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Source: Arctic, Antarctic, and Alpine Research, 46(1) : 272-285

Published By: Institute of Arctic and Alpine Research (INSTAAR), University of Colorado

URL: https://doi.org/10.1657/1938-4246-46.1.272

Arctic, Antarctic, and Alpine Research, Vol. 46, No. 1, 2014, pp. 272–285

Observed and Projected Climate Change in the Churchill Region of the Hudson Bay Lowlands and Implications for Pond Sustainability

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Abstract

There is concern over the fate of surface water bodies at high latitudes as a consequence of rising global temperatures. The goal of this study is to characterize climatic change that has occurred in the northern Hudson Bay Lowlands (HBL), Canada, from 1943 to 2009, to determine if this has resulted in a change to pond surface areas and to predict if changes may continue in the future. Climate change and changes to pond volume and size over the past ~60 years were examined using a combination of field methods/instrumental records (1943–2009), modeling (1953–2009; 1961–2100), and remote sensing/imagery analyses (1947–2008). Results demonstrate that temperatures are warming and breakup dates are earlier, but this has not significantly increased the duration of the open-water period or pond evaporation rates, which can be highly variable from year to year. Annual precipitation, primarily summer rainfall, has increased, lessening the summer moisture deficit and leading to wetter conditions. The observed changes of a smaller summer moisture deficit are predicted to continue in future, although there is less certainty with predictions of future precipitation than there is with predictions of air temperature. Thus, ponds are likely not at risk for drying and instead may be at risk for expansion. Despite the increases in summer rainfall, imagery analysis of 100 ponds shows that pond surface areas have fluctuated over the study period but have not increased in size.

DOI: http://dx.doi.org/10.1657/1938-4246-46.1.272

Introduction

Shallow water bodies are a ubiquitous feature of Arctic coastal plains of Siberia, northern Alaska, and Canada. In Canada, they are particularly prevalent in the Hudson Bay Lowlands (HBL) and the Mackenzie River Delta region. Within the subarctic environment, shallow pond features are estimated to occupy between 15% and 50% of the total land area (Duguay and Pietroniro, 2005) and contribute to regional chemical, energy, biological, and hydrological systems (Rouse et al., 1997; Prowse et al., 2006; White et al., 2007). These water bodies are often rich in biodiversity (Smol and Douglas, 2007) and are a habitat for wildlife as well as a breeding ground for migratory shorebirds. There has been growing interest from the scientific community regarding the status of shallow water bodies at high latitudes, as their growth or disappearance may be an indication of the effects of climate change (Schindler and Smol, 2006). Changes in pond extent and distribution may significantly impact the regional ecosystems and modify biophysical regimes (White et al., 2007; Rühland et al., 2013). Thus, more information is needed to assess trends in hydrologic storage in ponds and shallow lakes in the Canadian Arctic and subarctic, and to predict how these systems might respond under a changing climate.

It is anticipated that the effects of climatic change will be particularly significant at high latitudes (IPCC, 2007), with warmer temperatures and a longer open-water season (Serreze et al., 2000). The increased air and surface water temperatures and length of the open-water season are expected to lead to increases in evaporation (E) (e.g., Rouse et al., 1997). However, changes in precipitation (PPT) patterns are less clear (Serreze et al., 2000; Hinzman et al., 2005) and appear to vary regionally. In many regions across the Arctic and subarctic, the increased E caused by warming has not been balanced by increases in PPT, leading to more negative surface water budgets (i.e., PPT – E] (Hinzman et al., 2005; Woo and Guan, 2006; Smol and Douglas, 2007). This has the potential to lead to a widespread drying of shallow water bodies, as these systems are dynamic and highly responsive to changing climate conditions due to their high surface area–to–depth ratios (Rouse et al., 1997; Woo and Guan, 2006).

Several studies (Yoshikawa and Hinzman, 2003; Stow et al., 2004; Smith et al., 2005; Riordan et al., 2006; Labrecque et al., 2009) have used remotely sensed images to show decreasing lake surface areas over the past 50 years. Unfortunately, due to the resolution of the imagery used in most of these studies (about 15- to 200-m spatial resolution), only lakes have been studied and little is known about ponds. A Canadian study (Smol and Douglas, 2007), based on field observations over several decades, demonstrated that ponds on Ellesmere Island are drying. However, this study did not suggest the spatial extent to which this has occurred. Overall, there is a paucity of data on the fate of the many small, shallow water bodies at high latitudes, and variability across regions is not known. This is confounded by the fact that the extent of pond drying may also vary regionally with permafrost, soil type, and pond size/bathymetry.

Climate warming has accelerated permafrost thaw in many regions (Rouse et al., 1997; Frey and McClelland, 2009; Dyke and Sladen, 2010), leading to a drying of upland areas and an impounding of drainage in subsiding areas (Woo et al, 1990; Jorgenson et al., 2006). Surface water in the landscape presents a positive feedback that enhances permafrost degradation in wet areas (Jorgenson et al., 2006), which may either increase the availability of water to lakes and ponds (Payette et al., 2004; Smith et al., 2005; Prowse et al., 2006; Sannel and Kuhry, 2011) or increase hydrologic connections between lakes and their surroundings, causing them to drain (Woo et al., 1992; Yoshikawa and Hinzman, 2003).

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Much of the previous research on the fate of ponds has been conducted in Alaska, Siberia, and western Canada, but less is known about their fate in central Canada, which differs hydrologically and climatologically (Petrone et al., 2000). The Hudson Bay Lowlands (HBL) area spans central Canada and is the largest contiguous wetland in North America. The northern portion of the HBL is underlain by continuous permafrost, whereas discontinuous permafrost is more prevalent in the southern HBL. Shallow open-water features are prevalent in the region, covering 25% to 40% of the landscape (Bello and Smith, 1990; Duguay and Lafleur, 2003; Macrae et al., 2004; Dyke and Sladen, 2010). The microclimate of the northern HBL is heavily influenced by Hudson Bay (Bello and Smith, 1990; Petrone et al., 2008), which has warmed considerably over the past half century (Gagnon and Gough, 2005a; Kaufman et al., 2009; Hochheim and Barber, 2010; Hochheim et al., 2010), causing it be ice-free for a greater duration of the year (Gagnon and Gough, 2005b; Mallory et al., 2010). This warming pattern is anticipated to continue in the future (Gagnon and Gough, 2005a; Dyke and Sladen, 2010), increasing air temperatures and lengthening the open-water season over adjacent land areas (Sannel and Kuhry, 2011). Increased precipitation is expected in winter due to warmer air temperatures (IPCC, 2007), although increases in the magnitude and intensity of summer rainfall events are also possible (Serreze et al., 2000), particularly given the retreat of the sea ice cover on Hudson Bay (Gagnon and Gough, 2005a). The cumulative effects of climate change in the northern HBL on pond hydrologic storage are not known.

This study aims to answer three key questions regarding the potential responses of the water balances of subarctic ponds to climate change from the second half of the 20th century to the end of the 21st century: (1) Have there been changes in the 1947–2009 observational period in the Churchill region of the HBL? (2) What are the implications of climate change for pond sustainability in the northern HBL? (3) Are trends expected to continue until the end of the 21st century based on regional climate model scenarios?

Study Area

Research for this study was conducted within the Hudson Plains ecozone, near Churchill, Canada (58°44′N, 94°49′W, Fig. 1), along the western shore of Hudson Bay. The region has an Arctic climate that is heavily influenced by the presence of Hudson Bay (Rouse, 1998). Annual mean air temperature is –6.9 °C, with a mean July air temperature of 12 °C in July and –27 °C in January (Environment Canada, 2009). Annual precipitation averages approximately 432 mm, 44% of which falls as snow (Environment Canada, 2009).

