' 37.72"				
750	6600	3900	675	625	500	1150	7200	3425				
-	-	-	-	-	-	-	-	-	99	130	60	29
-	-	-	-	-	-	-	-	-	826	1088	460	226
-	-	-	-	-	-	-	-	-	7.1	7.5	6.6	0.4
-	-	-	-	-	-	-	-	-	276	439	186	94
-	-	-	-	-	-	-	-	-	11.3	12.0	10.4	0.5
-	-	-	-	-	-	-	-	-	41.0	58.8	27.7	11.0
-	-	-	-	-	-	-	-	-	8.8	11.7	6.2	2.3
-	-	-	-	-	-	-	-	-	13.7	19.2	8.5	3.7
-	-	-	-	-	-	-	-	-	3.0	3.8	2.1	0.8
181	405	365	310	230	263	334	500	441	279	541	66	131
474	677	477	380	389	293	489	430	362	512	893	283	151
8.8	8.9	8.9	8.8	8.9	8.8	8.9	9.0	8.9	8.9	9.0	8.8	0.1
294	275	283	325	310	316	177	348	302	323	726	177	102
5.2	4.4	4.5	5.3	5.2	5.1	5.2	4.8	4.5	4.0	5.5	3.0	0.9
61.8	24.1	55.1	58.7	61.2	60.1	23.6	54.0	22.8	53.5	128.1	20.1	23.5
7.6	7.1	5.6	11.0	10.9	11.2	10.5	11.9	6.2	9.1	14.3	5.5	2.3
15.9	31.7	23.2	16.9	15.9	16.2	31.6	20.1	28.5	23.3	47.1	13.1	8.0
1.5	1.6	1.3	0.8	0.7	0.6	1.3	1.3	1.2	1.2	2.5	0.2	0.5
-	-	-	-	-	-	-	-	-	289	410	95	93
-	-	-	-	-	-	-	-	-	577	921	323	194
-	-	-	-	-	-	-	-	-	324	559	217	94
-	-	-	-	-	-	-	-	-	13.6	17.8	10.8	2.4
-	-	-	-	-	-	-	-	-	29.5	55.5	18.0	10.2
-	-	-	-	-	-	-	-	-	6.1	10.6	3.0	2.2
-	-	-	-	-	-	-	-	-	11.2	16.9	5.9	4.6
-	-	-	-	-	-	-	-	-	1.5	2.2	0.9	0.4
-	-	-	383	305	346	417	474	448	383	505	222	89
-	-	-	382	416	418	448	580	507	549	1020	322	243
-	-	-	8.0	8.4	8.4	8.5	8.6	8.4	8.4	8.6	8.0	0.2
-	-	-	322	330	353	364	370	295	316	752	184	117
-	-	-	4.6	4.5	4.5	4.6	4.8	4.9	4.5	4.9	3.9	0.3
-	-	-	11.8	11.5	11.5	11.5	11.7	11.7	11.6	12.5	10.7	0.4
-	-	-	30.4	39.3	39.3	32.0	23.3	18.9	29.1	58.8	17.2	9.6
-	-	-	9.1	8.9	8.9	10.0	8.9	2.2	7.1	10.0	2.2	2.2
-	-	-	1.2	1.6	1.6	0.9	0.7	1.2	1.3	2.9	0.7	0.6

-	-	-	13.5	13.9	16.9	18.6	21.7	28.3	19.5	77.8	11.0	18.1
75	254	100	-	-	-	-	-	-	243	500	52	125
20	0	0	-	-	-	-	-	-	24	78	0	22
7.4	7.1	7.4	-	-	-	-	-	-	7.5	8.1	6.9	0.3
64	93	47	-	-	-	-	-	-	114	283	47	62
5.4	7.9	4.5	-	-	-	-	-	-	5.2	10.2	1.7	2.2
13.3	13.4	13.3	-	-	-	-	-	-	12.8	15.3	10.9	1.1
8.8	10.6	13.1	-	-	-	-	-	-	17.9	37.3	6.2	9.9
1.5	1.6	2.3	-	-	-	-	-	-	2.9	5.5	1.1	1.3
1.8	2.9	2.2	-	-	-	-	-	-	3.0	11.9	1.0	2.5
0.6	0.6	0.8	-	-	-	-	-	-	0.9	2.4	0.5	0.5

TABLE 4. Spearman correlation matrix identifying groups of significantly (0.01<P<0.05) correlated environmental variables based on hydro-chemical mapping of ponds across PBP on June 20, 2010.**

	WT	FT	CON	pH	DO	TEMP	Ca	Mg	Na
WT	1.000								
FT	-0.359*	1.000							
CON	0.158	0.056	1.000						
pH	0.109	0.150	0.809**	1.000					
DO	-0.187	-0.074	-0.500**	-0.511**	1.000				
TEMP	-0.100	0.117	0.533**	0.389**	-0.441**	1.000			
Ca	0.067	-0.050	0.849**	0.761**	-0.473**	0.589**	1.000		
Mg	0.225	-0.092	0.846**	0.770**	-0.462**	0.623**	0.910**	1.000	
Na	0.069	-0.104	0.642**	0.484**	-0.149	0.421**	0.826**	0.705**	1.000
K	0.045	0.025	0.696**	0.688**	-0.530**	0.401**	0.830**	0.690**	0.755**
Ba	0.064	0.025	0.790**	0.771**	-0.519**	0.638**	0.917**	0.869**	0.651**
Si	0.246	-0.240	0.774**	0.679**	-0.378*	0.353**	0.799**	0.801**	0.625**
Sr	0.039	0.065	0.775**	0.704**	-0.496**	0.412**	0.900**	0.754**	0.679**
HCO_3	0.090	-0.041	0.873**	0.839**	-0.514**	0.609**	0.923**	0.895**	0.696**
Cl	0.133	-0.218	0.707**	0.529**	-0.266	0.453**	0.813**	0.738**	0.877**
SO_4	0.396	-0.376	0.377*	0.201	0.054	-0.038	0.321**	0.393**	0.502**
DOC	-0.278	0.098	0.366*	0.241	-0.297	0.355*	0.400**	0.311*	0.294
DIC	0.078	-0.020	0.632**	0.598**	-0.338*	0.432**	0.732**	0.760**	0.516**

K	Ba	Si	Sr	HCO_3	Cl	SO_4	DOC	DIC
1.000								
0.757**	1.000							
0.785**	0.750**	1.000						
0.817**	0.889**	0.751**	1.000					
0.801*	0.896**	0.816**	0.827**	1.000				
0.735**	0.626**	0.695**	0.644**	0.749**	1.000			
0.288	0.077	0.468**	0.106	0.300	0.631**	1.000		

0.405**	0.269	0.389**	0.380*	0.385*	0.311*	0.061	1.000	
0.545**	0.729**	0.645**	0.650**	0.740**	0.487**	0.156	0.438**	1.000

TABLE 5. Spearman correlation matrix identifying groups of significantly (0.01 < P < 0.05) correlated environmental variables based on hydro-chemical mapping of ponds across PBP from August 26th – August 30th in 2009.**

	WT	SA	FT	CON	pH	DO	TEMP	Ca	Mg	K	Cl
WT	1										
SA	0.220	1									
FT	-0.294	0.415**	1								
CON	-0.395*	-0.572**	-0.004	1							
pH	-0.284	0.214	0.086	-0.058	1						
DO	-0.176	0.039	0.013	0.112	0.409**	1					
TEMP	-0.369*	-0.159	-0.182	0.026	0.419**	0.087	1				
Ca	-0.546**	-0.490**	-0.159	0.522**	0.310	0.084	0.593**	1			
Mg	-0.352*	-0.419**	-0.351**	0.583**	0.215	0.134	0.394**	0.684**	1		
K	0.084	-0.005	0.084	-0.170	-0.003	-0.125	0.041	-0.017	-0.354*	1	
Cl	0.266	0.056	0.167	0.158	-0.206	-0.086	-0.457**	-0.410**	-0.354*	0.292	1

Figures

Figure 1. Topographic map of the study area showing the wetland and wetland basin boundary of Polar Bear Pass, Bathurst Island, Nunavut (75° 40'N, 98° 30'W). *Control Ponds* (C) are shown with white circles, *Other Control Ponds* (C) and *Satellite Ponds* (S) are shown with black circles. Contour lines are spaced at 30-m intervals in the drainage basin area and at 10-m interval in the wetland area.

Figure 2. Examples of surface water bodies photographed across Polar Bear Pass wetland. (a) Centrally located *Control Ponds*, and upslope late-lying snowbed and hillslope creek; (b) south *Satellite Ponds*; (c) west *Satellite Ponds*; (d) east *Satellite Ponds*; (e) hydrologically connected *Control Pond* C1 ; (d) hydrologically isolated *Control Pond* C8; (e) polygonized terrain; (f) frost crack CF18.

Figure 3. Mean daily air temperature (Tair) and daily precipitation (P) at Polar Bear Pass and specific electric conductivity in selected ponds during study seasons 2007-2010.

Figure 4. Selected seasonal water tables in hydrologically connected *Control Ponds* C1 and C14 (a-b), isolated *Control Ponds* C8 and C10 (c-d), frost crack CF18 (e) and *Satellite Ponds* (S23, S33, S38, f) in 2007-2010.

Figure 5. Isotopic composition ($\delta^{18}O$ and δD) of the *Control Ponds* during the study seasons in 2007, 2009 and 2010 (a). Each distinct grouping between the linked and

isolated ponds is selected with an oval shape: solid black oval containing the linked ponds from snowmelt to early post-snowmelt period, solid grey oval containing the isolated ponds during the post-snowmelt period, dashed grey oval indicating the areas of overlap between the linked ponds sampled in late post-snowmelt period and the isolated ponds during the snowmelt. Solid, thick black or grey line: regression for the linked *Control Ponds*; dashed black or grey line: regression for the isolated *Control Ponds* (b).

Figure 6. Isotopic composition ($\delta^{18}O$ and δD) of the *Control Ponds* in 2007 (a); the *Control Ponds* and isotopic values of snow, rain and creek water in 2009 (b) and in 2010 (c); the *Other Control Ponds* in 2009 and 2010 (d); the *Control Ponds* and the *Satellite Ponds* in 2009 (e) and 2010 (f). All isotopic compositions of ponds are superimposed on the isotopic framework from Fig. 5.

Figure 7. Piper diagrams summarizing the relative concentrations of the major cations and anions in surface waters across PBP in 2010.

Figure 8. PCA biplot illustrating the relationships between 21 environmental variables and 40 sites from Polar Bear Pass during the snowmelt period in 2010 (Jun 19th – Jun 24th). Dashed and solid arrows represent passive and active environmental variable, respectively. Linked ponds are shown with crosses, isolated ponds are shown with squares, and satellite ponds are shown with circles. Physico-chemical variables used in the analysis include calcium (Ca), magnesium (Mg), potassium (K), sodium (Na),

Chloride (Cl), barium (Ba), iron (Fe), manganese (Mn), silicon (Si), strontium (Sr), fluoride (F), sulphate (SO4), hydrogen carbonate (HCO3), dissolved organic carbon (DOC), dissolved inorganic carbon (DIC), water table (WT), pH, conductivity (CON), dissolved oxygen (DO), water temperature (Temp) and frost table (FT).

Figure 9. Schematic diagram of Polar Bear Pass wetland illustrating the mechanism responsible for topographic hydrological connectivity between the hillslope sources and the wetland ponds. The study findings suggest that the terrain connectivity is the strongest within 500 m from the hillslope source which is identified as grey area in the diagram. Ponds located in this zone are identified to have strong hydrologic connectivity to their catchments, where as ponds located outside of this zone are the isolated water bodies receiving water inputs from precipitation only.

Figures

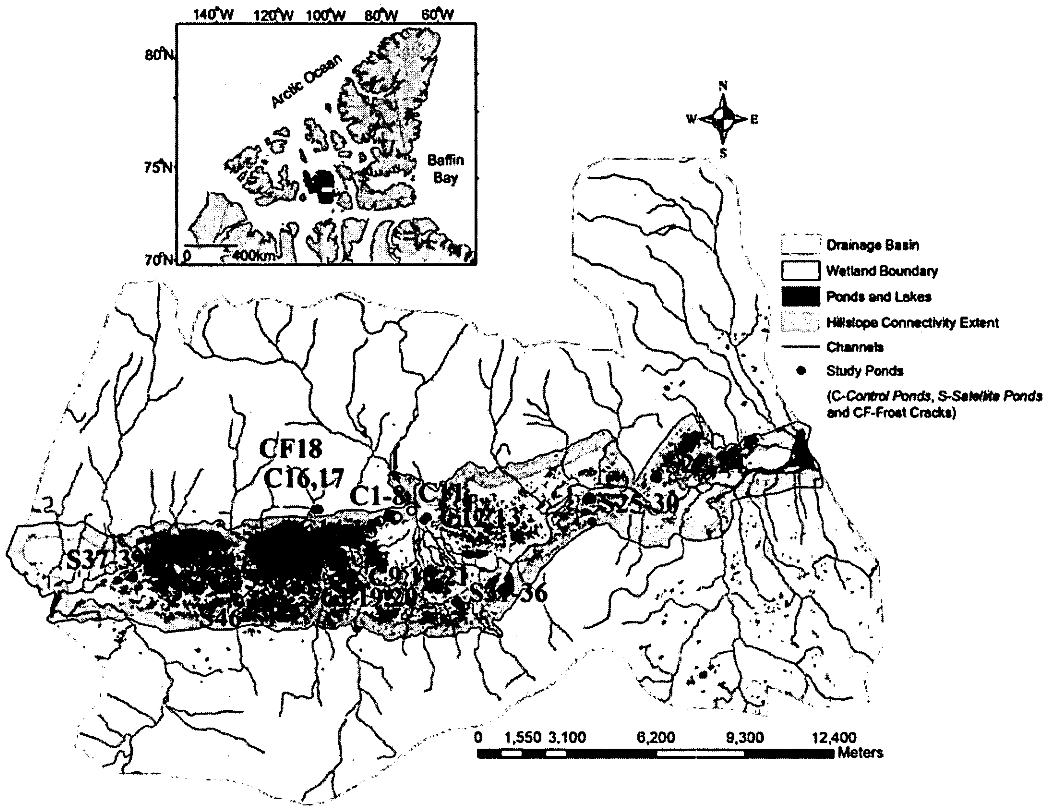


Figure 1

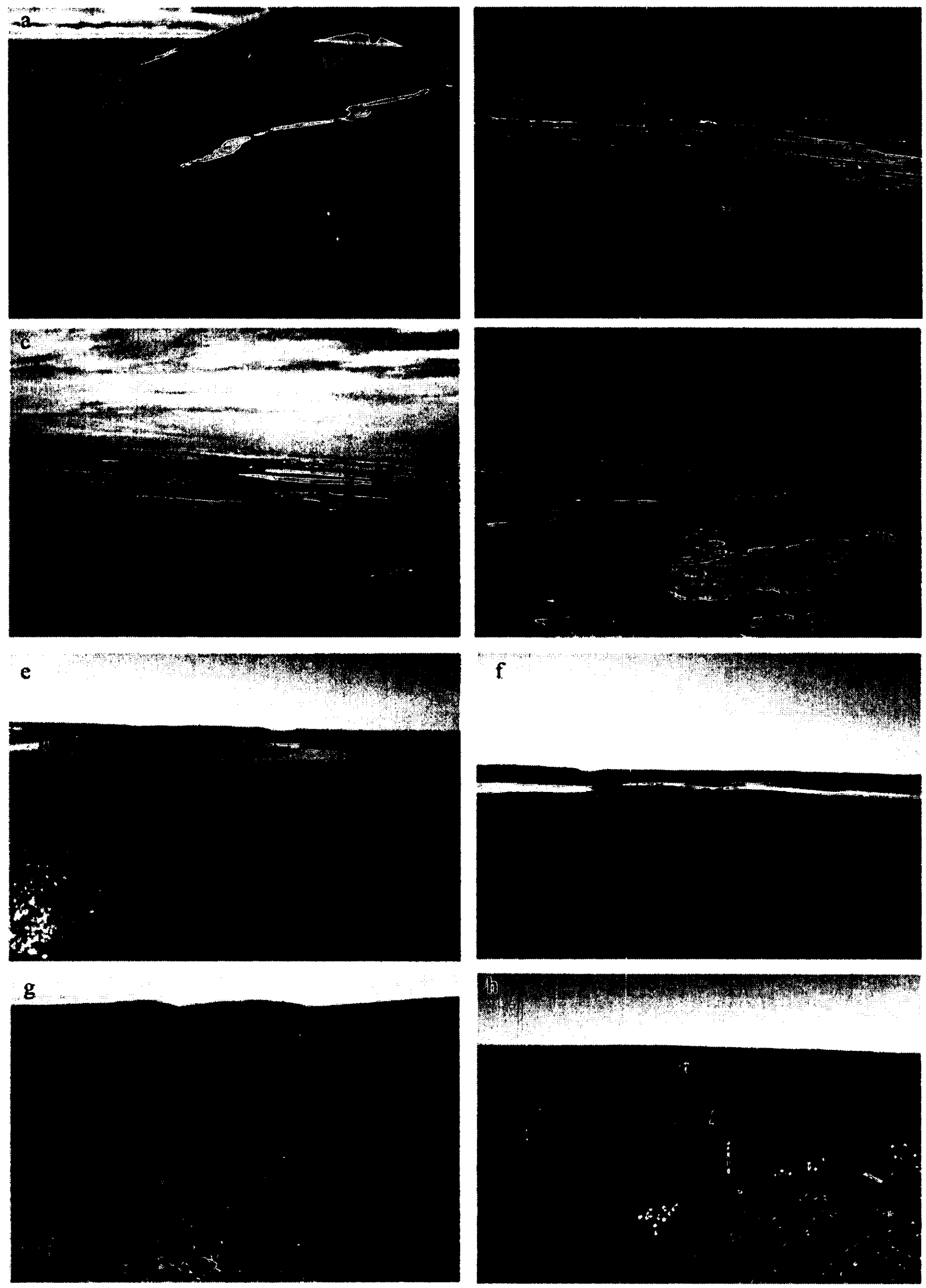


Figure 2

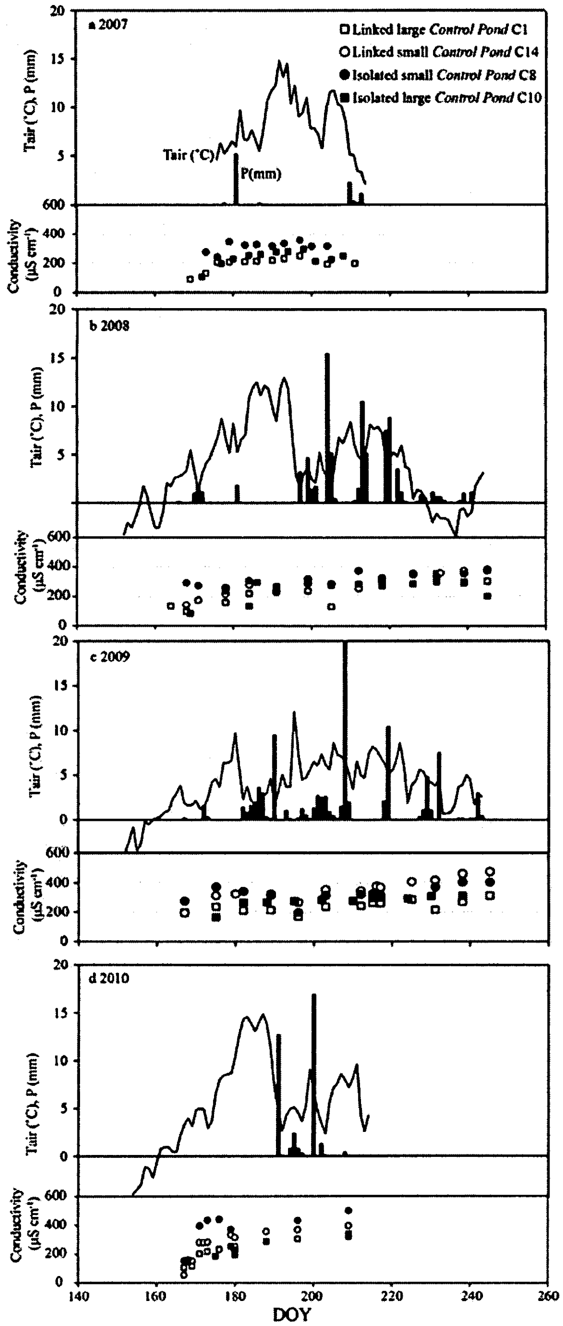
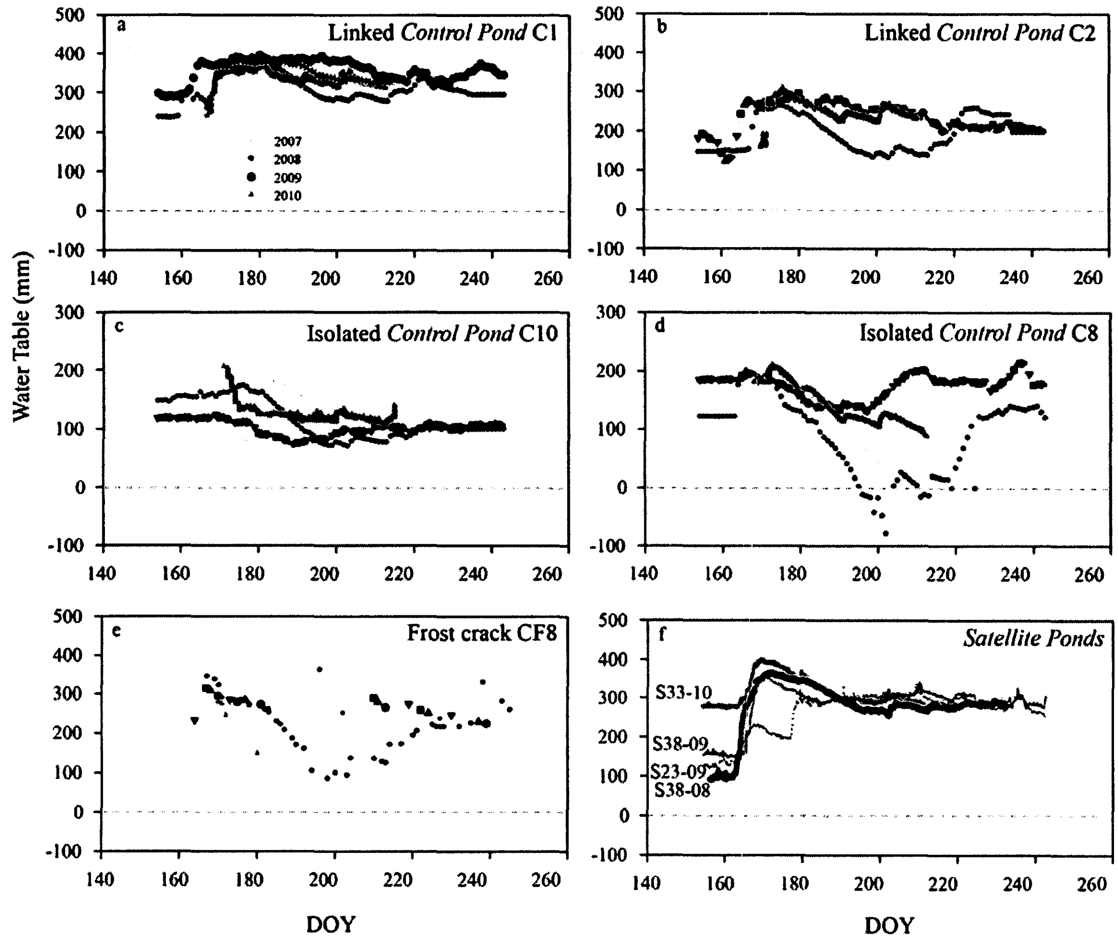

a 2007
Linked large Control Pond C1
Linked small Control Pond C14
Isolated small Control Pond C8
Isolated large Control Pond C10
Tair (°C)
P(mm)
b 2008
c 2009
d 2010
Tair (°C), P (mm)
Conductivity (µS cm⁻¹)
DOY

Figure 3



Water Table (mm)
a
Linked *Control Pond* C1
2007
2008
2009
2010
b
Linked *Control Pond* C2
c
Isolated *Control Pond* C10
d
Isolated *Control Pond* C8
e
Frost crack CF8
f
Satellite Ponds
S33-10
S38-09
S23-09
S38-08
DOY
DOY

Figure 4

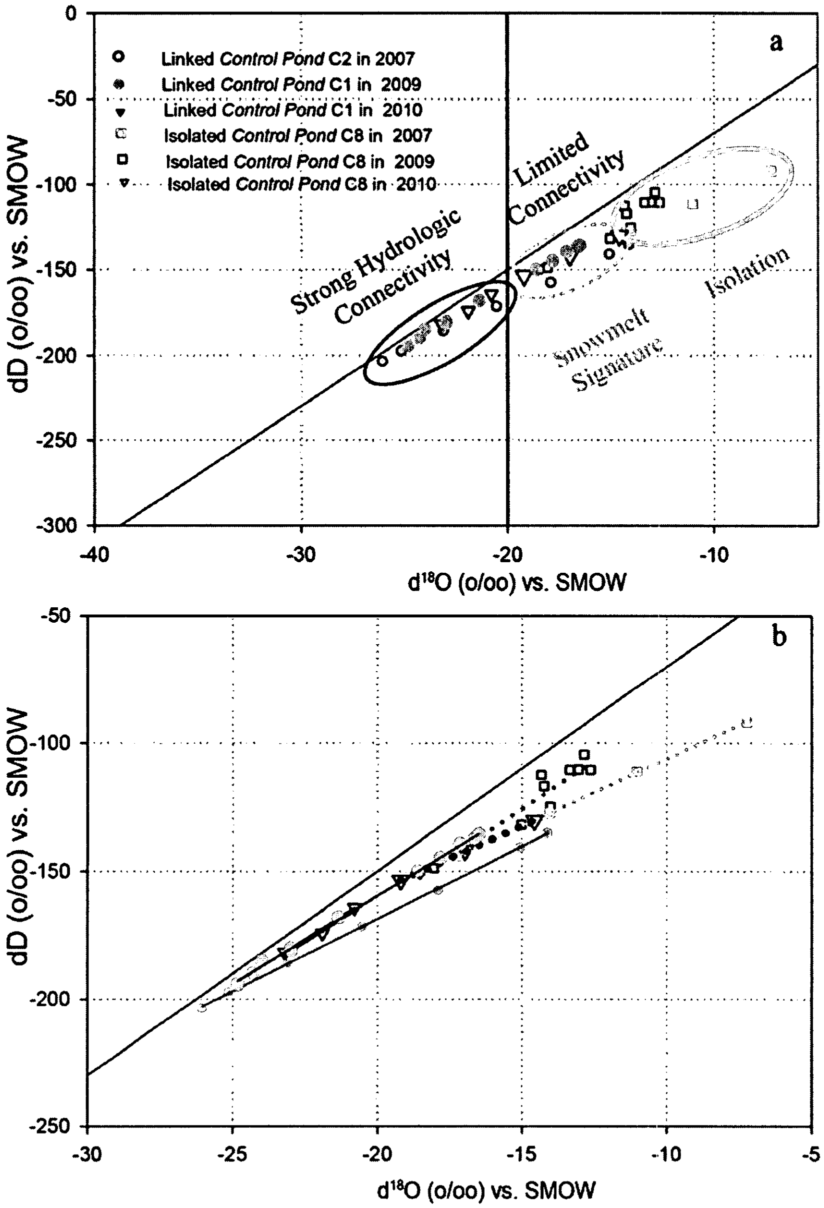

a
Linked Control Pond C2 in 2007
Linked Control Pond C1 in 2009
Linked Control Pond C1 in 2010
Isolated Control Pond C8 in 2007
Isolated Control Pond C8 in 2009
Isolated Control Pond C8 in 2010
Limited Connectivity
Strong Hydrologic Connectivity
Isolation
Snowmelt Signature
dD (o/oo) vs. SMOW
d18O (o/oo) vs. SMOW
0
-50
-100
-150
-200
-250
-300
-40
-30
-20
-10
b
dD (o/oo) vs. SMOW
d18O (o/oo) vs. SMOW
-50
-100
-150
-200
-250
-30
-25
-20
-15
-10
-5