The topography of the Churchill region is a gentle sloping plain (1 m km^{-1}) (Winter and Woo, 1990) that has been undergoing isostatic rebound since glacial retreat, at a rate of approximately 1 m $century^{-1}$ (Johnson et al., 1987). The HBL is underlain by Ordovician and Silurian limestone and dolomite, beneath which Precambrian rocks such as sedimentary gneiss, granitoid, and volcanic rocks are found (Dredge, 1992). A layer made up of a mixture of marine silt deposits and Pleistocene glacial till is found on top of the bedrock, which is overlain by peat that ranges in depth from a minimum of a few centimeters by the coast of Hudson Bay to a maximum of more than 4 meters in areas far inland from the coast (Dredge, 1992).

Much of the northern HBL is underlain by continuous permafrost, with active layer depths typically less than 1 m (Dyke and Sladen, 2010). The region is poorly drained and dominated by coastal wetlands, although some open-canopy spruce-lichen woodlands are present. The wetlands are predominantly a mixture of polygonal peat bog plateaus and channel fens. Numerous ponds are present, occupying up to 50% of the landscape in some areas (Dyke and Sladen, 2010). The ponds are situated close to one another, and many do not have well-defined catchment areas. Ponds in the coastal zone (i.e., the study ponds, located within 10 km of the coast) typically occupy depressions with restricted drainage from a mixture of permafrost, rocks, and marine silts (Dyke and Sladen, 2010), whereas ponds farther inland, south of the coast of Hudson Bay, are thermokarst features (Dredge, 1992). The coastal ponds range from 30 m to nearly 1 km in diameter (Macrae et al., 2004), and most are less than 1 m in depth and consequently freeze to the bottom in winter. These ponds also host thick deposits of highly organic sediments (Macrae et al., 2004) that also freeze in winter. Many of the ponds appear to be hydrologically disconnected from surrounding peatlands for a portion of the summer months, experiencing short periods of drying and sediment exposure due to large summer E and small amounts of PPT (Boudreau and Rouse, 1995; Macrae et al., 2004), although ponds receiving hydrologic inputs from channel fens are less prone to drying (Wolfe et al., 2011).

Methods

Data mining, modeling, and imagery analyses were used to address the objectives of this paper. Changes in climate between 1943 and 2009 were characterized using archived field data and modeled data where field data were not available (e.g., duration of open-water period, evaporation). The effects of climate change on pond hydrologic storage during the contemporary period (1947–2009) were characterized using both imagery analyses and basic water balance modeling. The characterization of a future climatic scenario was done using models (Canadian Regional Climate Model, CRCM 4.2.0; Canadian Lake Ice Model, CLIMo). Details are provided in the following sections.

DATA

Observations of Meteorological Data (1943–2009)

Data used in this study were obtained from the meteorological station at the Churchill airport, where data have been collected since 1943 (Environment Canada, 2009). Daily observations of precipitation (snow and rain), air temperature, wind speed, relative humidity, and cloud cover fraction were compiled.

Meteorological variables from the Churchill airport weather station (air temperature, humidity, cloud cover, wind speed, and snow accumulation) were used to force the Canadian Lake Ice Model (CLIMo) (Duguay et al., 2003), for the purpose of deriving the timing of ice cover freeze-up (ice-on) and breakup (ice-off) dates (and open-water season duration) and surface water temperature for a pond with a mean water depth of 1 m for the contemporary period (1953–2009). This period was chosen as several of the forcing variables required for the model were not available, or were incomplete, in the 1943–1946 period. CLIMo is a well-tested, one-dimensional thermodynamic model used for freshwater ice-cover studies (e.g., Ménard et al., 2002; Duguay et al., 2003; Jeffries et al., 2005; Morris et al., 2005; Labrecque et al., 2009; Brown and Duguay, 2011a) capable of simulating ice-on and ice-off, thickness, and composition of the ice cover (clear or snow ice). CLIMo has been shown to perform very well at simulating lake ice phenology when using input data that well represent the climate for the

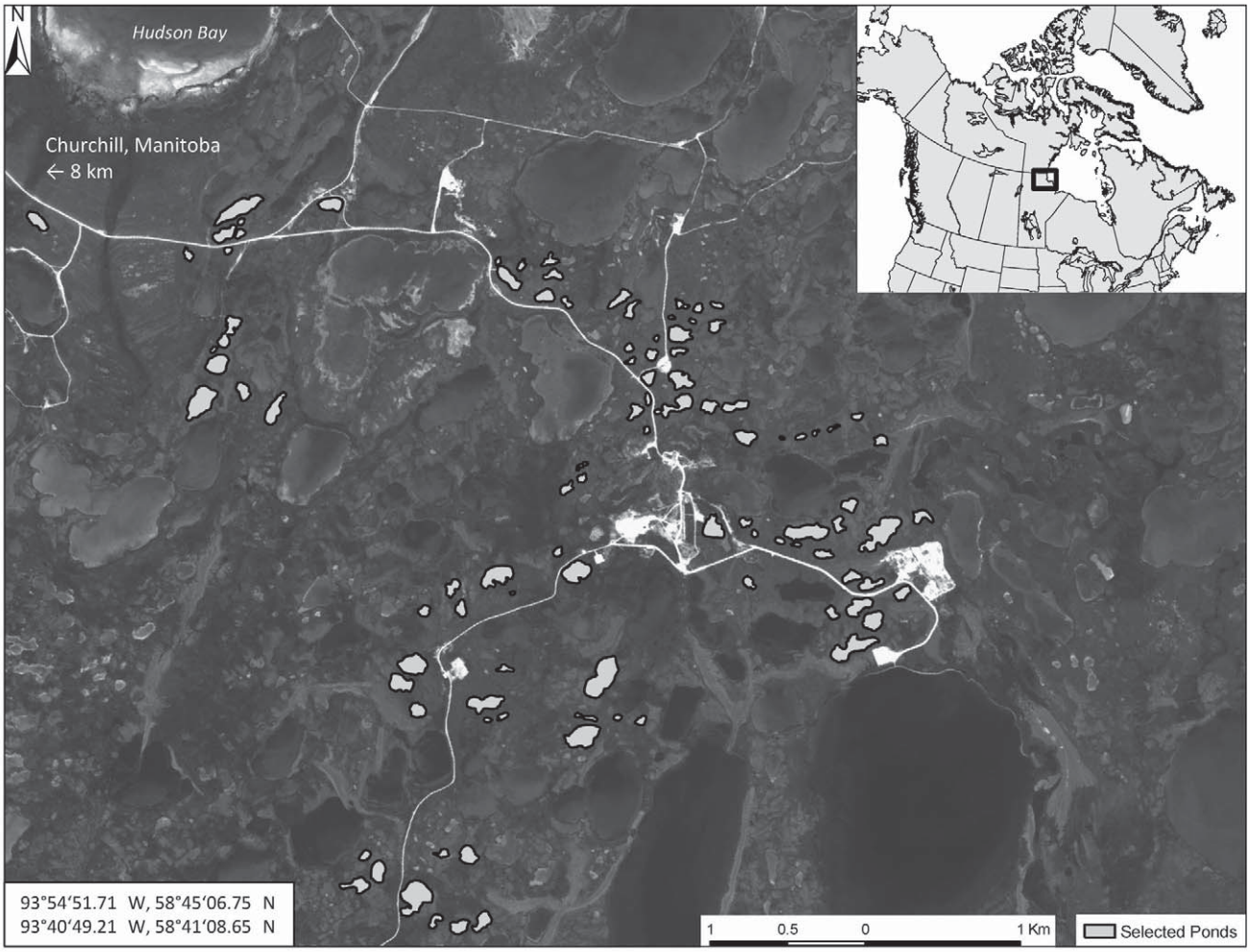

FIGURE 1. Location of the study area in the Hudson Bay Lowlands near Churchill, Manitoba, Canada. The subset of 100 ponds used in a change detection study (1947–2008) is shown with light gray shading and thick black outline.

lake—for example, from nearby meteorological towers (e.g., in the Churchill area: Duguay et al., 2003; Brown and Duguay, 2011a).