Figure 5

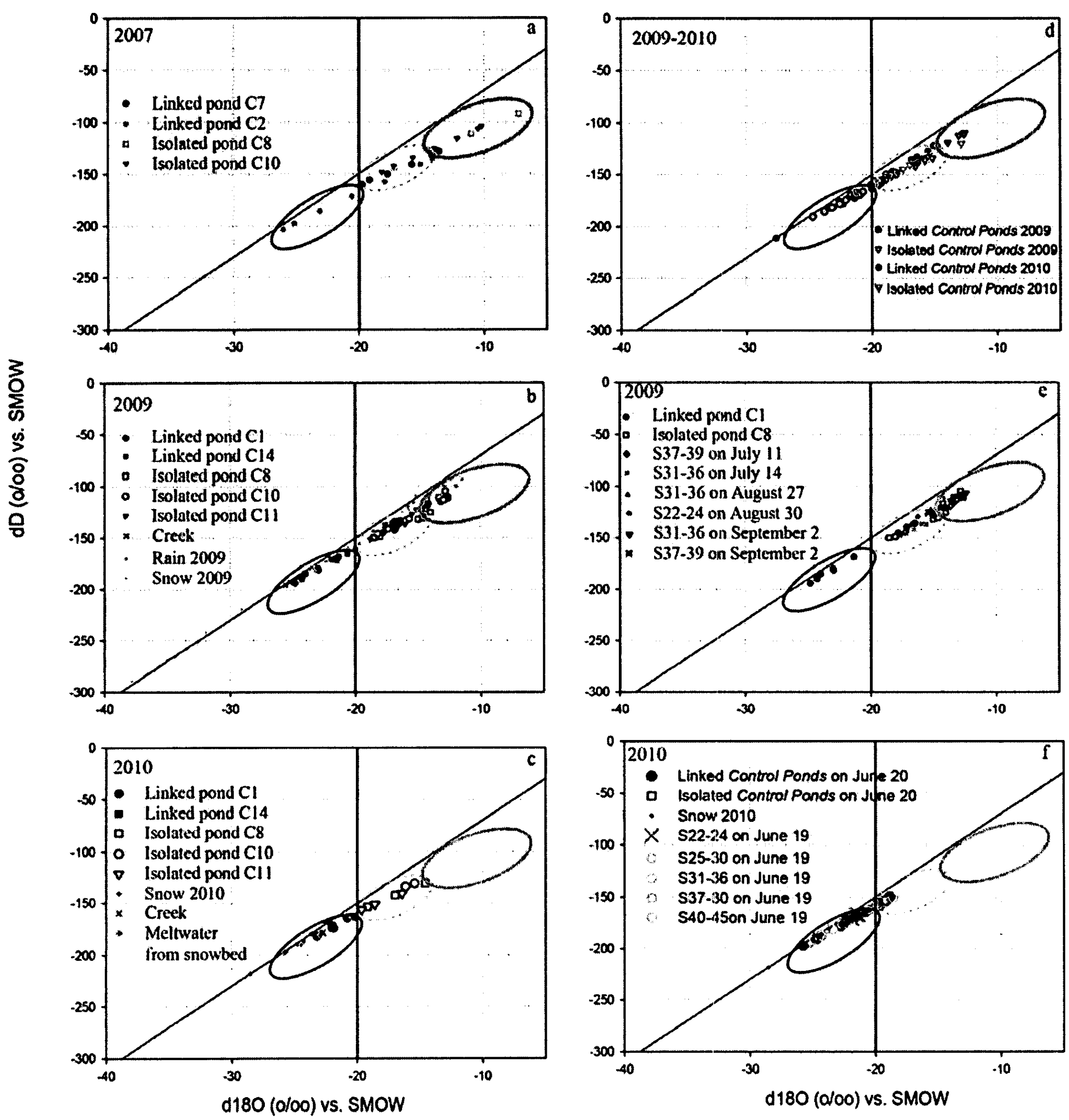

2007
a
Linked pond C7
Linked pond C2
Isolated pond C8
Isolated pond C10
2009-2010
d
Linked Control Ponds 2009
Isolated Control Ponds 2009
Linked Control Ponds 2010
Isolated Control Ponds 2010
2009
b
Linked pond C1
Linked pond C14
Isolated pond C8
Isolated pond C10
Isolated pond C11
Creek
Rain 2009
Snow 2009
2009
e
Linked pond C1
Isolated pond C8
S37-39 on July 11
S31-36 on July 14
S31-36 on August 27
S22-24 on August 30
S31-36 on September 2
S37-39 on September 2
2010
c
Linked pond C1
Linked pond C14
Isolated pond C8
Isolated pond C10
Isolated pond C11
Snow 2010
Creek
Meltwater from snowbed
2010
f
Linked Control Ponds on June 20
Isolated Control Ponds on June 20
Snow 2010
S22-24 on June 19
S25-30 on June 19
S31-36 on June 19
S37-30 on June 19
S40-45on June 19
dD (o/oo) vs. SMOW
d18O (o/oo) vs. SMOW

Figure 6

Piper Diagram

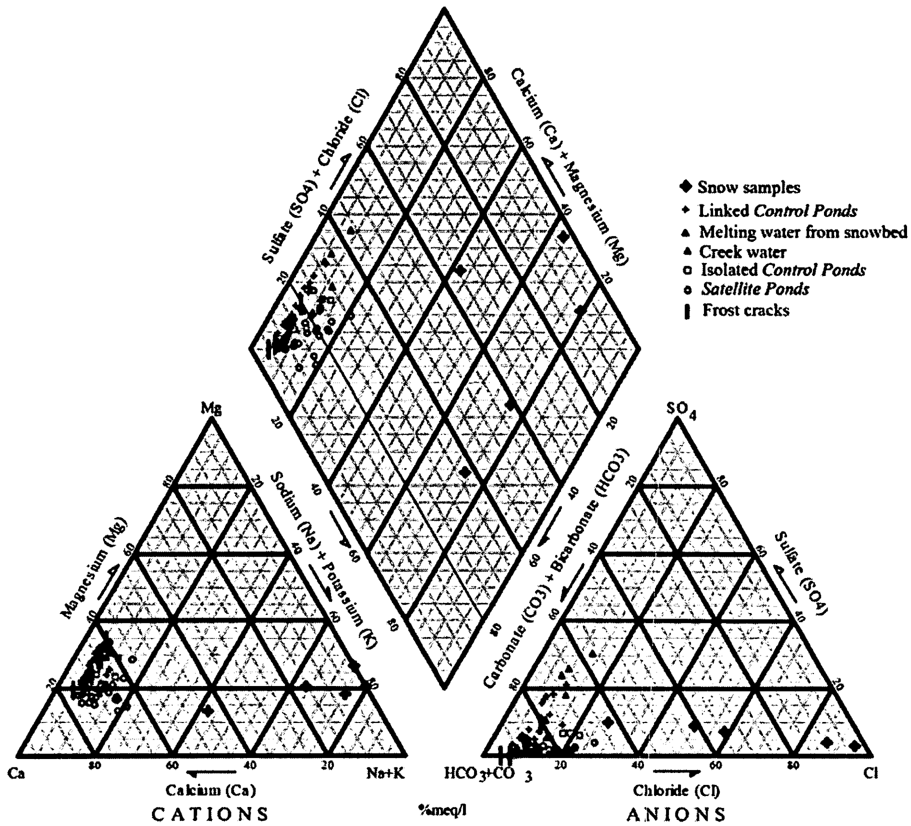
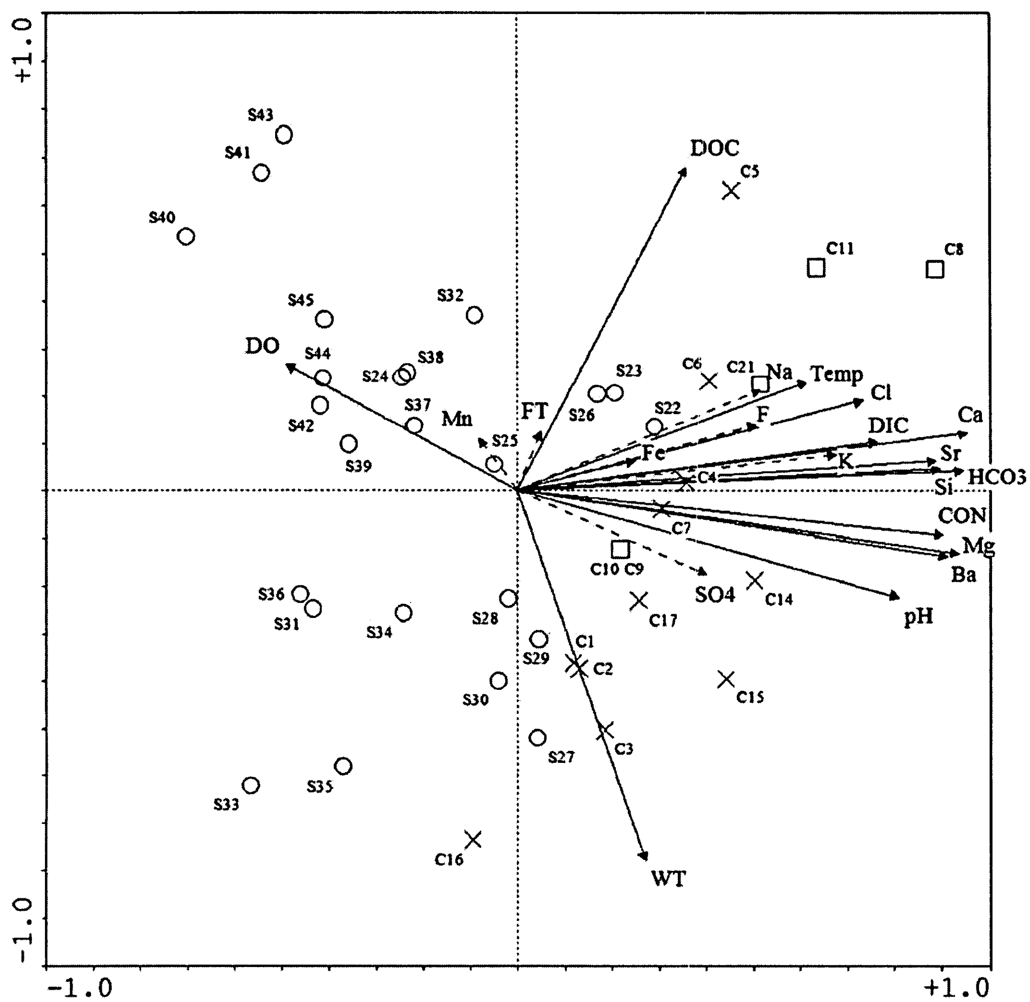
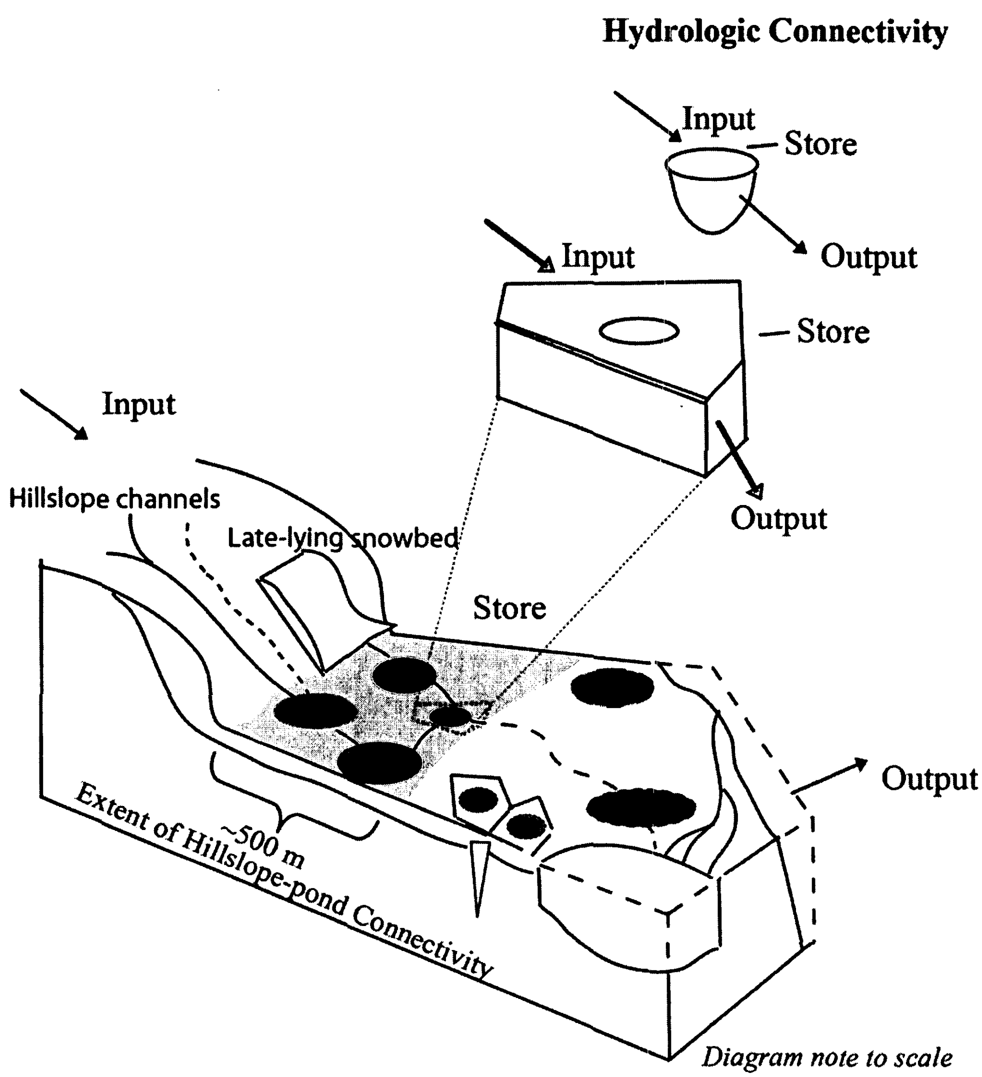