CLIMo has been modified from the one-dimensional sea ice model of Flato and Brown (1996), which was based on the one-dimensional unsteady heat conduction equation, with penetrating solar radiation, of Maykut and Untersteiner (1971) (see Duguay et al., 2003) and includes a fixed-depth mixed layer in order to represent an annual cycle. When open water is present, the mixed layer temperature is computed from the surface energy budget and hence represents a measure of the heat storage in the lake. When ice cover is present, the mixed layer is fixed at the freezing point. The water column of shallow lakes is typically isothermal and well mixed from top to bottom during the ice-free period, which allows the mixed layer depth to be a good approximation of the effect of lake depth leading to autumn freeze-up. A more detailed description of CLIMo can be found in Duguay et al. (2003).

Imagery (1947–2008)

Images from 1947, 1956, 1972 (mosaics of four aerial photographs, 1:35,000 scale; Natural Resources Canada), and 2008 (SPOT 5 panchromatic image, 0.48–0.71 μm, 2.5 m resolution) were compared to examine changes in pond surface area. One hundred ponds (surface areas 400–40,000 m^2, Fig. 1) were randomly selected and manually digitized on-screen at a scale of 1:2320 using ESRI ArcMap. Ponds were delineated using a set of rules regarding tone, size, shape, texture and orientation. In order to calculate digitization accuracy, a subset of 24 manually digitized surface areas from the 2008 SPOT image were compared with field-collected surface area data acquired one week prior to imagery acquisition (Trimble GeoXT GPS). A comparison of the two datasets revealed a strong relationship (r^2 = 0.92) with an overall root mean square error of 172 m^2. The areas of the digitized ponds were calculated and compared in order to determine feature growth or decay between 1947 and 2008.

Regional Climate Model Scenario (1961–2100)

The future scenario data (1961–2100) were produced by the Canadian Regional Climate Model (CRCM 4.2.0) (45 km true at 60°N) provided by Consortium Ouranos. CRCM is a limited-area model, originally developed at Université du Québec à Montréal

(UQAM), driven at the boundaries by GCMs or reanalysis data. CRCM uses the Canadian LAnd Surface Scheme (CLASS 2.7; Verseghy, 1991; Verseghy et al., 1993) to describe the water and energy exchanges between land surface and atmosphere (Music and Caya, 2007). For a detailed description of CRCM, see Caya and Laprise (1999) and Laprise (2008). CRCM was driven at the boundaries with the Canadian Global Climate Model (CGCM 3.1/T47 member 4), following the IPCC Special Report on Emission Scenarios A2 greenhouse-gas scenario of continually increasing CO_2 emissions. CGCM data are produced by the Canadian Centre for Modelling and Analysis (CCCma). CLIMo was used to produce the future ice cover simulations using daily CRCM data consisting of 2-m screen temperature (bias corrected following Brown and Duguay [2011b] using the Adjusted and Homogenized Canadian Climate Data [Vincent and Gullet, 1999; AHCCD, 2009]; available from Environment Canada), humidity (specific humidity converted to relative humidity using a calculated saturated vapor pressure as a function of temperature [Beljaars et al., 1989] and a fixed air pressure of 1015 mb), wind speed, water equivalent of snow, snow density, and cloud cover amounts.

ESTIMATES OF WATER BALANCE (1953–2009)

Pond water evaporation was estimated using the approach of Oswald and Rouse (2004), a mass transfer equation based on Fick's first law of diffusion via

$$E = K_e u_z \left(e_o - e_z\right) \tag{1}$$

where E is evaporation (mm d^{-1}), K_e is the coefficient of proportionality, u is the horizontal wind speed at height z, e_o is the vapor pressure of saturated air at the temperature of the water surface (estimated using CLIMo, as in Labrecque et al., 2009), and e is the vapor pressure of the air at the height above the water surface. A value of 0.81 was used for K_e, which was derived statistically using published values for summer evaporation rates from ponds collected over a 4-year period in the Churchill region (Boudreau and Rouse, 1995; Petrone et al., 2000; Yee, 2008). Results from this model provide an average, *regional* estimate of evaporation for shallow-water features (1 m mean depth) within the region. Individual pond evaporative fluxes are expected to vary with pond size, bathymetry, and underlying sediment composition. A mean pond water depth of 1 m was used because observed water levels for Golf Lake (approximately 1 m deep) were used to validate the models. Thus, findings are more representative of moderately sized ponds (e.g., 1 m deep). Additional work is needed for smaller, shallower ponds.

After the main meltwater period, precipitation and evaporation predominantly control pond water levels within the Churchill region (Boudreau and Rouse, 1995; Macrae et al., 2004), although some ponds receive runoff via channelized flow (Wolfe et al., 2011). Surface and subsurface flow contribute to the overall water balance during major rain events; however, these are relatively uncommon during summer months within the Churchill region (Environment Canada, 2009). Therefore, the water balances for ponds lacking surface connections with surrounding peatlands may be modeled using the equation

$$\Delta S = PPT - E, \tag{2}$$

where ΔS is change in pond water storage, PPT is precipitation and E is evaporation (Macrae et al., 2004). It is assumed that pond water levels begin at bankfull storage after the major melt period and oscillate due to atmospheric exchanges until freeze-up. It is understood that overland flow commences during all instances where total cumulative PPT exceeds E and any additional water is lost to the surrounding area. Although permafrost degradation is projected to occur under warmer temperatures in this region (Dyke and Sladen, 2010), which could supply runoff to adjacent ponds (e.g., Prowse et al., 2006; Woo and Guan, 2006), for simplification, our simple water balance model in Equation 2 does not include such lateral seepage.

STATISTICAL ANALYSES

Annual, seasonal, and monthly trends in air temperature, precipitation, ice cover, and evaporation were examined using the nonparametric Mann-Kendall test for detecting the presence of the monotonic increasing or decreasing trends and the nonparametric Sen's method for estimating the slopes of linear trends (significant trends $0.01 \leq \alpha \leq 0.1$).

Results and Discussion

VARIABILITY AND CHANGES IN CLIMATE (1943–2009)

Annual air temperatures have increased at a rate of +0.02 °C yr^{-1} (+0.2 °C per decade), or +1.02 °C over the 66-year period (p = 0.05; Fig. 2). These changes have been greatest during the winter (0.04 °C yr^{-1}, p = 0.05) and spring (0.03 °C yr^{-1}, p = 0.05) seasons, but have also been observed in summer (0.02 °C yr^{-1}, p = 0.01). No change has been observed during the autumn months ($p > 0.1$). Increases in air temperature have been most pronounced between 1998 and 2009 (Fig. 2), which followed a cooling period for approximately three decades prior (Rouse, 1998). Observed temperatures are similar to what has been described by others (e.g., Rouse, 1998; Kaufman et al., 2009; Rühland et al., 2013), although they demonstrate a slightly smaller change than reported by Gagnon and Gough (2005b) (0.5 °C per decade), who used climatic data from 1971–2000 in their study. A recent study demonstrated that the observed temperature increases have led to a significant shift in lake biological communities and suggested that the climate of the HBL has passed a tipping point (Rühland et al., 2013).