Figure 7



+1.0
-1.0
-1.0
+1.0
DOC
DO
FT
Mn
Na
Temp
Cl
F
Fe
DIC
Ca
K
Sr
Si
HCO3
CON
Mg
Ba
SO4
pH
WT
S43
S41
S40
S45
S32
S44
S38
S24
S42
S37
S39
S25
S23
S26
S22
S36
S31
S34
S28
S29
S30
S27
S33
S35
C5
C11
C8
C6
C21
C4
C7
C10 C9
C14
C17
C1
C2
C15
C3
C16

Figure 8



Hydrologic Connectivity
Input
Store
Output
Input
Store
Output
Input
Hillslope channels
Late-lying snowbed
Store
Output
~500 m
Extent of Hillslope-pond Connectivity
Diagram note to scale

Figure 9

CHAPTER 5: Conclusions

Research summary and future work

The research goal for this project was to provide a comprehensive characterization of the complex wetland system, Polar Bear Pass (PBP), by intensively investigating seasonal change in hydrology and carbon fluxes in ponds across the wetland during 2007–2010 expeditions.

The main findings of this study are:

1) PBP ponds fall into two categories: those that are hydrologically connected to their catchment by a linkage (*i.e.* receive additional sources of water from an upslope source like a late-lying snowbed, hillslope creek, or neighboring pond) and those that are hydrologically isolated (*i.e.* receive water from snow and rain only). This difference in hydrologic settings influences ponds' seasonal hydrologic stability and carbon dynamics. Linked ponds had low mean concentrations of dissolved organic and inorganic carbon and high carbon stocks identified as the storage of carbon in water column per unit area. The elevated carbon stocks indicated the importance of hydrologic connectivity as an essential carbon conduit to wetland ponds. Additionally, evaluation of carbon concentrations in the surface and subsurface waters of inflows revealed high concentrations and confirmed the importance of connectivity. Conversely, elevated dissolved carbon concentrations and small carbon stocks in isolated

ponds were a direct result of limited hydrologic connectivity and shrinkage of surface area, which led to the enrichment of solutes in the standing water column.

2) Detailed hydrologic and carbon monitoring during the snowmelt season illustrated that DOC concentrations varied, with maximum levels occurring during late snowmelt. Catchment-generated inflows of carbon into ponds accounted for these peaks. CO_2 concentrations in ponds were highest during the onset of snowmelt and in the early post-snowmelt season and were generated by water inflows enriched in DOC and DIC; the latter coincided with active layer thaw. These results highlight the complexity in physical and hydrologic processes taking place in ponds during the nival freshet.

3) This study provides the first estimates of GHGs and their variability in PBP ponds. Evaluation of GHG concentrations in all study ponds across the wetland revealed that in comparison to N_2O aquatic concentrations, all small surface waters were supersaturated in CO_2 and therefore represent strong sources of this GHG to the atmosphere. Since ponds at PBP receive their carbon from predominantly terrestrial sources that are intensified as a result of hydrologic linkages of ponds to their catchments, inflows rich in DIC and mineralization of terrestrial DOC in these ponds most likely contributes to the strong source of CO_2 to the atmosphere. The estimates of GHG emissions from our study ponds place them at the same level with most lakes and reservoirs in the arctic.

4) Intensive hydrologic and hydro-chemical monitoring and spatial surveys of ponds across the wetland indicated that the physico-chemical characteristics of PBP ponds are mainly influenced by a combination of seasonal variability of climatic conditions and hydrologic settings. There are clear differences between hydrologically linked vs. isolated ponds. While the physico-chemical characteristics of ponds across the wetland closely resemble the lithology of bedrock material found on Bathurst Island, the geochemical trends of the ponds' waters varied depending on the hydrological group and seasonal climatic conditions.

5) Upscaling from local, intensively studied ponds to infrequently monitored ponds located across the entire Pass demonstrated how the differences in hydrologic settings among the ponds, in combination with seasonal and annual variability in temperature and precipitation regimes can further alter the physico-chemical trends of pond waters. The magnitude of these controls is strongly affected by site-specific physical properties, such as hydrologic connectivity. These controls can indirectly affect pond carbon dynamics as they are heavily influenced by water levels and the influx of carbon and solutes entering the pond. Carbon delivery and storage in wetland ponds and the role of these ponds in carbon processing and contribution to the atmosphere in the form of CO_2 plays an important role, influencing both the local and wetland scale fluxes.

6) The results of this intensive hydro-chemical study expand the range of limnological data sets from the High Arctic.

The conclusions outlined in this manuscript will be improved by forthcoming research in the hydrology of this large wetland. More intensive biological investigations and nutrient sampling of PBP ponds will improve the understanding of these ecosystems' processes which are required for complete carbon balance estimates. It will be necessary to confirm the atmospheric carbon sequestration of wetland ponds during snowmelt and whether any dramatic emissions occur during the freeze-back period at this site. A biological investigation of sediment and water column will allow for identification of carbon processing mechanisms during different seasons in wetland ponds. Limited available information on frost crack hydrology and carbon dynamics points to a need for further intensive studies at this site.

It is important to note that all strategies require field-based knowledge of hydrology, observation of key processes and monitoring data of essential parameters despite the many difficulties. As Nibet (2007) says, "On-the-ground monitoring is unglamorous work, seldom rewarded by funding agencies or the science community". Understanding the influence of climate change in polar environments requires high quality in situ data that is essential for comprehension of greenhouse gas dynamics in these environments (Nibet 2007). By providing reliable data sets on the response of these sensitive small inland freshwaters to the ongoing impacts of climate change, better guidance for area management and improved sustainability strategies for the Arctic can be achieved.

References

Nibet E. 2007. Earth monitoring: Cinderella science. Nature 450(6):789-790. doi: 10.1038/450789a.

APPENDIX A: Small ponds with major impact: The relevance of ponds and lakes in permafrost landscapes to carbon dioxide emissions

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Abstract

Although ponds make up roughly half of the total area of surface water permafrost landscapes, their relevance to carbon dioxide emissions on a landscape scale has, to date, remained largely unknown. We have therefore investigated the inflows and outflows of dissolved organic and inorganic carbon from lakes, ponds, and outlets on Samoylov Island, in the Lena Delta of north-eastern Siberia in September 2008, together with their carbon dioxide emissions. Outgassing of carbon dioxide (CO_2) from these ponds and lakes, which cover 25% of Samoylov Island, was found to account for between 74 and 81% of the calculated net landscape-scale CO_2 emissions of 0.2–1.1 g C m^{-2} d^{-1} during September 2008, of which 28-43% was from ponds and 27-46% from lakes. The lateral export of dissolved carbon was negligible compared to the gaseous emissions due to the small volumes of runoff. The concentrations of dissolved inorganic carbon in the ponds were found to triple during freezeback, highlighting their importance for temporary carbon storage between the time of carbon production and its emission as CO_2. If ponds are ignored the total summer emissions of CO_2-C from water bodies of the islands within the entire Lena Delta (0.7–1.3 Tg) are underestimated by between 35 and 62%.

Introduction

The northern permafrost regions have been estimated to contain approximately 1700 Pg of organic carbon, about 90% of which occurs in permafrost deposits, representing approximately 50% of the estimated global below-ground organic carbon stock [*Tarnocai et al.*, 2009]. In a warming climate part of this large carbon stock could be released as carbon dioxide (CO_2) or methane (CH_4), thus generating a positive feedback to climate change [*McGuire et al.*, 2009]. Inland waters have been recognized as important participants in the carbon cycle, actively processing the carbon derived from terrestrial ecosystems that then makes its way into the atmosphere, oceans, and sediments [*Tranvik et al.,* 2009]. Most of the research into this aspect of the carbon cycle has addressed water bodies with a surface area of several hectares [*Kling et al.,* 1992; *Hamilton et al.*, 1994; *Duchemin et al.*, 1999; *Jonsson et al.,* 2003; *Åberg et al.*, 2004; *Repo et al.*, 2007; *McGuire et al.*, 2009], while ponds with surface areas of only a few square meters and depths measured in decimeters have received scant attention [*Boike et al.*, 2008; *Laurion et al.*, 2010] and are not taken into account for global estimates of CO_2 emissions because they are invisible to most satellites [*Muster et al.,* 2012]. However, substantial carbon emissions have been observed from small and medium-sized lakes in both sub-Arctic [*Hamilton et al.*, 1994; *Jonsson et al.,* 2003; *Huttunen et al.*, 2002a, 2003; *Repo et al.*, 2007] and Arctic environments *[Kling et al.*, 1992; *Blodau et al.*, 2008; *McGuire et al.*, 2009; *Shirokova et al.*, 2009]; a summary of CO_2 emissions from surface waters can be found in Table 1. Surface waters in Arctic and sub-Arctic environments emit on average 0.5 g CO_2-C m^{-2} d^{-1} (ranging between 0.0 and 3.0 g CO_2-C m^{-2} d^{-1}, with a standard deviation of 0.7 g CO_2-C m^{-2} d^{-1}) [*Kling et al.*, 1992; *Hamilton et al.,* 1994; *Duchemin et al.*, 1999; *Cole et al.,* 2000; *St. Louis et al.,* 2000; *Huttunen et al.*, 2002a, 2003; *Åberg et al.,* 2004; *Repo et al.,* 2007; *Blodau et al.*, 2008; *Shirokova et al.*, 2009] (Table1). Extrapolation of these fluxes to larger areas or over longer time periods remains challenging, however, because the total surface area of ponds that are invisible to satellites is virtually unknown and emissions during spring thaw and autumn freezeback periods are poorly constrained. This study aimed to fill the gap in our knowledge concerning the role of small ponds in the Arctic carbon cycle. The overall objectives were (i) to quantify the organic and inorganic carbon fluxes into/from water bodies of various sizes, and (ii) to evaluate their contributions to the larger scale CO_2 budget of a tundra region encompassing several square kilometers. The research was carried out in a typical tundra landscape on Samoylov Island, in the Lena Delta of north-eastern Siberia, during the late summer and early freezeback of 2008.

Site description

The study area is located within the continuous permafrost zone of north-eastern Siberia (Fig. 1a). This area has developed and preserved several hundreds of meters of deep, cold, permafrost as a result of the absence of extensive ice sheets during the Pleistocene and extremely cold climatic conditions (the present-day mean annual air temperature is about -14.9°C); it is believed to contain the world's largest carbon stocks in soils [*Romanovskii et al.*, 2000; *Tarnocai et al.*, 2009; *Boike et al.,* 2012]. The Lena Delta, which covers about 29,000 km^2, is the largest Arctic delta [*Gupta*, 2007] (Fig. 1b). The study area on Samoylov Island (4.3 km^2) is located in the central part of the Lena Delta (72°22' N, 126°28' E; Fig. 1c). Large numbers of (polygonal) ponds and thermokarst lakes dominate the polygonized landscape of the island (Fig. 1c-e). The ponds typically range in diameter from about 10 to 30 m, with water depths of 0.5 to 1 m. In contrast, the larger lakes have surface areas of several thousands of square meters and have water depths up to 5 m; they developed as a result of thermokarst processes and thawing of the underlying permafrost. Some lakes have developed outflow channels that eventually reach the main channel of Lena River.

The area acts as a carbon sink during July and August and becomes a CO_2 source from the beginning of September until winter [*Kutzbach et al.*, 2007]. Published methane emissions of 3.2 g CH_4-C m^{-2} yr^{-1} in small ponds and 1.6 g CH_4-C m^{-2} yr^{-1} in lakes here are relatively low [*Zhang et al.,* 2012] compared to other areas in the Arctic [*Walter Anthony et al.*, 2010], due to either limited nutrient availability or reduction of methane emissions as a result of oxidation by submerged brown mosses [*Liebner et al.,* 2011].

Methods

Water balance measurements

In order to cover a range of different freshwater systems on the island, we investigated selected ponds, thermokarst lakes, and island outflows (Fig. 1c). The water balances for lakes and ponds were assessed for the period from April to September 2008 using the following equation:

$$dS/dt = Precipitation - Evaporation + Inflow - Outflow \qquad (1)$$

where *dS/dt* is the change in water volume.

Snow surveys were conducted within the study area prior to snowmelt. Rainfall was measured with recording rain gauges (Tipping bucket 52203, RM Young, ±2%) set at a height of 0.3 m above ground level. Water levels were monitored continuously at all of the investigated lakes and ponds using pressure transducers (SensorTechnics BTE6000, ±0.2%) attached to Campbell dataloggers. The measurements were made every minute and averaged over 60-minute intervals, and their reliability was checked against manual water table measurements. Discharge rates (in m^3/s) from lake and floodplain outflows were monitored using gauging stations equipped with v-notch weirs (Thomson-V 70°, ±5-10%) established at the start of the season. Water level recorders at these stations provided regular measurements (in cm) at 60 minute intervals. The evaporation from water surfaces was evaluated using standard micrometeorological measurements: the applied methods are described in detail by *Muster et al.* [2012].

Water analyses

Water samples from lakes, ponds, and outflows were collected at weekly intervals during the study period (i.e. 6 August to 21 September 2008). Samples for dissolved organic carbon (DOC) analysis were filtered using a PE (polythene) syringe and GF (glass microfiber) filters (0.7 µm pore diameter) at the time of sampling, collected in 30 ml HDPE plastic (High-density polyethylene) containers, and acidified to pH 2 by adding 2 M HCl. Unfiltered samples for the analysis of dissolved inorganic carbon (DIC) and dissolved nitrous oxide (N_2O) concentrations were collected in headspace-free, sealed 20 ml glass vials. Samples in glass vials were stored in the dark at 4°C. Some of the glass vials were unfortunately damaged in transit, so that only the samples from September were available for the analysis of dissolved gases. Unfiltered samples for the determination of total organic carbon (TOC) concentrations were also collected in 30 ml HDPE plastic containers, and were kept frozen until laboratory analysis. Concentrations of DIC and dissolved N_2O were analyzed using a gas chromatograph (Shimadzu GC-2014AF) equipped with an AOC-5000 autosampler, a 1 m × 1/8" HayeSep Q 80/100-mesh column, an electron capture detector, and a flame ionization detector. Gas concentrations and total pressure of the headspace were analyzed after shaking the solutions at 90°C for 20 minutes on a rotary shaker. The gas concentrations in water were calculated from the headspace concentrations by applying Henry's law. Results were corrected for temperature, pressure-dependent residual gas concentrations, and pH-dependent carbonate equilibrium (only for DIC). We used the equations in *Plummer and Busenberg* [1982] for temperature adjustment of equilibrium constants.

The DOC and DIC stocks in lakes and ponds were estimated from the concentration (C) in mg/m^3 and volume (V) in m^3. The retention of carbon was calculated using the following equation:

$$Retention = C_{inflow}V_{inflow} + C_{precipitation}V_{precipitation} - C_{outflow}V_{outflow} - gaseous\ emission \quad (2)$$

The CO_2 emissions from lakes, ponds, and outflows were calculated according to *Repo et al.* [2007]:

$$\text{Gaseous Emission} = k_{gas} (C_{aq} - gas_{sat}) \tag{3}$$

where k_{gas} is the gas exchange constant, C_{aq} is the concentration of CO_2 + H_2CO_3 (H_2CO_3*) in the water, and gas_{sat} is the H_2CO_3* concentration of the water in equilibrium with the atmosphere calculated using temperature-adjusted Henry's law constants. The k_{gas} constant was estimated using the approaches of *Cole and Caraco* [1998] and *Crusius and Wanninkhof* [2003]. The dissolved concentrations of CO_2 (+ H_2CO_3) required for calculating the gaseous emissions were derived from DIC concentrations and temperature-corrected pH values [*Ben-Yaakov*, 1970], determined from the gas-tight vials in the laboratory, using temperature-adjusted dissociation constants [*Plummer and Busenberg*, 1982].

The isotopic composition of water (δD, $\delta^{18}O$) was determined for the surface water samples and for water samples obtained from a frozen soil core by an equilibration technique [*Meyer et al.*, 2000], using a mass-spectrometer (Finnigan MAT Delta-S).

CO_2 flux measurements

The net CO_2 flux (net ecosystem exchange) for the terrestrial tundra was obtained from high frequency measurements of wind speeds and CO_2 concentrations, using an eddy covariance system established within the polygonal tundra in the western part of Samoylov Island. The area around the eddy system features a comparably low density of ponds and lakes (Fig. 1c). The three-dimensional wind vectors and sonic temperatures were measured with a sonic anemometer (CSAT3, Campbell Scientific Ltd., UK), at a height of 2.4 m above ground level. The CO_2 concentrations were detected using an open path gas analyzer (LI-COR 7500). All measurements were conducted with sampling rate of 20 Hz and stored on a data logger (CR3000, Campbell Scientific Ltd., UK). A detailed technical description of the applied eddy system is provided in *Langer et al.* [2011]. The calculation of half hourly net ecosystem exchange (NEE) averages, including quality control on the stationarity and the integral turbulence characteristics, was performed using TK2 post-processing software [*Mauder et al.,* 2006]. The flux source area for each NEE value was estimated on the basis of the footprint model after *Schmid* [1994]. Those NEE values with a greater than 4% probability of being affected by lakes or ponds within the flux source areas were discarded. The meteorological station was located in the same vicinity as the eddy covariance station (Fig. 1c).

Up scaling of CO_2 emissions

Landscape-scale CO_2 emissions were calculated as a linear combination of emissions from terrestrial tundra, lakes, and ponds, as follows:

$$Landscape\ emission = A_{terr.} \times E_{terr.} + A_{lakes} \times E_{lakes} + A_{ponds} \times E_{ponds}, \quad (4)$$

where A is the dimensionless proportion of the total surface area that is made up of terrestrial tundra (*terr.*), lakes, and ponds, and E is the average CO_2 emission from terrestrial tundra, lakes, and ponds. The surface area proportions of the three landscape units were inferred from a supervised classification of high-resolution aerial images of the study area [*Muster et al.,* 2012].

Results and discussion

Water balance

Our estimation of the seasonal water budget for the studied ponds in 2008 showed that losses through evaporation were offset by similar precipitation inputs prior to the freezeback, resulting in a general state of equilibrium (Table 2). On average, evaporation rates were about 2 mm d^{-1}, with a maximum of 3 mm d^{-1}. Lake and pond water levels varied less than 10 cm during the study period. Overall, the water balance from April to September 2008 was in equilibrium, i.e. the precipitation input (233 mm) was only slightly higher than the evapotranspiration output (190 mm). The average snow water equivalent (SWE) was about 65 mm (average SWE taken from transects over the island), of which about half evaporated during the month of May (15 mm). The remaining water was most likely stored in ponds and lakes, and no visible runoff was generated (automated camera pictures). The summer rain (June-September) totaled 163 mm, most of which fell during the month of June (60 mm). The total runoff from the island during the entire summer period amounted to only about 10% of the total precipitation, illustrating the dominance of vertical water fluxes.

Carbon balance

The limnological characteristics of studied surface water bodies and chemical compositions of the freshwater samples are summarized in Table 3. Water samples collected from lakes, ponds, and outflows contained 7 to 45 mg DIC l^{-1} and were strongly supersaturated with CO_2 (Table 3). Depending on the approach used for estimating the transfer coefficient between water and atmosphere [*Cole and Caraco*, 1998; *Crusius and Wanninkhof*, 2003], this supersaturation with CO_2 resulted in CO_2 emissions of 1.4 or 2.1 g C m^{-2} d^{-1} for lakes, 1.5 or 2.2 g C m^{-2} d^{-1} for ponds, and 2.1 or 2.9 g C m^{-2} d^{-1} for outflows, prior to freezeback (Fig. 2). The DIC concentrations in ponds increased sharply during freezeback, from an average of 9–15 mg l^{-1} to 22–45 mg l^{-1}. This rapid increase in DIC concentrations resulted in peak CO_2 emissions from ponds of 10–12 g CO_2-C m^{-2} d^{-1}, levels that were not recorded for either lakes or outflows during the same period of time (Fig. 2). Although no direct measurements were made on CO_2 exchange to validate our estimates, these CO_2 emissions appear among the highest recorded to date from surface waters worldwide [*Kling et al.*, 1992; *Hamilton et al.,* 1994; *Roulet et al.*, 1997; *Cole and Caraco*, 1998; *Striegl and Michmerhuizen*, 1998; *Duchemin et al.*, 1999; *Riera et al.*, 1999; *Casper et al.*, 2000; *Cole et al.*, 2000; *St. Louis et al.*, 2000; *Huttunen et al.*, 2002a, 2002b, 2003; *Åberg et al.*, 2004; *Repo et al.*, 2007; *Blodau et al.*, 2008; *Shirokova et al.*, 2009] (see Table 1). Carbon dioxide emissions from lakes and ponds greatly exceeded the NEE of the terrestrial tundra on the island during the study period, which was approximately 0.2 g CO_2-C m^{-2} d^{-1} (Fig. 2), suggesting that small Arctic lakes and ponds represented particular hotspots for CO_2 emission.

Aerial photographs indicate that lakes and ponds cover 13% and 12%, respectively, of the island's surface, with the remaining 75% being either wet or dry tundra [*Muster et al.*, 2012] (Fig. 1c). Assuming that the island's net CO_2 emissions can be expressed as a linear combination of CO_2 emissions from tundra, lakes, and ponds (Equation 4), we can calculate an average landscape-scale CO_2 emission of 0.2–1.1 g C m^{-2} d^{-1} for September 2008 prior to freezeback, which is of the same order of magnitude as the NEE of –0.07 to 0.44 g CO_2-C m^{-2} d^{-1} previously determined for September 2003 [*Kutzbach et al.*, 2007]. The contribution to these emissions from lakes and ponds was between 74 and 81% (depending on the method used to calculate gas transfer coefficients), with roughly one half of the CO_2 originating from ponds. During freezeback the landscape-scale CO_2 emission equaled approximately 1.6 to 1.9 g C m^{-2} d^{-1}, of which about 70% was derived from ponds and about 7% from lakes. Compared to these gaseous emissions of CO_2 from the water surface, the export of carbon with flowing water during the study period was negligible because of the small volumes of runoff (Fig. 3, Table 2).

A tentative carbon budget for lakes and ponds revealed that CO_2 emissions to the atmosphere exceeded the retention of DIC and DOC as well as stocks of dissolved carbon in the water column (Fig. 3). This then raises the question of where all the CO_2 emitted to the atmosphere has come from. The required inputs of carbon could be explained by subsurface inflow, or by thawing of ground ice that is rich in dissolved carbon (average DOC concentration in ground ice: 155.07 ± 59.84 mg l^{-1}, J. Boike, unpublished data 2011), or by mineralization of organic carbon in the sediment of lakes and ponds. Similar absolute stable isotope ratios (δD and $\delta^{18}O$) and slopes of regression between δD and $\delta^{18}O$ for thermokarst lake water (slope estimate: 6.20) and permafrost ground ice (slope estimate: 6.16) suggest that thawed ice is the main source of lake water with potential mixing with precipitation (Fig. 4). The isotopic composition of pond waters is, however, determined by summer rain and evapotranspiration, as indicated by the slope of the regression line (Fig. 4). The CO_2 emissions from ponds are therefore unlikely to have been fuelled by carbon inputs from the thawing of ground ice or subsurface inflows, and sediment respiration is probably the main source of the CO_2 in these shallow ponds [*Kortelainen et al.*, 2006]. Furthermore, the oxidation of methane escaping from the sediment of ponds, for example in submerged layers of brown mosses [*Liebner et al.*, 2011], could provide an additional source of CO_2 in these water bodies.

Ponds with surface areas of less than 1000 m^2 cover 1577 km^2 of the Lena Delta, with a further 3008 km^2 covered by lakes that have areas of 3600 m^2 or more [*Muster et al.*, 2012]. By assuming an ice-free period of 100 days per year we can extrapolate the average CO_2 emissions measured from lakes and ponds in September 2008 to an overall annual emission of approximately 0.7 to 1.3 Tg CO_2-C from the entire area of islands within the Lena Delta, with the lower value calculated excluding freezeback data and transfer coefficient according to *Cole and Caraco* [1998], and the upper value including freezeback data and transfer coefficient according to *Crusius and Wanninkhof* [2003]. These emissions are underestimated by 35–62% if emissions from water bodies are neglected. In addition to the lakes and ponds within the Lena Delta, the river itself is supersaturated with CO_2 compared to the atmosphere by up to 1.5–2 fold in summer and up to 4–5 fold in winter [*Semiletov et al.,* 2011]. Therefore direct emissions of CO_2 from the river will increase the total CO_2 emissions from surface waters of the delta in addition to the emissions we report here.