Annual precipitation and rainfall were highly variable among years, but significant increases in precipitation were observed over the 66-year period (Fig. 3). The increases in annual precipitation (+1.55 mm yr^{-1}, p = 0.05) are attributed to increases in rainfall (+1.47 mm yr^{-1}, p = 0.01) as significant differences in annual snowfall were not observed (Fig. 3). The increased rainfall amounts received in the summer (June, July, August—JJA) (+0.78 mm yr^{-1}, p = 0.05) months were primarily driven by increased rainfall in August (+0.41 mm yr^{-1}, p = 0.05), and rainfall amounts did not change significantly in June, July, September, or October between 1943 and 2009, although significant increases in rainfall for the period of 1953–2009 were observed in July (+0.43 mm yr^{-1}, p = 0.05), August (+0.63 mm yr^{-1}, p = 0.01), and September (+0.43 mm yr^{-1}, p = 0.1). Increases in annual precipitation in North America have been observed and projected by others (e.g., McCabe et al., 2001; Gagnon and Gough, 2005b; Finnis et al., 2007; Rawlins et al., 2010; Shook and Pomeroy, 2012), but this has largely involved increases in winter precipitation due to increased cyclonic activity for the period between September and March in other regions (McCabe et al., 2001; Finnis et al., 2007). Thus, although increases in annual precipitation in the Churchill region have been observed,

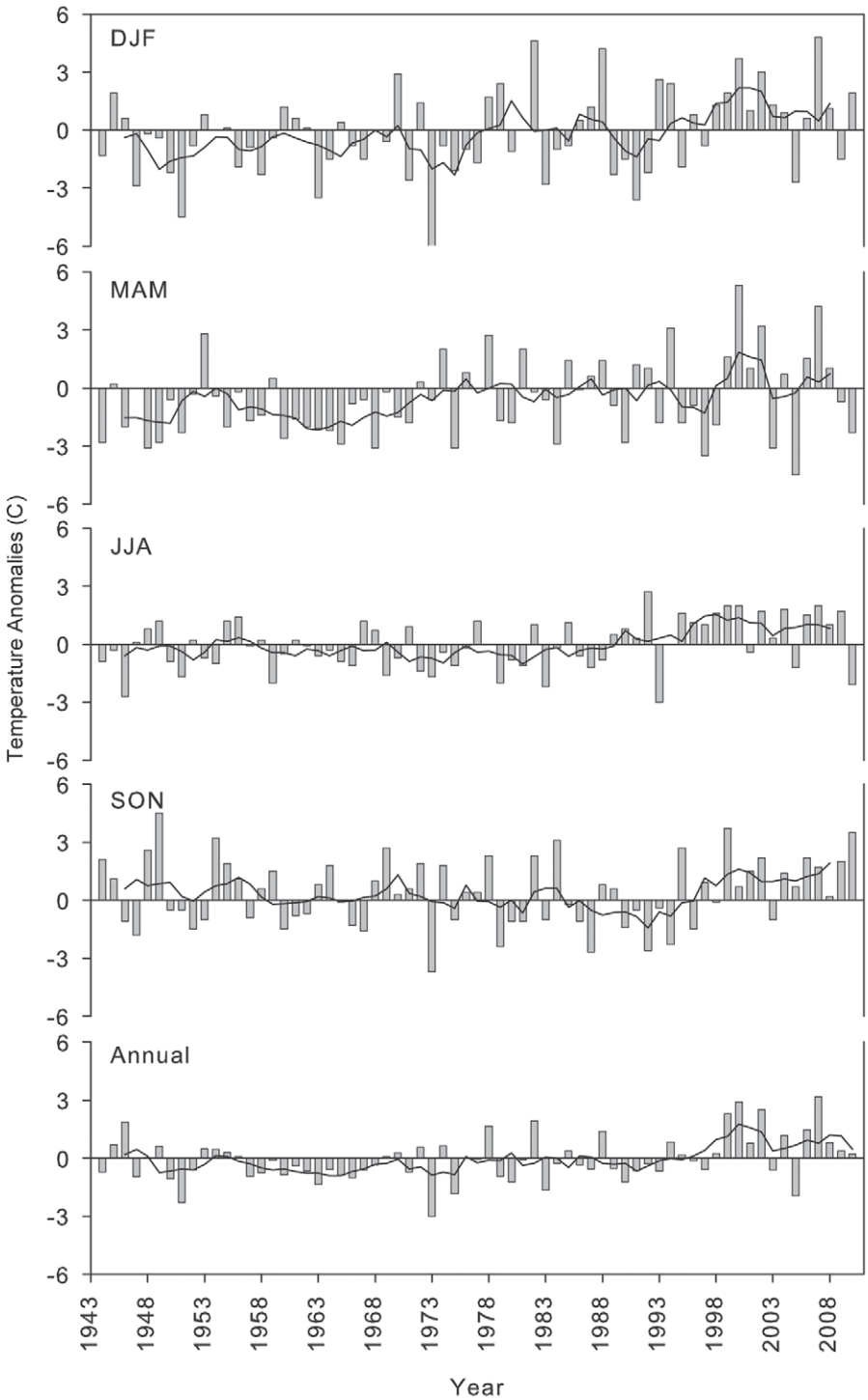

FIGURE 2. Normalized difference in seasonal and annual mean air temperatures recorded at Churchill, Canada, 1943–2009. Differences (degrees Celsius) are anomalies from the 1971–2000 mean. The solid line represents a five-year running mean.

the seasonal distribution of these increases is in contrast to what has been observed elsewhere. Determining the mechanism behind the increased summer rain events is beyond the scope of the current study but is likely due to Hudson Bay being ice-free for a longer portion of the year, providing a warmer surface and atmosphere in summer capable of holding and transporting more water vapor to adjacent areas (Gagnon and Gough, 2005b; Petrone et al., 2008; Hochheim et al., 2010; Hochheim and Barber, 2010; Mallory et al., 2010). A recent study examining the hydrologic character of rainfall in the Canadian prairies noted increased uniformity of summer rainfall, with a reduction in single-day summer rainfall events but an increase in multiple-day summer rainfall, and speculated that frontal precipitation events have increased (Shook and Pomeroy, 2012).

The duration of the pond open-water period (Fig. 4) did not increase in length between 1953 and 2009 ($p > 0.1$). However, ear-