Conclusions

Our results demonstrate that water bodies within polygonal tundra landscapes are hotspots for CO_2 emission from late summer until the beginning of the period of frozen ground. The CO_2 emissions from lakes and ponds are an order of magnitude higher than the net CO_2 flux observed above vegetated tundra surfaces. The findings of this study are limited to only one month due to loss of samples during transportation (September, 2008), however these high CO_2 emissions and their large areal extent indicate the importance of these water bodies in the carbon cycle of the polygonal tundra. This study has shown, for the first time, that not only lakes but also small ponds must be regarded as effective processors, transient stores, and conduits of carbon in permafrost landscapes, a conclusion that has wide-ranging implications for current evaluations of the Arctic carbon budget and its sensitivity to future changes in climate and landscape, since:

(I) The actual CO_2 emissions from vast areas of the Arctic are commonly underestimated since ponds are commonly not included in the global land surface classifications employed in estimating the Arctic carbon budget.

(II) Small variations in, for example, precipitation, evaporation, or drainage have the potential to produce considerable changes in the CO_2 emission potential of large Arctic regions as small changes in the water table level can have a major effect on the volumes and surface areas of ponds.

(III) Earth system modeling faces a formidable problem of scale since Arctic ponds are an important component in the carbon cycle of many Arctic regions but are far too small to be represented in land surface schemes used for global circulation models.

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References

Åberg, J., A. K. Bergström, G. Algesten, K. Söderback, and M. Jansson (2004), A comparison of the carbon balances of a natural lake (L. Örträsket) and a hydroelectric reservoir (L. Skinnmuddselet) in northern Sweden, *Water Res.*, *38*, 531-538, doi:10.1016/j.watres.2003.10.035.

Ben-Yaakov, S. (1970), A method for calculating the in situ pH of seawater. *Limnol.Oceanogr.*,*15*, 326–328.

Blodau, C., R. Rees, H. Flessa, A. Rodionov, G. Guggenberger, K.-H. Knorr, O. Shibistova, G. Zrazhevskaya, N. Mikheeva, and O. A. Kasansky (2008), A snapshot of CO_2 and CH_4 evolution in a thermokarst pond near Igarka, northern Siberia, *J. Geophys. Res.*, *113*, doi:10.1029/2007JG000652.

Boike, J., M. Langer, H. Lantuit, S. Muster, K. Roth, T. Sachs, P. Overduin, S. Westermann, and D. McGuire (2012), Permafrost - physical aspects, carbon cycling, databases and uncertainties, in *Recarbonization of the Biosphere – Ecosystems and the Global Carbon Cycle,* edited by R. Lal, K. Lorenz, R.F.J. Hüttl, B.U. Schneider, and J. von Braun, J. Springer, Dordrecht, The Netherlands.

Boike, J., C. Wille, and A. Abnizova (2008), The meteorology, and energy and water balances of polygonal tundra in the Lena Delta, Siberia during wet and dry years, *J. Geophys. Res.*, *113*, G03025, doi:10.1029/2007JG000540.

Brown, J., Jr. O. Ferrians, J. Heginbottom, and E. Melnikov (1997), Circum-Arctic map of permafrost and ground-ice conditions, National Snow and Ice Data Center/World Data Center for Glaciology, Digital Media, Boulder, CO.

Casper, P., S. C. Maberly, G. H. Hall, and B. J. Finlay (2000), Fluxes of methane and carbon dioxide from a small productive lake to the atmosphere, *Biogeochem.*, *49*, 1–19.

Cole, J. J. and N. F. Caraco (1998), Atmospheric exchange of carbon dioxide in a low-wind oligotrophic lake measured by the addition of SF6, *Limnol. Oceanogr.*, *43*, 647–656.

Cole, J. J., M.L. Pace, S.R. Carpenter, and J. F. Kitchell (2000), Persistence of net heterotrophy in lakes during nutrient addition and food web manipulations, *Limnol. Oceanogr.*, *45,* 1718–1730.

Crusius, J., and R. Wanninkhof (2003), Gas transfer velocities measured at low wind speed over a lake, *Limnol. Oceanogr.*, *48*, 1010–1017.

Duchemin, E., M. Lucotte, and R. Canuel (1999), Comparison of static chamber and thin boundary layer equation methods for measuring greenhouse gas emissions from large water bodies, *Environ. Sci. Technol.*, *33*, 350–357.

Gupta, A. (2007), *Large Rivers: Geomorphology and Management*, John Wiley and Sons, Chichester.

Hamilton, J. D., C. A. Kelly, J. W. M. Rudd, R. H. Hesslein, and N.T. Roulet (1994), Flux to the atmosphere of CH_4 and CO_2 from wetland ponds on the Hudson-Bay lowlands (HBLs), *J. Geophys. Res., 99*, 1495–1510.

Huttunen, J. T., T. S. Väisänen, M. Heikkinen, S. Hellsten, H. Nykänen, O. Nenonen, and P. J. Martikainen (2002a), Exchange of CO_2, CH_4 and N_2O between the atmosphere and two northern boreal ponds with catchments dominated by peatlands or forests, *Plant Soil, 242*, 137-146, doi:10.1023/A:1019606410655.

Huttunen, J. T., T. S. Väisänen, S. K. Hellsten, M. Heikkinen, H. Nykänen, H. Jungner, A. Niskanen, M. O. Virtanen, O. V. Lindqvist, and O. S. Nenonen (2002b), Fluxes of CH_4, CO_2, and N_2O in hydroelectric reservoirs Lokka and Porttipahta in the northern boreal zone in Finland, *Global Biogeochem. Cycles*, *16*, 1003, doi:10.1029/2000GB001316.

Huttunen, J. T., J. Alm, A. Liikanen, S. Juutinen, T. Larmola, T. Hammar, J. Silvola, and P. J. Martikainen (2003), Fluxes of methane, carbon dioxide and nitrous oxide in boreal lakes and potential anthropogenic effects on the aquatic greenhouse gas emissions, *Chemosphere*, *52*, 609-621, doi:10.1016/S0045-6535(03)00243-1.

Jonsson, A., J. Karlsson, and M. Jansson (2003), Sources of Carbon Dioxide Supersaturation in Clearwater and Humic Lakes in Northern Sweden, *Ecosystems*, *6*, 224-235, doi:10.1007/s10021-002-0200-y.

Kling, G. W., G. W. Kipphut, and M. C. Miller (1992), The flux of CO_2 and CH_4 from lakes and rivers in arctic Alaska, *Hydrobiologia*, *240*, 23-36, doi:10.1007/BF00013449.

Kortelainen, P., M. Rantakari, J. T. Huttunen, T. Mattsson, J. Alm, S. Juutinen, T. Larmola, J. Silvola, and P. J. Martikainen (2006), Sediment respiration and lake trophic state are important predictors of large CO_2 evasion from small boreal lakes, *Glob. Change Biol., 12*, 1554-1567, doi:10.1111/j.1365-2486.2006.01167.x.

Kutzbach, L., C. Wille, and E. M. Pfeiffer (2007), The exchange of carbon dioxide between wet arctic tundra and the atmosphere at the Lena River Delta, Northern Siberia, *Biogeosciences*, *4*, 869-890.

Langer, M., S. Westermann, S. Muster, K. Piel, and J. Boike (2011), The surface energy balance of a polygonal tundra site in northern Siberia–Part 1: Spring to fall, *The Cryosphere*, *5*, 151-171.

Laurion, I., W. F. Vincent, S. Macintyre, L. Retamal, C. Dupont, P. Francus, and R. Pienitz (2010), Variability in greenhouse gas emissions from permafrost thaw ponds, *Limnol. Oceanogr.*, *55*, 115-133.

Liebner, S., J. Zeyer, D. Wagner, C. Schubert, E. M. Pfeiffer, and C. Knoblauch (2011), Methane oxidation associated with submerged brown mosses reduces methane emissions from Siberian polygonal tundra, *J. Ecol.*, *99*, 914–922, doi:10.1111/j.1365-2745.2011.01823.x.

Mauder, M., C. Liebethal, M. Göckede, J. P. Leps, F. Beyrich, and T. Foken (2006), Processing and quality control of flux data during LITFASS-2003, *Bound.-Lay. Meteorol., 121*, 67-88, doi:10.1007/s10546-006-9094-0.

McGuire, A. D., L. G. Anderson, T. R. Christensen, S. Dallimore, L. Guo, D. J. Hayes, M. Heimann, T. D. Lorenson, R. W. Macdonald, and N. Roulet (2009), Sensitivity of the Carbon Cycle in the Arctic to Climate Change, *Ecol. Monogr.*, *79*, 523-555.

Meyer, H., L. Schonicke, U. Wand, H. W. Hubberten, and H. Friedrichsen (2000), Isotope studies of hydrogen and oxygen in ground ice-experiences with the equilibration technique, *Isotopes Environ. Health Stud.*, *36*, 133-150.

Muster, S., M. Langer, B. Heim, S. Westermann, and J.Boike (2012), Subpixel heterogeneity of ice-wedge polygonal tundra: a multi-scale analysis of land cover and evapotranspiration in the Lena River Delta, Siberia, *Tellus B*, *64*, 17301.

NASA Landsat Program (2000), Lena Delta, in *Landsat 7/ETM+, Visible Earth, v1 ID: 18024,* USGS EROS Data Center Satellite Systems Branch, 10/10/2011.

Plummer, L. N., and E. Busenberg (1982), The solubilities of calcite, aragonite and vaterite in CO_2-H_2O solutions between 0 and 90°C, and an evaluation of the aqueous model for the system $CaCO_3$-CO_2-H_2O, *Geochim. Cosmochim. Acta, 46,* 1011–1040.

Repo, M. E., J. T. Huttunen, A. V. Naumov, A. V. Chichulin, E. D. Lapshina, W. Bleuten, and P. J. Martikainen (2007), Release of CO_2 and CH_4 from small wetland lakes in western Siberia, *Tellus B*, *59*, 788-796, doi:10.1111/j.1600-0889.2007.00301.x.

Riera, J. L., J. E. Schindler, and T. K. Kratz (1999), Seasonal dynamics of carbon dioxide and methane in two clear-water lakes and two bog lakes in northern Wisconsin, U.S.A. *Can. J. Fish. Aquat. Sci., 56,* 265–274.

Romanovskii, N. N., H.-W. Hubberten, A. V. Gavrilov, V. E. Tumskoy, G. S. Tipenko, M. N. Grigoriev, and C. Siegert (2000), Thermokarst and Land-Ocean Interactions, Laptev Sea Region, Russia, *Permafr. Periglac. Proc.*, *11*, 137-152.

Roulet, N. T., P. M. Crill, N. T. Comer, A. Dove, and R. A. Boubonniere (1997), CO_2 and CH_4 flux between a boreal beaver pond and the atmosphere, *J. Geophys. Res.*, *102*, 29313-29319, doi:10.1029/97JD01237.

Schmid, H. (1994), Source areas for scalars and scalar fluxes. *Bound.–Lay. Meteorol.*, *67*, 293–318.

Semiletov, I.P., I.I. Pipko, N.E. Shakhova, O.V. Dudarev, S.P. Pugach, A.N. Charkin, C.P. McRoy, D. Kosmach, and Ö. Gustavson (2011), Carbon transport by the Lena River from its headwaters to the Arctic Ocean, with emphasis on fluvial input of terrestrial particulate organic carbon vs. transport by coastal erosion, *Biogeosciences*, *8*, 2407–2426.

Shirokova, L. S., O. S. Pokrovsky, S. N. Kirpotin, and B. Dupré (2009), Heterotrophic bacterio-plankton in thawed lakes of the northern part of Western Siberia controls the CO_2 flux to the atmosphere, *Int. J. Environ. Stud.*, *66*, 433-445, doi:10.1080/00207230902758071.

St. Louis, V. L., C. A. Kelly, É. Duchemin, J. W. M. Rudd, and D. M. Rosenberg (2000), Reservoir surfaces as sources of greenhouse gases to the atmosphere: A global estimate, *Bioscience*, *50*, 766-775, doi:10.1641/0006-3568(2000)050[0766:RSASOG]2.0.CO;2.

Striegl, R. G., and C. M. Michmerhuizen (1998), Hydrologic influence on methane and carbon dioxide dynamics at two north-central Minnesota lakes, *Limnol. Oceanogr.*, 1519-1529.

Tarnocai, C., J. G. Canadell, E. A. G. Schuur, P. Kuhry, G. Mazhitova, and S. Zimov (2009), Soil organic carbon pools in the northern circumpolar permafrost region, *Global Biogeochem. Cycles*, *23*, GB2023, doi: 10.1029/2008GB003327.