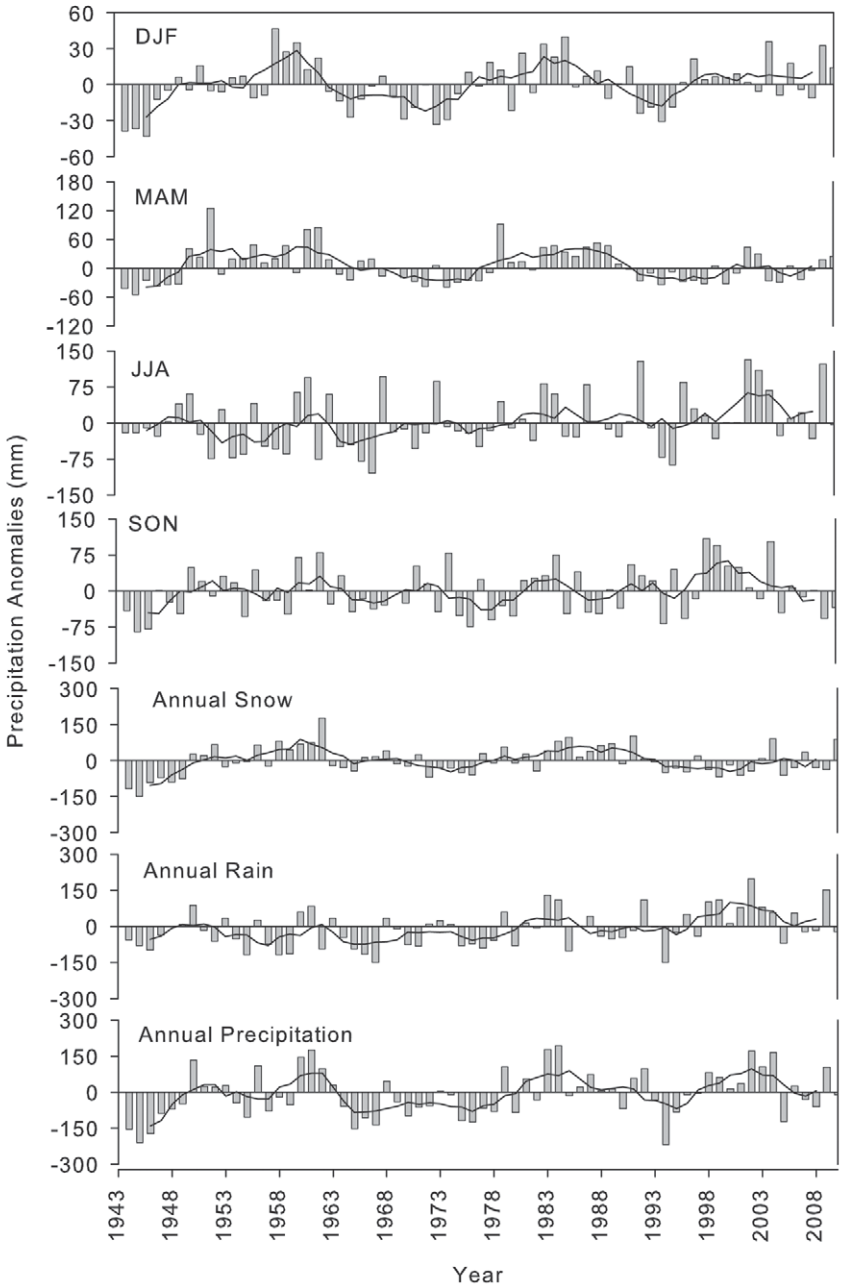

FIGURE 3. Normalized differences in seasonal and annual precipitation recorded at Churchill, Canada, 1943–2009. Differences are anomalies from the 1971–2000 mean. The solid line represents a five-year running mean. Different scales are used on the y-axes.

lier spring pond breakup dates were observed (6.1 days earlier from 1953 to 2009, or –0.11 days yr^{-1}, $p = 0.1$), whereas autumn freeze-up dates did not change significantly. The findings of a significant change in spring pond breakup dates but no significant change in the length of the ice-free season results from high variability in the data set, as the length of the open-water season fluctuated from 91 to 148 days over the record period. These patterns are similar to what has been observed over Hudson Bay near Churchill by Gagnon and Gough (2005a), who also observed earlier breakup dates (–0.8 days yr^{-1}) but did not find significant changes in the formation of ice cover. Often, high-latitude lakes host persistent ice long into the summer (Schindler and Smol, 2006), and ponds and lakes are only ice-free for 1–3 months each year. A lengthening of the ice-free season over the ponds is significant as it may lead to warmer water and sediment temperatures, and may affect pond ecology and biogeochemical processes (Rouse et al., 1997). A longer open-water season, combined with warmer air temperatures, may also increase evaporative losses from ponds and increase the potential for pond drying (Rouse et al., 1997; Smol and Douglas, 2007).

Despite the slightly warmer summer air temperatures and longer ice-free season, modeled pond evaporation (for ponds 1 m mean water depth) did not increase significantly between 1953 and 2009 (Fig. 5). Overall changes in pond evaporation (open-water period totals) were dampened due to opposing trends observed in the early and late parts of the season. Evaporation rates increased in June (0.44 mm yr^{-1}, $p = 0.05$) but decreased in October (–0.06 mm yr^{-1}, $p = 0.01$). No statistical differences in evaporation rates were found in July, August, or September. The increased June

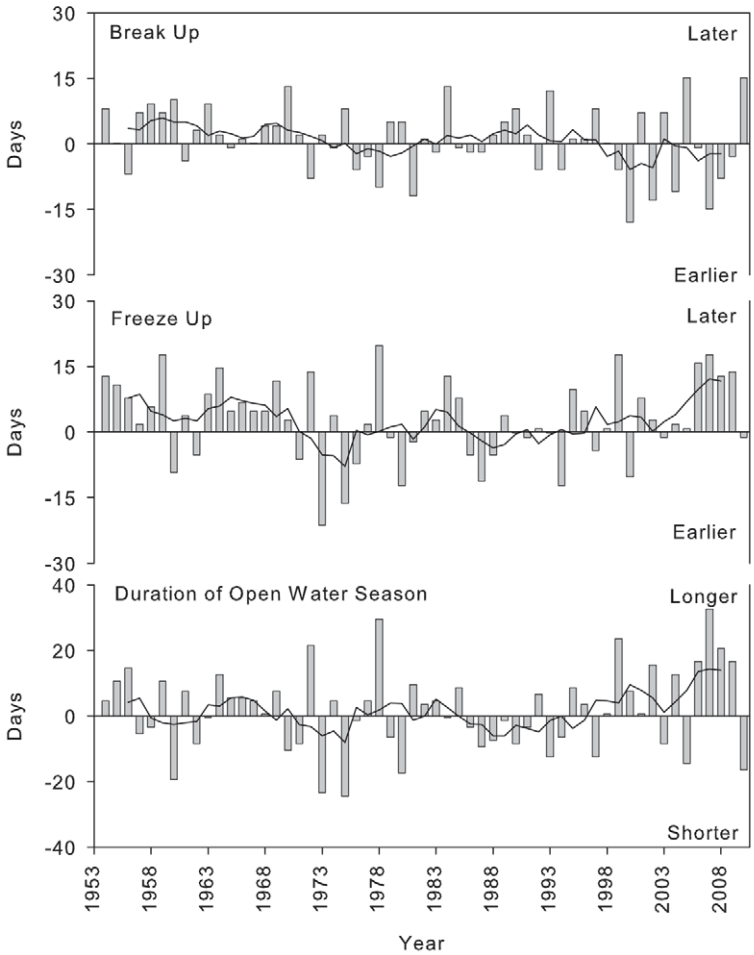

FIGURE 4. Normalized differences spring lake ice break-up and freeze-up dates, 1953-2009 (modeled using CLIMo model). Differences are anomalies from the 1971-2000 mean. The solid line represents a five-year running mean. Different scales are used on the y-axes.

evaporation rates likely result from the earlier breakup of ice cover and corresponding warmer air and pond water temperatures. The decreased evaporation rates in October may be a result of smaller vapor pressure gradients. Petrone et al. (2008) demonstrated that evaporation rates decrease when vapor pressure gradients are reduced during onshore winds from Hudson Bay, which are more persistent during the fall period.

TEMPORAL TRENDS IN THE WATER DEFICIT AND IMPLICATIONS FOR POND HYDROLOGIC STORAGE

The Churchill region has historically been an area that experiences a moisture deficit ($PPT < E$) during the post-snowmelt summer months (July–September) (Rouse, 1998; Macrae et al., 2004; Wolfe et al., 2011). However, the increases in summer precipitation are lessening the summer moisture deficit in the Churchill region (Fig. 6). The open-water period, although not significantly longer, appears to have shifted from experiencing a moisture deficit to one that experiences a surplus amount of moisture. This trend is not synchronous across the entire open-water period. The moisture deficit ($PPT - E$) did not change significantly in June or October (Fig. 6) from 1953 to 2009, whereas the moisture deficit became smaller in July (0.53 mm yr^{-1}, $p = 0.1$), August (0.79 mm yr^{-1}, $p = 0.05$) and September (0.79 mm yr^{-1}, $p = 0.05$) as a result of the increases in precipitation (Fig. 3). Thus, while the late summer and autumn months have historically been periods of time during which ponds continued to dry, these months are shifting to wetter conditions (Fig. 3). This trend is important for the regional hydrology and biogeochemistry, as ponds may fill with rainfall and possibly runoff from surrounding peatlands prior to freeze-up.