Tranvik, L. J., J. A. Downing, J. B. Cotner, S. A. Loiselle, R. G. Striegl, T. J. Ballatore, P. Dillon, K. Finlay, K. Fortino, L. B. Knoll, P. L. Kortelainen, T. Kutser, S. Larsen, I. Laurion, D. M. Leech, S. L. Mccallister, D. M. Mcknight, J. M. Melack, E. Overholt, J. A. Porter, Y. T. Prairie, W. H. Renwick, F. Roland, S. S. Bradford, D. W. Schindler, S. Sobek, A. Tremblay, M. J. Vanni, A. M. Verschoor, E. V. Wachenfeldt, and G. A. Weyhenmeyer (2009), Lakes and reservoirs as regulators of carbon cycling and climate, *Limnol. Oceanogr.*, *54*, 2298-2314.

Walter Anthony, K. M., D. A. Vas, L. Brosius, F. S. C. Iii, S. A. Zimov, and Q. Zhuang (2010), Estimating methane emissions from northern lakes using ice-bubble surveys, *Limnol. Oceanogr. Methods*, *8*, 592-609.

Zhang, Y., T. Sachs, L. Changsheng, and J. Boike (2012), Upscaling methane fluxes from closed chambers to eddy covariance based on a permafrost biogeochemistry integrated model, *Glob. Change Biol.*, *18*, 1428-1440, doi: 10.1111/j.1365-2486.2011.02587.x.

Tables

Table 1. Reported CO_2 fluxes from a range of lakes and ponds in forest (F), forest tundra (FT), tundra (T), peatland (P), forest peatland (FP) and fens and bogs (FB) landscapes.

Study	Location	Latitude	Land cover	Year/ period
Åberg et al. [2004]	Northern Sweden	Boreal	F	2001
Åberg et al. [2004]				
Blodau et al. [2008]	Northern Siberia	Boreal	FT	2006
Casper et al. [2000]	The English Lake District, U.K.	Temperate	FB	1997
Cole and Caraco [1998]	Central New Hampshire	Temperate	F	1991–95
Cole et al. [2000]	Wisconsin	Temperate	F	1991–98
Duchemin et al. [1999]	Taiga region of the Canadian shield	Boreal	F	1994
Hamilton et al. [1994]	Hudson Bay Lowlands	Boreal	P	1990
Huttunen et al. [2002a]	Northern Finland	Boreal	P	1994
			FP	1995
Huttunen et al. [2002b]	Northern Finland	Boreal	FP	1994–95
			FP	1994
Huttunen et al. [2003]	Northern Finland	Boreal	F	1996–98
			F	1996
			F	1997–98
			F	1997–98
			F	1997–98
Kling et al. [1992]	Arctic Alaska	Arctic	T	1988, 1990
Repo et al. [2007]	Northern Russia	Boreal	F	2005
			FT	2005
Rierra et al. [1999]	Northern Wisconsin	Temperate	F	1994
Roulet et al. [1997]	Thompson, Manitoba	Boreal	F	1996
Shirokova et al. [2009]	Western Siberia	Boreal	FT-T	2008
Striegl and Michmerhuizen [1998]	North-central Minnesota	Temperate	F	1992–93
St. Louis et al. [2000] given by *Huttunen et al.* [2003]				

Study site	Emission	
	(g CO_2 m^{-2} d^{-1})	(g C m^{-2} d^{-1})
Lake Skinmuddselet	1.06	0.29
Lake Öträsket	0.94	0.26
Ponds near Igarka	0.07	0.02
Priest Pot	1.76	0.48
Mirror Lake min	0.29	0.08
Mirror Lake max	0.51	0.14
Four lakes min	-0.66	-0.18
Four lakes max	2.20	0.60
La Grande 2 and Laforge-1 reservoirs	2.50	0.68
Lake de Voeuxt	0.16	0.04
10 ponds min	3.70	1.01
10 ponds max	11.00	3.00
Jänkäläisenlampi pond	0.53	0.14
Kotsamolampi pond	0.02	0.00
Reservoir Lokka	1.52	0.41
Reservoir Porttipahta	1.54	0.42
Lake Postilampi	0.79	0.21
Lake Heinälämpi	0.57	0.16
Lake Kevätön	0.64	0.17
Lake Vehmasjärvi	0.86	0.23
Lake Mäkijärvi	0.39	0.11
45 lakes	0.90	0.25
MT lake	0.50	0.14
MT pond	1.60	0.44
FT lake	1.50	0.41
Crystal bog	1.34	0.37
Trout bog	2.01	0.55
Crystal lake	0.02	0.00
Sparkling Lake	0.22	0.06
Beaver pond	6.20	1.69
Lakes and ponds	1.1	0.30
Williams Lake	0.02	0.01
Shingobee Lake	1.61	0.44
Temperate and boreal reservoirs excluding Lokka and Porttipahta	0.22	0.06

Table 2. Water balance estimates for Samoylov Island in 2008.

	Total precipitation (mm)	Evapotranspiration (mm)	Storage (mm)
April (snow)	65	0	65
May	6	-15	-10
June	60	-52	8
July	39	-56	-18
August	43	-44	-3
September	21	-23	-4
TOTAL:	233	-190	38

Table 3. Limnological and chemical characteristics as well as concentrations of carbon and N_2O, together with isotopic compositions, for ponds and lakes of Samoylov Island during the study period from 6 August to 21 September 2008 (mean values, with ranges in parentheses). Freezeback values from 21 September 2008 are italicized. Note: n/a stands for not available data, an asterisk (*) indicates analytes sampled only from September 5 to September 21.

	Thermokarst lake 1	Thermokarst lake 2	Pond 1	Pond 2	Lake outlet	Flood plain outlet
Trophic state	Oligotrophic	Oligotrophic	Oligotrophic	Oligotrophic		
Latitude N	8032363	8032778	8031992	8031981	8032840	8034141
Longitude E	414855	415607	414946	414928	416191	415027
Volume (m^3)	104820	96829	70	12	-	-
Surface area (m^2)	45982	40714	221	130	-	-
Area/Volume	0.44	0.42	3.16	10.54	-	-
Maximum depth (m)	6.40	5.7	0.18	0.5	-	-
Ca (mg/l)	7.5 *8.8*(2.4–9.2)	6.7 *6.9* (5.8–7.0)	4.6 *5.1*(2.4–6.2)	14.5 *15.4*(7.2–25.8)	6.2 *6.1*(3.1–7.9)	29.0 *n/a*(23.8–34.5)
Mg (mg/l)	3.4 *4.0*(1.0–4.1)	3.3 *3.5*(2.9–3.5)	3.2 *3.6*(1.6–4.1)	7.9 *8.7*(4.6–12.2)	3.0 *3.0*(1.5–3.8)	12.3 *n/a* (11.0–13.9)
Na (mg/l)	2.6 *3.1*(0.8–3.1)	0.6 *0.7*(0.5–0.7)	0.6 *0.7*(0.3–0.8)	1.1 *1.2*(0.8–1.3)	1.5 *1.5*(0.7–1.8)	1.6 *n/a* (1.2–2.1)
K (mg/l)	0.6 *0.6*(0.4–0.9)	0.5 *0.6*(0.4–0.6)	0.9 *1.2*(0.4–1.1)	0.7 *0.8*(0.3–0.9)	0.4 *0.5*(<0.2–0.4)	0.4 *n/a* (0.3–0.5)
Fe (µg/l)	105.0	21.3	211.4	907.9	301.0	359.3

	94.9(42.8–180.0)	*23.2*(<20–49.5)	*134.0* (114.0–506.0)	*231.0*(114.0–2660.0)	*118.0*(118.0–727.0)	*n/a* (234.0–570.0)
Mn (µg/l)	<20 *<20*(<20–<20)	<20 *<20*(<20–<20)	<20 *<20*(<20–<20)	413.1 *53.1*(<20–1500)	49.7 *<20*(<20–139.0)	343.5 *n/a* (32.4–782.0)
Si (mg/l)	0.3 *0.5*(<0.1–0.5)	0.3 *0.3*(0.2–0.5)	0.4 *0.7*(<0.1–0.7)	2.1 *4.7*(0.5–3.0)	0.6 *0.7*(0.2–1.0)	2.0 *n/a* (1.5–2.2)
Sr (µg/l)	43.1 *50.0*(<20–53.1)	33.5 *33.4*(29.4–35.4)	29.6 *31.0*(<20–42.2)	78.0 *84.5*(52.5–98.9)	31.8 *31.8*(<20–41.3)	155.8 *n/a* (109.0–198.0)
Cl (mg/l)	3.7 *3.8*(2.9–4.4)	0.7 *0.7*(0.6–0.7)	0.7 *0.9*(0.6–0.9)	0.9 *1.5*(0.4–1.1)	1.8 *1.9*(1.3–2.0)	0.4 *n/a* (<0.1–0.8)
SO_4 (mg/l)	1.8 *1.9*(1.2–2.0)	0.1 *<0.1*(<0.1–0.1)	<0.1 *<0.1*(<0.1–<0.1)	<0.1 *<0.1*(<0.1–<0.1)	0.7 *0.8*(0.5–0.9)	0.1 *n/a* 0.1–0.5)
Alkalinity (mmol/l)	0.8	2.3	0.9	1.5	0.7	*n/a*
Conductivity (µS/cm)	92.2 *92.0* (86.0–100.0)	65.5 *66.0* (63.0–67.0)	57.7 *60.0* (53.0–61.0)	121.6 *126.0* (78.0–225.0)	69.0 *65.0* (64.0–80.0)	255.4 *n/a* (208.0–307.0)
DO (mg/l)	7.8 *8.8*(6.8–8.9)	10.0 *9.1*(8.3–14.6)	7.2 *7.6*(4.5 –9.9)	5.2 *5.5*(2.1–8.3)	9.5 *9.4*(7.0–13.9)	6.1 *n/a* (3.1–9.7)
DOC (mg/l)	4.0 *4.0*(2.8–5.6)	2.1 *2.6*(1.7–2.6)	4.2 *4.2*(3.1–5.4)	6.8 *7.4*(4.2–14.4)	2.8 *3.3*(1.9–3.8)	4.3 *n/a* (2.6–6.5)
DIC*(mg/l)	13.1 *13.2*(11.6–14.4)	8.3 *7.8*(7.0–10.2)	13.6 *22.3*(9.2–22.3)	25.0 *45.4*(12.5–45.4)	11.6 *14.8*(9.3–14.8)	41.8 *n/a* (40.3–43.2)
pH	7.2 *7.2*(6.9–7.3)	7.2 *7.4*(7.2–7.4)	6.8 *6.5*(6.5–7.1)	7.2 *6.8*(6.8–7.5)	7.3 *7.8*(7.0–7.8)	7.5 *n/a* (7.5–7.5)
N_2O-N*(µg/l)	0.46 *0.53*(0.39–0.53)	0.42 *0.38*(0.38–0.45)	0.26 *0.00*(0.00–0.42)	0.31 *0.0*(0.00–0.47)	0.44 *0.41*(0.41–0.49)	0.43 *n/a* (0.39–0.47)
$\delta^{18}O$ (‰)	-17.4 *-16.3*(-17.8– -	-16.9 *-16.8*(-17.0– -	-14.7 *-15.4*(-15.4–	-15.9 *-17.7*(-17.7– -	-17.8 *-18.0*(-18.3– -	-16.8 *n/a* -

	16.3)	16.7)	14.3)	13.9)	15.7)	18.4– - 15.7)
□D(‰)	-136.4 *-129.1*(- 139.0– -129.1)	-133.5 *-132.6*(- 134.7– -132.2)	-122.2 *-124.9*(- 124.9– -120.7)	-125.9 *-136.2*(- 136.2– -119.3)	-140.1 *-140.9*(- 143.7– -124.7)	-131.0 *n/a* (- 137.9– - 125.0)

Figures

Figure 1. (a) Distribution of circumpolar permafrost [*Brown et al.,* 1997] and the Lena Delta. (b) Lena River Delta, Eastern Siberia [*NASA Landsat Program,* 2000], and the location of Samoylov Island. (c) Aerial mosaic image of Samoylov Island with visible thermokarst lakes, ponds, and outlets, showing micrometeorological station locations, polygonized ponds, and patterned ground. (e) Pond 1.

Figure 2. Outgassing of CO_2 from lakes, ponds, and water outflows at three points in time, in relation to net ecosystem exchange (NEE) determined from eddy covariance measurements. Daily values of NEE between 1 September and 30 September 2008 are presented as a box plot. The box frames values between the 25^{th} and 75^{th} percentiles, the horizontal line represents the median, and whiskers show the 10^{th} and 90^{th} percentiles. Extreme values are shown as crosses. Circles, squares, and triangles indicate emissions on 5, 12, and 21 September 2008, respectively, estimated using k_{gas} derived according to *Cole and Caraco* [1998] in black, and *Crusius and Wanninkhof* [2003] in white.

Figure 3. Carbon mass balance of the studied water bodies (lakes and ponds) from 1 August to 21 September 2008, in g C m^{-2}. Values in parentheses represent CO_2 emissions recorded on 21 September 2008, during freezeback.

Figure 4: Isotopic composition ($\delta^{18}O$ and δD) of lakes, ponds, and outflows during the study season from 6 August to 21 September 2008, and of permafrost ground ice. Solid,

thick black line: regression for lakes; dashed black line: regression for permafrost ice core; solid grey line: regression for ponds.