The seasonal pattern of the water deficit (WD, calculated by $PPT - E$) (7 day running means of 10-year periods, shown in Fig. 7, part a) changed in four ways over the study period: (1) the transition to a negative water balance (i.e., drying) occurred successively earlier in June; (2) the magnitude of the WD became smaller, both in the magnitude of the extent of drying observed, but also the duration of a negative water balance, where there was a much shorter period of drought and potential for pond drying and sediment exposure; (3) a mid to late summer wetting trend and return of the negative water balance to a positive water balance occurred progressively earlier in the mid to late summer, prior to freeze-up; and (4) there was more apparent surplus (i.e., a water balance that was more positive than at the timing of breakup). A change in the WD has significant implications for pond hydrologic storage. Previous research in this region (e.g., Macrae et al., 2004; Wolfe et al., 2011) and elsewhere (Woo et al., 1992; Hamilton et al., 1994; Quinton and Roulet, 1998; Quinton and Marsh, 1999; Woo and Guan, 2006) has shown that water levels in some ponds are largely balanced by the WD. Thus, changes in pond water levels (as a function of WD) over the same time period were estimated (Fig. 7, part b). Ponds

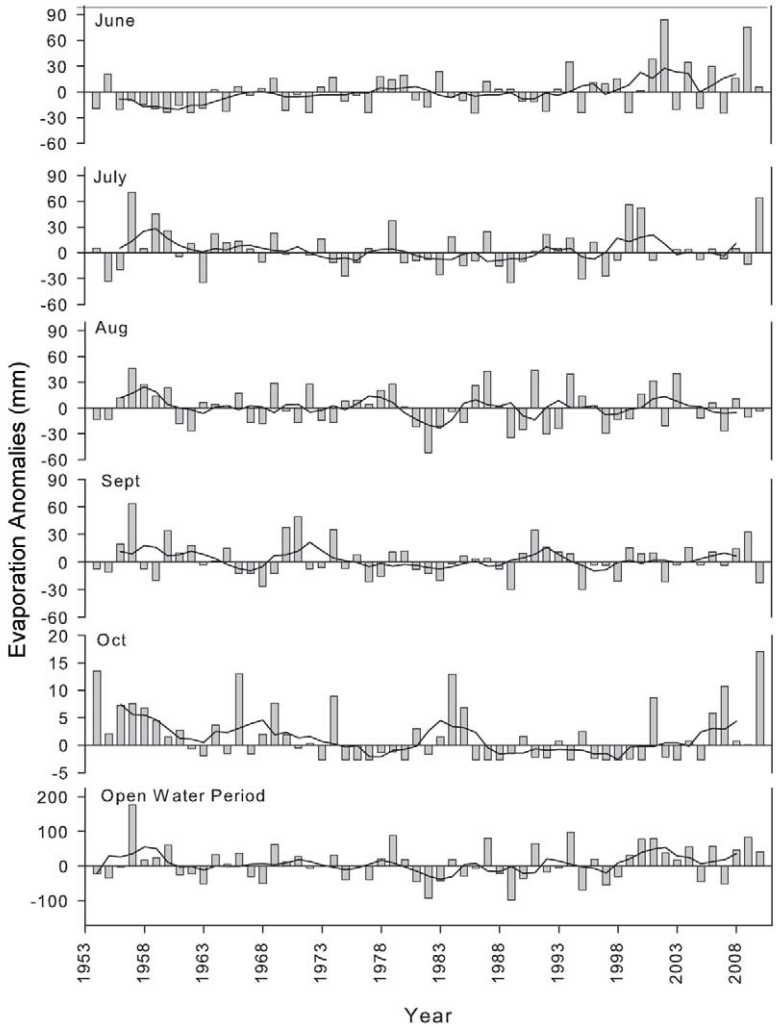

FIGURE 5. Normalized differences in modeled lake evaporation (1 m water column). Differences are anomalies from the 1971–2000 mean. The solid line represents a five-year running mean. Different scales are used on the y-axes.

were assumed to be at bankfull storage (capacity) at the time of spring breakup. A cumulative WD that exceeded this (i.e., more positive) was assumed to lead to overbank flooding and runoff (fill and spill) from the pond. It was assumed that there is minimal input to ponds from their surrounding catchments as was shown by Macrae et al. (2004) for the dry summer period. However, this assumption was only made to simplify the current discussion, as not all ponds are hydrologically disconnected (Wolfe et al., 2011), and most ponds are hydrologically connected to peatlands during wet periods when the storage capacity of the shallow active layer is exceeded (Macrae et al., 2004). An apparent increase in rainfall that leads to the filling of pond depression storage, and consequently generates runoff in the landscape, has important implications for pond-peatland connectivity (and general connectivity throughout the landscape) (Quinton and Roulet, 1998) as well as feedbacks to permafrost degradation that may be caused by an increase in latent energy in the saturated active layer (e.g., Jorgenson et al., 2006; Woo and Guan, 2006; Dyke and Sladen, 2010; Prowse and Brown, 2010). The many shallow ponds have historically been ephemeral, with sediment exposure/desiccation for extended periods in summer. The current results suggest that ponds (~1 m in depth) in this region are not at risk of drying and are instead at risk of transitioning from ephemeral systems to features that remain saturated throughout the open-water period. This has important implications for biogeochemical and ecological processes in both pond water and the thick layer of organic sediments present in ponds in the region. It is unclear what effects increased wetness will have on the waterfowl and other biota that rely on the ponds for habitat (Mallory et al., 2010; Bhiry et al., 2011). Alternatively, the increase in the availability of water to lakes and ponds may lead to a widespread drying of these features if it causes them to drain by enhancing permafrost degradation around the ponds, breaching pond banks, or creating subsurface connections beneath lakes (Woo et al., 1992).

CHANGES IN POND SURFACE AREA (1947–2008)

An analysis of 100 ponds using aerial imagery acquired in 1947 (dry), 1956 (dry), 1972 (wet), and 2008 (dry) shows that pond surface areas increased in wet years and decreased in dry years, but did not demonstrate a consistent directional change between 1947 and 2008 (Fig. 1). For example, the delineated ponds exhibited a 3% reduction in total surface area from 862,424 m^2 in 1947 to 837,516 m^2 in 1956. Pond surface areas increased in 1972 to a total of 862,142 m^2 for the 100 surveyed ponds (a change of <1%