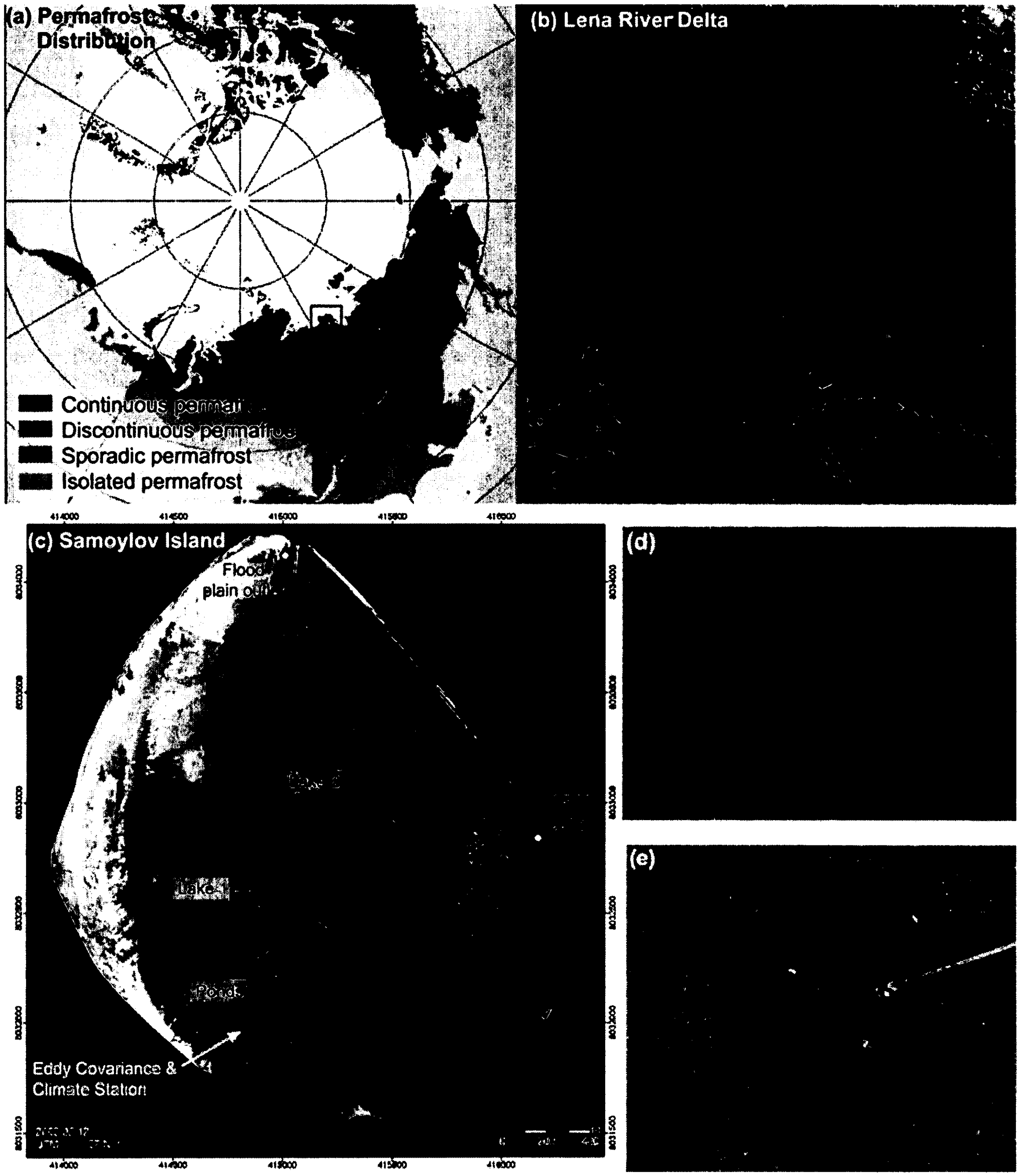
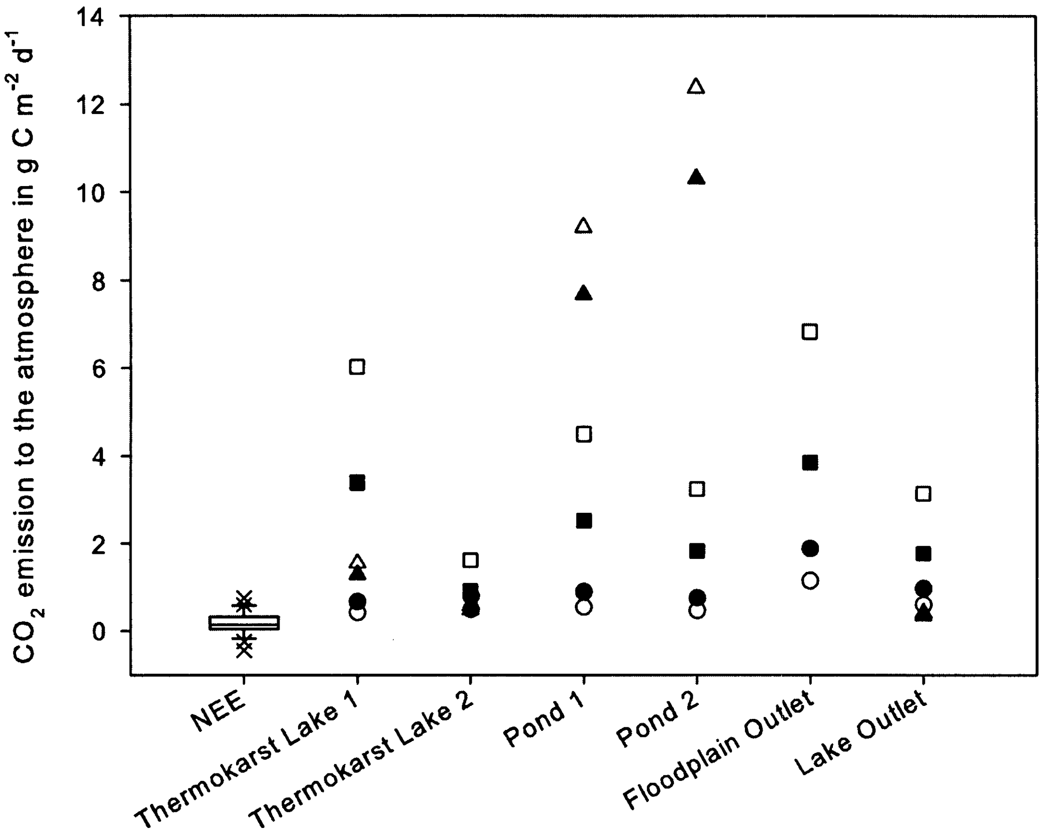

(a) Permafrost Distribution
Sporadic permafrost
Isolated permafrost
(b) Lena River Delta
(c) Samoylov Island
Lake 1
Pond
Eddy Covariance &
Climate Station
(d)
(e)

Figure 1



CO2 emission to the atmosphere in g C m-2 d-1
14
12
10
8
6
4
2
0
NEE
Thermokarst Lake 1
Thermokarst Lake 2
Pond 1
Pond 2
Floodplain Outlet
Lake Outlet

Figure 2

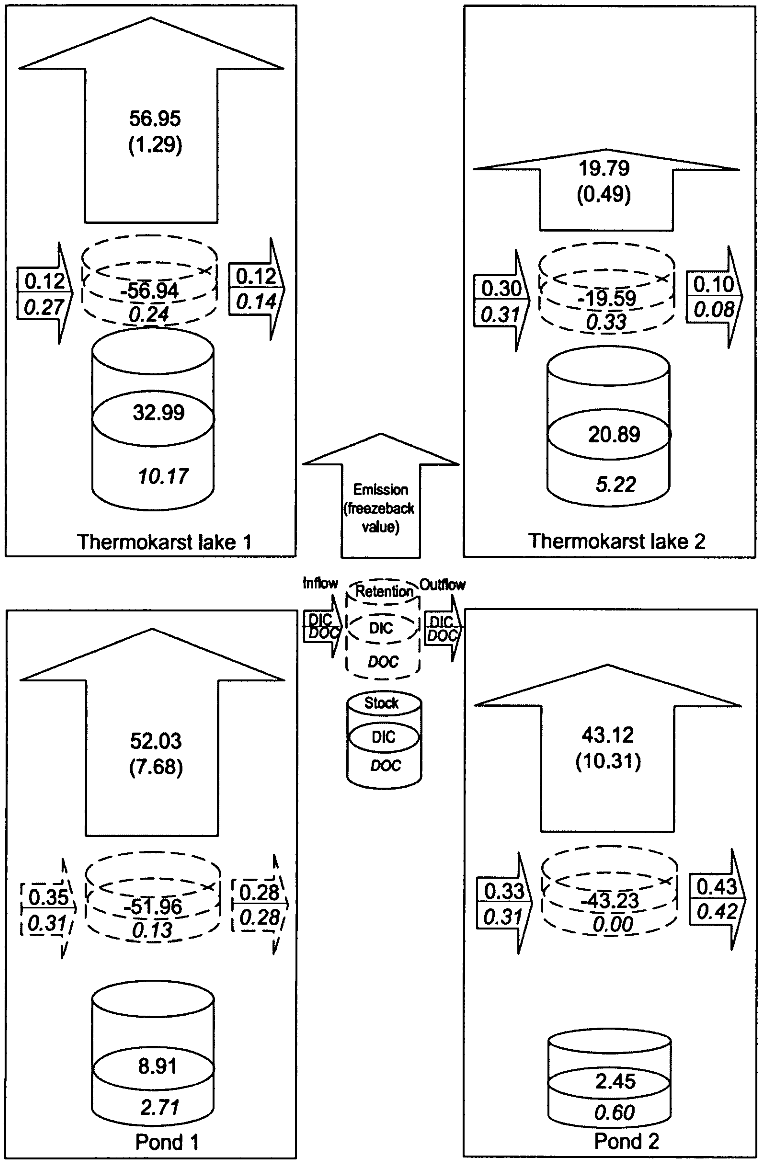
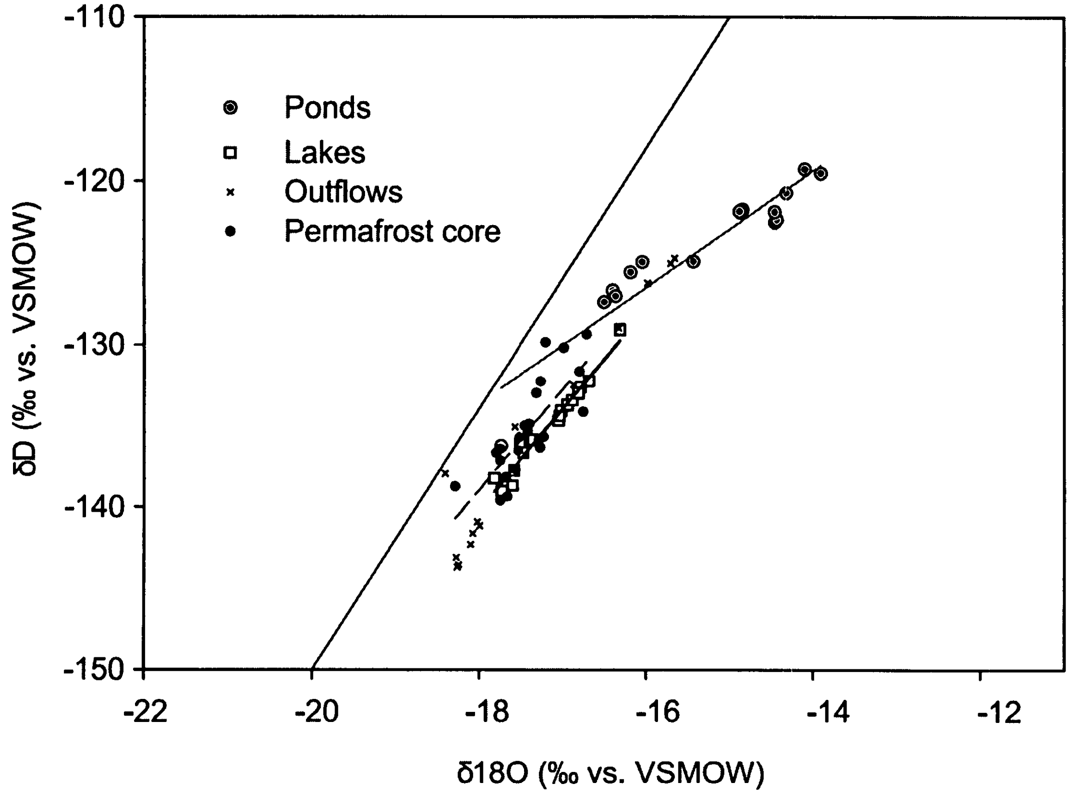

56.95
(1.29)
0.12
0.27
-56.94
0.24
0.12
0.14
32.99
10.17
Thermokarst lake 1
19.79
(0.49)
0.30
0.31
-19.59
0.33
0.10
0.08
20.89
5.22
Thermokarst lake 2
Emission
(freezeback
value)
Inflow
DIC
DOC
Retention
DIC
DOC
Outflow
DIC
DOC
Stock
DIC
DOC
52.03
(7.68)
0.35
0.31
-51.96
0.13
0.28
0.28
8.91
2.71
Pond 1
43.12
(10.31)
0.33
0.31
-43.23
0.00
0.43
0.42
2.45
0.60
Pond 2

Figure 3



Ponds
Lakes
Outflows
Permafrost core
-110
-120
-130
-140
-150
-22
-20
-18
-16
-14
-12
δD (‰ vs. VSMOW)
δ18O (‰ vs. VSMOW)

Figure 4