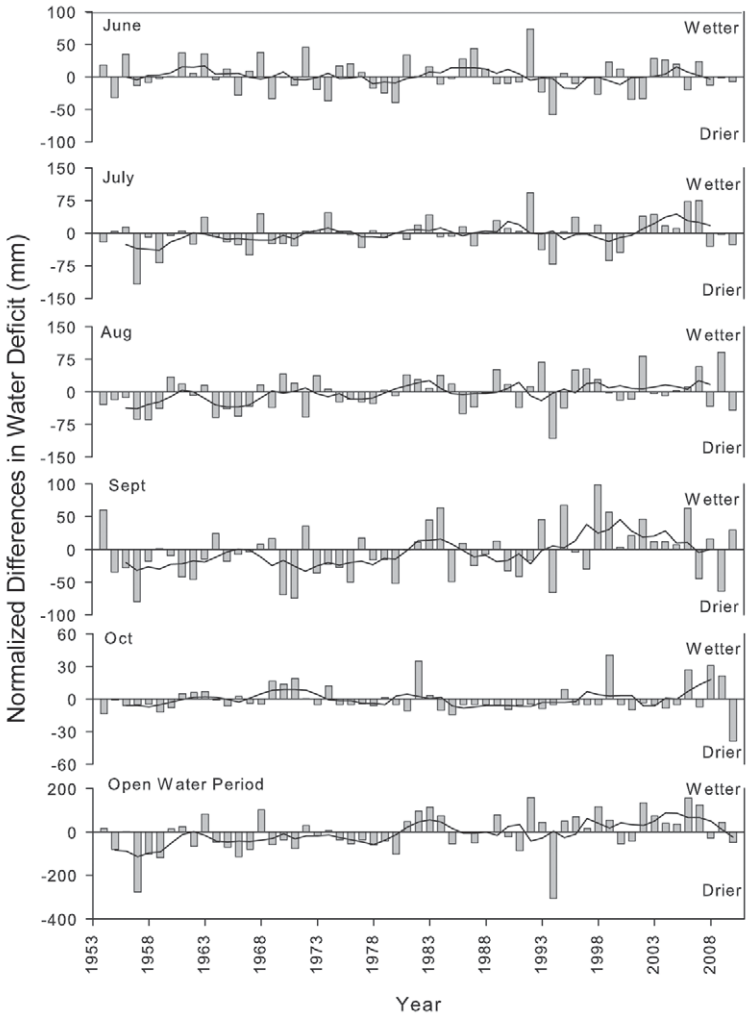

FIGURE 6. Normalized differences in the summer water deficit (difference between measured rainfall and modeled lake evaporation). Differences are anomalies from the 1971–2000 mean. The solid line represents a five-year running mean. Different scales are used on the y-axes.

from 1947 levels) but decreased again by 2008 to 773,611 m^2 (10% lower than 1947). Of the 100 ponds examined, 21 experienced consistent directional shoreline change between 1947 and 2008. An apparent shoreline retreat was observed for 20 features while consistent expansion was evident for one feature within the study site. Although directional shoreline changes were apparent, these changes were small and may have resulted from natural hydrologic variability in conjunction with human interactions or digitizing error. This apparent lack of change in pond surface area is in contrast to what has been observed elsewhere in the HBL (e.g., Payette et al., 2004), where pond expansion and coalescing was observed.

The failure of ponds to increase in size is in contrast to the increased summer rainfall amounts observed over the study period. We hypothesize that the images did not demonstrate change for several reasons. First, ponds in the Churchill region currently remain in a water deficit (negative water balance) over the summer months, despite the apparent increases in rainfall (Figs. 6 and 7). Second, due to the flat topography of the region, an exceedance of *PPT* over *E* will result in overbank flooding and runoff to adjacent depressions, and eventually to Hudson Bay. Finally, the images were captured in late July/early August (1947, 1972, 2008), with the exception of 1956 (August 28), which was a dry year. Ponds are typically at their maximum deficit at this time in the season (Figs. 6 and 7), whereas the increases in rainfall have typically occurred throughout August and September, after the images were captured. Thus, despite an apparent doubling of summer rainfall, the study ponds have not shown any shift in surface area, and thus the threshold moisture condition needed for pond growth and possibly the slumping and collapse of the surrounding peatland may not yet have been achieved. However, the increase in moisture availability in the region is expected to continue to increase based on the modeling work in this study. Thus, this region is at risk for pond expansion in the future.

PROJECTION OF FUTURE CHANGES FOR THE 21ST CENTURY

The increased temperature and precipitation observed over the past half century are projected to continue. Comparisons of air temperatures projected by the model for the 1961–2000 period with observations of air temperature at the Churchill Airport (Environment Canada, 2009) (Table 1) suggest that the model is performing well in all seasons of the year. Projections of precipitation are less certain, with the CRCM under-predicting precipitation (Table 1) for

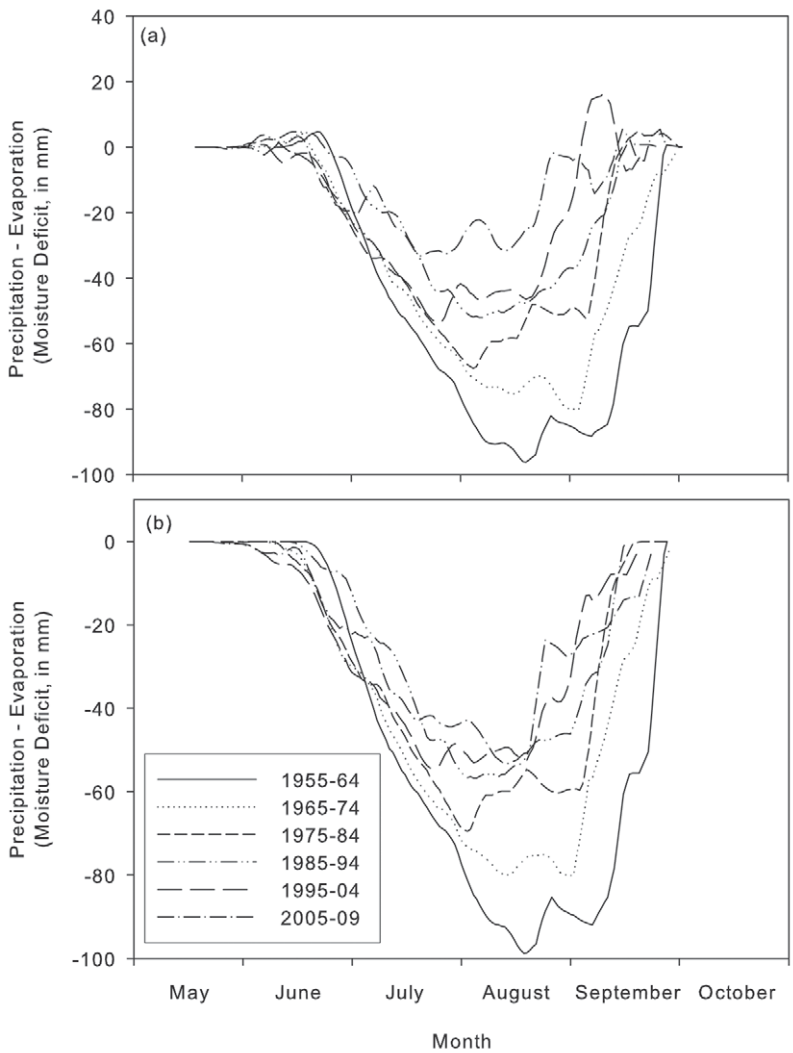

FIGURE 7. Atmospheric moisture deficit (observed precipitation – modeled evaporation) as an indication of pond water levels. Ponds are assumed to be full at break-up, although snowmelt and lateral seepage are not included in the model, and ponds are assumed to be hydrologically isolated from peatlands. The cumulative moisture deficit post–break-up is shown in (a). In (b), periods in summer and autumn during which the moisture deficit is more positive than at break-up are assumed to lead to overbank flooding (spill) from ponds, resetting the moisture deficit to zero. Values are 7-day running means for each day, averaged at 10-year intervals. Different scales are used on the y-axes.

the period between 1971 and 2000. Precipitation projections appear to be most uncertain for the spring (March, April, May—MAM) and autumn (September, October, November—SON) period (Table 1).

Due to the uncertainty of model output, 30-year means and standard deviations are provided for the projected data (Table 1). Mean annual temperatures are projected to rise from –6.6 ± 1.5 °C (1971–2000) to –3.5 ± 1.2 °C (2041–2070) and –1.4 ± 1.6 °C (2070–2100). This temperature increase will be a result of warmer winters, as mean December, January, February (DJF) temperatures are projected to rise from –24.2 ± 3.1 °C (1971–2000) to –19.1 ± 3.4 °C (2041–2070) and –17.7 ± 3.4 °C (2070–2100). The autumn period will also be warmer, as temperatures are projected to rise from –2.7 ± 1.8 °C (1971–2000) to –0.1 ± 1.8 °C (2041–2070) and +2.0 ± 1.6 °C (2070–2100). Minimal changes in spring (MAM) and summer (JJA) temperatures are projected (Table 1). Annual precipitation is also projected to increase from 1971–2000 levels by ~18% (2041–2070) and ~24% (2070–2100). These increases will occur across all seasons of the year but will be most significant in summer (JJA) (Table 1). This is in agreement with what has already been observed over the 66 years of observational data examined in this study and is in agreement with projections made by Gagnon and Gough (2005b).

The variables generated by the CRCM were used to force the CLIMo model to predict the duration of the ice-free period. The models predicted a longer ice-free season (127 days) than is currently experienced (115 days) for 1971–2000 (Table 2), suggesting that the projections should be treated with caution. The models project that the ice-free season will increase by an average of 18 days by 2041–2070 and 29 days by 2071–2100 (Table 2). The longer season is mainly a result of earlier breakup dates (JD 152 ± 10 days, 2041–2070; JD 144 ± 11 days, 2070–2100), but also results from later freeze-up dates (JD 297 ± 11 days, 2041–2070; JD 301 ± 8 days, 2071–2100) (Table 2). That is, later freeze-up (ice-on) implies delayed ice formation and thickening, resulting in thinner end-of-winter ice thickness (where thickness is not constrained by pond water depth), if all other climatic factors are not changed. However, if winter snowfall also increases, the ice growth rate could be further reduced. As a result, even if MAM temperatures do not change significantly, less energy will be required (i.e., fewer degree days in spring/early summer) to melt the ice cover, resulting in

TABLE 1

Mean (standard deviation) observed contemporary (Obs.) and projected (Proj.) changes in seasonal and annual mean air temperatures (T-air), mean annual total precipitation, and snowfall for the Churchill region for three time periods: 1971–2000, 2041–2070, and 2071–2100. Projections were generated using CRCM 4.2.0 data and observations were taken from the Churchill airport meteorological station.

		T-air (°C)		Precipitation (mm)		Snowfall (mm)	
Period		Proj.	Obs.	Proj.	Obs.	Proj.	Obs.
1971–2000	Annual	–6.6 (1.5)	–6.8 (1.3)	385 (94)	426 (88)	148 (65)	167 (47)
	DJF	–24.2 (3.1)	–24.6 (2.4)	48 (30)	51 (19)		
	MAM	–9.9 (3.1)	–9.9 (2)	44 (33)	64 (33)		
	JJA	10 (1.5)	10.1 (1.4)	166 (53)	167 (50)		
	SON	–2.7 (1.8)	–2.9 (1.9)	126 (59)	144 (53)		
2041–2070	Annual	–3.5 (1.2)		455 (117)		159 (54)	
	DJF	–19.1 (3.4)		68 (32)			
	MAM	–7.8 (3.3)		59 (46)			
	JJA	11.5 (1.8)		192 (78)			
	SON	–0.1 (1.8)		136 (60)			
2071–2100	Annual	–1.4 (1.6)		476 (113)		140 (33)	
	DJF	–17.7 (3.4)		66 (31)			
	MAM	–9.4 (2.9)		61 (37)			
	JJA	11.1 (1.6)		216 (77)			
	SON	2 (1.6)		134 (49)			

DJF = December, January, February.
MAM = March, April, May.
JJA = June, July, August.
SON = September, October, November.

earlier breakup (ice-off). Evaporation from a 1.0 m pond was also projected using the same approach outlined above (Oswald and Rouse, 2004). Projected rates of annual evaporation (254 ± 32 mm) were comparable to our modeled evaporation rates for the 1971–2000 period (245 ± 53 mm). Evaporation rates are projected to increase to 285 ± 32 mm (2041–2070) and 300 ± 31 mm (2071–2100).

The annual and summer water deficits (WD) were estimated using the projections of PPT and E. There is a large amount of uncertainty associated with these estimates, and they should therefore be treated with caution. The annual WD in the region is projected to decrease (i.e., a more positive water balance) by ~45 mm between the present and 2071–2100; however, this will not be caused by increases in snowfall and will instead be caused by increases in rainfall during the open-water period, which is projected to increase by ~49 mm (Table 2). A continued lessening of the WD, combined with permafrost degradation that is also projected to occur (Dyke and Sladen, 2010) is anticipated to lead to pond expansion and coalescence as has been observed in more southern areas of the HBL as a result of permafrost decay (Payette et al., 2004).

TABLE 2

Mean (standard deviation) contemporary (Est.) and projected (Proj.) changes in changes in pond break-up (BU) dates (Julian day), freeze-up (FU) dates, pond evaporation (E), and Annual and Open Water Season (OWS) moisture deficit (PPT-E). For projected BU, FU, and Evaporation, temperature, wind, and humidity projections were generated using CRCM 4.2.0. For the 1971–2000 period, BU and FU were estimated with CLIMO and evaporation from the method of Oswald and Rouse (2004), with Churchill airport meteorological station data as forcing variables.

	BU (JD)		FU (JD)		E (mm)		Annual PPT-E (mm)		OWS PPT-E (mm)	
	Proj.	Est.	Proj.	Est.	Proj.	Est.	Proj.	Est.	Proj.	Est.
1971–2000	164 (9)	171 (7)	291 (10)	286 (10)	254 (32)	245 (53)	131 (108)	108 (180)	–21 (75)	–38 (87)
2041–2070	152 (10)		297 (11)		286 (32)		169 (118)		–8 (106)	
2071–2100	144 (11)		301 (8)		300 (31)		176 (110)		28 (100)	

TABLE 3

Changes in surface areas of 100 ponds near Churchill, Manitoba (1947–2008).

Year	1947–1956	1956–1972	1972–2008	1947–2008
Mean Surface Area Change per Pond (m^2) (n = 100)	–249	+437	–688	–515
Mean and (Range) Surface Area Change (% of Individual Pond Surface Area, n = 100)	–3% (–79% to +323%)	+5% (–46% to +217%)	–8% (–50% to 39%)	–6% (–68% to +533%)
Total Change in Surface Area of All 100 Ponds (m^2)	–24,908	+24,626	–88,531	–88,813

Conclusions

This study has demonstrated a change in climatic conditions for the Churchill region of the HBL over a 66-year period. Air temperatures and the length of the growing season have both increased, leading to increases in modeled evaporation rates from ponds. However, these increased evaporation rates have been offset by increases in annual precipitation that are largely caused by a significant increase in rainfall in late summer/early autumn. The mechanism causing these increases in summer rainfall is unclear but should be investigated as the CRCM is not capturing the observed increases in its projections. Despite the increase in moisture availability in the region, the ponds still exhibit negative water balances in summer and consequently have not yet increased in surface area. Although CRCM projections of precipitation are less certain than those of temperature, the CRCM projects that increases in precipitation and evaporation will continue into the future. It is anticipated that if precipitation continues to increase at a faster rate than pond evaporation, ponds will begin to expand.

Our study did not examine runoff from peatlands into ponds. A reduction in the WD is anticipated to enhance permafrost degradation and increase the supply of runoff to ponds, which will reduce pond drying even more. Thus, future studies are needed that examine pond-peatland connectivity in response to both permafrost degradation and increases in rainfall in this region. Such studies should also be extended to include more inland thermokarst water bodies, which are more affected by the thaw of massive ice, as the geographic extent of our study was limited to coastal ponds, located within 10 km of Hudson Bay.

Acknowledgments

This project was funded and/or supported by the Canadian Foundation of Climate and Atmospheric Sciences (CFCAS grant to Macrae and Duguay), Churchill Northern Studies Centre Northern Research Fund, and the Northern Scientific Training Program (NSTP). Two anonymous reviewers are acknowledged for their constructive suggestions.

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MS accepted October 2013